Papers from the Third Gondwana Symposium
Canberra, Australia, 1973

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Gondwana Geology
Gondwana Geology

Papers presented at the Third Gondwana Symposium
Canberra, Australia, 1973

K. S. W. CAMPBELL, Editor

Australian National University Press
Preface

The Gondwana Symposia, held under the auspices of the International Union of Geological Sciences Sub-Commission for Gondwana Stratigraphy and Palaeontology, were designed to allow workers on the Gondwana System to meet, present the results of their researches, and plan future studies. The published proceedings of the first two symposia therefore show a preoccupation with the sediments, fossils and correlations of successions of Gondwana rocks on the various southern continents. There is a sprinkling of papers on broader topics such as the significance of the distribution of various kinds of organisms, the position of a continent within the Gondwanaland reassembly, or the difficulties of explaining the presence of contemporaneous ice caps over a super-continent the size of Gondwanaland. Various review papers published in the proceedings or as separate volumes, such as the I.U.G.S. reviews prepared for the First Symposium at Mar Del Plata and those published by the Brazilians for their own country at the same time, provide up to date accounts of regional geology.

Given this background data, and the enormous advances in the study of sea-floor spreading and associated topics, the organising committee of the Third Symposium felt that it should focus its efforts on an attempt to bring together studies of the Gondwana rocks on the various southern continents, and in their surrounding seas, in such a way that the stratigraphic problems could be seen in a wide palaeogeographic and tectonic framework, and various drift hypotheses could be tested. The largest section was therefore set aside for studies of this type; it was scheduled for the end of the symposium, and workers from each of the continents were asked to contribute. So that the widest coverage could be obtained, this section was expanded to include studies of tectonics, structural geology, igneous activity, and the geophysics of the continental margins.

Leading to this focal point were five other lines of study, each placed in its own section. These were to deal with specific problems in palaeogeography, palaeobotany, coal, glacial deposits, and advances in palaeontology and stratigraphy. Again, contributors were advised that these should be treated in a broad context, and that work which was of local significance only would not be accepted. The time interval to be covered was that of the Gondwana System and the immediately pre- and succeeding periods.

To cover these areas of study the following six sections were arranged, and the same groupings have been retained for the publication of the proceedings.

1. Palaeogeography
2. Gondwana Flora
### Preface

3. Environment and Origin of Gondwana Coal Deposits  
4. Age and Stratigraphical Relations of Glacial Deposits  
5. Advances in Stratigraphy and Palaeontology  
6. Tectonics, Igneous Activity, Geochronology, Structural Geology and Nature of the Continental Margins

In the event it was found impossible to achieve all the committee’s objectives. For some regions information is not yet adequate to produce syntheses of the type required, or no author was found who was willing to attempt the task. Contributors were continually confronted with the limitations of the available means of correlation, and for this reason relationships between events on the now dispersed elements of Gondwanaland were often not determinable. There was frequent conflict between workers on the palaeogeographic significance of a single piece of primary data. Some of the material presented did turn out to be of local significance only. And finally, some of the contributions were not prepared for publication, or were submitted in unpublishable form.

Cynics commonly aver that the main value of conferences is to indicate the areas where knowledge is weakest, thus providing prescriptions for further conferences. There is, of course, a grain of truth in this, and one benefit derived from the symposium was the setting up of several small working groups to define more clearly the difficult areas that had been identified, and to attempt to bring solutions to the next symposium. But this is far from being the whole story. A number of very significant advances have been recorded. We have fresh and stimulating reviews of what is known of the evolution of the Gondwana Basins of Antarctica, southern Africa, part of South America, and Australia set in their Gondwanaland contexts. There are attempts at reconstruction of the continental margins of large sectors of the Gondwana landmass. We have hopeful signs that a biostratigraphic zonation applicable to the whole Gondwana region, and based on both palynology and marine invertebrates, will be available in the not too distant future, and that it will be possible to tie this in to the world time scale. A clear pattern is beginning to emerge from the rather confused mass of information on the late Palaeozoic glaciation, and the over-simplified models of Gondwana coal formation have been filled out. All these are clear gain, and along with other advances are set out in these proceedings.

It is now clear, at least to me, that there is a great deal of useful information known to a number of workers in each of the continents, but generally not available to workers elsewhere, either because it has not been organised into a form suitable for publication, or because it is in the files of a company or government instrumentality. Further, there is little incentive to publish much of this because it cannot be incorporated into a hypothesis by any worker or group of national workers. The next step in Gondwana studies therefore seems obvious. After the problem areas have been identified and defined, national representatives should be appointed to glean all the information available from whatever source, and then it should be pooled at a small international meeting. From this a joint report should be prepared in which all sources of information should be acknowledged. Only after this has been done will we be ready for the next round of broad review-type studies. The remaining problems are too complex for any one person or any group of persons to tackle in isolation.
In preparing the work for publication I have had the invaluable assistance of the convenors of the various sections—Drs A. R. Crawford, B. E. Balme, C. T. McElroy, B. C. McKelvey, B. N. Runnegar, R. E. Wass and I. McDougall, and Professor H. J. Harrington. The Chairman and members of the Organising Committee, Drs J. M. Dickins, M. J. Rickard and M. W. McElhinny have lent considerable support. Professor D. A. Brown has given advice on numerous editorial problems and has attended to many time-consuming matters, such as handling correspondence during my absence from the A.N.U. Each of the papers has been read by at least one referee. Mr G. Harper has assisted with the numerous drafting problems. To all these fellow workers, named and anonymous, I extend my grateful thanks.

Canberra, 1974

K.S.W.C.
Organising Committee of Symposium

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          M. J. Rickard, Australian National University, Canberra

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Contributors to Publication of the Proceedings

Australian Academy of Science; Australian National University; Broken Hill
Proprietary Company Limited; Esso Australia Limited; Joint Coal Board, N.S.W.;
United States Steel International (New York), Inc.; Utah Development Company.
List of Contributors

S. K. Acharyya
Geological Survey of India, Calcutta, India

Ann M. Anderson
Bernard Price Institute, University of Witwatersrand, South Africa

Harlan P. Banks
Division of Biological Sciences, Cornell University, Ithaca, N.Y., U.S.A.

M. R. Banks
Department of Geology, University of Tasmania, Hobart, Tasmania, Australia

A. M. Barnaby
Department of Geology, University of Cape Town, Rondebosch, South Africa

Peter J. Barrett
Department of Geology, Victoria University of Wellington, New Zealand

T. H. Barry
South African Museum, Cape Town, South Africa

U. K. Basu
Geological Survey of India, Calcutta, India

A. J. R. Bennett
CSIRO Division of Mineralogy, North Ryde, New South Wales, Australia

H. J. Blignault
Department of Geology, University of Stellenbosch, South Africa

R. A. Britten
Joint Coal Board, Sydney, New South Wales, Australia

M. K. Roy Chowdhury
Geological Survey of India, Calcutta, India

M. J. Clarke
Geological Survey of Tasmania, Hobart, Tasmania, Australia

Jacqueline Conrad
C/o Centre de Recherches sur les Zones Arides, Centre National de la Recherche Scientifique, Paris, France

Campbell Craddock
Department of Geology and Geophysics, University of Wisconsin, Madison, Wisconsin, U.S.A.

J. C. Crowell
Department of Geological Sciences, University of California, Santa Barbara, California, U.S.A.

N. J. de Jersey
Geological Survey of Queensland, Brisbane, Queensland, Australia

T. Delevoryas
Department of Botany, University of Texas, Austin, Texas, U.S.A.

C. F. K. Diessel
Department of Geology, University of Newcastle, Newcastle, New South Wales, Australia

A. R. Drysdall
Geological Survey Department, Lusaka, Zambia

A. du Plessis
Department of Geology, University of Cape Town, Rondebosch, South Africa
<table>
<thead>
<tr>
<th>Name</th>
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</tr>
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<tbody>
<tr>
<td>David H. Elliot</td>
<td>Department of Geology and Mineralogy and Institute of Polar Studies, The Ohio State University, Columbus, Ohio, U.S.A.</td>
</tr>
<tr>
<td>P. R. Evans</td>
<td>Esso Australia Ltd, Sydney, New South Wales, Australia</td>
</tr>
<tr>
<td>L. A. Frakes</td>
<td>Department of Geology, Australian National University, Canberra, Australia</td>
</tr>
<tr>
<td>Jose Frutos</td>
<td>Instituto di Investigaciones Geológicas, Antofagasta, Chile</td>
</tr>
<tr>
<td>P. K. Ghosh</td>
<td>Coal Board and Coal Controller, Calcutta, India</td>
</tr>
<tr>
<td>S. G. Ghosh</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>R. E. Gould</td>
<td>Department of Geology, University of New England, Armidale, New South Wales, Australia</td>
</tr>
<tr>
<td>J. R. Griffiths</td>
<td>CSIRO Mineral Physics Division, North Ryde, New South Wales, Australia</td>
</tr>
<tr>
<td>G. E. Grikurov</td>
<td>Research Institute for the Geology of the Arctic, Leningrad, U.S.S.R.</td>
</tr>
<tr>
<td>Rafael Herbst</td>
<td>Secretaria de Estado de Cultura y Educacion, Universidad Nacional de Nordeste, Corrientes, Argentina</td>
</tr>
<tr>
<td>Elizabeth M. Kemp</td>
<td>Bureau of Mineral Resources, Canberra, Australia</td>
</tr>
<tr>
<td>Barry P. Kohn</td>
<td>Department of Geology, Victoria University of Wellington, New Zealand</td>
</tr>
<tr>
<td>Rosemary Kyle</td>
<td>Department of Geology, Victoria University of Wellington, New Zealand</td>
</tr>
<tr>
<td>William S. Lacey</td>
<td>School of Plant Biology, University College of North Wales, Bangor, Wales, U.K.</td>
</tr>
<tr>
<td>B. Laskar</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>B. G. Lopatin</td>
<td>Research Institute for the Geology of the Arctic, Leningrad, U.S.S.R.</td>
</tr>
<tr>
<td>Graham McClung</td>
<td>Geological Survey of Queensland, Brisbane, Queensland, Australia</td>
</tr>
<tr>
<td>Eurico Romulo Machado</td>
<td>Instituto de Geociencias, Universidad Federal do Rio Grande do Sul, Porto Alegre, Brazil</td>
</tr>
<tr>
<td>M.-Th. Mackowsky</td>
<td>Bergbau-Forschung GMBH, Forschungsinstitut des Steinkohlenbergbauvereins, Essen, West Germany</td>
</tr>
<tr>
<td>I. R. McLachlan</td>
<td>Southern Oil Exploration Corp., Johannesburg, South Africa</td>
</tr>
<tr>
<td>Henno Martin</td>
<td>Geologisches Institut, Göttingen, West Germany</td>
</tr>
<tr>
<td>N. D. Mitra</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>N. J. Money</td>
<td>Geological Survey Department, Lusaka, Zambia</td>
</tr>
<tr>
<td>Noreen Morris</td>
<td>Department of Geology, University of Newcastle, Newcastle, New South Wales, Australia</td>
</tr>
<tr>
<td>B. R. J. Rao</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>Contributor</td>
<td>Institution/Location</td>
</tr>
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<tr>
<td>C. N. Rao</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>Bruce Runnegar</td>
<td>Department of Geology, University of New England, Armidale, New South Wales, Australia</td>
</tr>
<tr>
<td>Izak C. Rust</td>
<td>University of Port Elizabeth, Port Elizabeth, South Africa</td>
</tr>
<tr>
<td>M. V. A. Sastry</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>R. A. Scrutton</td>
<td>Grant Institute of Geology, West Mains Road, Edinburgh, U.K.</td>
</tr>
<tr>
<td>S. C. Shah</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>M. Shibaoka</td>
<td>CSIRO Division of Mineralogy, North Ryde, New South Wales, Australia</td>
</tr>
<tr>
<td>E. S. W. Simpson</td>
<td>Department of Geology, University of Cape Town, Rondebosch, South Africa</td>
</tr>
<tr>
<td>Gopal Singh</td>
<td>Geological Survey of India, Calcutta, India</td>
</tr>
<tr>
<td>M. Smyth</td>
<td>Joint Coal Board, Sydney, New South Wales, Australia</td>
</tr>
<tr>
<td>K. R. Surange</td>
<td>Birbal Sahni Institute of Palaeobotany, Lucknow, India</td>
</tr>
<tr>
<td>F. L. Sutherland</td>
<td>Department of Mineralogy and Petrology, The Australian Museum, Sydney, New South Wales, Australia</td>
</tr>
<tr>
<td>Paul Tasch</td>
<td>Department of Geology, Wichita State University, Wichita, Kansas, U.S.A.</td>
</tr>
<tr>
<td>J. C. Theron</td>
<td>Department of Geology, University of Fort Hare, Alice, Cape Province, South Africa</td>
</tr>
<tr>
<td>J. N. Theron</td>
<td>Geological Survey, Stellenbosch, South Africa</td>
</tr>
<tr>
<td>Alvaro Tobar</td>
<td>Instituto de Investigaciones Geológicas, Santiago, Chile</td>
</tr>
<tr>
<td>Daniel Alberto Valencio</td>
<td>Departamento de Ciencias Geológicas, Universidad de Buenos Aires, Argentina</td>
</tr>
<tr>
<td>J. J. Veevers</td>
<td>School of Earth Sciences, Macquarie University, North Ryde, New South Wales, Australia</td>
</tr>
<tr>
<td>Michel Westphal</td>
<td>C/o Centre de Recherches sur les Zones Arides, Centre National de la Recherche Scientifique, Paris, France</td>
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* Deceased.
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Section 1

Palaeogeography
The Palaeomagnetism of South American Rocks and its Significance for the Fragmentation of Gondwana Land

DANIEL ALBERTO VALENcio

ABSTRACT

Recently published South American palaeomagnetic data are critically analysed. The Early-Middle Carboniferous to Early Cretaceous section of the South American polar wandering curve is now fairly well defined. It suggests an almost continuous polar shift relative to South America which spanned at least the time from the late Palaeozoic to the Early Cretaceous. This episode of drift would have been interrupted in the Permo-Triassic. As a consequence of this drifting episode South America moved away from the South Pole reaching latitudes very close to the present ones. The discrepancy between the positions of Kiaman geomagnetic poles of South America and Australia suggests that the initial fragmentation of Gondwanaland started in the Early Permian. The comparison of Mesozoic geomagnetic poles of South America and Africa suggests that the South Atlantic Ocean was open in the Early Cretaceous.

INTRODUCTION

Since the Second International Gondwana Symposium held in South Africa in 1970, new palaeomagnetic data of Palaeozoic, Mesozoic and Cainozoic rocks from South America have been published. At the same time radiometric dates have been obtained from rocks on which palaeomagnetic studies have been previously available. The significance of these data for the history of the fragmentation of Gondwanaland, and particularly for the age of separation of South America and Africa, are discussed in this paper.

NEW PALAEOMAGNETIC-RADIOMETRIC SOUTH AMERICAN DATA

Palaeozoic

Embleton (1970) has suggested that the former geomagnetic pole for Los Colorados red-beds of the Middle Section of the Paganzo Group (Creer et al., 1969) be replaced by two separate geomagnetic poles: one for the lower levels (SAP3, 59.5°S, 357.5°E) and the other one for the upper levels (SAPo, 74°S, 308°E). Thompson (1972) subsequently proposed new co-ordinates for the position of the geomagnetic pole SAP2 (74°S, 313°E, α95 = 4.5) and a geomagnetic pole (SAP3, 82°S, 349°E, α95 = 4°) for the levels of the Paganzo Group at Los Colorados higher than those used by Embleton. Valencio (in press) has suggested a Late Carboniferous age for the lower levels of the Middle Section of the Paganzo Group at Los Colorados (pole SAP3). The same author has indicated that the direction of the magnetic remanence of the upper levels of the Paganzo Group at Los Colorados (poles SAP2 and SAPo) suggests an age younger than late Westphalian but older than Middle Permian, and probably close to the Permian-Carboniferous boundary.

Magnetic remanences of normal and reversed polarity were used in computing the
polar position for the Middle Section of the Paganzo Group exposures in Río Chaschuil (SAP₉, 85°S, 64°E, $\alpha_{95} = 8^\circ$), (Thompson, 1972; Valencio, in press). This polar position could be affected by the non-dipolar components of the geomagnetic field. An age equivalent to that of the geomagnetic poles SAP₈ and SAP₉ has been assigned to this pole (Valencio, in press).

Thompson (1972) has also computed the polar position for the lower levels of the Middle Section of the Paganzo Group exposures in Paganzo Village (SAP₇, 81°S, 327°E, $\alpha_{95} = 4^\circ$); Valencio (in press) suggested an age equivalent to that of geomagnetic poles SAP₂ and SAP₃ for this pole.

On the basis of their magnetic remanences Valencio (in press) has suggested a probable Stephanian age for some red-beds of the Middle Section of the Paganzo Group from Huaco (SAP₄, 63°S, 356°E, $\alpha_{95} = 9.5^\circ$, Embleton, 1970) and from the northwestern Argentina (SAP₅ₐ, 65°S, 347°E, $\alpha_{95} = 6^\circ$, Creer, 1965).

Palaeomagnetic studies of igneous rocks of the Serie Porfiritica or Choioyilitense from Nihuil and Cuesta de los Terneros, Province of Córdoba, Argentina, were carried out by Valencio (1969) and Vilas (in press a); the rocks from the mafic and silicic facies of the Serie Porfiritica in Nihuil were subdivided into two formations: the Quebrada del Pimiento Formation and the Cerro Carrizalito Formation (Gonzalez Diaz, in press). These formations are included within the Cerro Carrizalito Group. Samples of the Quebrada del Pimiento Formation give a mean K-Ar age of 263 ± 5 m.y. (Valencio and Mitchell, 1972a, 1972b; Creer et al., 1971). This new evidence suggests a Middle Permian age for the Serie Porfiritica and this is therefore the age of the geomagnetic poles computed for this formation by Valencio (SAP₅, 81°S, 280°E, $d\psi = 7^\circ$ $d\chi = 5^\circ$) and Vilas (SAP₆, 80°S, 228°E, $d\psi = 11^\circ$ $d\chi = 9^\circ$).

Figure 1.1 shows that the geomagnetic poles for the Paganzo Group SAP₅, SAP₂ and SAP₃ and the geomagnetic pole for the Piaui Formation (SAC₅) form a Late Carboniferous age group. The mean of these four geomagnetic poles (SAC₅ₑu) is 63°S, 352°E, $\alpha_{95} = 4^\circ$.

The geomagnetic poles for the Paganzo red-beds SAP₂, SAP₇, SAP₅, and SAP₉ form a Permo-Carboniferous age group (Fig. 1.1); the mean of these four geomagnetic poles (SAPᵐ) is 82°S, 335°E, $\alpha_{95} = 9^\circ$.

The mean of the Serie Porfiritica or Choioyilitense Middle Permian geomagnetic poles SAP₅ and SAP₆ (SAPᵐ) is 82°S, 253°E, $\alpha_{95} = 19^\circ$.

These mean geomagnetic poles are shown in Figure 1.1 together with the Middle-Lower Carboniferous Taiguati (SAC₂, 38°S, 326°E) and the Upper Permian Las Tunas and Bonete (SAP₁, 78°S, 219°E) geomagnetic poles. The Late Palaeozoic section of the polar wandering curve for South America is also shown in Figure 1.1.

**Mesozoic**

The first Middle Jurassic geomagnetic pole for South America (84°S, 56°E, $d\psi = 6^\circ$ $d\chi = 7^\circ$, 161 m.y.) was computed on the basis of the cleaned magnetic remanences of igneous rocks of the Chon Aike Formation from
Puerto Deseado, Province of Santa Cruz, Argentina (Valencio and Vilas, 1970). Two years later, Creer et al. (1972) suggested that this is a virtual geomagnetic pole and computed a new geomagnetic pole for the Chon Aike Formation on the basis of the magnetic remanences of igneous rock exposures of this formation close to Camarones, Province of Chubut, Argentina (82°S, 231°E, $\alpha_{95} = 12^\circ$). More recently, Vilas (1973) carried out a palaeomagnetic study of the Chon Aike Formation exposures in the Estancia La Reconquista, Province of Santa Cruz, and computed a new geomagnetic pole for this formation using all the available palaeomagnetic data (Puerto Deseado, Camarones and Estancia La Reconquista). This geomagnetic pole SAJ$_1$ (85°S, 197°E, $\alpha_{95} = 6^\circ$) is considered as representative of the Middle Jurassic polar position for South America (see Fig. 1.2).

![Fig. 1.2. Mesozoic geomagnetic (circles) and mean geomagnetic poles (triangles) for South America. The solid curve is the Late Palaeozoic-Lower Cretaceous section of the polar wandering curve for South America.

Palaeomagnetic studies were recently carried out with Cretaceous South American rocks. McDonald and Opdyke (1972) computed a virtual geomagnetic pole for the Cretaceous La Teta lava, Colombia, and Opdyke and McDonald (1972) presented a virtual geomagnetic pole for the Late Cretaceous Pocos de Caldas Alkaline Complex, Brazil. These virtual geomagnetic poles do not allow us to improve our knowledge about the Cretaceous polar position for South America defined by the geomagnetic pole for the Lower Cretaceous Serra Geral Formation, Brazil (Creer, 1962) (SAK$_1$, 78°S, 54°E, $\alpha_{95} = 4^\circ$). Valencio (1972) computed a geomagnetic pole for the Lower Cretaceous Vulcanitas Cerro Colorado Formation of the Sierra de los Cóndores Group, Province of Córdoba, Argentina (SAK$_2$, 81°S, 14°E, $\alpha_{95} = 13^\circ$, 118.5 ± 6 m.y.), and Linares and Valencio (in press) presented a geomagnetic pole for the Cretaceous-Early Cainozoic trachybasaltic dykes from Río de los Molinos, Province of Córdoba, Argentina (SAK$_3$, 77°S, 16°E, $\alpha_{95} = 8^\circ$). The mean of these geomagnetic poles and the Serra Geral Formation pole (SAK) is: 79°S, 27°E, $\alpha_{95} = 8^\circ$; the author thinks that this mean pole defines with more confidence the Cretaceous polar position for South America. This mean Cretaceous pole is shown in Figure 1.2, together with the Middle Jurassic Chon Aike geomagnetic pole (SAJ$_1$) and the mean of the three South American Triassic and Permo-Triassic geomagnetic poles (SATr, 78.6°S, 231°E, Valencio and Vilas, 1972). The late Palaeozoic-Early Cretaceous section of the polar wander curve for South America is also shown in Figure 1.2.

Cainozoic

Mazoires et al. (1971) computed virtual geomagnetic poles for the Quixaba Formation (2.38 to 3.12 m.y.) and the Remedios Formation (8.02 to 10.67 m.y.) from the Fernando de Noronha Island, Brazil.

Fleck et al. (1972) carried out K-Ar and palaeomagnetic studies of Cainozoic basalt flows from Cerro del Fraile, Province of Santa Cruz, Argentina, obtaining eight virtual geomagnetic poles whose ages lie between 1.03 and 2.06 m.y.

Pacca and Valencio (1972) computed virtual geomagnetic poles for igneous rocks from the Abrolhos Islands, Brazil, whose ages are 46.6 to 52.4 m.y.
Figures 1.1 and 1.2 show that the polar wander curve for South America between the Early-Middle Carboniferous (SAC₂) and the Early Cretaceous (SAK) is fairly well defined. Within this interval five age groups of geomagnetic poles are defined for the Late Carboniferous (SACₚ), Permo-Carboniferous (SAPC), Middle Permian (SAPₚₘ), Triassic and Permo-Triassic (SATₚ) and Early Cretaceous (SAK). There is some doubt about the real position of SATₚ (Valencio and Vilas, 1972) and we must consider the possibility that the geomagnetic poles of this group (SATₚ₁, Las Cabras Formation; SATₚ₄, Giron Formation and SATₚ₅, Paganzo Group, Upper Section) form a 'time group' with the Late Permian pole for the Las Tunas and Bonete Formations (SAPₚ). Therefore the palaeomagnetic data suggest a gradual polar shift relative to South America from the Early-Middle Carboniferous to the Early Cretaceous, and that within this long interval there would have occurred a 'quasi-static interval' in Permo-Triassic times. During this overall period of time, South America moved away from the South Pole, and reached latitudes very close to the present ones in the Late Permian.

Valencio and Vilas (1972) have suggested that the late Palaeozoic (Early-Middle Carboniferous, SAC₂, to the Late Permian, SAPₚ₁) section of the polar wandering curve of South America (Fig. 1.1) also defines the polar shift relative to Africa and that during that interval South America and Africa were joined by their Atlantic littorals.

The Australian geomagnetic poles also define a polar shift relative to this continent in Late Carboniferous times (McElhinny and Briden, 1972); these authors have shown that this polar shift is common to the present continents that formerly constituted the Gondwanaland supercontinent. This Late Carboniferous shift is equivalent to that one here defined between the Early-Middle Carboniferous geomagnetic pole SAC₂ and the mean of the geomagnetic poles of the Late Carboniferous 'age group' of South America (SACₚ, Fig. 1.1). Also the last quoted authors have suggested a Late Palaeozoic polar shift for Australia between the Upper Carboniferous Paterson Toscanite geomagnetic pole and the 'time group' of Australian Kiaman geomagnetic poles. The authors suggested that this late Palaeozoic polar shift is common to all the Gondwanic continents.

The post-Late Carboniferous polar paths relative to South America and Australia are different. Figure 1.3 shows the discrepancy between these polar paths; in this figure the late Palaeozoic geomagnetic poles of South America and Australia have been drawn on a map of Gondwanaland reconstructed from palaeomagnetic data (Valencio et al., 1971). This discrepancy could imply that the rupture of Gondwana started in the Late Permian when two continental blocks were formed—South America-Africa and Australia-Antarctica-India.

Combination of palaeomagnetic results from African rocks whose ages range from 209 to 190 m.y., yields an estimate of the polar position for the Triassic of Africa (Valencio and Vilas, 1969). This polar position overlaps with that of the mean of the Triassic and Permo-Triassic South American geomagnetic poles (SATₚ) when rotated with South America to Africa so as to reconstitute Gondwanaland. This indicates that in the Triassic South America and Africa were joined by their Atlantic littorals.

By an analogous reasoning it is shown that the geomagnetic polar position defined on the basis of the Chon Aike Formation exposures at Puerto Deseado, Camarones and Estancia La Reconquista, Argentina (SAJ₁, Vilas, 1973) overlaps with that of the mean of the Jurassic (170 to 140 m.y.) African geomagnetic poles (Valencio and Vilas, 1969), when both continents and their polar positions are rotated so as to reconstitute Gondwanaland. This suggests that the South Atlantic was not formed in Middle Jurassic times.

Combination of palaeomagnetic results from the Mlanje Massif and Lupata Alkaline Volcanics (122 and 109 m.y., respectively), yields an estimate of the Early Cretaceous polar position of Africa (Briden, 1967). This differs from that of the mean of the Early Cretaceous South American geomagnetic poles (SAK) when rotated with South America to Africa so as to reconstitute the
The South American geomagnetic poles have been indicated by the same symbols as in Fig. 1.1. The Australian geomagnetic poles (diamonds) have been named following McElhinny and Briden, 1972: \( Au_{3-5} \) is the mean of Australian Middle Carboniferous poles; \( Au_4 \) is the geomagnetic pole for the Upper Carboniferous Paterson Toscanite and \( Au_{7-12} \) is the mean of the Kiaman geomagnetic poles.

Therefore, Antarctica would have moved away from Africa after the former fragmentation of Gondwanaland suggested by palaeomagnetic data; this is now supported by the Triassic tetrapod fauna found in these continents.

**ACKNOWLEDGMENTS**

The author wishes to thank the Universidad de Buenos Aires and the Consejo Nacional de Investigaciones Científicas y Técnicas of Argentina for the support which enabled the work described to be carried out.
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2 Palaeomagnetic Results from the Carboniferous and the Jurassic of the Saharan Platform

JACQUELINE CONRAD and MICHEL WESTPHAL

ABSTRACT

The palaeomagnetic study of Saharan Carboniferous clays and siltstones gives a virtual geomagnetic mean pole at 25°N, 235°E. Jurassic igneous and sedimentary rocks give a virtual geomagnetic mean pole at 72°N, 281°E. These results are in agreement with those previously obtained from Morocco and other parts of Gondwanaland.

INTRODUCTION

A study has been carried out in the northwest of the African Platform, between the South Moroccan (Anti-Atlas and the northern margin of the Tindouf Basin) and the coal basins to the northwest of the Hoggar (Fig. 2.1).

The coal-bearing series of the Central Sahara embraces marine and continental formations (Conrad, 1973), which terminate the deposits forming the Palaeozoic cover of the Precambrian massif of the Hoggar. These covering deposits are, on the whole, continuous from the Cambrian-Ordovician to the Carboniferous in spite of local unconformities, which sometimes result from reactivation of the shield, and sometimes from climatic or eustatic influences. A red continental formation, with marine incursions, ended this Palaeozoic sedimentation in the late Namurian. The cover was folded overall for the first time during a Hercynian orogeny of post-Namurian age.

Doleritic intrusions, sills and dykes, emplaced during a phase of post-Hercynian tension, dated as Jurassic by recent radiometric measurements, transect the entire Palaeozoic.

Two detrital continental formations lie unconformably on the Palaeozoic cover and are themselves mutually unconformable:

1. a lower formation referred to the Upper Jurassic, well displayed in the Reggan Basin and folded during an Eo-Cretaceous tectonic phase;
2. an upper formation of Early Cretaceous ('continental intercalation') age, dated by its fauna and flora, which lies almost undisturbed in the Central Sahara.

Other doleritic intrusions, also dated as Jurassic, mark the Ougarta zone (Touat dolerites) and the Bécher-Abadla Basin, and indicate continuation of the phase of tension between the Central Sahara, the northern flank of the Tindouf Basin, and the Moroccan Anti-Atlas (Hailwood and Mitchell, 1971; Conrad, 1972 (with bibliography); Bardon et al., 1973; Leblanc, 1973). These dykes are everywhere overlain by the Cretaceous.

INVESTIGATION OF PALAEOZOIC SAMPLES

Three oriented samples, some 15 m apart, have been selected from the same Tournaisian horizon at the top of the Khenig sands, on the eastern limb of the Azzez Matti syncline (Fig. 2.1), the zone separating the Reggan Basin from the Bled-el-Mass-Sebkha Mekerrhane antecline. They consist of a ferruginous oolite of marine facies with regressive features (JC-1729, A, B, C).

Two samples have been taken from the
Fig. 2.1. Outline geological map of north-west Africa, showing locations of samples studied. 1: Oum Oulili; 2: Guelb-el-Aouda; 3: Tazoult; 4: Ain-ech-Chebbi and Hassi Taibine regions; 5: Azzel Matti.

The continental Namurian of the Ain-ech-Chebbi Formation, on the eastern margin of the Reggan Basin. They consist of fine sandy red argillites, located 1 km southeast of Ain-ech-Chebbi (JC-1727) and 30 km south of Hassi-Taibine (JC-1654).

**PALAEOMAGNETIC RESULTS**

(Fig. 2.2A)

We have measured the natural magnetisation of these samples and have analysed it during demagnetisation of the rocks by progressive heating, in a zero field or in an alternating magnetic field.

The directions obtained from the Tournaisian samples are strongly dispersed and two of them have a magnetisation whose direction, after tectonic correction, is close to the present Earth’s field.

<table>
<thead>
<tr>
<th>Samples</th>
<th>Localities</th>
<th>Direction of magnetisation after demagnetisation</th>
</tr>
</thead>
<tbody>
<tr>
<td>JC-1729 A</td>
<td>Azzel Matti (25.4°N; 0.5°E)</td>
<td>decl. incl. decl. incl.</td>
</tr>
<tr>
<td>JC-1729 B</td>
<td>98° +35° 105° +47°</td>
<td></td>
</tr>
<tr>
<td>JC-1729 C</td>
<td>336° +45° 347° +51°</td>
<td></td>
</tr>
<tr>
<td>JC-1654</td>
<td>135° +27° 135° +27°</td>
<td></td>
</tr>
<tr>
<td>JC-1727</td>
<td>Ain-ech-Chebbi (26.5°N; 0.3°E)</td>
<td>111° — 2° 137° +52°</td>
</tr>
</tbody>
</table>
The mean pole obtained for the two Namurian samples lies at 25°N and 235°E. The mean poles obtained by different authors for the end of the African Palaeozoic are (Creer, 1970):

- Lower Carboniferous: 27°N 207°E
- Permo-Carboniferous: 40°N 244°E
- Permian: 27°N 269°E

Our result is comparable with those of the Carboniferous and beginning of the Permian (Fig. 2.3).

INVESTIGATION OF JURASSIC SAMPLES

We have made a similar study of samples from Jurassic dolerite veins and sills (3 sites, 7 separate samples):

- JC-1725 A, B: sill in lower Namurian at Tazoult (Touat).

We have also studied a fine, sandy, red argillite sample (JC-1726), from the continental Upper Jurassic, 1 km southeast of Ain-ech-Chebbi.

RESULTS AFTER DEMAGNETISATION

(Fig. 2.2B)

The mean pole for the Early Jurassic is (circle 1 of Figure 2.3):

\[ N = 4, \quad K = 18, \quad \alpha_{95} = 22° \]
\[ \lambda = 72°N, \quad \phi = 261°E \]

<table>
<thead>
<tr>
<th>Samples</th>
<th>Localities</th>
<th>N</th>
<th>D</th>
<th>I</th>
<th>k</th>
<th>\alpha_{95}</th>
<th>\lambda</th>
<th>\phi</th>
<th>age</th>
</tr>
</thead>
<tbody>
<tr>
<td>JC-1400</td>
<td>Oum Oulili (29.2°N; 6.6°W)</td>
<td>3</td>
<td>4°</td>
<td>33°</td>
<td>71</td>
<td>8°</td>
<td>78°N</td>
<td>154°E</td>
<td>180 m.y.</td>
</tr>
<tr>
<td>JC-1644</td>
<td>Ain-ech-Chebbi (26.5°N; 0.5°E)</td>
<td>2</td>
<td>342°</td>
<td>40°</td>
<td>(700)</td>
<td>(10°)</td>
<td>74°N</td>
<td>260°E</td>
<td>166 m.y.</td>
</tr>
<tr>
<td>JC-1725</td>
<td>Tazoult (27.2°N; 0.2°W)</td>
<td>2</td>
<td>319°</td>
<td>36°</td>
<td>(370)</td>
<td>(13°)</td>
<td>52°N</td>
<td>267°E</td>
<td>188 m.y.</td>
</tr>
<tr>
<td>JC-1726</td>
<td>Ain-ech-Chebbi (26.5°N; 0.3°E)</td>
<td>1</td>
<td>334°</td>
<td>3°</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>Upper Jurassic</td>
</tr>
</tbody>
</table>
These results are comparable with those for the same epoch in Africa:

**Pole**

  - Samples shown in Fig. 2.2B
  - 67°N 283°E (1 site, 3 samples).
- Morocco (circle 2): dolerites
  - 71°N 216°E (27 sites, 160 samples).

Morocco (circles 3, 4): Foum Zguid, dolerites
- 65°N 236°E (16 sites, 96 samples).
- 58°N 259°E (5 sites, 27 samples).

Sierra Leone (circle 5): dolerites
- 81°N 223°E (10 sites, 50 samples).

South and East Africa:

- Trias (circle 6):
  - 65°N 245°E
- Jurassic (circle 7):
  - 68°N 256°E

Means calculated by K. M. Creer.

**Fig. 2.3. Virtual geomagnetic poles. Triangles:**

- Palæozoic: 1: present study; 2: Lower Carboniferous.

**Circles:**

- Jurassic: 1: present study; 2: Morocco (Bardon *et al.*); 3, 4: Morocco (Hailwood and Mitchell); 5: Sierra Leone (Briden *et al.*); 6: South Africa, Trias (Creer); 7: South Africa, Jurassic (Creer).
CONCLUSION

The sedimentary samples have given significant results for ferruginous, argillaceous, and continental siltstones, in which hematization is original. On the other hand, in the ferruginous sands of marine origin, such as those from the Tournaissian of Azzel Matti, it appears as though the ferruginous minerals have been subjected to later reactions in the deposit since the magnetization directions found are generally close to those of the present Earth’s field.

The similarity between the poles obtained and those of distant formations (Morocco, Sahara, Sierra Leone, South and East Africa) enables us to compare the magnetic field to that produced by a geocentric dipole for both the end of the Palaeozoic and the Early Jurassic.

This paper was translated from the French by D. A. Brown.

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3 Changing Sediment Transport Directions from Devonian to Triassic in the Beacon Super-Group of South Victoria Land, Antarctica

PETER J. BARRETT AND BARRY P. KOHN

ABSTRACT

The Beacon Super-Group in south Victoria Land comprises the Taylor Group (Devonian and possibly older), 1100 m mainly of quartzose sandstone, and the Victoria Group (Permian and Triassic), 1000 m of more feldspathic and carbonaceous alluvial plain sediments. Orientations of about 1200 large-scale cross-beds from an outcrop belt of Beacon strata 200 km long have given 268 palaeocurrent directions. Palaeocurrent directions from equivalent strata in the central Transantarctic Mountains are compared with the south Victoria Land data to show the regional palaeogeography.

Most of the Taylor Group was deposited on a northeast-sloping floodplain with the sediment derived from an area of the East Antarctic craton west of the present Ross Ice Shelf. However, the basal New Mountain Sandstone was deposited mainly from west-flowing currents, probably in a shallow marine environment. New Mountain deposition ended with major uplift to the north, where in places the New Mountain Sandstone was completely stripped away, and grit and conglomerate of the overlying Altar Mountain Formation rest directly on basement. The sediments become finer southwards, and the palaeocurrent data suggest deposition from low sinuosity streams flowing to the southwest. The upper part of the Altar Mountain Formation is transitional in lithology and palaeocurrent direction to the thick quartzose sandstone sequence, comprising the Arena Sandstone and the Beacon Heights Orthoquartzite, which were deposited by braided streams flowing northeast. The stream pattern changed to meandering with the deposition of the youngest formation for the Taylor Group, the Aztec Siltstone, but the palaeoslope remained unchanged.

No regional flow direction could be established for the ice that deposited the Metschel Tillite at the base of the Victoria Group, but a drainage divide for the lower part of the overlying Permian Weller Coal Measures has been located at 78°52'S. North of this the palaeocurrent direction is northwestward, but to the south it is southward. The southward palaeoslope continues for 1300 km at least as far as the Ohio Range. The Weller Coal Measures appear to be mainly outwash material—fine to coarse sandstone with only a few percent of shale in the lower and middle parts. However, the measured variability in palaeocurrent direction is high, in contrast to the low sinuosity of modern outwash streams. The high variability is attributed to control of stream direction by the ice front and by the hummocky topography exposed as the ice margin retreated.

A major change in palaeoslope took place between deposition of the lower part of
the Feather Conglomerate, which trends west-northwest like the underlying Weller Coal Measures, and deposition of the upper part, which trends north. This change is correlated with an even larger shift in a palaeocurrent direction in the central Transantarctic Mountains from southeast in the Permian to northwest in the Early Triassic. As deposition proceeded in the Triassic Lashly Formation of south Victoria Land the palaeoslope gradually returned to the northwest direction that characterised Permian deposition.

Taylor Group sediments in both south Victoria Land and the central Transantarctic Mountains were derived from continental areas of low relief on both sides of the present Transantarctic Mountains. Victoria Group sediments in south Victoria Land were eroded from a less mature cratonic highland to the east, whereas sediments of similar lithology and age to the south came from the northwest and southeast. These data suggest that in Late Palaeozoic and Early Mesozoic time the Ross Ice Shelf area was similar in its geologic character to the present East Antarctic craton.

INTRODUCTION

The Beacon Super-Group is a relatively thin sedimentary sequence exposed on the rim of the East Antarctic craton. The sequence is most completely developed in the central Transantarctic Mountains and in southern Victoria Land (Fig. 3.1), where it has been subdivided into two groups (Table 3.1). The older Taylor Group is mainly well-sorted quartzose sandstone deposited on a gently undulating surface cut in the granitic and metamorphic basement. In south Victoria Land most of this sediment was derived from the southwest. By contrast, the Victoria Group is mainly an alluvial plain sequence of quartzose and feldspathic sandstone and shale derived mainly from the southeast, in the area of the present Ross Ice Shelf. Palaeocurrent data are presented for each formation of the Beacon Super-Group in south Victoria Land. Regional palaeocurrent directions are then compared with similar measurements in other areas to provide a synthesis of sediment transport directions in the Transantarctic Mountains from Devonian to Triassic times.

PALAEOCURRENT MEASUREMENTS

Palaeocurrent measurements were made mainly to determine sediment transport direction, and hence infer source direction, for each formation in the Beacon Super-Group of south Victoria Land. Most measurements were made on trough cross-beds 10 to 50 cm thick, but small-scale cross-beds, ripple marks and parting lineation were also measured (Table 3.2). Measurements by Matz (1968) have been included with our own data as they are well located stratigraphically. To average out errors in measurement and local variation in current flow we aimed to measure at least three or four cross-beds at each station. However, we have accepted single measurements of well-exposed cross-beds at 15 per cent of the stations, where no other cross-beds were available.

Our interpretation is based mainly on the large-scale cross-beds because they were the most commonly measured and because they acquire their orientation from relatively strong currents. Other structures have been kept separate because they formed under different flow conditions. Large-scale trough cross-beds form from migrating dunes in the upper part of the lower flow regime, and thus represent flood stage currents on a floodplain, or currents with equivalent power in marine situations. Coleman (1969: 229) has shown that such cross-beds are normally oriented within a few degrees of the flood stage current direction. Most of the palaeocurrent directions are separated by tens of metres stratigraphically (equivalent to hundreds or thousands of years), and by kilometres geographically. Thus they are regarded as a sample taken without significant
<table>
<thead>
<tr>
<th>Age</th>
<th>Rock Unit</th>
<th>Description</th>
<th>Max. Thickness (Metres)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Triassic</td>
<td>Lashly FM</td>
<td>Member D, Member C, Member B, Member A</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sandstone like Member B. Interbedded fine sandstone, siltstone and thin coal beds. <em>Dicroidium</em>.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Massive feldspatic and lithic sandstone with a few logs and stems.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Greyish grey fine sandstone, siltstone and claystone in fining-upwards cycles. Abundant roots and a few calamitid stems.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Feather Conglomerate</td>
<td>Upper part</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Quartzoic grit, sandstone and siltstone in fining-upwards cycles.</td>
<td>100</td>
</tr>
<tr>
<td></td>
<td>Feather Conglomerate</td>
<td>Lower part</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Quartzose sandston and grit with abundant white vein quartz pebbles south of Taylor Glacier. Mainly sandstone to north.</td>
<td>150</td>
</tr>
<tr>
<td>Permian</td>
<td>Weller Coal Measures</td>
<td>Quarztose and feldspatic sandstone and minor carbonaceous shale. Pebbles and boulders scattered and in lenses especially near base. Coal, logs and stems and <em>Glossopterys</em>.</td>
<td>250</td>
</tr>
<tr>
<td></td>
<td>Metschel Tillite</td>
<td>Pyramind Erosion Surface. Tillite, with scattered mainly granitic clasts up to 1.4 m across. Minor sandstone, siltstone, conglomerate.</td>
<td>70</td>
</tr>
<tr>
<td>Mid-Upper</td>
<td>Aztec Siltstone</td>
<td>Maya Erosion Surface. Greyish-red and greyish-green siltstone and light coloured sandstone. Fish fossils, plant roots, mud-cracks and ripple marks common. Rare plant stems.</td>
<td>220</td>
</tr>
<tr>
<td>Devonian</td>
<td>Beacon Heights Orthoquartzite</td>
<td>Orthoquartzite, with occasional quartz grit lenses. Rare lycod stems. <em>Beaconites</em> trails throughout.</td>
<td>220</td>
</tr>
<tr>
<td></td>
<td>Arena Sandstone</td>
<td>Yellowish-grey sandstone. Ferruginous layers, burrows and trails (including <em>Beaconites</em>) common.</td>
<td>360</td>
</tr>
<tr>
<td>Devonian</td>
<td>Altar Mountain FM</td>
<td>Sandstone, siltstone. Burrows and trails (including <em>Beaconites</em>).</td>
<td>240</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Arkose grit and conglomerate.</td>
<td>(50)</td>
</tr>
<tr>
<td>Older (?)</td>
<td>New Mountain Sandstone</td>
<td>Heimdall Erosion Surface. Quartzose sandstone, minor siltstone.</td>
<td>270</td>
</tr>
<tr>
<td></td>
<td>Terra Cotta Siltstone</td>
<td>Shaly siltstone, minor sandstone</td>
<td>(60)</td>
</tr>
<tr>
<td></td>
<td>Boreas</td>
<td>Argillite, sandstone, conglomerate</td>
<td>(20)</td>
</tr>
<tr>
<td></td>
<td>Windy Gully Sandstone</td>
<td>Pebble quartzose sandstone</td>
<td>(50)</td>
</tr>
<tr>
<td></td>
<td>Subgreywacke</td>
<td>Kukri Surface</td>
<td></td>
</tr>
<tr>
<td>Ordovician</td>
<td>Basement Complex</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Precambrian</td>
<td></td>
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</tbody>
</table>
bias from a vast population of flow directions, and are used to estimate both average flow direction and variability in direction of the depositing palaeocurrent system. Provided no regional variation in palaeocurrent direction is evident, interpretation of the floodplain systems in the Beacon Super-Group is based on the expectation that large standard deviations indicate high sinuosity streams. Within-channel variability has been greatly reduced by averaging sets of cross-bed measurements so that the main factor causing variability in the palaeocurrent directions should be changes in channel direction.

Each arrow on each palaeocurrent map (Figs. 3.2 to 3.9) represents the vector mean of all similar palaeocurrent directions at a locality. Student's t-test ($P_{0.05}$) was used to decide whether differences between groups of directions were significant. All single directions more than $40^\circ$ apart were considered significantly different and are plotted as separate arrows.
Table 3.2. Source and Numbers of Palaeocurrent Directions (Vector Means of Sets of Palaeocurrent Measurements) used in this Paper

<table>
<thead>
<tr>
<th></th>
<th>This paper</th>
<th>Other</th>
<th>Matz (1968)</th>
<th>Large-scale</th>
<th>Other</th>
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<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>cross-beds</td>
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<td></td>
<td>structures</td>
<td></td>
<td>structures</td>
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</tr>
<tr>
<td>Victoria Group</td>
<td>81</td>
<td>41</td>
<td>59</td>
<td>22</td>
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<tr>
<td>Taylor Group</td>
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<td>32</td>
<td>21</td>
<td>9</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>196</td>
<td>73</td>
<td>80</td>
<td>31</td>
<td></td>
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</tbody>
</table>

Total directions from large-scale cross-beds = 276
Total measurements of large-scale cross-beds = 1298

Fig. 3.2 Palaeocurrent map of the New Mountain Sandstone (Devonian)
The Taylor Group ranges from post-Ordovician to Devonian in age (Webb in Barrett et al., 1972). It has been regarded by most writers (Vialov, 1962; Shaw, 1962; Harrington, 1965; Matz, 1968; Gevers et al., 1971) as mainly a nearshore marine sequence, chiefly on account of the lack of non-marine fossils and the abundance of burrows and trails, but no body fossils have been found beneath the Aztec Siltstone. The evidence now available suggests to us that most of the sequence above the Heimdall Erosion Surface (Table 3.1) is non-marine and accumulated under floodplain conditions. Mud-cracks are now known from every formation above the Heimdall surface (though they are abundant only in the Altar Mountain Formation and Aztec Siltstone), indicating that the surface of deposition was exposed to the air many times. The formation and preservation of the red-beds in the Altar Mountain Formation and Aztec Siltstone requires a non-reducing, and hence non-marine, environment in a region and slow accumulation (Dunbar and Rodgers, 1957: 212). The most direct evidence for environment of deposition exists for the Aztec Siltstone (Table 3.3).

Palaeocurrent measurements indicate that the Arena Sandstone and the Beacon Heights Orthoquartzite were deposited by a current system similar to, but less variable than, the meandering floodplain system that deposited the Aztec Siltstone. The few *Haplostigma* impressions collected from the Beacon Heights Orthoquartzite show that plant stems several centimetres across were deposited and subsequently oxidised away. Burial and oxidation of such material is more consistent with a floodplain than a shallow marine environment of deposition. Nevertheless, parts of the New Mountain Sandstone may be of shallow marine origin. Shell fragments of brachiopods or pelecypods have been reported from the basal Windy Gully sandstone member (Angino and Owen, 1962), and the large persistent cross-beds in the overlying New Mountain Sandstone may be submarine sand waves.

New Mountain Sandstone

The Beacon strata of south Victoria Land everywhere rest on a fresh surface (Kukri Peneplain of Gunn and Warren, 1962) cut in granitic and metasedimentary basement rocks. The surface has only slight local relief in most places, but up to 30 m has been noted at Windy Gully (Zeller et al., 1961). South of the Taylor Glacier the surface is overlain by the Windy Gully Sandstone and Terra Cotta Siltstone members (Zeller et al., 1961) of the New Mountain Sandstone (Table 3.1), and around Mount Boreas by the Boreas Subgreywacke Member (Webb, 1963). No directional sedimentary structures have been described from these beds.

The New Mountain Sandstone proper is a uniform fine- to medium-grained quartzose sandstone. Mostly it is indistinctly bedded and massive, but at several levels at each locality there are huge well-developed flaggy cross-beds from 0.5 to 2.5 m thick. The bedding planes in section (Plate 3.1) are generally long gently curved arcs that can be traced for several metres down dip to their tangential base. Foreset dips normally range up to 15°. Where they can be seen in plan each set has the form of a large trough cross-bed 5 to 10 m across. The uppermost foreset beds in several sets appear to be oversteep-

<table>
<thead>
<tr>
<th>Abundant mud-cracks</th>
<th>McKelvey and Webb, 1959; Webb, 1963</th>
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</thead>
<tbody>
<tr>
<td>Abundant red siltstone and claystone</td>
<td>Harrington, 1965; McElroy, 1969; Barrett et al., 1971</td>
</tr>
<tr>
<td>Soil features, such as root beds, vein networks, etc.</td>
<td>Barrett et al., 1971</td>
</tr>
<tr>
<td>Conchostracans and freshwater fish</td>
<td>Gunn and Warren, 1962; White, 1968; Barrett et al., 1971</td>
</tr>
<tr>
<td>High strontium 87/86 ratios</td>
<td>Faure and Barrett, 1973</td>
</tr>
</tbody>
</table>
Plate 3.I Large trough cross-beds in the New Mountain Sandstone.
A. Three large sets at Knobhead, on southern side of Taylor Glacier. The circled figure indicates scale.

B. Folded and locally oversteepened foreset beds at West Beacon (B2, Fig. 3.2), indicating subaqueous rather than aeolian deposition.
Plate 3.11 Large trough cross-beds typical of flood-plain deposits in the Beacon Super-Group. A. Cross-bedded grit at the base of the Odin Arkose on the south side of Wright Valley. Current flow was perpendicular to the face of the outcrop. B. Trough cross-beds on a sandstone platform in the Aztec Mountain (AM, Fig. 3.6). Current direction was parallel to the trough axis and to the upper right.
ened and contorted (Plate 3.1), due to pene-contemporaneous slumping, indicating that the cross-beds are subaqueous, not aeolian, despite their size. Most of the orientations indicate a remarkably consistent transport direction to the northwest over a distance of 100 km (Fig. 3.2). However, at the thick section at Mount Jason (J), transport direction alternated northwest and northeast in the lower 150 m, and changed to south for the upper 100 m. The character of the cross-beds and of the current system is unlike that of any of the floodplain sequences in the Beacon Super-Group (cf. Plates 3.1 and 3.11).

The structures may have formed from migration of large sand waves in a shallow sea.  

**Altar Mountain Formation**

The lower 10 to 50 m of the Altar Mountain Formation (the Odin Arkose Member) are mainly feldspathic sandstone and grit with abundant trough cross-beds 10 to 20 cm thick (Plate 3.1I). In the north the Odin Arkose rests on a well-defined surface, the Heimdal Erosion Surface (McKelvey et al., 1970) with a maximum relief of 16 m over 60 m (Mount Boreas, BS), and the pockets of New Mountain Sandstone appear as rem-
nants preserved in hollows in the older Kukri Surface. The basal beds form a 20 m thick conglomerate of quartz, quartzite and granite up to 20 cm across at Mount Suess (SU), but this thins rapidly southwards. South of the Taylor Glacier the pebbles are restricted to lenses a few centimetres thick. The pebbles are all less than 3 cm across, and are almost entirely vein quartz. The Odin Arkose also becomes finer-grained southwards, and is very similar in lithology to the underlying New Mountain Sandstone. Reworking has so diluted the new detritus of the Odin Arkose south of Beacon Heights that it has not yet been possible to identify the Heimdall Erosion Surface clearly. This trend in decreasing grain and pebble size and the cross-bedding suggests that the Odin Arkose was deposited from low sinuosity streams flowing southwestward in response to uplift in the area around Mount Suess (Fig. 3.3).

In the north the Odin Arkose is succeeded by yellowish-grey ferruginous sandstone that is difficult to distinguish from the overlying Arena Sandstone. Scattered trough cross-beds, mostly from 10 to 50 cm thick, are variable in direction (Fig. 3.3), but there is a trend to the northwest, perhaps transitional.

Fig. 3.4 Palaeocurrent map of the Arena Sandstone (Devonian)
between the strong southwest direction for the Odin Arkose and the more persistent northeast trend for the upper part of the Taylor Group. The upper part of the Altar Mountain Formation south of Beacon Heights consists mainly of thin beds of burrowed red and green siltstone with mudcracks, indicating a non-marine, though possibly coastal, environment of deposition. The thin beds of quartzose sandstone increase in thickness upwards to constitute the overlying Arena Sandstone, which has only a few thin beds of greenish siltstone.

**Arena Sandstone and Beacon Heights Orthoquartzite**

Throughout the area there is a gradation upwards from the yellowish-grey friable sandstone of the Altar Mountain Formation and the Arena Sandstone (which are difficult to distinguish north of the Taylor Glacier) into the pure white bluff-forming sandstone of the Beacon Heights Orthoquartzite. The change is from a clay matrix in the former to a quartz cement in the latter. Both formations maintain their character.
from the Mackay Glacier south for 180 km, though the boundary is commonly hard to pick because of its transitional nature.

Both formations have the same well-defined northeast palaeocurrent direction for the 60 km over which measurements were made (Figs. 3.4 and 3.5). The trend must extend as far as the southern Boomerang Range 100 km further south, where cross-beds in the Beacon Heights Orthoquartzite indicate an easterly palaeocurrent flow also. Both formations are here interpreted as alluvial plain deposits (though for a marine interpretation see Vialov, 1962, and Gevers et al., 1971), for the mean directions of the unimodal palaeocurrent patterns lie within a few degrees of that of the overlying Aztec Siltstone, which is demonstrably an alluvial

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**Aztec Siltstone**

- **PARTING LINES A FREE DIRECTION**
- **RIPPLE DIRECTION**
- **LARGE CROSS BEDS**
- **SMALL CROSS BEDS**

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**Statistics for large cross beds**

<table>
<thead>
<tr>
<th>RIPPLE MARKS</th>
<th>LARGE CROSS BEDS</th>
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<tbody>
<tr>
<td>X = 089°</td>
<td>N = 11</td>
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<tr>
<td>S = 45°</td>
<td>n = 38</td>
</tr>
<tr>
<td>C1, 52 = 37°</td>
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**Fig. 3.6 Palaeocurrent map of the Aztec Siltstone (Mid-Upper Devonian)**
plain deposit (Table 3.3). In this context the small standard deviations for the Arena Sandstone and Beacon Heights Orthoquartzite (40° and 41°) indicate deposition from low sinuosity streams whose average flow direction varied little. According to Schumm (1963) low sinuosity in streams is inconsistent with the large proportion of clay (as much as 20 per cent in the Arena Sandstone) though in this case the clay may have a post-depositional origin, like the quartz cement in the formation above. The low sinuosity and lack of fines in at least the Beacon Heights Orthoquartzite suggest that the streams were braided.

**Aztec Siltstone**

The Beacon Heights Orthoquartzite grades up over a few metres into the Aztec Siltstone, a sequence of massive greyish-red and greenish-grey siltstone with interbedded channel sandstone. The palaeocurrent trend continues northeast, though the variability in cross-bed orientations increases markedly. However, we do not infer a change in floodplain gradient. Schumm (1972) has shown that hydraulic parameters of channels are largely determined by type of sediment load and by flood discharge. Sinuosity increases with increasing wash load (silt and clay) and decreasing flood discharge. The marked change in sinuosity took place about the time that plants appear in the stratigraphic record in the Taylor Group. Plant stems have been found at several localities in the Aztec Siltstone, and a single stem has been found 60 m below the top of the Beacon Heights Orthoquartzite (Askin et al., 1971). We suggest that development of a plant-covered land surface and the formation of the first humic soils increased production of fines and decreased runoff and level of flood discharge sufficiently to have effected the change in the stream pattern from braided to meandering.

**Victoria Group**

The Victoria Group in south Victoria Land ranges in age from Early Permian (Barrett and Kyle, this volume) at least to late Middle Triassic (Helby and McElroy, 1969). Coal, root and plant beds, stems and logs in the sandstones, and the absence of marine fossils all point to floodplain deposition. The sequence (Table 3.1) begins with glacial beds which rest on the Maya Erosion Surface cut in Aztec Siltstone. After the ice sheet retreated rivers reworked a great deal of the glacial debris, and in a number of places exhumed and cut through the Maya Erosion Surface, forming the younger Pyramid Erosion Surface (McKelvey et al., 1970). Subsequently a sequence of Permian coal measures, non-carbonaceous sandstone and conglomerate, and Triassic coal measures was deposited without any obvious physical break. The Weller Coal Measures are entirely Permian on the basis of the abundant Glossopteris leaves that occur in lenses to within 12 m of the overlying Feather Conglomerate. The latter has yielded no datable fossils to the writers, but Pinet (1969) has reported glossopterid plants from beds equivalent to the upper part of the Feather Conglomerate and Member A of the Lashly Formation. However, the presence of Dicroidium odontopteroides 16 m above the base of the Lashly Formation at Portal Mountain (Barrett et al., 1971) convinces us that the lower Lashly at least is Middle or Upper Triassic. The upper Lashly Formation is late Middle Triassic on palynological grounds (Helby and McElroy, 1969). Victoria Group sedimentation in south Victoria Land was ended by a period of erosion prior to the eruption of doleritic breccias of the Jurassic Mawson Formation, and intrusion and extrusion of tholeiitic dolerite and basalt.

**Metschel Tillite**

Southern Victoria Land was one of the main late Palaeozoic ice centres for Gondwanaland, with the ice sheet spreading southward along the Transantarctic Mountains and northward onto Australia and Tasmania (Lindsay, 1970; Crowell and Frakes, 1971). Many indicators of ice flow direction have been found in glacial beds to the south (Long, 1965; Frakes et al., 1966, 1971;
Lindsay, 1970), but only three features that clearly indicate the trend of ice movement have so far been found in south Victoria Land. Striations on a well-exposed glacial pavement at nearby Mount Metschel strike at $120^\circ$ (Barrett and Kyle, this volume). An esker at Mt Ritchie (A4) has a similar strike ($150^\circ$). By contrast, a steep-walled valley at Alligator Peak, cut by ice and filled with slumped glacial debris, trends northeast (Barrett, 1971). More data are needed to give a clear indication of regional ice flow direction.

The northwest palaeoslope for the overlying coal measures in south Victoria Land (Matz and Hayes, 1966), and the southeast

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**Weller Coal Measures**

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![Diagram of Weller Coal Measures](image)

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Fig. 3.7 Palaeocurrent map of the Weller Coal Measures (Permian)
Changing Sediment Transport Directions

palaeoslope for equivalent beds in the Beardmore Glacier area (Barrett, 1970), led Lindsay (1970) to believe that there was a 'Permian drainage divide in south Victoria Land [which] . . . in turn suggests that the centre for ice dispersal for the late Palaeozoic ice sheet may also have been in this area'. Our palaeocurrent data for the Weller Coal Measures allow this divide to be located more precisely (Fig. 3.7) just north of latitude 78°. However, the granitic clasts in the tillite show that basement was exposed in the source area, which was therefore beyond the present extent of the Taylor Group, that is, either north of the Mawson Glacier (Fig. 3.1, 76°20'S) or possibly to the east or west of the present Transantarctic Mountains.

Weller Coal Measures

The Weller Coal Measures were deposited on an alluvial plain with abundant vegetation. Plant debris occurs as coal beds up to 7 m thick, as coal streaks and stems in the sandstone, and as finely comminuted plant debris in the mudstones. The sediments are mainly fine- to coarse-grained sandstones with finer sediments forming only 5 to 30 per cent of the sequence, mainly in the upper 50 m. However, scattered carbonaceous shale and siltstone lenses, representing ephemeral ponds, are scattered throughout the formation, and commonly yield glossopterid leaves. Erosion surfaces are common, and well-defined 'fining upwards' cycles occur in the upper part of the formation.

The formation has a thin discontinuous pebble bed at the base. The clasts are mainly granitic, as in the underlying glacial beds, but only quartz clasts remain after several metres. The lower part of the formation also includes locally coarse arkosic sandstone with highly weathered feldspar. The altered feldspar and the crumbly nature of the granite boulders indicates the substantial effect of post-glacial weathering on the debris that was reworked to form the lower Weller Coal Measures.

Sandstones in the middle of the formation also have pebbles and boulders at some localities. The small pebbles are almost all vein quartz, but the cobbles and boulders include quartz sandstone, acid volcanics, granite, schist and vein quartz. This variety of lithologies may have resulted from reworking of glacial debris, but contrasts with the granite-dominated tillite beneath. It is more likely to have come from rapid post-glacial erosion of a complex basement terrain.

The palaeocurrent directions from the Weller Coal Measures considered as a whole are spread around the compass, but significant trends appear if geographic position is considered (Fig. 3.7). For the lower 80 m the average palaeocurrent direction is northwest for localities north of 77°51'S, but is south for the southern localities. The difference is significant (P<0.01), but the 'watershed' so defined is difficult to interpret, because the strata in the watershed area are as thick or thicker than those to the north and south. The thickness data suggest alluvial fan deposition, but the palaeocurrent data indicate a reasonably distinct drainage divide. In either event the main source of Weller sediment was to the east, beneath the present Ross Ice Shelf.

The northwest trend continued through the Weller Coal Measures in the north, but outcrops do not extend far enough south to locate a drainage divide in the study area. Nevertheless equivalent strata in the Darwin Mountains, only 150 km south, have a southeast palaeocurrent trend, placing a southern limit on the position of the divide.

The variability in palaeocurrent direction for the Weller Coal Measures is as large as that from deposits of modern high sinuosity streams. However, the sediment is coarse, even conglomeratic in places, the proportion of siltstone and shale is small, and most modern outwash plains or streams from any rapidly eroding area are built up by low sinuosity braided streams. Control of stream direction by the ice margin, and by an irregular hummocky topography exposed by the retreating ice sheet, must have contributed substantially to palaeocurrent variability. However, we do not yet have sufficient data to compare channel sinuosity with these variations in palaeocurrent direction over a larger area.
**Feather Conglomerate**

The Feather Conglomerate (McElroy, 1969) is a massive non-carbonaceous pebbly sandstone unit that overlies the Weller Coal Measures. It has an abundance of small rounded white vein quartz pebbles in the type area at Mount Feather 10 km east of TI (Fig. 3.8), but the pebbles decrease rapidly in abundance north of the Taylor Glacier so that over a large part of the outcrop area the formation is almost entirely fine to very coarse quartzo-feldspathic sandstone.

Between the Taylor and Mackay Glaciers the upper part of the formation comprises a series of 'fining upwards' cycles, from coarse to fine sandstone, and locally siltstone. These beds were named the Fleming Formation by McKelvey et al. (1970), for at Mt Fleming they are distinctly different from the pebbly sandstone of the Feather Conglomerate below and the fine sandstone and carbonaceous siltstone of the Lashly Formation above. McElroy's 'Feather Conglomerate' for the entire interval is retained here because the fine beds that characterise the 'Fleming Formation' disappear southward and the unit can-
not be reliably distinguished in the field from the beds below. However, the base of the 'Fleming Formation' marks an important palaeocurrent boundary, and has been used to divide the Feather Conglomerate into an upper and a lower part (Fig. 3.8). Where the 'Fleming Formation' could not be recognised, the boundary was taken to be the base of the upper 80 m of the Feather Conglomerate.

The palaeocurrent data (Fig. 3.8) for both lower and upper parts of the formation give a strong unimodal pattern, with no significant areal change in direction. The standard deviations are similar and remarkably small (33° and 35°), indicating deposition from low sinuosity, probably braided, streams. However, the mean palaeocurrent directions differ by 90°. The lower part of the formation was deposited from streams flowing west-northwest, in a similar direction to the underlying Weller Coal Measures. The upper part of the formation was deposited after a dramatic shift in palaeoslope to the north-northeast, and this new direction continued into the overlying Lashly Formation.

LASHLY FORMATION

The transition upwards into the lowest
beds of the Lashly Formation (Member A) is marked by a sudden decrease in grain size. The fining upwards cycles in the Lashly Formation comprise thin fine-grained sandstone beds followed by much thicker carbonaceous siltstone and claystone containing root beds and calamitid stems, suggesting a swampy environment crossed by small meandering streams. The overlying 450 m (Members B to D) has much thicker sandstone beds with coalified stems, but from about 150 to 220 m above the base of the formation plant-bearing siltstones and thin coal beds are common (Member C).

The palaeocurrent data indicate deposition from streams of moderate sinuosity flowing to the northwest. There is no significant difference in average palaeocurrent direction between adjacent stratigraphic intervals \((P_{0.10})\), that is, between the Feather and lower Lashly or between the lower and upper Lashly Formations. However, the difference between Feather and upper Lashly Formations is significant \((P_{0.01})\), suggesting that the gradual swing in palaeoslope back to the northwest is real.

**REGIONAL CONSIDERATIONS**

The Taylor Group is more or less continuously exposed for 800 km from the Mawson Glacier to just south of the Beardmore Glacier (Fig. 3.1). The palaeocurrent data (Fig. 3.10) show that the bulk of the sediment filling this basin came from the southwest in south Victoria Land and from the west fur-
ther south, that is, from the East Antarctic craton west of the Ross Ice Shelf. However, some of the quartzose sediment was brought in from the east, and the similarity in lithology of the sediments from both directions, that is, supermature quartzose sandstones, suggests that a similar long-standing terrain of low relief existed on both sides of the present Transantarctic Mountains during the early Palaeozoic. The present Ross Ice Shelf must then have been underlain by a stable platform like the East Antarctic craton.

A notable event during deposition of the Taylor Group of south Victoria Land was the uplift in the north, recorded by the Heimdall Erosion Surface and the overlying Odin Arkose with its short-lived southwest palaeoslope. This probably resulted from the mid-Palaeozoic folding in the Borchgrevink Orogen (Craddock, 1970) which extends across northern Victoria Land, 400 km north, and Marie Byrd Land 1000 km west. The time of folding must predate the flat-lying Gallipoli Porphyries of northern Victoria Land (375 m.y., Faure and Gair, 1970), and is indicated by radiometric ages of 410-420 m.y. in phyllites (Gair et al., 1970; Ravich and Krylov, 1964).

The palaeoslope for the Permian strata of the Transantarctic Mountains and the presence of a drainage divide had been established previously (Matz and Hayes, 1966; Barrett et al., 1967), but the location of the divide is now better defined (Fig. 3.7). South of the divide the floodplain extended for at least 1300 km down the palaeoslope to beyond the Ohio Range. The floodplain north of the divide was probably only 200 km across, being restricted by the topographic high north of the Mawson Glacier. This high, which extended for several hundred kilometres into northern Victoria Land, may have existed throughout late Palaeozoic and early Mesozoic times, for the oldest sediments on the basement high are of Late Triassic age (Norris, 1965).

The 90° shift in palaeoslope between the lower and upper parts of the Feather Conglomerate is believed to be related to the Late Permian or Early Triassic reversal of palaeocurrent direction in the Beardmore Glacier area (Barrett, 1970). In that area the Permian sediments form a sequence of lacustrine shale, massive fine-grained sandstone deposited from braided streams and a thick floodplain sequence dominated by carbonaceous shale. The coal measures are much finer grained than those in south Victoria Land, and there is no conglomeratic unit like the Feather Conglomerate. In contrast, the units above the change in palaeoslope are quite comparable. The cycles of the upper Feather Conglomerate, beginning with coarse quartzose sandstone, are similar to the cycles of the lower Fremouw Formation. The Lashly Formation, with cyclic sandstone and root-bearing siltstone and claystone followed by thick sandstone with logs and stems, compares closely with the middle and upper parts of the Fremouw Formation (Barrett, 1969: 51-2).

An important consequence of correlating the marked changes in palaeoslope is that the boundary between the lower and upper parts of the Feather Conglomerate may include the Permian-Triassic boundary, for the equivalent surface in the Beardmore Glacier area is underlain by Glossopteris-bearing coal measures and is overlain by Lystrosaurus-bearing strata (Elliot et al., 1970; Kitching et al., 1972). Vertebrates have been looked for in the upper Feather and Lashly Formations, but have not yet been found.

The reversal in palaeoslope in the Beardmore Glacier area probably resulted from the Weddell Orogeny (Ford, 1972), which folded Permian and older strata in the Ellsworth and Pensacola Mountains. An upper age limit for the orogeny is provided by a Middle Jurassic age for the unfolded Dufek intrusion. The Pensacola Mountains contain primary phosphate beds and rederived phosphate pebbles in Devonian (?) quartzose sandstone (Schmidt et al., 1965; J. B. Cathcart, pers. comm.). Phosphate pebbles also occur in quartzose sandstone in Lower Triassic beds 2000 km down the palaeoslope in the Beardmore Glacier area (Barrett, 1969), but at no other level in the sequence, supporting the hypothesis that reversal resulted from uplift in the Pensacola Mountains in the Late Permian or Early Triassic. If this interpretation is correct the Triassic floodplain extended at least 1800 km from near
the Pensacola Mountains at least to south Victoria Land.

The transport direction of the post-glacial sediment in south Victoria Land for most of the Permian and Triassic was to the northwest and the dominant source area is inferred to be beneath the present Ross Ice Shelf. The mainly quartzose and feldspathic sediments had a continental source (Matz, 1968: 91), and support the inference made earlier from Taylor Group sediments that in the past the Ross Ice Shelf area was similar in its geologic character to that of the East Antarctic craton.

ACKNOWLEDGMENTS

We wish to thank fellow members of Victoria University of Wellington Antarctic Expeditions 15 and 16 (1970-71 and 1971-72) for willing assistance in the field and for subsequent discussions. Dr D. H. Elliot, Ohio State University, kindly commented on the manuscript. We also thank the University Grants Committee and the Victoria University Council for financial assistance, Antarctic Division, D.S.I.R. for logistic support, and the U.S. Navy for air transport to and within Antarctica. Mr E. Hardy draughted the diagrams.

REFERENCES


Structural and Palaeogeographical Evidence for an Upper Palaeozoic Sea between Southern Africa and South America

HENNO MARTIN

ABSTRACT

The Atlantic margin of southern Africa largely coincides with the Late Precambrian-Cambrian Pan-African fold belt, considerable parts of which must also underlie the shelf and the continental slope. The metamorphic grades indicate great uplift and profound denudation during the early Palaeozoic. By the late Palaeozoic, a westward directed drainage system had been established across the metamorphic backbone of the fold belt. This drainage, which was remodelled by the Permo-Carboniferous ice sheet, has been largely exhumed today. It seems unlikely that the erosion products reached South America. They may have been trapped in fault troughs which are today buried underneath the thick sediment wedge of the outer shelf and continental slope. The existence of Palaeozoic basins along the Pan-African fold belt may have greatly influenced the later rifting of Gondwanaland.

During the Mesozoic the westward directed drainage was blocked, probably by up-tilting of the rift shoulders. This led to an aggradation of sediments and basalt flows which overstepped and buried all the older topography. A comparison with the young Red Sea and Rhine rifts makes it probable that the rift shoulders were up-tilted to elevations of 1000 to 2000 m, and were initially supported by cushions of low density upper mantle material. The crustal thickness of the uplifted shoulders was then reduced by erosion. As the active spreading ridge moved away and the upper mantle cushion cooled the thinned crust of the former shoulders may have subsided to a considerable depth below sea level to be buried underneath the Upper Cretaceous to Recent shelf and continental rise sediments.

Three important conclusions are drawn: (1) Gondwanaland may not have been a single continental plate, but was probably subdivided by marine basins during the Upper Palaeozoic. (2) The present-day 1000-m bathymetric contours may, in spite of their good fits, deviate considerably from the lines of Mesozoic rifting. (3) The complex history of the continental margin was repeatedly controlled by vertical uplifts and subsidences caused by upper mantle processes.

INTRODUCTION

The break-up of Gondwanaland is usually discussed without due consideration of the resultant continental margins. This can be ascribed to the fact that large portions of these margins were, and still are, only superficially known. As knowledge increases the details of the geology and the geomorphology of the marginal belts can throw light on numerous questions concerning the original shape of Gondwanaland and the history of its break-up. Regarding the shape of Gond-
Fig. 4.1. Palaeozoic palaeogeographical features of the Atlantic margin of southern Africa. A comparison with Fig. 4.3 shows that the palaeozoic valleys cross the most deeply eroded part of the Pan-African fold belt. The positions of the sections of Fig. 4.2 are indicated.
wanaland before the break-up two conflicting possibilities have to be considered: Was Gondwanaland a single plate as assumed in the classical reassemblies and as suggested by the good fit of the 500 fathom depth contour; or was Gondwanaland physiographically more diversified, and were the present continental areas partly separated by epicontinental seas or even small oceanic basins? It is the purpose of this paper to discuss first the geology and the palaeogeography of the Atlantic margin of southwestern Africa with respect to these questions and then to analyse the influence of the Mesozoic rifting on the geology and the physiography of the continental margin.

THE LOWER PALAEOZOIC PALAEOGEOGRAPHY OF NORTHWESTERN SOUTH WEST AFRICA

It has been known for some time that the northwestern margin of South West Africa (see Figs. 4.1 and 4.2) is drained by river systems that were glacially shaped in the late Palaeozoic (Martin, 1953, 1968). These deep and broad valleys must have been incised by pre-glacial rivers on a westward sloping land surface. It is certain that a great volume of sediment must have left the present-day continental area even before the onset of the Permo-Carboniferous glaciation by way of this drainage system. Where has this material been deposited? And when was the westward directed drainage established? I shall first discuss the latter question.

The continental margin of Africa between the Cape of Good Hope and Gabon forms a part of the Late Precambrian-Cambrian Pan-African fold belt (Cahen and Snelling, 1966). Its folded and metamorphosed rocks partly either hug the present-day continental area even before the onset of the Permo-Carboniferous glaciation by way of this drainage system. Where has this material been deposited? And when was the westward directed drainage established? I shall first discuss the latter question.

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The isotopic ages of the metamorphic rocks indicate a metamorphic peak 520 to 500 m.y. ago, i.e. during the Late Cambrian (Nicolaysen, 1962). This should also be the time of the onset of the main phase of uplift and denudation of the fold belt. Sediments produced by this denudation should therefore have Cambrian to Silurian ages. Is it possible to estimate the amount of this early Palaeozoic denudation? A fair estimate can be made for the coastal belt between lat. 23° and 17°S, which is also the area of the well-preserved Palaeozoic valleys (see Fig. 4.3). There, rocks of the higher amphibolite grade and migmatites occur at the present coast and for variable distances inland (Martin, 1965). The p-t conditions of critical metamorphic mineral parageneses show that these rocks were formed at depths of 8 to 15 km minimum (Puhan and Hoffer, pers. comm.).

It is important to stress that the highest metamorphism, and this means the greatest early Palaeozoic uplift, occurred in the present-day coastal margin, for this fact leads to the unavoidable conclusion that the belt of elevated metamorphism and subsequent uplift and denudation must originally have extended westwards, probably at least as far as the present-day shelf and continental slope. This conclusion makes the question of the site of deposition of this enormous volume of early Palaeozoic denudation products still more important. The following considerations seem pertinent.

A mountain belt—a denudation of 8 to 15 km implies a former mountain belt—would have shed sediments towards both the east and the west. It seems possible, though not yet proved by isotopic or palaeontological dating, that the Ovambo Basin (see Fig. 4.3) may have received a large portion of the eastward directed erosion products. Underneath its blanket of Tertiary Kalahari Beds this basin contains a Karroo sequence, with tillites and interbedded varves at its base, which is unconformably underlain by some 2000 m of unmetamorphosed siltstones, fine-grained sandstones, shales and thin interbeds of stromatolitic limestones. This sequence has been marginally affected by the last phase of the Damaran (Pan-African) folding. It is still an open question whether these sediments are an epicontinental facies of the geosynclinal Damara Super-Group or whether they represent erosion products of the Damaran fold belt. A similar uncertainty exists with regard to the undated sediments of the Nama Basin, which is situated to the south of the transcontinental branch of the Pan-African fold belt. There is a possibility that the Fish River beds, the uppermost unit of the Nama Super-Group, may have been derived from the rising Damara mountain
Fig. 4.2. The two profiles parallel to the coast (see also Fig. 4.1) show that the main features of the present-day geomorphology have been inherited from the Upper Palaeozoic and Mesozoic topography. The positions of the numerous remnants of glacial Dwyka Beds make it probable that all the major valleys were established already during the Palaeozoic. (Vertical scale in metres.)
Fig. 4.3. Schematised map of the Pan-African fold belt on the Atlantic margin of southern Africa. Between lat. 16° and 34°S. The distribution of high grade metamorphism, migmatites and deep-level granites indicates that the present coastal belt has been deeply denuded since the orogeny. This belt must therefore, after the orogeny, have formed part of a mountain belt of very considerable relief.
range (Germs, in prep.). However that may be, neither the Ovambo Basin nor the Nama Basin seem adequate depositories for the very large volume of sediment which must have been shed by the Damara belt during the early Palaeozoic. Where are these sediments? Where in Africa do we find large volumes of early Palaeozoic sediments?

There are only two regions with such sediments, the Cape Super-Group about 1600 km to the south of the area in question and the Palaeozoic basins of northern Africa about 4000 km to the north. It is very unlikely that sediments derived from the Damara belt could have reached the North African basins, but it is not impossible that some of the erosion products could have contributed to the filling of the Capé Basin, though this basin seems mostly to have been filled with sediments derived from closer at hand (Rust, 1967), probably from an elevated part of the southern branch of the Pan-African fold belt.

Thus no adequate depositories have been found on the African continent for the bulk of the denudation products which must have been shed from the early Palaeozoic Pan-African mountain ranges of southwestern Africa.

The problem is further aggravated if one takes into consideration that by the late Palaeozoic the mountain range had been eroded down to near the present level of exposure, that a westward directed drainage had been established across the metamorphic backbone of the range, and that a large part of this metamorphic belt must underlie the present-day shelf. A great volume of early Palaeozoic clastic sediment must therefore have been transported to somewhere west of the present-day shelf. Two possibilities suggest themselves. The sediments could have found their way into South America or they could have been trapped underneath the continental slope.

The following arguments support the first possibility. (1) The South American coastal belt from Paraná northwards is largely built by rocks which strike subparallel to the coast and suffered their last metamorphism at the time of the Pan-African event. This belt could therefore originally have formed the western margin and foreland of the Pan-African fold belt, in spite of the fact that the Brazilian Minas Super-Group is probably older than the Damara Super-Group of South West Africa. (2) There are early Palaeozoic sediments in parts of the Paraná and the Maranhão basins. At present there seems to be no serious objection to this possibility. But it should be kept in mind that in still intact orogens a considerable portion of the coarser denudation products is, as a rule, trapped in fault-bounded basins close to the zone which has suffered the highest metamorphism and the greatest subsequent elevation. The Permian basins in the Hercynian fold belt of Europe and the Devonian Midland Valley of Scotland are good examples.

Regarding the possibility that the early Palaeozoic denudation products might be found underneath the continental slope, there would be no difficulty if all the arguments for ocean-floor spreading and plate tectonics were disproved. This not being the case, there remains the further possibility that at the same time as the uplift of the mountain range a number of intermontane fault troughs may have come into existence either within or just to the west of the belt of greatest uplift, and that these troughs may have trapped a considerable portion of the coarser erosion products. I think that the formation of such intermontane troughs (internal molasse basins) can be explained by the assumption that the rise of the geosiotherms in a metamorphic belt may cause an expansion of the upper mantle and intrusions of material from the low-velocity layer into the sub-Moho crust, thereby causing a higher isostatic uplift. In such a case erosion may reduce the sialic crust to less than its average continental thickness. This seems to be happening at present in the southern part of the New Zealand geosyncline (Bradshaw, 1973). The subsequent gradual cooling of the upper mantle may then cause subsidences to depths of several kilometres. I think it likely, although there is at present no proof, that such troughs underlie parts of the shelf and the continental slope of the Atlantic seaboard of Africa and also the ‘marginal plateau’ (Heezen and Tharp, 1961) attached to the Brazilian shelf. Similar conditions
Fig. 4.4. On these profiles the highest basement elevations, the altitudes of the Upper Palaeozoic Dwyka Beds and of the Mesozoic formations of the continental margin have respectively been superimposed on single sections normal to the coast. A comparison of the two uppermost profiles shows to what degree the summit level of the pre-Palaeozoic basement (broken double line) has been dissected by Palaeozoic erosion. The position of the basement surface underneath the Ovambo and the South Kalahari Basins is indicated by hatched lines. A comparison with the third profile shows that the Palaeozoic topography was completely buried by Mesozoic aggradation. An averaged position of the Mesozoic erosion surface is indicated by the dotted line.

A comparison of the uppermost profile with a similarly constructed profile of the western margin of Scandinavia, drawn to the same scale, shows a surprising similarity in the overall configuration of these two parts of the Atlantic seaboard. (Vertical scale in metres.)
seem to exist off Norway where deep basins, probably partly filled with Palaeozoic sediments, have been located on the shelf which is otherwise underlain by Caledonian metamorphic rocks (Talwani and Eldholm, 1973). The existence of such fault-bounded troughs associated with the Pan-African fold belt may have greatly facilitated the Mesozoic separation of Africa, which is largely framed by this fold belt (Clifford, 1967), from the

### Fig. 4.5. Three profiles across the shoulders of the Red Sea rift and the Rhine Graben. A comparison with Fig. 4.4 shows that the uptilted shoulders of these Tertiary rifts are much narrower than the elevated parts of the present Atlantic margin. (Vertical scale in metres.)
other Gondwana continents. The great dependence of the African rift valleys on older structures and metamorphic belts, which has been so well demonstrated by McConnell (1972), shows a comparable relationship.

Figure 4.6 gives a schematic interpretation of the Late Precambrian to early Palaeozoic structural and crustal development of the later continental margin of northwestern South West Africa. This tentative interpretation is an attempt to correlate the palaeogeomorphology of this region with changes of crustal thickness caused by orogeny, denudation and isostatic response.

The Late Ordovician glaciation, which may have affected the whole of Africa, may well have rejuvenated some of the earlier faults.

THE LATE PALAEOZOIC GEOMORPHOLOGY OF THE CONTINENTAL MARGIN OF SOUTH WEST AFRICA

Regarding the Atlantic margin of South West Africa the conclusion seems unavoidable that the well-preserved Palaeozoic valleys had a common base level somewhere to the west of the present-day coast (see Figs. 4.1, 4.2, 4.3 and 4.4). The Paraná Basin could have received the African denudation products as advocated by Maack (1963). As, however, none of the diagnostic African erratics have been found in the Permo-Carboniferous tillites of the Paraná Basin there is no evidence for this assumption (Martin, 1961). It is therefore more likely that the glacial debris was deposited in a separate basin which is today buried underneath the younger sediments of the shelf, the continental slope and the marginal plateau of Brazil. Is there any supporting evidence for this assumption? And if there is, was it a terrestrial or a marine basin?

Glacially-shaped, westward-draining valleys containing remnants of Permo-Carboniferous glacial deposits, and evidence that the ice was flowing from east to west (Martin and Schalk, 1957), were until recently only known from the Kunene River to the Huab River (see Fig. 4.3 and Martin, 1968). It is certain that these features extend northwards into Angola, perhaps as far as Gabon (Micholet, Molinas and Penet, 1970). Southwards, the western part of the central highland of South West Africa also contains westwards draining valleys showing geomorphological features similar to those for which a Palaeozoic origin can be proved. However, no Karroo age sediments having been found in the central highland, speculations concerning the late Palaeozoic geomorphology of this region had to remain mere conjectures. This has been changed by the discovery (J. Faupel, pers. comm.) of a tiny patch of Karroo sediment in one of these valleys on the farm Komuanab in the western margin of the Khomas Highlands (see Fig. 4.3, Profile A-B). The patch of indurated conglomeratic sediment, lying unconformably on Damara mica schists, has been preserved because it was intruded by a dolerite sill. This relationship proves the Karroo age of the sediment. The occurrence is too small to allow a conclusion regarding a glacial or non-glacial origin, but the Nausgomab Valley, a tributary of the Kuiseb River, shows the geomorphological characteristics of the glacially shaped Kaokoveld valleys. Because the Khomas Highlands is traversed by numerous faults, the question has to be asked whether the Nausgomab Valley could be a post-Karroo fault trough instead of an exhumed Palaeozoic valley. Some of these faults are found to the east and the west of the Nausgomab Valley, but the valley itself is not bounded by faults. The evidence therefore greatly favours a Palaeozoic origin for this valley and by inference also for the origin of the Swakop and the Kuiseb valleys and some of their tributaries which, like the broad Windhoek Valley, though partly incised along faults, are not fault troughs. The profiles of Figure 4.3 show also that the major geomorphological features remain unchanged from the Kunene to the Gamsberg.

The new discovery extends the belt of westward-directed late Palaeozoic drainage by 200 km, as far south as lat. 23°S. This fact has an important bearing on the late Palaeozoic palaeogeography, because the southern part of the central highland contains remnants of a southward directed drainage system with patches of tillite, and this drainage is connected with the South Kalahari Basin, which was invaded by a marine transgression.
Fig. 4.6. Hypothetical model for the development of the Atlantic margin of northern South West Africa. The assumed position of the present-day coast has been marked by an arrow on each section.
Evidence for an Upper Palaeozoic Sea from the west during the waning stage of the Dwyka glaciation (Martin and Wilczewski, 1970). The glacial marine beds occur at Schlip at a present elevation of 1300 m above sea level, i.e. at a higher elevation than most of the glacial deposits in the westward draining valleys (see Figs. 4.3 and 4.4). It seems possible therefore that some of these valleys, especially the Kunene Valley, could have been fjords at the time of the deglaciation. A very limited search has failed to produce any marine fossils, but a remnant of glacial boulder mudstones in the Kunene Valley contains large marlstone concretions which are very similar to those found in the marine glacial fossiliferous boulder mudstones of the South Kalahari Basin. The probability that the marine ingression into the South Kalahari Basin came from the west makes it highly probable that the westward directed late Palaeozoic valleys were linked to a marine basin situated off the present-day west coast. A good impression of the late Palaeozoic geomorphology of the continental margin can be gained by comparing the uppermost profile on Figure 4.4, on which the highest basement elevations have been superimposed on a section normal to the coast, with the second profile, on which this has been done for the remnants of glacial deposits. The east-west gradients have probably been altered by later warping of the continental margin, but the relative elevations have been altered only locally and only insignificantly by later faulting.

There is also some geophysical indication that beds of Karroo age may underlie the Cretaceous and Cainozoic sediment wedge underneath parts of the shelf off Walvis Bay (Du Plessis et al., 1972). The formation of such a basin over parts of the Pan-African fold belt may be attributed to the same kind of upper mantle processes—cooling following on uplift and erosion of a metamorphic belt—that caused the marine transgressions over parts of the Hercynian belt not long after it had been folded. The formation of late Palaeozoic basins within the realm of Gondwanaland may also have been promoted by the Hercynian orogeny which affected large parts of the globe. Some small oceanic basins may even have been formed during this time between the different parts of Gondwanaland (K. M. Creer, pers. comm.). Figure 4.6 gives a diagrammatic representation of this stage of development of the continental margin.

The above explanation should, however, not be applied to the short-lived marine ingression into the South Kalahari Basin. This ingression is attributed to the isostatic depression of southern Africa under the weight of the Permo-Carboniferous ice cap. But the large-scale isostatic adjustments may well have rejuvenated older faults and thereby contributed to the later break-up of Gondwanaland as suggested by Gough (1970).

The Mesozoic Palaeogeography of Northwestern South West Africa

There are sufficient remnants of Mesozoic beds to allow a fair reconstruction of the geological events that have affected the continental margin during this time.

During the (?) Early Permian waning stage of the glaciation the valleys were filled with glacial and periglacial sediments. During the isostatic rebound of the subcontinent most of this infilling was again removed by erosion but there were considerable remnants, especially in those parts of the topography which had been over-deepened by glacial gouging. The overall gradients must have remained low, otherwise the incision would have cut deeper.

The next event is represented by the deposition of shales, impure coal, sandstones, fine-pebble conglomerate and interbedded concretionary limestones in the lower, western parts of the Huab Valley and the wide Ugab River depression (see Fig. 4.3). These beds were thought to have an upper Dwyka age (Reuning and Martin, 1957), because at Doros a thin conglomerate contains rolled bones which von Huene (Reuning and von Huene, 1925) assigned to the genus Meso-saurus. The bones are, however, not nearly well enough preserved for a valid identification, and the general aspect of the partly red-coloured sequence, combined with Reuning's find of a tooth with two roots, makes a younger Triassic or Jurassic age more likely (von Huene, 1925).

The results of a sedimentological study of
these beds carried out by Hodgson (1970) are important for the interpretation of the Mesozoic development of the continental margin. The cross-bedding directions in the sandstones indicate sediment transport from the north, south and east, in agreement with the basement topography. The important point is that aggradation did occur at all, for this means that the valleys which before had had outlets towards the west, must now have somehow been closed in the west. This is also indicated by the layers of concretionary, partly stromatolitic limestone, and thin limestone beds which are probably of lacustrine origin. The blocking of the westward directed drainage must have been a regional event, because the aggradation eventually smothered the whole old basement topography as shown by a comparison of the third profile of Fig. 4.4 on which the occurrences of Mesozoic rocks have been superimposed on a section normal to the present coast, with the uppermost profile on which the highest basement elevations have been combined. The aggradational sequence contains an unconformity separating the fluviatile and lacustrine beds from the overlying aeolian Etjo Sandstone which was deposited by westerly winds (Bigarella, 1970; Hodgson, 1970). The aeolian sandstone is conformably overlain by and locally interbedded in the Lower Cretaceous plateau basalt flows of the Kao-koveld (Siedner and Miller, 1968). Where these flows are not lying on aeolian sandstone they overstep with subhorizontal attitude onto and over the older basement topography (Reuning and Martin, 1957). The maximum remaining thickness of this pile of lava flows is 800 m. The original thickness must have far exceeded this figure, because the lavas must originally have mantled the large Mesozoic granite plutons (see Fig. 4.4) (Korn and Martin, 1954). This thick sequence of basalt flows could not have accumulated if it had not been confined by some barrier somewhere to the west of the present-day coast. There is thus a long history of aggradation (? Triassic to Cretaceous) over a terrain with a pronounced palaeoslope of long standing. How can this change of drainage conditions be explained?

Under the classical continental drift interpretation this sequence is assigned to the former easternmost rim of the Paraná Basin. Under this assumption the aggradation needs no special explanation. However, the foregoing discussion of the Palaeozoic geology and geomorphology having cast doubt on the classical interpretation, it has to be asked what other explanation might fit the facts? In order to answer this question I shall discuss the probable consequences of rifting and the beginning of ocean-floor spreading on a continental margin.

Geophysical investigations of the Rhine Graben (Mueller, 1970) and the Gregory Rift Valley (Griffith et al., 1971), show that the rifts and the up-tilted rift shoulders are underlain at the crust-mantle boundary by cushions of low-density and warmer upper mantle material. It is not yet clear whether the formation of such a cushion is the result of the rifting or vice versa, but the antithetic up-tilting of the rift shoulders seems to be the direct consequence of the abnormal upper mantle cushion. Such up-tilting might well create a barrier behind which sediments and lava flows could be impounded and aggraded (see Fig. 4.6). Profiles of the shoulders of the Rhine Graben and the Red Sea rift (see Fig. 4.5) show that the shoulders can be elevated sufficiently to impound thicknesses of sediments and lavas of the order of 1000 to 2000 m. Without a barrier of some kind the lava flows could not have accumulated to any great thickness, because flood basalts can spread, even over subhorizontal surfaces, for distances of more than 100 km (Danes, 1972). The pile of basalt flows could therefore not have accumulated if an oceanic basin had existed to the west of the present-day shelf. The choice therefore lies with either contiguity of the lava flows with those of the Paraná Basin or damming behind a high barrier which did not exist during the Palaeozoic. Both interpretations, therefore, imply ocean-floor spreading.

There is some structural evidence for the rift shoulder model. Near the coast the lavas have been displaced by faults which strike subparallel to the coast. Along these faults the lava flows have been antithetically tilted eastwards (R. McG. Miller, pers. comm.). This is just the kind of structure to be ex-
Evidence for an Upper Palaeozoic Sea

So far the model seems to be consistent. But what has happened to the rift shoulders which, in the case of the southwestern African continental margin, should have been situated in the region of the outer shelf or the continental slope?

An answer to this question is suggested by the development of oceanic rifts as interpreted by the ocean-floor spreading concept. The oceanic rift zones have, like the continental rifts, broad up-tilted shoulders which are supported by cushions of hot, low density upper mantle material. As spreading proceeds, and the shoulders migrate away from the active spreading ridge, the upper mantle cools and the crust sinks back to normal oceanic depths. Upper mantle cushions underneath intracontinental non-spreading rifts seem to undergo a similar development. This is indicated by gravity profiles across the Permian Oslo Graben (Ramberg, 1972), which is underlain by an elevation of the upper mantle of normal sub-Moho density. It is probable that this mantle protrusion represents the remnant of a former low density upper mantle cushion which on cooling reverted to normal density. If this interpretation is correct then continental margins up-tilted at the time of the initial rifting should subside again when the spreading ridge has moved away.

Such a sequence of events apparently cannot account for all the major features of the southwestern African continental margin, because beneath the shelf the Karroo-basement interface has probably subsided to well below the depth at which it would be expected, if its gradient on land is extended seawards (Simpson, 1971; Du Plessis et al., 1972; van Andel and Calvet, 1972; see Fig. 4.4). But this feature can also be explained without difficulty. In contrast to the oceanic crust, which along the oceanic ridges remains below sea level, the continental crust along the initial rift was tilted up high above sea level. The rift shoulder may have been elevated by as much as 2000 m above sea level, as shown by the young Red Sea Rift. Therefore the rift shoulders will be eroded down to near sea level, with concomitant isostatic compensation, before the slow cooling of the upper mantle causes subsidence. This means that just along the newly formed continental margin the continental crust will be thinned by several kilometres and will therefore subside well below its original depth (see Fig. 4.6).

A comparison with the young rift shoulders of the Red Sea and the Rhine Graben may give an indication of the approximate width of this deeply sunken belt. Figure 4.5 shows that the most highly elevated part of the rift shoulders is about 100 km wide. As this part will suffer crustal thinning by denudation, it seems reasonable to assume that a sunken belt of continental crust of about this width is underlying the outer shelf and the upper parts of the continental slope.

During the phase of subsidence induced by upper mantle cooling a further process is likely to come into play enhancing the subsidence and causing a certain amount of extension and thinning of the continental crust along its boundary with the oceanic crust. This is to be expected because along such a boundary the continental crust is subjected to tensional forces (Bott, 1971, 1972). There is thus a combination of several processes which will contribute to the formation of fault troughs and basins on the shelf and the continental slope. I regard it as highly probable that these structures will largely coincide with earlier, Late Precambrian and Palaeozoic structures for the existence of which a case has been made in the first part of this paper.

This model for the evolution of the continental margin has one important consequence: it makes it probable that thinned continental crust may in some regions extend far beyond the 500 fathom depth contour.
which is being used for continental reassemblies. As suggested by Talwani and Eldholm (1973) the boundary of magnetic smooth zones may better define the limit of the continental crust than a depth contour (see also Mascle and Phillips, 1972). If these deductions are correct then the former position of Africa relative to South America must have differed to some extent from the reassembly found in most Gondwanaland reconstructions. This may also be the case for the parts of Gondwanaland surrounding the Indian Ocean, and may explain the rather unsatisfactory fits of that region.

CONCLUSIONS

The foregoing discussion has led to the following conclusions:

1. The Atlantic margin of southern Africa coincides over the greater part of its length with portions of the Late Precambrian-Cambrian fold belt. Thickening of the crust accompanied by high-grade metamorphism, perhaps connected with intrusions of material from the low-velocity zone into the sub-Moho upper mantle, have caused great isostatic uplift and deep-reaching denudation. It seems probable that parts of the metamorphic belt, situated to the west of the present-day coast, were so deeply eroded that they sank below sea level when the upper mantle cooled again (Fig. 4.6).

2. During the Palaeozoic a westward directed drainage was established across the fold belt. It was probably linked to the basins and fault-troughs formed as a result of deep erosion and upper mantle cooling (see Fig. 4.6). The Permo-Carboniferous ice cap discharged its glacial debris westwards (Fig. 4.6) and southwards into marine basins. Isostatic adjustments may well have rejuvenated existing structures and thus facilitated the later break-up of Gondwanaland.

3. The South Atlantic rift between Africa and South America, largely following older structures, may have developed during the Triassic though active spreading may have begun much later. The rifting was caused or accompanied by the formation of a hot upper mantle cushion above which the rift shoulders were uplifted to elevations of 1000 to 2000 m above sea level. The long established westerly drainage was blocked by this uplift. As a consequence sediments began to accumulate and later, when the spreading began, lava erupted and eventually covered most of the subcontinent (Fig. 4.6).

4. The most highly elevated parts of the rift shoulders were deeply eroded and subsided below sea level when the spreading ridge receded and the upper mantle cushion cooled. The subsiding continental margin with its complex older structures was buried underneath a thick wedge of Cretaceous to Recent marine sediments (Fig. 4.6). There have been only slight shifts of the coastline since the Late Cretaceous. During this time the coastal belt was graded to the Atlantic base level, and the Great Escarpment was formed as an erosional feature, accentuated by further up-warping, partly at least as an isostatic response to the denudation. The present-day geomorphology is largely dependent on the directions of the late Palaeozoic drainage with respect to the later Atlantic rift. The overall geomorphology closely resembles that of the Atlantic margin of Scandinavia (see Fig. 4.4), where the Caledonian structures may have influenced later events in a manner similar to that of the Pan-African structures along the coasts of Africa.

5. The complex history of the continental margin was repeatedly controlled by vertical uplifts and subsidences caused by upper mantle processes.

6. It is probable that the boundary between the continental and the oceanic crust does in some regions depart considerably from the position of the 500 fathom depth contour. There is therefore greater latitude than is usually assumed for the reconstruction of Gondwanaland.

7. It should eventually be possible to check these deductions by geophysical means or by off-shore drilling.

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Pattern of Clastic Dispersal in the Lower Gondwana Coalfields of Peninsular India

N. D. MITRA, B. LASKAR AND U. K. BASU

ABSTRACT

The Lower Gondwana Coal Measures, comprising mainly the Karharbari and Barakar Formations, display crudely developed fining upward cycles, each with a scoured surface at the base. Regional and basinal variation in gross lithology is analysed using sandstone-shale ratios because of the dominance of these lithotypes in the cycles. In the Damodar Valley basin belt the top stratum and backswamp deposits form a significant proportion of the lithic fill, whereas in the Pranhita-Godavari, Son-Mahanadi and Satpura belt the cycles are more arenaceous with high sandstone-shale ratios.

The high variation in palaeocurrent directions in the Damodar Valley basins, together with the high proportion of the finer clastics in the sedimentary cycles, suggest that these basins were drained by meandering streams of high sinuosity flowing along local hydraulic gradients. The high frequency of sandstone dominated cycles, and low variability of palaeocurrent vectors in the Son, Pranhita-Godavari and Satpura basin belts, indicate deposition of sediments in braided streams of low sinuosity.

Reconstruction of the palaeo-flow regions suggests that the streams of the Son-Mahanadi and Pranhita-Godavari Valley basins flowed in deeper and less sinuous channels than those of the Damodar Valley basins.

The basins in the Damodar Valley were of flexure-bound intermontane type with multiple sources of supply. The Son-Mahanadi, Pranhita-Godavari and Satpura Basins were, on the contrary, drained only by northwesterly flowing river systems. The stability of northwesterly palaeoslopes in the central part of Peninsular India resulted from downwarps along the northeast-southwesterly trending Son-Narmada lineament. The northwesterly flowing river systems probably discharged into Tethys along this zone of negative gravity anomaly.

INTRODUCTION

The Gondwana coalfields of India lie in linear belts in the Precambrian platform. A regional analysis of the dispersal pattern of the coal measure sediments should provide information not only on the flow patterns and the nature of the depositing fluvial systems, but also on other geomorphic and palaeogeographic features.

FACIES VARIATIONS IN THE DIFFERENT BASIN BELTS

The Lower Gondwana Coal Measures of Peninsular India are mainly in the Karharbari and Barakar Formations. These are crudely cyclic sequences of arkosic sandstone, siltstone, shales and coal seams. The cycles are sometimes similar to the typical fining upward cycle reported by Allen (1965) in
modern river sediments. The cyclic pattern in the coal measures, however, is more often abbreviated or modified.

**Sandstone units**

The sandstone units of the coal measures are the most important members of the cycles. Their base is often a scoured erosional surface. The coal measure sediments, however, exhibit a wide range in composition and sedimentary structures, and may be subdivided into the following lithologically and structurally distinctive units.

1. Massive-bedded coarse sandstones and pebbly sandstones. These are developed either as thick blanket type deposits or as lenticular units at the erosive base of some coal measure cycles. The massive pebbly sandstones represent the antidune phase of flow in the upper flow regime. Intergradational cross-bedded pebbly sandstones, however, were deposited as channel bars in braided streams.

2. Large-scale cross-bedded sandstones. These sandstones have foresets more than 4 cm thick and are the most abundant sandstone types in the coal measures. Thicker foresets are developed in the coarse-grained sandstone and sometimes the pebbles are aligned along the foreset plane. Tabular cross-bedding in solitary sets up to a thickness of 2 m have been recorded in the coarse Barakar sandstones, and these in general are of Alpha cross stratification type which is considered to be the product of outward building of deltaic sedimentation units into bodies of standing water (Collinson, 1968).

3. Small-scale cross-bedded sandstones. Cross-bedding shows a decrease in scale vertically towards current-rippled or ripple-drift cross-laminated sandstone. There is an accompanying decrease in grain size which is in the range of fine sand to coarse silt. These correspond to Lambda or Kappa cross stratification. It is suggested that ripple-laminated sandstones above the large-scale cross-bedded sandstones are the result of overbank flooding close to the river channel.

4. Flat bedded sandstones. These units are interbedded with the cross-bedded sandstones and are generally best developed in the finer-grained micaceous sandstones.

5. Laminated siltstone. These siltstone beds are disposed in horizontal or ripple-laminated units. Where the horizontal lamination is discontinuous and the ripples are erosive, the rocks have a streaky appearance. This facies marks the transition between the sandstones and the overlying coal-shale sequence. Such units are primarily the product of suspended load evidently deposited from a low flow regime in tranquil conditions.

6. Shales and fire clays. These units succeed the laminated siltstones or at places directly overlie the small-scale cross-bedded sandstones. The bedding is generally poorly developed, but appearance of a few silty interbands gives a striped appearance. These units are the top-stratum deposits, being the product of flood basin accretion.

**Coal Seams**

The cyclic pattern, where fully developed, begins with a pebbly sandstone and ends with a coal seam. Such complete cycles are not common within the Barakar Coal Measures, but can be observed in the Bansloi River section of the Panchwara Basin in the Rajmahal Hills. The Lower Gondwana basins exhibit a varied development of coal seams. In the South Karanpura Basin as many as forty-four coal seams have been recorded within 900 m of section. Coal seams are generally intercalated with shale bands and exhibit a tendency to split laterally. The formation of coal seams in the fluvial cycle indicates a return to backswamp conditions.

**REGIONAL VARIABILITY IN FACIES ORGANISATION**

Exploratory drilling in the different coalfields has brought to light a wealth of information on the regional facies of the Barakar
Table 5.1. Sandstone/Shale Ratios in Coal Measures

<table>
<thead>
<tr>
<th>Name of basin belt</th>
<th>Name of Gondwana basin and sector explored</th>
<th>Maximum thickness of coal measures explored (Barakar and Karharbari Formations taken together)</th>
<th>Mean sandstone:shale ratio of coal measures sediment (Barakar and Karharbari Formations taken together)</th>
<th>Average variations in thickness of sandstone units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Damodar Valley</td>
<td>Raniganj</td>
<td>650</td>
<td>1.4 : 1</td>
<td>2-9</td>
</tr>
<tr>
<td></td>
<td>West Bokaro</td>
<td>400</td>
<td>1 : 2</td>
<td>0.5-7.5</td>
</tr>
<tr>
<td></td>
<td>North Karanpura</td>
<td>300</td>
<td>1.1 : 1</td>
<td>2-5</td>
</tr>
<tr>
<td></td>
<td>Ramgarh</td>
<td>515</td>
<td>1.3 : 1</td>
<td>2-11</td>
</tr>
<tr>
<td></td>
<td>Daltonganj</td>
<td>140</td>
<td>3 : 1</td>
<td>3-4</td>
</tr>
<tr>
<td>Mahanadi Valley</td>
<td>Talchir Central Sector</td>
<td>650</td>
<td>1.5 : 1</td>
<td>2.0-10</td>
</tr>
<tr>
<td></td>
<td>1b-river</td>
<td>300 (Barakar Formation only)</td>
<td>1 : 1.7</td>
<td>One 40-60 base Barakar</td>
</tr>
<tr>
<td>Son Valley</td>
<td>Sisapgar</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kotma Sector</td>
<td>200</td>
<td>8 : 1</td>
<td>10-35</td>
</tr>
<tr>
<td></td>
<td>Bijuri Sector</td>
<td>308</td>
<td>4 : 1</td>
<td>5.0-44</td>
</tr>
<tr>
<td></td>
<td>Batura Sector</td>
<td>270</td>
<td>14 : 1</td>
<td>11.0-26</td>
</tr>
<tr>
<td></td>
<td>Birampur</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Khargaon Sector</td>
<td>275</td>
<td>7 : 1</td>
<td>13-36</td>
</tr>
<tr>
<td></td>
<td>Songara Sector</td>
<td>270</td>
<td>9 : 1</td>
<td>10-21</td>
</tr>
<tr>
<td></td>
<td>Lakhapur</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Western Sector</td>
<td>300</td>
<td>8 : 1</td>
<td>12-40</td>
</tr>
<tr>
<td></td>
<td>Eastern Sector</td>
<td>205</td>
<td>9 : 1</td>
<td>11-28</td>
</tr>
<tr>
<td></td>
<td>Sonhat</td>
<td>900</td>
<td>3 : 1</td>
<td>5-17</td>
</tr>
<tr>
<td></td>
<td>Chirimiri</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>TISCO-I Sector</td>
<td>180</td>
<td>4.5 : 1</td>
<td>4-30</td>
</tr>
<tr>
<td></td>
<td>TISCO Western Sector</td>
<td>200</td>
<td>11 : 1</td>
<td>8-24</td>
</tr>
<tr>
<td>Rajmahal Hills</td>
<td>Mahuagari</td>
<td>570</td>
<td>3 : 1</td>
<td>0.5-10</td>
</tr>
<tr>
<td></td>
<td>Chuperbhita</td>
<td>550</td>
<td>3.5 : 1</td>
<td>1-10</td>
</tr>
<tr>
<td></td>
<td>Panchwara</td>
<td>310</td>
<td>3 : 1</td>
<td>2-14</td>
</tr>
<tr>
<td>Pranhita Godavari</td>
<td>Chanda</td>
<td>220</td>
<td>7.5 : 1</td>
<td>7-30</td>
</tr>
<tr>
<td></td>
<td>Majri Sector</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Godavari Valley</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bedadanuru Sector (southernmost part of Godavari Coalfield)</td>
<td>270</td>
<td>5 : 1</td>
<td>2-98</td>
</tr>
<tr>
<td></td>
<td>Golet</td>
<td>225</td>
<td>6 : 1</td>
<td>3-27</td>
</tr>
<tr>
<td></td>
<td>Kamptee</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Silewara</td>
<td>300</td>
<td>3 : 1</td>
<td>2-15</td>
</tr>
<tr>
<td></td>
<td>Umrer</td>
<td>200</td>
<td>2 : 1</td>
<td>2-3</td>
</tr>
<tr>
<td></td>
<td>Pench-Kanhan-Tawa</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Damuda Sector</td>
<td>250</td>
<td>5 : 1</td>
<td>2-5 in general a few units 10-24 thick</td>
</tr>
<tr>
<td></td>
<td>Tandsi Sector</td>
<td>380</td>
<td>4 : 1</td>
<td>2-9</td>
</tr>
<tr>
<td></td>
<td>Pathakhera</td>
<td>475</td>
<td>4 : 1</td>
<td>2-28</td>
</tr>
<tr>
<td></td>
<td>Mohpani</td>
<td>260</td>
<td>2 : 1</td>
<td>2-3 in general</td>
</tr>
</tbody>
</table>

and Karharbari coal measures. Because of the dominance of sandstone and shale, variation in gross lithofacies is well reflected in the pattern of variation of sandstone-shale ratios, shown in Table 5.1. Though the sandstone-shale ratio shows significant intrabasinal variations, a mean value for the coal measures of an individual basin has been worked out for inter-regional comparison.

The most important conclusion to be drawn from the above synthesis is that in the Damodar Valley basins and in the Talchir...
Basin of the Mahanadi Valley, the modal cycles include a significant proportion of top-stratum deposits. The top-stratum and back-swamp deposits together constitute about 50 per cent of the lithic fill. In contrast, in the Son, Pranhita-Godavari and Satpura basin belts, the large-scale cross-bedded sandstones form the dominant unit of the modal cycle. In these areas, the high sandstone-shale ratio of 4:1 to 14:1 is accompanied by an increase in thickness of large-scale cross-bedded sandstone members in an individual cycle. It is quite possible that these thick sandstone units result from the superposition of the sandstone bodies of two cycles, or they may represent an anastomosing complex of a large fluvial network. The sandstone units are thus arranged in the manner of multistorey and multilateral bodies (Potter, 1963).

INTERPRETATION OF PALAEOFLOW DIRECTIONS

The dispersal pattern of the coal measures sediments in the different basin belts is worked out mainly from the study of the cross-bedding dip azimuths. Cross-bedding is the most profusely developed primary sedimentary structure in the sandstones, and the mean current directions determined from these azimuths is the best indicator of the regional palaeoslope available.

In the Damodar Valley basin belt the palaeocurrent pattern shows wide regional variability. The individual basins are of 'intermontane' type, with sediment supply from the north, east and west (see Fig. 5.1). They formed depressions in which there was centripetal drainage, and periodically they became the sites of extensive marshy conditions in which thick top-stratum and back-swamp deposits could form. An analysis of the palaeocurrent pattern in the Damodar Valley basins indicates that they had very little fluvial inter-communication. If a stream draining the entire east-west trending Damodar Valley basin belt existed, it would have flowed westerly close to the southern boundary fault. If they ever existed, traces of this pattern of fluvial dispersal have been obscured, probably as a result of later movements along the boundary fault.

The high variability of palaeocurrent vectors suggests that the Damodar Valley basins were drained by meandering rather than by braided streams. This is in conformity with the high proportion of top-stratum deposits recorded in these basins, because it is now well established that mean percentage of silt-clay rises as stream sinuosity increases. Leopold and Wolman (1960), on the basis of their studies on Recent sediments, concluded that the top-stratum deposits form a minor proportion of floodplain deposits. It is, therefore, clear that thick silt and clay units of the coal measures in the Damodar Valley belt originated from a large network of channel belts with high sinuosity draining across an alluvial plain. The presence of a large amount of overbank sediment in the Damodar Valley further testifies to the growth of appreciable alluvial relief. Such relief is the result of restriction of movement of channel belts by clay-plugged former channels—a feature of channels of high sinuosity (Allen and Friend, 1968).

In the Rajmahal basins, the palaeocurrent patterns show an east-northeasterly trend and the vector mean varies from 64° to 75°. This evidently indicates discharge of this drainage to the sea close to the present Himalayan foothills at Darjeeling.

As can be seen from Figure 5.1, palaeocurrent directions in the Mahanadi Valley and the Pranhita-Godavari Valley basins are uniformly west-northwesterly throughout. In the Son Valley coalfields, which are widely separated and have a considerable east-west extent, a northwesterly direction is the most common; though at Tatapani-Ramkola at the eastern extremity of the basins the sector veers toward the north, and at Umaria at the western extremity, it veers toward the west. The east-west elongated Satpura basin belt also has uniform northwesterly oriented palaeocurrents.

The preferred direction of sediment transport in the Son, Mahanadi, Satpura and Pranhita-Godavari basin belts is a reflection of the stability of the northwesterly inclined palaeoslope in Central India at this time. As sedimentation progressed, the maintenance of a uniform palaeoslope resulted in the deposition of many multistorey sand bodies. The coal measures of this area, with a predominance of sandstone units along with a
corresponding decrease in the overbank sediments, point to a regime characterised by repeated reoccupation by the channel belts of their former positions. Such a pattern of channel migration could occur in the case of low sinuosity streams, whose channels could wander freely from side to side (Allen and Friend, 1968).

**ESTIMATION OF FLOW PARAMETERS**

In recent years there has been a growing emphasis on the reconstruction of the hydraulics of ancient flow regimes. It has been suggested that in environmental reconstruction the size of the bed material and the velocity and depth of the stream flow are the three fundamental parameters from which other parameters can be systematically extrapolated (Jopling, 1966).

The sandstones of the Lower Gondwana coal measures show wide lateral and vertical variations in grain size. Nevertheless in most of the basin belts medium-grained arkosic sandstones predominate. Such sandstones, characterised by large-scale cross-bedding, have a mean grain size of approximately 0.5 mm. From Sundborg's (1936) competency diagrams, the competent velocity for such a grain size range would be 0.24-0.27 m/sec. Since the Barakar and Karharbari sandstones contain a significant proportion (about 20-25 per cent) of fine sand and silt, which were transported as suspended load, the velocities of the streams required to cause an
appreciable amount of suspended transport would be 2-2.5 times the competent velocities of 0.24-0.27 m/sec. In summary, the mean velocities of the streams which laid down the sediments of the coal measures in Peninsular India were little more than 0.6 m/sec. The internal sedimentary structures, such as sharply defined laminations in sandstone and relatively steep angle of the foreset laminae (20°-25°), also indicate that the stream velocity was not particularly high.

The large-scale cross-bedded sandstones also provide an index of stream depth. In the case of point-bar deposits, co-sets of cross-bedded strata may attain a thickness equivalent to maximum water depth (Royse, 1970). Moody-Stuart (1966) also suggests that the thickness of cross stratified units is equivalent to the channel depth at bankfull stage. Thus in the Damodar Valley basin belt, where the mean thickness of cross-bedded sandstone is 3-4 m, the stream depth at meander loop was approximately 3-4 m. As sedimentation progressed, the depth of the stream channels fluctuated both in time and space. In the Son-Mahanadi, Pranhita-Godavari and Satpura Valley, on the contrary, the tabular cross-bedded sandstones in the individual cycles attain an average thickness of 10-15 m, and hence streams in this basin flowed in deeper channels.

Another line of analysis of the hydraulic regimen of the palaeostreams stems from the quantitative approach adopted by Schumm (1968). Preliminary work using this approach provides a confirmation that the Damodar Valley streams were more sinuous than those of the Son Valley, but the analytical method and the assumptions necessary to apply the method need more detailed analysis.

PALAEOGEOGRAPHIC SUMMARY

The east-west chain of Damodar Valley basins, extending from Raniganj Basin in the east to Auranga in the west, were drained by meandering streams of high sinuosity flowing along local hydraulic gradients. Due to the centripetal pattern of the drainage system, coupled with periodic subsidence, these basins were periodically the locales of marshy conditions in which a thick pile of back-swamp sediments was deposited. Such a model does not preclude the possibility of some kind of a common drainage link between the separate intermontane basins.

Evidently the northeasterly trending rivers draining the Rajmahal Basin, and possibly the hidden Purnea Basin, discharged into the sea somewhere near the foot of the Darjeeling Himalayas. The Precambrian uplands of Santhal Parganas formed a water divide between the rivers draining the Raniganj and Rajmahal Basins.

The Son-Mahanadi Valley sandstones were mainly deposited from braided streams of low sinuosity flowing in the northwesterly direction. The stabilisation of the northwesterly palaeoslope and uniformity of lithic fill in this basin belt suggest no significant lateral change in tectono-sedimentological regime over this wide area of Central India.

The Pranhita-Godavari and Satpura basin belts had similar and similarly oriented drainage systems.

The stability of the northwesterly palaeoslope during the sedimentation of the Lower Gondwana Coal Measures probably resulted from continued downwarp along the Narmada-Son lineament. In all probability the network of northwesterly flowing river systems draining the Central Indian Gondwana basins ultimately discharged to the Tethys along this zone of negative gravity anomaly. This postulate is supported by the presence of a marine bed at the base of the coal measures in Umaria Coalfield, located very close to the Son-Narmada lineament. The river system there evidently flowed westerly, following the path of the marine regression.

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Sedimentological Evidence for the Extension of the African Continent Southwards During the Late Permian-Early Triassic Times

J. C. THERON

ABSTRACT

The Beaufort Group of the main Karroo Basin extends over an area of some 200,000 km².

In the Orange Free State, Natal and the Cape Province, a persistent northerly trend in the direction of sediment transport is shown by palaeocurrent analysis of most of the formations of this group. It is inferred that a source area of large dimensions, providing the sediments of the southerly derived formations, lay to the south of the present coastline.

Heavy mineral analyses show that rocks derived from granitic terrains within the basin at that time provided assemblages comparable to those of the sediments derived from the southern source. The conclusion is drawn that the rock types exposed in the southern provenance areas had a composition corresponding to those lying below the Karroo sequence.

This in turn suggests the continuation of the continent southwards. Furthermore, evidence in the form of pebbles in the Beaufort Group near East London points to a source devoid of rocks belonging to the Cape Super-Group. This, together with the palaeocurrent and mineralogical evidence, indicates that the Cape Folded Range (Samfrau Geosyncline) did not exist much further to the east of its present-day outcrop. A large area of positive relief composed essentially of Precambrian rocks constituted the area to the southeast of Southern Africa.

INTRODUCTION

Beaufort Group sediments are mainly alternating layers of mudstone and sandstone. Fossil evidence shows that they were deposited from Middle Permian to Early Triassic times (Romer, 1970: 112-15). Their outcrop area is some 200,000 km² in extent, and a thickness of more than 3000 m is given for the Eastern Cape where their development was at a maximum (Winter and Venter, 1970: 397).

The sedimentological evidence to be presented here is based mainly upon a detailed palaeocurrent analysis of the entire outcrop area, and is supported by data obtained from heavy mineral analysis, particle size determinations, and field observations.

The fact that sandstone occurs in the Beaufort sequence throughout the entire outcrop area facilitates an investigation of the ancient sediment dispersal systems operating at the time. Whereas other investigations (i.e. Ryan, 1967; Stratten, 1970) were limited by absence of directional structures in some areas, poor exposures and relatively narrow outcrop widths of the lower groups, readings were obtained at most sites selected for the present study.
Although the percentage of sandstone in sections varies from more than 60 per cent down to a few per cent, the thickness of the beds of sandstone is rarely more than 30 m. One notable exception is the Katberg Sandstone which attains a thickness of some 600 m (Johnson, 1966), but it is limited to about one per cent of the total outcrop area of the Beaufort. According to du Toit (1954: 294) some of the thicker sandstone bodies occur over about 40 per cent of the total outcrop area, and these may therefore be regarded as sandstone sheets. Sampling by the writer has indicated that the outcrops are areally more restricted than du Toit surmised, and that his correlation of such units as the upper sandstone of the Middle Beaufort Stage (old terminology) is not reliable.

Most of the sandstone bodies are lenticular when seen in cross-section, and it is probable that most of them are shoestrings formed by meandering streams. A fluviatile dispersal system is visualised for the greater part of the Beaufort Basin.

SAMPLING

Stratigraphical columns and detailed geological maps are not available for the greater part of the outcrop of the Beaufort Group. The presence of a myriad dolerite intrusions and the absence of marker beds in the majority of the formations prohibits the subdivision of this group without detailed mapping. For these reasons sampling was done on a grid system and with complete disregard for the stratigraphic position of the sampling site.

The method of sampling is not thought to be detrimental to the outcome of this study, because where the stratigraphic positions of the samples were known, specimens from all levels could be shown to yield similar results. For example, in an area of about 2500 km² and representing a stratigraphical thickness of some 500 m, the sandstones had a mean grain-size ranging from 0.212 mm to 0.115 mm for thirty-two samples. The heavy mineral assemblages showed no statistically significant differences between samples (Theron, 1965: 51, 61-3). It was argued that, as the rocks are so alike in texture and composition, the source and basinal structural elements retained their relative positions and importance throughout the greater part of the depositional interval. The drainage and aggrading networks in a broad sense, therefore, maintained their relation during subsidence and contemporaneous sedimentation. The supply of sediment was greater than the requirements of the subsiding basin in the south so that a palaeoslope was maintained in a northerly direction.

An analysis of the basin as attempted in this study should show the influence of the major structural elements, whereas minor tectonic events cannot be identified without more precise stratigraphic control. This limitation should be kept in mind.

PALAEOCURRENT ANALYSIS

The palaeocurrent analysis forms the more important section of the research. A very large sample of readings on directional structures was drawn from the whole Beaufort outcrop area. This sample consists of about 11,000 readings taken at more than 1000 localities. The sampling of the basin is to a certain extent non-uniform as different areas were investigated at various times over a period of nine years. The highest sampling density is in the Orange Free State, while the least is in the basin west of long. 24°E. The sampling density in all areas is adequate to show the more salient features of the ancient dispersal system.

The direction of sediment transport was calculated for each sampling site using the vector mean statistic described by Potter and Pettijohn (1963: 264). The moving average directions, shown on Figure 6.1, were calculated from these directions for each degree square. The resultant direction was plotted as an arrow at the centre of the square. The progression was in steps of half degree units. The moving average direction of the dispersal system is indicated by the arrow, and the length of the arrow represents the vector strength (reliability) of the direction.

An average of about eleven readings per locality has been shown to be adequate for the method of sampling (Theron, 1970: 65-72). Most of the readings were taken on well-exposed cross-bedded structures. All except an insignificant number of cross-bedded
Fig. 6.1. Moving average palaeocurrent map of the Beaufort Group

units were of the trough type, and in all these cases the direction of the axis was measured. Where directions of dip of the foreset planes are recorded a much greater number of readings has to be taken (J. J. Bigarella, pers. comm.).

All except one of the recognised sedimentary units of the Beaufort Group had source areas situated in the south. This anomalous
unit, called here the Mooi River Formation, though extending over a large area, constitutes a small part of the volume of the strata of the Beaufort Group.

An area of low vector strength occurs in the western part of the basin while another extends from the central part to the northern perimeter of the basin (Fig. 6.2). It is noteworthy that this is not a marginal feature but lies well within the Beaufort Basin.

A large area of high vector strength is indicated in the vicinity of East London along the southeastern margin of the basin (Fig. 6.3). Smaller areas are present at different locations throughout the basin, but display a preference for marginal areas.

Examining the directional data of the areas of high vector strength it is apparent that for some there is a marked deviation between directions indicated. This may be due to a low sampling density for certain marginal areas. Two areas, however, are considered to be of importance; the East London area, and one approximately halfway between Cape Town and East London.

In summary, the following features should be particularly noted (see Fig. 6.1).

1. A weak but consistent dispersal system existed in the western part of the Beaufort Basin.

2. A strong radiating pattern is present in the southeastern part of the outcrop and may be traced as far north as Bethlehem and Harrismith in the Orange Free State.

3. A third major dispersal system, resulting in the sedimentation of the Mooi River Formation, was active in the northern part of the basin. This system was active only for a short period during the sedimentation of the Beaufort Group.

HEAVY MINERAL ANALYSIS

An analysis of more than 100 slides shows that a characteristic feature of the heavy mineral assemblages is the high proportion of garnet present in the sandstones of the Beaufort Group. Investigations of the lower groups have also indicated that garnet is a common mineral throughout the Karroo strata, with a few exceptions.

The garnet content of the Dwyka is very high (Rust, 1962: 193; de Villiers and War- daugh, 1962). The rocks of the Cape Super-Group lying below the Dwyka are almost totally devoid of garnet (Rust, 1962). The most characteristic heavy mineral species of the Ecca is also garnet (Nel, 1962: 93, 97), but the proportion is less than in the underlying Dwyka.

The Molteno Formation of the Stormberg Group, resting upon the Beaufort, has an almost total absence of garnet above the contact in the northern part of the basin (Theron, 1970: 132), but garnet again appears in the upper beds (Kingsley, 1964). Rust (1962) showed that certain horizons in the garnet deficient sequence contain appreciable amounts of garnet. In turn the overlying Red Beds have garnet as a common accessory mineral (le Roux, 1969: 37-8). Le Roux (pp. 61-3) also shows that the Molteno and Red Beds had a southerly source, which is in agreement with a palaeocurrent analysis on the Molteno by Botha and Theron (1967).

The sediments of the Beaufort, therefore, represent a phase in the denudation of a southerly provenance. Garnet, although being the predominant accessory mineral in this group, varies in respect of its proportion to the other minerals. This is mainly governed by the provenance. Climate, topography, palaeoslope and distance over which the sediments were transported are also considered to be the major controlling factors. The Mooi River Formation with its proximal source areas has a much higher proportion of garnet than its southern counterparts, reflecting its immaturity.

Although the object of the analysis was to identify zircon, rutile, tourmaline and garnet, and group the remainder into three classes, namely, translucent, opaque and authigenic, the remaining translucent detrital minerals consisted principally of apatite, monazite, and titanite.

GRAIN SIZE DISTRIBUTION

The maximum grain size in most of the Beaufort Group is rarely larger than 1 mm, and the mean diameter is usually less than 0.25 mm. The rocks of the Katberg Sandstone and the Mooi River Formation are the only two known exceptions. The largest clast
recorded in the Mooi River Formation has a maximum diameter of about 18.6 cm and a weight of more than 4 kg. A lighter flat boulder has a diameter of about 19.2 cm. The rudaceous material of this formation (including the two clasts described above) is mostly angular to sub-rounded.

The clasts of the Katberg Sandstone are smaller and more rounded than those of the Mooi River Formation, the maximum to date being 12.3 cm. These may often be classified as well-rounded and only the more fissile types are more angular. The rudaceous material of the Katberg Sandstone is found in a very limited area along the coast near East London. Outcrops carrying pebbles were recently found further inland (100 km ±) by C. J. Gunter (pers. comm.).

The rudaceous material rarely appears as conglomeratic lenses or layers, occasionally as pebble-washes but more commonly as dispersed clasts in the coarse- to medium-grained sandstone and arkoses of both formations described above. Most of the arenites of the Mooi River Formation may be classified as arkoses (Theron, 1970), and the Katberg Sandstone has a high percentage of feldspar and may often be classified as arkosic (C. J. Gunter, pers. comm.).

The rudaceous constituents of the Mooi River Formation proved to be few in number and closely related. Gneiss is the main rock type, often containing relatively large amounts of a red garnet which may be observed with the unaided eye. Milky quartz and large fragments of microcline are also present, with some of the feldspars having a diameter of several centimetres, suggesting a pegmatitic origin (Theron, 1970: 140). Pebbles of red granite and quartzite were found in more restricted areas. This suggests a metamorphic terrain as the principle source for the sediments of this formation.

The rudaceous material of the Katberg Sandstone is more polymictic. According to Mountain (1939) and Johnson (1966) the clasts are fine- to coarse-grained red granite, white granite, gneiss, quartz-feldspar porphyry, quartzite (varying from fine- to coarse-grained), sandstone, conglomerate and fossilised wood. It is noteworthy that there is a relatively large number of fissile fossilised wood clasts present. Furthermore there appears to be a rapid reduction in the size of the pebbles in a northerly direction.

DISCUSSION

Palaeocurrent Analysis

The consistent trend of the palaeocurrents indicates that a northerly directed palaeoslope existed for the duration of the Beaufort sedimentation. No evidence of a break in sedimentation has yet been proved in the southern part of the basin. The Mooi River Formation represents a period of change in the northern part of the basin, but the extent and importance of this break has not been fully assessed.

The possibility that the contact between the Ecca and Beaufort Groups is diachronous has been voiced by Ryan (1967: 53) and this has been endorsed by the writer (Theron, 1970: 187). This view was presented on grounds that the basal units of the Beaufort overlap northwards. It would appear as if there was a slow regression of the Ecca basinal area northwards, while a fluviatile association, represented by the more arenaceous Beaufort, developed in the south. This advanced northwards, at first as deltas and later as an ever broadening coastal plain, with abundant vegetation and aquatic and terrestrial life. This low-lying land eventually extended far to the north. The sediments continued gravitating northwards, indicating that a basinal area existed to the north, probably well beyond the present northern limit of the Beaufort Basin. Argillaceous sedimentary rock north of the Karroo Basin, previously correlated with the Ecca, may in fact be the time equivalent of the Beaufort in the south, as Beaufort rocks generally appear to be absent. However, Dsrydall and Weller (1966: 63) remark that the Madumabisa Mudstone of Zambia ranges from Ecca to lower Beaufort in age, and this could represent such a basinal area. Though the presence of rocks equivalent to the Beaufort have not been reported from Malagasy by Flores (1970: 7-8), there are indications of a marine environment during the Permian and again in the Triassic.

An entry point to the Beaufort Basin was present off the present coast near East Lon-
Fig. 6.2. Areas of low vector strength in the Beaufort Group indicating the moving average values of the vector strengths of the palaeocurrent directions. The values were calculated for each degree square and plotted at the centre of the square. In this diagram areas having a vector strength of 45 per cent and less are shaded. This limit was chosen arbitrarily.
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don (see Fig. 6.1). A fan-shaped distributary
carried the sediment into the eastern part of
the basin, and far to the north. This pattern
probably represents the dispersal pattern of a
well established delta draining a large area
to the southeast. Evidence that a provenance
area could have existed to the southeast of
East London as early as Witteberg times is
given by Mountain (1964: 205). Teichert
(1970: 130) is of the opinion that the upper
Witteberg could be of Carboniferous age.
Stratten (1970) and Ryan (1967) indicate
that a depositional area existed to the south­
east during Dwyka and Ecca times (Carboni­
ferous-Permian). This is questioned in the
light of the evidence presented here, and
data from a detailed investigation carried out
by Kingsley (pers. comm.) on the Ecca west
of East London indicating that the palaeo­
current pattern corresponds closely with that
of the overlying Beaufort in this area. Out­
crops of Dwyka tillite are not present in that
area near East London where Stratten
(1970: 483, fig. 1) indicated an exit from the
Dwyka Basin.

It is therefore suggested that terrains in
the southeast were yielding detritus from as
early as the Carboniferous and as late as the

An area of high vector strength (Fig. 6.3)
is also indicated to the north of Mossel Bay.
The direction of transport is parallel to the
strike of the Cape Fold Belt. Although it is
not apparent from the moving average map
this direction is recurrent along the southern
margin of the Beaufort Basin. An entry point
in the southwestern part of the basin is in­
dicated. Because the palaeocurrent directions
are parallel to the fold range and as these
beds are slightly folded, it is argued that
these lowermost beds were deposited prior to
the folding of that part of the geosyncline.
The current directions possibly represent flow
longitudinal to the axis of the basin as has
been found in the turbidite basins described
by Potter and Pettijohn (1963: 241). The
sequence consists primarily of mudstone with
minor very fine-grained dirty sandstone
(greywacke?) displaying micro-cross-bedding
and current lineation. Sole structures were
observed at one locality. A flysch-like en­
v
vironment could have existed at this stage
of Beaufort sedimentation along the southern
margin of the basin and in close proximity to
an area of strong negative relief (Cape
Trough).

From the vector strengths (Figs. 6.2 and
6.3) it is apparent that the directions are
more randomly distributed in the western
part of the basin. Sandstones are usually very
fine-grained and contain much clayey mat­
terial. Most of the directional structures, such
as abundant micro-cross-bedding, indicate
that the velocity of the sediment laden dis­
tributing streams was low (lower part of the
lower flow regime). Shale or mudstone are
the predominant rock types. However, at
certain locations in this part of the basin
sections include an appreciable thickness of
arenaceous material and cross-bedding is
present in some areas, indicating more rapid
flow. Further examination of this part of the
basin is required to unravel the finer details
of sedimentation.

The western part of the Beaufort Basin
probably represents an embayment where
quiet conditions existed. Apparently the sub­
sidence was more active in the eastern and
northern parts of the basin along a line strik­
ing approximately north-northeast.

At a certain stage in the history of the
basin, previously thought to be at the close
of the Permian, tectonism resulted in the ex­
posure of intrabasinal Precambrian source
areas in the northern part of the basin. The
deposition of the Mooi River Formation fol­
lowed. The tectonic activity may have been
in the form of folding or faulting, though
little evidence was found to support the first
possibility. This event appears to have oc­
curred in the Late Permian as fossils belong­
ing to the old *Cistecephalus* Zone have been
found in this coarse-grained sequence. The
names *Daptocephalus* or *Whaitsia* have been
proposed by Kitching (1970: 310) for this
zone. In the extreme north the fossil evid­
ence is somewhat inconclusive as fossils be­
longing to this and the *Lystrosaurus* Zone
have been found at different locations in the
area underlain by the Mooi River Formation.

An area of low vector strength in the
eastern and northern parts of the basin is
thought to represent the positions of the
source areas providing the detritus of the
Fig. 6.3. Areas of high vector strength in the Beaufort Group. The lower limit of this class was chosen as having a vector strength of 80 per cent. The limit was chosen arbitrarily.

Mooi River Formation. Theoretically an area of low vector strength may be obtained where centripetal, radiating or otherwise opposed drainage patterns are present, or in an area with a very low palaeoslope. As the sedi-
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ments found in this area of low vector strength are coarse-grained the latter possibility may be excluded. The suggestion that the area of low vector strength is a high in the pre-Karroo floor is based also on other sedimentological considerations.

Grain Size

The presence of large clasts in the Mooi River Formation is suggestive of the local source areas discussed above.

In the south (East London area) the presence of pebbles in the Katberg Sandstone indicates that a proximal area in the southeast was being actively denuded during the period of sedimentation of this unit. The different rock types indicate that varied rock types were exposed in the provenance area. The predominance of quartzite pebbles does not necessarily indicate that this was the dominant rock type, but is rather an indication of its relative resistance to abrasion during transport.

The presence of gneiss and fossilised wood (often carbonaceous) is of particular interest. According to Krumbein and Sloss (1963: 425) coal may be formed on unstable shelf areas. The fissile nature of the fossilised wood clasts, as well as their poor rounding and relatively low hardness, is evidence that these fragments were not transported over a great distance. A landmass with its margin not far south of the present coastline near East London was apparently being rapidly eroded at this time. The stripping of coal beds is evidence of the upward (epeirogenic?) movement of this landmass.

It is suggested that the entry point of a large river, draining this southern landmass, was close to the present African continental margin. A high palaeoslope could not have been maintained over any appreciable distance from the point of entry, as the sedimentary fill contains too high a proportion of sand to have become sufficiently consolidated to maintain a high slope.

The presence of gneiss is important from the point of view that this is the predominant rock type of the floor of the Karroo Basin, as is indicated by the clasts in the Mooi River Formation.

Heavy Minerals

The close resemblance between the assemblages of heavy minerals present in the arenaceous members of the Mooi River Formation and that of the formations derived in the south is remarkable. As mentioned earlier, the ratio of garnet varies greatly, probably reflecting the degree of maturity attained by the sediments. The writer (Theron, 1965) has demonstrated that there is little variation in the composition of the assemblages of the sandstones derived in the south, which suggests that a state of equilibrium was achieved. Similar results were later obtained by le Roux (1969).

It is concluded that the source rocks in the south are comparable to those supplying the material of the Mooi River Formation, where garnetiferous gneiss appears to have been the principal source rock.

Geographical Setting

From the evidence presented here it is evident that a vast amount of sediment was transported past a mobile entry point south-east of the present Beaufort outcrop. There is a similarity between this postulated drainage dispersal system and that of the Brahmaputra River described by Coleman (1969). Comparison of the volume of sediment in the Beaufort and that deposited by the Brahmaputra indicates that a very large provenance area once existed south of the present limit of the African continent.

The evidence of the presence of a source area to the southeast of the present coast, its close relationship with the basin and its inferred vast size precludes the possibility that in Beaufort times the Cape Fold Belt could have been extended far east of its present extremity. This is also corroborated by the evidence that the amplitude of the folds appears to decrease from a position to the north of Port Elizabeth in an easterly direction. Furthermore the Karroo formations, as well as the southern boundary of the Karroo dolerites, appear to strike southwards around the eastern limit of the folds. If this source was active from as early as the Carboniferous it is unlikely that the Samfrau Geosyncline ever existed in the position du Toit depicts it (1937: 80). The reality of the
Gondwanide Orogen as shown by du Toit (p. 93, fig. 11) is also open to the same criticism.

The feldspathic nature of the Katberg Sandstone and also the Molteno in the east and southeast (Botha and Theron, 1967) shows that the area of positive relief was granitic. This would hardly have been the case if the sediments were derived from the Gondwanide foldings.

Several studies have indicated that the maximum thickness of the Karroo strata is attained in the so-called Natal Trough. The conclusion is drawn that this trough was actively subsiding during Beaufort (and Stormberg) sedimentation, but that the Cape Trough (a part of du Toit’s Samfrau Geosyncline) was relatively immobile or beginning to deform at the time. Evidence for this is the unimportant position of the western part of the basin in the dispersal system.

**RECONSTRUCTION OF THE SOUTHERN LANDMASS**

Metamorphic rocks, lying below the rocks of the Cape Super-Group, are exposed along the eastern margin of the African continent and were described by McIver (1966). He states that the oldest rocks present are pyroxene-garnet granulite and garnet-biotite schist. Other rocks present are hypersthene-quartz diorite, granodiorite and various granites, gneisses and granulite. His description of the various rocks shows that garnet is a common accessory mineral in the rocks of the southern Natal 1000 m.y. granitic basement. Dryden and Dryden (1946: 92) have shown that garnet weathers readily, and Pettijohn et al. (1972: 305) indicate that garnet is only moderately stable, but that the stability of heavy minerals is affected by a large number of variables. Rapid denudation and a suitable climate might account for the high garnet content present in the rocks of the Beaufort Group. According to Stavrakis (pers. comm.) and McIver (1966) pegmatites are also present in the Precambrian rocks of Natal.

It appears that the Precambrian rocks exposed in Natal and those postulated to be present in the sub-Karroo surface to provide material for the Mooi River Formation, were similar. The heavy minerals of the southern source also suggest similarity to the Mooi River Formation. The Precambrian rocks in Natal, the Orange Free State and the source in the south possibly formed part of the same surface at the time the Beaufort was deposited.

**CONCLUSIONS**

1. Palaeocurrent evidence indicates that the major source of the detritus supplied to the Beaufort Basin lay to the southeast of the basin.

2. The southern source provided material with a composition similar to the sediments provided by source areas within the northern extensions of the basin. Continuity of the pre-Karroo floor southwards is postulated.

3. The southern provenance lay close to the present continental margin, suggesting that the Samfrau Geosyncline did not exist in this area at that time.

4. The southern source provided sediments over a long period of time and probably was of large dimensions.

5. In a reconstruction of Gondwanaland a large area probably consisting predominantly of Precambrian rocks should be placed to the southeast of Africa.

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Section 2

Gondwana Flora
Palaeographic Implications of Some Silurian-Early Devonian Floras

HARLAN P. BANKS

ABSTRACT
Late Silurian (Pridolian) and Early Devonian floras are analysed on a worldwide basis using Cooksonia, the oldest proven land plant, and members of the subdivision Zosterophyllophytina (Division Tracheophyta) as examples. The latter group has yielded more anatomical and morphological detail over a wider range of Earth’s surface than any other group yet documented in the Devonian. Salient details are illustrated. Zosterophylls extend from Australia to the Canadian Arctic, U.S.A., eastern Canada, Spitsbergen, Britain, western Europe, Poland, and western Siberia. On the basis of their uniformity over this wide range, petrographic studies of rocks around the present North Atlantic, detailed sedimentological studies in the same area, and similarity of preservation of the plants in these sediments, the plants lend support to the hypothesis of a single pre-Atlantic land mass which has subsequently been split apart by continental drifting.

INTRODUCTION
Some of the more enthusiastic supporters of plate tectonics have suggested that continental drift is now proved and there is no need to attempt to deduce the former positions of continents from the distributions of fossil organisms. Nevertheless it may be instructive to examine the distribution of early land plants as a test of the hypothesis that continents in Devonian time were juxtaposed. Assuming the truth of the hypothesis, some projection of the position of land masses and epicontinental seas may result. The following account is a look at selected Early Devonian floras from the viewpoint of a worker in the northern hemisphere.

Two of the earliest papers on Devonian plants were those of Göppert (1852) describing the lycopod Drepanophycus spinaeformis and Dawson (1859) on Psilophyton princeps. In 1871 Dawson added P. princeps var. ornatum. It remained for a series of five papers by Kidston and Lang between 1917 and 1921 to establish the real significance of this early land flora. The wide impact of Kidston and Lang’s contribution resulted from the completeness of the anatomical and morphological detail revealed by sections of silicified fossils of Rhynia, Horneophyton, Asteroxylon and associated non-vascular plants. The authors erected (1917) the new order Psilophytales for leafless plants with terminal sporangia. Thenceforth Devonian floras and psilophytes became almost synonymous in the minds and writings of many workers. The obvious result was that despite occasional publications in which distinctly different species were described in considerable detail, the group psilophytes became a repository for numerous totally unrelated organisms. Only within the past decade has order begun to emerge from this chaos. The emergence of order is a major determinant in the selection of material to be presented in this account. I want specifically to show the kind of work that is being done in order
Fig. 7.1. \textit{Gosslingia breconensis}. Reconstruction of plant with its lateral sporangia and circinate apices. Axillary tubercles indicated by black spot below some branchings. \(x c. 3/8\). (From Edwards, 1970a.)

Fig. 7.2. \textit{Cooksonia caledonica}. Reconstruction of dichotomising axis with terminal sporangia. \(x c. 1 1/8\). (From Edwards, 1970b.)

Fig. 7.3. \textit{Zosterophyllum llanoveranum}. Reconstruction of a longitudinal section of sporangium showing thickened margins of valves and depression between them. \(x c. 70\). (From Edwards, 1969a.)

Fig. 7.4. \textit{Zosterophyllum mryetonianum}. Reconstruction of plant with spikes of radially arranged, lateral sporangia and basal, H-type branching. \(x c. 1/2\). (Redrawn from Walton, 1964.)

Fig. 7.5. \textit{Rebuchia ovata}. Reconstruction of plant showing spikes of sporangia that are bilaterally arranged. \(x c. 1/2\). (From Hueber, 1972.)

Pl. 7.I. Fig. 1. \textit{Cooksonia} sp. from New York State. \(x 34\). Fig. 2. \textit{Cooksonia pertoni} from Wales. \(x 4\). Fig. 3. \textit{Cooksonia hemisphaerica} from Wales. \(x 6\). Fig. 4. Spore of \textit{Cooksonia}. \(x 1200\). Fig. 5. Film pull of epidermal cells of \textit{Cooksonia}. \(x 150\). Fig. 6. Film pull of cortex of \textit{Cooksonia}. \(x 140\). Fig. 7. Film pull of tracheids of \textit{Cooksonia}. \(x 370\).
to bring about orderliness and to produce plants and floras that can be analysed in a meaningful way, and whether to test hypotheses or simply to portray the course of evolution of land plants.

EARLIEST DEMONSTRABLE VASCULAR LAND PLANTS—PRIDOLIAN

I have reviewed recently the distribution of *Cooksonia* and related organisms in the Pridolian (Banks, 1972). In brief, Lang (1937) erected the genus *Cooksonia* for plants found throughout the Downtonian (= Pridolian, Upper Silurian) of Wales and the Welsh borderlands (Pl. 7.I, figs. 2-3). He demonstrated its epidermis, hypodermis, a vascular strand of annular tracheids, and trilete spores (Pl. 7.I, figs. 4-7). Lang’s paper is an example of the kind of investigation that I consider a major contribution to our knowledge and understanding of a taxon. Such detail is essential for plants that are to be used in palaeogeographic or comparative studies. Lang applied a series of techniques to his material and extracted the maximum of data. In contrast to the Welsh *Cooksonia*, other reports of the genus in Pridolian strata are based wholly on gross morphological appearance. Histological details are completely lacking. Obrhel (1962) reported *C.* sp. cf. *C. hemisphaerica*, *C.* sp. and ?*C.* sp. from Czechoslovakia. Ishchenko (1969) described *C. pertoni* and *C. hemisphaerica* from Podolia. Banks (1972, 1973) reported *Cooksonia* sp. from New York State (Fig. 7.1). Recently Daber (1971) has described what he considers to be a *Cooksonia* from the Pridolian of Libya. Preservation is poor and the attribution of the single specimen to *Cooksonia* can be regarded as uncertain. If all these reports are accepted as valid, the oldest vascular land plant flora ranged from North America to the U.S.S.R. and perhaps extended to North Africa, all in Pridolian time. A second valid genus of vascular plant in the Pridolian is *Steganotheca* Edwards (1970b). It, however, is found only in Wales and does not yet add to a discussion of worldwide distributions. It does contribute to variation in the Pridolian flora.

References to earlier occurrences of macrofossils of vascular land plants are here ignored either because the material may not be plant at all or because no tissues whatsoever have been demonstrated that would suggest vascular plants. Reports of microfossils such as isolated individual spores bearing trilete markings may refer to spores of bryophytes rather than vascular plants. Up to now we have no simple criterion to distinguish the spores of one group from the other. However, detailed analyses of the spores in a vertical succession of strata such as that by Richardson and Lister (1969) in the Siluro-Devonian (Wenlockian to Siegenian) of the Welsh Borderlands and South Wales do suggest that spores like those of vascular plants were produced before Pridolian time. One explanation for the apparent discrepancy between evidence from spores and evidence from macrofossils on the time of first appearance of fully demonstrable vascular land plants is given by Banks (1972). Basically he suggests the possibility of independent origin of the several characters associated with vascular land plants—cuticle, epidermis, stomates, trilete spores.
and vascular tissue. These several characteristics may have been ‘combined’ into a single organism only in Pridolian, or perhaps Ludlovian time. Another kind of fossil found in Pridolian and older Silurian strata is flattened coalified remains from which cuticle-like sheets of cells, elongate tubes often with internal spiral or annular thickening, and spores can be demonstrated (Nematophytales of Lang, 1937). No one considers these remains to be vascular plants but they do simulate some characteristics of vascular plants. One also finds in macerations of rock both sheets of cells or cell outlines and short lengths of tubes with internal thickening. In my opinion both might originate from a Nematothallus-like organism or might come separately from some totally different source. They should not be regarded as evidence of vascular plants older than Pridolian time unless the organism whence they originated is demonstrated.

The subject ‘Origin of Vascular Plants’ may now be clarified by distinguishing two aspects of the problem. The more obvious is the determination of the time of appearance of the first organism so well preserved that it demonstrates several of the requisite characteristics and is acceptable to any purist as a true vascular plant. To me, as one purist, Cooksonia in the Pridolian of Wales is such a plant (Pl. 7.I, figs. 2-7). The second problem concerns the evolutionary origin of source and development of that first vascular organism. This is the more complex of the two problems. Many of us still assume an origin of vascular plants from filamentous green algae but we are completely ignorant of actual stages of this transition in the fossil record. Thus the interpretation of fragments of plants such as cuticle-like remains, spirally thickened tubes, and trilette spores will continue to be the subject of controversy. Possibly only newer biochemical, biophysical, electronmicroscopic or other techniques will ultimately produce a solution.

Cooksonia continued to live at least through Emsian time in such disparate places as the Gaspé Peninsula, eastern Canada (Hueber, 1964a), Arizona, U.S.A. (Cartright, 1970), Wales (Croft and Lang, 1942), Scotland (Edwards, 1970a), central Kazakhstan (Yurina, 1964, 1969) and western Siberia (Anan’ev, 1960; Anan’ev and Stepanov, 1969).

Cooksonia consists of a naked axis that forks dichotomously an unknown number of times and bears single, terminal sporangia. Its basal parts are unknown. Anan’ev and Stepanov (1969) have attempted a reconstruction in which they show a straight, slender main axis bearing spirally arranged lateral branches each of which dichotomises a few times. All tips end in sporangia. Edwards (1970b) has reconstructed Cooksonia caledonica Edwards without a main axis but only successive dichotomies (Fig. 7.2). Her drawing is somewhat more in accord with specimens I have seen and fits Kidston and Lang’s description of Psilophytales perfectly. This means that the earliest vascular land plants are truly simple plants and that Rhynia, now considered to be of Siegenian-Emsian age, may very well represent the most perfectly preserved of the primitive land plants rather than a morphologically reduced...
organism as Scott (1920: 422) suggested. *Cooksonia* may just settle the prolonged argument over primitiveness versus reduction in the discussion of *Rhynia*.

**DEVONIAN PLANTS**

In 1892 Penhallow described some axes from the Gedinnian of Scotland that earlier were thought to resemble leaves of the marine angiosperm *Zostera*. Little attention was paid to his genus, *Zosterophyllum* Penhallow 1892, until Lang (1927) reinvestigated the original material. As I emphasised when discussing Lang's study of *Cooksonia*, he must here also be given credit for extracting maximum details from his specimens. He demonstrated cuticle, the outlines of some epidermal cells including possible stomata, possible hypodermal cells, tracheids, the reniform shape of sporangia (Pl. 7.IV, fig. 33), dichotomous and H-type of branching (Fig. 7.4), the lateral position of (?) sporangia, and the tufted habit of growth. Lang failed to obtain spores from the structures he believed to be sporangia but he did recognise that folding of these reniform bodies during compression could result in the varied shapes sometimes encountered. *Zosterophyllum* thus was a plant of tufted habit, branching profusely near its base, and with sparsely branched, erect axes up to 15 cm high. The latter bore spirally arranged lateral bodies (now proven to be sporangia) of reniform shape and with strikingly thickened convex margins (Pl. 7.IV, fig. 33). The basal portions of the plant bore no stomates, a point demonstrated convincingly by Lele and Walton (1961) over thirty years later.

Today it would seem that Lang's analysis of *Zosterophyllum myrtonianum* would have sufficed to convince morphologists of the existence of a second distinct type of early vascular land plant, one with lateral sporangia as opposed to the first type which bore terminal sporangia, the Psilophytales. It did not, and the group known as psilophytes was simply broadened a little. This story is typical of the whole history of early land plants and is one reason for selecting *Zosterophyllophytina* Banks, 1968, as the major subject for discussion here.

By the mid-1930s one was beginning to see the outline of three Devonian floras called by Kräusel (1937) the *Rhynia*, the *Hyenia* and the *Archaeopteris* floras of Lower, Middle and Upper Devonian strata respectively. Leclercq (1940) used the terms *Rhynia*, *Protopteridium*, and *Archaeopteris* floras. But, it must be emphasised, few of the genera recognised in Devonian floras were understood in significant detail. Many were only fragments of plants. Most were based on gross morphological appearance alone. For example, *Archaeopteris* was then recognised as a fern whereas it is now treated as a progymnosperm (Beck, 1960, 1962; Carluccio, Hueber and Banks, 1966) because it has been found attached to the stem *Callixylon* whose anatomy is gymnospermous. For every genus of Devonian plant known as well as *Archaeopteris* now is, there are several for which almost no data are available, and for every quarry or area from which plants with these details have been derived there are dozens of collections around the world for which details have not been sought. In some cases the details exist if one has the patience, skill, and time to get them and permission to study the specimens.

Additional problems are encountered when attempting to analyse the stratigraphic occurrence and precise age of floras on different continents whether currently or through the literature of the past. For example, the
Rhynie chert, once regarded as Middle Devonian, is now thought to be Siegenian or Emsian, the Downtonian containing Lang's Cooksonia is now Upper Silurian rather than Lower Devonian (McLaren, 1972), and the Australian Baragwanathia-containing beds have been raised up from Silurian to Siegenian-Emsian levels. One has high hopes for
the success of students of *Sporae Dispersae* in markedly improving intercontinental stratigraphical correlations. I think this effort is already well under way. It is significant work because the best fossil plant remains occur in continental strata where useful marine index fossils are unavailable. Spores may well supply the badly needed data. Yet lest I give the impression that spores are the one final answer, I must point out that range of variation of spores in one sporangium of one species throughout its ontogeny is unknown. When therefore a person asks why there are so many species and genera of spores in the Devonian when the number of genera of macrofossils is so small, the reply might be that many species of spore (named from *Sporae Dispersae*) are the product of a single species of macrofossil. Opportunity to test such a hypothesis is a rare occurrence. But it can happen (Banks, Bonamo, and Grierson, 1972). *Leclercqia complexa* described therein does present wide variation in the morphology of spores preserved in situ in sporangia. The work is currently being carried further (Bonamo and Grierson, 1973).

Equally interesting is the occurrence of one genus of *Sporae Dispersae* in unrelated macrofossils, for example, *Retusotriletes* in sporangia of *Zosterophyllum* sp. cf. *Zosterophyllum sp. fertile* (Edwards, 1969b) and in *Psilophyton princeps* (Hueber, 1964b). In spite of any drawbacks palynology appears capable of contributing much to the precision of work on Devonian floras.

Thus although one could readily paint a broad picture of worldwide Devonian floras, essentially uniform in generic content, a view often expressed in the literature, I prefer to dwell on one group of plants. This category, *Zosterophyllophytina*, has become markedly strengthened since its erection (Banks, 1968b), is worldwide in its distribution, occurs in many areas where stratigraphy is moderately well understood, and is replete with botanical detail. Comparable data are becoming available for some other Devonian plants for example, the progymnosperms (Beck, 1970a, 1970b, 1971; Bonamo and Banks, 1967; Matten and Banks, 1966, 1967; Matten, 1968; Scheckler and Banks, 1971a, 1971b) but for none of these are the data available over the wide geographic range covered by the zosterophylls. Individual genera too have been worked out in great detail but usually they have been studied at one locality only. Comparisons of them with plants on other continents depend on gross morphological similarity. The question then becomes 'what Devonian plants continue the palaeogeographic story started by *Cooksonia* in the Silurian?'

**Zosterophyllophytina—A second type of early vascular land plant**

It took from the time of Lang's (1927) careful analysis of *Zosterophyllum myretonianum* until 1968 (Banks, 1968b) to break up the fossil representatives of Psilopsida formally into separate major categories. Important indications that this must happen were seen in Leclercq (1954) and Hueber (1964b). The former pointed out that two sporangial positions, terminal and lateral, existed among the then psilophytes. The latter added the observations that terminal sporangia tended to be fusiform and to dehisce longitudinally whereas the lateral sporangia were globose to reniform and dehisced distally. Banks (1968b) added the character of centric maturation of xylem in the plants with terminal sporangia and exarch maturation in those with lateral sporangia. As a result he (Banks, 1968b; Banks and Davis, 1969) erected three major subdivisions of vascular plants including some of the many genera previously considered to be Psilopsida or Psilophytales. These subdivisions of Tracheophyta are Rhyniophytina including such genera as *Cooksonia* and *Rhynia*; Zosterophyllophytina including *Zosterophyllum* and related forms; Trimerophytina including *Dawsonites*, *Trimerophyton*, *Psilophyton* and *Pertica*. The complete anatomical structure of *Psilophyton* is now known (Banks, Leclercq and Hueber, in press).

*Zosterophyllophytina* as originally published (Banks, 1968b) and as modified (Banks and Davis, 1969) has been treated variously by Edwards (1970a) and by Hueber (1972). Both accounts contain valuable information and points of view. Currently the group includes *Zosterophyllum*, *Sawdonia* (syn. *Psilophyton princeps* var. *ornatum*),
**Accounts Based on Histological Detail**

In view of my earlier comments about the meticulous work of Lang (1937) on Cooksonia and on Zosterophyllum (1927) from Scotland, I should mention a dozen other papers that document vital details of some of these genera and which therefore go far toward making Zosterophyllophytina a substantial and a natural group. Only representative characters of the group will be illustrated. Subsequent to the work of Lang, Lele and Walton (1961) documented cuticle, epidermal cell patterns, stomates, branch traces and annular tracheids of Zosterophyllum myretonianum. Their confirmation of the absence of stomates below and their presence above on aerial branches suggested anew that the plant inhabited marshy areas. Edwards (1969a) then demonstrated for Z. llanoveranum from Wales a circular, exarch xylem strand in vegetative axes and an elliptic strand in fertile axes (cf. Pl. 7.II, fig. 21), a hypodermal cortex, scalariform tracheids and highly modified cells around the distal, convex surface of the sporangium along the line of dehiscence (Fig. 7.3). She also obtained spores out of sporangia. They bore some characteristics of each of the two genera of Sporae Dispersae, Punctatisporites and Retusotriletes. Edwards (1969b) studied Z. sp. cf. Z. fertile Leclercq 1942 from Wales. Here she again found an exarch, terete or elliptical protostele, a hypodermal cortex, a dehiscence zone probably similar to that of Z. llanoveranum and spores in situ referable to Retusotriletes. She also found the sporangial wall to be composed of two zones, an outer zone of hypodermal tissue like that of the stem and an inner zone of flattened cells.

The story for Sawdonia ornata Dawson Hueber 1971a (syn. Psilophyton princeps var. ornatum Dn. 1871) is longer but basically similar (Pl. 7.II, figs. 8, 9). Not until 1924 when W. N. Edwards macerated speci-
spines, cuticle, epidermis, stomata, hypodermis (Pl. 7.II, figs. 10, 14, 16, 17) and tracheids that earlier workers had seen. She studied stomata especially and found that guard cells are at the same level as other epidermal cells but are ridged over by cuticular extensions. These features probably led Edwards (1924) to conclude that guard cells are sunken. No one has found anticlinal walls between the ends of two guard cells. This suggests the failure of a mother cell to divide during ontogeny into two guard cells. The apparent stomatal apparatus may consist therefore of a single cell which may or may not have been able to open and close when living. Precisely the same situation holds in *Zosterophyllum*.

*Gosslingia* (Pl. 7.III, fig. 20) was erected by Heard (1927) for plants lacking leaves and roots, possessing stomata and hairs, branching dichotomously and by unequal dichotomy, and bearing lateral sporangia on specialised branches which he thought arose near the axil of a dichotomy. Scars often called axillary tubercles occurred near each branching (Pl. 7.III, fig. 24). These represented the point at which Heard thought the fertile branches were attached. Heard prepared polished and etched surfaces of pyritic petrifactions and found a hypodermal cortex and an elliptic, exarch protoxylem strand composed of spiral and reticulate tracheids. Edwards and Banks (1965) briefly extended Heard's anatomical observations and Edwards (1970a) published a thorough analysis of the anatomy of *Gosslingia* demonstrating that the xylem strand of an axis branches twice successively, the first branch continuing into a lateral and the second supplying the axillary tubercle (Pl. 7.III, fig. 27). She also demonstrated spores in the sporangia and a thickened outer rim on the convex margin of sporangia. It seems to be characteristic of several zosterophylls. Edwards's reconstruction of *Gosslingia* is seen here in Figure 7.1.

*Crenaticaulis* Banks and Davis (1969) is the fourth genus of zosterophylls for which an abundance of detail is available. It bears two rows of teeth on opposite sides of the stem and papillae on some epidermal cells (Pl. 7.III, figs. 23, 28). It has a hypodermal cortex (Pl. 7.III, fig. 26) and an elliptic, exarch protoxylem like those of *Zosterophyllum, Sawdonia, and Gosslingia*. Its lateral, reniform sporangia (Pl. 7.IV, figs. 29, 31) dehisce distally into two unequal segments (Pl. 7.III, fig. 32), and branches are actually attached in the position of axillary tubercles (Pl. 7.III, figs. 25, 28).

**Accounts Based Primarily on Gross Morphology**

Once it is established that there was a natural group of organisms (Zosterophyllphyta) based on a range of characteristics one can evaluate more satisfactorily the other papers that extend the geographic and stratigraphic range, the morphologic variation, and the number of species in the group, but lack much evidence of histologic detail.

Anan'ev and Stepanov (1968) illustrated *Sawdonia ornata* with attached lateral sporangia under the name *Psilophyton princeps* (Pl. 7.IV, fig. 30). *Rebuchia ovata* (Dorf) Hueber 1970 (syn. *Bucheria ovata* Dorf 1933) can now be recognised as another taxon among zosterophylls (Hueber, 1972). Hueber has described the vegetative as well as fertile parts of the plant, confirmed the position and arrangement of its sporangia (Pl. 7.IV, fig. 34, Fig. 7.5), extracted unornamented spores, and determined that sporangia dehisce distally. *R. ovata* occurs in the State of Wyoming, U.S.A., and in Germany. Hueber (1971b) has erected a second species, *R. capitanea*, for elongate spikes of reniform sporangia with distal, basipetal dehiscence found on Bathurst Island in the Canadian Arctic. Other reports of *Rebuchia (Bucheria)* deal with plants that are better classified elsewhere (Hueber, 1972), for example, the *Bucheria* with a terete xylem strand reported by Lepekhina, Petrosyan and Radchenko (1962).

*Bathurstia denticulate* Hueber (1971b) is another taxon based on compressions. It too was found on Bathurst Island. It has an elongate axis with occasional dichotomies, hair-like teeth, a compact, two-rowed spike, a cuticle, elongate reniform sporangia and circinate apices. *Bathurstia* may be related most closely to the subgenus *Platyzosterophyllum*. 

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*Gondwana Flora*
Species of Zosterophyllophytina

Gosslingia, Bathurstia and Crenaticaulis are monotypic genera. The two valid species of Rebuchia (Hueber, 1972) are mentioned above. Sawdonia is more of a problem. S. ornata was described as a Psilophyton, P. princeps var. ornatum Dawson 1871. In fact it was held by some workers to be the type species of that genus (cf. Hueber and Banks, 1967; Anan'ev and Stepanov, 1968). The selection of a neotype for Psilophyton princeps and the recognition of an exarch protostele and lateral sporangia on specimens of the variety (Hueber and Banks, 1967) along with a more elaborate description of P. princeps itself (Hueber, 1968) made it possible to set the variety aside in a new genus as Sawdonia ornata (Dawson) Hueber (1971a). It is now clear that either preserved anatomical structure or attached sporangia is required to make a positive separation between the two genera Sawdonia and Psilophyton. No monographic treatment of the numerous species ascribed to the genus Psilophyton is yet available. Therefore the present account is limited to specimens (formerly considered to belong to Psilophyton) for which histological data are available. All are mentioned above. Sawdonia is monotypic.

Zosterophyllum, in contrast, includes several valid species. Differences are chiefly in arrangement and distribution of sporangia on the fertile branches. Zosterophyllum myrotonianum Penhallow 1892, in addition to localities given above, has been found in the U.S.S.R. in western Siberia (Anan'ev, 1959, 1960). The species has been reviewed by Walton (1964) who drew a reconstruction to replace the classical one of Kräusel and Weyland (1935). Walton shows a tufted (Fig. 7.4) plant that might occupy a muddy environment whereas Kräusel and Weyland depicted a fully submerged aquatic. The position of stomata as shown by Lele and Walton (1961), the terete shape of petrified specimens, the thick hypodermal cortex, and the terete to elliptic xylem strand (Edwards, 1969a, 1969b) remove all support from the concept of Zosterophyllum as an aquatic. So far as I can tell that hypothesis was based only on the fact that compressed stems were flattened, ribbon-like, and resembled the leaves of the marine angiosperm Zostera. Petrified stems, on the contrary, are terete in cross-section.

Zosterophyllum australianum Lang and Cookson 1930 was erected for tight 'spikes' of spirally arranged sporangia found in the Centennial beds of the Walhalla Series, Victoria, Australia. Cookson (1935) found unattached H-branched vegetative axes associated with fertile axes. Yurina (1969) reported the species near Karaganda in Kazakhstan and Croft and Lang (1942) found it in Wales as Zosterophyllum sp. cf. Z. australianum.

Zosterophyllum rhenanum Kräusel and Weyland 1935 was described from late Siegenian strata at Münchsecke in the Wahnbach in the Rhineland, Germany. Hueber (1972) considers it to be close to Rebuchia ovata, Edwards (1969a) believes it may be synonymous with Z. myrotonianum, and Hoeg (1967) accepts it as a valid species. Yurina (1969) has found it in Kazakhstan. Anan'ev (1959, 1960) has found it in western Siberia.

Zosterophyllum minor Anan'ev 1960 was found in the Altai-Sayan Mountain area of western Siberia. It is well illustrated in Høeg (1967: 262).

Zosterophyllum longum (Høeg) 1967 was originally described by Hoeg (1942) as a species of Bucheria from the Lower Devonian of Spitsbergen.

Zosterophyllum fertile Leclercq 1942 came from Nonceveux in the Belgian Ardennes from upper Gedinian strata. The sporangia appeared to be in two rows but Leclercq demonstrated that they were inserted spirally. The species has been found in the Gedinian of Scotland (Edwards, 1972) and Z. sp. cf. Z. fertile was described from Wales by Edwards (1969b). Edwards in the same paper described the anatomy, dehiscence, and spores of this species (see above).

Zosterophyllum llanoveranum Croft and Lang 1942 was recognised as a distinct species because its sporangia appeared to be in two rows rather than arranged spirally. The authors therefore erected a subgenus Platyzosterophyllum for it and suggested that species with radially arranged sporangia be
placed in the subgenus *Euzosterophyllum*. Hueber (1972) has revised the latter to subgenus *Zosterophyllum* to conform to International Rules. Edwards (1969a) added much to the understanding of this species by demonstrating its anatomical structure, its dehiscence mechanism, and its spores (see above). She also reported uncovering sporangia with needles and a dissecting microscope in such a way as to demonstrate a spiral rather than a strictly two-rowed arrangement. Continued efforts of this type could lead to abandonment of the two subgenera. One may continue to be faced, however, with hand specimens some of which appear to be bilaterally arranged. Each specimen will then need careful uncovering to determine what is real and what is artifact of preservation or of ontogenetic twisting during life. Edwards (1972: 82) made this point more obvious when she reported seeing specimens of *Z. myretonianum* whose sporangia seem to be in a single row. This is a species normally regarded as a clear-cut case of spiral arrangement. Anan’ev (1959) has reported *Z. llanoveranum* from the Altai-Sayan Mountain region in western Siberia.

*Zosterophyllum artesianum* Danzé-Corsin 1956 was found in the Lower Devonian of the Pas-de-Calais area, France. Banks (1968a) considered this species to be a strange zosterophyli, possibly intermediate between the Rhyniophytina and the Zosterophyllophytina. Its reniform sporangia and distal dehiscence along an obviously thickened convex margin recall zosterophylls. Its long-stalked sporangia look more like those of some rhynia-type. Edwards (1969a) thought *Z. artesianum* resembled her *Cooksonia* from the Gedinnian of Scotland which was described subsequently (Edwards, 1970b). Høeg (1967) and Hueber (1972) both treat the species as a member of *Platy-zosterophyllum* without comment.

**Geographic and Stratigraphic Occurrences**

Zosterophyllophytina range widely around the globe (Table 7.1). They are found in the Canadian Arctic Islands, southern Canada, and the U.S.A., across Spitsbergen, Britain and western Europe to eastern Europe and around to Siberia. *Zosterophyllum* and *Sawdonia* are the two best documented genera and of these *Sawdonia* is known in greater detail from a greater number of widely separated areas. Edwards’s (1924) first report of the details of its epidermal surface and Lang’s (1931) discovery of its spines both dealt with specimens from the type locality on the Gaspé Peninsula, Province of Quebec, Canada. Lang followed (1932) with identical details of Scottish material. Then came Hueber and Grierson’s (1961) demonstration of its persistence into Upper Devonian strata in New York State. They succeeded in again proving histological identity. Hueber (1964b) followed with macerations of similar spines, epidermal cells, stomates and, newly discovered, lateral sporangia on specimens from the James Bay Lowlands at the southern end of Hudsons Bay, Ontario, Canada. When illustrations of these specimens are published there will remain no question of their authenticity. In the same paper Hueber also reported the discovery of a fertile specimen of *Sawdonia* with lateral sporangia at Dawson’s type locality on the Gaspé Peninsula. Anan’ev and Stepanov (1968) completed this demonstration of sporangia by illustrating beautiful specimens of fertile *Sawdonia* (their *Psilophyton princeps*) from the Altai-Sayan Mountain region of western Siberia. But Zdebska added the final touch by macerating *Sawdonia* from the Lublin Uplands of eastern Poland and duplicating once again the epidermal, hair, stomatal, and tracheidal pattern found at James Bay, New York, Gaspé and Scotland. I have studied Zdebska’s material in Poland and marvelled both at its beauty and its identity to western material. Here we have a Devonian taxon that can be documented with reasonable completeness, and that extends over a vast range of Earth’s surface. I have no doubt that additional studies carried out with equal care could show this species to be even more widely distributed. Perhaps enough has been said to indicate that one taxon, *Sawdonia ornata*, and one subdivision of vascular plants, *Zosterophyllophytina*, suggest a striking uniformity in the flora of the lowlands in Devonian time.
Table 7.1. Geographic and Stratigraphic Occurrence of a Selection of the Best Documented Specimens of Zosterophyllphytina in Devonian Strata

<table>
<thead>
<tr>
<th>Stage</th>
<th>Locality</th>
<th>Plants</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gedinnian</td>
<td>Scotland</td>
<td><em>Zosterophyllum myrelonianum</em> Penhallow 1892</td>
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<tr>
<td></td>
<td></td>
<td>(also Lang, 1927; Lele and Walton, 1961)</td>
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<tr>
<td></td>
<td></td>
<td><em>Z. fertile</em> (Edwards, 1972)</td>
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<tr>
<td></td>
<td>Spitsbergen</td>
<td><em>Zosterophyllum sp.</em> (Hoeg, 1942)</td>
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<tr>
<td></td>
<td>Belgium</td>
<td><em>Z. fertile</em> Leclercq 1942</td>
</tr>
<tr>
<td></td>
<td>Canada-Arctic, Bathurst Island</td>
<td><em>Rebuchia capitanea</em> Hueber 1972 (in 1971b)</td>
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<tr>
<td></td>
<td></td>
<td><em>Bathurstia denticulata</em> Hueber 1972 (in 1971b)</td>
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<tr>
<td></td>
<td></td>
<td><em>Sawdonia ornata</em> (Dawson) Hueber 1971a</td>
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<tr>
<td></td>
<td>France</td>
<td><em>Zosterophyllum artesianum</em> Danze-Corsin 1956</td>
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<tr>
<td></td>
<td>Germany</td>
<td><em>Z. rhenanum</em> Krausel and Weyland 1995</td>
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<td></td>
<td>Wales</td>
<td><em>Z. llanoveranum</em> Croft and Lang 1942 (also Edwards, 1969a)</td>
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<td></td>
<td></td>
<td><em>Z. sp. cf. Z. australianum</em> (Croft and Lang, 1942)</td>
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<tr>
<td></td>
<td></td>
<td><em>Z. sp. cf. Z. fertile</em> (Edwards, 1969b)</td>
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<tr>
<td></td>
<td></td>
<td><em>Gosslingia breconensis</em> Heard 1927 (also Edwards, 1970a)</td>
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<td></td>
<td>Australia</td>
<td><em>Z. australianum</em> Lang and Cookson 1935 (also Cookson, 1935)</td>
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<td></td>
<td>Victoria</td>
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<td></td>
<td>Kazakhstan</td>
<td><em>Z. rhenanum</em> (Iurina, 1969)</td>
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<td></td>
<td>S. W. Siberia</td>
<td><em>Z. llanoveranum</em> (Anan’ev, 1959)</td>
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<tr>
<td></td>
<td>Canada</td>
<td><em>Sawdonia ornata</em> (Dawson) Huerber 1971a (also Edwards, 1924; Lang, 1931; Hueber, 1964b; Hueber and Banks, 1967)</td>
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<tr>
<td></td>
<td>Gaspé Peninsula</td>
<td><em>Crenatiaulis verruculosus</em> Banks and Davis 1969</td>
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<tr>
<td></td>
<td>James Bay</td>
<td><em>Sawdonia ornata</em> (Hueber, 1964b)</td>
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<td></td>
<td>Lowlands</td>
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<tr>
<td></td>
<td>Scotland</td>
<td><em>Sawdonia ornata</em> (Lang, 1932)</td>
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<td></td>
<td>E. Poland</td>
<td><em>Sawdonia ornata</em> (Zdebska, 1972)</td>
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<td></td>
<td>Lublin Uplands</td>
<td><em>Rebuchia ovata</em> (Dorf) Huerber 1970 (also Huerber, 1972)</td>
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<td></td>
<td>U.S.A.</td>
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<tr>
<td></td>
<td>Wyoming</td>
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<tr>
<td>Frasnian</td>
<td>U.S.A.</td>
<td><em>Sawdonia ornata</em> (Hueber and Grierson, 1961)</td>
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<td></td>
<td>New York State</td>
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</table>

Zosterophyllphytina appear first in Gedinnian strata, apparently close to the Pridolian boundary in Spitsbergen, continue through Siegenian and Emsian and on into Upper Devonian Frasnian strata. The one acceptable report in the Frasnian is that by Hueber and Grierson (1961) on *Sawdonia* in New York State. Stockmans (1968) referred some Middle Devonian plants to the still invalid genus *Serrulacaulis*. When finally validated, it will add a Middle Devonian zosterophyll (Belgium) and a second Upper Devonian one (New York State). Present evidence points to Emsian as the time when zosterophylls reached their peak, and Late Devonian as the time when their numbers waned prior to extinction.

Summary of Zosterophyllphytina

The zosterophylls appear to be a natural group of six genera to which at least two others may be added soon. Only one species,
Zosterophyllum myretonianum, is known as a whole plant. Walton (1964) reconstructed it with a basal tuft of upwardly and downwardly growing branches. It is quite possible that the downwardly directed ones were part of a horizontal rhizomatous mass from which some erect branches grew upward at right angles to the rhizomes. The appearance of the fossils might result from tearing the plant free from its substrate and flattening the aerial branches during deposition so that aerial and horizontal branches appear to be in one plane.

Two genera, Rebuchia and Bathurstia, have not been proven to have vascular tissue. Bathurstia does have a cuticle. Their association with four genera of vascular land plants is based on the remarkable similarity of their sporangia to those of plants in which vascular tissue has been demonstrated. In this connection it must be remembered that many specimens of Devonian plants fail to yield vascular tissue even when it is sought carefully. Also true is the fact that plants with vascular tissue were relatively common by Siegenian time when Rebuchia and Bathurstia lived. I am willing to assume, with appropriate caution, that all six genera were vascular plants and to hope that ultimately tracheids will be found in newer collections of Rebuchia and Bathurstia.

Edwards (1969a) speculated that the shape of the xylem strand in some zosterophylls mirrored the external symmetry of the plants. She found an elliptic strand in the fertile part of Zosterophyllum llanoveranum where the sporangia appear to be in two rows and a terete xylem strand in the lower vegetative parts that are radially symmetrical. Unfortunately we still have too few specimens with preserved strands and too few with the xylem preserved at more than one level in the plant to comment usefully on this hypothesis. Edwards suggested that Sawdonia, whose elliptic strand was reported by Hueber and Banks (1967), might be an exception. But this genus also has too little xylem preserved to permit argument. Edwards's significant contribution is to indicate how much remains to be learned about zosterophylls and to caution against over-generalisation from limited evidence. Furthermore, if Edwards's suggestion (1969a: 202) that some 'spikes' of sporangia of Zosterophyllum llanoveranum were spirally arranged rather than two-rowed, the question would arise 'did this specimen therefore have a terete xylem strand?'

The most remarkable aspect of zosterophylls is the uniformity among their sporangia. The reniform to sometimes globose sporangia that dehisced along their distal, convex margin are distinctive. All genera except Crenaticaulis dehisced into two equal valves. Crenaticaulis dehisced into two unequal valves. In 1930 Lang and Cookson (p. 145) suggested that the thickened distal margin of sporangia of Z. australianum 'gives the appearance of two ridges with a groove between'. Remarkable confirmation of this suggestion was provided by Edwards (1969a) when she described petrified sporangia of Z. llanoveranum. The distal margin is indeed composed of two ridges of thick-walled cells with a groove between (Fig. 7.3). In Zosterophyllum sp. cf. Z. fertile Edwards (1969b) found a dehiscence mechanism. Thickened distal rims on other species indicate the possibility of similar mechanisms but proof is wanting. Lang (1927) also recognised that sporangia of Zosterophyllum when viewed adaxially or abaxially in their normal position were reniform. When viewed from the side their appearance is often very different because the two halves of the reniform body may have been folded together in the adaxial direction during or prior to fossilisation. The appearance then simulates 'a curved linear structure with an oval body placed adaxially to its distal portion' (Lang, 1927: 449). Quite independently subsequent workers have come to the same conclusion, thus avoiding a series of descriptions of different kinds of sporangia among zosterophylls.

Histologically the epidermal cells and stomates of zosterophylls, where known, are similar. The guard cells are not marked by anticlinal walls at their ends. Therefore one cannot refer to a pair of guard cells. Perhaps there was but one ring-like cell. Perhaps the anticlinal walls of two guard cells were unpreserved. The cortex where known is composed of thick-walled cells, a hypodermis. Maturation of the xylem strand was exarch
but it is still uncertain whether protoxylem was equally spaced around the margin or was confined to one portion of the periphery of the strand. Scalariform tracheids constituted the bulk of the xylem.

_Gosslingia_ and _Crenaticaulis_ both had axillary tubercles (scars located in the axil between the main axis and lateral branches). Edwards (1970a) demonstrated that a xylem strand extended into the tubercle of _Gosslingia_. Banks and Davis (1969) showed branches attached in the location of the tubercle of _Crenaticaulis_. On the basis of similarity of the two genera, the attached branches in one, and the vascular supply to the scar in the other, Banks and Davis speculated that these branches might resemble the roots or rhizophores of a _Selaginella_.

Banks (1968b) instituted two families in _Zosterophyllophytina_, _Gosslingiaceae_ and _Zosterophyllaceae_. Heuber (1972) thinks the evidence too limited to justify more than one family, the _Zosterophyllaceae_, suggesting in particular that unpublished data show that sporangia of _Crenaticaulis_ and _Sawdonia_ may be borne in spikes whereas they had been described as 'in groups' or 'random'. Edwards (1970a) is inclined to place only _Gosslingia_ and _Crenaticaulis_ in _Gosslingiaceae_; _Sawdonia_ and _'Serrulacaulis'_ she would place elsewhere, e.g. _Incertae Sedis_. Families of fossil plants are not subject to rules of priority under the International Code of Botanical Nomenclature. Hence they are used at the discretion of each author to express his opinion of the most useful way to sort out groups of organisms. As data increase in volume or in detail, changes in family alignments can be expected. Perusal of Edwards (1969a, 1969b, 1970a, 1972), Hueber (1971b, 1972) and Lele and Walton (1961) shows that the number of characters known among zosterophylls is increasing. In time it will be possible to evaluate more objectively the need for one or for more than one family of zosterophylls.

Both Edwards and Hueber have been concerned about one zosterophyll character in particular, the position of the sporangium whether radial ( _Zosterophyllum_ ) or bilateral ( _Platyzosterophyllum_ ). Edwards's discovery of _Z. llanoveranum_ with a possible third row of sporangia rather than the usual two and her observation (1972) that the usually radially arranged sporangia of _Z. myretonianum_ may sometimes occur in a single row suggest that even this character may not be definitive.

Finally, the ornamentation of the stem of the zosterophylls varied widely from multicellular spines in _Sawdonia_, through tooth-like outgrowths on _Bathurstia_ and _Crenaticaulis_, to naked axes in _Zosterophyllum_ and _Rebuchia_. _Gosslingia_ has protuberances of unknown morphology. Perhaps they are comparable to the epidermal papillae found on many epidermal cells of _Crenaticaulis_. It is possible that the darkened areas seen on the outer surface of some epidermal cells of _Sawdonia_ are also equivalent to papillae. By now it should be obvious that disparity in preservation and in techniques of study make precise comparisons of species of Devonian plants a difficult task.

**Relationship of Northern and Southern (Australian) Hemisphere Floras in Early Devonian**

Lang and Cookson (1930, 1935) and Cookson (1935) have described Early Devonian floras of Victoria in some detail. The one zosterophyll, _Z. australianum_, links the floras of the two hemispheres closely. It is supplemented by axes with axillary tubercles that suggest _Crenaticaulis_ or _Gosslingia_ of Gaspe and Wales and by _Flostimella_ which is frequent in the northern hemisphere. The affinities of _Hostinella_ are generally unknown although Banks (1968c) described specimens from Rørogen, Norway that had a centrarch xylem strand like that of _Psilophyton_. The non-vascular plants _Sporogonites_ and _Pachytheca_ similarly link the two floras.

Four other species, _Baragwanathia longifolia_ Lang and Cookson 1935, _Hedeia corymbosa_ Cookson 1935, and _Yarravia subspheca_ and _Y. oblonga_ both described by Cookson (1935) are usually considered to be restricted to Australia. However, there are four reports of this group in the northern hemisphere. Danzé-Corsin (1956) described _Y. minor_ from the Pas-de-Calais area, France. It seems to me a doubtful specimen of the genus. _Y. gorelovii_ Anan'ev 1960 from the
Altai-Sayan Mountain region, Siberia looks more like a *Yarravia* than that from France but it is not fully convincing. More convincing is Canright's (1970) identification of *Yarravia* sp. cf. *Y. gorelovii* from the Salt River Canyon in central Arizona. Some years earlier Teichert and Schopf (1958) had mentioned the occurrence of *Hedeia* at the Salt River locality, although they gave no illustrations. The age given for these Arizona specimens has ranged from Emsian-Eifelian to Givetian-Frasnian. This wide range is not helpful in the present discussion but it is significant that *Hedeia* and *Yarravia* may not be unique to Australia. If better material forces the erection of new genera for the specimens of *Hedeia* and *Yarravia*-like plants in the northern hemisphere, they will be most closely related to their Australian counterparts and will be of comparable or younger age. *Baragwanathia* is unknown in the northern hemisphere but the somewhat similar lycopod *Drepanophycus spinaeformis* ranges from Siegenian through Frasnian (Banks and Grierson, 1968). The xylem strand of *D. gaspianus* (Grierson and Hueber, 1968) is comparable to that of *Baragwanathia*, as is also the strand of *D. spinaeformis* (Fairon-Demaret, 1971).

A brief glance at the Siegenian-Emsian flora of Victoria, Australia indicates that it is closely related to floras of comparable age in the northern hemisphere.

**Palynological Evidence**

Long vertical series of strata that are regarded as securely dated and that have been carefully analysed for Early Devonian spores are uncommon. One of the best is that in Wales and the Welsh Borderlands by Richardson and Lister (1969). They studied Wenlockian through Siegenian strata and found a steady increase in number and diversity of spores. McGregor, Sanford and Norris (1970) studied the fossils of the Moose River Basin on the James Bay Lowland, northern Ontario. McGregor was able to correlate the spores of the Kenogami River Formation with Gedinnian spores described by Richardson and Lister and some described by Allen (1967) from the Gedinnian of Spitsbergen. Immediately overlying beds contained spores comparable to those from the Siegenian in Spitsbergen and on the Gaspé Peninsula, eastern Canada. They agree also with data for the Siegenian in the reviews by Chaloner (1967) and Richardson (1969). Spores from higher up in the Stooping River Formation are clearly Emsian assemblages. They resemble collections from the Gaspé Peninsula, the Sahara, the Eifel region of Germany, Belgium, and Ellesmere Island.

The succession of spores in the James Bay area appears comparable to successions elsewhere. The ages are confirmed by invertebrates. Thus these spores can be used to date strictly non-marine strata when they are found in the latter. The Sextant Formation in the James Bay region is an example. It contains *Sawdonia ornata* and can be dated as Emsian on the basis of its contained spores. It can be equated also with the continental Battery Point Formation on the Gaspé Peninsula. That is the formation in which *Sawdonia* and other elements of the classic Gaspé Emsian flora of Dawson occurs.

This extremely limited sampling of palynological data points to the same conclusion as do macrofossils. Early Devonian floras were widespread and relatively uniform. Were time and space available, Middle and Upper Devonian palynological evidence would expand the analysis markedly. One paper cited above (McGregor, Sanford and Norris, 1970) includes Middle Devonian assemblages of spores and relates them to collections in other areas. McGregor (1967) and Kerr, McGregor and McLaren (1965) both deal with spores from the Canadian Arctic of Mid or Late Devonian age. Both papers discuss relationships between these Canadian spores and spores from areas as widely separated from Canada as Australia and the U.S.S.R. McGregor (1967) concludes on the basis of his studies of spores that 'broad areas seem to have supported similar vegetation in the Devonian'. He notes also the resemblance of an assemblage of spores from Cornwallis Island (Canadian Arctic) to one described by Balme and Hassell (1962) from the Famennian of Western Australia. Balme and Hassell themselves related their spores to assemblages in North Africa and the Russian Platform. Additional notes on the spores of
the Canadian Arctic Islands are given in McGregor and Uyeno (1972) where spores from Lower, Middle and Upper Devonian strata are compared with assemblages elsewhere. In short, palynological data as presently understood parallel data from macrofossils in suggesting widespread uniformity of Devonian floras.

EVIDENCE FROM SEDIMENTATION AND PALAEOGEOGRAPHY

Given a marked similarity between the Devonian plants collected at distant localities, it may be instructive to examine the conditions under which they were deposited. This, in turn, might suggest something about the location of continents at the time when the plants lived. For convenience we can look just at the present fringes of the North Atlantic and across Arctic Canada. This account is based principally on two papers, Allen and Friend (1968) and Allen, Dineley and Friend (1968). Devonian sediments in the area concerned were deposited on the seaward side of an orogenic belt, often referred to as the Caledonian orogeny. This belt was forming from Late Precambrian to, or through, Devonian time. Much of the sediment in the Appalachians and northern Canada is related more to the Late Devonian Acadian orogeny than to earlier events. Non-marine sediments were deposited in two environments, alluvial plains that bordered the orogenic belts interfingering with marine environments, and intermontane basins. The former were the more extensive, including especially the Catskill sediments in eastern U.S.A., the Canadian Arctic (McGregor and Uyeno, 1972) and eastern Canada, southern Britain, and western Europe. The latter are found in Scotland, western Norway, northern Ireland, eastern Greenland, and Spitsbergen. I would make the tentative suggestion that the genera of plants in the two areas, plains and intermontane valleys, are not markedly different, at least in so far as we now understand them.

The rocks found on the former alluvial plains are characterised by cyclothems 5-15 m thick. Their sediments are finer upward. Each cyclothem commences on an eroded surface and sandstones or conglomerates, usually cross-bedded, are deposited first. They are followed by interbedded sandstones and siltstones. These are scoured and then a new series is initiated. Variation in the cyclothsms led to the interpretation that close to the Devonian highlands there was an area of steeper slopes and braided streams, followed by lesser slopes and meandering streams, finally giving way to tidal flat and related environments fringing an epicontinental sea. This was illustrated diagrammatically by Allen and Friend (1968: 10). Fossil plants are preserved similarly in similar cyclothems in North America and in Wales. Palaeocurrent data indicate that the direction of flow of sediments during Devonian was in essentially opposite directions in North America and Wales. Finally, the uplands that supplied the sediments in these two places were similar in abundance of metamorphic rock (see Petrology in Allen and Friend, 1968), in 'altitude, relief, predominance of mechanical over chemical weathering, and relative nearness to depositional sites' (Allen and Friend, 1968).

The facts given above are consonant with the hypothesis of a single pre-Atlantic orogenic belt which was displaced by the movement apart of continents. Lateral remnants of this belt are found in the present Appalachians of eastern North America extended across eastern Canada, Britain, Greenland and Norway to Spitsbergen. If the present continents be fitted together at the 500 m submarine contour, there is produced an area that when uplifted during Devonian would have produced an abundance of sediment along both sides, continental strata closer to and marine strata further from the area of uplift. Intermontane basins as in Spitsbergen, Norway and Scotland might have existed within the uplifted area. Rocks now found along the eastern Baltic region, the Rhine-Ardenne, Ardenne, southern Britain, Appalachian United States and Canada, and the Canadian Arctic were deposited as sediments on the alluvial plains and in the epicontinental seas that fringed the uplifted area (Fig. 7.6).

If one wonders at the similarity of Devonian plants that grew under such disparate climatic conditions as now exist in the Cana-
Fig. 7.6. Basins of deposition bordering the North Atlantic in Devonian time. Continental positions based on Bullard et al. (1965) and equator from Opdyke (1962). (From Allen and Friend, 1968.)

One can certainly not argue that Devonian plants, regardless of the perfection of their preservation, constitute any form of proof of continental drift. However, their identity over wide areas and the similarity of their preservation under comparable conditions of preservation do fit the hypothesis that land masses were once much closer together and were subjected to similar climatic conditions. It may never be possible to rule out the possibility of air-borne transport of spores from continent to continent across oceans, an alternative explanation for the widespread occurrence of the fossil plants if drift is rejected. But the evidence presently available, including climatic conditions in Devonian time, points in the direction of a former single land mass in the northern hemisphere.

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The Rhacopteris Flora in New South Wales

NOREEN MORRIS

ABSTRACT

Although the Rhacopteris flora in New South Wales is found in rocks of Late Carboniferous age, the commonest plants in it have primitive structure and venation and appear identical with taxa which in Europe are restricted to Early Carboniferous rocks. A species of Sigillaria (Subsigillaria), a Late Carboniferous to Early Permian genus, is, however, a minor component of the flora. Very few species make up the flora and in general there is very little variation in its composition. In the top 300 m of its range two new taxa appear which seem to be identical with American and European species of Middle to Late Carboniferous age. In parts of the region the commonest plant fossils in this top 300 m are forms intermediate between these two taxa. A severe climate is thought to be responsible for the poverty of the flora and the absence of most species of the northern hemisphere coal flora.

INTRODUCTION

The age of the Rhacopteris flora in Australia and South America has been a matter for debate for many years. Palaeobotanists familiar with the Late Carboniferous coal floras of the northern hemisphere almost always consider the Australian flora to be of Early Carboniferous age. Stratigraphers who have collected the flora in the field consider it to be of Late Carboniferous age.

The problem has arisen because two of the commonest plants in the flora, Rhacopteris ovata (M'Coy) Walkom and the one described by Dun (1899) as a species of Cardiopterus, bear a very close resemblance to the European Early Carboniferous species Rhacopteris circularis Walton and Cardiopterus frondosa Goeppert.

There are three possible solutions to the problem.

1. The rocks are of Early Carboniferous age. This is definitely not the case in New South Wales. The rocks containing R. ovata are younger that late Viséan and older than Early Permian.

2. The rocks are Late Carboniferous, and the similarity of the plants to European species is the result of heterochronous parallel evolution. Rigby (1973a, b) considers that the floras of Gondwanaland and the northern hemisphere developed along separate lines after Early Devonian time. This hypothesis cannot explain the existence of a diverse Viséan flora with strong European affinities underlying the Rhacopteris flora in New South Wales.

3. Because of tectonic and climatic changes, a depauperate Early Carboniferous flora became isolated on the Gondwana continent and survived well into the Late Carboniferous. Very few Late Carboniferous species were able to cross the barrier which developed between the northern and southern floras. This hypothesis is favoured herein.

STRATIGRAPHIC SEQUENCE OF FLORAS

The Carboniferous succession in New South Wales is now relatively well understood and there is no doubt that the Rhacopteris flora is found in rocks of Late Carbon-
iferous age. A flora containing typical Viséan lycopods such as *Lepidodendron spetsbergense* Nathorst found in rocks 1000 m below the lowest *Rhacopteris* bearing beds is associated with an abundant fauna of marine invertebrates of undoubted Viséan age. The topmost *Rhacopteris* bearing beds are only 300 m below coal seams associated with an Early Permian flora containing *Glossopteris*, *Gangamopteris* and *Noeggerathiopsis*. No unconformity has been recognised.

The Viséan flora consists for the most part of driftwood in marine sediments; *Lepidodendron* sp. and small shrubby lycopods are the plants most commonly found, but fine marine sediments of early to middle Viséan age contain 'fern' plants, principally *Rhodea* spp. Plant fragments belonging to the form genera *Cardiopteridium*, *Sphenopteridium*, *Sphenopteris* and *Adiantites* have also been found but they are very rare. The pinnules of these plants are all very small, usually less than a centimetre across. On one horizon in the Rouchel area numerous lycopod logs have been preserved in terrestrial sediments associated with volcanics. The majority of these logs are less that 10 cm in diameter.

At the end of Viséan time volcanic activity reached a peak in the southern part of the basin and much of the area was raised above sea level. The sediments associated with the volcanics contain few lycopods, but there are abundant silicified logs of *Pitus süssmilchi* Walkom on some horizons. The largest have a diameter of 60 cm or more.

A widespread marine horizon, thought to be Westphalian in age because of the presence in it of vast numbers of the productid *Levipustula levis* Maxwell, together with the characteristic *L. levis* assemblage, occurs a few hundred metres above the volcanics. *Rhacopteris* is found on several horizons within the succeeding 2000 m of sediment. It has been reported also from two localities which may be below the *Levipustula* horizon, possibly extending its range downwards to the Namurian.

In the southern part of the area, i.e. closer to the Carboniferous shoreline, acid volcanics and a considerable thickness of varved shales with interbedded conglomerates separate the *Rhacopteris* flora from rocks of undoubted Permian age. In the north, up to 100 m of conglomerates and diamictite lie between the upper *Rhacopteris* beds and the Alum Mountain Volcanics which contain coal bands associated with a typical Early Permian

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Table 8.1. Correlation Chart, Carboniferous Rocks North of Newcastle

<table>
<thead>
<tr>
<th>Stroud-Gloucester syncline</th>
<th>Paterson District</th>
<th>Commonest plant fossils</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alum Mountain Volcanics</td>
<td>Dalwood Fm</td>
<td><em>Glossopteris</em>, <em>Gangamopteris</em></td>
</tr>
<tr>
<td>Johnsons Creek Conglom.</td>
<td>Seaham Fm</td>
<td></td>
</tr>
<tr>
<td>McInnes Fm</td>
<td>Paterson Toscanite</td>
<td><em>Rhacopteris ovata</em></td>
</tr>
<tr>
<td>Cut Hill/Booral Fm</td>
<td>Mt Johnstone Fm</td>
<td>Pitus <em>süssmilchi</em></td>
</tr>
<tr>
<td>Faulkland/Karuh Fm</td>
<td>Gilmore Volcanics</td>
<td>Lepidodendron spp.</td>
</tr>
<tr>
<td>Berrico Crk Fm</td>
<td>Wallaringa/Flagstaff Fm</td>
<td></td>
</tr>
<tr>
<td>Buggs Creek Volcanics</td>
<td>Bonnington Fm</td>
<td>Ararat Fm</td>
</tr>
<tr>
<td>Carsonville/Conger Fm</td>
<td>Bingleburra Fm</td>
<td></td>
</tr>
</tbody>
</table>

Formation names which do not appear on the Correlation Chart for the Carboniferous System in Australia of Jones et al. (1973) are reserved nomenclature for the Gloucester region. The numbers refer to the Faunal Zones of Campbell and McKellar, 1969.
Glossopteris-Gangamopteris flora. No unconformity has been observed.

**PALAEOBOTANY**

*Rhacopteris ovata* (M'Coy) Walkom is a very common and widespread fossil and is found well-preserved in many localities. The remains of associated plant species are, however, exceedingly rare in all but the top 300 m of *Rhacopteris*-bearing beds.

Plate 8.I. a. x 1 *Cardiopteris frondosa* Westbrook. b. *Sigillaria* (*Clathraria*) sp. nov. x 1 showing seasonal banding. b'. The same x 2 showing double vascular trace and parichnos. Stroud Road. c. *Triphylopteris austrina*. Sugarloaf Creek, Stroud. d. *Daetylophyllum digitatum* with aphlebia and fructification (upper left) d'. Another specimen showing structure of pinnules. Both from Dingo Creek, Stroud. e,f,g,h,i. simple pinnae of intermediate forms. Sugarloaf Creek. j,k,l. Compound pinnae of intermediate forms. j,l. Sugarloaf Creek; k. Stroud Road.
Only six species have been found in the lower beds. These are

*Rhacopteris ovata* (M'Coy)
*Cardiopteris frondosa* Goeppert

calamitean stems (? *Paracalamites* sp.)

*Lepidodendron steinmanni* Jongmans

*Sigillaria* (Subsigillaria) group *Clathraria* sp. nov.

A few additional species occur in the upper beds, but the total number is still only about ten for the whole of Late Carboniferous time. In comparison, the Canadian flora of the same age contains ninety-one species (Bell, 1943).

Both *Rhacopteris ovata* and *Cardiopteris frondosa* have primitive radiating bifurcating venation, and pinnules constricted at the base. These are characteristics of Early Carboniferous plants. By Late Carboniferous time many plants in the northern hemisphere had developed a true midrib and a few had reticulate venation. Many pinnules had broad or confluent bases.

Walton (1927) considered *Rhacopteris ovata* (M'Coy) to be the same as *Rhacopteris circularis* Walton from the Lower Carboniferous of Great Britain. The two species are certainly very similar; both have more or less symmetrical pinnules with two or three orders of radiating veins and an entire or crenate margin. The only significant difference appears to be in the size of the pinnules. *R. ovata* is considerably larger than *R. circularis*; the average length of the pinnules is 15 mm and the maximum 25 mm, whereas the maximum pinnule length of *R. circularis* is only 12 mm.

The specimens of *Cardiopteris frondosa* do not appear to differ at all from the Viséan fossil from Europe. Unlike *Gondwanidium*, which it strongly resembles in size, venation, and the once-pinnate structure of the frond, its pinnules are symmetrical and consistently heart-shaped at the base. The pinnules average 20 mm-30 mm in length, but some are very much larger. *C. frondosa* is relatively rare in the lower beds but is much more common in the upper, particularly near Paterson.

The calamitean stems have opposing ribs at the nodes. They resemble *Archaeocala-

*mites scrobiculatus* Schlotheim but the ribs are closer together than in this species. A single verticil of leaves found in the upper beds closely resembles *Stellotheca robusta* Surange and Prakash, and indicates that the stems may belong to the form genus *Paracalamites* Zalessky; this form genus is not uncommon in the Lower Permian in Australia (Rigby, 1966a, 1966b) and could conceivably have an ancestor in the Upper Carboniferous.

All of the lycopods are very rare. Two of them, *Cyclostigma austral* and *Lepidodendropsis steinmanni* are rather primitive and would not be out of place in a Lower Carboniferous flora. *Sigillaria* (Subsigillaria), however, is not known from the Lower Carboniferous of the northern hemisphere and is rare until the Stephanian and Permian. The significance of this is probably ecological rather than biostratigraphic. *Sigillaria* (Eu-sigillaria) thrived in the Westphalian coal swamps and the xeromorphic *Sigillaria* (Sub-sigillaria) in the much drier Stephanian and Permian.

*Sigillaria* (Subsigillaria) group *Clathraria* sp. nov. occurs also in the South American *Rhacopteris* flora (Paganzo Series—Fren-guelli, 1946), but because of its rhomboidal leaf cushions it has been confused with *Leptophloeum austral*.* It can be readily distinguished from this plant, however, by the double vascular trace which is found in the centre of the leaf scar. The single oval vascular trace of *Leptophloeum* is usually found at the top of the leaf scar. The leaves of *Sigillaria* (Clathraria) sp. nov. are frequently preserved on the ends of the branches. They are very long (8.5 cm+) and rigid. The leaves of *Leptophloeum* are very rarely preserved; they are small and scale-like and have usually fallen from the leaf scars before they are 3 mm across. The stems of *Sigillaria* (Clathraria) sp. nov. are also distinguished by a pronounced horizontal banding caused by alternating zones of large distinct leaf cushions and much shorter ill-defined ones. This banding may be due to cessation of growth during severe winters. The plant was apparently slow-growing, as the annual bands are only 5 cm long. It was rather like a grass tree (*Xanthorrhoea*) in
habit, and appears to have been a rarely branching shrub or small tree up to about 2 m in height.

The flora in the top 200 m-300 m of the Rhacopteris-bearing beds is rather different from that in the lower part. Rhacopteris itself is much rarer and is absent in some localities. The other five species are relatively much more abundant. Stems of Sigillaria (Subsigillaria) group Clathraria sp. nov. reached a greater diameter, and rare wood fragments and winged seeds indicate that Cordaites may have been present, although in small numbers. The commonest species in these upper beds are not known in older rocks. As is the case with Rhacopteris ovata, outcrops consist frequently of a single species only, which may be distributed over an area of many square metres on a single horizon or on several closely spaced ones. One of these species, common in the Paterson area, was named Rhacopteris digitata by Etheridge jun. The pinnules of this plant, although very primitive in structure, have a shape not usually associated with Carboniferous plants.

A very deep primary division cuts the pinnule into two deltoid lobes which are further subdivided by regular dichotomies into eight, sixteen or thirty-two long linear segments. The venation is very coarse and dichotomises simply so that each segment of the pinnule contains one or two veins. This leaf shape is usually associated with Ginkgo (Baiera) but in this case the leaves are attached to a bipinnate frond and closely associated with a sphenopterid type of fructification resembling Dictyothyllum Goeppert. They are most probably seed ferns. Read (1934-35) gave the name Dactylophyllum to similar pinnules found in Namurian rocks in Colorado.

A very large and variable aphlebia (or protective organ), Rhacophyllum diversiforme of Etheridge jun., is found associated with Dactylophyllum digitatum. It appears to have been a modified frond and varies in shape from an almost entire sub-elliptical structure to a once or twice pinnate frond with very irregular ultimate divisions, which approaches Dactylophyllum in size and shape. The venation is coarse, consisting of several parallel strands about 1 mm apart in the central part of the structure, and simply dichotomising veins in the lateral parts. A single vein enters a lobe or pinna of the divided aphlebiae at an angle of 30° and branches several times within it. The almost entire aphlebiae are rather like the Permian form genus Rubidgea Tait in structure, and indicate a possible evolutionary pathway for the development of the Rubidgea-Gamopteris-Glossopteris type of leaf frond from a more primitive once or twice pinnate frond. In a growing season which was becom-

<table>
<thead>
<tr>
<th>Properties of pinnule</th>
<th>T. australina Score</th>
<th>Intermediate forms Score</th>
<th>D. digitatum Score</th>
</tr>
</thead>
<tbody>
<tr>
<td>venation</td>
<td>fine</td>
<td>2</td>
<td>coarse</td>
</tr>
<tr>
<td>symmetry</td>
<td>asymmetrical</td>
<td>2</td>
<td>symmetrical</td>
</tr>
<tr>
<td>primary division into 2 lobes</td>
<td>absent</td>
<td>2</td>
<td>deep</td>
</tr>
<tr>
<td>division into further lobes</td>
<td>absent</td>
<td>4</td>
<td>half pinnule length</td>
</tr>
<tr>
<td>width of pinnule determined by summation of segments</td>
<td>90° or more 2</td>
<td>75° 1</td>
<td>45° 0</td>
</tr>
<tr>
<td>elongation of segments relative to width</td>
<td>no division into segments 4</td>
<td>1:1-&gt;2:1 3,2,1</td>
<td>&gt;2:1 0</td>
</tr>
<tr>
<td>third lobe present</td>
<td>yes</td>
<td>2</td>
<td>no</td>
</tr>
<tr>
<td>Total</td>
<td>18</td>
<td>5-9</td>
<td>0</td>
</tr>
</tbody>
</table>
ing increasingly short, the entire, possibly deciduous aphlebia (cataphyll) may eventually have been the only type of frond produced.

*Triphyllopteris austrina* (Etheridge jun.) was first described from the Joe-Joe Formation in Queensland. The type specimen is an almost complete bipinnate frond which shows clearly the great variation in the shape of the pinnules on different parts of the frond. The majority of the pinnules are entire, about 1.5 cm long, and asymmetrical. On the lower pinnae some of the pinnules are very large and lobed. The pinnules on the ends of the pinnae are very often divided into three lobes; this is a diagnostic characteristic of the form genus *Triphyllopteris*. *T. australa* is very similar to and may be identical with *T. rhomboidea* Ettingshausen from the Westphalian of Canada and Eastern Europe.

Until recently only small fragments of *T. australa* had been found in New South Wales, but a very extensive outcrop has recently been discovered in the dacite quarry at Raymond Terrace. Associated with it is a large fragment of an aphlebia similar to the ones found with *Dactylophyllum digitatum* but with much coarser venation, and a well-preserved fructification resembling *Dictyothalamus* Goeppert. No other plants occur in this quarry with the *T. australa* leaves and so these three almost certainly belong to the same plant species.

Palaeozoic plants are placed in form genera based on the shape and venation of the pinnule and the structure of the frond, because the fructifications, on which a more meaningful classification could be based, are usually not found in contact with the leaves. *Dactylophyllum digitatum* and *Triphyllopteris australa* are placed in different form genera because of different leaf shape, but the fact that both have large aphlebiae of similar structure and similar fructifications indicate that they were probably related at the generic level.

In the Stroud-Gloucester region both *D. digitatum* and *T. australa* are found in small numbers, but the dominant vegetation consists of plants with asymmetrical divided pinnules. There is a great variety in the shape of the pinnules, which show a gradation from narrow dissected forms close to *Dactylophyllum* to gently lobed or crenate forms close to *Triphyllopteris*.

Feistmantel (1890) placed some of these forms in the genus *Archaeopteris* Dawson because of their similarity to *Archaeopteris dissecta*, now *Sphenopteridium dissectum* (Goeppert). Other forms with more divided pinnules have been described as *Rhacopteris septentrionalis* Feistmantel and *Sphenopteridium cuneatum* Walkom.

The area in which Feistmantel's specimens were collected was lost until recently (Kisi, 1970). Although abundant material similar to the type material has been found it has proved impossible to classify it satisfactorily using the old nomenclature. If sufficient specimens are collected, they can be seen to form a complex of intergrading forms with *T. australa* as one end member and *D. digitatum* as the other. The majority of the intermediate forms fall within the form genus *Sphenopteridium* but some with symmetrical divided pinnules with very short lobes fall within *Dactylophyllum* and others with entire but elongated pinnules should be included in *Triphyllopteris*. There are some extremely large-leaved forms present, not only of the intermediates but also of *Dactylophyllum digitatum*. It is possible that these are polyploids ('gigas' forms).

The truly intermediate nature of these pinnules can be illustrated by constructing a hybrid index histogram (Fig. 8.7) (Riley, 1938, in Briggs and Walters, 1969). The characters in which *D. digitatum* and *T. australa* differ from one another are listed and a numerical value given to each character. This value is 0 for the characters of *D. digitatum* and 2 or 4 for the correspond-
Fig. 8.3. a-k. simple pinnae from base of frond.
I-v. compound pinnae from middle part of frond.
a. *Dactylophyllum digitatum*. k and u. *Triphylopteris australis*; all others intermediate forms. b and l,
g and r, h and p, k and u could be expected to occur on same frond. Note similarity in shape but difference in size of the pairs c and f, h and j, r and f. All x 1. Localities: a, q Paterson; b, l, m Stroud Rd; k Raymond Terrace. All others Sugarloaf Creek, Stroud.

Fig. 8.4. a-d. Aphlebiae of *Dactylophyllum digitatum* (*Rhacophyllum diversiforme* of Etheridge jun.) showing variation from almost entire to much divided structure. e. frond of *Dactylophyllum digitatum*. All x 1. All specimens on same slab of rock. Locality: Dingo Creek, Stroud.

ing character of *T. australis*. Intermediate characters score intermediate values. For example if a pinnule is split deeply into two ± equal lobes it resembles *D. digitatum* and scores 0, a shallow division is an intermediate character and scores 1, if no division is present as in *T. australis* the score is 2. The fifty specimens used to produce the histogram
were all collected in Sugarloaf Creek.

The two end members are here considered to be species because plants very similar to or identical with them are known from other countries. It is possible, however, that they were ecotypes of a variable species, the broad-pinnuled form being adapted to wet and the narrow-pinnuled to drier environments. Obligative outcrossing and/or the ability to reproduce vegetatively enabled the intermediate forms to persist throughout the time represented by 260 m of sediments, the equivalent of hundreds of plant generations.

The distribution of *T. australis-D. digitatum-Sphenopteridium* complex in the Late Carboniferous basin is complex. The following floral lists are from localities considered to be on or near the same horizon.

(* signifies abundant, † very rare)

**Paterson (south-west of basin)**  
*Cardiopteris frondosa*  
*Dactylophyllum digitatum* (small form)  
*Sphenopteridium* complex  
*Rhacopteris ovata*  
*Calamitean stems*

**Sugarloaf Creek, Booral (north-east of basin)**  
*Sphenopteridium* complex  
*Dactylophyllum digitatum* (large form)  
*Triphyllopteris australis* (small form)  
*Sigillaria* (Subsigillaria) Group *Clatharia* sp. nov.  
† *Cyclostigma australe*  
† *Lepidodendropsis steinmanni*  
*Calamitean stems*

Sugarloaf Creek is on the eastern side of the Stroud-Gloucester syncline. Many other localities in which *Sphenopteridium* is the dominant or only fossil present are known along strike from this locality.

**Near Bulahdelah (far north-east)**  
*Rhacopteris ovata*  
† *Dactylophyllum digitatum*  
† *Lepidodendropsis steinmanni*

**Raymond Terrace (south-east of basin)**  
*Triphyllopteris australis* (large form)  
*Calamitean stems*

**COMPARISONS WITH NORTHERN HEMISPHERE**

If the *Rhacopteris* flora is compared with the contemporaneous floras from Europe and North America some marked differences emerge.

1. In New South Wales the transition from a Lower to an Upper Carboniferous flora is abrupt. There is no species known to occur in both floras.

2. Although conditions for preservation are often good, sometimes ideal, the number of species found fossilised is very small (six to ten) and the number of species in one outcrop is usually only one or two.

3. Two of the most commonly occurring plants appear identical with taxa which in the northern hemisphere are restricted to rocks of Lower Carboniferous age. One much rarer species, however, belongs to a group known only from Upper Carboniferous and Permian rocks.

4. During the greater part of the Late Carboniferous *Rhacopteris ovata* was the dominant species. Fructifications are virtually absent from rocks of this age. Winged seeds of the type associated with *Cordaites* are unknown, and wood fragments are rare and usually small. This is in marked contrast with both the Early Carboniferous and Early Permian floras.

These differences appear to be due to the formation of a partial barrier between the northern hemisphere and eastern Australia during Middle Carboniferous time. It is possible that this barrier was climatic in nature. This partial barrier appears to have persisted until very late in the Carboniferous, as two of the new taxa appearing at this time in the upper *Rhacopteris* beds appear to be identical with European and North American taxa of Namurian and Westphalian age. The appearance of these additional species may have been due to a slight amelioration of conditions, the flora became more diverse, several fructifications including winged seeds are known, woody stems were much more abundant and the stems of calamites and lycopods attained a greater diameter.

The climate then apparently deteriorated as is shown by the presence of varved shales, conglomerates and diamictites containing very rare plant macrofossils and an im-
poverished pollen flora. The overlying Alum Mountain Volcanics contain a typical Glossopteris flora. In this region there is no transitional stage containing elements of both Glossopteris and Rhacopteris floras.

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ABSTRACT
Discoveries of petrified plant remains in the Upper Permian of the Bowen Basin, Queensland, have provided silicified material yielding anatomical details of the Gondwana Glossopteris flora; comparable deposits also occur in Antarctica. Most information to date on the morphology of Vertebraria indica Royle has been obtained from carbonised compressions or impressions of this plant organ; uncompressed petrified specimens now show the complete anatomical arrangement as well as various growth stages. Some specimens exhibit considerable development of secondary wood around the basic septate axis; this wood is almost indistinguishable from that of associated silicified trunks which have been referred to Araucarioxylon arberi (Seward) Maheshwari. It is inferred that V. indica was part of the underground system of these trees, and possibly represents an adaptation to a semi-aquatic environment.

INTRODUCTION
The investigation of petrified fossil plants yields more detailed information on their anatomical structure than can be obtained from carbonised compression material or impressions. Unfortunately, from most Permian plant-bearing deposits of Gondwanaland only compressions or impressions are available, so the occurrence of varied and well-preserved silicified plant remains in Upper Permian strata of the Bowen Basin of Queensland is quite exceptional. These exquisitely preserved petrifactions have been noted previously (Gould, 1967), and details of some osmundaceous stems reported (Gould, 1970). Very similar petrifactions have also been discovered in Antarctica (Schopf, 1965, 1970a, 1970b). One of the more outstanding features of the Queensland material is the prominence of specimens of Vertebraria indica Royle, 1833, and these are used in this paper to show the information that can be obtained from the study of petrified plants.

OCCURRENCE
The petrified material occurs at many localities in the Upper Permian coal measures of the Bowen Basin (Gould, 1970: fig. 1). Specimens discussed here come from the southwestern boundary of Homevale Station, Winchester Downs, and just west of Utah Development Company's Blackwater Mine; these localities are shown on Fig. 9.1. All localities are within the Blackwater Group as shown on the geological map of the basin compiled by Malone, Olgers, Mollan and Jensen (1967). The Blackwater Group is generally considered to be of Late Permian age (Hill, Playford, and Woods, 1972).

PRESERVATION
The petrifactions have two modes of occurrence which appear to intergrade. Individual or groups of silicified specimens are found in sandstone; these commonly also show much iron oxide which may obliterate the plant structure to some extent. Plants
found in this material include *Palaeosmunda* Gould 1970, *Glossopteris* spp., *Vertebraria*, and trunks of gymnospermous wood.

The best material occurs as silicified layers of plant material, very similar to that described from Antarctica by Schopf (1970b, 1971) and interpreted as permineralised peat; this type of deposit contains many different kinds of plant organs, including countless leaves of *Glossopteris* spp., small

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**Fig. 9.1. Locality map**

Pl. 9.I. Figs. 1, 2. Permineralised peat. 1, petrified plant remains including layers of innumerable *Glossopteris* leaves in section and a cordaitalean-type stem. x 1. 2, group of seeds in the petrified material. UNEF13494. x 1.8. Figs. 3-12. *Vertebraria indica* Royle. 3, cross-section of *Vertebraria* axis. Note radiating xylem arms, and periderm. UNEF13495. x 1.3. 4, growth ring in secondary xylem arm. UNEF13496. x 50.5, cross-section of central portion of small axis showing primary xylem and four radiating arms of secondary xylem. UNEF13497. x 50. 6, detail of protoxylem showing annular thickening. UNEF13498. x 126. 7, tangential longitudinal section of axis showing transverse xylem platforms and periderm. UNEF13498. x 1.8. 8, detail of xylem platform. UNEF13498. x 8. 9, xylem platform showing root trace. UNEF13498. x 8. 10, detail of periderm showing well-developed cork. Large cells at periphery of space between xylem arms on left. UNEF13498. x 50. 11, cross-section showing root atypically connected to end of xylem arm. UNEF13499. x 1.8. 12, transverse section showing root extending from xylem platform. UNEF13500. x 12.6.
Fig. 9.2. Stylised reconstruction showing proposed relationship of *Vertebraria* and gymnospermous trunk.

cordaitalean-type stems (Pl. 9.1, fig. 1), *Vertebraria*, trunks of gymnospermous wood, and numerous seeds (Pl. 9.1, fig. 2), sporangia, and rootlets.

The preservation of these Permian plants as petrifactions as opposed to the generally occurring compressions and impressions, means that their investigation will provide valuable information on the Gondwanaland *Glossopteris* flora in much greater detail than was hitherto thought possible. Indeed, these silicified petrifactions have been compared in importance to the well-known Carboniferous coal-balls of the northern hemisphere (Schopf, 1970b).

**TECHNIQUES**

Details of petrified plants are studied in polished sections, ground thin-sections, or acetate peels. The silicified material from the Bowen Basin was found suitable for study by all three techniques, but the ground thin-sections and acetate peels revealed the most information. Ground thin-sections were prepared by normal petrographic slide-making techniques. Acetate peels were made by first etching a smoothed surface of the silicified specimen with 40 per cent hydrofluoric acid for 1-2 minutes, and then embedding the organic material which was left standing in relief in a sheet of cellulose acetate using acetone as a solvent; when dry, this sheet was then 'peeled' off the specimen. Specimens which also contained much iron oxide were treated for 30-45 seconds in 20 per cent hydrochloric acid following the treatment with hydrofluoric acid. The thin-sections and peels obtained yielded superb anatomical details when examined microscopically.

**STRUCTURE OF *VERTEBRARIA***

The Permian genus *Vertebraria* Royle 1833, was characterised by elongate, flattened, segmented cylindrical casts, which are simple or branched, often bearing root-like organs at the transverse ridges; in cross-section the axes show wedge-like sectors radiating from the centre (Arber, 1905). Specimens of *Vertebraria* are found preserved in positions both parallel and perpendicular to bedding, and the actual nature of these axes was quite controversial for some time (Walkom, 1922; Pant, 1968). Walton and Wilson (1932), working from carbonised compressions, interpreted the basic structure of the xylem, including description of the tracheidal pitting; they also showed how compression of the original cylindrical axis resulted in the forms generally found. Following the study of further carbonised specimens, Sen (1958) suggested cordaitalean affinities for the genus. A reconstruction showing branches, roots, and some outer layers was presented by Plumstead (1962). Schopf
Pl. 9.II. Figs. 1-6. *Vertebraria indica* Royle. 1, large specimen showing considerable secondary growth around the central septate axis. UNEF13501. x 0.6. 2, cross-section of axis showing secondary growth. Central region of primary xylem not preserved. UNEF13496. x 1.8. 3, radial longitudinal section of secondary xylem in radial arm. UNEF13498. x 126. 4-6, radial longitudinal sections of secondary xylem in zone outside basic septate arrangement. UNEF13502. 4,5, x 126. 6, detail of the slit-like cross-field pits. x 320. Figs. 7, 8. Trunk of gymnospermous wood. UNEF13503. x 126. 7, radial longitudinal section through growth ring. Late wood *Araucarioxylon arberi* (Seward) Maheshwari; early wood *A. bengalense* (Holden) Maheshwari. 8, radial longitudinal section showing groups of pits.
(1965) reported a petrified specimen with considerable secondary growth from Antarctica, but unfortunately, like some of the Bowen Basin material (Pl. 9.II, fig. 2), this specimen lacked the central area of primary xylem. Further details of the anatomy of carbonised compressions were reported by Pant and Singh (1968) and some more details of the Antarctic petrifactions of *Vertebraria* have been shown by Schopf (1970a, 1971).

In the silicified deposits the layers of leaves, twigs, and other plant material are all arranged distinctly horizontally (Pl. 9.I, fig. 1), but the specimens of *Vertebraria* and rootlets permeate through these in all directions indicating that the *Vertebraria* were most probably growing in the accumulated plant material.

The petrified specimens show that the structure determined by Walton and Wilson (1932) was correct. In cross-section (Pl. 9.I, fig. 3), specimens consist of a central region of exarch primary xylem (Pl. 9.I, fig. 5), with four to seven radiating arms of secondary xylem showing distinct growth rings (Pl. 9.I, fig. 4); this septate arrangement is surrounded by a cylindrical periderm with well-developed cork (Pl. 9.I, fig. 10), leaving spaces between the xylem arms. In longitudinal section (Pl. 9.I, fig. 7), the secondary xylem arms are connected at varying intervals by transverse platforms also composed of secondary xylem tracheids, but generally containing a root trace (Pl. 9.I, fig. 9). These septate axes are up to 7 cm in diameter.

The exarch protoxylem tracheids show annular thickening (Pl. 9.I, fig. 6) and are situated between the radiating arms of secondary xylem; this arrangement is very similar to that in the roots of higher plants (Esau, 1965). The hollow regions between the xylem arms and platforms sometimes contain a few scattered parenchyma cells, and the cells at the outer periphery of the hollows appear to have been collapsing; the xylem itself often shows development of callus-like groups of cells into these spaces (Pl. 9.I, fig. 9) adding additional evidence that the hollows were present in the living axis and are not a result of incomplete preservation. The numerous rootlets arising from the axes and the growth directions mentioned above indicate that *Vertebraria* was an underground axis or root. Presumably the permineralised peat accumulated under a semi-aquatic or swamp environment and it would thus not be unreasonable to assume the hollows are some sort of aerenchyma, reminiscent of that in aquatic angiosperms where the internal aeration system is also often lysigenous (Esau, 1965).

The root traces extend from the protoxylem groups out through the adjacent xylem platforms into the rootlets (Pl. 9.I, fig. 12). In some sections the rootlets appear to be associated with the ends of the xylem arms (Pl. 9.I, fig. 11), but a platform is generally present just above or below. The smaller rootlets are diarch, others triarch to pentarch. Secondary growth in the various stages of the triarch to pentarch forms follows the same pattern as that of the original *Vertebraria* axis, the rootlets thus being merely branches although apparently exhibiting determinate growth.

Some specimens exhibit a considerable thickness of gymnospermous secondary wood around the basic septate axis (Pl. 9.II, figs. 1, 2); this wood shows distinct growth rings and may be up to 9 cm thick. The pitting in the tracheids is somewhat variable; the radial tracheid walls of the xylem arms and this surrounding wood exhibit opposite or alternate, bordered pits in up to five or six vertical rows (Pl. 9.II, figs. 3, 4), although pits are often in groups of two to seven in one or two horizontal rows (Pl. 9.II, fig. 5). These groups of pits sometimes appear to be associated with the medullary rays. The rays are uniseriate, one to twenty cells high, the higher rays occurring toward the outside of the axes. Cross-field pits vary from slightly oblique and slit-like (Pl. 9.II, fig. 6) to taxodioid or cupressoid, and range in number from one to twelve.

Associated with these deposits are numerous large trunks of gymnospermous wood up to about 40 cm in diameter. These have been referred to *Araucarioxylon arberi* (Seward) Maheshwari, 1972 (Beeston, 1972). The wood shows considerable variation in structure and varies within one trunk from forms resembling *A. arberi* to *A. ben-
Petrified Axes of *Vertebraria* 115

galenose (Holden) Maheshwari, 1972 (Pl. 9.II, figs. 7, 8). Because of the known variation of wood structure in living examples (Bailey and Faull, 1934) care should be taken in studying and comparing fossil gymnospermous woods. However, the structure of wood in parts of these trunks is exactly like that exhibited by *Vertebraria* (cf. Pl. 9.II, figs. 4, 5 and 7, 8), strongly suggesting that the *Vertebraria* were the underground axial system of these trunks (Fig. 9.2). Evidence from association also supports this conclusion, as well as the suggestion that these trees may have borne *Glossopteris* leaves (Schopf, 1970a).

ACKNOWLEDGMENTS

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On Osmundacaulis carnieri (Schuster) Miller and Osmundacaulis braziliensis (Andrews) Miller

RAFAEL HERBST

ABSTRACT

The discovery some 30 km east of Villarica, southeastern Paraguay, of a well-preserved specimen of Osmundacaulis carnieri (Schuster) Miller together with Dadoxylon-type wood and, in the same sediments, impressions of arborescent lycopods and fresh-water pelecypods belonging to one of the assemblages described for the Estrada Nova Formation of Brazil, show that these sediments are Permian in age, and probably Late Permian. The Permian age of the Independencia Series has not been accepted by palaeobotanists because of what was considered a ‘too advanced stelar structure’ for Osmundacaulis carnieri (Schuster) Miller, the only previously studied fossil. After a close comparison with Osmundacaulis braziliensis (Andrews) Miller, it is concluded that this species is a junior synonym of O. carnieri, and the few apparent differences are reconciled.

It is concluded that although the creation of a new sub-family is not yet justified, it is necessary to separate O. carnieri from the typical members of the other two accepted sub-families, and place it either as Incertae Sedis in the Osmundaceae (sensu lato) or integrate it into a ‘Permian Osmundalian complex’ with certain other species.

INTRODUCTION

A well-preserved specimen of Osmundacaulis carnieri (Schuster) Miller, and a large number of pycnoxylic-type stems which are generally known as ‘Dadoxylon’ have been found near Colonia Independencia east of Villarica, Paraguay. All the fossils were found loose in the beds of two nearby forks of a creek on the road from Mbojayaty to Colonia Independencia, 7 km west of Melgarejo (a detailed location map in Herbst, 1972). The host sediments could not be far upstream, i.e. in the Independencia Series of the northern slope of the Cordillera de Ybituruzu. Evidence for this is the small degree of rounding as well as the size of some of the specimens, and the local morpho-geographical conditions. Four or five km to the east, and in the same strata, I found impressions of arborescent lycopods which have been described elsewhere (Herbst, 1972); also in the same group of sediments, 25-30 km further northeast, I found impressions and valves of pelecypods which have not yet been studied but which appear to belong to one of the assemblages known from the Permian Estrada Nova Formation of Brazil, recently summarised by Runnegar and Newell (1971). Permian pelecypods have been recorded previously in Paraguay (Beder, 1923; Harrington, 1950) but they need critical revision. The invertebrates, the arborescent lycopods, a ‘psaroniaceous’ fern that is being described elsewhere, and the general geology of the region, leave little doubt that the sediments involved are Permian, probably Late Permian, in age. This was suggested many years ago by Beder (1923).
and accepted by others including Harrington (1950), Eckel (1959) and Putzer (1962) and recently by myself (Herbst, 1972). In reaching this judgment I have accepted the account of the stratigraphy of the 'Independencia Series' given by Harrington. He has pointed to the lithological similarity and stratigraphic position of the Independencia Series and the Estrada Nova Formation of Brazil.

Consequently the Permian age of *O. carnieri* is well established, but those referring to it from the palaeobotanical standpoint did not recognise this, and implicitly or explicitly accepted the age proposed by Schuster (1911) i.e. 'between Jurassic and Tertiary', based mainly on the apparent structural complexity of the stem which did not 'allow' a greater age for this plant. A very similar case is that of *O. braziliensis* which was originally referred to Permian sediments of the 'Municipio de Rio Pardo, Rio Grande do Sul, Brazil', but which, because of its advanced stelar structure when compared with other Osmundaceae could 'not be older than Jurassic' (Andrews, 1950). This parallelism is of interest because it will be proposed herein that *O. carnieri* and *O. braziliensis* are conspecific.

**SYSTEMATICS**

*Osmundacaulis carnieri* (Schuster) Miller 1911, *Osmundiates* carnieri Schuster: 538, pl. 20, figs 1-3
1914, *Osmundites* carnieri, in Kidston & Gwynne-Vaughan: 475, pl. 44
1950, *Osmundites braziliensis* Andrews: 32, figs 1-4
1967, *Osmundacaulis* carnieri, in Miller: 146
1967, *Osmundacaulis braziliensis*, in Miller: 146
1971, *Osmundacaulis* carnieri, in Miller: 138, fig. 1C
1971: *Osmundacaulis braziliensis*, in Miller: 137

*Type-specimen*: slide in the Kidston collection, University of Glasgow.

*Description of the new specimen*

Stem arborescent, about 90 mm diameter, surrounded by a mantle of adventitious roots 30 mm broad. Stele 33-35 mm diameter, type uncertain; it is an 'amphiphloic siphonostele' or 'dissected ectophloic siphonostele' (see below, and also Miller, 1971, for definitions). The pith, 18-20 mm in diameter, is composed exclusively of parenchyma cells 70-110 μ in diameter by 300-350 μ long. The xylem cylinder is 4-5 mm thick; it is 50-58 tracheids wide radially, composed of 10-14 strands of variable size, separated by leaf-gaps 1.2-1.5 mm wide. Protoxylem is very difficult to recognise. Metaxylem tracheids are polygonal in outline (5-7 sided), 70-105 μ when isodiametric and 50-60 μ X 130 μ when radially elongated. In longitudinal section they show typical scalariform thickenings.

Phloem and pericycle are not preserved.

A narrow reddish to dark brown band (one cell thick?) can be observed completely surrounding each strand. The band is interpreted as corresponding to the endodermis (see later discussion).

The cortex, 27-30 mm wide, is homogeneous and consists of slightly polygonal parenchyma cells, generally 70-100 μ in diameter by 220-330 μ long (ratio 1:3), similar to those of the pith. Parenchyma cells located in the adaxial concavity of leaf-traces are a bit smaller and more clearly polygonal in outline. No sclerenchyma elements could be found in any part of the cortex.

At the outermost part of the cortex, forming a neat boundary between it and the root mantle, is a continuous layer, 12-18 cells thick, of moderately thick-walled cells. They are 60-80 μ diameter by 150-180 μ long. According to Miller (pers. comm.) this tissue should be clearly distinguished from a true osmundaceous sclerenchymatic outer cortex; he considers it between a 'thick-walled parenchyma' and a 'thin-walled sclerenchyma' and is inclined to think of it as a sort of hypodermis.

Pl. 10.I. *Osmundacaulis* carnieri (Schuster) Miller. Fig. 1. Portion of stele, in which leaf-gaps and pith cells are clearly seen. x 6. Fig. 2. Detail of same with xylem clearly defined. x 15. Fig. 3. Longitudinal section to show the rather acute angle at which roots depart from leaf bundles. x 11. Fig. 4. Portion of the stem, from Andrews's specimen from Brazil (formerly *Osmundacaulis braziliensis*). (Photo Dr C. N. Miller.)
Osmundacaulis carieri and O. braziliensis
There are 38-46 leaf traces in the known sections of the cortex. They are mainly C-shaped, rather open near the stele (3.5-4 mm long by 2 mm high) and mostly rounded (4.0-4.8 mm long by 5-6 mm high) to V-shaped near the root mantle; occasionally they look slightly Y-shaped because of the budding of a root from a central position. Each leaf-trace consists of a C-shaped central xylem strand 3-4 cells thick with the cells usually radially elongated, 100-110 μ by 40-50 μ wide in the main part of the C, and more isodiametric, 70-80 μ diameter, at the tips (ends) of the traces; the leaf-traces are usually thicker at these tips. Protoxylem groups are difficult to recognise; in a few instances two or three slightly protruding xylem cells located at various points in the adaxial concavity are thought to correspond to them.

No phloem or pericycle can be seen in the leaf-traces, but a narrow reddish to dark brown band completely surrounds the xylem, and is again interpreted as an endodermis.

A root mantle up to 30 mm thick immediately surrounds the cortex. Individual roots, 0.5-1.5 mm diameter, are composed of a central ovoid diarch stele with 5-8 metaxylem cells 300-350 μ in diameter and 3-5 protoxylem cells 160-200 μ in diameter at the tips; the cortex is differentiated into an inner compact band of smaller cells, usually of 5-7 rows, and an outer less compact band with bigger cells, usually of 7-9 rows. The outer cortex insensibly grades into an external, 2-3 rowed sclerenchyma sheath.

**Material**: CTES-PB n 2878; thin sections CTES-PMP 830-837.

**Locality**: Arroyo Vista Alegre, 7 km west of Melgarejo, on the road between Mboicayaty and Colonia Independencia, Depto. Guaira, Paraguay. (Lat. 25°40’S, long. 56°10’W).

**Horizon and age**: Upper third of the Independencia Series (cf. Harrington, 1950; Putzer, 1962); Permian.

**DISCUSSION**

There is no doubt that this specimen belongs to *Osmundacaulis carnieri* (Schuster) Miller, as described by Schuster (1911), Kidston and Gwynne-Vaughan (1914) and Miller (1971). A few previously unknown or poorly known characters are added here. No essential diagnostic new characters that would allow a definitive classification of this plant could be established.

From a comparison of *O. carnieri* with *O. braziliensis* (Andrews 1950; Miller 1971; and observation of the holotype) I conclude that they are conspecific. The apparent discrepancies are easy to explain.

1. The ‘islands’ at the end of the C-shaped xylem of the leaf-traces (occupied in life by parenchyma tissue according to Andrews) are just empty spaces originally occupied by xylem. Not all the strands in the holotype of *O. braziliensis* possess these ‘islands’ and in some of the strands of the new specimen of *O. carnieri* a few tracheids are missing in the same position.

2. *O. braziliensis* has C-, V-, and Y-shaped leaf-traces while *O. carnieri* has only C- (predominantly) and V-shaped ones. The new specimen and also Schuster’s original drawing show that these outlines grade from the centre to the periphery, where the V-shape tends to become predominant; the Y-shape, as stated in the description, is simply due to the budding of a root from a central abaxial position of the leaf-trace, or it may also be due to the oblique position of the strand as it traverses the cortex near its exit; both can be seen.

3. Small quantitative differences: number of leaf-traces in a given section, size and extension of the stele, cortex, etc. may simply reflect different parts of the stem; they could also be explained by assuming a different individual age for the specimens; they may reflect slightly different microclimatic conditions.

Having explained the ‘discrepancies’, and considering that all the other characters

Pl. 10.II. Osmundacaulis carnieri (Schuster) Miller. Fig. 5. Detail of a leaf-trace. A dark line, representing the probable ‘endodermis’ can be seen in the concave side. x 15. Fig. 6. Intracortical (lower) and extracortical (upper) roots. The tissue interpreted as a ‘hypodermis’ can be clearly observed. x 16. Fig. 7. Detail of an extracortical root. x 35. Fig. 8. Portion of stem showing the shape variability in leaf-traces, in Andrews’s specimen from Brazil. (Photo Dr C. N. Miller.)
Osmundacaulis carieri and O. braziliensis
Gould (1969) strongly suspected that both species were in fact one and Miller (pers. comm.) agrees completely with the present course. While analysing the phylogenetic relations of the Osmundaceae, Miller (1971: 160) says: 'The two species of this group, O. braziliensis and O. carnieri are distinct from all other members of the Osmundaceae (Pl. 10.II, fig. 6) and neither of their axes contains features which definitely establish their relationship to this family'. His assertion is a consequence of his careful analysis of all the anatomically well-preserved species available. It is worthwhile, therefore, to point out the main characters which separate this, now single species, from the typical members of the family Osmundaceae:

1. Absence of petioles is typical of all the known specimens and so no details of stipular expansions can be ascertained. All other members of the Osmundaceae, excluding one or two Palaeozoic genera, possess these expansions, which are useful for species recognition (Gould, 1970). Their absence is not the result of poor preservation, but may indicate that the specimens are all older parts of the plant, such as the base of the stem.

2. Most species have a well-defined heterogeneous cortex: an inner fleshy parenchymatic layer and an outer sclerenchymatic one. This does not occur in O. carnieri. Sclerenchymatic elements are not known, either isolated or in groups, in any part of the cortex. The outermost sclerenchyma-like tissue is not to be considered as part of the cortex.

3. Another distinctive character, although not restricted to O. carnieri, is its stelar structure: it is a 'dictyostele' with discontinuous phloem and endodermis which totally surrounds each strand of 'meristele'; these are separated by well-developed leaf-gaps. Phloem could not be observed in any of the specimens, and the endodermis has not been definitely recognised. It is supposed to be the narrow reddish to dark brown band which occupies the place where the endodermis should be, but no cells could be distinguished (Andrews 1950: 29; this paper p. 118). Miller (1971: 118) called this stelar structure a 'dissected siphonostele' and thinks that, though it is not a dictyostele in the sense of Esau, the endodermis still surrounds completely the phloem and xylem.

The presence of an internal and external endodermis ('dictyostele') is shared with a few species of the O. skidegatensis group (sensu Miller), which owing to this character (but not to others) is closest to O. carnieri among the Osmundaceae.

4. The great size (thick xylem cylinder) and its arborescent habit are uncommon in the family. This character is shared to a certain extent again only with the O. skidegatensis group. No other known Palaeozoic genus has attained this size or habit, although they are known on the extant genus Lepopteris.

5. The departure of the roots from a central, median, abaxial position of the leaf-trace is unusual in other members of the family. In general roots depart in pairs, laterally from each leaf-trace. Miller (1971) says quite correctly that detection of this character depends to a certain extent on where the transverse section is cut; its observation then could be a matter of chance and it could be more frequent and widespread in the other species than is believed.

In view of the above considerations it is clear that O. carnieri represents a special member of the Osmundaceae (sensu lato) and has, unlike all other Permian members a more highly evolved anatomical structure. Thus, it cannot be considered a link between the two accepted sub-families Thamnopteroideae and Osmundoideae, nor is it probably on a direct line of descent from a primitive Osmundaceous archaetypal ancestor as viewed by Miller (1971: 164). It cannot be placed in either of the mentioned sub-families if we respect their present definitions, but the creation of a third sub-family to receive it is not yet warranted. The two possibilities are to leave it as an 'Incertae Sedis' member of the family Osmundaceae (sensu lato) or to place it in a 'Permian Os mundalian complex' together with some other poorly known and/or odd genera previously described.
ACKNOWLEDGMENTS

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REFERENCES


Some Problems of 'Mixed' Floras in the Permian of Gondwanaland

WILLIAM S. LACEY

ABSTRACT

This paper deals with current views on 'mixed' floras in the Permian of Gondwana. 'Mixed' floras are defined as those having an admixture of taxa found in two or more synchronous palaeofloristic regions. Gondwana during the Permian is defined as including the Hazro area of Anatolia (southeastern Turkey) and New Guinea, in addition to the usual components of Australia, India, Arabia, Madagascar, Africa, South America and Antarctica. Details are given of ten occurrences of 'mixed' floras, or indications of such floras, falling within the definitions adopted. Occurrences in Rhodesia, Portuguese East Africa, Argentina, Anatolia and New Guinea are given special consideration. Tentative explanations of the occurrence of 'mixed' floras are put forward involving combination in varying degrees of both migration of floras and their parallel evolution.

INTRODUCTION

In 1897 Seward identified a lycopod associated with Glossopteris in the Transvaal as Sigillaria brardi, a European Permocarboniferous species. Soon afterwards Amalitzky (1901) described Glossopteris leaves, associated with pareiasaurid reptiles, from the Upper Permian of the northern U.S.S.R. These two early descriptions raised the question as to whether migration of elements of northern hemisphere and Gondwana floras in opposite directions could have taken place. However, the leaves originally identified by Amalitzky as Glossopteris were later transferred by Zalessky (1933) to a new genus Pursongia, a course of action supported by Meyen (1967) who showed that they had nothing to do with Glossopteris; and Seward's Transvaal Sigillaria was renamed Lycopodiopsis by Edwards (1952) who showed that there was no satisfactory evidence for the presence of any northern fossil lycopod genus in the southern hemisphere. If these two now discounted instances had been the only ones the idea of mixing of floras might not have persisted, but the account by Walton (1929) of a flora in the Upper Wankie Sandstone of Rhodesia containing not only numerous species of Glossopteris but also presumed northern hemisphere species of Sphenophyllum and Pecopteris seemed to support the idea of the migration of some species from the north.

Opposing views of migration and of parallel evolution have been proposed to explain the Rhodesian situation. The controversy has continued unabated, the more so as further examples of such 'mixed' floras have come to light elsewhere in South Africa, in South America, Anatolia (southeastern Turkey), New Guinea, India, and possibly in Australia. Some of these more recent occurrences are mentioned by Chaloner and Lacey (1973). The present paper reviews the position more fully and provides data for a more detailed consideration of the problems involved.
SOME DEFINITIONS

Definition of 'Mixed' Floras

Opposition to the idea of migration to account for 'mixed' floras and a firm belief in an alternative theory of separate development of Gondwana floras (with implications of homoplasy or parallelism where necessary) are put forward by Plumstead (1962), who speaks of 'the myth of mixed floras' and states (p. 106) that

the acceptance of a theory of separate development implies that the term 'mixed' flora should no longer be used to mean a mingling of northern and southern elements, but rather to imply an environment in which the early moisture-loving plants could survive with the new and hardier Glossopteridae.

This new use of the expression 'mixed' flora in an ecological rather than a derivational sense is further expanded by Plumstead (1966). In this paper the expressions 'mixed' Glossopteris zone and stage are also used. Thus the position has now been reached where the adjective 'mixed' has been applied to floras, environments, zones and stages, that is, to taxa, their surroundings and time. Plumstead (1966) explains that, for her, a 'mixed' flora represents a position in the evolution of Carboniferous and Permian floras in Africa where representatives of families other than the Glossopteridae still occur in quantity, together with representatives of the Glossopteridae themselves, before the development of 'a pure Glossopteris flora' (p. 8) in Middle and Late Permian times. This definition, still used by Plumstead (1973), and so different from that of most authors, constitutes the first problem in dis-

Fig. 11.1. Pre-drift continent positions and floras in Early Carboniferous time (after Chaloner and Lacey, 1973). ← ? possible migration route.
cussing ‘mixed’ floras in Gondwanaland. It necessitates a definition of the meaning of the term in the present context.

For the present purpose a definition modified from that of Archangelsky and Arrondo (1967) is adopted. A ‘mixed’ flora is one containing an admixture of morphographically-defined taxa (species or genera) found in and characteristic of two or more synchronous palaeofloristic regions (the phytocoria of Meyen, 1969), e.g. Eurameria, Angara, Cathaysia, Gondwana.

Such a definition admits of other mixtures, in addition to an exchange of northern hemisphere and southern hemisphere taxa, and also leaves open for later discussion the question whether what may be called the ‘critical’ or ‘contaminating’ taxa occurring in the localities of the ‘mixed’ floras have arrived there as a result of migration or as a result of homoplasy (parallel or convergent evolution) from earlier common ancestral stocks.

**Definition of Gondwanaland in Permian Times**

For the purpose of the present paper Gondwanaland in the Permian is defined as indicated in the map provided by Wagner (1962), that is, including Anatolia (southeastern Turkey), and New Guinea, in addition to the usual components of Australia, Peninsular India, Arabia, Madagascar, South Africa, South America and Antarctica. This is approximately as given in the pre-drift positions of the continents suggested by Smith and Hallam (1970) used as the basis for Figure 6 of Chaloner and Lacey (1973), and reproduced as Figure 11.3 in the present paper.

**Records of ‘Mixed’ Floras in Gondwanaland**

In the Gondwana region, as defined above, eleven occurrences of ‘mixed’ floras, or suggestions of such floras, are now known. Details are available for ten of them. In four the typical *Glossopteris* flora is ‘contaminated’ by only one presumed northern taxon, and in three of these (Zambia, Brazil, Australia) the identity of this single taxon is not yet established beyond doubt. These cases will not figure largely in the subsequent discussion. Indeed, some authors would probably not accept them as examples of ‘mixed’ floras at all. For example, Meyen (1970: 554), while allowing that there was some penetration of some northern (Euramerian-Cathaysian) elements into both Gondwana and Angara floras, states that this ‘has not discomposed the general order of local plant assemblages, so one cannot observe here the formation of a true mixed flora’. Be that as it may, details are given for the ten occurrences for which data are available. Except in the special cases of Anatolia and New Guinea, where full species lists are given, only the ‘critical’ or ‘contaminating’ taxa are listed.

**South Africa**

1. Rhodesia. Wankie. (c. 18°40’S, 26°50’E) (after Walton, 1929; revised by Huard-Moine, 1964a, 1964b, 1965; Lacey and Huard-Moine, 1966)

   *Sphenophyllum thonii* Mahr. s.l.
   *S. oblongifolium* (Germar & Kaulfuss) Unger
   *S. cf. verticillatum* (Schlotheim) Brongniart
   *S. wankianum* Huard-Moine (this species also occurs in China)
   *Pecopteris arcuata* Halle
   *P. unita* Brongniart
   *Asterotheca hemitelioides* Brongniart
   *Chansitheca cf. kidstoni* Halle

The above list represents about one-fifth of the Wankie flora which now totals approximately thirty-eight taxa, including ten species of *Glossopteris* and *Gangamopteris*, *Sphenophyllum speciosum* Royle and *Noeggerathiosis hislopi* (Bunbury) Feistmantel—see Lacey and Huard-Moine (1966).


   *Sphenophyllum thonii* Mahr
   *S. oblongifolium* (Germar & Kaulfuss) Unger
   *S. cf. verticillatum* (Schlotheim) Brongniart


   *Annularia sphenophylloides* (Zenker)


   *Sphenophyllum thonii* Mahr
I2Ö Gondwana Flora

N. American & Cathaysian

Mixed Cathaysian/Euramerican

Mixed Glossopteris/Euramerican

Angara

Fig. 11.2. Pre-drift continent positions and floras in Late Carboniferous/Early Permian time (after Chaloner and Lacey, 1973). ←→ Possible migration route.

South America


*Sphenophyllum* cf. *oblongifolium* (Germar & Kaulfuss) Unger


*Sphenophyllum* thonii Mahr

*Pecopteris anderssonii* Halle

*P. unita* Brongniart

cf. *Asterotheca hemitelioides* Brongniart

7. Argentina. Chubut. (c. 44°32'S, 70°22'W) (after Archangelsky, 1960)

*Sphenophyllum* cf. *oblongifolium* (Germar & Kaulfuss) Unger

S. cf. *cuneifolium* (Sternberg) Zeiller

As in the case of the Rhodesian flora, these lists combined represent a fairly high proportion of the total Argentine flora which includes nine species of *Glossopteris*, three species of *Gangamopteris*, *Sphenophyllum speciosum* Royle and *Noeggerathiopsis hislopii* (Bunbury) Feistmantel. With reference to the species of *Sphenophyllum* (other than *S. speciosum*) Archangelsky and Arrondo (1970) make the point that 'some species cannot be differentiated from “northern” types'.

Australia

8. Western Australia. Irwin River Area, Perth Basin. (c. 33°15'S, 116°5'E) (after Rigby, 1966)

*Sphenophyllum rhodesii* Rigby (1966)

This species seems very close to, if not identical with, *S. thonii* Mahr, but Rigby does not make any reference to *S. thonii* in the discussion of his new species.
Some Problems of 'Mixed' Floras

Turkey
Full list of taxa; C, G, A, and E indicate Cathaysian, Gondwana, Angara and Euramerian elements.
Gigantopteris nicotianaefolia Schenk C
Glossopteris cf. stricta Bunbury G
Taeniapteris sp.
Zamiopteris sp.
Gondwanidium validum (Feistmantel) Gothan G
This is given as Dicroidium? vel Thinnefeldia? sp. by Wagner (1962). The comparison with Gondwanidium validum made by Lacey (1962) is supported by Archangelsky (1971, pers. comm.)
cf. Angaropteridium cardiopteroides:
(Schmalhausen) Zalessky A
Cladophlebis roylei Arber G
Pecopteris tenuicostata Halle C
'Validopteris' sp. (cf. Pecopteris arcuata Halle pars) C
Pecopteris cf. wongi Halle C
P. phegopteroides (Feistmantel) G
P. jongmansii Wagner E
P. tenuidermis Wagner

New Guinea
10. Western New Guinea (c. 5°0'S, 135°50'E) (after Jongmans, 1940; Hopping and Wagner, in Visser and Hermes, 1962; and Kon’no, 1966)
Jongmans (1940) described a Cathaysian flora (Locality A list), resembling the Djambi flora of Sumatra, and a Glossopteris flora, based on fine specimens of Vertebraria, from two localities about 10 km apart, near the southern coast of West New Guinea in the area of the Otakwa and Akimeugah rivers, near Apimoe. Both of these localities are south of the Tertiary collision zone between the Australian block and the Indonesian arcs—that is, they are on the Gondwana landmass. Hopping and Wagner (1962), in Visser and Hermes (1962), described a Glossopteris flora from two more localities a considerable distance west of the Jongmans localities. Their records are presented as a

Fig. 11.3. Gondwanaland in Permian time, showing localities for Palaeozoic floras (after Chaloner and Lacey, 1973). ± possible migration routes.
composite list (Locality B list), and came from the area of the Aria and Aidoena rivers, south of Geelvink Bay, and from South Vogelkop. These localities are just about in the region of the collision zone (Aria/Aidoena) and clearly north of it (South Vogelkop) respectively. Consequently it does not seem possible to explain the mixing of floras in this region by resort to plate movement.

**Locality A.** Cathaysian flora (after Jongmans, 1940)
- *Sphenophyllum verticillatum* Schloth
- *Pecopteris unita* Brongniart
- *P. cf. arcuata* Halle
- *P. cf. paucinervis* Jongmans
- *P. cf. orientalis* Schenk
- *Taeniopteris cf. multineris* Weiss
- *T. cf. taiyuanensis* Halle

**Locality B.** Glossopteris flora (after Hoppening and Wagner, in Visser and Hermes, 1962)
- *Glossopteris cf. browniana* Brongniart
- *G. cf. indica* Schimper
- *G. aff retifera* Feistmantel
- *Vertebraria* sp.
- *Taeniopteris cf. hallei* Kawasaki
- *Cladophlebis cf. australis* (Morris)
- *Pecopteris monyi* Zeiller
- *Validopteris sp.* (cf. *P. arcuata* Halle)
- *Pecopteris unita* Brongniart
- *Sphenophyllum cf. speciosum* Royle

**DISCUSSION**

Although Plumstead (1962, 1966, 1973) has consistently taken the view that the Gondwana flora was a separate development and that the presence of certain taxa resembling northern species was probably due to parallelism, several other authors have recently expressed the view that some mixing of floras may have occurred. For example, Archangelsky and Arrondo (1967: 76) state that ‘the presence of “northern” taxa together with Glossopteridales in certain areas of the Gondwana Region may now be accepted as a fact’. Meyen (1967: 149) believes that it is not necessary to ‘exclude floral interchange from reconstructions of plant life of the Upper Palaeozoic’ and goes on to say I consider that floral interchange might well play a more significant role in the Late Palaeozoic. There was some interchange between Angara, Cathaysian and Euramerian floras. Gondwana plants hardly penetrated northwards. The latter penetration might have been hindered not only by the Tethys but also by a hypothetical belt of thermophylic and hydrophylic floras.

In more recent papers (Meyen, 1970; Chaloner and Meyen, 1973), as Plumstead (1973) points out, a somewhat more cautious line is taken as to the explanation of the occurrence of ’mixed’ floras, although the fact of their existence is not questioned.

In this paper the data presented in the previous section are condensed into three main groupings for more detailed consideration. These are South Africa and South America taken together, Anatolia, and New Guinea.

**South Africa and South America**

The Wankie flora of Rhodesia and the Tete flora of Portuguese East Africa, with at least ten species in common, are so similar that for practical purposes they can be considered as representing one ’mixed’ flora occurring in the Zambezi Basin area. This Rhodesia/P.E.A. ’mixed’ flora, as Archangelsky (1958) has already pointed out, is very similar to the mixed flora of Argentina, having at least sixteen species in common, including both ’northern’ and *Glossopteris* flora elements. South Africa and South America are therefore here considered together as one land mass in the context of ’mixed’ floras.

In this South Africa/South America land mass the localities for ’mixed’ floras occur near to or not far within the limits of glaciation and not far from localities for Early Carboniferous (*Lepidodendropsis*) floras lying just outside the limits of glaciation (see Fig. 11.3). Chaloner and Meyen (1973) point out that all the examples of ’northern’ genera in these localities are pteridophytic, spore-producing and probably homosporous plants, and imply that their presence in southern localities might have been effected by the dispersal of wind-borne spores. An alternative and not necessarily mutually exclusive suggestion is provided by Chaloner and Lacey (1973), namely that these
'mixed' floras might have originated by the migration inwards of 'northern' genera from peripheral Early Carboniferous stocks and outwards of *Glossopteris* flora after the retreat of the ice in those areas. This suggested explanation invokes (a) relatively short-distance migration of 'northern' genera (certainly aided by the dispersal of light propagules) and of *Glossopteris* flora in opposite directions, and (b) homoplasy, or parallel evolution, in the South Africa/South America land mass of 'northern' taxa from Early Carboniferous peripheral stocks, rather than long-distance migration from northern Africa, Europe or northern India. An objection to this suggestion is the complete absence, so far, of any indication of sphenopsids or pecopterid pteridophylls as supporting evidence in the peripheral Early Carboniferous localities.

**Anatolia**

According to Wagner (1962) the Hazro flora includes Gondwana and Cathaysian elements. This view is supported by several authors. For example, Archangelsky and Arrondo (1967: 81) state that the figured specimens of *Glossopteris* are correctly identified and subsequently Archangelsky (1971, contribution to discussion, Krefeld Meeting) supported the suggestion made by Lacey (1962) that the *Dicroidium* ? sp. of Wagner (1962) could be the basal part of *Gondwanidium validum* (Feistmantel) Gothan, a species originally described from India. The Gondwana element in the Hazro flora therefore consists of *Glossopteris* cf. *stricta* Bunbury, cf. *Gondwanidium validum* (Feistmantel) Gothan, *Cladophlebis roylei* Arber and *Pecopteris phegopteroides* (Feistmantel), the last-named also known from India. Speaking of the Cathaysian taxa, Meyen (1970) states that 'their presence in Anatolia is confirmed by good illustrations (Wagner, 1962)'.

How is the occurrence of Gondwana and Cathaysian elements in the Hazro flora to be explained? It seems reasonable to suggest that the Gondwana element arrived by migration northwesternwards from northern India. Such a source is perhaps more likely than the almost equally-distant currently known northernmost African source at Entebbe in Uganda, since the latter would entail migration of a flora including seed-plants across a land mass, while migration from India would allow the possibility of movement along a low-lying coastal route (see Figs. 11.1 to 11.3).

It would also seem reasonable to suppose that the Cathaysian element arrived by migration along a northern shore route through Central Asia, for at first sight this is supported by the occurrence of a large number of Cathaysian plants at Madygen in South Fergana (Sixtel, 1960). But the total absence of any Cathaysian plants in European U.S.S.R. or Europe and the revision of the Madygen flora, with re-dating to Triassic (Meyen, 1970), make this route unlikely.

Could it be that the Cathaysian element arrived by a route in part similar to that for the Gondwana element, involving Sumatra, New Guinea, Western Australia and northern India? If this were so, then migration along a southern shore route could not have taken place in Permian time, since New Guinea and Australia were by then already well separated from Sumatra, Malaysia and China (see Fig. 11.2).

An alternative possibility is that migration from China of Early Carboniferous (*Lepidodendropsis*) flora stocks, with the genetic potentiality subsequently to evolve Upper Carboniferous/Permian Cathaysian taxa, took place along a southern shore route in Early Carboniferous time (Fig. 11.1). This suggestion receives a measure of support from the occurrence of Early Carboniferous *Lepidodendropsis* flora in Australia and northern India, and the possibility that *Sphenophyllum rhodesii* Rigby in Western Australia, apparently very similar to *S. thomii* Mahr, might indicate a weak development there of a Cathaysian element. Until recently the most serious objection to this suggested southern shore route has been the absence of any records of Cathaysian elements in northern India. Indeed, Archangelsky and Arrondo (1967) have already pointed out the need for search for 'northern' (Eurameric and Cathaysian) taxa in other parts of Gondwanaland and particularly in India.

Now comes news from Meyen (pers. comm.)
not only that he has seen in the Geological Survey of India, in a collection made by Kapoor, elements of the Cathaysian flora together with Glossopteris from Kashmir, but also that some typically Cathaysian plants, including Lobatannularia heianensis (Kodaira), have been collected by Czechoslovak geologists from the Ga’ara Formation in Iraq and identified by Meyen himself. With this valuable additional evidence the idea of a southern coastal migration route gains considerable support.

New Guinea

The stratigraphic relationship of the localities with the Cathaysian flora to those with the Glossopteris flora remains unknown. Hence it is not clear whether the two floras are contemporaneous or whether, as Kon’no (1966) believes, the Cathaysian flora is Early Permian and the Glossopteris flora Late Permian.

In any event, the occurrence of an important flora of Late Carboniferous (Westphalian E) age at Djambi in Sumatra (Jongmans, 1937) having much in common with the Cathaysian flora of Shansi (China) on the one hand and the New Guinea flora on the other, strongly suggests a route for migration via these islands in Early Carboniferous times (Fig. 11.1) before the separation of New Guinea from Sumatra by Permian time (Fig. 11.2).

The apparently isolated occurrence of a Cathaysian flora in West New Guinea would then be seen as the result of evolution of Carboniferous stocks received from China via Sumatra, while the presence of the Glossopteris flora in West New Guinea might be explained by migration northwards from northwest Australia, which in recent reconstructions is closely associated with New Guinea and where not-far-distant localities for Glossopteris exist.

It may well be, therefore, as Kon’no (1966) believes, that the New Guinea Glossopteris flora is younger than its Cathaysian flora, but this does not invalidate the argument that migration southwards of Early Carboniferous stocks with the potentiality for the production of Late Carboniferous/Permian Cathaysian floras elsewhere took place before the destruction of this migration route. The migration northwards from Australia to New Guinea could have been contemporaneous with the southbound migration or could have been later.

Summary and Conclusions

It is suggested that
1. the ‘mixed’ floras of South Africa and South America resulted from a combination of the migration inwards of ‘northern’ elements from peripheral Early Carboniferous stocks and a migration outwards of Glossopteris flora.
2. the Gondwana element in the Anatolia ‘mixed’ flora arrived by migration from northern India.
3. the Cathaysian element in the Anatolia ‘mixed’ flora originated from Early Carboniferous stocks with Cathaysian potentialities in China migrating and evolving en route via Sumatra, New Guinea, Western Australia, northern India and Iraq.
4. the ‘mixed’ flora (or, more correctly, the two floras) of West New Guinea resulted from a combination of migration southwards from China via Malaya and Sumatra of Carboniferous stocks with Cathaysian potentialities and a contemporaneous or later migration northwards of Glossopteris flora from Australia.

These suggestions involve combinations of migrations and homoplasy in varying degrees in different parts of Gondwanaland. There are many gaps and deficiencies in such suggestions, notably the absence of any indication of sphenopsids and pecopterid pteridophylls in the Early Carboniferous localities of South Africa and South America, the absence of any Early Carboniferous floras in Sumatra and New Guinea, and the absence of reliable evidence of ‘mixed’ floras in Australia. Furthermore, the conception of the Early Carboniferous Lepidodendropsis flora as a uniform worldwide flora is called into question. Meyen (1969) allows some differentiation of Early Carboniferous floras. It may well be the case (and would fall into line with suggestions made above) that the apparently isolated Chinese Early Carboni-
Europe, North America, Africa and South America.

ACKNOWLEDGMENTS

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Indian Lower Gondwana Floras: A Review

K. R. SURANGE

ABSTRACT

The Gondwana sediments and their distribution on the Indian sub-continent are reviewed in a number of articles published on this and preceding symposia. They contain abundant fossil plants that serve to subdivide different groups of exposures. The Lower Gondwana, with which we are concerned, contains a rich *Glossopteris* flora. A succession of distinct mega- and microfloras has been worked out in recent years from the various stages of the Indian Lower Gondwana System: Talchir, Karharbari, Barakar, Barren Measures and the Raniganj.

THE TALCHIR STAGE

The Talchir Stage occurs at the base of the Gondwana sequence and begins with the deposition of glacial sediments in different Gondwana basins of India. The strata rest directly on the Precambrian basement and are overlain by the coal-bearing Karharbari Formation. According to Ghosh and Mitra (1970) the Talchir is characterised by a varied assemblage of lithofacies such as tills, fluvio-glacial outwash sandstones and conglomerates, glacio-lacustrine shales, rhythmites and turbidites.

The exact relation of the Lower Gondwana flora to the late Palaeozoic glaciation has been much debated (Sahni, 1938). According to Sahni, the *Glossopteris* flora had already come into existence before the land was free of ice and the widespread glaciation could hardly have destroyed all traces of life over the whole of the Gondwana continent. On the other hand, Fox (1931, 1934), who made careful study of the Lower Gondwanas, believed that in India the *Glossopteris* flora came in long after the ice had disappeared. In support of his contention Sahni (1938) cited Virkki's (1937) discovery of typical Lower Gondwana spores from two horizons at Kathwai in the Salt Range in Pakistan, only about 0.45 m and 1.25 m above the boulder beds. Sahni further stated that the time interval represented by the overlying 0.45 m of sediments must have been so short that the lowermost of the spore-bearing horizons may be taken to be of approximately the same geological age as the glacial bed.

In 1956 Surange and Lele discovered some stunted forms of *Gangamopteris* and *Noeggerathiopsis* in the needle shales of the Talchir Stage collected from 3 m above the boulder bed in Giridih coalfield, which also yielded many monosaccate spores. The needle shales are believed to be the end product of crushed and powdered rocks of various kinds—the result of the abrasive action of the slow moving ice sheets. It is believed that this rock powder was deposited as the ice sheets thawed on waters of extensive lakes. The preserved hardy plant remnants must have been thriving nearby and were buried by the suspended silt released by the thawing ice sheets. This discovery, therefore, indicates that the early phases of *Glossopteris* flora overlapped with the last phases of glaciation.

Until recently there was no evidence of plant fossils from the Talchir Boulder Bed itself. In 1970 Lele and Karim described about forty miospore genera from two intercalated Talchir Boulder Beds in the Jayanti
The flora is dominated by monosaccate taxa like Virkkipollenites, Plicatipollenites and Parasaccites and has twenty-three new elements. However, in view of the faulted nature of the Talchir metamorphic contact, the two Talchir Boulder Beds would appear to be much above the true base of the Talchir Stage. This would suggest that the microflora reported from Jayanti coalfield, although coming from the boulder bed itself, cannot be considered as the earliest representative of the Glossopteris flora.

Lele (1966) had earlier reported spores, which included ?Punctatisporites, Plicatipollenites and Virkkipollenites from siltstones lying immediately above the Talchir-metamorphic contact in the West Bokaro coalfield. This remains the earliest record of plant life, almost at the beginning of the Gondwana sedimentation in India.

Needle shales, which generally follow the Talchir boulder beds, have yielded mega- as well as microfossils in a number of coalfields. Megafossils and spores were reported from shales in the Giridih Coalfield and Johilla Coalfield in the South Rewa Basin (Surange and Lele, 1955, 1956). Spore assemblages in the Johilla Coalfield were dominated by the monosaccate forms, such as Plicatipollenites, Virkkipollenites and Stellapollenites. Thus the evidence is accumulating for the presence of Glossopteris flora at different levels in the Talchir Stage—the boulder beds, siltstones lying just above them, and the needle shales.

There is also evidence that Tethys transgressed the Indian peninsula during the Permian as far southwest as Manendragarh and Umaria in Central India. The earlier contention was that an arm of the sea ran across the Vindhyan, having a connection to the west and northwest. A new interpretation was offered by Ahmad (1961) on the basis of close faunal relations of the Umaria fauna with those of Assam and Australia (Lyons Group). According to Ahmad ‘the connection of the Umaria fossil bed was to the South East where marine conditions might have prevailed over a fairly extensive area’. More recently a different interpretation has been given by Sastry and Shah (1969), who postulate that the Manendragarh transgression was older (early Talchir Stage) and came from the northeast, whereas the Umaria transgression came much later (?early Karharbari) from the west.

The marine beds near Manendragarh have recently yielded characteristic Talchir miospores in association with a few marine acritarchs (Lele and Chandra, 1972). In the Umaria marine beds the same workers found a number of acritarchs such as Leiosphaeridia and Foveofus. Unlike Manendragarh, spores are few at Umaria and they include some monosaccate types.

The progressive increase in the number and variety of fossil plants and spores from the lower to the upper part of the Talchir is a reflection of the return of more favourable conditions. Fossil plants showed relatively greater abundance and variety in the Talchirs in the South Rewa Basin, Singrauli, Auranga and Deogarh Coalfields. The maximum development of the flora is seen in the Rikba Beds (North Karanpura Coalfield) which represent the uppermost part of the Talchir flora. There is also an abundance of Samaropsis type of seeds, besides occasional equisetaceous plants and conifers such as Paranoclados. Glossopteris is absent in the Talchir except in the uppermost Rikba Beds, where it is represented only by two species. Glossopteris may, therefore, be taken as an indicator of more favourable conditions, as it proliferates steadily through the Karharbari, to rise to a dominant position in the subsequent floras. At the same time Gangamopteris, which declines above the Karharbari Stage, possibly required an environment much cooler than that of the succeeding Barakar and Raniganj. The Talchir and Karharbari record a distinct flora, which may be termed the Gangamopteris-Noeggerathiopsis flora.

The miospores, preserved mainly in the shale facies, also indicate a distinct flora. Towards the upper part of the Talchir, dominant monosaccates were associated irregularly with a few trilete, alete, disaccate and monocolpate miospores, such as Horriditrilletes, Jayantisporites, Callumisporea, Suamipollenites, Lunatisporites and Ginkgoecycadophyllum. The occurrence of these taxa is, however, not uniform in the Talchir of various Lower Gondwana basins. Alete forms such
as *Quadrisporites* and *Leiosphaeridia* are also met with in the upper Talchir miofloras. Megaspores have recently been discovered in the Talchir, the most prevalent genus being *Duosporites*. The most abundant miospore genus in the Talchir is *Parasaccites*, and it is commonly associated with *Plicatipollenites*, *Potonieispores*, *Caheniasaccites*, *Virkkipollenites* and *Callumispora*. This association occurs in the Talchir of Giridih, South Jhilla, Raniganj, Korba, Jayanti and North Karanpura basins.

**KARHARBARI STAGE**

The Talchir Stage is succeeded by a group of coalbearing strata known as the Karharbari Formation. It is 60-120 m thick and is of wide geographical occurrence. The rocks are grits, conglomerates, sandstones and shales, containing seams of coal. Blanford (1876) first established this as a distinct stage on the evidence of plant fossils such as *Gangamopteris*, *Noeggerathiopsis*, *Gondwanidium* and *Buriadia*, which were absent in the true Damuda flora. He placed it along with the Talchir Stage under the Talchir Series. Fox (1931), on the other hand, placed Karharbari under the Damuda Series and regarded it as the basal beds of the overlying Barakar Stage. Bharadwaj (pers. comm.) has recently suggested that the basal part of the Karharbari should be included in the Talchir Series and the upper part should be regarded as the basal horizon of the Damuda Series.

The type area for the Karharbari lies in the Giridih Coalfield. Plant fossils from this area have been investigated by many authors including Feistmantel (1879), Zeiller (1902), Seward and Sahni (1920), Sahni (1928), Lele and Maithy (1964) and Maithy (1966, 1970).

The Karharbari fossil plant assemblage is rich in variety, and according to Maithy (1966, 1970), comprises at least twenty-three genera. The important genera are as follows: *Schizoneura* (2 spp.), *Phylotheca* (2 spp.), *Gondwanidium* (2 spp.), *Sphenopteris* (1 sp.), *Gangamopteris* (17 spp.), *Glossopteris* (7 spp.), *Rubidgea* (2 spp.), *Vertebraria* (2 spp.), *Noeggerathioptis* (11 spp.), *Euryphyllum* (2 spp.), *Buriadia* (1 sp.), *Cordaicarpus* (2 spp.), *Samaropsis* (6 spp.), *Ottokaria* (1 sp.), *Arberia* (2 spp.).

Pant and Singh (1968) described cuticle of five species of *Gangamopteris* from Giridih and Kurasia coalfields. Pant and Nautiyal (1967) have described in detail the morphology of *Buriadia sewardii*.

The plant assemblage in the Karharbari is still dominated by *Gangamopteris*, with *Glossopteris* as a close second. *Noeggerathioptis*, with eleven species, attained its maximum development in this stage, and *Gondwanidium* and *Buriadia* are restricted to it. Quite a few seed genera also occur in the Karharbari, along with new gymnospermous forms such as *Rubidgea* and *Euryphyllum*. Some new forms such as *Ginkgophyton*, *Palmatophyllites* and *Dolianitia* have recently been reported by Maithy (1965a, 1965b).

On the whole the Karharbari flora is distinct from the underlying Talchir and the overlying Barakar flora. It has been worked out mostly from the type area, the Giridih Coalfield, where it is best preserved. But in recent years, the characteristic genera *Gondwanidium* and *Buriadia*, along with other Karharbari forms, have also been discovered widely distributed in a number of other Lower Gondwana basins (Maithy, 1969) such as Jayanti, South Karanpura, North Karanpura, Raniganj, Jhilla, Chirmiri, Umaria, Mohpani, Hutar, Auranga and Daltonganj. Although in many of these areas it is still disputed whether Karharbari should be recognised as a distinct stage, floristically at least the plant-bearing strata are equivalent to the Karharbari of the Giridih Coalfield.

The underlying Talchir flora shows a closer relationship with the Karharbari than with the Damuda flora. It is as if the Talchir flora diversified further in the Karharbari, bringing in new elements but at the same time maintaining the dominance of the *Gangamopteris-Noeggerathioptis* complex. The miospore assemblages of the Talchir and Karharbari show the same relationship. The dominant miospore forms in both are very nearly the same.

The Karharbari mioflora is best known from the Giridih Coalfield (Maithy, 1965c).
and the equivalent strata in North Karanpura, Korba, Jayanti and Mohpani Coalfields (Bharadwaj and Anand Prakash, 1972; Bharadwaj and Srivastava, 1973, Lele and Makada, in press). The monosaccates of the Talchir continue to dominate the lower Karharbari mioflora, but they are associated with large numbers of the trilete genus Callumispora. The spore assemblage of the lower Karharbari is Callumispora complex dominant and Parasaccites complex sub-dominant, as was discovered in the carbonaceous strata including the lower Karharbari seam overlying the Talchir sediments in the Giridih Coalfield. A similar assemblage has been reported from equivalent stratigraphic positions in the sub-surface sections in the Korba and the North Karanpura coalfields, and surface exposures in the Jayanti and the Raniganj coalfields. Besides, some taxa like Crucisaccites appear to be peculiar to the Karharbari Formation. In the North Karanpura sub-surface studies, the lower Karharbari flora has been recognised (Kar, 1973) by the dominance of the triletes Punctatisporites (=Callumispora), Cyclobaculispora, Microbaculispora, Indotriradites and Lacinitriletes, along with the sub-dominance of Parasaccites, Virkkipollenites, Plicatipollenites and Caheniscaccites.

In the upper Karharbari, the monosaccates appear to decline and the dominating elements are the non-striate disaccates, notably Sulcatisporites. Striate disaccates (Striatites, Faunipollenites, Lunatisporites) and some trilete miospores also appear in recognisable proportion. In North Karanpura it has been found that the upper Karharbari is characterised by the dominance of monosaccates, with a sub-dominant population of disaccates, and some triletes. According to Bharadwaj, the upper sub-stage, overlying the Lower Karharbari seam, comprising sandstone and upper Karharbari seam, contains an assemblage dominated by Illinites and others of the Sulcatisporites complex, associated with an appreciable proportion of the Parasaccites complex. However, in Korba and North Karanpura sub-surface, according to Bharadwaj, the Callumispora complex dominated the lower Karharbari sub-stage. It is directly overlain by a Parasaccites dominated complex, followed by a Sulcatisporites dominant complex and Parasaccites sub-dominant complex.

The Parasaccites complex dominated assemblage, which initiated the upper Karharbari sub-stage, has been interpreted by Bharadwaj (pers. comm.) as corresponding with the second phase of glaciation in the lower Gondwana of India, on the evidence that the radial monosaccates have always been found associated with glacigene sediments in India and other Gondwana countries.

Karharbari miofloras are being intensively studied at present with a view to distinguishing this unit from the overlying Barakar Formation. So far the general composition of the Karharbari miofloras appears to be sufficiently distinct to identify the unit in a number of basins where lithological characters are not diagnostic.

BARAKAR STAGE

The Talchir Series is succeeded by the Damuda Series, which is divided into three stages: Barakar, Ironstone Shale or Barren Measures, and Raniganj. The Barakar Stage is widely distributed in the Gondwana basins outside Bengal in the Satpura and the Mahanadi and the Godavari Valleys. The Barakar Stage rests conformably upon the Talchir Series and consists of coarse, soft, white, massive sandstones and shales with coal seams. This is the chief coal-bearing formation in India and in recent years has been extensively studied palynologically for the correlation of coal seams in different coalfields.

The megaflora of the Barakar is characterised by the dominance of the genus Glossopteris and the absence of the Gangamopteris-Noeggerathiopsis complex which was dominant in the Talchir and Karharbari. Gangamopteris might be present in the basal part of what is called the Barakar (but which may represent the Karharbari) in some coal basins, but it is absent in a large number of coalfields of true Barakar age. Glossopteris is represented by at least ten species; there may be many more (Srivastava, 1957; Kulkarni, 1971). The recognition of Glossopteris species on the external features of fossil leaves is a tricky problem,
and disagreement between authors is not uncommon. It is, therefore, difficult to say which species are confined to the Barakar. Many of them, no doubt, continue into the Raniganj Stage (see Table 12.1) and some species such as *G. indica* and *G. communis* have a wide range, extending from the Talchir to the Raniganj. But, as proved in recent years by the work on epidermal studies, these two 'species' (and many others) are made up of more than two species. Only if we succeed in precisely defining *Glossopteris* species on the external characters of the leaves, will it be possible to know their precise ranges. However, the dominance of *Glossopteris* in the Barakar is unmistakable.

Table 12.1. Distribution of Plant Fossils in the Lower Gondwana Formations of India

<table>
<thead>
<tr>
<th>Name of plant fossil</th>
<th>Talchir</th>
<th>Karharbari</th>
<th>Barakar</th>
<th>Barren</th>
<th>Measures</th>
<th>Raniganj</th>
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<td><strong>A. LYCOPODS</strong></td>
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- *G. karharbariens*  
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- *G. macronata*  
- *G. oblaneelata*  
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Pteridophytic remains are few in the Barakar as compared to the Raniganj, but many equisetaceous genera are present. They are represented by five species of Phyllotheca and one species each of Schizoneura and Stellotheca. The last mentioned species is only from the Barakar Stage in the Rajmahal Hills of Bihar. Phyllotheca angusta Surname and Kulkarni (1968) has crowded stomata present on only one surface of the leaf. The two guard cells are completely surrounded by two subsidiary cells and do not show the characteristic rib-like thickenings present in the modern Equisetum. Sphenophyllum in the Indian Lower Gondwana is monotypic and is also reported from the Barakar. The fern-like plants are represented by Sphenopteris, Pecopteris and Alethopteris, each genus with one or two species. The conifers are represented by Walkomiella, which is not known from the Raniganj.

A number of petrified woods belonging to extinct trees are known from the Barakar Stage. They have no stratigraphical significance but are interesting botanically. All of them were previously grouped under the form genus Dadoxylon. The secondary xylem in most of them is relatively uniform, but some show pith and primary xylem. A number of genera have been created for wood, utilising various characters including those of the pith and primary xylem. The Barakar contains three genera and nine species. They are: Araucaryoxylon bengalense, Araucaryoxylon barakarense, Araucaryoxylon gondwanense, Araucaryoxylon kharhariense, Damudoxylon parenchymosum, Damudoxylon indicum, Polysolenoxylon jhariense, Polysolenoxylon krauselii and Polysolenoxylon canalosum (Maheshwari, 1972).

The Barakar Stage, being the chief source of coal in India, has been extensively studied palynologically. In the Giridih Coalfield, the sequence overlying the upper Karharbari seam is characterised by a Sulcatisporites dominated miospore assemblage. A similar type of assemblage is obtainable in the subsurface samples of the Korba Coalfield (Bhardwaj and Srivastava, 1973) and the North Karanpura Coalfield (Kar, 1973). In the North Karanpura Coalfield radial monosaccates dominate the upper Karharbari assem-

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blage at the 481.8 m level. The Barakar Stage begins at 444 m and at 405.6 m the spore assemblage contains 38 per cent non-striate disaccates, 37 per cent striate disaccates and 17 per cent radial monosaccates. At 393 m, the mioflora shows a distinct change, containing 82 per cent striate disaccates, 6 per cent non-striate disaccates and 7 per cent radial monosaccates. This flora continues unchanged up to the 178.6 m level at which the Ironstone Shale Stage commences.

Palynologically, therefore, the Barakar Stage is divisible into two biozones. The lower one is dominated by a non-striate disaccate complex, prominently associated with striate disaccate spores, and the upper one is dominated exclusively by striate disaccates. Furthermore, in the lower Barakar facies, zonate triletes and varitriletes are often numerically over-represented, while in coals of the upper Barakar the apiculate triletes are significantly associated with the dominant striate disaccates.

Some of the important lower Barakar mio-spore genera are as follows: Hennelysporites, Microfoveolatispora, Microbaculispora, Lacinitriletes, Brevitriletes, Indotriradites, Dentatispora, Parasaccites, Barakarites, Potonieisporites, Sulcatisporites, Ibisporites, Rhizomaspora, Strotersporites, Striatitites, Hamiapollenites.

The Upper Barakar genera are: Cyclogranisporites, Lophotriletes, Horriditriletes, Apiculatisporis, Sulcatisporites, Ibisporites, Rhizomaspora, Strotersporites, Striatitites, Hamiapollenites.

The palynological limit between the Barakar and the Barren Measures is again not well understood. In the North Karanpura Basin, where successional study has been made by Kar (in press), the characteristic genera of the uppermost Barakar seem to persist in the lower Barren Measures, but the presence and rising incidence of the monosaccate genus Densipollenites seems to distinguish the Barren Measures. On the whole the Barren Measures differ palynologically from the Barakar in the abundance of saccate miospores and the negligible number of triletes.

RANIGANJ STAGE

This stage is typically developed in the Raniganj Coalfield and equally well developed in the Satpura area, where it is known as the Bijori Stage.
The Raniganj Stage has rich well-preserved megafossils, and so has received more attention from the palaeobotanists. Although lycopods are absent, the Raniganj flora is rich in other pteridophytic remains. Although no reproductive organs have so far been found, the equisetalean genera are now known in greater detail. The external morphology (Surange, 1955), epidermal structure and other anatomical characters of Phyllotheca are now known. It has been pointed out that although the genus is artificial, its Gondwana species may form a natural group which presents synthetic characters between Asterocalamitaceae and Calamitaceae (Pant and Kidwai, 1968). Another Articulate genus, Raniganjia bengalensis (Feistmantel) Rigby (Actinopteris indica of Feistmantel) has been described in detail. It shows close affinities with Sphenopsida, more particularly with Equisetales (Pant and Nautiyal, 1967a). The epidermis of Sphenophyllum speciosum (Royle) Zeiller has sinuous walled cells and apparently haplocheilic stomata (Pant and Mehra, 1963). Maheshwari (1968) is of the opinion that this southern form of Sphenophyllales should retain its original name, Trizygia speciosa Royle, until we know its fructification. Ferns and fern-like plants are represented by the genera Sphenopteris, Alethopteris, Pecopteris, Ptychocarpus and Merianopteris.

Glossopteris no doubt attained its maximum development in the Raniganj Stage. At least forty species are described on the external morphological and epidermal characters (Srivastava, 1957; Maheshwari, 1965a, 1966; Pant and Gupta, 1971). Surange and Srivastava (1957) have found that on the basis of epidermal characters, the species of Glossopteris, Gangamopteris and Palaeovitaria fall into six different groups, each probably representing a taxon of generic rank. All these genera are, therefore, form genera containing species of more than one genus.

A number of detached seeds and sporangia have been described in detail from the Raniganj Stage (Pant and Nautiyal, 1960, 1965). But recently a number of fructifications assignable to Glossopterideae have been discovered from the Damodar Valley basins and from some Raniganj exposures in Orissa (Maheshwari, 1965b; Surange and Maheshwari, 1970; Surange and Chandra, 1973a, b, c).

Glossotheca and Eretmonia are two types of male fructifications assignable to Glossopterideae (Surange and Maheshwari, 1970). Sporangia are borne on a branch system. In Glossotheca three to four pedicels spring from the abaxial side of the petiole of a fertile scale-leaf. Each pedicel bifurcates into two branches and then each daughter branch divides further by dichotomy four to five times until each slender ultimate branch bears one sporangium. Sporangia are held on either side of the petiole in closely placed clusters (Surange and Chandra, 1974c). In Eretmonia there is only one pedicel which bifurcates only once and the sporangia are borne in two clusters, one on either side of the petiolate fertile leaf.

Regarding the female reproductive organs, Scutum (Surange and Chandra, 1974a) has been interpreted as a female reproductive organ of Glossopteris, consisting of a bilaterally symmetrical, oval to round ovule-bearing receptacle, which is borne in the axil of a protective scale-leaf with Glossopteris type of venation. Both the ovule-bearing receptacle and its protective scale-leaf are carried on a common pedicel, which, in its turn, is attached to the petiole of a vegetative leaf of Glossopteris species. The receptacle is lens shaped and the marginal ovules are arranged closely in a row which gives a 'wing-like' rim to the oval to round receptacle. The scale-leaf must have opened widely or fallen off when the fructification matured, exposing the ovules to fertilisation by wind-borne winged pollen grains. Cistella (Maheshwari, 1965b; Surange and Chandra, 1974a) is more cylindrical and does not show a 'wing-like' rim. It is suggested that at least some of the Cistella type of fructifications might be the seedless receptacles of Scutum. Dictyopteridium (Surange and Chandra, 1973a; Maheshwari, 1965b) poses a similar type of organisation. It is a female reproductive organ consisting of a cylindrical receptacle bearing small, oval ovules on round cushions arranged in close spirals. This ovule-bearing receptacle is borne in the axil of a small
scale-leaf, which being protective, closely fits on one side of the fructification.

Certain cupulate organs have also been discovered attached to the fertile scale-leaves possessing *Glossopteris* type of venation. *Denkania* (Surange and Chandra, 1973b) is such a type in which a fertile scale-leaf with anastomosing venation bears five, six or more cupules on long, slender stalks which are attached to its petiole. Each cupule contains probably one large seed. *Partha* (Surange and Chandra, 1973c) is another female reproductive organ. The petiolate scale-leaf here has no midrib, but a few veins run straight in the middle and give out anastomosing secondary veins. Two to four pedicels spring from its petiole and each pedicel carries two to four cupules (or seeds ?) at its apical end. One winged seed, *Indocarpus*, might have been borne in some such way. *Lidgettonia mucronata* (Surange and Chandra, 1974b) is also another female reproductive organ. The petiolute scale-leaf here has no midrib but a few veins run straight in the middle and give out anastomosing secondary veins. Eight short pedicels, four on each side, are attached on the lower portion of the scale-leaf. Each pedicel carries one umbrella type cupulate disc, on the underside of which are probably attached small unwinged seeds. *Ottokaria* is also a seed bearing reproductive organ.

With so many reproductive organs now known, we shall soon have a better idea about the gymnospermous plants which constituted the dominant elements in the *Glossopteris* flora of the southern hemisphere.

A number of petrified woods of botanical interest are known from this stage. Maheshwari (1972) regrouped them recently under different genera and species. They are *Dadoxylon jamudhiense*, *Araucaryoxylon parbeliense*, *Araucaryoxylon ninghahense*, *Damudoxylon waltonii*, *Damudoxylon jamuriense*, *Megaporoxylon kraueselii*, *Kaokoxylon zaleskyi* and *Trigonomyelon raniganiense*. Many of these woods were formerly placed under various species of the form genus *Dadoxylon*. The woods most probably represent the trees which bore the leaves of *Glossopteris* species, as well as some other gymnospermous genera.

Some species of *Gangamopteris* have been reported from the Raniganj Stage. Maithy (1966) and Maheshwari (1966), who re-examined them, feel that they should be assigned to *Glossopteris*. *Gangamopteris* in all probability is absent in the Raniganj Stage.

Palynological assemblages from the Raniganj Stage have been studied mainly from the coalbearing strata of the Raniganj Coalfield (Bharadwaj, 1962; Bharadwaj and Salujha, 1964; Salujha, 1965; Bharadwaj and Tiwari, pers. comm.) as well as from bore cores from the North Karanpura Coalfield (Kar, 1969a, 1969b, 1973). The transition from the Barren Measures into the overlying Raniganj Stage, according to Bharadwaj (pers. comm.), is characterised by the sudden increase of apiculate triletes and monoletes from almost zero to over 10 per cent of the flora. Striate disaccates continue to be dominant throughout the Raniganj Stage. In the coal facies, the pollen assemblages contain a much higher percentage of triletes and monoletes.

Three biozones could be recognised in the Raniganj Formation (Bharadwaj, 1970-1, Srivastava, pers. comm.). These are characterised by the relative dominance of striate disaccates and trilete spores. The lowest zone (Kar, 1973) shows the prevalence of disaccates which is continued from the Barren Measures, but some new types also make their appearance, e.g. *Verticipollenites*, *Lahirites* and *Hindipollenites*. Trilete and monolete spores also increase considerably. Some of the genera, which add distinction to the lower Raniganj mioflora, are *Apiculatisporis*, *Lophotriletes*, *Cyclogranisporites* and *Laeviagatisporites*.

In the middle Raniganj, the triletes show further proliferation (42 per cent-50 per cent), and the non-striate disaccates increase in proportion to the striate-bisaccates.

The top part of the Raniganj Stage, studied by Kar (1970) and Bharadwaj and Tiwari (pers. comm.), is dominated by striate disaccates, whereas the cryptogamic spores are low in proportion. Kar (1970) described the mioflora from what may represent the youngest zone of the Raniganj. This horizon occurs just below the Panchet
Formation in the Raniganj Coalfield. In this zone trilete miospores notably decrease once again (19 per cent) while the striate disaccates attain the highest proportion (73 per cent).

**PANCHET STAGE**

The Raniganj Stage is overlain by the Panchet Stage of the Triassic age. The Panchet marks the decline of the *Glossopteris* flora and the appearance of the *Dicroidium* flora. This significant change in the vegetation is reflected in the mioflora also. The lithological change from the Raniganj Stage to the Panchet Stage is a conformable sequence and is accompanied by the gradual miofloral change from striate disaccate to non-striate disaccate domination.

In the Raniganj Coalfield the Panchet is characterised by the overwhelming presence of trilete spores, some of which are unknown in the Raniganj, e.g. *Dictyophyllidites, Divaripunctites, Decisporis* and *Rimaspora* (Kar, 1970). This assemblage has been referred to as the ‘*Decisporis* mioflora’ by Bharadwaj (1969). It is followed by an assemblage dominated by non-striate disaccates (*Alisporites* complex) which are believed to be of corystospermaceous affinity (*Dicroidium* etc.). Thus mega- and microfloral evidence clearly shows a major floral change at the close of the Raniganj Stage, which naturally sets the upper limit for the Lower Gondwana of India.

**ACKNOWLEDGMENTS**

I would like to express my thanks to Drs D. C. Bharadwaj and K. M. Lele for supplying information on the miofloras and Dr P. K. Maithy for his help in formulating the table on distribution of megafossils in the Indian Lower Gondwana.

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— and Makada, R., in press. Palaeobotanist in Madras.


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Triassic Floral Succession in the Gondwana of Peninsular India

M. K. ROY CHOWDHURY, M. V. A. SAstry, S. C. SHAH, GOPAL SINGH AND S. G. GHOSH

ABSTRACT

Four floral assemblages in the Triassic Gondwana of Peninsular India have been recognised. Flora I comes from the basal Panchet Formation of earliest Scythian age; Flora II, from the Upper Panchet Formation of early Scythian age; Flora III, from the Nidpur Beds of possibly late Scythian-early Anisian age; and Flora IV, from the Parsora Formation of late Carnian-Rhaetian age. Plant fossils from the late Anisian-early Carnian are very rare. The study indicates that there were no sweeping floral changes at the base of the Triassic. The floral succession in the Indian Gondwana supports a threefold classification and the three subdivisions are referred to as the Permian, Triassic and Jurassic-Early Cretaceous Gondwana sequence.

INTRODUCTION

Although the *Dicroidium* flora (of Triassic age) previously named as *Thinnfeldia* by Du Toit, is well developed in South Africa, South America, Australia and New Zealand, it is poorly known in India. Recently, only the Rhaetian flora was referred to as the *Dicroidium* assemblage in India.

Triassic sediments are well developed in both Extra-Peninsular and Peninsular India. In the former region, the sediments were deposited under marine conditions and no associated flora is known so far. The latter region has a non-marine sequence with no marine sediments within or immediately above or below it. The non-marine Triassic sequence is developed in practically all major Gondwana basins except the Rajmahal area. Here the Dubrajpur Formation was regarded as Triassic, but Sah and Shah (in press) concluded that it was partly Permian and partly Jurassic. The Triassic sequence is well developed in the Damodar Basin, Koel Valley, South Rewa, Satpura, Godavara Valley (including Pranhita-Godavari and Wardha Valleys), and Mahanadi Valley (Fig. 13.1). The successions in these areas (Fig. 13.2) have been based mainly on the information as given by Cotter (1917, 1938); Fox (1931); Gee (1932); Crookshank (1936); Sahni and Rao (1956); Pascoe (1959); Shah et al. (1971); Dutta and Ghosh (1971); Kutty and Roy Chowdhury (1970); Mitra (1972).

TRIASSIC FLORA

The Triassic flora is represented by macroflora and mioflora, tabulated in Tables 13.1 and 13.2 respectively. They have been compiled from various sources. The floral assemblage listed by Wadia (1926) under the Pachmarhi Sandstone of Triassic age comes from the Kamthi Formation of Late Permian age. Hence this flora is excluded here. The Triassic macroflora consists of thirty-five genera and fifty-four species and the mioflora of fifty-eight genera and seventy-two species. The macroflora is represented by pteridophytes including three genera of equisetales, one of lycopodiales,
Gondwana Flora

ten of filicales, and six of pteridosperms; and gymnosperms including four genera of cycadales, two of cordaitales, five of coniferales and two of ginkgoales. Two seed genera are also known. The mioflora is represented by sporites comprising seven genera of triletes, eleven of apiculati, two of aletes, three of monoletes, and one of ornati; and pollenites comprising three genera of saccites, twenty of disaccites, two of costati, one of polysaccites, one of praeolpites and four of monocolpatae. Three megaspore genera are also known.

FLORAL SUCCESSION

No single basin has a complete sequence on which the floral succession can be based. This difficulty is further increased by the paucity of the associated marine fossils. Other associated fossils like vertebrates and estheriids have been relied upon. Under these circumstances, the correlation of these fossiliferous zones lacks precision with reference to the Standard Scale.

Four different floral assemblages including macro- and mioflora can be differentiated in the Peninsular Triassic Gondwana sequence. Of these, one unit is entirely based on macroflora and another on mioflora. The other two are based on both macro- and mioflora.

_Flora I_

This floral assemblage comes from the basal part of the Panchet Formation and lies below the _Lystrosaurus_ Zone in the Raniganj Coalfield of the Damodar Valley. The basal part of the Panchet Formation has a gradational contact with the underlying Late Permian Raniganj Formation. The macrofloral assemblage is characterised by the abundance of _Glossopteris_ (holdovers from the Permian) along with _Cyclopteris pachyrachis_, doubtful _Dicroidium_ and _Podozamites_ (Satsangi, in press).-The incoming of

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Fig. 13.1. Triassic Gondwana occurrences in Peninsular India

Fig. 13.2. Triassic Gondwana succession in India
<table>
<thead>
<tr>
<th>GONDWANA BASINS</th>
<th>DAMODAR COALFIELD</th>
<th>KOEL COALFIELD</th>
<th>REWA COALFIELD</th>
<th>SATPURA GODAVARI VALLEY</th>
<th>MAHANADI VALLEY</th>
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<td>RIDIPUR</td>
<td>PRANHITA-GODAVARI</td>
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INDEX
- Shales / Red Clays
- Reworked Shale / Clay
- Sandstones
- Limestone bands
- Plant Fossils
- Esthers
- Vertebrate Fossils
- Unionids

Triassic Floral Succession in Peninsular India
# Table 13.1. Distribution of Important Macrofloral Elements from Triassic Gondwana of India

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Notes: A—Lower Panchet Formation of Raniganj Coalfield; B—Panchet Formation of Auranga Coalfield; C—Mahadeva Group of Auranga Coalfield; D—Panchet Formation of Ramkola-Tatapani Coalfield; E—Nidpur beds of South Rewa; F—Daigaon Formation of South Rewa; G—Tiki Formation of South Rewa; H—Parsora Formation of South Rewa.

The new elements is clearly noticeable and significant. The miofloral assemblage is also similar to that of the Raniganj Formation. The assemblage is dominated by disaccate genera, but quantitatively and qualitatively it is reduced from more than sixty genera in the Raniganj Formation, to only thirteen genera in the basal Panchet (Satsangi et al., 1972; Gopal Singh and Shah, 1972).

The associated estheriids give a Triassic aspect to this zone (Tasch et al., this volume) and since it underlies the early Scythian Lystrosaurus fauna, the basal Panchet with its flora can safely be considered as of earliest Scythian age. Flora I is therefore the oldest unit in the Triassic of the Indian Gondwana sequence.

Similar floral assemblages are known from the Ramkola-Tatapani Coalfield (Feistmantel, 1882, 1886), Auranga Coalfield (Banerjee, in press), and Pali-Daigaon beds (Saksena, 1952, 1961; Lele, 1969). Probably the miospore assemblage described by Trivedi and Misra (1969) from Nidpur belongs to this zone (see discussion under Flora III also).

**Flora II**

This comes from the Upper Panchet Formation of the Damodar Basin. No macroflora is known; the mioflora comes only from the borehole samples (Shrivastava and Pawde, 1962; Kar, 1970a, 1970b; Sarbadhikari, 1972). It is characterised by the abundance of trilete spores, and appearance of several new Mesozoic miospores. Though not associated with the samples, vertebrate fossils known from the Upper Panchet indicate that it is not younger than early Scythian. This assemblage is not known from any other basin in India.

**Flora III**

This floral assemblage at Nidpur, first discovered by Satsangi (1964), comes from the South Rewa Basin. Its stratigraphic position in relation to the first two floral zones is controversial. It directly overlies the Raniganj Formation with a vertebrate fauna of Late Permian age collected from a locality ten miles from the plant locality, and it is overlain by the unfossiliferous Mahadeva. Two miofloral assemblages have been described from Nidpur and both have been assigned to the Triassic. The assemblage described by Trivedi and Misra (1969) is characterised by dominance of striated disaccate over other disaccate grains. Striated grains are profuse in the mioflora from the Raniganj Formation. It may be noted here
Table 13.2. Distribution of Microflora from Triassic Gondwana of India

<table>
<thead>
<tr>
<th>Name of microspore</th>
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<td>Talchirella spp.</td>
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<td>Duoiporites sp.</td>
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Note: A—Lower Panchet Formation of Raniganj Coalfield; B—Upper Panchet Formation of Raniganj Coalfield; C—Nidpur beds of Madhya Beds (Trivedi and Misra, 1969); D—Nidpur beds of Madhya Pradesh (Bharadwaj and Srivastava, 1969).
that the Permian mioflora corresponding to
the Raniganj Formation was recorded from
this area (Maheshwari, 1967) though it may
represent the older unit, Barren Measures
(Bharadwaj, 1970). In another Triassic mio-
floral assemblage, first recorded by Chandra
and Satsangi (1965) and described in detail
by Bharadwaj and Srivastava (1969), non-
striated saccates are dominant. It is not clear
whether these two miofloras are different
expressions of the same assemblage or from
different stratigraphic levels of a condensed
sequence. The macroflora was not studied
by Trivedi and Misra (1969) but they made
the statement (p. 14): ‘The shales revealed
the presence of Glossopteris, Schizoneura,
Vertebraria and equisetaceous stems, mostly
as impressions. Some of these have, in addi-
tion, impressions of Dicroidium as reported
by Satsangi.’ But the macroflora from the
collection which Bharadwaj and Srivastava
(1969) have studied for mioflora, consists
of three new species of Dicroidium, Glossop-
teris, and other forms. Bharadwaj and
Srivastava (1969) and Bharadwaj (1970)
have compared the mioflora they had des-
cribed with that from the Raniganj Forma-
tion and Upper Panchet (which actually they
referred to as Lower Panchet) as well as
other Triassic mioflora, and concluded that
it is younger than Upper Panchet (i.e.
Lower Panchet in their sense). The mioflora
of the Raniganj Formation has been quite
extensively studied, and similar floral assem-
blages to those described by Bharadwaj and
Srivastava are not known from a single
locality. Hence the possibility that the two
miofloras described by Trivedi and Misra,
and Bharadwaj and Srivastava are different
expressions of the same is ruled out. The
former mioflora may be either from the beds
equivalent to the Raniganj Formation or
possibly to some part of the Panchet For-
mation. The mioflora described by Bharad-
waj and Srivastava is taken as characteristic
of the Nidpur assemblage because in this
zone the Dicroidium had been firmly estab-
lished. Nonstriated disaccates, which consti-
tute more than 50 per cent in the assem-
blage at Nidpur (Bharadwaj, 1970), are
presumed to be derived from Dicroidium
(Anderson and Anderson, 1970). Glossop-
teris does continue in this floral unit. It
is not considered older than Flora I because
the Flora I comes from the basal Panchet
which has a gradational contact with the
Raniganj Formation of Late Permian age.
The absence of nonstriated disaccates in
Flora II gives a younger aspect to Flora
III which has more than 50 per cent of
these elements. Hence, the Flora III is
younger than early Scythian and may vary
in age between late Scythian and early
Carnian. Most likely it is of late Scythian
age. This flora is not known from any other
basin in India.

**Flora IV**

This comes from the South Rewa Basin
and occurs in the upper part of the Tiki
Formation and the Parsora Formation, al-
though the floral assemblage from the former
unit is poorly known. On the basis of the
vertebrate fauna which occurs mostly in the
upper part of the Tiki Formation, the floral
assemblage coming from this unit is dated
as late Carnian-Rhaetian age. Dicroidium
continues to occur in Fauna IV but in addi-
tion there are elements like Araucarites,
Pterophyllum and Marattiopsis which are
the forerunners of the Jurassic floras (Lele,
1969).

A similar flora is also recorded from the
beds near Acheli and Ganaram (Rao and
Shah, 1960; and their unpublished reports)
of the Pranhita-Godavari Basin. These beds
are probably equivalent to the Dharmaram
Formation.

These four floral zones do not cover the
entire Triassic period as practically nothing
is known from the late Anisian-early Carn-
ian.

**PERMIAN-TRIASSIC BOUNDARY IN PENINSULAR
INDIA**

In the Peninsular Gondwana, the Late
Permian and Early Triassic are represented
by non-marine rocks and the floral changes
in these beds are important in recognising
the boundary. The best areas for boundary
study are those where there was continuity
of sedimentation. The Raniganj Coalfield of
the Damodar Valley is the best area available.
The Raniganj Formation has an abundance
of Glossopteris and its invertebrate faunas
indicate that it is of Tatarian age. Though there are no significant vertebrates in the formation in the Damodar Basin, it does contain Tatarian vertebrates in other basins. This formation has a gradational contact with the overlying Panchet Formation of earliest Scythian age (as discussed earlier). The floral changes observed in this basin together with those of the South Rewa and Koel Valley can be summarised as follows.

There is not a complete extinction of the rich Permian flora at the boundary between the Raniganj and Panchet Formations, nor is there evidence of any catastrophic events at or near this stratigraphic level (Mitra, 1972). Some Permian floral elements lingered on into the Early Triassic and a few new elements appeared, resulting in a limited macroflora. The incoming of fewer elements is significant in comparison with the dominance of the earlier floral elements. The floral changes were gradual but not as sweeping as in the Salt Range, Pakistan (Balme, 1969). The elements that survived as 'holdovers' could not flourish and were reduced in size, as is well illustrated by species of *Glossopteris*. The climatic changes, though gradual, caused a severe decline in development of *Glossopteris*, but the genus lingered on into Flora IV (of late Carnian-Rhaetian age).

*Dicroidium* along with other forms appeared for the first time at the base of the Triassic. The first *Dicroidium* was of smaller size than those found higher up in the sequence.

**CLASSIFICATION OF INDIAN GONDWANA SEQUENCE**

Feistmantel (1876: 78-9) suggested that the Gondwana may be divided into two: Lower and Upper. The Lower Gondwana included the Permian and Early Triassic. Followers of this scheme laid stress on the preponderance of the Permian elements in the Early Triassic sequence and suggested that the *Glossopteris* flora was succeeded subsequently by the *Ptilophyllum* flora. The sedimentological break above the Lower Triassic (Panchet) and Mahadeva Group gave additional support to this view (Mitra, 1972).

Feistmantel (1880) also initiated a three-fold scheme when studying the Triassic plants from the South Rewa basins collected by Hughes (1884). He found an admixture of *Glossopteris* and *Ptilophyllum* floral elements which he grouped under the 'Transitional Beds' and suggested that it might eventually be classed as 'Middle Gondwana'. The presence of red-beds and the Triassic reptiles in the Panchet gave support to the threefold scheme (Vredenburg, 1910; Wadia, 1926). Floral aspects were further studied and the Middle Gondwana was supported by Sak-sena (1952, in press) and Lele (1969). The recent analysis of the flora taken together with the study by Shah et al. (1971) clearly indicates that the Triassic Gondwana has a distinctive floral composition separable from the older *Glossopteris* flora and the younger *Ptilophyllum* flora. Because the Lower, Middle and Upper Gondwana correspond to the Permian, Triassic and Jurassic-Lower Cretaceous of the Standard Scale, the usage of the terms Lower, Middle and Upper should be dropped. These should be referred to as the Permian, Triassic and Jurassic-Lower Cretaceous Gondwana sequence.

**REFERENCES**


—, in press. The Gondwana classification. Autumn School of Palaeobotany, KodaiKanal, India.
Miospore Zones in the Lower Mesozoic of Southeastern Queensland

N. J. DE JERSEY

ABSTRACT

Within 100 km of Brisbane, continental sediments of Middle Triassic to Middle Jurassic age have yielded abundant spores and pollen. On the basis of detailed study of the miospore assemblages, three zones and one subzone are now formally proposed within this succession. Older sediments from the Esk Trough contain assemblages here referred to the Duplexisporites problematicus Microflora. A formal biostratigraphic unit is not proposed because of incomplete knowledge of the lateral extent of this microflora.

Species that are stratigraphically significant for the recognition of these units are figured, and their distribution is illustrated diagrammatically in relation to sedimentary groups and formations in which they occur. Evidence on the age of the zones, based on comparison of the assemblages with other, accurately dated microfloras, is discussed. This discussion is particularly relevant to location of the Triassic-Jurassic boundary within the sequence.

Recognition of this sequence of miospore zones in Queensland will provide a basis for comparison with other Gondwana microfloras of similar age. Evidence already available from South America, Antarctica and New Zealand is suggestive of parallel development of miospore sequences in other parts of Gondwanaland.

INTRODUCTION

Mesozoic sediments in southeastern Queensland are known to range in age from Middle Triassic (the Esk Formation) to Middle Jurassic (the Walloon Coal Measures). As this succession is developed within 100 km of Brisbane, the stratigraphic relationships of the constituent formations have been studied for many years and are now known in some detail. The writer has pointed out (de Jersey, 1971a: 24) that, because of the presence of lower Rhaeto-Liassic sediments not known from other parts of Queensland, this succession provides the most comprehensive record of Upper Triassic and Lower Jurassic strata in the state. It can, accordingly, be regarded as a reference section for correlation of sediments of this age within eastern Australia.

The sediments in this succession are regarded as very largely, or entirely, of continental origin. Although acritarchs have been recorded from two horizons in the Esk Formation (de Jersey, 1972b: 16) and from a thin oolitic section in the Marburg Formation (de Jersey, 1971a: 22), no other fossils indicative of a marine environment have been found. It can only be suggested, as indicated by Reiser and Williams (1969: 21), that these records of acritarchs denote changes in environment, but do not definitely establish the occurrence of marine incursions.

All the formations from the Esk Formation upwards have yielded spores and pollen. The abundance and diversity of miospore
species have rendered them more useful than plant megafossils, in biostratigraphic studies in this sequence. Other fossils, such as pelecypods, insects, and vertebrates, are too sparsely distributed to be of much significance for establishing correlation. Previous investigations (de Jersey, 1970a, 1971a-d, 1972a-d, 1973a-b) have provided a record of miospore distribution in various parts of this succession. The purpose of this paper is to utilise this knowledge to propose a formal zonation, which will also provide an opportunity to assess evidence on the ages of the assemblages, and to discuss the location of the Triassic-Jurassic boundary within the sequence. Finally, comparison of the Queensland microfloras with those of similar age from other Gondwana areas, furnishes evidence bearing on the Gondwanaland concept.

STRATIGRAPHIC BACKGROUND

Swindon (1971) has published a general review of the Mesozoic stratigraphy of the Moreton District, covering approximately the area considered in this paper. Reference should be made to this review, modified by recent revision of the stratigraphic nomenclature (Cranfield and Schwarzbock, 1972) and re-interpretation of the age of the Esk Formation (de Jersey, 1972b) to provide a summary of current stratigraphic knowledge of the area. The principal modifications to the nomenclature proposed by Cranfield and Schwarzbock (1972) are the institution of formal nomenclature for the Triassic units of the Esk Trough, and the abandonment of the term 'Flelidon Sandstone' which they consider has been used in various senses by different authors, and so has become a source of confusion. They have proposed a new term 'Woogaroo Sub-Group' for the lower part of the Bundamba Group (equivalent to 'Helidon' in the broad sense). In re-interpreting the age of the Esk Formation, the writer (de Jersey, 1972b: 18) has assigned it to the Middle Triassic, regarding it as significantly older than the Ipswich Coal Measures, with which it was previously correlated by some authors (e.g. Swindon, 1971: 109, 117).

MIOSPORE SEQUENCE AND ZONES

A general account of the zonal concept, particularly as applied in Australia to zonation by miospore species, is given by Dettmann and Playford (1969: 180-2) in introducing their zonation of the Australian Cretaceous. Their interpretation is accepted and followed in this paper, so that 'zone' is used in the sense of Teichert (1958: 108) as 'a bed or group of beds which contains a defined assemblage of fossils, one of which is selected for the naming of the zone'; the nominate fossil 'may also be a member of earlier or later fossil assemblages, and it may in places be absent from the typical assemblage which characterises the zone'. Like Dettmann and Playford, the writer uses 'subzone' as a 'subdivision of zone' and 'microflora' to denote a biostratigraphic category whose spore-pollen contents are inadequately known.

The succeeding discussion on zonation in southeastern Queensland is divided into two parts: (a) selection of species for use in zonation, and (b) definition and distribution of the miospore zones. In the first part the reasons for the selection of species used in zonation are indicated. This section also serves as a summary of the most significant changes in the microfloral succession. In the second part the zones are formally defined and described, data on reference sections, distribution and age being included.

Selection of Species for Use in Zonation

One of the most striking features of the microfloral sequence in the succession studied is the abundance of *Classopollis* in the upper Woogaroo Sub-Group and lower Marburg Formation. At first sight it would appear that the incoming of this genus could be used to separate this part of the succession from the underlying Raceview Formation, on the basis of its distribution recorded in light-microscope studies (de Jersey, 1971a: fig. 3). However, recent studies with the scanning electron microscope have led to a re-interpretation of the morphology and circumscriptio of the genus and species assigned

Fig. 14.1. Moreton District regional geology (based on Swindon, 1971: fig. 1)
Illustrating general relationship between biostratigraphic units defined in the Surat Basin (Reiser and Williams, 1969.) and those proposed for the Moreton Basin.

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**Fig. 14.2.** General relationship between biostratigraphic units defined in the Surat Basin (Reiser and Williams, 1969) and those proposed for the Moreton Basin.

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**Fig. 14.3.** Vertical ranges of selected species in relation to proposed zones, composite lithologic sequence and ages (based on miospore assemblages).
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most persistent of the three species, particularly in the lower part of its range. Accordingly the first appearance of this species has been used to define the *Trisaccites variabilis* Zone, formally proposed in the succeeding section. The top of this zone is readily defined by the introduction of *Zonalapollenites*, which constitutes a persistent marker in both the Surat and Moreton Basins (Reiser and Williams, 1969: 19-21; de Jersey, 1971a: 22, 23, 1971d: 471, 1973a).

Below the *Trisaccites variabilis* Zone, *Classopollis* (the species *C. meyeriana*) persists to the base of the Bundamba Group, but is relatively rare and sporadically distributed in the lower part of its range in the Aberdare Conglomerate and Raceview Formation. The species *Polycingulatisporites crenulatus* is proposed as a zonal index, as it is relatively persistent in the lower Bundamba Group (de Jersey, 1970a: 19, 20) and may be used in distinguishing this part of the succession from the underlying Ipswich Coal Measures. Consequently the base of the *P. crenulatus* Zone (formally proposed below) is defined by the first appearance of this species. Within this zone, the restricted range of *Ceratosporites helidonensis* enables a subzone to be recognised on the basis of the presence of that species.

In the older (Triassic) part of the succession, strata assigned to the *P. crenulatus* Zone are underlain disconformably by the Ipswich Coal Measures. The microflora of the latter is relatively uniform and it was noted (de Jersey, 1971b: 18) 'The available evidence does, in fact, favour the view that only one biostratigraphic unit, based on microspore assemblages, can be recognised for the whole sequence of the Ipswich Coal Measures'.

In defining a biostratigraphic unit for the Ipswich sequence, the species selected should be one of those listed (de Jersey, 1971b: 19), which are not known from older microfloras. A feature particularly characteristic of the Ipswich assemblages, in contrast to those from older sediments, is the widespread distribution of spores with distal circumpolar muri (*Annulispora* spp. and *Polycingulatisporites densatus*). However, such species are excluded from consideration as zonal indices, because rare specimens assigned to, or compared with, *Annulispora* have been recorded from the Middle Triassic of the Wianamatta Group of the Sydney Basin (Helby, in press), and from the Middle Triassic of the Moolayember Formation (de Jersey and Hamilton, 1967: 11). The species selected to define a zone encompassing the Ipswich Coal Measures is *Craterisporites rotundus* de Jersey, 1970. It is a relatively persistent component of the Ipswich microflora from the basal sediments of the Mount Crosby Formation upwards; it is also persistent laterally, as indicated by its occurrence in the Tarong Beds of similar age (de Jersey, 1970c).

The oldest Mesozoic assemblages recorded from southeastern Queensland are those of the Esk Formation (de Jersey, 1972b). They lack *Craterisporites rotundus* and *Annulispora* spp., which characterise the Ipswich microflora. Like assemblages from the Moolayember Formation, with which this Esk sequence has been correlated, they contain *Duplexisporites problematicus*, the presence of which distinguishes them from older microfloras. As the material studied came from three drillholes, all in the southern part

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*Pl. 14.I. All specimens are re-figured from de Jersey (1970a, 1971b, 1972b) where slide numbers, stage co-ordinates and sample localities are recorded. All photographs approximately x 1000.*

*Fig. 1. Annulispora microannulata de Jersey, 1962 (median focus). Craterisporites rotundus-Trisaccites variabilis Zones. Fig. 2. Annulispora folliculosa (Rogalska) de Jersey, 1955 (distal focus). Craterisporites rotundus-Trisaccites variabilis Zones. Fig. 3. Polycingulatisporites densatus (de Jersey) Playford & Dettmann, 1965 (median focus). Craterisporites rotundus-Trisaccites variabilis Zones. Fig. 4. Stereisporites perforatus Leschik, 1955 (distal focus). Craterisporites rotundus-Trisaccites variabilis Zones. Fig. 5. Duplexisporites problematicus (Couper) Playford & Dettmann, 1965 (equatorial focus). Duplexisporites problematicus Microflora, Craterisporites rotundus-Trisaccites variabilis Zones. Fig. 6. Equisetosporites stereos (Jansoniuss) de Jersey, 1968 (equatorial focus). Duplexisporites problematicus Microflora. Fig. 7. Acanthotriletes bradiensis Playford, 1965 (equatorial focus). Craterisporites rotundus Zone. Fig. 8. Accinitesporites ligatus Leschik, 1955 (equatorial focus). Duplexisporites problematicus Microflora. Fig. 9. Voltziaceasporites heteromorpha Klaus, 1964 (median focus). Duplexisporites problematicus Microflora.*
of the Esk Trough, which extends over a length of at least 250 km, evidence on the lateral persistence of the assemblages within the Esk Formation is incomplete. Until more such evidence is available these assemblages are assigned informally to the *Duplexisporites problematicus* Microflora.

**Definition and distribution of the miospore zones**

In ascending stratigraphic order, the biostratigraphic units recognised in southeastern Queensland are as follows:

*Duplexisporites problematicus* Microflora

Craterisporites rotundus Zone

Polycingulatisporites crenulatus Zone (including the Ceratosporites helidonensis Subzone)

**Trisaccites variabilis Zone**

These units are defined below, in terms of diagnostic species; species characteristic of each are also recorded and their areal distribution is discussed. Ranges of these species are indicated in Figure 14.3, which also illustrates the relationship of these biostratigraphic units to lithologic units. Correlation with biostratigraphic units in the Surat Basin is indicated in Figure 14.2. Most of the characteristic species listed are figured in Plates 14.I-III.

The *Duplexisporites problematicus* Microflora

This microflora is characterised by the presence of *D. problematicus* and absence of *Craterisporites rotundus*. The latter appears at the base of the *C. rotundus* Zone. Associated species, not present in the latter, include *Accinctisporites ligatus* Leschik, 1955, *Equisetosporites stevesi* (Jansonius) de Jersey, 1968, *Guttatisporites viesscheri* de Jersey, 1968, and *Voltziaceaesporites heteromorpha* Klaus, 1964.

**Reference Section:** G.S.Q. Ipswich 17, 248-880 m (Esk Formation).

**Distribution:** The microflora is also present in the Esk Formation of P. S. Baylam I and P. S. Lockrose I (de Jersey, 1972c: 17). It also occurs in the Moolayember Formation of the Bowen Basin and the Triassic of the Abercorn Trough (de Jersey, 1972d: 385).

**Age:** Middle Triassic (see de Jersey, 1972c: 18).

The *Craterisporites rotundus* Zone


**Reference Section:** N.S. 272 (Ipswich Coalfield), 563.1 m-732.1 m (Blackstone Formation, de Jersey, 1970a). The zone

Pl. 14.II. All species are re-figured from de Jersey (1970a, 1971a, 1971b), where slide numbers, stage co-ordinates and sample localities are recorded. All photographs approximately x 1000. Fig. 1 *Anapiculatisporites dauesonensis* Reiser & Williams, 1969 (distal-equatorial focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 2. *Anulisporites variaramulatus* (Levet-Carette) Reiser & Williams, 1969 (distal focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 3. *Ceratosporites helidonensis* de Jersey, 1971 (distal-equatorial focus). *Ceratosporites helidonensis* Subzone (*Polycingulatisporites crenulatus* Zone). Fig. 4. *Clas­sopolis meyeriana* (Klaus) de Jersey, 1973 (proximal focus). *Polycingulatisporites crenulatus-Trisaccites vari­abilis* Zones. Fig. 5. *Lycopodiumsporites rosewoodensis* (de Jersey) de Jersey, 1963 (distal focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 6. *Polycingulatisporites mooniensis* de Jersey & Paten, 1964 (median focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 7. *Crater­isporites rotundus* de Jersey, 1970 (median focus). *Craterisporites rotundus* (Polycingulatisporites crenulatus Zone). Fig. 8. *Polycingulatisporites crenulatus* Play­ford & Dettmann, 1965 (median focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 9. *Semiretisporis denmeadi* (de Jersey) de Jersey, 1970 (distal focus). *Craterisporites rotundus* Zone. Fig. 10. *Aratrisporites flexibilis* Playford & Dettmann, 1965 (equatorial focus). *Craterisporites rotundus* Zone.
comprises the entire section of the Ipswich Coal Measures; no single drillhole covers the complete section. The section penetrated in N.S. 272 forms the upper part of the zone, and its junction with the overlying *Polycingulatisporites crenulatus* Zone. Lower portions of the zone are present in N.S. 566 (Ipswich) and G.S.Q. Ipswich 16 (de Jersey, 1971b).

Distribution: In southeastern Queensland, this zone is represented in the type area of the Ipswich Coal Measures and extends into the Brisbane area (de Jersey, 1972d: 457), and to Stradbroke Island (de Jersey, 1971c). To the west, it comprises the Tarong Beds (de Jersey, 1970b). It is probably also represented in the Brady Formation and ‘Feldspathic Sandstone’ of Tasmania (Playford, 1965), and the older (Triassic) portion of the Leigh Creek Coal Measures of South Australia (Playford and Dettmann, 1965).

Age: Carnian (Late Triassic) (see de Jersey, 1971b: 20-2).

The *Polycingulatisporites crenulatus* Zone

This zone is defined by the presence of *P. crenulatus* and the absence of *Trisaccites variabilis* (Dev) Haskell, 1968. Other species that have ranges commencing within the zone include *Anapiculatisporites dawsonensis* Reiser and Williams, 1969, *Antulsporites variigranulatus* (Levet-Carette) Reiser and Williams, 1969, *Ceratosporites helidonensis* de Jersey, 1971, *Classopollis meyeriana* (Klaus) de Jersey, 1973, *Classopollis* sp. cf. *C. chateaunovi* Reyre, 1970. *Indusiisporites parvisaccatus* (de Jersey) de Jersey, 1963, *Lycopodiumsporites austroclavatidites* (Cookson) Potonie, 1956 (distal focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 1. *Lycopodiumsporites austroclavatidites* (Cookson) Potonie, 1956 (distal focus). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 2. *Classopollis* sp. cf. *C. chateaunovi* Reyre, 1970 (equatorial view). *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 3. *Classopollis* sp. cf. *C. chateaunovi* Reyre, 1970 (equatorial view). Slide S3095, stage co-ords. 21.0, 83.0; co-ords of lower right hand corner of cover glass 26.3, 80.2; G.S.Q. Ipswich 18, 381.5 m. *Polycingulatisporites crenulatus-Trisaccites variabilis* Zones. Fig. 4. *Trisaccites variabilis* (Dev) Haskell, 1968 (equatorial focus). Slide S4241, stage co-ords 10.6, 103.7; co-ords of lower right hand corner of cover glass 26.8, 85.4; G.S.Q. Ipswich 14, 111.9 m. *Trisaccites variabilis* Zones. Fig. 5. *Trisaccites variabilis* (Dev) Haskell, 1968 (equatorial focus). Slide S4040, stage co-ords 13.3, 88.8; co-ords of lower right hand corner of cover glass 27.7, 83.5; G.S.Q. Ipswich 19-22R, 429.7 m. *Trisaccites variabilis* Zones. Fig. 6. *Zonalapollenites damptieri* Balme, 1957 (equatorial focus). (Above *Trisaccites variabilis* Zones). Fig. 7. *Zonalapollenites segmentatus* Balme, 1957 (median focus). (Above *Trisaccites variabilis* Zones). Fig. 8. *Ischyosporites marburgensis* de Jersey, 1963 (equatorial focus). *Trisaccites variabilis* Zones. Fig. 9. *Nevesissporites vallatus* de Jersey & Paten, 1964 (proximal focus). *Trisaccites variabilis* Zone.

Reference Sections: For the *P. crenulatus* Zone, G.S.Q. Ipswich 18 (Walloon), 764.9 m-1154.6 m (Woogaroo Sub-Group and lower Marburg Formation). For the *C. helidonensis* Subzone, G.S.Q. Ipswich 18 (Walloon), 937.9 m-1087.6 m (Woogaroo Sub-Group). The stratigraphy and palynology of these sections are described in de Jersey (1971a, 1971d).

Distribution: Available evidence indicates that the base of this zone coincides with the base of the Bundamba Group (the base of the Aberdare Conglomerate, or where that unit is not recognised, the base of the Raceview Formation). Its upper limit is within the upper Woogaroo Sub-Group or lower Marburg Formation, depending on local facies (de Jersey, 1971d, 1973a).

This zone is widespread in the Moreton Basin. Elsewhere in Queensland it occurs in the lower Precipice Sandstone of the Surat Basin (Reiser and Williams, 1969).
the Razorback Beds (Playford and Cornelius, 1967), and the Mulgildie Basin (de Jersey, 1972b). It is probably represented in part of the upper section of the Leigh Creek Coal Measures of South Australia, described by Playford and Dettmann (1965).

Age: From comparison with other microfloras, assemblages from this zone were regarded as Rhaeto-Liassic (de Jersey, 1971a: 24). Evidence on the position of the Triassic-Jurassic boundary, within the zone, is discussed in the succeeding section of this paper.

The *Trisaccites variabilis* Zone

This zone is defined by the presence of *Trisaccites variabilis* (Dev) Haskell, 1968, and the absence of *Zonalapollenites segmentatus* Balme, 1957. Species which first appear within the zone include *Ischyosporites marburgensis* de Jersey, 1963 and *Nevesisporites vallatus* de Jersey and Paten, 1964. Most of the assemblages are dominated by the genus *Classopollis*.

Reference Section: G.S.Q. Ipswich 18 (Walloon), 398.6 m-746.4 m (Marburg Formation). Stratigraphic and palynological details of the section have been recorded previously (de Jersey, 1971d).

Distribution: This zone is widely distributed in the Moreton Basin. Its upper limit is within the Marburg Formation; its base is either within the Marburg Formation or the upper Woogaroo Sub-Group, depending on local facies (de Jersey, 1971d, 1973a). On the basis of Surat Basin data recorded by Reiser and Williams (1969) for the first appearances of *N. vallatus* and their 'Podosporites sp.' (≡ *T. variabilis*), the *T. variabilis* Zone can be equated approximately with their *N. vallatus* Sub-zone. (As indicated above, the available evidence indicates that *N. vallatus* is relatively rare and sporadically distributed in the lower part of its range in the Moreton Basin.) Elsewhere in Australia, the zone is represented in the Leigh Creek area of South Australia, as indicated by the presence of *I. marburgensis* in the uppermost assemblage recorded by Playford and Dettmann (1965: 162).

Age: Evidence for the Liassic age of the section in the Lowood area, assigned to this zone, has been quoted previously (de Jersey, 1971a: 24). More recent data on its age, within the Liassic, are discussed below.

EVIDENCE FROM MIOSPHERE SEQUENCES ON LOCATION OF THE TRIASSIC-JURASSIC BOUNDARY

In Australia, the oldest Jurassic microfloras dated by associated marine faunas are those from the Bajocian of Western Australia, described by Balme (1957). Determination of the age of earlier assemblages from sediments largely, if not entirely, of non-marine origin and lacking other diagnostic fossils, must accordingly be based on comparison with Jurassic microfloras from other parts of the world.

In such comparison, an obvious datum horizon is provided by the first appearance of *Zonalapollenites dampieri*, as previously indicated (de Jersey, 1971a: 24). In Europe, Schulz (1967: 593) recorded the initial appearance of this species in the late Toarcian (i.e., latest Liassic, accepting his assignment of the succeeding Aalenian to the Dogger). In the southern hemisphere, Goubin (1965) recorded the species from the upper Liassic onwards in Madagascar, and Volkheimer (1972) has recently found that it makes its first appearance in the upper Toarcian in a sequence in Argentina, dated by ammonite faunas. From this evidence for the introduction of *Z. dampieri* at about the same time in both hemispheres, the top of the *Trisaccites variabilis* Zone is regarded as late Toarcian in age. This indicates assignment of most, if not all, of the Marburg Formation to the Toarcian, although Volkheimer (1972: 353) has suggested that the top of this formation may extend into the Aalenian.

Lower in the Liassic, the same author has found evidence for dating the first appearance of *Nevesisporites vallatus* in the Argentine sequence as early Pliensbachian or late Sinemurian. Correlation with this sequence would thus indicate an early Pliensbachian or late Sinemurian age for the base of the *Trisaccites variabilis* Zone.

Microfloras equivalent in age to the basal part of the Moreton Basin succession (the
lower part of the *P. crenulatus* Zone) are absent from the sequence in Argentina. However, some recent evidence on the age of this zone is forthcoming from comparison with microfloras from New Zealand, where Triassic and Jurassic sediments are dated by marine faunas. Investigation of the miospore content of the latter is in its preliminary stages, but Dickson (1972) has recently recorded a Rhaetian assemblage, which contains *P. mooniensis* but lacks the abundance of *Classopollis* which is characteristic of the upper part of the *P. crenulatus* Zone. In these respects it is similar to assemblages from the Raceview Formation and the type section of the Ripley Road Sandstone, which were previously assigned to the Late Triassic (de Jersey, 1971a: 24, 25). From comparison with this New Zealand assemblage, and the earliest assemblage in the Argentine sequence, it is suggested that the Triassic-Jurassic boundary lies within the *P. crenulatus* Zone, probably in the section immediately overlying the Raceview Formation.

The recent evidence quoted has reduced uncertainty concerning this boundary; further records of accurately dated microfloras from New Zealand and South America could lead to refinements in dating the Queensland sequence.

**MICROFLORAL EVIDENCE BEARING ON THE GONDWANA CONCEPT**

These Queensland microfloras show greater similarity to other Gondwana microfloras than to those of northern (Laurasian) regions. Volkheimer (1972: 355) has recently commented on the close relationships between Liassic miospore sequences from Queensland and Argentina. The similarity is also evident in the Middle Triassic, where a notable example is *Cadargasporites senectus* de Jersey and Hamilton, 1967, recorded from the lower Moolayember Formation of Queensland (de Jersey and Hamilton, 1967), the Wianamatta Group of New South Wales, and the Lashly Formation of Antarctica (Helby and McElroy, 1969) and the Choiyoi Formation in the southern Mendoza Province of Argentina (R. Herbst, pers. comm.). This distinctive species is unknown in Laurasian assemblages of similar age. In the Late Triassic, *P. mooniensis*, a characteristic species of the *P. crenulatus* Zone, has also been reported from the Rhaetic of New Zealand (Dickson, 1972), but is unknown in northern microfloras. Also in the Triassic, Mrs R. Kyle (pers. comm.) has observed the distinctive species *Annulispora microannulata*, *Hamiapollenites insculptus* and *Nevesisporites limatus* in the Beacon Super-Group of Antarctica.

Volkheimer (1972) has studied a sequence of Early Jurassic microfloras (in the Nequén Basin of Argentina) in which *Classopollis*, *Nevesisporites vallatus* and *Zonalapollenites* appear in the same order as in the Moreton Basin. Here again, *N. vallatus* is confined to these Gondwana occurrences. The same author (Volkheimer, 1968: 346) has also recorded the distinctive Australian species *Ischyosporites marburgensis* from the Middle Jurassic of the Nequén Basin, Argentina.

Detailed comparison is hampered by the lack of close study of the extra-Australian sequences. However, as indicated by the above examples, the available data indicate that this Queensland miospore sequence is more closely related to other Gondwana assemblages than to those of the Northern (Laurasian) regions, thus providing support to the Gondwana concept. While there are these general relationships among the Gondwana microfloras, recent studies of assemblages from Antarctica (R. Kyle, pers. comm.) and Argentina (R. Herbst, pers. comm.) have shown that there are locally restricted species in each region. It seems likely that, in the future, local provinces may be recognised on the basis of miospore assemblages in the lower Mesozoic of Gondwanaland.

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ABSTRACT

Cycadophytes apparently had their origin in the late Palaeozoic, and by the middle of the Mesozoic had established themselves as one of the dominant vegetational forms in many areas of the world. On the basis of a survey of fossil cycadophytes and recent collections of fossil plants, it is becoming apparent that the growth habit of extant members of the Cycadales does not represent that of most of the early Mesozoic cycadophytes. During the Triassic and Jurassic periods there were many forms with slender stems, often branched, with leaves that may have been deciduous. Leaves of present day Cycadales, as a rule, shrivel before they drop off and there is no way by which they could become flattened in accumulating sediments to become fossilised in the form of Mesozoic cycadophyte leaves. The latter are most often found complete or nearly complete (except when broken during excavation), with little or no evidence of shriveling, suggesting that there must have been an abscission mechanism. Furthermore, the densely packed nature of the fossilised leaves in certain areas suggests abscission of leaves in great profusion. This same habit seems to have occurred in Mesozoic cycadophytes throughout the world, and evolution of the group continued along similar lines on all continents even after separation had progressed to a considerable degree in the Mesozoic.

INTRODUCTION

The term 'cycadophytes' has been used in a variety of ways, and the connotations continue to change. Older systems of classification (e.g. Delevoryas, 1961) used the designation of class 'Cycadophyta' to include three orders, the seed ferns (Cycadofilicales or Pteridospermales), the Cycadeoidales (or Bennettitales), and the true cycads (Cycadales). Recently, however (e.g. Bold, 1973), the trend has been to consider the seed ferns as a separate class because it is becoming evident that not only do the Mesozoic seed ferns represent more than one order, but the late Palaeozoic ones may, also. Furthermore, even the concept of seed ferns makes it difficult to accommodate all Palaeozoic and Mesozoic forms in the same class. That leaves the class Cycadopsida to include the cycadeoids and cycads. This, too, is really only because of tradition and probably has no basis in fact. When faced with delimiting the Cycadopsida one is hard pressed to find characters that circumscribe both groups, the Cycadales and Cycadeoidales, accurately. In fact, except for superficial resemblances, the cycads and cycadeoids have little in common and probably represent two groups that diverged before the onset of the Mesozoic. Bold (1973) recognised this and listed the two classes, Cycadeoidopsida and Cycadopsida, within the division Cycadophyta. Admittedly, this move raises the problem of identifying the two classes to one level higher (the division), but it is a move in the right direction.

Because cycads and cycadeoids are so common in the Mesozoic, and because they ex-
isted in the same sediments, I will use the term 'cycadophytes' (with lower case 'c') in the traditional sense, and simply for convenience, and do not imply that both orders are naturally allied within the same class. There is also a practical reason for maintaining the term 'cycadophytes'; distinction of leaves of the two orders is based largely on differences of cuticular structure, and there are many instances where lack of cuticular preservation makes it impossible to distinguish between leaves of the two orders.

In this discussion I have drawn on works of many palaeobotanists who have investigated cycads and cycadeoids of the Mesozoic, particularly those of the Triassic and Jurassic. Although the forms included have come from widely scattered localities, it is significant that there tends to be considerable homogeneity among the cycadophytes, and from these reports of fossils from separated areas there is emerging a clearer picture of the evolutionary levels attained by the cycadophytes during the Mesozoic Era.

Both orders, Cycadales and Cycadeoidales, were in existence at the onset of the Mesozoic. It is natural, then, to look to the late Palaeozoic for signs of the appearance of one or both groups. Indisputable Palaeozoic representatives are rare, however. In 1969, Taylor described a petrified pollen cone that was considered to have had some cycadalean features. Later (Taylor, 1970), however, he admitted the possibility that the cone could have belonged to another group of gymnospermous plants.

Cridland and Morris (1960) described a plant, *Spermopteris*, with foliage of the *Taeniopteris* type from the upper Pennsylvanian of Kansas, U.S.A., and on these leaves, on the abaxial surface near the margins, were borne seeds. The seed arrangement is that which would be expected if our interpretation of cycadeid megasporophylls based on extant members of the order is correct. The genus *Cycas* demonstrates that the megasporophyll is a modified leaf and in that genus the sporophyll is more leaflike than in any other. Cridland and Morris, however, preferred to regard *Spermopteris* as a pteridosperm.

Mamay (1969) described fragments of two additional kinds of megasporophylls from the Permian. One of these, from the Lower Permian of Kansas, consisted of a rachis with ovules borne along both edges. On the basis of surface features, he suggested that *Taeniopteris*-like laminae found in association may represent distal parts of the sporophyll. Another Early Permian megasporophyll with possible cycadalean affinities has a distal lamina with seeds borne on the lower side of the proximal part of the leaf. The seeds were partially covered by an extension of the lamina curving abaxially.

A rather large number of additional late Palaeozoic fronds with possible cycadophyte affinity have been reported (see Seward, 1917; Anon., 1968), but in no instances are there corroborating fertile structures. Admittedly, certain late Palaeozoic ovules have much in common with those of cycadeids and, conceivably, could represent reproductive structures of some of the plants that bore the cycadophyte foliage.

**TRIASSIC AND JURASSIC CYCADALEANS**

Although the Cycadales probably originated in the late Palaeozoic, the first ones that make it possible to understand the growth form of the earliest members of the group are Triassic in age. For many years, the oldest completely reconstructed cycadeal was *Bjuvia simplex* (Fig. 15.1) from the Rhaetian of Bjuv, Sweden (Florin, 1933). The plant was supposed to have had sporophylls (called *Palaeocycas*), each with a distal lamina and proximally borne ovules (Fig. 15.1a). Large leaves of the *Macrotaeniopteris* type were associated (Fig. 15.1b). On the basis of epidermal structure, Florin was able to demonstrate that the megasporophylls and leaves were parts of the same kind of plant. No evidence was shown, however, that these organs were attached to the type of stem shown in the familiar reconstruction (Fig. 15.1c). *Bjuvia simplex*, then, while an important record of an early Mesozoic cycadeal, still does not provide insight concerning the appearance of the whole plant.

Fig. 15.1. *Bjuvia simplex* reconstruction. a. Megasporophyll (*Palaeocycas integer*); b. Leaf (*Macrotaeniopteris* type); c. Hypothetical restoration of plant. After Florin, 1933.
The first relatively complete cycadalean is the Late Triassic *Leptocycas gracilis* described by Delevoryas and Hope (1971). Although *Leptocycas* is antedated by other fossil cycadaleans, it is the first one to give us a glimpse of the habit of early Mesozoic cycadaleans. Material consists of numerous leaves of the *Pseudoctenis* type and stem remains. In one instance a stem fragment,
leaves, cataphylls, and what is presumably a pollen-bearing cone, are attached (Pl. 15.1).
Stomata are haplocheilic and typically like those in the order Cycadales.

Another fairly complete cycadalean is based on fossil remains from the Middle Jurassic of Yorkshire, England. The plant reconstructed by Harris (1961) has foliage of the *Nilssonia tenuinervis* type, pollen cones of *Androstrobus wonnacotti*, and seed cones of *Beania mamayi*. The restoration shows a slender plant, branched, with tufts of leaves at the tips of branches. It should be men-

Fig. 15.2. *Cycadwoidea* reconstruction. After Delevoryas, 1971.

tioned here that Harris did not have actual fossil remains of the stems of this plant, but based his reconstruction on indirect evidence that will be cited below.

Harris (1961) raised some valid questions concerning the restoration of a presumed Late Triassic cycad, *Dioonitocarpidium*, from Bavaria. The plant is shown with a *Dioon*-like stem with taeniopterid leaves and supposed megasporophylls, with ovule-like bodies borne proximally, and with pinnately divided distal portions (Lilienstern, 1928). It must be noted that no attachment was demonstrated and that there are certain aspects of the supposed megasporophyll that
need to be cleared up before it can be accepted without reservations as a cycad megasporophyll.

TRIASSIC AND JURASSIC CYCADEOIDALEANS

The best known of the cycadeoidaleans (bennettitaleans) are members of the genus *Cycadeoidea* (*Bennettites*) and restorations have been presented in numerous places (Fig. 15.2). These plants are characteristically Jurassic and Cretaceous, although Sidney Ash has shown me a cycadeoid of probable Triassic age. These plants had squat, globose, barrel-shaped, or columnar trunks with a terminal crown of pinnately compound leaves. Reproductive structures in the form of fleshy cones (not flower-like as shown in earlier reconstructions) were scattered on the trunk among the persistent leaf bases (Delevoryas, 1971). Although an image of these plants is what immediately comes to mind when one thinks of cycadeoidaleans, the growth habit is not most representative of the order. There are a number of other members that differ considerably in habit.

The reconstruction of the Middle Jurassic plant *Williamsonia gigas* is based on fossil remains from Yorkshire, England. (Harris, 1969, retains the generic name *Williamsonia* for ovulate cones; leaves of the plant are *Zamites gigas.* ) The plant, as restored, shows an erect, unbranched stem with a terminal crown of leaves and cones. Only a small fragment of the stem is known (named *Bucklandia*), and little is known about the actual habit of the plant. This is an important plant because it represents the first attempt to portray a fossil cycadeoidalean (Williamson, 1870).

In 1909 Nathorst presented a reconstruction of a cycadeoidalean (bennettitalean) from the Rhaetian of Scania, Sweden. Fragments of stems, leaves, and cones were available and the plant, called *Wielandiella angustifolia*, was shown as a slender, branching one. Leaves were of the *Anomozamites* type and were borne in tufts at the regions where the stem branched.

*Westerheimia pramelreuthensis* is an unusual fossil fragment described by Krasser (1918) and figured later in more detail by Kräusel (1943). It is from the Middle Keuper (Upper Triassic) of Lunz, Austria and consists of an axis with a slender lateral stalk on which are borne at least four tufted ovulate cones. The stem fragment is narrow, with loosely spaced leaf scars. A figure shows leaves of *Pterophyllum* nearby, but they are not attached and cannot be assumed to have belonged to the rest of the plant fragment.

The reconstruction of *Williamsonia sewardiana*, a cycadeoidalean from the Jurassic Rajmahal Hills of India, has been widely illustrated (Sahni, 1932) (Fig. 15.3). Leaves of the *Otozamites* type are arranged in a crown at the apex of the stem and on the tips of the presumed occasional branches. Cones are of the *Williamsonia* type. Although that reconstruction is the most familiar of the *Williamsonia* group, there are some uncertainties concerning its form; stems (*Bucklandia indica*), for example, have not been found with attached leaves.

The reconstruction of the Middle Jurassic *Williamsoniella coronata* from Yorkshire is, of course, well known. The first reconstruction, by Thomas (1915), shows bisporangiate cones borne at positions of forking of the stem axes. Zimmermann (1933) in his later reconstruction, implied that cone position was axillary. Leaves were entire, of the *Nilssoniopteris* type.

Harris (1969) presented a suggested restoration of another bennettitalean plant from the Middle Jurassic of Yorkshire. Stems were of the type referred to as *Bucklandia pustulosa*, leaves of *Ptilophyllum pecten*, and ovulate cones of *Williamsonia leckenbyi*. The restoration shows a branched plant with loosely arranged leaves. In the same publication, Harris referred to other bennettitalean remains and suggested that they were part of the same kind of plant. These remains consist of a *Bucklandia* type of stem bearing leaves of *Ptilophyllum pectinoides* and cones known as *Williamsonia hildae*. These apparently had scales around the fertile part of the cone known as *Cycadolepis hypene*.

A purely fanciful reconstruction of a plant called *Weltrichia mirabilis* was presented by Schuster in 1911. It seems to combine characters of cycadeoids and williamsonians, but the suggested reconstruction has no basis in fact.
These listed reports of Triassic and Jurassic Cycadales and Cycadeoidales are not the only ones in the literature, but they represent attempts by the authors to reconstruct the possible habits of the plants involved. There are countless other descriptions of isolated cycadophyte stems, leaves, and fertile structures. A few of these will be mentioned below.

**CONCLUSIONS CONCERNING MESOZOIC CYCADALES AND CYCADEOIDALES**

The most obvious conclusion we may draw is that knowledge of the structure of cycadaleans and cycadeoidaleans in the Mesozoic is far from extensive; and of these two groups, cycadaleans are even less well known than the cycadeoidaleans. In fact, *Leptocycas gracilis* (Delevoryas and Hope, 1971) is the only fossil member of the Cycadales that has allowed a reasonably complete reconstruction of the entire plant. The other fairly completely restored member of the Cycadales is...
that of Harris (1961) who showed a plant combining the parts Beania mamayi, Androstrobus wonnacotti, and Nilssonia tenuinervis. The stem, however, is imaginary.

Members of the Cycadeoidales are slightly better known, but again there are a number of uncertain points. Cycadeoidea is the best understood, partly because of the abundance of petrified stems that have been collected and partly because of the relatively excellent state of preservation of so many of them. Furthermore, the cones were borne close to the stem and were surrounded by persistent leaf bases, thereby enhancing chances of preservation. Of the remaining cycadeoidaleans, Williamsonia gigas, Wielandiella angustifolia, and Williamsoniella coronata have reconstructions based on fragments that show at least some evidence of connection of parts. Many stem fragments and numerous fossils of leaves and cones are known and from them we may postulate the kinds of growth habits other cycadeoidaleans might have had, but no precise reconstructions can be

Fig. 15.4. Leptocyas gracilis. Reconstruction by Delevoryas and Hope, 1971.
presented until more convincing evidence of connection of parts is presented. Nevertheless, valuable information available from these parts gives us a basis for suggestions concerning the structure and evolutionary potential of Mesozoic cycadophytes.

HABIT OF MESOZOIC CYCADOPHYTES

Because so few Mesozoic cycadophytes are known in a relatively complete state, the two orders, Cycadales and Cycadeoidales, are grouped together here. Furthermore, members of both orders grew together, and together they contributed a significant part of the vegetation of the Triassic and Jurassic.

Leptocycas gracilis, the only fairly complete cycadalean plant known from the Mesozoic, is important because it presented in Late Triassic times a kind of plant growth-habit quite distinct from that of modern Cycadales. The stem was slender, not more than 5 cm in diameter, relatively smooth (i.e., not covered by persistent leaf bases), and bore leaves rather widely spaced (Fig. 15.4). A number of stem fragments have been found (identified on the basis of cuticular features), and none indicates branching, so the reconstruction presented by Delevoryas and Hope shows an unbranched stem. It is recognised, however, that additional finds of...
the material could indicate that branching did exist, although it would appear that even if the stem were branched, it was not profusely so. The most significant aspect of Leptocycas is that no present-day member of the order has a stem that is so elongated and slender, with such loosely arranged leaves. Although certain modern Cycadales have elongated, erect stems (e.g., Macrozamia hopei, Dioon spinulosum, Microcycas calocoma), they are considerably thicker than those of Leptocycas and have the characteristic, persistent leaf bases that are closely spaced. In general, however, cycadalean stems are short and squat. Zamia pigmee has a narrow stem (about 2.5 cm), but it never exceeds several centimetres in length.

When he presented a reconstruction of a plant bearing the parts Nilssonia tenuinervis, Androstrobus wonnacotti, and Beania mamayi, Harris (1961) had to use indirect evidence to suggest a slender, branching stem (Fig. 15.5) because all the parts were not found connected. He explained that the stem was 'fairly thick', but indicated that by 'thick' is meant 'at least 2 cm'. That is not thick compared with modern cycads, nor even when compared with the slender Leptocycas. He wrote that because the base of the leaf of Nilssonia tenuinervis is broad, the stem must have been that wide to accommodate it. Although he showed closely spaced leaf scars on the stem surface, there is no supporting evidence. No persistent leaf bases were present, according to Harris, because the leaf is thought to have fallen off cleanly. This differs from Leptocycas, which certainly did have persistent petiole bases (although not so regularly disposed in size and arrangement as in modern cycads).

Thus, although remains that allow the entire plant of a Mesozoic cycadalean to be reconstructed are still meagre, the evidence at hand points to a habit involving a slender stem without the characteristic closely spaced leaf bases so commonly associated with present-day cycads. More evidence to support this assertion will be presented below.

Among the cycadeoidaleans (bennettital-
(bennettitaleans) enough fossil material exists of the Rhaetian *Wielandiella angustifolia* (Nathorst, 1909) to allow a fairly complete reconstruction. Stem axes were generally less than 1.5 cm thick, and branched in a pseudo-dichotomous fashion. In the forks were cones and tufts of leaves of the *Anomozamites* type (Fig. 15.6). Elsewhere the stems seemed to have been naked, although polygonal leaf scars are often found just below a fork and, occasionally, on other areas. Thus emerges a picture of a cycadophytic type also different in habit from that in modern cycadaleans.

The little known plant from the Upper Triassic of Lunz, Austria, *Westersheimia pramelreuthensis*, had enough of the stem preserved to indicate the existence of another slender cycadophyte in the early part of the Mesozoic. The material shows only a fragment of a plant, but enough to indicate that part of the stem system did not exceed 4 cm in diameter, and leaf scars suggested uncrowded leaves, and a lack of persistent leaf base armour. Cones attached to a slender branch show relationship to those of *Williamsonia* (Fig. 15.7). Reconstructions of other cycadeoidaleans (bennettitaleans) point to the same kind of conclusions concerning growth habit. *Williamsonia gigas* reconstructions are based on stem fragments to which are attached leaves of *Zamites gigas* and cone peduncles. Although the stems are not complete, the parts that are known seem to range from 3.5 cm in diameter. Although surface details are often unclear, Seward (1917) felt that persistent rhomboidal leaf bases were found on stems. Peduncles of cones, some about 3 cm in diameter, branched off the stem near the apex, and some of these peduncles themselves were branched. No branching has been observed in the stem axes, so it was assumed, as recorded in the reconstructions, that the stem was unbranched (Fig. 15.8). In general habit, then, *Williamsonia gigas* looks much like *Leptocycas gracilis*, except that in the former, leaves must have been more crowded and did leave persistent bases after abscission.

Fig. 15.7. *Westersheimia pramelreuthensis* as figured by Kräusel, 1943
The restoration of a cycadeoidalean by Harris (1969), based on association of stems of *Bucklandia pustulosa*, leaves of *Ptilophyllum pecten*, and ovulate cones of *Williamsonia leckenbyi*, also presents a habit quite different from that usually associated with cycadophytes (Fig. 15.9). Harris emphasised that he has seen no organic connection of parts but felt strongly that consistent association of parts cannot be ignored. Furthermore, the leaf scars on the *Bucklandia* are suitable for the leaves thought to have been attached. While the principal axis in the restoration is broad (up to 15 cm in diameter), laterals are much more slender, as little as 7 mm thick. The case for leaves and cones belonging to the same plant is stronger than that for the stems, because *Williamsonia leckenbyi* is always found in association with *Ptilophyllum pecten* but not with *P. pectinoides*. The latter is always found in association with another *Williamsonia*, *W. hildae*. The two cones do not occur together. Harris visualises the plant as a tree, much branched, with laterals of progressively smaller size. Obviously, with more cambial activity the lateral branches would have increased in
Mesozoic Cycadophytes

Fig. 15.9. Reconstruction of plant combining stems of *Bucklandia pustulosa*, leaves of *Ptilophyllum pecten*, and cones of *Williamsonia leckenbyi*. From Harris, 1969.

diameter, but the end result still would not resemble the trunk of a present day member of the Cycadales, in which plants the trunk, even though there is secondary vascular development, remains more or less columnar for its entire length (except, of course, in the smaller, more bulbous forms).

A similar, branched habit must have existed in the plant that probably combined the parts *Ptilophyllum pectinoides*, *Williamsonia hildae*, and stems indistinguishable from those in the plant reconstructed above. Harris found the same kind of *Bucklandia pustulosa* with these leaves and cones as he did with *Ptilophyllum pecten* and *Williamsonia leckenbyi*. No figured restoration of the plant was presented, however. Even if these two types of plants were not exactly as imagined by Harris, certainly some plants had stems such as those of *Bucklandia pustulosa* with a habit quite unlike that of modern Cycadales.

*Williamsoniella coronata*, as reconstructed by Thomas (1915) (Fig. 15.10) is closely similar in habit to *Wielandiella angustifolia*. Stems of *Williamsoniella* (so identified because of association with cones and leaf scars that match the base of the leaf, *Nilssonopteris viitata*) are slender, 0.7-2 cm thick, and branched in a forking fashion. Thomas considered the cones to have been borne in the angles of the forks as in *Wielandiella*, but Zimmermann (1933) felt they were axillary (Fig. 15.11). Nevertheless, the resemblance to *Wielandiella* is striking and *Williamsoniella* provides still another example, this one Middle Jurassic, of a cycadophyte different in habit from that of present-day Cycadales. Leaves of *Williamsoniella* were apparently more widely scattered than in *Wielandiella*, and did not leave the characteristic cycadalean armour when they abscissed.

Although no formalised reconstructions of other cycadophytes from the Triassic and Jurassic have been presented, I wish to include descriptions of remains since these fossils offer additional evidence that Triassic and Jurassic cycadophytes were generally slender plants, often branched.

The Middle Jurassic Rajmahal Hills in India have yielded a number of cycadophytic stem remains that give considerable insight into the general habit of this group of plants. Bose (1953a) described *Bucklandia sahnii*, which has a stem diameter ranging from 0.7-2.8 cm. In another paper (Bose, 1953b) he reported on a number of additional cycadophytic stems. One fragment about 7 cm long had a diameter of 6 cm, tapering to 4.5 cm. Another slender fragment measured 1.5 cm thick. A third, branched stem was about 2.5 cm thick, a fourth 3.5 cm, and a fifth, 5 cm. The precise affinities of these fragments are not known, of course, but there is little doubt that they may be assigned to the cycadophyte complex.

Cycadophytic stems from the Jurassic of Oaxaca, Mexico, are not especially abundant, but Wieland (1914-16) described some and assigned them to *Williamsonia*. The greatest diameter of these fragments is about 6 cm. They are covered with closely spaced, persistent scars that Wieland interpreted as
Fig. 15.10. *Williamsonia coronata*. Reconstruction by Thomas, 1915.

Fig. 15.11. *Williamsonia coronata*. Reconstruction by Zimmermann, 1933.
sites of abscission of scales or cataphylls of the *Cycadolepis* type. Some of these fragments were reported to have had nodes up to 7 cm apart. This is misleading; Wieland felt that expanded foliage was borne at these intervals and did not consider the scars left by scales (which are morphologically the equivalents of leaves) as representing nodes. At any rate, the habit, with widely spaced leaves (exclusive of scales) was quite different from all other cycadophytic forms, including early and middle Mesozoic ones.

More evidence of the habit of cycadophytes comes from the Late Triassic of North Carolina, U.S.A. A recently discovered stem fragment, less than 2 cm in diameter, bears leaves with a spacing very much like that in *Leptocycas gracilis*. Cuticular analysis of leaves, however, indicates a cycadeoidealean (bennettitalean) affinity, and leaves are of the *Otozamites* type (Pl. 15.II). Investigation of this plant is not complete (it is being studied by Delevoryas and Hope), but enough evidence exists to demonstrate another slender early Mesozoic cycadophyte.

Undoubtedly there are other examples of Mesozoic cycadophytes that would add direct evidence of a predominantly slender growth habit of these plants. There is also indirect evidence that is sufficiently convincing not to be ignored. This evidence is in the form of cycadophyte leaf fossils. The first point is that the way in which leaves were borne and the manner by which they fell off the stem were different from those of the present-day Cycadales. In the Cycadales, the usual situation appears to be one in which a leaf, after it has ceased to be functional, shrivels up while still attached to the stem. When it drops off it is certainly different in shape from what it was while functional on the plant. It would be difficult to visualise a leaf of a modern cycad being found as a fossil in a form that approximates the shape of the leaf while still on the stem. Mesozoic cycadophyte leaves, on the other hand, are usually found in an expanded form, almost never shrivelled, and closely approximating the condition they must have had while borne on the stems. Many of the leaves, especially the smaller forms, are found in their entirety. This fact emphasises another point concerning early and middle Mesozoic cycadophytes. Many of these plants had small leaves. A few modern Cycadales have relatively small leaves, but none so diminutive as many of the Mesozoic forms. Many of the leaves of the genera *Ptilophyllum* and *Otozamites* described by Wieland from the Jurassic of Oaxaca, Mexico, measure less than 10 cm in length. Others approach 20 cm, but they are slender and delicate, again showing a leaf type quite different from that of many modern cycads. Collections of cycadophyte leaves from Upper Triassic beds in North Carolina, U.S.A., yield remains of leaves 10 cm or less in length. Leaves from the Jurassic of Yorkshire, England (e.g., *Ptilophyllum pecten, P. pectinoides*) are rather small and delicate, generally about 20 cm in length.

Also interesting is the manner in which cycadophyte leaves occur in Mesozoic sediments. Even if leaves of modern Cycadales did abscise in an expanded form, the number that drops off at a given time would be small, and if they were preserved as fossils, they would occur sparingly in the rocks. In a good many instances in Mesozoic rocks cycadophyte leaves occur in great abundance, closely packed together, reminiscent of the way angiosperm leaves may be found in certain Tertiary sediments (Pl. 15.III). This phenomenon could reflect the greater abundance of cycadophytic plants during the Mesozoic, but I am inclined to think that this frequency and packing of leaves in sediments suggest that they were borne in greater numbers than the leaves on modern cycadeoideans. A counter argument may be that accumulation in the rocks is accentuated by current transport and deposition in certain areas. This, it seems to me, would strengthen the original argument, since there must have been many leaves from many sources that were gathered by the moving streams. Even if transport by water had been involved, in many areas it must not have been too significant a factor because the nature of the sediment suggests little or slow water movement.

The aggregation of leaves in the sediment also suggests a periodic or cyclic time of abscission, again reminiscent of the habit of
angiosperms as reflected by leaves preserved as fossils. The leaves are often cut off cleanly at the base as would occur if an abscission zone were produced.

The occurrence of many small leaf types would also point to their attachment on slender axes. Admittedly, many large plants (e.g. *Sequoia*) bear small leaves, but these leaves are attached to slender, distal branches. Because it seems that many leaves were borne on individual plants, the logical way for this to have occurred is as tufts on many branches. The dropping off of leaves cleanly, rather than after withering on the branches, also adds evidence that many stems were left without persistent leaf bases as in many modern cycads.

To present a balanced picture it should be pointed out that not all early and middle Mesozoic cycadophytes had small leaves and, most likely, not all Mesozoic cycadophytes had slender, generally branched stems. *Cycadeoidea* is the outstanding exception, but there must have been others. Some Mesozoic cycadophyte fronds were impressively large. I have collected *Otozamites* fronds in the Jurassic of Mexico that were more than 60 cm in length. The petiole was massive, and it must have taken a robust stem to have borne such leaves. The main point I have been trying to make is that the more frequent habit of Triassic and Jurassic cycadophytes is the slender, often branched, type; other habits must have existed, but they were in the minority.

**Cycadophyte Distribution in the Mesozoic**

An interesting point seems to have surfaced in this study and it concerns the makeup of Mesozoic floras throughout the world. Much emphasis has been placed on similarities of plants and animals of the same age in the now separated components of Gondwanaland. Certainly, these distributions played a large part in earlier ideas about positions of continents. Just as interesting, however, is the fact that floras at various times of the Mesozoic (and earlier) were amazingly similar throughout vast areas of the Earth. I can think of no place on Earth at the present time that reflects such a uniformity in floral makeup as existed in the Palaeozoic and Mesozoic Eras. For my examples in this paper I have drawn on reports of Triassic and Jurassic cycadophyte fossils from such widely separated places as Mexico, England, and India. Similarity of forms, even correspondence in genera, is striking. Cones of *Williamsonia*, for example, may be virtually identical in the Jurassic of Oaxaca, Mexico, and the Rajmahal Hills, India. *Williamsonia* cones described from Mexico by Delevoryas and Gould (1973) could just as easily have come from Yorkshire, England. Leaf genera such as *Piilophyllum* and *Otozamites* are extremely widespread. *Cycadeoidea*, too, has counterparts in both North America and various parts of Europe. This point has been discussed in a recent publication (Delevoryas, 1973) so I will not belabour it here. But there are interesting problems that must be considered if we are to understand how it was possible for such a widespread distribution of similar plants to have existed in the past. Certainly there must have been physical barriers encouraging isolation and diversity similar to those that exist today. Yet they must have been less effective than they seem to be at the present time. In my paper (cf. 1973) I felt that perhaps it was an increase in the numbers of kinds of biological forms that came about late in the Cretaceous that provided the stimulus for more local diversification. And it was biological competition rather than physical barriers that were most important for the development of regional differences in plants.

Finally, a word explaining the emphasis on Triassic and Jurassic cycadophytes in a paper concerned with Mesozoic cycadophytes. The simplest explanation is that it is mainly in these two periods that enough fossil remains are found to allow reconstructions of whole or partial plants. The exception is *Cycadeoidea*, of course, that reached its maximum development during the Cretaceous. Its usually bulbous trunk is more characteristic of modern cycadophytes, and it perhaps reflects a trend that occurred...
Gondwana Flora
during the Cretaceous but is unrecorded in the fossil record. There are abundant remains of foliage and cones of cycadophytes in the Cretaceous, but not enough to give us insight into the general growth habit. It is interesting to observe that the cone of *Williamsonia bulbiformis* from the Cretaceous of Santa Cruz, Argentina, is like that of williamsonians in all other parts of the world (Menéndez, 1966). Cycadophyte leaf genera such as *Dictyozamites* and *Otozamites*, also from the Cretaceous of Argentina, are remarkably like those in other parts of the world (Archangelsky, 1970). It would seem, then, that similar evolutionary events were occurring on a worldwide scale during the Cretaceous, even after significant continental separation had occurred, and the extinction of the Cycadeoidea by the end of the Cretaceous apparently occurred everywhere almost simultaneously as did the subsequent evolution of modern cycadalean forms from their earlier Mesozoic progenitors.

**ACKNOWLEDGMENT**

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Section 3

Environment and Origin of Gondwana Coal Deposits
Comparative Petrography of Gondwana and Northern Hemisphere Coals Related to Their Origin

M.-Th. Mackowsky

Abstract

Four coalfields have been contrasted and compared on the basis of qualitative and quantitative studies of macerals, microlithotypes and minerals. The differences are such that a clear distinction between Gondwana and northern hemisphere coals is not possible with petrographic methods alone, and other geological parameters have to be considered.

Introduction

The present paper aims at a comparison of Gondwana and northern hemisphere coals on the basis of their respective petrographic composition. This is a considerable task that can only be an attempt as so many problems concerning maceral genesis remain unsolved even in northern hemisphere coals, let alone those of the southern hemisphere. I have therefore reduced the all-encompassing theme to a consideration of European coals on the one hand and those of Australia and South Africa on the other, highlighting the genetic differences between the two groups.

The coals of the northwest German coal belt were formed during the Late Carboniferous in two different geological settings. One was the fore-deep (marginal basin) and adjacent shelf north of the Hercynian Fold Belt, and the other the intra-montane basins within the latter (Fig. 16.1). The climate during coal formation was largely subtropical. Gondwana coals range from Permian to Triassic in age. The climatic conditions appear to have been temperate to cool humid. In Australia, as in northwest Europe, coal deposition occurred to a large extent in a fore-deep. Diessel (1971) compared the Sub-Alpine Molasse in Bavaria with Australian coal-forming environments, and Kneuper (1973) also referred the coals of the intra-montane Saar-Lorraine Basin to a molasse environment. South African coals, on the other hand, were formed in continental troughs on an old cratonic basement. In the following discussion then, we will compare coals from similar fore-deeps or marginal basins but different climatic environments (eastern Australia and northwestern Europe, especially the Ruhr) from similar climatic but different geological environments (South Africa and eastern Australia) and from similar geological (limnic) but different climatic environments (South Africa and the Saar-Lorraine Basin).

The relations between petrographic features and the different modes of origin of the coals is presented in the form of a discussion of qualitative and quantitative differences in maceral compositions, microlithotypes, and the particularly interesting syngenetic minerals. The type and distribution of the latter often has a greater bearing on the solution of environmental problems than differences in both composition and layout of macerals and microlithotypes.

Comparison of Composition and Layout of Macerals

In Figure 16.2 results of maceral analyses of the two groups of Gondwana coals available to us have been contrasted with
### Geologic Age and Condition

<table>
<thead>
<tr>
<th>Geological age</th>
<th>Gondwana coal</th>
<th>European coal</th>
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<tbody>
<tr>
<td></td>
<td>Australia</td>
<td>South-Africa</td>
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<tr>
<td>Permian - Triassic</td>
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<td>Geological condition</td>
<td>fore deep paralic-limnic</td>
<td>continental trough limnic</td>
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<tr>
<td>Climate</td>
<td>temperate</td>
<td>cool</td>
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### Flora

- **Gondwana coal**: Glossopteris-Flora
- **European coal**: Lepidophytes-Flora

### Type of Coal Genesis

- **Gondwana coal**: hypautochthonous-autochthonous
- **European coal**: autochthonous-hypautochthonous

### Redox Potential and pH Value

- **Gondwana coal**: 0 - negative
- **European coal**: 0 - positive

**Fig. 16.1. Comparison between Gondwana coals and European coals**

The differences between the two groups of European coals. The two European coals have an average vitrinite content markedly above the two Gondwana coals but the maceral composition of Australian coals is particularly variable. On the other hand, Gondwana coals display a higher average inertinite content, while paralic coals contain less exinite than is found in limnic environments. This could be explained by the respective hydrogen ion concentration. Peat logs are commonly acidic, but under the influence of sea water they can become alkaline. Exinite, which is chemically quite resistant under acid conditions, may then be destroyed.

The differences between the two Gondwana coals on the one hand, and the two European coals on the other are highlighted further when not only maceral groups but also individual macerals are considered.

### Vitrinite

Figure 16.3 shows that the maceral group vitrinite consists of the two macerals telinite and collinite. The term vitrodetrinite is used only when dealing with vitrinite fragments of such small size that they cannot be referred to as either telinite or collinite with sufficient confidence.

Macerals can be further subdivided into maceral types, maceral varieties and cryptomacerals. The distinction between various maceral types is based on small genetic differences while maceral varieties are distinguished on the basis of noticeable differences in source material. The diagnosis of cryptomacerals requires additional analysis techniques, e.g. etching of polished coal blocks by acidified potassium permanganate.

No different types of telinite have been described so far, though a considerable number of varieties have been recognised. In my opinion distinct differences must exist between the lepidophytes-telinite typical of the European coals, and the *Glossopteris*-telinite that characterises Gondwana coals. In the latter, annual growth rings, which are proof of climatic fluctuations and are often found in petrified wood of the Gondwana regions, might be recognised.

No marked differences in the collinite...
from the various coal-forming environments can be found. Brown et al. (1964) noticed differences in reflectance between thick vitrinite layers and the collinitic cement of microlithotypes, respectively. The type called Vitrinite A, which is largely equivalent to the microlithotype vitrinite, has a reflectance that is significantly above that of Vitrinite B (Fig. 16.4). Similar observations on coals from Lorraine were published by Alpern (1966), and our investigations support this for the Saar coals. Alpern's homocollinite is analogous to Vitrinite A and heterocollinite to Vitrinite B. Attempts to recognise similarly large reflectance differences between the collinite cement of microlithotypes and vitrinite have not been successful, as shown in Figure 16.5. We have no knowledge of similar investigations in South Africa. It is therefore certain only that differences in reflectance between Vitrinite A and B, which in the international nomenclature proposed by the I.C.C.P. are called telocollinite and desmocollinite, cannot be used to distinguish European and Gondwana coals. There is some uncertainty concerning the cause of the above mentioned reflectance differences. Etched coals (Pl. 16.1) show that telocollinite does not always consist of cryptotelinite. Desmocollinite consists of telonitic detritus which is usually strongly gelified. This is shown by the vague outline of cell wall fragments which are not as clearly defined as those of the cryptotelinite or telocollinites. Such detrital habit of the cryptotelinite is, however, not the cause of the differences in reflectance because detrital habit is likewise typical for the collinitic matrix of clarites and trimacerites in the coals of the Ruhr Basin. Electronmicroscopic and autoradiographic investigations by Alpern and Pregermann (1956a) as well as by Alpern and Quesson (1956) suggest the possibility of minute inorganic impurities, e.g. clay minerals, or inorganic-organic complexes, are the
<table>
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<th>Group Maceral</th>
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<th>Submaceral</th>
<th>Maceral Variety</th>
<th>Cryptomaceral</th>
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<td>Vitrinite</td>
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| Exinite       | Sporinite     |                                           | Tenuisporinite                           | Cryptoexosporinite                      |
|               |               |                                           | Crossisporinite                          |                                         |
|               |               |                                           | Microsporinite                           |                                         |
|               |               |                                           | Macrosporinite                           |                                         |
|               |               |                                           |                                         |                                         |
|               |               |                                           |                                         |                                         |
|               |               |                                           |                                         |                                         |
|               |               |                                           |                                         |                                         |
|               |               |                                           |                                         |                                         |

| Inertinite    | Micrinite     | Pyrofusinite                 | Plectenchyminite                         |                                         |
|               | Macrinite     | fusinite                    | Corporsclerotinite                       |                                         |
|               | Semifusinite  |                            | Pseudocorposclerotinite                  |                                         |
|               | Fusinite      |                            |                                          |                                         |
|               | Sclerotinite  |                            |                                          |                                         |
|               | Inertodetrinite |                        |                                          |                                         |

Fig. 16.3. Table of macerals and group of macerals (International Handbook, 1963)

cause of the reduced reflectance. However, the possibility of diffused impregnations by plant excretions being the cause cannot be discarded. The agreement between Australian Gondwana coals and those of the Saar-Lorraine Basin with respect to the occurrence of the two collinite types is thus only noted here without trying to invoke a genetic relationship between the two.

Fig. 16.4. Frequency diagram of reflectance values for vitrinite A and B (Fiery seam, Australia, Brown et al., 1964).
As shown in Figure 16.2 the exinite group constitutes a distinctly smaller percentage in Gondwana coals than in their European counterparts, and the Australian coals are lower in exinite than the South African ones. Among the European coals those from the Saar and Lorraine average a higher exinite content than those of the Ruhr Basin. Apart from those quantitative differences there are likewise qualitative contrasts which are primarily related to differences in the vegetable source. However, such variations sometimes seem to be related to genetic differences. In all European humic coals of the Carboniferous, sporinite consists of micro- and macrosporinite (Pl. 16.II). Although microspores considerably outnumber macrospores there are no seams that contain microspores only. Macrospores are so important in the northwest European coal belt that in the United Kingdom attempts were made some years ago to use macrospores in coal seam identification.

Only few macrospores have been described from Gondwana coals. According to Plumstead (1957) their rarity is related to the predominance of the pollen-bearing Glossopteris flora. Likewise related to differences in source material is the occurrence of crassispores, first described by Stach (1954) in European Carboniferous coals (Pl. 16.III) but regarded as an exception in Gondwana coals by Taylor and Warne (1960). These authors also point out that sporinites of a single coal 'are of variable reflectance', which, according to Murchison and Jones (1962), is due to the effects of oxidation during the biochemical stage of coalification. Apart from some brown colouring of macrospores in weakly coalified sub-bituminous coals, the sporinite of European Carboniferous displays uniform reflectance. Only in association with clay minerals, e.g. in claritic trimacerite-bearing and duritic carbargilites, do white spores occur. By slightly changing Murchison and Jones's interpretation it may be argued that this variable reflectance of spores in one coal seam could also be related to pre-diagenetic oxidation since neither spores nor cuticles are resistant to atmospheric attack. If this assumption is true, it would explain in terms of dry environmental conditions, not only the low percentage of exinite in Gondwana coals, but also their variation in reflectance and the poor preservation of spores and cuticles. The white spores in the carbargilites of the Ruhr district, which are admittedly rare, may be explained in the same way.

Though Gondwana coals are relatively poor in sporinite and cutinite, they carry a higher proportion of resins and waxes due to the characteristics of the floral source. There are, however, differences in the proportion of the maceral resinite which, by definition, displays a lower reflectance than the vitrinite of the same seam. Resinite occurs as infillings in cell lumens of telinite, as isolated resin bodies or as diffuse impregnations of collinite. The reflectance characteristics of both resins and waxes are altered by oxidation and polymerisation. Present knowledge suggests that corpocollinite has been derived from resin bodies in a similar way to a large proportion of corpocollinite. The latter is therefore called pseudo-corpocollinite. It is interesting to note that the coals from the Saar-Lorraine Basin, which are particularly rich in vitrinite, are also rich in resinite but very low in corpocollinite and even lower in pseudo-corpocollinite. They are followed by the coals of the Ruhr Basin and possibly
Pl. 16.I. Etched collinite: cryptotelinite and cryptocorpocollinite. x 250, oil immersion.
Pl. 16.II. Macrosporinite and microsporinite in a high volatile Ruhr coal. x 250, oil immersion.
Pl. 16.III. Crassistporinite in a durite, seam J, Ruhr coal. x 250, oil immersion.
Pl. 16.IV. Light grey macrospore in a trimacerite (with about 10-15 per cent clay minerals).

The sporinite in this high volatile coal is in general dark grey. x 250, oil immersion.
Australian coals, whereas the South African coals seem to carry high proportions of crypto-corpocollinite and pseudo-corpoclerotinite. If this observation, which so far has been based on few observations only, proves correct, then it would seem that similar vegetable source material yielded high resinite contents when aerobic attack was prevented, but low resinite contents when the plant material was kept under prolonged aerobic conditions prior to burial.

**Inertinite**

The most obvious differences between Gondwana coals on the one hand and European hard coals on the other are displayed in the ratio of vitrinite to inertinite, as is shown on Figure 16.3. In addition to having a higher carbon and lower volatile matter content, Figure 16.6 shows that inertinite is also richer in oxygen than vitrinite or exinite. This relatively high oxygen content confirms the general opinion that aerobic conditions of decomposition play a more important part in the formation of inertinite than in vitrinite. High inertinite contents are therefore proof of a low ground water table, the more so, when dealing with degradofusinite or semifusinite which according to Teichmüller (1944) is often associated with pyrofusinite. Degradofusinites are also often associated with sclerotinite, which is proof of the importance of fungal attack in their origin. Fusinite and semifusinite do not appear to be suitable parameters when differentiating between Gondwana and European coals, although it appears that larger areas and thicker bands of the two macerals are more common in the coals of the Ruhr Basin than in Gondwana coals. This fact might indicate a higher proportion of pyrofusinite in European coals and shorter distance of transportation. The much higher proportion in Gondwana coals of inertodetrinite, the individual particles of which differ markedly in reflectance (Pl. 16.V), indicates longer distances of transportation and the consequent degradation of inertinite materials. As was indicated by Taylor and Cook (1962) the sclerotinite, so common in Gondwana coals, does not always consist of fungal remnants. This refers particularly to corposclerotinite. In the International Glossary on Coal Petrography of the I.C.C.P. a distinction is made between a corposclerotinite which has certainly been

Fig. 16.6. Oxygen and carbon content of the maceral groups with increasing rank
Pl. 16.V. A special type of inertinite in Gondwana coal. The inertinite particles show rather strong differences in reflectance. x 500, oil immersion.
Pl. 16.VI. Degradomicrinite (upper line) and detritomicrinite (stereoscan)
Fig. 16.7. Table of microlithotypes (International Handbook, 1963)

Fig. 16.8. Microlithotype composition of Ruhr, Saar-Lorraine, South African and Australian coals derived from fungal spores, and a pseudo-corposclerotinite, which is more frequent in Gondwana coals than in the Ruhr or Saar Basins. It is almost totally absent in the Saar. I am not in a position to judge whether the sclerotinite variety called plectenchyminitine is partly or wholly of fungal origin in Gondwana coals, or whether it has been derived from degraded plant tissues. In the coals of the Ruhr Basin it has been assumed to be fungal.

The genesis of macrinite, which occurs in trimacerites and durites of both Gondwana and Ruhr coals, is still so doubtful that I do not intend to discuss this maceral of the inertinite group any further. On the other hand, it seems that the maceral micrinite is of special interest. It consists of small inertinite particles with a diameter ranging from less than 5 mm to 10 mm. As shown in Plate 16.VI it occurs as infillings in cell lumens of telinite in both Australian and European coals. South African coals are, as far as I know, devoid of micrinite. Recent investigations with the scanning electronmicroscope have shown (Mériaux, 1969), that even-grained micrinite...
is generated during the decay of certain cell walls (Fig. 16.7). Likewise, it is possible to obtain micrinite from the decay of the cell content, and in many cases, both varieties occur side by side. I would like to call this type of micrinite degrado-micrinite. It is particularly even-grained and occurs occasionally disseminated in desmocollinite. Its mode of origin is as unknown as its chemical composition. The micrinite often observed by Stach and Alpern (1966) in the immediate vicinity of sporinite is in my opinion of different origin and it is thus a different micrinite type which I intend to call detritomicrinite. Figure 16.8 shows that this is very finely fragmented inertinite which shows a great variation in both particle shape and size. This degrado-micrinite is common in African coals and is particularly rare in the Saar coals which, in any case, have the lowest percentage of inertinite.

**COMPARISON OF COMPOSITION AND LAYOUT OF MICROLITHOTYPES**

As shown in Figure 16.9 the three maceral groups give rise to three mono-maceral, three bi-maceral and one tri-maceral microlithotypes. Results of microlithotype analyses obtained from the coals of the above mentioned occurrences have been compiled in Fig. 16.10. The compilation has been based on three variables—vitrinite plus clarite, durite plus trimacerite, and inertinite—and triangular diagrams have been obtained for coals from the Saar-Lorraine, the Ruhr and Australia. Lack of our own analyses prevented us doing the same for South African coals. The results show that vitrite and clarite are most prominently represented in the Saar-Lorraine coals, followed by the Ruhr coals which also show a wider variation. Apart from the higher proportion of inertinite, Australian coals are distinctive in their wide range of microlithotype composition. The variation shown for Australian coals in Figure 16.10 does not refer to individual seams but to average coalfield values; the lateral variation in microlithotype composition within one seam is no greater than in European coals. When, however, the results of analyses of different Australian seams are compared, differences in microlithotypes are considerable, even when the comparison is restricted to coals of low-to-medium rank (Ro max < 1.4 per cent) in which exinite is still clearly distinguishable from vitrinite. This seems to indicate considerable variation in the en-

![Fig. 16.9. Coals from some Australian seams plotted on the classification diagram (CSIRO, 1970)
<table>
<thead>
<tr>
<th>Mineral Group</th>
<th>Syngenetic Formation (intimately intergrown)</th>
<th>Epigenetic Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Transported by water or wind</td>
<td>Deposited in fissures cleats and cavities (coarsely intergrown)</td>
</tr>
<tr>
<td>Clay Minerals</td>
<td>Kaolinite, Illite</td>
<td>Ankerite</td>
</tr>
<tr>
<td></td>
<td>Sericite, Clay</td>
<td>Calcite</td>
</tr>
<tr>
<td></td>
<td>Minerals with mixed-layer structure, 'Tonstein'</td>
<td>Dolomite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ankerite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Siderite</td>
</tr>
<tr>
<td>Carbonates</td>
<td>Siderite-Ankerite concretions, Dolominate, Calcite, Ankerite</td>
<td>Ankerite</td>
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<tr>
<td></td>
<td></td>
<td>Siderite, Calcite, Ankerite in Fusite</td>
</tr>
<tr>
<td>Sulphide Ores</td>
<td>Pyrite concretions Melnikowitz-Pyrite Coarse Pyrite (Marcasite) Concretions of FeS2-CuFeS2-ZnS</td>
<td>Pyrite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marcasite</td>
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<tr>
<td></td>
<td></td>
<td>Zinc Sulphide (Sphalerite)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lead Sulphide (Galena)</td>
</tr>
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<td></td>
<td></td>
<td>Copper Sulphide (Chalcopyrite)</td>
</tr>
<tr>
<td>Oxide Ores</td>
<td>Haematite</td>
<td>Goethite, Lepidocrocite ('Needle Iron Ore')</td>
</tr>
<tr>
<td>Quartz</td>
<td>Quartz grains</td>
<td>Chalcedony and Quartz from the weathering of Felspar and Mica</td>
</tr>
<tr>
<td>Phosphates</td>
<td>Apatite</td>
<td>Phosphorite</td>
</tr>
<tr>
<td>Heavy Minerals and Accessory Minerals</td>
<td>Zircon, Rutile, Tourmaline, Orthoclase, Biotite</td>
<td>Chlorides, Sulphates and Nitrates</td>
</tr>
</tbody>
</table>

Fig. 16.10. Minerals in coal (Mackowsky, 1968)

Environmental conditions of coal formation, perhaps primarily related to differences in ground water table, but also to differing rates of subsidence.

In the South African coals, which are low in vitrinite, spore rich clarite appears to be fairly common, whereas in Australian coals clarite constitutes hardly more than 5 per cent. On the other hand, South African coals are low in durite, which constitutes up to more than 20 per cent of Australian coals. A comparison between the Ruhr and Saar coals shows that the latter are very rich in clarite and poor in durite. Of the trimacerites, clarodurite predominates in both Gondwana coals, whereas in those from the Saar duroclarites dominate and in those from the Ruhr clarodurite and duroclarite are both common. Pure inertinite is most frequent in South African and very rare in Saar coals whilst both Australia and the Ruhr show varying proportions in the trimacerite types. A better knowledge of maceral proportions within the microlithotypes and detailed analysis of pillar samples would certainly elucidate much further the problem of facies differences between the coal deposits dealt with.
Finally, we comment on the minerals associated with coal. In my opinion, the differences between them are so great that they outweigh the absence of analyses of pillar samples. In Plate 16.VIII the most common minerals in coal are displayed. In connection with questions of coal formation the syn-genetic minerals, especially clay minerals, carbonates, sulphides and quartz, are of particular interest. The main mineral impurities in the coal of most coalfields are clay minerals, such as kaolinite, illite, montmorillonite, mixed layer clays and sericite. Such minerals may be either washed or blown into the coal swamp as clastic material or they may form there as authigenic minerals. In Australian and the two European coals, minute clay minerals occur predominantly in vitrinite-bearing microlithotypes, e.g. vitrite, clarite and vitrinertite, and, more rarely, in durite and inertinite. This seems to suggest that the clay minerals were either formed authigenically in the swamp or constitute the finest clastic outwash. The latter idea is supported by the fact that, for instance in the Kreftenscheer Seam in the Ruhr Basin, a quartzose sandstone bed changes into a carbargilite with increasing distance from the source area.

In South Africa the inertinite-rich microlithotypes are particularly rich in clay minerals. This might suggest that the clay minerals and at least a portion of the inertinite are not washed by water but blown by wind into the depositional site. Further support for this suggestion is obtained from the lack of syngenetic carbonates and sulphides, which can form only under strongly anaerobic conditions. On the other hand, Australian and European coals, in both carbargilites and vitrinitic microlithotypes contaminated by clay minerals, quite frequently carry siderite concretions which, in view of their spherical shape (Pl. 16.IX), must have been formed in the early stages of diagenesis and peatification. The siderite concretions that were later formed have been interfered with in their growth, and consequently do not display a smooth shape or show the Brewster Cross in polarised light (Pl. 16.X).

The second mineral group that requires anaerobic conditions is the iron sulphides, either in the form of pyrite, marcasite or melnikovite. Pyrite is most common in coal seams that have been affected by marine
conditions, and is sometimes so common that it has been used in seam correlation. Since the Saar-Lorraine, the South African, and the Newcastle (or equivalent) Coal Measures of Australia are regarded as freshwater deposits, it is not surprising to find a low concretionary pyrite content in these coals. On the other hand, there is hardly a coal seam in the Ruhr Basin that is free from syngenetic pyrite. This marine influence is also seen in the dolomite nodules in both the seam roof (e.g. Katharina Seam) and within the seam (Pls. 16.XI, 16.XII). For many years they have been used in coal seam correlation.

Nowadays attempts are made to correlate between coal seams in different coal basins, e.g. limnic and paralic, on the basis of tonstein (claystone) horizons. There are several types of tonsteins, e.g. crystal-bearing, pellet tonstein, pseudomorph-bearing and vermicular (Fig. 16.11). The same types have been described from Australia, but I do not know whether similar tonsteins occur in South Africa.

It has been mentioned before that the coal seams affected by marine conditions contain much fine concretionary pyrite. Marcasite, on the other hand, is rare. The formation of pyrite took place under anaerobic conditions in the pH range of 6-9. More interesting than the pH is the redox potential, which can vary from $-300 \text{ mV}$ to about $+600 \text{ mV}$. In layers with much organic substances the redox potential is usually negative although the pH may be neutral or even in the alkaline range (Pl. 16.XIII). All three ranges are of interest when considering pyrite forma-
Pl. 16.IX. Siderite concretions with irregular forms. x 250, polarised light.
Pl. 16.X. Dolomitised peat (transmitted light). $\times 100$. 
Pl. 16.XI. Idiomorphic crystals of dolomite, reflected light. x 250, oil immersion.
Pl. 16.XII. 'Kaolin worm', reflected light. x 250, oil immersion.
Pl. 16.XIII. Spheres of pyrites in coal (see Fig. 16.11c)
Petrography of Gondwana and Northern Hemisphere Coals

1 — 150 yjm

1 — 10 yjm

PI. 16.XIV. Compressed spheres of pyrites

Pyrite concretions as shown in Plate 16.XIV are typical of Range C in which the water under the sediment has a redox potential of 0. Such spherical pyrites are later, during diagenesis, converted to lenticular pyrite so that individual crystals are difficult to identify. In Range B individual crystals are formed but no spheres. Perhaps the acicular crystals, which may have originated as marcasite, indicate a shift of the pH into the acid range (Pl. 16.XV). Range A is typical for the first stages of diagenesis, in which the redox potential is always negative. When lime is present the pH can be raised above 7 so that the spores, which are not stable under alkaline conditions, are replaced by pyrite (Pl. 16.XVI). If this is all correct the paucity of pyrite in the Saar coals may be related to a low H₂S content compared with CO₂ in spite of the negative redox potential as indicated by high vitrinite. In South Africa the high inertinite content does likewise indicate low H₂S, and in addition a positive redox potential is suggested, while in Australia in the Greta and Tomago Coal Measures, as well as in the marine influenced coals of the Ruhr district, the redox potential was negative.
Pl. 16.XV. Formation of pyrites— normally single crystals (see Fig. 16.11b)

The question whether bacteria reduced any $\text{SO}_4$ present in sea water remains unanswered since so far no bacteria have been diagnosed in coal. The Newcastle Coal Measures, which are limnic, are lower in pyrite and therefore more comparable with the Saar-Lorraine coals.

ACKNOWLEDGMENTS

The author is indebted to the Joint Coal Board, The Broken Hill Proprietary Co. Ltd, and the Utah Development Board for the generous financial support which gave me the chance to join the Third International Gondwana Symposium. Many thanks are expressed to Professor C. E. Marshall, Sydney University, and Professor C. F. Diessel, Newcastle University, for many fruitful discussions and the translation of this paper from German into English, and to Dr C. T. McElroy, Convener of Section 3, for assistance generally.

REFERENCES

Pl. 16.XVI. Formation of pyrites; pH value > 7, macrospores are replaced by pyrites (see Fig. 16.11a)


The Environment of Coal Formation in the Peninsular Gondwana Basins of India

P. K. GHOSH

ABSTRACT

Coal-forming environments were different in different basin belts of Peninsular India owing to the complex interplay of regional subsidence, tectonic setting, geomorphology and climatic change.

In the basins of the Damodar Valley belt, a convergent drainage pattern, coupled with rapid subsidence that accompanied sedimentation, periodically transformed the basins into coal swamps. The general deterioration of coal quality towards the basin periphery can be correlated with the introduction of clastics by the radial drainage.

The geological setting in Son, Mahanadi, Narmada and Pranhita-Godavari basin-belts presents a sharp contrast to that of the Damodar Valley areas. These were the centres of very large alluvial plains which were drained by a braided northwesterly flowing fluvial network. In these basin-belts, the supply of sediments by aggrading streams frequently outpassed the basinal subsidence and this imbalance resulted in the accumulation of a thick pile of coarse sediments and frequent scouring of the coal seams formed in the local backswamps by wandering river channels before these could be preserved by regional subsidence. Frequent splitting and digitation of the coal seams, their widespread regional variations both in thickness and quality, and even complete erosion of the coal seams in these basin-belts testify to frequent rejuvenation of the drainage system within the flood basins of these areas. The depositional model of the coal swamps of the peninsular Gondwana basins of India corresponds to that of hypautochthonous origin which involves mixing of plant debris by the intermittent action of running water within the general area of plant growth.

Recent advances in the study of coalification in the Damodar Valley basins indicate that the central part of the basin-belt (viz. Jharia) had been subjected to higher geothermal gradients resulting from the interplay of tectonics and the intrusion of igneous bodies radiating from a focus in the Rajmahals. Exposure to relatively higher temperatures has been instrumental in increasing the rank as well as coking properties of the coal in this part of this basin-belt.

INTRODUCTION

Since the Hercynian orogeny the Indian craton has been a stable landmass. Some pre-existing lineaments represented by shear and/or fault zones within the Precambrian plat-
basin-belts which coincide, in general, with some of the present-day river valleys—Damodar, Son, Mahanadi, Pranhita-Godavari, Narmada, etc. Gondwana basins spatially distributed in an *en echelon* fashion within these river valleys are generally referred to by the name of the valley in which they are developed (Ghosh and Mitra, 1970).

The coal measures occur in three separate stratigraphic units—the Karharbari, Barakar and Raniganj Formations—within the Permian succession. These three Lower Gondwana coal measures in turn have distinctive lithological attributes of their own and have been accorded different lithostratigraphic status within the Gondwana Super-Group.

The coal measures have been the object of much attention since the middle of the last century. However, a systematic study of the environment of sedimentation of the coal measures has been attempted only in recent years (Banerjee, 1960; Niyogi, 1966; Casshyap, 1970; Ghosh and Mitra, 1970). This paper embodies the recent trends of thought on the evolution of the different Gondwana basins of Peninsular India and the varying interplay of tectonism and geomorphology on the formation of the Lower Gondwana coal measures.

**TECTONIC SETTING OF GONDWANA BASINS AND BASIN DEVELOPMENT**

Most of the Gondwana basins show the 'half graben' structure with one side marked by a normal gravity fault or the boundary fault, while the other side is defined by the unconformable contact between the Gondwanas and the Precambrian rocks. In a few cases as in the Giridih Coalfield, Bihar and parts of Godavari Valley Coalfield, both the sides of the concerned basins are delineated by boundary faults and as such the structure is of 'rift type'. The 'half graben' configuration of the basins has caused homoclinal tilting of the basin floor towards the boundary fault. The boundary fault, in general, marks the limit of the deposition of the coal measures, though in a few cases older Talchir sediments and basal units of Karharbari Formation are found as outliers in the upthrown block of Precambrian rocks.

Besides the boundary fault, the Gondwana sediments are traversed by a network of normal faults which are either intrabasinal or basin marginal faults. Recent exploratory drilling in Karanpura and Ramgarh (Savanur, 1966) has proved beyond doubt that these faults have assumed an arcuate nature with the dip of the fault decreasing with depth. Sometimes the faults have even assumed a subhorizontal attitude as in the Sudamdih area, Jharia Coalfield. The overwhelming preponderance of normal faults and absence of structural elements like tight folds or prominent reverse faults resulting from compressional tangential stress convincingly demonstrates that dislocative structures of the peninsular Gondwana basins had taken shape in a tensional regime.

Systematic drilling investigations indicate that the Gondwana basins evolved as embryonic depressions within the pre-existing weak zones. From the subsurface data it is seen that the glacigene Talchir Formation forming the basal unit of the Gondwana succession has patchy distribution in Jharia and Karanpura fields of the Damodar Valley. This formation is mainly developed in bedrock depressions and wedges out over the basement prominences (Ghosh and Mitra, 1970). Evidently the domain of Talchir sedimentation was primarily confined to erosional depressions on the Precambrian rocks at the beginning of Gondwana sedimentation and the tectonic control of sedimentation was not imposed at this early stage. With passage of time these embryonic Gondwana basins developed into master basins with progressive subsidence which went *pari passu* with Karharbari, Barakar and Raniganj sedimentation.

Deposition of the full Gondwana sequence other than the Talchir was controlled by this tectonic phenomenon. Further, it is seen that the boundary faults exercised dominant control on the regional subsidence as maximum subsidence and sedimentation have often taken place close to the boundary fault as in Jharia Coalfield. Tectonic movement along the boundary fault was also the primary factor for progressive shifting of a basin axis and redefining
Environment of Coal Formation in India

the domain of sedimentation. In the Rani- ganj Coalfield the basin axis has progressively shifted south towards the boundary fault as sedimentation progressed. Although earlier views were that the Gondwana sediments were laid down in pre-faulted rift valley basins (Fox, 1934), or that present-day Gondwana basins represent the faulted remnants of extensive basins (Gee, 1932; Ahmad, 1961), recent thinking suggests that the boundary faults and some major interbasinal faults owe their origin to reactivation of pre-existing Precambrian faults during sedimentation (Chaterji and Ghosh, 1967). The overall tectonism involved in the process of growth and evolution of the Gondwana basins is an example of resurgent tectonism in which weak zones of antiquity are reactivated in a newly imposed stress regime and attain mobility. It is, however, imperative to note that most of the major Gondwana basins are composite in nature, with component sub-basins, troughs and downwarps each having its independent evolutionary history. The basin floors in the different peninsular belts are also structurally and geomorphologically inhomogeneous. Moreover, the nature of epi-rogenic movements in the widely separated basin-belts are variable in space and time. This tectonic movement was more pronounced in the Damodar Valley fields during the deposition of the coal-bearing Gondwana sediments and began to lose its intensity during Early Triassic Panchet sedimentation. Such tectonic movements were, on the other hand, less pronounced in the outlying fields during this period. Movement during the Panchet and post-Panchet times, however, played a significant role in the deposition of the Upper Gondwanas in the outlying fields. These complex regional variations of tectonics and geomorphologic setting have been reflected in the pattern of arrangement of lithic fill in the different basin-belts.

INTERPRETATION OF THE DEPOSITIONAL ENVIRONMENT OF LOWER GONDWANA COAL MEASURES

The Lower Gondwana Coal Measures occur in three stratigraphic units, the Karharbari (Lower), the Barakar (Middle) and Raniganj (Upper). The Raniganj Measures are separated from the Barakar by an inter- vening unit of Ironstone Shales or Barren Measures (100-610 m in thickness). The Karharbari Formation, varying in thickness from 50-150 m, is characterised by coarse greywacke, fireclay and coal seams. The sandstones are very poorly sorted with subangular grains of quartz and feldspar embedded in argillaceous matrix. Owing to extensive scouring of the underlying Talchir Formation, pellets of Talchir sediment are dispersed within the sandstones. Occasionally the base of the Karharbari Formation is marked by 'recomposed granite', as in the Karanpura Basin, indicating very little transport and reworking of the granitic detritus from the Precambrian positive area. The coal seams are usually dull, non-banded, and usually have relatively low ash and phosphorous content (Ghosh and Basu, 1969).

The Barakar Formation is the major coal-bearing unit in Peninsular India and has widespread development in the different basin-belts. It exhibits a variable thickness of 250 to 1000 m. The greater thickness has been recorded mainly in the Damodar Valley Coalfields—Jharia, Raniganj and South Karanpura Basins. This formation is characterised by an alternating sequence of coarse arkosic sandstones, siltstones, and well banded coal seams, and has a transitional boundary with the underlying Karharbari Formation.

The Raniganj Measures (100-1000 m in thickness) are primarily developed in the Damodar Valley belt, where they consist of fine-grained, calcareous, laminated sandstones, siltstones, shales and high volatile coal seams. In the outlying fields, except that of Singrauli, and some of the Central Indian fields where there is incipient development of the unit, it generally is a ferruginous sandstone-shale facies and is termed the Kamthi Formation.

FACIES ORGANISATION OF THE COAL MEASURES

The Lower Gondwana coal measures display a vertical alternation of sandstones and finer sediments. This rhythm in the lithic arrangement of the coal measures is manifested not only in the outcrops, but also at depth, as has been proved in a large number of boreholes drilled in the various peninsular coalfields. Krishnan (1960) suggested that the re-
petition of the coal seams and the associated sandstones and shales indicates a cyclic pattern of sedimentation. Banerjee (1960) and Ghosh (1970) discounted the concept of cyclic repetition of the litho-units of the coal measures. Niyogi (1966) has, however, described cyclic repetition of:

- channel sandstone
- laminated shale
- coal
- carbonaceous shale
- kaolinitic shale
- laminated shale

in the Barakar Coal Measures of Saharjuri Coalfield, Bihar, and has explained such lithic arrangement as due to a tectonic-geomorphic cycle. It was further suggested that the locus of cyclic sedimentation was in intermontane valleys and that the cycle appears to have been initiated by strong uplift which produced a youthful topography. At that stage the basin was filled by immature arkosic sandstones deposited as a piedmont alluvial sheet with superimposed channel bodies. With passage of time and tectonic quiescence, the size of the channel bodies progressively decreased while the shale percentage increased. The process of peneplanation continued and the landscape was covered with thick kaolinitic soil when the finer-grained sediments were deposited in the central swamps. Gradually tectonism was accelerated and a strong uplift caused deposition of sandstones of the succeeding cycle.

A large amount of borehole section, and the outcrop sections in various coalfields including Saharjuri Basin, however, does not exhibit the total cycle. Frequently the cyclic pattern is truncated, abbreviated, or slightly irregular. However, in the sedimentary organisation of the coal measures a certain degree of ‘fining upwards’ tendency is generally well manifested. In an individual ideal cycle, the base is sometimes defined by cross-bedded sandstones which grade upward in ripple drift cross-laminated sandstones and siltstones, and this in turn grades upward into carbonaceous shales and coal seams. Such a cyclic arrangement has been noted in the Bansloi River section in Rajmahal Coalfield, Bihar, and in the Jamunia River section in Jharia Coalfield, Bihar, etc.

The main feature of the fining upward sequence of the coal measures leads to a general interpretation in terms of water velocity. The basal, coarser-grained, cross-bedded sandstones, followed by deposits of decreasing grain size, denote decreasing velocity of flow, and the array of sedimentary structures in the fining upward sequence also provides a general trend from large to small bed shear stress and scour and deposition in relatively upper flow regime to sedimentation in lower flow regime as suggested by Allen and Friend (1968). The recurring succession of the fining upward sequence in the Lower Gondwana coal measures demonstrates periodic fluctuation of water velocity and depth.

ENVIRONMENT OF DEPOSITION OF COAL MEASURES

The lithological attributes of the coal measures, the primary sedimentary structures, and the fining upward units, suggest a fluvial environment of deposition. Therefore, analogues may be sought among the recent alluvial sediments studied by Allen (1965).

Most modern alluvial sediments have been differentiated into channel or substratum deposits, and finer overbank or top stratum deposits. Such a subdivision fits in well with the standard fining upward sequence of the coal measures. The basal coarse-grained members of the fluvial facies are typified by channel sandstones. In most observed cases the channel sandstones display co-sets of trough cross-stratification which are succeeded by horizontal lamination or ripple drift cross-lamination. These cross-bedded channel sandstones resemble pointbar deposits. The thick sandstone units consist of complex channel infillings which had either cut down in the underlying sandstones to form multistory sandstone units or in the upper argillaceous member of the underlying cycle. The thickness of the channel deposits in an individual cycle is related to complex sets of variable factors which include rates of channel migration, rates of relative subsidence of basin floor vis-à-vis the uplift in the positive area, the rate of supply of detritus and pattern of vegetal cover in the source area, etc.

The channel sandstones are succeeded by
Fig. 17.1. Clastic ratio map of XVI-XV interseam sediments in the central and southeastern part of Jharia Basin

ripple cross-laminated siltstones, carbonaceous shales and coal seams. Siltstones and lutites were deposited in the lower half of the lower flow regime by the construction of levees at bankful stage of the river, or by accumulation of sediments in the low-lying interchannel flood basins. Therefore, the fining upward cycles of the coal measures demonstrate the following sequence of events. The lower cross-stratified sandstone members were deposited within fluvial channels. Then the channels migrated laterally into an area of earlier overbank deposition, or a catastrophic change of course took place probably due to some tectonic disturbance. With that, the depositional area became a domain of overbank sedimentation when the thick sequence of fine-grained sediments and coal accumulated. The river channel then again migrated, reoccupying the earlier position to produce the channel sandstones of the next cycle. This fluvial cycle was repeated several times during the formation of the coal measures. The subsidence of the basin floor also went pari passu with sedimentation so that a thick pile of shallow water fluvial sediments could accumulate. In general, the subsidence was controlled by movement along the boundary faults. The mode of deposition of the sediments of the coal measures in the different peninsular Gondwana belts, however, needs to be further elucidated keeping this broad framework of fluvial sedimentation pattern in view.

Since the palaeocurrent pattern provides the integrating framework in which the different facies arrangements are set, it is imperative to outline the dispersal pattern of coal measures sediments in the different peninsular coal-belts for basin analysis. In the Damodar Valley coal-belt—Jharia, Bokaro and Ramgarh Basins—palaeocurrent data
show a centripetal type of sediment transport (Ghosh and Mitra, 1970). In these basins, small streams originating from surrounding uplands have discharged into the central depression. The generalised facies model, with such dispersal framework in the Damodar Valley basins, consists of prominent coarse basin marginal facies of channel sandstone with occasional pebble-beds in the Barakar and Karharbari formations which were deposited by the meandering channels of radial drainage pattern. There is a progressive increase in the proportion of finer-grained sediments towards the centre of the basin. An analysis of the variation of sandstone-shale ratio of the Barakar Formation in the Jharia Coalfield, Bihar (Fig. 17.1), likewise shows a regular decrease in the sandstone-shale ratio towards the basin centre (N. Sen-gupta, pers. comm.). Moreover, thick sequences of top stratum sediments including siltstones, shales and coal seams have been recorded in the Karharbari, Barakar and Raniganj Formations in the Damodar Valley Basins. Evidently, these basins have experienced repeated subsidence so that the thick pile of finer argillaceous members and coal seams could be buried before they could be removed by erosion and scouring by the migrating river channels which separated the overlying sandstones.

In contrast, in the outlying area of the Son, Mahanadi, Pranhita-Godavari and Narmada Valley basin-belts, the palaeocurrent data of the coal measures show a uniform northwesterly direction of sediment transport (Ghosh and Mitra, 1970). The overall coarse-grained nature of the sediments, and the unimodality of the dispersal pattern, suggest that the sediments were deposited in braided streams, flowing along the northwesterly palaeoslope in these extensive basin-belts. Moreover, the dominance of coarse sandstones in the sedimentary cycles of these areas, and the paucity of top stratum deposits, further indicate that in these basin-belts catastrophic change of course of the river system took place frequently, and the regime was characterised by quick reoccupation of their former positions by the channel belts. In such a sedimentary organisation, the records of the deposition of the overbank sediments are frequently erased. In Lakhanpur and Sonhat Coalfields the borehole data indicate that the coal seams and the associated argillaceous sediments are frequently completely scoured by a thick unit of multistorey sandstones. Evidently the subsidence did not keep pace with sedimentation in these basin-belts, and ingress of coarse clastic detritus was more than could be trapped by regional subsidence. This imbalance between the supply and accumulation of sediments resulted in the formation of a number of multistorey sandstone units with consequent removal of coal seams and associated argillaceous beds.

MODE OF DEPOSITION OF COAL SEAMS

The Gondwana coals of Peninsular India were generally believed to be of drifted origin. The main arguments favouring this view include (a) pronounced regional variation in thickness and quality of the seams; (b) high ash content of most of the seams; (c) lateral transition of coal to carbonaceous shales; (d) absence of seat-earth and upright stems in the underclays; (e) scarcity of plant leaf impressions in the roof of coal seams; and (f) bifurcation and branching of coal seams. Using the above observations, Fox (1934) and Gee (1932) suggested that the coal accumulated as drifted vegetation in the flood-basin swamps and in abandoned channels of aggrading rivers. Abundant impurities in the coal seams resulted from the periodic influx of suspended sediments during the periods of flood.

The accumulating new finds in the different coalfields call for a reappraisal of the classical ‘drift theory’ of the origin of Gondwana coal. Niyogi (1966) first reported the occurrence of upright stems and roots below the coal seams of the Saharjuri Basin, Bihar. Based on this finding, as well as on the constancy of thickness and the sheet-like nature of the coal seams, Niyogi postulated that these coals originated by in situ growth from decayed vegetation. Of late, Chandra (pers. comm.) has recorded rootlets in the floor sediments of coal seams in the Damodar Valley as well as in the outlying basins. These observations from isolated areas, however, do not convincingly demonstrate the in situ
origin of all Indian coal. 'Seat earths' with rootlets have been recorded from a few places only, and there are also a number of instances, particularly from Son Valley Coalfields, where the coal seams rest directly on arkosic sandstones without any recognisable traces of rootlets. Moreover, the individual seams show pronounced variability due to lateral and vertical gradations of coal into shaly coal or coaly shale. The thickness of clean coal may vary remarkably over short distances as in Seam-I of the North Karanpura Basin in which the clean coal thickness varies from 9 to 13 m out of a total section of 20 to 25 m.

In addition, the interaction between fluvial sedimentation and peat deposition caused frequent splitting and digitation of coal seams, which is well documented in the case of the thick Kargali seam in the Bokaro Coalfield. This seam, which is a composite 30 m unit in the central part, splits into two to four seams in the western and eastern parts of the same basin. Such splitting indicates proximity to the source of clastic detritus. However, on a regional scale some seams remain fairly constant in thickness as in the case of Turra Seam in Singrauli Coalfield, though the thickness of the clean coal may vary to some extent.

The conclusion from all these observations is that the Gondwana coals in Peninsular India generally show evidence of formation in both allochthonous and autochthonous conditions. It is, therefore, reasonable to infer that these coal seams were of hypautochthonous origin which involved very limited transportation of vegetation within the general area of their growth (Hacquebard and Donaldson, 1964). Although little transportation and mixing of plant debris under hypautochthonous conditions is necessary, the process does not involve rafting of a great mass of vegetal debris from the surrounding upland into the coal swamps as envisaged earlier.

The generalised depositional model for Lower Gondwana coal swamps can be reconstructed as follows. Extensive swamps were established on the flood-basins and these were sufficiently remote from the active river channels to preclude any large influx of detrital sediments. The cut-off of clastic sediments was effected by climatic changes, thick vegetation cover in source areas, tectonic movements in the provenance areas, or possibly changes in sites of deposition. In such completely enclosed basins with no outlets for drainage, the peat accumulated under hypautochthonous conditions due to mixing of plant debris by intermittent action of running water. Small streamlets originating from the surrounding uplands debouched into the central swamp. In such a geomorphic setting, coarser clastic detritus brought by the streams was laid down near the swamp margin, whereas finer clastics were carried in suspension towards the centre of the basin. Under such conditions relatively superior quality coal seams could sometimes form near the basin centre, and there was deterioration of quality near the basin periphery. The iso-ash map of the Dishergarh seam in the Raniganj Coalfield (Fig. 17.2) demonstrates this point well. Likewise the VII and VI seams are of superior quality in the centre of the basin near Ghato in the West Bokaro Basin.

**GEOCHEMICAL EVOLUTION OF LOWER GONDWANA COAL**

In spite of the greatly increased accumulation of data on the physical and chemical properties of the Gondwana coals of India, the mechanism of geochemical evolution is candidly admitted by most authorities to be poorly known. Most of our present knowledge results from the pioneering work carried out in the laboratories of the Central Fuel Research Institute in India. This has been discussed by Lahiri and Choudhury (1968) and Lahiri (1969).

The process of coal formation usually involves two stages. The first is the 'diagenetic' stage in which biochemical agencies transform vegetable matter to coal, the process being arrested to some point. The second stage may be referred as 'dynamochemical' during which \( \text{CH}_4, \text{CO}_2 \) and \( \text{H}_2\text{O} \) are eliminated leading to growth of condensed ring structures.

It is postulated that the major reactions in normal coalification possibly involve decarboxylation and dehydroxylation from the
brown coal stage of 60-62 per cent carbon to the 84-85 per cent carbon level in the bituminous range. Some high-moisture, high-volatile, low-rank coal such as that in the Dishergarh, Burradhemo, Hirakhun seams of the Raniganj Coalfield, have undergone significant structural changes during the course of this metamorphism. The intensity of the process increased with depth as is evidenced by carbon enrichment, which shows an increase of 0.5 to 0.8 units per 100 m of vertical depth (Mukherjee et al., 1972). Deoxygenation appears to be the chief mechanism in the formation of coals of sub-bituminous ranks. The evolution of bituminous coal involves higher energy levels, and appears to be a function of condensation through dehydroxylation and progressive conversion of hydroaromatic carbon to aromatic. Dehydrogenation appears to set in soon after the 85 per cent carbon level is reached.

In India, coals with coking characters are primarily confined to the Jharia, Raniganj, Bokaro, Karanpura and Ramgarh Basins of
the Damodar Valley, and to the Kanhan Valley area of Satpura Basin. Evidently such higher rank coals are known to occur either in association with highly tectonically disturbed areas such as the Kanhan Valley where a number of subparallel strike faults have affected the coal measures, or in the neighbourhood of igneous intrusions as in the Damodar Valley Coalfields. The igneous intrusives in the Raniganj, Jharia and Bokaro Coalfields are mostly represented by dykes and sills of lamprophyre intrusives. These probably radiate from the Rajmahal Hills—a major centre of plateau basalt eruption. Similar lamprophyre intrusives have also been noted in the Himalayan foothill belt of Darjeeling to the northeast of the Rajmahal Hills. The frequency of the lamprophyre bodies also progressively decreases towards the west in the Damodar Valley, showing that they had their origin somewhere to the east of the Damodar Valley Basins. It has been suggested (Lahiri, 1969) that in the Jharia Coalfield, which is the major source of prime coking coal in the country, regional metamorphism was induced by magmatic heat of the lamprophyre bodies and this has led to rapid devolatilisation and increase of rank of coal from 87.5 per cent to 93.4 per cent carbon level. In Jharia Coalfield, coals were evidently ‘soaked’ under a higher geothermal gradient. This stage of metamorphism, where most bituminous caking and coking coals occur, presumably involves dehydrogenation of alicyclic structures and simultaneous ring growth through dehydroxylation of the aromatics (Lahiri and Choudhury, 1968). It has been concluded that although several distinct steps of metamorphism in the evolution of coal are there, two major trends in metamorphism of coal are involved, (a) deoxygenation between 60 and 85 per cent carbon and (b) dehydrogenation through total transformation of aliphatic structures to aromatic and multiple condensation of aromatic groups through dehydroxylation.

Besides the dynamochemical factors like pressure and temperature resulting from igneous activity and tectonic disturbances which were instrumental in increasing the rank of coal, some biochemical factors may independently be responsible for variation of rank. An example is the Dishergarh seam in the Raniganj Coalfield where there is an increase in rank in the western side of the basin accompanied by an increase of calorific value. The calorific value of the Dishergarh seam in the eastern sector is 7685-7775 Kcal/kg, whereas the corresponding figure in the western sector is 8215-8260 Kcal/kg. Since the field has a uniform tectonic setting and a uniform distribution of intrusive bodies in both the eastern and western halves, the variation in rank has been ascribed to biochemical factors. It has been assumed (Ghosh et al., 1966) that in the western part the basin was closed and deeper, and there the seam attained progressively a greater thickness than in the central part of the swamp (Fig. 17.3). Biochemical reactions took place in this area in more reducing conditions, whereas in the eastern side the depth of water standing over the mother coal was less, or there was more access to fresh water which made the environment more oxygenated, thereby reducing the rank of coal.

A complex interplay of biochemical and dynamochemical actions took place in the various peninsular Gondwana basins as plant debris of hypautochthonous origin was progressively transformed into coal of varying rank depending on the extent of igneous activity, and the geomorphologic and tectonic settings of the different basin belts.

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Fig. 17.3. Sketch map showing isopach line of Dishergarh seam, western sector, Raniganj Coalfield

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Environmental Interpretations of Gondwana Coal Measure Sequences in the Sydney Basin of New South Wales

R. A. BRITTEN, M. SMYTH, A. J. R. BENNETT AND M. SHIBAOKA

ABSTRACT

Gondwana coal measure sequences of the Sydney Basin are characterised by the structural features and differing coal types which occur within each sequence. The nature of coal splitting is an important environmental indicator in the coal measure sequences. In the Singleton Coal Measure the coal seams frequently develop 'zig-zag' splits in a process of multidirectional splitting, indicating an environment where sediments were emplaced from river channels switching back and forth across the peat swamps. Seam splitting in the Greta Coal Measures is often unidirectional and the splits diverge progressively in one direction towards a subsiding trough zone. In the Illawarra Coal Measures there is almost complete absence of splitting, indicating much slower and more stable subsidence than in the two above environments.

The petrographic compositions of the coals from these coal measure sequences fall into two main groups: coals from the Singleton and Greta Coal Measures are rich in vitrite plus clarite; those from the Illawarra Coal Measures are poor in vitrite plus clarite.

These environmental interpretations are supported by statistical analyses of the transitions between the sediments and by a detailed study of an individual seam. Contemporary tectonism influenced coal measure deposition; thicker sedimentary sequences containing relatively more coal seams occur in synclinal areas.

INTRODUCTION

Gondwana coals have generally been considered to be different from Carboniferous coals in the northern hemisphere; however, both of these coal types vary widely, and it is expected that similar coals would result from similar environments, whether they are northern or Gondwana.

Wide differences of coal type occur throughout the Sydney Basin, and the purpose of this paper is to explain these differences with reference to the depositional environments associated with different seams and different coal measure sequences. An interpretation of significant depositional environments has therefore been made for a number of important Permian coal measure sequences within the Sydney Basin of New South Wales.

Each of the coal measure sequences is characterised by environmental and related structural features which may be interpreted by considering the nature of seam splitting, the type of coal seams they contain, the nature and frequency of the transitions between lithologies, and the pattern of total sedimentation within certain of the sequences.
Fig. 18.1. Location and distribution of some Gondwana coal measure sequences in the Sydney Basin of N.S.W.

**GEOLOGICAL SETTING**

The Sydney Basin is, economically, one of the most important coalfields in Australia. It is situated in a foredeep, or sediment-filled linear downwarp separating an old land mass to the west from a rising orogenic belt to the east, and the coal seams occur mainly in Permian sediments. The Sydney Basin is a structural basin, with a central cover of Triassic rocks surrounded by outcropping Permian rocks on the north, west and south as indicated in Figure 18.1. The depositional basin apparently formed as an asymmetric depression, thin sediments being deposited on the western and southern margins bordering the basement rocks, while thick sediments were deposited against the mobile New England region to the north-east.

The sequences selected include the Greta Coal Measures, which have the stratigraphic status of a group, the Singleton Super-Group (also known as the Singleton Coal Measures), and the Illawarra Coal Measures, which is also a group. The Greta Coal Measures and the Singleton Super-Group are located in the Hunter Valley near the northeast margin of the Sydney Basin, whilst the Illawarra Coal Measures are located toward...
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Gondwana Coal Measure Sequences in the Sydney Basin 235

NARRABEEN GROUP

MUSWELLBROOK ANTICLINE

LODER - CAMBERWELL ANTICLINE COMPLEX

HUNTER THRUST LINEAMENT

GREGA COAL MEASURES

DALWOOD GROUP (MARINE)

MAITLAND GROUP (MARINE)

SINGLETON SUPER GROUP

GRETA COAL MEASURES

Fig. 18.2. Generalised section east-west through the Hunter Valley region of N.S.W. showing various coal measure sequences present

and lower boundaries correlate with those of the Singleton Coal Measures (1400 m) in the north, and the intervening strata are equivalent. At the top of the Illawarra Coal Measures lies the Bulli Coal seam, from which the structure of the measures has largely been determined.

ENVIROMENTAL INTERPRETATIONS OF SOME TYPICAL SYDNEY BASIN COAL MEASURE SEQUENCES

The environmental differences apparent within and between the major coal measure sequences of the Sydney Basin have, in some instances, resulted from contemporaneous tectonic movements. These may have been widespread, due to regional subsidence or tilting about a hinge line, or in some cases the effect of the movements may have been of minor extent, due to a localised compaction. Except in the rare instance of seams that are entirely allochthonous, the coal forming environments compete with the depositional environments of the intercoal sediments. Wherever the water level and other factors are suitable for the accumulation of peat, the growth of peat will transgress any underlying depositional environment. This interaction between coal forming and inorganic sedimentary environments can be seen from the way splits are formed in the sequence. The nature of coal splitting is therefore an important environmental indicator, examples of which are briefly outlined.

Seam Splitting within some Gondwana Coal Measure Sequences

The coal seams within the Singleton, Greta and Illawarra Coal Measures each show a

Fig. 18.3. Diagrammatic illustration of seam splitting characteristic for some coal measure sequences of the Sydney Basin
characteristic type of splitting.

In the Singleton Coal Measures the coal seams frequently develop zig-zag splits in which the splits join overlying and underlying coal beds in a process of multi-directional splitting.

Seam splitting in the Greta Coal Measures is often unidirectional and the splits diverge progressively in one direction towards a zone which is subsiding more rapidly than that required for the continuous accumulation of peat.

In the Illawarra Coal Measures there is almost a complete absence of splitting and the interseam strata are disposed in more or less parallel beds between the various seams.

These modes of splitting are illustrated diagrammatically in Figure 18.3.

Zig-Zag Splitting in the Singleton Coal Measures

The zig-zag splitting in the Singleton Coal Measures indicates a rapid and frequent oscillation between the accumulation of organic peat and that of the mechanically deposited inorganic sediments.

Plate 18.1 illustrates a zig-zag seam split and shows the typical development of a significant part of the associated structural compaction near the split junctions.

In the vicinity of the lowermost leaf of a zig-zag split the structure will be characterised by three features.

1. The stratum in contact with the roof of the coal seam meets it tangentially but curves upwards away from the coal roof towards the point of initial compaction.
2. The bedding of the strata overlying the coal seams becomes bent downwards away from the point of initial compaction, accompanied by a certain amount of shearing in the beds, which is most intense in the vicinity of the seam and split junctions.
3. The angle of inclination of the beds above the coal seam increases relative to the coal as the distance of the bed from the coal increases.

The splits often converge or diverge from adjacent coal beds at angles of up to $45^\circ$; but because of the necessity to accept that
each peat bed is laid down horizontally in a coal swamp, an unusual sequence of events has to be invoked to explain how the seams and splits have come to be so steeply inclined to each other. This sequence of events and their explanation, as far as is known, have not been described for Gondwana coal measure sequences outside Australia.

The processes causing this multi-directional seam splitting are explained and illustrated in Figure 18.4. In this reconstruction of the depositional events three important features stand out.

1. The peat environment tends to persist throughout the depositional basin of the coal measures.
2. Interruptions to the prolonged formation of peat are intermittent, localised, and occur within a very short time compared with the time occupied by accumulation of peat.
3. The thick peat beds absorb the interruptions locally through their capacity for compaction.

This is illustrated in Figure 18.5 which shows in detail the early steps (Steps 1 to 3 of Figure 18.4) in the development of what is here called a mobile compaction structure, i.e. the compacted zone in the peat bed moves outwards. It can be seen that the cross-section of the beds tends to be thicker at their centre; and that in the early stages of emplacement these beds sag increasingly at their centre, pressing deeper and deeper into the compacting peat, because of their own weight.

The mobile compaction structure is typically developed through a vertical interval of about 30 m and a lateral interval of between 0.1 and 10 km. This lateral interval will be governed by the distance between the axes of the wide shallow channels from which the sedimentary beds are deposited.

Figures 18.4 and 18.5 show an idealised mobile compaction structure which is symmetrical, but the actual structures will at best approximate such models.

It should be emphasised that it is the compactability of the peat beds which produces the characteristic sedimentary structure.

Relationship Between Seam Splitting and Subsidence

The multi-directional splitting in the Singleton Coal Measures and the unidirectional splitting in the Greta Coal Measures are both indicative of rapid accumulation and burial of the coal. The former occurred in an environment where the sediments interrupting the peat were deposited from channels and their overbank floods which were switched back and forth across the peat swamps. The latter occurred in an environment where the channels and their associated structures migrated in one direction towards a subsiding trough zone.

By way of contrast, the virtual absence of splitting in the Illawarra Coal Measures is indicative of greater stability. The peat accumulation and subsidence took place slowly and uniformly throughout a broad tranquil area that probably had low relief.

This stability is particularly apparent from a detailed section of the Bulli Seam, which has been extensively worked and is well known. This section is depicted in Figure 18.6. It illustrates the consistency of the seam over a distance of 50 km and shows how even individual plies and the alternations of dull and bright coal in the seam can be recognised extensively. The stability and consistency of the coal and interseam strata are interpreted as reflecting their association with widespread, low energy depositional environments close to sea level.

These interpretations are further supported from the coal types in the sequences and from a statistical analysis of the transitions occurring in the sequences.

Relationship of Coal Type and Environment

Many of the seams within the coal measures have been analysed petrographically and their petrographic composition is shown in Figure 18.7. In this figure the seams are identified with a particular sequence and also as individual seams. Three very significant features stand out.

1. The coals fall into two main groups; one has coals rich, and the other depleted in vitrite and clarite.
2. Of the twelve coals from the north of the Sydney Basin, which include practically
A thick peat bed, A, accumulates and is subjected to its first incipient sedimentation at P & Q.

Peat bed A yields by compaction to successive sedimentation at P & Q.

Peat bed A reaches its limit of compaction locally in response to sedimentation at P & Q.

Peat growth is resumed and peat bed B is emplaced.

Incipient sedimentation starts at R, S & T in zones of greatest compactive response. This has resulted in the channels repositioning themselves.

Sedimentation at R, S & T proceeds to the limit of compaction locally levelling peat bed A and bending peat bed B. The sediments deposited within the beds P & Q.

Peat growth resumes and peat bed C is emplaced.

Coal measures or other sedimentation proceeds steadily increasing load and general compaction.

Ultimate compaction of peat 10:1, inter seam sediment 3:2, together with coalification, yields seam splits and mobile compaction structure.
Fig. 18.4. Processes of multi-directional seam splitting with the development of a typical mobile compaction structure

BED 1 EMPLACED

BED 1 SAGS AT ITS CENTRE CREATING SPACE FOR BED 2

BED 2 EMPLACED

BED 2 SAGS AT ITS CENTRE BED 1 SAGS EVEN MORE SPACE IS CREATED FOR BED 3

BED 3 EMPLACED

PROCESS CONTINUES

LATER SURFACE XYZ IS RESTORED TO HORIZONTAL AND SURFACE X O Z IS BENT IN THE PROCESS.

Fig. 18.5. Details of emplacement and later distortion of sedimentary beds during the development of a mobile compaction structure

Fig. 18.6. Section of the Bulli Seam across the southern part of the Sydney Basin showing the constancy of plies within the seam

all the seams from sequences A and B of the Singleton Coal Measures and those of the Greta Coal Measures, only two fall into the vitrite-clarite-depleted group and these occur adjacent to a widespread environmental change of exceptional extent and persistence.

3. Of the five coals from the south of the Sydney Basin all except one are in the group depleted in vitrite and clarite.

In the north of the Sydney Basin it is clear that the vitrite and clarite rich peats within the coal measure sequences were buried rapidly and preserved quickly (Mackowsky, 1968).

The exceptionally low vitrite content of the Bayswater seam has been explained in terms of the influence of the underlying environments (Britten and Smyth, in press). Evidence of bioturbidity and the presence of nodular pyrite suggests deposition in a widespread marine embayment, just prior to the emplacement of this seam. After closure of the embayment the environment became lagoonal. Later, in response to eustatic changes, a widespread sand facies was deposited in a lacustrine environment, and this sand formed the stable, widespread foundation of the dull, vitrite-depleted, Bayswater seam. The Ravensworth seam is less depleted in vitrite than the Bayswater seam. It lies above the Bayswater seam and its petrographic composition is intermediate between that of the Bayswater seam and the vitrite-rich seams that comprise most of the other seams of the coal measure sequences in the

SW

ROOF

FLOOR

0 1 2 3 4 5 MILES
0 2 4 6 8 KILOMETRES

NE

FT M
10 3
7 2
3 1
0 0

POOR RICH IN VITRINITE
north of the Sydney Basin.

In the south of the Sydney Basin the coals are depleted in vitrite and are associated with stable depositional conditions linked with slow uniform subsidence throughout widespread areas. In these environments the relative lack of vitrite and clarite is ascribed to a greater degree of degradation of the peat due to slow burial and preservation. The Wongawilli seam, which is richer in vitrite and clarite than the other coal seams of the Illawarra Coal Measures, is also very thick and banded. This indicates that it accumulated during a regionally increased rate of subsidence, with more rapid burial and preservation of the coal-forming peat.

Environmental conditions proposed for the deposition of the Illawarra Coal Measures in the south of the Sydney Basin include both lower fluvial and deltaic facies (Bunny, 1972) and they comprise a lower energy depositional regime than is generally found in the northern sequences. Although the environments of coal deposition in the north are essentially different from those in the south, transitions can be recognised throughout the Sydney Basin (Stuntz, 1972). The comparison between the Singleton and Illawarra Coal Measures given in Figure 18.8 illustrates many of the known environmental differences.

**Relationship between Lithological Transitions and Coal Measure Environments**

Both of these coal measures contain fluvial components, but a statistical analysis of the transitions between lithologies, revealed by a typical vertical section selected for each meas-
Fig. 18.8. Diagrammatic comparison between the principal environments interpreted for the Singleton Coal Measures of the Sydney Basin of N.S.W.
Fig. 18.9. Isopachs of total sedimentary component. Sequence A—Singleton Super-Group.
tation, no significant second preference transitions emerged from the analysis of the Illawarra Coal Measure sequence. This showed that the environments in these measures were independent of each other and that the existence of the peat forming environment tended to exclude the existence of inorganic deposition, and vice versa, with no continuous switching between the two. This supports the idea of parallel, independent, stable environments for the Illawarra Coal Measures.

**Tectonic Influence on Coal Measure Sedimentation**

The Singleton and Greta Coal Measures, near the northern margin of the Sydney Basin, both bear the imprint of local contemporaneous tectonism, which is shown to be an early manifestation of the existing meridional structures of the Hunter Valley region. These structures include the Muswellbrook Anticline, the Loder Dome, the Belford Dome and the Lochinvar Anticline (Fig. 18.9).

The Greta Coal Measures are well known in small, localised zones around the Muswellbrook Anticline and the Lochinvar Dome. The data available from these areas indicate that the coal seams have split away from the crestal zones of the present anticlines, although during the time of peat deposition these zones were subsiding slowly to yield the thick peat beds which later became coal seams. This splitting of the seams from the crestal zones indicates a more rapid subsidence of trough zones, located towards the flanks of the anticlines which are now prominent structural features.

The Singleton Super-Group has been explored more extensively than the Greta Coal Measures, and this allows an interpretation of some gross depositional features of the complete coal measure sequences A and B of Figures 18.1 and 18.2.

Sequences A and B are bounded at their top and base by non-coal measure transitional environments, so that the relationship of coal and sedimentary components of these sequences can be determined. Isopachs of the total sedimentary component of the sequences A and B are depicted on Figures 18.9 and 18.11 respectively, and the ratio lines of total sediment to coal are given in Figures 18.10 and 18.12. These figures reveal the following features:

1. Sequence A thins southward and thickens towards the Hunter Thrust Lineament beyond which it may have attained its maximum development, and the zones containing the greatest accumulation of sediment occur between major anticlines (Fig. 18.9).

2. The ratio of sedimentary component to coal for Sequence A is greater in the vicinity of anticlinal structures (Fig. 18.10).

3. Sequence B has its maximum development in a zone between the projections of major anticlines, and its known development suggests the form of an elongated meridional lobe thinning towards the north (Fig. 18.11).

4. The ratio of sedimentary component to coal for Sequence B is greatest in a zone towards the southern extension of the Muswellbrook Anticline and in a zone to the east, which may reflect an anticlinal structure so far undisclosed (Fig. 18.12).

These features indicate the influence of contemporaneous tectonism in the vicinity of the known anticlines during the deposition of both Greta and Singleton Coal Measures.

**Conclusions**

The study of some Gondwana coal measure sequences occurring in the Sydney Basin of New South Wales has disclosed environmental differences that can account for differences in coal types and in the dispositional of the seams.

Sequences in the north of the basin are thick and contain numerous vitrite and clarite-rich coals, which are often characterised by erratic zig-zag splitting of thick coal seams and the presence of structures, likely to be unique to coal measure sequences, called mobile compaction structures. These structures may also be unusual in Gondwana coal measures elsewhere, no previous records of them having been noted outside Australia.

In the north of the Sydney Basin subsidence was relatively rapid, particularly in trough zones between existing anticlines, and thick sedimentary sequences, sometimes containing coarse sediments, were deposited. The streams were carrying much material, at an
Fig. 18.10. Zones of high sediment/coal ratios. Sequence A—Singleton Super-Group.
Fig. 18.11. Isopachs of total sedimentary component. Sequence B—Singleton Super-Group.
Fig. 18.12. Zones of high sediment/coal ratios. Sequence B—Singleton Super-Group.
energy level sufficient to cause the frequent incision of meanders and/or channel switching which in turn contributed to the frequent splitting and coalescing of the coal seams.

Sequences in the south of the Sydney Basin are thinner, and contain predominantly vitrite-poor coals, emplaced during stable conditions and characterised by the widespread constancy of the contained coal seams.

Exceptions to the general characteristics in both the northern and southern coal measure sequences are explained by recognition of a different environmental setting for the exceptions in each sequence.

A simple cause of the difference in conditions to the north and south of the Sydney Basin is likely to be their positions with respect to the active New England Fold Belt. Proximity to this fold belt is associated with higher energy environments that include fluvial facies with a heavy sedimentary load. To the south the influence of the belt was much less, and mechanical deposition was in a lower energy domain never far removed from the paludal environment of the widespread swamps and marshes in which the peat beds accumulated.

ACKNOWLEDGMENTS

Acknowledgments are due to the Joint Coal Board for permission to publish and to the staff of Clutha Development Pty Ltd for their help and co-operation in obtaining the illustration presented as Plate 18.1.

REFERENCES


ABSTRACT

The post-Katangan sequence in the mid-Zambezi Valley consists of the sediments of the Sinakumbe Group, Lower Karroo, Upper Karroo and Tertiary-Quaternary. The Lower Karroo rocks consist of the Siankondobo Sandstone Formation, the Gwembe Coal Formation and the Madumabisa Mudstone Formation. The base of the Gwembe Coal Formation contains a relatively thick (5-10 m) coal seam, the Main Seam. A number of minor seams, generally less than half a metre thick, occur in the upper parts of the formation. The Zambian coal is similar to other Gondwana coals except for the high inertinite maceral content of over 90 per cent. It is also high in ash and sulphur. The Main Seam, which rests on an irregularly bedded carbonaceous sandstone—Maamba Sandstone—is variably developed; it thins down-dip, is uniform laterally over long distances, contains wash-outs and elongated inter-seam sandstone bodies. The general linear distribution of the ash values is independent of seam thickness, but largely conforms to the inter-seam sandstone geometry, pointing to well-defined seepage channels for the source of silt and clay, whereas the sandstone bodies represent channel infillings. Although there is no unequivocal evidence for an autochthonous origin, such as rootlet underclay horizon or upright stems in position of growth, most of the characteristics of the Main Seam point to an in situ rather than a drift origin. The Main Seam at the base was essentially autochthonous, but the minor seams above were largely hypautochthonous. The sedimentary characteristics of the Main Seam, coupled with its petrographic peculiarity, indicate an autochthonous origin in a floodplain with a low water-table, probably along the margins of a series of northeasterly trending echelon bodies of water, which probably originated during the Dwyka glaciation which affected the region prior to the deposition of the coal. The separate bodies of water subsequently became larger and linked together to form the Madumabisa Mudstone lake subsequent to the deposition of the Gwembe Coal Formation.

GEOLOGY

The mid-Zambezi Valley of Zambia is floored by Karroo and older sediments resting upon the crystalline basement; Table 19.1 summarises the stratigraphy of the region. The basement geology of the area is being described by Matheson (in prep.), and consists predominantly of schists and gneisses, probably older than 1000 m.y., which have been intruded by a variety of granitic rocks ranging in age between 600 and 1000 m.y. The oldest unmetamorphosed sediments are the well-lithified, brightly coloured, arenaceous sequence of the Sinakumbe Group, representing deposits along a continental margin. They are correlated with the Sijarira Group of Rhodesia and are probably early Palaeozoic in age (Denman and Money, 1970).
<table>
<thead>
<tr>
<th>System</th>
<th>Series</th>
<th>Group/Formation</th>
<th>Lithology/Subdivision</th>
<th>Thickness (in metres)</th>
<th>Environment</th>
<th>Period</th>
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</thead>
<tbody>
<tr>
<td>Recent</td>
<td>Tertiary</td>
<td>Gondwana Coal Deposits</td>
<td></td>
<td></td>
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<td></td>
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<td>UPPER</td>
<td>STORMBERG</td>
<td>Batoka Basalt</td>
<td>Flows/Pyroclastics</td>
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<td>Volcanic</td>
<td>Lower Jurassic (Lias)</td>
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<td></td>
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<td>640</td>
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<td></td>
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<tr>
<td>LOWER</td>
<td>ECCA</td>
<td>Sandstone E</td>
<td></td>
<td>0 - 8</td>
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<td>GWEMBE COAL</td>
<td>Carbonaceous Mudstone</td>
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<td>43</td>
<td>Flood Plain</td>
<td>Permian</td>
</tr>
<tr>
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<td></td>
<td>0 - 9</td>
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<td></td>
<td>37</td>
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<td></td>
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<td></td>
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<td>25 - 43</td>
<td>Flood Plain</td>
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<td></td>
<td>Sandstone B</td>
<td></td>
<td>1 - 8</td>
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<tr>
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<td></td>
<td>40 - 51</td>
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<td></td>
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<td></td>
<td>0 - 30</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Main Seam</td>
<td></td>
<td>0 - 14.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Maamba Sandstone</td>
<td></td>
<td>0.3 - 15</td>
<td></td>
<td></td>
</tr>
<tr>
<td>KARROO</td>
<td>ECCA-DWYKA</td>
<td>Siankondobo Sandstone</td>
<td>Conglomerate/Mixtite/</td>
<td>0 - 90</td>
<td>Fluvio-Glacial</td>
<td>Permo-Carboniferous</td>
</tr>
<tr>
<td></td>
<td>PRE-DWYKA</td>
<td>Sinakumbe Group</td>
<td>Sandstone/Quartzite</td>
<td>0 - 61</td>
<td>Shelf</td>
<td>Devonian</td>
</tr>
<tr>
<td></td>
<td>BASEMENT</td>
<td></td>
<td></td>
<td></td>
<td>Continental margin</td>
<td>Ordovician</td>
</tr>
<tr>
<td></td>
<td>COMPLEX</td>
<td></td>
<td></td>
<td></td>
<td>Metamorphic</td>
<td>Precambrian</td>
</tr>
</tbody>
</table>
The Sinakumbe rocks are overlain unconformably by the Karroo succession, which is subdivided into lower and upper parts. The distribution of Karroo rocks in the mid-Zambezi Valley, the coalfields and other localities mentioned in this paper is shown in Figure 19.1. The Lower Karroo rocks, consisting of the Siankondobo Sandstone, the Gwembe Coal Formation and the Madumabisa Mudstone, outcrop along the western
margin of the Zambezi trough, where the beds are tilted and faulted against the basement. The Upper Karroo rocks, consisting of the Escarpment Grit and Upper Sandstone, occupy the central and eastern parts of the valley, and the Batoka Basalt occurs to the southwest.

The Siankondobo Sandstone, the oldest Karroo formation in the area, consists predominantly of arenaceous, non-carbonaceous beds resting unconformably on the basement or on the Sinakumbe Group. It is overlain unconformably by the dominantly carbonaceous beds of the Gwembe Coal Formation.

Fig. 19.2. Lower Karroo Succession (Mid-Zambezi Valley)
It has an average thickness of 10-20 m and consists of ill-sorted conglomerates, mixtites and sandstones of glacial and fluvo-glacial origin, deposited on an irregular glaciated land surface of pre-Karroo rocks. The greater part of the formation consists of fine-grained buff to white, uniform, horizontal or cross-laminated silty sandstone, although in places the beds are massive. It is devoid of fossils, except for a few burrows in the upper parts. The formation is correlated with the Lower Wankie Sandstone of Rhodesia, which is thought to be Dwyka-Ecca in age (Money, Denman and Drysdall, in prep.; Bond, 1967; Watson, 1960).

The Gwembe Goal Formation, which overlies the Siankondobo Sandstone Formation unconformably, is a succession of carbonaceous sandstones and siltstones with interbedded coal seams and sandstones. It appears to attain its maximum development in the Siankondobo Coalfield, where it is up to 280 m thick. The succession in the southwest and northeast of the Siankondobo coalfield and in the Mulungwa region is shown in Figure 19.2; it illustrates the importance of Sandstone A and the impersistence of the other arenaceous horizons. The formation thins southwestwards and is between 100-150 m thick in the Mulungwa area.

The Maamba Sandstone is the basal member of the formation and is immediately overlain by the Main Seam. Minor coal seams, generally less than 25 cm and rarely more than a metre thick, are developed at higher horizons in the lower part of the formation.

In the southwestern part of the Siankondobo region the Gwembe Coal Formation contains up to six sandstone horizons, including the Maamba Sandstone and the Gwembe Sandstones A to E (Fig. 19.2). In the northwest the Gwembe Sandstones are absent, with the local exception of Sandstone A, and the sequence is one of carbonaceous mudstones and siltstones. The carbonaceous beds pass conformably upwards into the non-carbonaceous Madumabisa Mudstone through a transitional zone—the Izuma Beds (Radosevic, Money and Denman, 1968). Sandstone A persists to the southwest and is present throughout the Mulungwa and Maze areas.

None of the Gwembe Sandstones are present to the north in the Nkandabwe area.

Plant remains occur throughout the formation, but are for the most part fragmentary, and are seldom identifiable. The formation is generally regarded as mainly Middle and Upper Ecca in age, with the possible exception of beds above Sandstone D, in which freshwater molluscs of probable lowermost Beaufort age occur. Recent palynological studies (Utting, 1970) indicate the dominance of trilete spores in the Maamba Sandstone and an abundance of disaccate (non-striate) pollen and trilete spores in the overlying coal formation beds.

The Madumabisa Mudstone, consisting of a thick sequence of non-carbonaceous, grey-green mudstone, with occasional calcilutite horizons which are fossiliferous in places, succeeds the Izuma Beds paraconformably. It is considered to be Lower Beaufort in age.

This is overlain in turn by the Escarpment Grit and by an undifferentiated sequence of arenaceous sediments. The total thickness probably exceeds 2300 m, and the succession consists of red, buff and pink cross-bedded feldspathic to arkosic grits and sandstones of continental origin. The red beds are overlain by pyroclastics and lava of the Batoka Basalt Formation, which is probably 500 m thick (Matheson, in prep.; Gair, 1959; Money, in press).

Tertiary conglomerates, sandstones and gravels overlie the basalts.

**GWEMBE COAL FORMATION**

**Main Seam**

The coal seams of the Gwembe Coal Formation may be divided into the Main Seam and the Upper Seams, the former being the only ones exploited at present.

The Main Seam usually overlies the Maamba Sandstone conformably and is generally present throughout the Siankondobo area, although it shows considerable variation in thickness and may be laterally replaced by muddy coal and coaly mudstone. In the northern part of the Kazinze area it is split by several horizons of sandstone and siltstone resembling Maamba Sandstone rock-types. The average thickness of the Main Seam is 6 m and it attains a maximum
Fig. 19.3. Siankondobo-Kazinze Region. (a) Isopach map of the Main Seam (excluding interseam Sandstone Body). (b) Isopach map of the Main Seam (excluding interseam Sandstone Body). Contour interval = 1 metre □ = Shafts ● = Boreholes
Control points (Boreholes and Shafts)

Isopachs at 1 metre intervals.
thickness of 14.6 m in the Kazinze region (Fig. 19.3). It thins down-dip, and is replaced laterally to the southwest and southeast by coaly mudstone similar to that which normally overlies the seams and into which it grades. At Nkandabwe to the north, the Main Seam ranges in thickness from 1 to 7 m, the average being 3.7 m (Taverne-Smith, 1960; Cornwall, 1965) whereas in the Mulungwa area, to the southwest of Kazinze, the Main Seam varies between 5.8 m and 9 m (Radosevic, 1968).

The bulk of the Main Seam consists of massive durain with a characteristic dull metallic lustre and a conchoidal fracture, but some horizons of finely laminated coal with a higher vitrain content are present, and interbedded lenses of sandstone occur in certain localities. In the open-pit area (Maamba Mine) cross-bedded sandstone units interdigitate with the Main Seam and as many as four divisions of the Main Seam have been mapped. The interbedded sandstones vary in thickness but are uniform in character; the thicker parts show well-developed cross-bedding and fining-upward sequences, beginning with a pebbly lag-conglomeratic base. The total thickness of sandstone present in the Main Seam decreases to the south and west (Fig. 19.4). The geometry of the sandstone units and their apparent absence further west and south in the Kazinze area are important factors for palaeogeography and provenance.

Petrography

Petrographically the Main Seam is significantly different from the other Gondwana coals. A preliminary maceral analysis of the Zambian coal from Shaft 1 (Maamba Mine) yielded 15 per cent vitrinite, 2 per cent exinite and 83 per cent inertinite. The only detailed petrographic analysis was made on samples from Shaft 3 by Alpren (1968) on behalf of Sofremines. The full details of this study, together with information on coals from Wankie, Rhodesia, and Witbank, South Africa, are presented in Table 19.2; Figure 19.5 illustrates the compositional character-istics of the Zambian coal in relation to other Gondwana coals and to the European Carboniferous coals.

In general, over 95 per cent of the Main Seam consists of inertinite (I), together with 3 per cent vitrinite (Vt) and 2 per cent exinite (E). Two bulk samples of fine washed coal of specific gravity less than 1.5, yielded 5.3 per cent and 3.8 per cent of vitrinite respectively. (The vitrinite maceral group comprises collinite and telinite, both of which have been recorded in Zambian coal, although the amount of telinite is small.) As seen in Table 19.2 and Figure 19.5, the most characteristic maceral group is inertinite, which includes micrinite, semisemifusinite, fusinite, sclerotinite, macrinite, detritoinertinite and semidetritoinertinite. The exinite maceral group comprises sporinite, cutinite, detritoeoxinite, alginite and resinoid macerals; all except the last two have been recorded from the Zambian coal. Alpren (1968) also records the microlithotypes of the Zambian coal (Table 19.3).

Within the Main Seam, argillaceous laminae, siderite specks (roses and spherulites) oxidised to limonite and to other iron oxides, pyrite and, more rarely, marcasite flowers and thin films of calcite and barytes occur on joint and bedding planes; horizons of carboniserite, mainly carbopyrite, may also be present (Money, Drysdall and Denman, in prep.). The total amounts of car-
### Table 19.2. Maceral Analysis
(Zambian coal compared to Rhodesian and South African coals)

<table>
<thead>
<tr>
<th>Maceral analysis</th>
<th>Zambia(^1)</th>
<th>Rhodesia(^2)</th>
<th>South Africa(^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Siankondobo</td>
<td>Wankie</td>
<td>Witbank</td>
</tr>
<tr>
<td></td>
<td>Shaft III</td>
<td>Sample I</td>
<td>Seam 2</td>
</tr>
<tr>
<td></td>
<td>Shaft I</td>
<td>Sample II</td>
<td>Seam 4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Seam 5</td>
<td></td>
</tr>
<tr>
<td>Vitrinite GP (vt)</td>
<td>%</td>
<td>%</td>
<td>%</td>
</tr>
<tr>
<td>Collinite</td>
<td>3.3</td>
<td>14.0</td>
<td>34.6</td>
</tr>
<tr>
<td>Telinite</td>
<td>14.0</td>
<td>54.2</td>
<td>30.2</td>
</tr>
<tr>
<td>Inertinite GP (I)</td>
<td>1.1</td>
<td>61.7</td>
<td>44.3</td>
</tr>
<tr>
<td>Micrinite</td>
<td>25.4</td>
<td>4.2</td>
<td>55.7</td>
</tr>
<tr>
<td>Semifusinite</td>
<td>79.5</td>
<td>0.4</td>
<td>2.6</td>
</tr>
<tr>
<td>Sclerotinite</td>
<td>7.0</td>
<td>2.2</td>
<td>1.5</td>
</tr>
<tr>
<td>Macrinite</td>
<td>6.0</td>
<td>7.0</td>
<td>8.2</td>
</tr>
<tr>
<td>Detritoinertinite</td>
<td>4.4</td>
<td>4.4</td>
<td>12.5</td>
</tr>
<tr>
<td>Exinite GP (E)</td>
<td>0.4</td>
<td>0.4</td>
<td>6.2</td>
</tr>
<tr>
<td>Cutinite</td>
<td>0.2</td>
<td>0.4</td>
<td>13.3</td>
</tr>
<tr>
<td>Resinite</td>
<td>2.0</td>
<td>2.0</td>
<td>12.5</td>
</tr>
<tr>
<td>Sporinite</td>
<td>0.2</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Detritoxinite</td>
<td>1.3</td>
<td>4.6</td>
<td></td>
</tr>
<tr>
<td>Alginate</td>
<td>1.3</td>
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<td></td>
</tr>
<tr>
<td>Minerals</td>
<td>3.7</td>
<td>4.5</td>
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</tr>
<tr>
<td>Pyrite</td>
<td>2.0</td>
<td>6.2</td>
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<td>Other minerals</td>
<td></td>
<td>13.3</td>
<td></td>
</tr>
<tr>
<td>Maceral Groups</td>
<td>Vitrinite (vt)</td>
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<td>34.6</td>
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<td>45.6</td>
</tr>
<tr>
<td></td>
<td>Exinite (E)</td>
<td>2.0</td>
<td>4.9</td>
</tr>
</tbody>
</table>

1 After Alpren (1968), and Sofremine Project Report (1967).
2 After Chandra and Seyler in Watson (1960). (Mineral Free basis.)

Bominerite and pyrite in a sample from Shaft 3 were approximately 7 per cent and 3 per cent respectively. Other detrital mineral matter in Zambian coal includes clay (see Table 19.3), carbonate (mainly siderite) and quartz. The clay constituents occur occasionally as horizontal partings or bands but generally they are in finely disseminated form. They occur in thin vertical desiccation cracks in vitrain bands, and in the cavities within bands and lenses of durain. Clay has been observed also in the more conspicuous cleat joints in association with carbonates—particularly barites—extending through various thicknesses of the bed. Siderite occurs as an integral part of the predominantly durain coal. It is present either as aggregates of small nodules or as thin layers and is largely syngenetic in origin.

**Chemical and Heating Characteristics**

The chemical and heating (calorific value) characteristics of the Zambian Main Seam based on over 2000 individual analyses and comparisons with the other nearby African coals are given in Table 19.4 and Figure 19.6. The most striking feature of the mid-Zambezi Main Seam is its high ash content compared with European coals, but it is even more striking when compared with other African Gondwana coals. The ash value fluctuates between 18 and 25 per cent with an average of 22 per cent for the main coalfields. The ash values from the Sian-
Table 19.3. Microlithotype Analyses of Zambian Coal
(After Alpren, 1968)

<table>
<thead>
<tr>
<th>Lithotype</th>
<th>Microlithotype</th>
<th>Percentage combination</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vitrain</td>
<td>Vitrite</td>
<td>0.9 Vt</td>
</tr>
<tr>
<td></td>
<td>Vitrinertite</td>
<td>2.3 Vt + I</td>
</tr>
<tr>
<td>Fusain</td>
<td>Microite</td>
<td>55.9 I</td>
</tr>
<tr>
<td></td>
<td>Fusite</td>
<td>13.3 I</td>
</tr>
<tr>
<td>Sporite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clarain</td>
<td>Durite</td>
<td>12.6 I + E</td>
</tr>
<tr>
<td></td>
<td>Duroclarite</td>
<td>1.1 Vt + E + I</td>
</tr>
<tr>
<td></td>
<td>Clarodurite</td>
<td>1.8 I + E + Vt</td>
</tr>
<tr>
<td>Other minerals</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Carbonminerite</td>
<td>6.9</td>
</tr>
<tr>
<td></td>
<td>Pyrite</td>
<td>3.0</td>
</tr>
<tr>
<td></td>
<td>Kaolinite</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>Carbonate</td>
<td>0.2</td>
</tr>
<tr>
<td></td>
<td>Quartz</td>
<td>0.2</td>
</tr>
</tbody>
</table>

The kondobo region are lower than those of the Nkandabwe area and there is a close relationship between ash values and specific heat. Although the regional variation is approximately 3-4 per cent there is considerable fluctuation within an area. The ash isoline map for the Kazinze area (Fig. 19.7) shows the lateral variation; the vertical variation is equally marked, as shown by values from Shafts 2 and 3 (Figs. 19.8 and 19.9).

The ash is very finely disseminated and the relatively high percentages of silica and alumina in the ash (Money et al., in prep.) suggest that it probably consists predominantly of argillaceous detritus. Apart from the finely disseminated clastic matter within the durain-fusain coal, the Main Seam also includes thin, impersistent intercalations of coaly mudstone, some of which are finely banded and laminated, containing a high proportion of vitrinite. These grade into coal of similar appearance. The high ash content, largely attributable to fine clastic detritus, precludes any substantial improvement by washing. Washed coal samples gave ash values which were 30-50 per cent lower than the raw coal, but the total ash content was still above 9 per cent. The high ash content is a function of the fine particle size of the detritus and the maceral composition, whereas the distribution of ash values is largely a result of the sedimentary environment.

The average content of volatile matter in the raw coal ranges from 17.5-22.5 per cent (Fig. 19.6), whereas the washed samples yield an average of 25.3 per cent, which would correspond to a coking coal on the European scale. A thin band of high ash/high volatile coal was encountered during mining at Nkandabwe and was thought to be a cannel type. On the basis of volatile content the Zambian coal lies on the boundary between hard coal and anthracite. However, the calorific value of a coal of this class is over 8650 cal/gm (15,500 btu/lb), which is greater than that of the Main Seam coal. The latter is a low volatile non-coking coal and its characteristics are attributable to its low vitrinite and very high inertinite.

The average fixed carbon content of the mid-Zambezi Main Seam is 55.7 per cent, which is similar to the Witbank coal of South Africa, but significantly lower than that of the Wankie coal of Rhodesia (63.8 per cent). Washed samples from the Kazinze area yielded values in the range of 85-88 per cent for fixed carbon. As might be expected, there is close correspondence between ash content, calorific value and fixed carbon. On the basis of fixed carbon the Zambian Main Seam would fall within the category of the hard bituminous, moderately compact, coals of Europe (Francis, 1961: 362), but the latter have calorific values between 8800 and 9300 cal/gm, whereas the Zambian coal has an average calorific value of only 6000 cal/gm.

The moisture content of the mid-Zambezi Main Seam is generally uniform both vertically and laterally, ranging between 1.6 and 2.3 per cent. The values obtained from the Maamba township area, Mulungwana and from Wankie, are at the lower end of the scale (Table 19.4) and it is probable that the lower inherent moisture of some coals may be a function of the degree of heat flow from the underlying basement rocks, as suggested by Smith (1970) for the Central Witbank coalfield of South Africa. In this re-
Fig. 19.6. Mid-Zambezi coal. Proximate analysis compared to other African Gondwana coals.
It is interesting to note that, whereas much of the mid-Zambezi coal area is underlain by schistose basement, in the vicinity of Maamba Township and Mulungwa the common basement rock is amphibolite gneiss. It is possible that the relatively higher heat flow from the amphibolites may explain the lower moisture content of the coals in these areas.

The average sulphur content of the Main Seam, the bulk of which is present as pyrites, is generally between 2 and 3 per cent, which is higher than for the other African Gondwana coals. There is considerable local and regional variation in sulphur content; the Kazinze area has a higher value than either Nkandabwe or Mulungwa, the latter area especially having an exceptionally low percentage (Table 19.4). Within the Kazinze region there appears to be a steady increase in sulphur along strike from southwest to northeast, and a thin film of creamish-yellow sulphur is a common feature on most bedding planes within the Main Seam where it is exposed. Apart from regional and local variations, there is also considerable vertical variation (Figs. 19.8 and 19.9). Although in general there is a tendency for the basal sections to be rich in sulphur, there are also subsidiary concentrations of sulphur at higher horizons. The extent to which the total sulphur can be reduced by washing is limited, as a proportion of it is syngenetic and organic in origin. The irregularity of the lateral and vertical distribution of sulphur is again a function of the environment of deposition of the coal.

The average calorific value of the Zambian coal at 6000 cal/gm is much lower than that of the Wankie or Witbank coals and there is a marked tendency for the coal to improve in rank from north to south (Table 19.4). Although there is no great lateral variation in calorific content within an area, there is considerable difference vertically, as amply illustrated by Shafts 2 and 3 (Figs. 19.8 and 19.9). Research on the heating characteristics of the Main Seam indicates that the specific
heat varies directly with the ash content, so that by determining the latter the specific heat of the coal may be estimated to within approximately 147 calories (Sofremines report, 1968; Money et al., in prep.). The overall average calorific values of the Main Seam depend on the relative proportions of their coal seams and muddy coal horizons, but the average value is usually in the range of 6400 to 6600 cal/gm. The latter tend to be more numerous with increase in thickness of the Main Seam. The calorific value and the volatile content of the Zambian coal do not correspond to the standard European classification system.

The mean values obtained from ultimate analyses of the Main Seam in the eastern part of the Kazinze area, on a moisture-ash-free basis (m.a.f.) are carbon 85.9 per cent, hydrogen 4.2 per cent, oxygen 6.8 per cent, sulphur 1.17 per cent, nitrogen 1.77 per cent and chlorine 0.16 per cent. On the basis of Seyler’s classification (in Krevelen, 1961: 20) the Zambian Main Seam would be classed as Sub-Para-Bituminous to Sub-Ortho-Bituminous. According to the carbon-oxygen ratio chart of Hickling (1931), illustrating the continuous variation in the composition of the coals from lignites to anthracites, the Main Seam coal plots between bituminous and anthracite, that is at a higher rank than is actually the case.

CLASSIFICATION OF ZAMBIAN COAL

It is apparent from the above that nearly all permutations of the properties measured by ultimate, proximate or miscellaneous analyses used in the classification of Zambian coal are unsatisfactory. This is largely because there has been a pronounced tendency among American and European coal geologists to classify coals on a limited territorial
Fig. 19.9. Proximate analyses of Main Seam in No. 3 shaft, Kazinze area. Sample intervals every 15 cm. Average value for seam: from 56.7m-61.7m, calorific value = 4850 cal/gm (8790 btu/lb), fixed carbon 52.5%, ash 27.8%, volatiles 17.3%, moisture 2.5%, combustible sulphur 14%. The same but excluding mudstone: calorific value 5933 cal/gm (10,680 btu/lb), fixed carbon 54.7%, ash 24.2%, volatiles 18.6%, moisture 2.4%, combustible sulphur 1.5%
Table 19.4. Summary of Proximate Analyses of Mid-Zambezi and other African Gondwana Coals

<table>
<thead>
<tr>
<th>Areas *</th>
<th>Proximate analyses</th>
<th>H2O %</th>
<th>Ash %</th>
<th>Volatile %</th>
<th>Fixed C %</th>
<th>S %</th>
<th>Cal/gram</th>
<th>Btu/lb of analyses</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Nkandabwe (Tavener-Smith, 1960)</td>
<td>2.0</td>
<td>22.2</td>
<td>22.6</td>
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<td>1.29</td>
<td>5945</td>
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<td>20.2</td>
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<td>5555</td>
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<td>21.4</td>
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<td>0.06</td>
<td>5445</td>
<td>9 800</td>
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<td>21.7</td>
<td>51.5</td>
<td>1.00</td>
<td>5835</td>
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<td>150</td>
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<td>Nkandabwe Open-Pit (Selective Mining)</td>
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<td>22.6</td>
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<td>1.11</td>
<td>7030</td>
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<td></td>
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<tr>
<td>5 Nkandabwe (average all boreholes)</td>
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<td>24.8</td>
<td>22.4</td>
<td>57.7</td>
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<td>5835</td>
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<td>22.4</td>
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<td>7 Nkandabwe Maamba (Intermediate area)</td>
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<tr>
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<td>6805</td>
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<tr>
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<td>14.2</td>
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<td>32 Mbuyura Dept, 1960</td>
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<td>0.09</td>
<td>5790</td>
<td>10 425</td>
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</tbody>
</table>

* Location Numbers (See Fig. 19.6).
† Sources: Drysdall et al., 1967a, b and c; Money, Denman and Drysdall, in prep; Sofremines Report, 1968).

basis, so that systems found to work satisfactorily within one country are unsatisfactory when applied to coals of other countries and continents. Consequently all such classifications are, at best, only approximations and, at worst, do not fulfil the basic purpose of a good system of classification, namely to enable the nature and properties of a substance to be estimated from its position in the classification. However, if only for reasons of...
convenience, a classification is essential and the best available is probably the Interna­tional classification of hard coals. According to this system the code number of the Zambian coal is 400-411. The first figure (4) denotes the class of coal, determined by volatile matter, and in the case of the Zambian Main Seam the value is 25 per cent (m.a.f.); the second digit, denoting the group, is zero (0) or unity (1), as the coking ability of the Zambian coal is almost nil, and finally the third digit of the code, representing the subgroup to which it belongs, is also zero (0) or unity (1) as the coking properties of the Main Seam also are almost nil.

It is generally pointed out by authors of previous classifications that their systems are largely restricted to the so-called 'bright coals of normal composition'. However, in the case of coals that are either young or immature there is no such thing as a 'bright coal'. In fact the Zambian Main Seam is described as a dull coal. Moreover, there are no marked or abrupt differences in the constitution and properties between the various coals and cannel, bogheads, torbanites or oil shale. Indeed, all these coals or carbonaceous matter occur in mixed types and there is gradual merging of one type into another. It is difficult, therefore, to classify coals adequately without paying full regard to the chemical constituents of the parent material of the decaying vegetation and subsequent maceral groups, and to the biochemical, diagenetic and metamorphic changes that took place after deposition.

The difficulty in adequately classifying the Gondwana coals in general and the Zambian Main Seam in particular, stems from the failure to take into account the high content of inertinite. To dismiss this as abnormal is unsatisfactory and does not contribute towards the development of an acceptable international system of classification. Furthermore, a classification cannot be considered satisfactory or acceptable when it is based on the characteristics of material which constitutes less than 5 per cent of the whole. For these reasons it is suggested that the Zambian coal, and indeed all Gondwana coals, be described in terms of the ash and inertinite maceral percentages, since these are their dominant features, in addition to the international standard classification terms and code numbers. By this system, the Zambian coal would be defined as High Ash, Inertinite-90 Bituminous with a code 400-411. In regard to the ash content, less than 10 per cent is defined as low, 10-15 per cent as medium and above 15 per cent as high. The advantage of such a classification is that it pinpoints the significant characteristics of the coal, and hence its possible uses—the most crucial factor in any definition or classification.

PALAEOGEOGRAPHY AND ORIGIN OF COAL

The sedimentary structures and the distribution of the sediments of the Lower Karroo rocks in the mid-Zambezi Valley suggest that the Siankondobo Sandstone Formation is a quiet-water deposit, the Gwembe Coal Measures are the product of a deltaic and fluviatile environment and the Madumbisa Mudstone is probably a lacustrine sequence (Money, Denman and Radosevic, 1968; Money, Denman and Drysdall, in prep.).

The Siankondobo Sandstone, beginning with a relatively thin mixtite horizon and boulder conglomerates, passes upward into a uniform, fine-grained sandstone containing only minor sedimentary structures, but with a characteristic rhythmic banding reminiscent of varves. Its petrology points to deposition under conditions of increasing stability and to a seasonal climate.

This comparatively stable environment ceased to exist during a period of uplift which preceded the deposition of the Gwembe Coal Formation, which shows many features resembling those of modern floodplains and deltas. The coarse-grained basal sediments of the Maamba Sandstone probably accumulated as a result of slumping in an unstable environment along the delta front (Money, Denman and Radosevic, 1968). This instability may have been in part tectonic, resulting from spasmodic isostatic readjustment after the retreat of the Dwyka ice. The upper beds of the Maamba Sandstone, comprising rapid alternations of sandstone and siltstone with horizons of mudstone
and thin coal seams, suggest deposition in a constantly changing environment, similar to that of the lower reaches of a floodplain. This environment was not very different from the present-day Niger delta area where ‘the channel floors are a patchwork of mudbanks subjected to repeated erosion, and ripples of dune covered mounds and sand with clay pebbles and plant debris’ (Allen, 1964: 179).

In such an environment establishment of marshy back-swamps would be expected. Such was the case in the Siankondobo area, where narrow, linear back-swamps with a southwesterly regional slope developed, in which organic detritus accumulated on a larger scale than in the vicinity of the main river itself. The vegetation probably consisted mainly of small plants and shrubs, with few trees, resembling in certain respects present-day vegetation of the upper-Zambezi floodplain between Mongu and Senanga. The palaeorelief map (Fig. 19.10) shows the major rivers and relief of the area at the beginning of the coal formation period.

The relatively steep palaeoslope of the area would have resulted in a well-drained terrain, except during times of flood. Thus, at times, the vegetation was accumulating and growing in an oxygen-deficient, but largely aerobic environment. Within the floodplain complex there were almost certainly bodies of semi-permanent to permanent water, such as ox-bow lakes and meander loop cut-offs, in which reducing conditions prevailed. Thus, the vegetation which eventually gave rise to the Main Seam established itself on the Maamba Sandstone floodplain in an environment which was not always anaerobic. Furthermore, as the region was recovering from an ice age, the climate was temperate and seasonal and the vegetation probably deciduous, resulting in large accumulations of leaf and plant debris every year on the floodplain. Oxidation of the organic matter took place during prolonged exposure in an aerobic environment. The isopach-map of the Main Seam (Fig. 19.3), in conjunction with the palaeorelief map (Fig. 19.10) shows that the distribution of the thicker coal was within the floodplain, but away from the active rivers. The size of the floodplain and its relief obviously determined the extent of the growth of vegetation and hence of the development of coal.

The interseam sandstone bodies in the Main Seam exposed in the Maamba open pit in the northeast, suggest that periodic clastic sedimentation took place in the area of coal formation. The sandstone bodies, which overall trend south-southwesterly (Fig. 19.4) and conform essentially to the proto-drainage and physiography of early Gwembe Coal Formation times, have all the characteristics of channel sandstones. Their form gives further support to the view that the Zambian coal developed in a channelled floodplain environment.

Whereas the high ash content of the Zambian Main Seam is a function of the fine particle size of the detritus and maceral composition, its distribution is largely dependent on the topography of the area of deposition. The ash isoline map of the Kazinze area (Fig. 19.7), the palaeo-relief map (Fig. 19.10) and the map of the distribution of the interseam sandstone (Fig. 19.3), show that the high ash values are distributed along the original ‘valleys’, and that the axis of the interseam sandstone body tends to lie within the proto-valley floor. These relationships suggest that the well-defined palaeorelief tended to persist, perhaps with minor modifications, and affected sedimentation throughout the period of deposition of the Gwembe Coal Formation. The river valleys and other depressions in the floodplain were favourable sites for the transport and deposition of clastic detritus, with the result that the lateral variation in ash values in the Kazinze area is a function of the palaeorelief, and in particular of the position of the main river channel.

The largely inertinite composition of the Zambian coal also indicates a floodplain origin. During peat formation, wood, bark, leaves and roots are transformed to humic substances which form vitrinite macerals. The chemically resistant hydrogen-rich plant components, such as pollens, spores, leaf epidermis, resin and waxes, result in exinite.

Fig. 19.10. Siankondobo-Kazinze Region. Palaeo-relief map at the beginning of Gwembe Coal Formation times. □ shafts; • boreholes. Scale 1: 20,000. Contours at 10m intervals.
macerals, but these are generally of minor importance compared to humic substances. The major factor in the development of vitrinite is an anaerobic medium. However, the carbon-rich macerals, such as fusinite, micrinite, etc., which give rise to the inertinite maceral group, originate from strong aerobic decomposition of plant residues at the peat surface. In the Siankondobo area, as well as in the other mid-Zambezi coal deposits, the environment obviously was aerobic. After the development of each peat layer and accumulated plant debris, a prolonged period of exposure over a dry well-drained terrain would result in inertinite maceral development, giving rise to all the peculiarities of the Zambian coal. In isolated areas such as lakes, where prolonged anaerobic conditions prevailed, syngenetic sulphide was formed. The resulting, apparently unsystematic distribution of sulphur values in the Main Seam thus lends further support to the floodplain theory.

It has long been recognised that there are major differences in the composition and character of the coal seams of the northern and southern hemispheres, and various authors (Steart, 1919; Lamplugh, 1907) have suggested that these differences may be explicable in generic terms, that is the northern coals were formed in situ (autochthonous), whereas those of Gondwanaland are predominantly of drift origin (allochthonous). Wybergh (1922), in his very extensive review of the coal resources of the Republic of South Africa, points out that the coal seams of that country are so extensive and so thick that a purely drift origin seems highly improbable. Plumstead (1966) compared the northern and southern hemisphere coal-measure environments, and emphasised that, whereas the seams of Europe and North America were formed from mainly evergreen vegetation in the nearly stagnant waters of extensive coastal swamps under humid tropical conditions, the Gondwana coals were laid down in lacustrine swamps and were formed from deciduous plants which grew in a comparatively cool, seasonal climate during and following the retreat of the Dwyka ice. To Lamplugh (1907), the Wankie Main Seam in Rhodesia is of drift origin, although this conclusion is queried by Campbell (1915). Lightfoot (1929) is also of the opinion that, as there is no underclay, seat-earth or ganister at Wankie, and the coal consists mainly of finely banded durain and vitrain representing 'alternating layers of shale and vegetable debris', it is allochthonous. This conclusion was discussed in detail by Watson (1958), who argued that the criteria usually accepted as indicative of an in situ origin for the coals of the northern hemisphere are not necessarily valid in southern Africa, in view of the differences of vegetation, climate and environment at the time of formation.

The Gwembe Valley coals, with their high ash content, contain an even higher percentage of mineral detritus than those of Wankie or Witbank. It seems improbable, however, that a seam up to 15 m thick could be entirely of drift origin, considering the immense thickness of the original accumulation of organic debris that would have been required. It seems more likely that the water in the original floodplain swamps was not always stagnant and that fine mineral detritus in suspension entered them; also some detritus may have been wind-borne. It may be noted that peat from present-day bogs may contain more than 30 per cent of ash and that 'coal' formed from such peats would probably have an even higher ash content. Some redistribution of the plant debris may have taken place, particularly if oxidation reached an advanced stage prior to burial; but there would seem to be no valid reason for suggesting that any significant amount was transported into the basins of deposition from outside areas. Although the vertical distribution of ash in the Main Seam is complex, and the bottom of the seam is almost invariably well-defined, the top is interbanded with coaly and carbonaceous mudstone. In these respects, therefore, the coals of the Gwembe Valley resemble those of the Transvaal and Wankie, but the included horizons of high-volatile, cannel-like coal appear to be haphazardly distributed and are not confined to lenses at the base of the seam. They may represent, therefore, accumulations of algal scum in open pools, as has been suggested by Plumstead (1966). Although the Zambian Main Seam does not possess a single feature proving an in situ origin, none
of its characteristics is inconsistent with one. The absence of seat-earths could be a result of the comparatively slow growth of a distinctive type of vegetation. The Maamba Sandstone shows all the textural characteristics of a normal seat-earth, even though no roots or up-standing stems have been found. High inertinite durain coal has been shown to be due to prolonged oxidation and fragmentation of plant debris prior to its burial, and to the high proportion of Bryophytes in the flora. Bearing in mind that the climate of southern Africa during Ecca times was temperate and seasonal rather than tropical, and that coal development took place on a relatively well-drained fluviatile floodplain, the absence of seat-earths does not preclude an in situ origin for the Zambian coal. On the contrary, the palaeogeographical, sedimentological and maceral characteristics of the Main Seam point to an in situ origin.

It would seem, therefore, that for the most part the basal seam of the Gwembe Valley shows features consistent with an autochthonous origin, but towards the top there is increasing evidence of an hypautochthonous mode of accumulation. It is suggested that the two processes are not mutually exclusive, but that most of the plant debris accumulated in situ to form the greater part of the seam, and that some represents redistributed plant material within the floodplain. Transported plant debris forms a higher proportion of the upper part of the seam and the overlying beds than the lower part of the seam, and the carbon content of the overlying mudstones is probably entirely allochthonous.

The deposition of Sandstone A above the Main Seam marked a temporary re-establishment of more active fluviatile conditions. This sandstone is a typical lateral accretion deposit, formed within the confines of the meandering channel in the floodplain. The succeeding Gwembe sandstones again mark the periodic re-establishment of a river complex.

The thickness of Sandstone A and the Main Seam vary inversely. In the southwestern extremity of the Kazinze area the seam is very thin or absent in areas where the sandstone is thickest, whereas its thickness is greatest where sandstone is absent, as in the Kazinze River area. This suggests that, even during periods when clastic sedimentation was almost at a standstill, for example during the formation of the Main Seam, fine-grained argillaceous sediments accumulated in the areas of active river channels.

In brief, the complex sedimentary sequence beginning with the graded or fining-upward units of coarse Maamba Sandstone, through intimately interbedded coaly mudstone, coal and sandstone, passes upwards to the relatively organic-free, greyish mudstones of the Izuma Beds. The sediments are characterised by vertical lithological variations on more than one scale, and the lateral physical and compositional variations are even more marked. Splitting and thinning of seams is as important as the gradual passage into coaly mudstone and sandstone. One of the characteristic features within this fine clastic sequence is the occurrence of sandstones that represent channel deposits cutting across the muddy alluvial floodplain.

The heavy mineral suites and the palaeocurrent directions in the arenaceous sediments indicate that the provenance of the Siankondobo and Maamba Sandstones and of Sandstone A was a gneiss and quartzite terrain, lying to the north-northeast. Sandstones C and D include heavy minerals derived from basic rocks, which were probably also the source of their iron content, but the heavy mineral suite of Sandstone E is of granitic origin (Radoevic, Money and Denman, 1968; Money, Denman and Drysdall, in prep.).

REFERENCES


—— , in prep. The geology of Western Province and parts of North-Western Province. Mem. geol. Surv. Zambia, 3.


Late Palaeozoic coal in Brazil is distributed in three distinct regions: the Eastern Region of Parnaiba Basin where there are thin seams of no economic significance in the Mississippian Poti Formation; the Western Region of Parnaiba Basin with unworkable coals in the Pennsylvanian Piaui Formation; the Southern Brazil Region which contains all the workable coal reserves of the country in the Rio Bonito Formation of the Paraná Basin Gondwana stratigraphic sequence. The total inferred reserves are about three thousand million tons and the region produces about 5.6 million tons of run-of-mine coal annually. These coal deposits are described in broad outline.

Current research suggests that as the late Palaeozoic ice sheet retreated across the Paraná Basin from south to north, a large body of water covered the stable platform. This was isolated from the sea in the southwest by a low barrier. These changes, together with the amelioration in the climate, produced abundantly vegetated paludal environments along the southern margins of this water body, and in these the coal of the Rio Bonito Formation formed. The delicate balance of environments in this region has produced the complexity of the stratigraphy in the lower part of the Tubarão Group described in this paper.
Fig. 20.1. Neo-Palaeozoic coal occurrences in Brazil (after Machado, 1970)

thick coal seams occur, whereas reddish calciferous sandstones occur in the lower part.

Southern Brazil Region

These lie in a belt through the States of Rio Grande do Sul, Santa Catarina, Paraná and São Paulo (Fig. 20.1). The existing coal layers belong to the top of the Carboniferous or the Early Permian, and they lie within the basal group (Guatá Group, Rio Bonito Formation) of the Gondwana column of the so-called Paraná Sedimentary Basin. The total inferred reserve is about three thousand million tons and nowadays the mines produce about five million tons per annum.
Table 20.1. Stratigraphy of the Palaeozoic Era in the West of Piaui State

<table>
<thead>
<tr>
<th>Period</th>
<th>Stage</th>
<th>Formation</th>
<th>Member</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carboniferous</td>
<td>Upper</td>
<td>Piaui</td>
<td>—</td>
<td>Continental sediments with rare marine intercalations</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>Poti</td>
<td>—</td>
<td>Continental sediments in the upper zone and marine sediments in the lower zone</td>
</tr>
<tr>
<td></td>
<td>Upper</td>
<td>Longá</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Ipiranga</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Oeiras</td>
<td></td>
</tr>
<tr>
<td>Devonian</td>
<td></td>
<td></td>
<td>Passagem</td>
<td>Cycle of marine sediments</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Picos</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>Pimenteiras</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Itaim</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Serra Grande</td>
<td>—</td>
<td></td>
</tr>
</tbody>
</table>

Pre-Devonian and Crystalline Basement

Table 20.2. Coal Reserves in Southern Brazil
(all figures quoted in tons)

<table>
<thead>
<tr>
<th>State</th>
<th>Deposit</th>
<th>Blocked reserve</th>
<th>Total estimated reserve</th>
<th>Type of coal</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>125,000,000</td>
<td>725,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Hulha Negra</td>
<td>10,000,000</td>
<td>100,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Sã0 Sepé</td>
<td>3,500,000</td>
<td>7,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td>Rio Grande</td>
<td>Iruí</td>
<td>109,000,000</td>
<td>330,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Leão-Butiá</td>
<td>50,000,000</td>
<td>80,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Arroio dos Ratos</td>
<td>Mined</td>
<td>Mined</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Charqueadas</td>
<td>177,500,000</td>
<td>890,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td>Gravataí</td>
<td>12,000,000</td>
<td>15,000,000</td>
<td>Sub-bituminous</td>
</tr>
<tr>
<td></td>
<td></td>
<td>487,000,000</td>
<td>2,147,000,000</td>
<td>—</td>
</tr>
<tr>
<td>Santa Catarina</td>
<td>Santa Catarina</td>
<td>270,000,000</td>
<td>1,105,000,000</td>
<td>Bituminous</td>
</tr>
<tr>
<td></td>
<td></td>
<td>270,000,000</td>
<td>1,105,000,000</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Rio Tibagi</td>
<td>5,800,000</td>
<td>—</td>
<td>Bituminous</td>
</tr>
<tr>
<td></td>
<td>Rio do Peixe</td>
<td>25,000,000</td>
<td>35,000,000</td>
<td>Bituminous</td>
</tr>
<tr>
<td></td>
<td>Ibití</td>
<td>150,000</td>
<td>1,373,000</td>
<td>Anthracite</td>
</tr>
<tr>
<td></td>
<td>Wenceslau Braz</td>
<td>—</td>
<td>90,000</td>
<td>Bituminous</td>
</tr>
<tr>
<td></td>
<td>Barbosas</td>
<td>180,000</td>
<td>600,000</td>
<td>Anthracite and Bit.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>31,130,000</td>
<td>37,065,000</td>
<td>—</td>
</tr>
<tr>
<td>TOTAL</td>
<td></td>
<td>788,130,000</td>
<td>3,289,065,000</td>
<td>—</td>
</tr>
</tbody>
</table>

Sources: Works of DACM, DNPM, DPM, COPELMI, IBPT and Personnel.
The distribution of the main coal deposits in southern Brazil, as well as the geology of the associated rocks, is shown on Figure 20.2. Generally speaking, it can be said that the belt of Gondwana sediments, in which coal deposits occur, behaves as a peripheral depression between the Coastal Crystalline Shield and the interior basaltic plateau. The Coastal Crystalline Shield has mountainous topography and is interrupted in the area around the boundary of the States of Rio Grande do Sul and Santa Catarina. The interior basaltic plateau, whose rocks conceal the Gondwana column of the Parana Basin, forms the largest field of lava in the world, occupying an area of about 1,200,000 km$^2$. This column, which lies directly on rocks of the shield or on Devonian sediments, is formed by the following groups, cited in descending order:

- *São Bento Group*—Triassic-Jurassic sediments covered by basalts up to 1000 m thick in some areas and ranging up into the Early Cretaceous.

- *Passa Dois Group*—Permian sediments.

- *Tubarão Group*—sediments whose age is not yet precisely known, but which is probably Late Carboniferous to Early Permian.

The coals are in the Tubarão Group, which is subdivided as follows:

- **Tubarão Group**:  
  - *Guatá Sub-Group*  
  - *Palermo Formation*  
  - *Rio Bonito Formation*

- *Itararé Sub-Group*

  In this classification, the glacial sediments are in the Itararé Sub-Group and the main coals are included in the Rio Bonito Formation.

During the course of research, it transpired that the stratigraphy of the Tubarão Group was more complex than expected, but the details are not the concern of this work. Nevertheless, it should be noted that interglacial coal beds have been demonstrated, the number and ages of the tillites change from south to north, and there are restricted marine transgressions within the Itararé Sub-Group as well as the Rio Bonito Formation. As a result of these discoveries some workers believe that the basal glacial sediments (with or without inter-glacial coal beds) can be separated from the sediments of the overlying Guatá Sub-Group, while others believe that within the Tubarão Group there is considerable lateral and vertical intertonguing, at least on the eastern side of the sedimentary basin.

However, in the area of the main coal deposits, the Tubarão Group is relatively simple and the column previously cited seems to be applicable. Only the post-glacial coals, that is those within Rio Bonito Formation, are economic. In this region, intercalations of marine sediments are absent. Hence, in what follows, the formations of the Guatá Sub-Group may be dealt with without consideration of these complications.

### Rio Bonito Formation

This formation, whose thickness in the area of the deposits is normally less than 90 m, consists basically of intercalations of quartzose sandstone, grey or cream shales, carbonaceous shales and coal seams, the latter being derived from a *Glossopteris-Ganagmopteris* flora. Sedimentary structures of shallow water origin, such as ripple marks and cross-bedding, as well as some deltaic structures, occur in the sandstones. The clay fraction is mainly kaolinitic (Formoso and Figueiredo, 1967). The upper contact is conformable with the Palermo Formation and the lower contact is either with rhythmites, damicitites, conglomerates, etc. of the Itararé Sub-Group or, in many regions of Rio Grande do Sul State, with crystalline basement.

### Palermo Formation

This formation, whose thickness in the area of the deposits is commonly less than 100 m, consists predominantly of sandy siltstones and thin silty sandstones, classifiable as subgreywackes, typically characterised by a thin, wavy and irregular stratification, thin interlamination of silt and fine sand and streaks or striae of dark carbonaceous material. As a result the rock is commonly striped. Very fine cross-bedding, some pyrite, worm tubes and delicate primary features characteristic of hydroplastic sediments deposited on tidal-flat plains and inner shelves have been found. Previously no conclusive
The clay fraction, unlike that of the Rio Bonito Formation, is predominantly montmorillonite (Formoso and Figueiredo, 1967). There are occasional thin and lenticular beds of whitish quartzose sandstones, commonly very calcareous, in the formation.
CHARACTERISTICS OF COAL BEDS

Only the following deposits are still mined (see Fig. 20.2): Candiota, Leão-Butiá and Charqueadas (Rio Grande do Sul State), Santa Catarina (in the State of the same name), Rio Tibagi e Rio do Piexe (Paraná State). We shall discuss the basic behaviour and the characteristics of the Southern Brazil coal seams, using them as patterns of reference.

Rio Grande do Sul State

Candiota Deposit. This deposit is situated in the vicinity of Dario Lassance, Municipality of Bagé. Its total area is not yet known and only its eastern limit can be traced in the contact between the Tubarão Group and the rocks of the shield. A general idea of its setting is given in Figure 20.3. Coal occurs in all outcrops of Gondwana sediments. Due to the presence of overburden less than 16 m
Candiota coal mine. Partial structural section at Malha II.

thick, the areas called 'Malha I, II and III' were selected for open-pit mining.

In the Rio Bonito Formation, there are about ten coal seams. Only the so-called 'Candiota Bed' (see Fig. 20.4) can be mined. Characteristically, this unit is subdivided in two coal seams, each one with average thickness of 2 m separated by a layer of grey claystone averaging 0.90 m thick. The coal is covered by another claystone with similar ceramic and refractory properties, and averaging a little more than 1 m in thickness.

Morphologically, the deposit occurs as more or less elongated ridges with relatively plain or wavy tops, separated by more or less deep depressions where the bed is absent due to erosion. Such a setting produces irregular mining areas as is illustrated in Figure 20.3.

Structurally the measures lie in a broad meridional syncline whose northern flank lies in Hulha Negra (Fig. 20.2). There is, however, a series of local structures, as is the case of Malha II, for example, where a gentle southerly plunging anticline is cut by a fault with a throw of about 50 m, and forms its edge (Fig. 20.3).

The coal of Candiota Bed is of inferior quality, but open-pit mining allows low cost extraction, so that the powerhouse there produces the cheapest thermally generated electricity in Brazil. Coal quality can be expressed by the average numbers shown in Table 20.3.

---

**Table 20.3. Candiota Seam**

Proximate Analysis of R.O.M. Coal (Dry-Basis)*

<table>
<thead>
<tr>
<th>Items</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volatiles</td>
<td>22.2</td>
</tr>
<tr>
<td>Fixed carbon</td>
<td>27.9</td>
</tr>
<tr>
<td>Ash</td>
<td>49.9</td>
</tr>
<tr>
<td>Sulphur</td>
<td>1.9</td>
</tr>
<tr>
<td>Sup. calor. power</td>
<td>3,402 c/g</td>
</tr>
</tbody>
</table>

* Hygroscopic moisture 10.2%.

Ybert (pers. comm.) has described the miospores of Candiota-Hulha Negra; much of his material, which we supplied, came from these coal measures. He reports that


L'apparition des formes telles que Balmella et Quadrisporites semble également appuyer cette hypothèse. Il faut noter toutefois, que nous n'avons jamais retrouvé la composition indiquée par TIWARI et NAVAL pour leur échantillon 1911 où l'ensemble des genres Balmella, Congoites, Brazilea, Pilasporites, Quadrisporites, et Dissectispora dépasse 80% alors que les formes correspondantes n'atteignent jamais 30% avec une moyenne de 6,5% dans l'ensemble de nos échantillons. Ce fait nous conduit à douter de la
représentativité de l'échantillon étudié par les auteurs indiens. Nous avons eu l'occasion d'observer les charbons en provenance d'autres bassins du Rio Grande do Sul, du Santa Catarina et du Paraná et, nous avons constaté une grande similitude de composition, indiquant des conditions climatiques et écologiques identiques sur toute l'étendue de l'aire de dépôt de ces charbons. Ceci nous montre une fois de plus, le danger qui existe à tirer des conclusions de l'étude d'échantillons trop peu nombreux et de provenance douteuse.

Sur le plan local, les données quantitatives peuvent aider à établir des corrélations et, notamment, à différencier le Banc Inférieur de Banc Supérieur de la Veine Candiota par la prédominance du groupe des Zonati-Cavati dans le premier cas et des Cingulati dans le second. Nous avons pu de cette façon, reconnaître des deux bancs de la Veine Candiota dans les échantillons du Río Jaguarão et, rapporter la veine inférieure de la mine de Hulha Negra au Banc Supérieur de la veine Candiota. Toutefois, certains échantillons présentant des anomalies (Sondagens 805, 807 et 811), il est nécessaire de s'appuyer sur les données géologiques et d'étudier plusieurs veines ou passées successives pour diminuer les risques d'erreurs. En outre, il est pratiquement impossible de différencier les diverses passées de la Veine Appleby car celles-ci sont très irrégulières et n'ont pas été échantillonnées totalement dans aucun des sondages.

A study of the minor elements of the Candiota Seam carried out by Urdininea and Pintaude (1972) is summarised in Table 20.4. Petrographically, Candiota coal has been only recently studied. The data obtained by Nahuys et al. (1968) in the analysis of the cores of the drilling F3-AC, performed in Malha III are given in Table 20.5.

Finally it is noted that the Candiota seam contains thin whitish beds that are very regular in their distribution and position within the seam. Their thicknesses are of the order of centimetres. Two of them occur in the Upper Bank, the other in the Lower Bank (Fig. 20.4). These are apparently 'Tonstein' according to the results of Corrêa da Silva (1971). They are coherent massive layers, sometimes with microcrystalline appearance, grey-whitish, containing carbonaceous fragments, leaf impressions, megaspores and microspores. Chemical analysis showed that the clay is almost pure kaolinite, and X-ray diffraction indicated small quantities of mixed layer illite-montmorillonite. According to Corrêa da Silva (1971), they are 'Stratotonstein' in Bouroz's classification (1964) and, by their petrographical characters, the two upper layers can be a 'Kristalltonstein' and the lower layer a 'Grauptentonstein', in Schuller's classification (in Scheere, 1958).

Table 20.4. Concentration of Minor Elements in the Coal Ash of Candiota
(in ppm)

<table>
<thead>
<tr>
<th>Elements</th>
<th>Upper Bank (12)</th>
<th>Lower Bank (12)</th>
<th>Upper Bank mean sample</th>
<th>Lower Bank mean sample</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mn</td>
<td>190</td>
<td>320</td>
<td>200</td>
<td>250</td>
</tr>
<tr>
<td>Zn</td>
<td>nd</td>
<td>nd</td>
<td>nd</td>
<td>nd</td>
</tr>
<tr>
<td>Y</td>
<td>80</td>
<td>50</td>
<td>70</td>
<td>33</td>
</tr>
<tr>
<td>Zr</td>
<td>325</td>
<td>259</td>
<td>190</td>
<td>200</td>
</tr>
<tr>
<td>Ga</td>
<td>24</td>
<td>19</td>
<td>21</td>
<td>21</td>
</tr>
<tr>
<td>V</td>
<td>79</td>
<td>66</td>
<td>60</td>
<td>56</td>
</tr>
<tr>
<td>Nb</td>
<td>75</td>
<td>24</td>
<td>25</td>
<td>21</td>
</tr>
<tr>
<td>Ti</td>
<td>6200</td>
<td>5325</td>
<td>5400</td>
<td>5400</td>
</tr>
<tr>
<td>Ge</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>Cq</td>
<td>16</td>
<td>14</td>
<td>15</td>
<td>13</td>
</tr>
<tr>
<td>Ba</td>
<td>240</td>
<td>200</td>
<td>170</td>
<td>180</td>
</tr>
<tr>
<td>Sc</td>
<td>16</td>
<td>13</td>
<td>12</td>
<td>11</td>
</tr>
<tr>
<td>Cr</td>
<td>56</td>
<td>55</td>
<td>40</td>
<td>60</td>
</tr>
<tr>
<td>Sr</td>
<td>157</td>
<td>117</td>
<td>90</td>
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<td>Pb</td>
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<td>Co</td>
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<td>11</td>
<td>10</td>
</tr>
<tr>
<td>Be</td>
<td>12</td>
<td>11</td>
<td>10</td>
<td>10</td>
</tr>
</tbody>
</table>

(12) — Average of samples
nd — not detected
tr — traces
Table 20.5. Analysis of the Macerals in the Drilling F3–AC

<table>
<thead>
<tr>
<th>Intervals</th>
<th>Analysis of the macerals</th>
<th>R.O.M. coal</th>
<th>Pure Coal</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Vitrinite</td>
<td>Exinite</td>
</tr>
<tr>
<td>12.61 – 14.69</td>
<td>29  7  23  31</td>
<td>56  11  33  0.54</td>
<td></td>
</tr>
<tr>
<td>12.61 – 17.98</td>
<td>28  4  18  50</td>
<td>56  8  36  0.56</td>
<td></td>
</tr>
<tr>
<td>15.58 – 17.98</td>
<td>37  1  17  45</td>
<td>66  3  31  0.53</td>
<td></td>
</tr>
<tr>
<td>23.90 – 25.10</td>
<td>25  3  16  56</td>
<td>57  6  37  0.58</td>
<td></td>
</tr>
<tr>
<td>28.23 – 29.58</td>
<td>34  6  19  41</td>
<td>58  10  32  0.55</td>
<td></td>
</tr>
</tbody>
</table>

Leão-Butiá Deposit
This deposit is situated in Butiá municipality and for some time it has been mined by underground methods. Open-pit mining areas have been exhausted. The general development of the basin is shown on Figure 20.5. The mine shafts vary in depth from 30-127 m.

The deposit is defined in all directions except the northwest, where its limits are quite unknown. It has upper and lower seams that are variously split and are separated by an interval of shales whose thickness varies from 2-15 m; the lower seam rests on a greywacke (called ‘Pedra Areia’ in the mine), which was considered by Putzer (1952: 145) to be tillite (see Machado and Castanho, 1957: 10).
Fig. 20.6. Partial structural section through the Leão-Butia Basin (Leão coal mine)

Table 20.6. Proximate Analysis of Leão-Butiá R.O.M. Coal (Dry-Basis)*

<table>
<thead>
<tr>
<th>Items</th>
<th>Values %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volatiles</td>
<td>25.6</td>
</tr>
<tr>
<td>Fixed carbon</td>
<td>34.8</td>
</tr>
<tr>
<td>Ash</td>
<td>39.6</td>
</tr>
<tr>
<td>Sulphur</td>
<td>0.7</td>
</tr>
<tr>
<td>Sup. calor. power</td>
<td>4,323 c/g</td>
</tr>
</tbody>
</table>

* Hygroscopic moisture 5.00%.

This lithology (see Figs. 20.8 and 20.9) commonly associated both with coal and 'Palermo' lithology (Grangeon, 1960; Dorokhin et al., 1969: 208). There are many coal seams, intercalated among the 'Pedra Areia' strata, but these are of no economic significance because of their limited extent and the remarkable irregularities in their general behaviour.

The mined coal seam, Bed I, belongs to the lower set and has an average thickness of 1.30 m. In the ML-24 drilling, the coal is close to the granitic basement; in the coal measures of the Rio Grande do Sul it is common to find coal beds only a few metres above the crystalline basement.

Structurally, Figure 20.5 shows that the Leão-Butiá deposit is intersected diagonally by a large, periodically reactivated, regional fault along which the eastern side is upthrown.
Table 20.7. Petrographic Composition of Leão Coal (Bed 1)

<table>
<thead>
<tr>
<th>Reflectance</th>
<th>Vitrinite</th>
<th>Exinite</th>
<th>Inertinite</th>
<th>Sterile</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Collinite</td>
<td>Telinite</td>
<td>Total</td>
<td>Micrinite</td>
</tr>
<tr>
<td>0.51</td>
<td>31</td>
<td>17</td>
<td>48</td>
<td>13</td>
</tr>
<tr>
<td>0.51</td>
<td>31</td>
<td>18</td>
<td>49</td>
<td>13</td>
</tr>
</tbody>
</table>

Fig. 20.7. Charqueadas coal deposit. Partial structural section through the centre of the basin.

about 40 m, and there is also some horizontal movement.

The coal of this deposit is the best in the State and its composition can be expressed by the average values shown in Table 20.6.

Trindade (1962) determined the following spores: Lagenoisporites brasiliensis, Lagenoisporites sinuatus and Cystosporites sp. Petrographically, this coal was studied by Nahuys (1966), from whose work the data in Table 20.7 are abstracted.

Minor element studies on the ash after combustion have been done by Urdiníncia and Pintaúde (1972), and the results shown in Table 20.8 have been obtained.

In comparison with Candiota coals these values are, in general, higher, especially Ge, Co and Ni.

Charqueadas Deposit

This deposit (see reserves in Table 20.2) occupies areas of São Jerônimo and Triunfo municipalities and is mined by underground methods at a depth of 270 m. The general development of the basin is shown on Figure 20.5.

There are five coal seams, but only the so-called I1F and I2B Seams can be mined. At present only I1F is worked. This seam generally has three different coal units: the first is about 40 cm thick, the second varies from 1 m to 2 m, and the third between 10 cm and 30 cm. The dip in the whole mine area is to the north, but there are numerous small faults that cause minor disturbances. Such faults are not surprising, because diabase intrusives are very abundant at several levels. They are contemporaneous with the extensive Jurassic-Cretaceous basalts of southern Brazil.
The quality of the IIF seam is represented by the average values shown in Table 20.9. Charqueadas spores were studied by Trinidade (1959, 1964), who reported the following species: Lagenoisporites brasiliensis, L. sinuatus, Triiletes nitens, T. tenius, T. trivialis, T. trivedii, T. vulgatus, T. labiosus and Calamospora sp. Petrographic details reported by Nahuys (1966) are summarised in
LEGEND

Structural Contours of "Barro Branco" Seam
Outercrop Line of "Barro Branco" Seam
0 to 50 m Overburden Area
"Barro Branco" Seam Natural Limit
Geologic Section

Fig. 20.9. Coal basin of Santa Catarina
Table 20.8. Concentration of Minor Elements in the Coal Ash of Leão (in ppm)

<table>
<thead>
<tr>
<th>Elements</th>
<th>Min.</th>
<th>Max.</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>24</td>
<td>74</td>
<td>55</td>
</tr>
<tr>
<td>U2O8</td>
<td>4.1</td>
<td>10.3</td>
<td>6.3</td>
</tr>
<tr>
<td>Mn</td>
<td>200</td>
<td>300</td>
<td>260</td>
</tr>
<tr>
<td>Zn</td>
<td>nd</td>
<td>150</td>
<td>—</td>
</tr>
<tr>
<td>Y</td>
<td>52</td>
<td>135</td>
<td>101</td>
</tr>
<tr>
<td>Zr</td>
<td>140</td>
<td>1,600</td>
<td>447</td>
</tr>
<tr>
<td>Cu</td>
<td>23</td>
<td>72</td>
<td>37</td>
</tr>
<tr>
<td>V</td>
<td>70</td>
<td>240</td>
<td>126.5</td>
</tr>
<tr>
<td>Ni</td>
<td>31</td>
<td>240</td>
<td>90</td>
</tr>
<tr>
<td>Ti</td>
<td>6,000</td>
<td>12,000</td>
<td>9,400</td>
</tr>
<tr>
<td>Ge</td>
<td>52</td>
<td>420</td>
<td>85</td>
</tr>
<tr>
<td>Ga</td>
<td>15.5</td>
<td>41</td>
<td>22</td>
</tr>
<tr>
<td>Ba</td>
<td>10</td>
<td>320</td>
<td>192</td>
</tr>
<tr>
<td>Sc</td>
<td>10.5</td>
<td>68</td>
<td>26</td>
</tr>
<tr>
<td>Cr</td>
<td>27</td>
<td>170</td>
<td>84.5</td>
</tr>
<tr>
<td>Sr</td>
<td>150</td>
<td>600</td>
<td>251</td>
</tr>
<tr>
<td>Pb</td>
<td>20</td>
<td>92</td>
<td>30.5</td>
</tr>
<tr>
<td>Co</td>
<td>5</td>
<td>48</td>
<td>19.5</td>
</tr>
<tr>
<td>Be</td>
<td>10</td>
<td>30</td>
<td>20</td>
</tr>
<tr>
<td>Mo</td>
<td>—</td>
<td>—</td>
<td>nd</td>
</tr>
<tr>
<td>La</td>
<td>—</td>
<td>—</td>
<td>tr</td>
</tr>
<tr>
<td>Nb</td>
<td>—</td>
<td>—</td>
<td>tr</td>
</tr>
</tbody>
</table>

nd – not detected
tr – traces
No. of samples – 15

Table 20.9. Proximate Analysis of Charqueadas R.O.M. Coal (Dry-Basis)*

<table>
<thead>
<tr>
<th>Items</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>%</td>
</tr>
<tr>
<td>Volatiles</td>
<td>19.5</td>
</tr>
<tr>
<td>Fixed carbon</td>
<td>26.9</td>
</tr>
<tr>
<td>Ashes</td>
<td>53.6</td>
</tr>
<tr>
<td>Sulphur</td>
<td>0.7</td>
</tr>
<tr>
<td>Sup. calor. power</td>
<td>3,151 c/g</td>
</tr>
</tbody>
</table>

* Hygroscopic moisture 6.70%.

Table 20.10. Concentration of Minor Elements

Table 20.10. Petrographic Composition of Charqueadas Coal

Analysis of the macerals

<table>
<thead>
<tr>
<th>Bed</th>
<th>R.O.M. coal</th>
<th>Pure coal</th>
<th>Mean reflectance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Vitrinite</td>
<td>Exinite</td>
<td>Inertinite</td>
</tr>
<tr>
<td>I1</td>
<td>23</td>
<td>9</td>
<td>29</td>
</tr>
<tr>
<td>I2B</td>
<td>29</td>
<td>2</td>
<td>15</td>
</tr>
</tbody>
</table>
than that given above, and some are even in the opposite direction (Figs. 20.10 and 20.11). Dips of 2° to 3° are most common. There are rare mines which manage to develop all their galleries with inclines of less than 3 per cent. Inclines up to 10 per cent are not uncommon, and some of 18 per cent have been observed.

The seam is thus gently folded. Faults are frequent, but normally they are short and
Gondwana Coal Deposits

have throws of 0.3 to 1.5 m—rarely 2 or 3 m. Dolerite sills and dykes are frequent. The sills are generally up to 1 m thick, rarely 2 or 3 m; dykes are more common, and are variable in thicknesses between 0.2 and 1.0 m. There are rare dykes of 2 to 5 m. The thickest sills are marked on Figure 20.8.

There are hundreds of kilometres of galleries where one can observe the coal, and the following features can be seen: in each 1000 m of gallery, in any direction, about ten small faults, with throws varying from 0.3 to 1.5 m, average 0.6 m; from 0.2 to 1.0 m thick, average 0.4 m.

The sills have been introduced preferentially at the contact of the Itararé with the basement; in certain siltstone levels near the top of the Itararé; in the carbonaceous layers, especially when they contain coal and not only carbonaceous shales; in a siltstone horizon near the top of the Rio Bonito Formation, about 20 m above Barro Branco coal seam; in the Itatí Formation, especially where they contain a lens of siliceous-calcareous rocks. It is interesting to note that, as also happens in Rio Grande do Sul, the Palermo Formation is never intruded. The faults and dykes have little effect on the coal. The sills, however, are much more destructive, and in places replace the whole seam.

The basin originally extended to the east and north beyond the present line of outcrops, except in certain areas near Icara, where the natural border of the basin was found. Southward, the coal passes below sea level. To the northeast it plunges below the Serra Geral scarp, but to the southwest the deposit border was discovered in many places. At the natural margin of the basin there is a rapid reduction in thickness of the bed; at first, the Veia do Forro disappears; next Quadracão begins to disappear, and the bed is almost reduced to the Banco; the Alevante becomes more clayish and its thickness increases to 3-5 m, while the Barro Branco Superior sandstone becomes thin, finer-grained clayey and grey to dark brown in colour. At last, the bed is reduced to a layer of the Banco; the Alevante interfocus with Barro Branco Superior sandstone in a sandy shale or a very clayish sandstone, grey to dark brown in colour, with a thickness of 8, 10 or more metres. Beyond the edge of the coal-bearing part of the basin this sandy shale lies directly over the basal sandstone.

The deposit limits (Fig. 20.9) are well known eastward; northward, there are areas yet to be tested, and westward the limits are largely inferred from sparse drillings. In the northwest, where the coal dips below the Serra Geral scarp, there is no economic interest in verifying the continuity of the coal. Toward the south the coal is below sea level.

Table 20.11 presents some typical proximate analyses of the coal of the Barro Branco bed (dry-basis analysis of R.O.M. coal).

Finally, the total reserves of this bed in the whole deposit was estimated by Putzer (1952) as being about 900,000,000 tons. Current research by CPRM considers the useful reserves to be much less.

Irapuá Seam

This is next to the Barro Branco bed in importance because it contains a similar bituminous and coking coal, and it is a large producer (Fig. 20.10). It is economically mined in two special areas: Rio Maina and Siderópolis, occurring as restricted basins hundreds of metres wide and a few kilometres long.

<table>
<thead>
<tr>
<th>Items</th>
<th>Metropolitana</th>
<th>Criciuma</th>
<th>Própresa</th>
<th>Urussanga</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture (%)</td>
<td>2.7</td>
<td>3.2</td>
<td>2.5</td>
<td>3.0</td>
</tr>
<tr>
<td>Volatiles (%)</td>
<td>27.1</td>
<td>25.2</td>
<td>24.8</td>
<td>25.5</td>
</tr>
<tr>
<td>Fixed carbon (%)</td>
<td>40.7</td>
<td>39.3</td>
<td>40.9</td>
<td>33.8</td>
</tr>
<tr>
<td>Ash (%)</td>
<td>29.5</td>
<td>32.3</td>
<td>31.8</td>
<td>37.7</td>
</tr>
<tr>
<td>Sulphur (%)</td>
<td>3.6</td>
<td>5.5</td>
<td>4.5</td>
<td>5.1</td>
</tr>
<tr>
<td>Sup. calor. power</td>
<td>5,575 c/g</td>
<td>5,000 c/g</td>
<td>5,220 c/g</td>
<td>5,750 c/g</td>
</tr>
</tbody>
</table>
(Fig. 20.12). In other areas, it is reduced to a few centimetres in thickness or it begins to form a carbonaceous shale. Reserves in the areas above mentioned were estimated by Machado et al. (1962) at 7,200,000 tons.

**Ponte Alta Seam**

This bed is normally a carbonaceous shale, with thin levels of coal. In the past it has been mined locally, but there are no present workings.

**Bonito Seam**

Coal mining in Santa Catarina began in this bed more than half a century ago. It was mined especially in the surroundings of Lauro Müller town (Figs. 20.8 and 20.9), where the coal was thick, although of inferior quality; it is also worked in certain areas to the south where the unit begins to form, a carbonaceous shale with thin coal beds. Consequently this bed has limited economic importance. The total reserve in the deposit was estimated by Putzer (1952) as being about 260,000,000 tons.

Finally, there are several palynological studies of the Santa Catarina coal. Trindade (1970: 307) has given the following conclusions on the basis of megaspores.

<table>
<thead>
<tr>
<th>Bed</th>
<th>Westphalian C-D — Stephanian</th>
<th>Westphalian A-B</th>
<th>Westphalian A-B</th>
<th>Namurian C</th>
<th>Namurian C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barro Branco</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Irapuá</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ponte Alta</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bonito</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pré-Bonito</td>
<td>Namurian C</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Megaspore assemblages analysed so far indicate that the Permo-Carboniferous of southern Brazil has a monotonous flora of lycopoids.

Daemon (1966: 211) analysed the distribution and zoning of the sporomorphs in the upper Palaeozoic of the Paraná Basin. Nahuys (1966) has published petrographic studies on the Santa Catarina coal. The partial data in Table 20.12 were taken from his work.

**Paraná State**

In Paraná, only the coal of the deposits of Rio do Peixe and of Rio Tibagi is mined (Fig. 20.2). In both basins, reserves are small and only one of the seams present is economically minable. It is normally less than 1 m thick. In the Rio do Peixe the unit is usually composed of two coal seams separated by a bed of shale of variable thickness, whereas in the Rio Tibagi there is a single seam sometimes with thin interlaminations of shale. These deposits are structurally similar to those of the Rio Grande do Sul, and so no comment is needed. The mined coal has good coking properties but the coke cannot be used in steel making, because of the high percentage of organic sulphur. In Table 20.13, typical results of some proximate analyses of the R.O.M. coal are listed.

The Rio Tabagi coals have a high variability in volatiles (normally low), because of the occurrence of a considerable series of thick diabase dykes cutting the coal beds on northeast or southeast trends. It is noted also that the coals are post-glacial, tillites and interglacial sediments occurring beneath them.

Finally, in Figure 20.13 the position of the

---

**Table 20.12. Petrographic Composition of Santa Catarina Coal**

**Analysis of the macerals**

<table>
<thead>
<tr>
<th>Bed</th>
<th>Vitrinite Reflectance</th>
<th>Exinite Collinite</th>
<th>Exinite Telinite</th>
<th>Exinite Total</th>
<th>Inertinite Micrinite</th>
<th>Inertinite Semi Fusinite</th>
<th>Inertinite Fusinite</th>
<th>Inertinite Exclerotinite</th>
<th>Sterile Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barro Branco</td>
<td>0.91</td>
<td>58</td>
<td>3</td>
<td>61</td>
<td>7</td>
<td>8</td>
<td>11</td>
<td>6</td>
<td>1</td>
</tr>
<tr>
<td>(Cia. CB-CA)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Irapuá (Cia.</td>
<td>0.92</td>
<td>60</td>
<td>13</td>
<td>73</td>
<td>6</td>
<td>8</td>
<td>6</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>Metropolitana</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

---

Gondwana Coal in Southern Brazil 287
Fig. 20.12. Situation of Irapua Bed in areas of Cias Metropolitana and Catarinense (modified after Geosol S.A., 1962).

South Brazilian coals in the metamorphic series of Nahuys (1966) is shown.

POINTS CONCERNING COAL GENESIS

The following points merit special consideration.

1. The coal deposits discussed here are preserved in gentle synclines or structural basins. Such structures are mainly due to the higher amount of compaction of the peat originally deposited in the centres of the basins relative to the clastics that predominate around the basin margins (Machado, 1961). The general tectonic environment was one of slow epeirogenic movement. Consequently, normal faults occur in the deposits, the great majority being relatively small (Putzer, 1953).

2. The coals of southern Brazil are of inferior type. In our opinion, there are two main contributing factors—they were formed

<table>
<thead>
<tr>
<th>Table 20.13. Proximate Analysis of Paraná R.O.M. Coal</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Deposits</strong></td>
</tr>
<tr>
<td><strong>Items</strong></td>
</tr>
<tr>
<td>Moisture</td>
</tr>
<tr>
<td>Volatiles</td>
</tr>
<tr>
<td>Fixed carbon</td>
</tr>
<tr>
<td>Ash</td>
</tr>
<tr>
<td>Sulphur</td>
</tr>
<tr>
<td>Sup. calor. power</td>
</tr>
</tbody>
</table>
from a cold climate flora which had special structural and chemical characters, and they developed in an area of slow epeirogenic movement.

3. In the Leão Mine, the contact between the so-called Lower Suite of coal and the greywacke Pedra Areia was the principal stratigraphic level for correlation. Under that contact, coal beds and Pedra Areia occur over about 100 m (Fig. 20.6). This was proved in the mines. The bottoms of the galleries of the Lower Suite are established on the highest level of Pedra Areia. Presumably, in the first stages of the evolution of the Leão-Butiá deposit, periodic active erosion and redeposition from the surrounding source areas, maybe even in the form of mud flows, controlled the recurring development of Pedra Areia and lenticular coal layers. Once these factors stabilised, peat formation extended across the whole depositional area, giving rise to the above-mentioned Lower Suite of coal, the barren beds representing smaller influxes of thin elastics on the peat flow and temporary subsidence followed allowing the accumulation of clays and of silts with an average thickness of about 6 m, after
which once more the area stabilised, and the Upper Suite of coal was formed. Subsequently the area evolved towards the development of the Palermo Lake producing the transgression of the Palermo Formation across the Rio Bonito.

4. In the large Irui deposit there are also two coal suites similar to those of Leão-Butiá, with the difference that, in many places, the interval of clastic sediments that separates them (also of about 6 m) is occupied by a sandy siltstone of Palermo type. In addition, in other areas of coal occurrences in Rio Grande do Sul, as pioneer drillings proceeded deeper into the sedimentary basin the Rio Bonito became progressively more similar to the Palermo. This suggests that our southern deposits behave as if they were paralic. They are not generally termed paralic because there are characteristically marine beds. However, in the central region of Santa Catarina, intercalated between the Rio Bonito and the Palermo, there are the thin marine layers of the Taio Formation. Only one interpretation of this set of facts has suggested itself. The swamps that originated the southern deposits were situated in low areas marginal to large water masses, precariously isolated from a sea to the southwest, but with possible easy connection through Argentina. While the Rio Bonito lithology was deposited on the margins and coal was formed, the Palermo lithology was deposited in the epi-continental internal sea. An analogy might be found in Lakes Algonquino and Agassis, that resulted from the Pleistocene thawing in the USA, which occupied the same area as the southern region of the Paraná Basin, and could easily have been linked to the sea eastward and southward. Afterwards, by slow transgression, the Palermo also transgressed the Rio Bonito. It is felt that such considerations compelled Gordon (1947) to establish his Guatá Group. It would be surprising if a fluvial-lacustrine facies such as the Rio Bonito was extended uniformly at the same stratigraphical level around the enormous area of the Paraná Basin.

5. There are many doubts about placing the marine Budó Formation of Rio Grande do Sul, correlated for many years with the important unit called Passinho Formation of Paraná State, within the Rio Grande do Sul Itararé Group. The following comments, summarised from Machado (1966), are still relevant:

It is very common in Rio Grande do Sul, to find between the typical lithology of the Rio Bonito Formation and the basement, a sequence of variable thickness (maximum 30.25 m in a drilling of the Irui Deposit), mainly of siltstone or claystone, purple, greenish, or a mixture of both colours. This is intercalated in the lower portion of the column with whitish quartzose sandstone (often feldspathic) and conglomerates with a sandy matrix and often containing pieces of crystalline basement. These rocks are usually placed with the Itararé glacial facies, which would then form the basal part of the Tubarão Group in the area under discussion. The thinner sediments and some heterogeneous conglomerates a few metres thick would be considered fluvio-glacial.

It is the present opinion, however, that not enough is known of the coal-bearing area to judge such deposits as undoubtedly of glacial origin. The few outcrops of the lithology more frequently mentioned in the literature as tillites of Rio Grande do Sul are restricted to basement contacts, and the author has not yet seen in the field a direct contact with the rest of the column of the Tubarão Group. On the other hand, there are hundreds of drillings in which no marine beds of the so-called Budo Facies (correlated with the Teixeira Soares Formation of the Paraná), are found associated with the coal measures. This so-called Budó Facies apparently occurs in small structures faulted or folded into the crystalline basement. We agree, then, with the many authors who stress the need for caution in applying to Rio Grande do Sul the column of the Tubarão Group from Paraná State, as has been proposed by Lange (1954). Support for this view comes from the work of Mendes (1960: 5, 1962: 73), and Delaney and Goni (1963: 6-9).

H. Martin (pers. comm.) believes that these so-called glacial sediments were deposited in preglacial valleys, with relatively high initial dips, and that later the ice compressed these sediments against the walls of the valley. Martin has mapped similar situations in southwestern Africa.

There are therefore two distinct lithologies in Rio Grande do Sul in the Itararé Group:
a. the tillites, varvites and fluvioglacial deposits;
b. the sandy fossiliferous siltstones.

Lange (1954), in Paraná State, referred the glacial strata to the Palmeiras Formation, and the fossiliferous strata to the Teixeira Soares Formation. However, in this State there are three tillites under the fossiliferous strata, and one tillite above them. Therefore, we suggest that, when dealing with these lithologies in Rio Grande do Sul, we should use local names until more exact correlations have been established.

No outcrop in Rio Grande do Sul shows clearly which of the lithologies is the oldest and we therefore consider it preferable to divide the Itararé Group into two facies, instead of dividing it in members or formations. We propose for them the following designations: Suspiro (glacial) and Budó (sandy fossiliferous siltstone). The names are taken from the places where they are best known and developed.

We believe, then, that the Budó Formation of Rio Grande do Sul is in many places older than the Tubarão Group.

6. The foregoing points have to be reconciled with the interfingering observed in the north (Paraná and São Paulo) between the Guatá and the Itararé. Only one explanation seems possible and this has already been proposed by other authors, using different lines of argument (see Mendes, 1962; De Loczy, 1964). The ice mass would have come from the north and thawing would have progressed from south to north. This view is adopted because the greatest thickness of glacial deposits and the greatest number of tillites are in the north. As the result of thawing, large intracratonic lakes were established in the south with paludal environments on their southern margins. The progressive regression of the ice to the north, with some oscillations, meant that the top of the Itararé became younger northwards towards Guatá. As there are occurrences of intertillitic coal in Southern São Paulo, the secondary phases of retreat and advance of the ice had to be relatively fast, but the intervals involved must have been more prolonged than suggested by De Loczy (1964).

7. As we have seen there is no physical continuity between coals in the several deposits. We believe, therefore, that it is meaningless, for example, to search in the Leão-Butiá deposit for the bed corresponding to the Barro Branco or Irapuá of Santa Catarina. The only satisfactory approach is to correlate the sections using palynological and petrographical methods.

8. The coal seams mined in southern Brazil present characteristic regularities in internal structure that remain constant over large areas. This is possible only in autochthonous

Table 20.14. Brazilian Production of R.O.M. Coal (tons)

<table>
<thead>
<tr>
<th>Years</th>
<th>R. G. Sul</th>
<th>Santa Catarina</th>
<th>Paraná</th>
<th>S. Paulo</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>1925</td>
<td>305,682</td>
<td>85,197</td>
<td>—</td>
<td>—</td>
<td>391,879</td>
</tr>
<tr>
<td>1930</td>
<td>335,739</td>
<td>45,409</td>
<td>3,000</td>
<td>—</td>
<td>385,148</td>
</tr>
<tr>
<td>1935</td>
<td>689,200</td>
<td>150,888</td>
<td>—</td>
<td>—</td>
<td>840,088</td>
</tr>
<tr>
<td>1940</td>
<td>1,065,488</td>
<td>265,638</td>
<td>2,733</td>
<td>2,402</td>
<td>1,336,301</td>
</tr>
<tr>
<td>1945</td>
<td>1,139,858</td>
<td>815,678</td>
<td>98,342</td>
<td>19,002</td>
<td>2,072,881</td>
</tr>
<tr>
<td>1950</td>
<td>834,758</td>
<td>1,005,174</td>
<td>98,717</td>
<td>—</td>
<td>1,958,649</td>
</tr>
<tr>
<td>1955</td>
<td>948,297</td>
<td>1,325,512</td>
<td>74,903</td>
<td>—</td>
<td>2,348,712</td>
</tr>
<tr>
<td>1960</td>
<td>646,646</td>
<td>1,439,434</td>
<td>74,534</td>
<td>—</td>
<td>2,160,232</td>
</tr>
<tr>
<td>1966</td>
<td>731,206</td>
<td>2,451,775</td>
<td>187,439</td>
<td>—</td>
<td>3,308,420</td>
</tr>
<tr>
<td>1967</td>
<td>925,888</td>
<td>3,097,300</td>
<td>315,599</td>
<td>—</td>
<td>4,338,787</td>
</tr>
<tr>
<td>1968</td>
<td>995,543</td>
<td>3,489,543</td>
<td>344,504</td>
<td>—</td>
<td>4,827,590</td>
</tr>
<tr>
<td>1969</td>
<td>1,065,861</td>
<td>3,706,728</td>
<td>414,762</td>
<td>—</td>
<td>5,187,351</td>
</tr>
<tr>
<td>1970</td>
<td>965,010</td>
<td>3,944,777</td>
<td>361,888</td>
<td>—</td>
<td>5,271,673</td>
</tr>
</tbody>
</table>

Sources: CPCAN reports.
Ed. da Eletrobrás.
IBGE.
Table 20.15. Consumption of Coal in Brazil (%)

<table>
<thead>
<tr>
<th>Uses</th>
<th>National steam coal</th>
<th>National metallurgical coal</th>
<th>Foreign metallurgical coal</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thermoelectricity</td>
<td>53.37</td>
<td>97.62</td>
<td>—</td>
</tr>
<tr>
<td>Metallurgy</td>
<td>—</td>
<td>—</td>
<td>86.97</td>
</tr>
<tr>
<td>Railroad</td>
<td>41.29</td>
<td>2.10</td>
<td>—</td>
</tr>
<tr>
<td>Navigation</td>
<td>2.86</td>
<td>2.00</td>
<td>—</td>
</tr>
<tr>
<td>Gas production</td>
<td>—</td>
<td>—</td>
<td>13.03</td>
</tr>
<tr>
<td>Industry, several purposes</td>
<td>2.48</td>
<td>0.28</td>
<td>—</td>
</tr>
<tr>
<td>Total (tons)</td>
<td>821,564</td>
<td>1,564,357</td>
<td>455,193</td>
</tr>
</tbody>
</table>

Sources: Rev. 'Carvão de Pedra', CPCAN, IBGE.

Table 20.16. Coal Power Plants in Southern Brazil (Data of 1967)

<table>
<thead>
<tr>
<th>Power plant</th>
<th>State</th>
<th>Production (Mwh)</th>
<th>Capacity (MW) Actual</th>
<th>Proposed capacity to 1973</th>
</tr>
</thead>
<tbody>
<tr>
<td>Candiotas</td>
<td>RS</td>
<td>89,071</td>
<td>20</td>
<td>126</td>
</tr>
<tr>
<td>Charqueadas</td>
<td>RS</td>
<td>252,735</td>
<td>72</td>
<td>72</td>
</tr>
<tr>
<td>S. Jeronimo</td>
<td>RS</td>
<td>126,559</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Capivari (CSN)</td>
<td>SC</td>
<td>84,090</td>
<td>24</td>
<td>24</td>
</tr>
<tr>
<td>Sotelca</td>
<td>SC</td>
<td>183,182</td>
<td>100</td>
<td>232</td>
</tr>
<tr>
<td>Figueira</td>
<td>PR</td>
<td>65,920</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Pres. Vargas (Klabin)</td>
<td>PR</td>
<td>154,383</td>
<td>27</td>
<td>27</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td>955,941</td>
<td>283</td>
<td>521</td>
</tr>
</tbody>
</table>

REMARKS: In 1966 the net production (815,660 Mwh) represented 2.59 per cent from the Brazilian total (32,654,137 Mwh) and 28.2 per cent from the south Brazil region (3,000,073 Mwh).

Source: CPCAN.

PRODUCTION AND USES OF THE COAL

Tables 20.14, 20.15 and 20.16 show details of the production and uses of the Gondwana coal of southern Brazil.

In conclusion, two important facts should be noted. In 1973, the Aços Finos Piratini S.A. in Rio Grande do Sul State, close to the Charqueadas Mine, will begin operations to produce special steel by the SL/RN process of direct reduction. Also in Ibituba (Santa Catarina), the Indústria Carboquímica Catarinense S.A., will use pyritic wastes from the washing of the Santa Catarina coal to produce sulphuric acid and phosphoric acid.

REFERENCES


Houiller Beige.


Coalification Trends in the Sydney Basin, New South Wales

C. F. K. DIESSEL

ABSTRACT

Improvements in instrumentation have made it possible to obtain accurate reflectance values from very small particles, so that coalification studies can be extended beyond coal measure sediments into any strata that contain small dispersed plant fragments (phytoclasts). Examples of such studies are given in this paper.

Sampling has been done in three stages in order to conduct photometric analyses, 1. in different stratigraphic horizons at equal topographic elevation, 2. in similar stratigraphic horizons at different depth of burial, and 3. down the section in deep wells. The results show that there is little correlation between coal rank and stratigraphic position. Hunter Valley samples taken at about sea level, and ranging from the base of the Permian to Middle Triassic, fluctuate about 0.8 per cent mean maximum reflectance (oil), which indicates largely post-tectonic coalification.

Ro-depth factors (rate of coalification expressed in reflectance increase per 100 m thickness of overburden) of individual coal seams have been compared with those obtained from nearby boreholes. They support the general notion that during coal measure sedimentation the northern margin of the Sydney Basin was tectonically unstable. Orogenic movements started soon after the deposition of the Greta Seam so that about 90 per cent of its coalification took place after deformation. In the nearby Lake Macquarie Syncline as well as in the South-Western Coalfield approximately 50 per cent of the coalification happened before tectonic movements ceased while in the Southern Coalfield over 70 per cent of the coalification was pre-tectonic.

Since coalification is a function of temperature and time, Ro-depth factors differ with geothermal gradients and the diagenetic maturity reached by the coal. In low rank coal the increase in rank with increasing depth of burial is small compared with the higher coalification gradients observed in high rank coals. Any comparison of Ro-depth factors must therefore be based on similar rank intervals. Some data are presented which illustrate regional variations in Ro-depth factors down to 1 per cent mean maximum vitrinite reflectance over parts of the Sydney Basin. They range from 0.028/100 in the Upper Hunter Valley to 0.073/100 south of Sydney, and it is thought that higher than usual heatflows, possibly due to igneous activity, are responsible for the high values.

INTRODUCTION

Since D. White published the carbon ratio theory (1915), which states that the gravity of mineral oil varies inversely as the carbon ratio of any associated coal, analyses of coal and carbonaceous matter have been successfully employed in oil exploration. Their main contribution has been the definition of a basal hydrocarbon deadline below which petroleum is unlikely to occur, and only re-
cently it was demonstrated by Hacquebard and Donaldson (1970) that failure to check the siting of oil bores in relation to the position of the hydrocarbon deadline can lead to unnecessary expenditure.

The increasing application of coal analytical methods to oil prospecting has not only brought economic benefits to the industry but the gathering of rank data from depths commonly not reached by coal bores has greatly extended our knowledge of the processes and causes of coalification.

As long as coal rank determinations had to rely entirely on chemical methods, the application of the carbon ratio theory was limited to strata containing coal seams so that sufficiently large samples could be obtained from the drill cuttings or sidewall cores. With the development and perfection of microscope photometres a new tool in rank determination has been made available that is based on reflectance measurements. Hoffmann and Jenkner (1932), Seyler (1941, 1952), Stach (1949), Mackowsky (1950), Dahme and Mackowsky (1950), Wege (1954), Huntjens and van Krevelen (1954), McCartney and Hofer (1955), Murchison (1958), Köttler (1960), and others have conclusively shown that the reflectance of coal increases with increasing carbon and decreasing volatile matter content. Because of the rapidity with which the measurements can be made, and because they can be applied to single macerals, reflectance measurements convey a more accurate picture of the rank of most coals than can be obtained by more conventional analytical methods based on multi-maceral samples. Since modern instruments are capable of recording reflectance values in extremely small particles the occurrence of coal seams is no longer a prerequisite for rank analyses. Minute coalified plant fragments (phytoclasts), as they occur in most marine and many terrestrial sediments, constitute quite adequate material for accurate rank determination, thus enabling the investigator to broaden the spectrum of coalification studies far beyond the boundaries of common coal measures.

The purpose of the present paper is the presentation and interpretation of coalification data obtained from several deep oil exploration bores in the Sydney Basin (see locality map in Fig. 21.1). Most of them penetrated Triassic and Permian strata including fluvial, lacustrine, paludal, deltaic and marine sediments. Any restriction of the reflectance measurements to coal seams would have severely curtailed the scope of the studies; however, small phytoclasts up to 0.01 mm in size were found in many strata, and in the bluish-grey sandy mudstones of the marine Permian (Dalwood and Maitland Groups) they have proved to be quite ubiquitous. The largely terrestrial Triassic sediments, particularly the Hawkesbury Sandstone, are on the other hand often devoid of phytoclasts. Instead, they contain sporadic small stringers of coalified cortical tissues associated with coalified tree trunks. When found in bore cores or sampled in outcrops, such material represents a telinite concentrate; however, because of syn- and post-Mesozoic weathering, reflectance values tend to fluctuate. No such material is usually recovered from percussion bores.

**METHODS OF INVESTIGATION**

The reflectance measurements have been carried out in oil immersion on polished sample blocks by using a Carl Zeiss-MPM-Photometer mounted on a Zeiss-Photomicroscope. Commonly a magnification of 800× has been employed, the area measured being 2.5 microns in diameter. Plane polarised light of a wavelength of 546 nm has been used for all photometric analyses. The readout has been calibrated by several standards of known reflectance.

The number of measurements per sample has varied according to the availability of suitable material. In samples obtained from coal seams up to sixty readings of maximum and minimum vitrinite reflectance have been obtained by rotating the microscope stage a full circle. When, however, the measurements were taken on phytoclasts fewer readings were possible. An example of how the number of measurements per sample varies with depth and stratigraphy is the reflectance curve obtained from Esso Howes Swamp No. 1 Bore shown in Figure 21.2. On account of their small size and particulate nature, phytoclasts present some problems...
concerning vitrinite identification. In a normal coal slide, maceral diagnosis is usually accomplished by comparison of several physical properties in addition to reflectance, such as relative hardness, colour, and the degree of gelification. Such determinations are difficult when macerals are segregated in a rock matrix, and the difficulty increases with coal rank. It is thought that the larger than usual spread of sample means in the upper portion of Figure 21.2 is probably due to wrong maceral identification, for example, some massive micrinite might have been mistaken for vitrinite. Examples of phytoclasts are shown in Plate 21.1, 1-2.

Another source of variation in the results is related to sample types. Apart from a few surface samples most of the material studied has come from diamond drill cores and percussion chips. The latter display a slightly wider spread of data, probably as a result of contamination from occasional cavings of borehole walls.

The sampling procedure did not follow any fixed pattern but was governed by the availability of suitable material. Measurements of the reflectance of vitrinite only have been made, and care has been taken to further restrict the measurements to the structured form of vitrinite, i.e. telinite, when dealing with low rank coals. Such a procedure has been found to be essential since in sub-bituminous and high volatile bituminous coals noticeable differences in reflectance values occur between the various forms of vitrinite, the variety collinite usually yielding lower values. In high rank material these differences disappear because of the convergence of the coalification band.
Fig. 21.2. Reflectance versus depth diagram based on coal and phytoclast samples obtained from drill cuttings of an oil exploration well. The figures alongside the curve indicate the number of measurements per sample.

Pl. 21.I-2. Phytoclasts of micrinite (left) and vitrinite (right) in sandy mudstone. Branxton Formation, Maitland Group, reflected light, 600x, oil immersion.
Fig. 21.3. Relationship between subsidence, duration of burial, temperature and degree of coalification shown by two coals formed in a cratonic shelf setting. A = Upper Miocene, U.S.-Gulf Coast; B = Westphalian, North-West German Basin. Modified after M. and R. Teichmüller (1966b).

Fig. 21.4. Comparison of coalification curves (depth versus mean maximum vitrinite reflectance in oil) between several Australian sedimentary basins. A = North-West Shelf, B = Cooper Basin, C = Southern Bowen-Surat Basin, D = Sydney Basin (open circles refer to northern portion, full circles to southern portion). Data in A to C after Shibaoka, Bennett and Gould (1973).
COMPARISON OF COALIFICATION GRADIENTS IN SOME AUSTRALIAN COAL BEARING BASINS

The work of many authors (e.g. Patteisky and M. and R. Teichmüller (1962), M. and R. Teichmüller (1966a, b), Kisch (1968), Teichmüller (1970, 1971), Hacquebard and Donaldson (1970), Damberger (1971)) and others, has revealed considerable variations in the rates at which coalification has proceeded in different parts of the world. A number of reasons have been proposed for such differences; however, the works of Huck and Karweil (1953, 1955), Huck and Patteisky (1964), Karweil (1965, 1966) and Damberger (1968) have left no doubt that once the rank of lignite has been passed, temperature and the duration of exposure to such temperatures are the principal agents of coalification. An illustration of this relationship is shown in Figure 21.3, in which some features pertaining to the post-depositional history of two coals have been contrasted with one another. The two coals have been formed in similar geological environments—mobile shelf on cratonic margins. Both have been obtained from similar depth, namely, 5100 m and 5440 m respectively and both were subjected to similar end-temperatures, such as 147°C and 141°C, thus suggesting similar geothermal gradients. The two coals differ, however, in age and degree of coalification. One comes from the Carboniferous of the Northwest-German Basin and is a meta-bituminous coal of 16 per cent volatile matter, while the other is a high volatile bituminous coal of 37 per cent volatile matter, coming from the Upper Miocene of the U.S.-Gulf coast (after M. and R. Teichmüller, 1966b). In view of the tectonic similarity of the post-depositional history of both coals, the higher rank of the German coal must be related to the longer period of time it had been exposed to the same end-temperature.

Similar variations in coalification gradients can be found in Australia. Shibaoka, Bennett and Gould (1973) have published coal reflectance data obtained from a number of Australian sedimentary basins, and some of their results have been incorporated in Figure 21.4. As in the example of Figure 21.3, some of the rank differences at equal depth are the result of age differences, but others are not. For instance, the lowest coals of the Northwest Shelf (Fig. 21.4A) yield reflectance values of less than 0.6 per cent at a depth of 3000 m. This is considerably less than the 1.0 per cent found in the Cooper Basin (Fig. 21.4B), the 2 per cent in the southern Bowen-Surat Basin (Fig. 21.4C), or the 3 per cent in the southern portion of the Sydney Basin (Fig. 21.4D), all obtained from the same depth. Although it could be argued that the Jurassic coals of the Northwest Shelf are approximately 100 m.y. younger than the others and therefore comparatively immature, this would not explain the differences between the basal coals of the other basins since they all come from Permian strata. Moreover, there are considerable differences even within one basin. An indication of this is given in Figure 21.4D by the different slope angles of the curves referring to the northern and southern portions of the Sydney Basin, respectively. In view of the age equivalence of the sediments analysed, and keeping in mind that coalification is a function of temperature and time, the different coalification gradients obtained can only be explained as a result of different geothermal gradients.

COAL RANK IN RELATION TO STRATIGRAPHIC POSITION AND DEPTH OF BURIAL

In order to test the effect of time and tectonic deformation on the development of coal rank, two sets of photometric analyses have been carried out. In one series of reflectance measurements samples of different geological age but of similar present-day topographic position (i.e. at or just above sea level) were used, and in another series a group of samples from the same stratigraphic level but variable depth of burial was used. The results of the tests are graphically displayed in Figures 21.5 and 21.6.

The stratigraphic section in Figure 21.5 cuts transversely across the Lower Hunter Valley and extends into the Carboniferous rocks of the southern extension of the New England Fold Belt. Superimposed on their respective stratigraphic positions are the reflectance values obtained from near surface samples. The results indicate that there is very little proportional rank increase with
Fig. 21.5. Stratigraphic section through the northern portion of the Sydney Basin and adjacent New England Fold Belt. Superimposed are reflectance data obtained from coal and phytoclast samples in near surface positions. Stratigraphic stratigraphic depth in the Triassic and Permian sediments of the Sydney Basin. The samples fluctuate between Ro max. = 0.72 and 0.82 per cent. There is a slight tendency towards the higher value in the lower units except for the Greta Coal Measures which, in spite of their low stratigraphic position, show a surface reflectance of only 0.65 per cent. This means that coalification occurred largely after the Lower Hunter Valley portion of the Sydney Basin had been tectonically differentiated into the controlling Lochinvar Anticline and Lake Macquarie Syncline because a pre-tectonic coalification would have resulted in a greater systematic differentiation of reflectance values between successive stratigraphic units. Only in the Carboniferous of the New England Fold Belt north of Newcastle do the reflectance values obtained from surface samples vary markedly. They show a strong high between Gresford and Carsons ville, reaching over 6 per cent mean maximum reflectance (oil). However, these differences are likewise not related to stratigraphy but to metamorphic zoning (Offler, 1973; Diessel, 1973).

It is shown in Figure 21.6 that coalification could not have been exclusively post-tectonic either. The trend lines displayed demonstrate that rank increases with respect to depth of burial in several well-known coal seams representative of the major coalfields of the Sydney Basin. Most of the reflectance values displayed have been obtained from measurements in outcrop, mine and bore samples, supplemented in the case of the Bulli Seam by data published by Cook and Wilson (1969).

Correlation between the degree of coalification and depth of burial of individual
seams is quite good, minor fluctuations probably being related to variations resulting from the wide spacing between some sample points within the respective coalfields. Most interesting is the result that the highest reflectance values have been measured in the topographically elevated coals from the southern part of the Sydney Basin, whereas in the north similar ranks are encountered only at great depths. An example is the Greta Seam for which a reflectance of 1.24 per cent was obtained from a depth of almost 1500 m in the Planet East Maitland No. 1 Bore, whereas similar and higher values can be measured

Fig. 21.6. Reflectance versus depth curves obtained from measurements in individual coal seams. Full circles = Greta Seam; open circles = Borehole Seam; squares = Wongawilli Seam; triangles = Bulli Seam.
in outcrop samples of the Bulli Seam along the escarpment above the city of Wollongong.

In the case of a completely post-deformational coalification one would expect coals that occur at similar topographic levels within a basin to have similar ranks. This is not so as is shown in Figure 21.6. In fact coalification proceeded further in the southern portion of the Sydney Basin than in equivalent positions in the north, either because of previously deeper subsidence and subsequent higher elevation or because of a higher heat-flow possibly due to widespread igneous activity in that area. Even so, tectonic tilting must have post-dated a large part of the coalification process.

**COALIFICATION IN RELATION TO TECTONISM**

The question of the relation of coalification to tectonic deformation can be answered with the aid of Ro-depth factors. These are measures of coalification gradients, and according to Hacquebard and Donaldson (1970) they indicate the rate of rank increase in percent reflectance per unit depth, commonly 100 m. The Ro-depth factor can be used as a tool for the solution of a number of geological problems, independently of the actual degree of coalification. By comparing Ro-depth fac-

---

**Fig. 21.7. Reflectance versus depth curves measured down the hole of some deep wells in the Sydney Basin.**
1 = A.O.G. Dural South; 2 = Esso Howe's Swamp; 3 = Swansea; 4 = Esso Jerry's Plains; 5 = Planet East Maitland; 6 = Alkae Terrigal; 7 = A.O.G. Kirkham.
Coalification Trends in the Sydney Basin

Tors obtained from analyses of individual coal seams, e.g. those in Figure 21.6, with coalification gradients measured down boreholes, such as shown in Figure 21.7, the relative timing of coalification can be quantified. If, for example, all the coalification suffered by a group of coal seams happened before the seams were folded, the iso-rank lines would appear folded too, and each coal seam would retain its rank independent of its post-tectonic depth of burial (Fig. 21.8A). Stratigraphic differences would show up as rank differences, and the highest coalification gradient would be obtained from bores sunk along crest or trough lines of anticlines or synclines. The Ro-depth factors of individual coal seams would be nil. An actual example approaching this situation is the Carboniferous coal measures of Westphalia in Germany. From the data published by Patteisky and Teichmüller (1960) and Patteisky et al. (1962), an overall Ro-depth factor of 0.085 per cent per 100 m can be calculated for the Ruhr Basin. This is a very high figure when compared with individual coal seams. The Sonnenschein Seam, for example, has a gradient of 0.004 per cent per 100 m down to a depth of 1000 m below sea level and to a maximum reflectance of 1.6 per cent. The large difference of 0.081 per cent between the Ro-depth factors can only mean that in the coals which are currently up to 1000 m deep, about 96 per cent of the rank increase took place before tectonic deformation while the remainder happened syn- and post-tectonically. The difference between the two Ro-depth factors decreases at greater depths because coalification is still in progress at elevated temperatures.

Complete post-tectonic coalification is probably very rare because it would involve very rapid deposition and folding. According to M. and R. Teichmüller (1966a) iso-rank lines in such a case would not parallel folded strata but would be horizontal as shown in Figure 21.8B. Examples of combined pre- and post-tectonic, as well as syn-tectonic coalification are more common. In such instances the iso-rank lines bulge up in anticlines and are slightly depressed in synclines. The Saar Basin in Germany approaches a situation of post-tectonic coalification, and from the data published by Damberger et al. (1964) and M. and R. Teichmüller (1966b) the Ro-depth factors of individual seams can be calculated as 0.021 per cent per 100 m, while the general coalification curve for the Saar Basin yields a figure of 0.026 per cent per 100 m. In an ideal case of post-tectonic coalification there should be no difference between the two values because both coalification curves should coincide; however, the difference of 0.005 per cent in the above example is very small and indicates that only 19 per cent of the rank increase was accomplished before tectonism began.

Fig. 21.8. Two schematic examples illustrating the relationship between coalification and tectonic deformation. A = pre-tectonic coalification; B = post-tectonic coalification.
A comparison between reflectance versus depth curves obtained from individual seams in the Sydney Basin (Fig. 21.6) and down boreholes (Fig. 21.7) reveals differences in slope angles. On correlating Ro-depth factors of individual seams with those measured down the hole of nearby bores, the calculation shown in Table 21.1 can be made.

The most obvious result of the calculation is the striking difference between the northern and southern coastal portion of the Sydney Basin, and it appears likely that such variations are due to different temporal relationships between tectonic deformation and coalification. It seems that around the Lochinvar Anticline in the north about 90 per cent of the coalification took place after tectonic movements were well advanced or had almost terminated, and in order to have remained a sub-bituminous coal for over 250 m.y. the low-rank portion of the Greta Seam must have stayed high on the flank of the developing anticline. This concept of early tectonism is supported independently by Stuntz (1972) who considers the Lochinvar Anticline as having been a growing structure since Upper Permian time. A sharp increase in uplift occurred during the deposition of the Tomago Coal Measures, the older parts of which are unconformably overlain by younger members of the same group around the flanks of the anticline. At the beginning of the Triassic much of the Permian Newcastle and Tomato Coal Measures, and probably older sediments, were eroded from the crest of the rising anticline, which like most northern portions of the Sydney Basin were tilted against the uplifting New England Fold Belt. This leaves barely 10 m.y. for the coalification of the coals outcropping near the surface around the flanks of the Lochinvar Anticline, which explains their low rank.

In the Newcastle Coalfield, situated in the Lake Macquarie Syncline, the difference is 52.6 per cent between the Ro-depth factors based on individual seams and down the hole, whereas in the Maitland/Cessnock area, on the flank of the Lochinvar Anticline, the difference between the two gradients is only 9.5 per cent. These figures are in good agreement with the differences in the respective geological settings which would call for an earlier cessation of coalification on a growing anticline than in a subsiding syncline.

During Triassic time, when the north was already dragged up against the border thrusts (Hunter Thrust System) separating the Hunter Valley Fore-Deep from the New England Fold Belt, the central and southern portions of the Sydney Basin were still subsiding. The connection between the basin and the sea was probably not far from the present Southern Coalfield at Wollongong (Diessel and Moelle, 1970) as is indicated by the convergence of palaeocurrents in that area. Tectonic deformation remained weak compared with the Hunter Valley and coalification was well advanced (according to Table 21.1 over 70 per cent of it was completed) before any substantial movements occurred. They took the form of an upwarp of the southern basin margin which, on the basis of palaeocurrent studies, can be placed as Middle to Late Triassic (base of Hawkesbury Sandstone),

Table 21.1. Comparison of Ro-depth Factors measured in Individual Coal Seams and 'Down the Hole'

<table>
<thead>
<tr>
<th>Bore/Seam</th>
<th>Ro-depth factor in % per 100 m</th>
<th>Difference in Ro-depth factor</th>
<th>Per cent diff. to bore values</th>
</tr>
</thead>
<tbody>
<tr>
<td>E. Maitland No. 1</td>
<td>0.042</td>
<td>0.004</td>
<td>9.5</td>
</tr>
<tr>
<td>Greta Seam</td>
<td>0.038</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Swansea bores</td>
<td>0.056</td>
<td>0.029</td>
<td>52.6</td>
</tr>
<tr>
<td>Borehole Seam</td>
<td>0.027</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kirkham No. 1</td>
<td>0.061</td>
<td>0.023</td>
<td>52.4</td>
</tr>
<tr>
<td>Wongwilli Seam</td>
<td>0.038</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Woronora No. 1</td>
<td>0.110</td>
<td>0.081</td>
<td>73.6</td>
</tr>
<tr>
<td>Bulli Seam</td>
<td>0.029</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
thus leaving at least 20 m.y. longer than in the uplifted portions of the Hunter Valley for coalification to proceed without substantial interruption.

VERTICAL AND LATERAL VARIATIONS IN COALIFICATION GRADIENTS

Reflectance versus depth curves, such as those in Figures 21.2, 21.4, 21.6 and 21.7, display a variety of different shapes. However, in most cases they are steep near the top and flatten in the lower portion. It follows that increments in reflectance are small in the initial stages of coalification followed by some later acceleration which causes a downward increase in Ro-depth factors. An analysis of the change of Ro-depth factors of a number of reflectance versus depth curves has been made in terms of reflectance intervals of 0.2 per cent. The results are graphically displayed in Figure 21.9. They show a wide scatter which is not surprising since local data have been combined in the diagram with those taken from the literature, e.g. Hacquebard and Donaldson (1970), thus representing a variety of environments with a wide range of geothermal gradients. In spite of the scatter it is obvious that a definite trend from low to high Ro-depth factors exists. This parallels increments in coal rank as expressed in mean maximum vitrinite reflectance. Any comparison of Ro-depth factors between different bores or regions must therefore be based on similar rank intervals. As an example Figure 21.10 illustrates regional variations in Ro-depth factors, down to 1 per cent mean maximum vitrinite reflectance in oil, over some portions of the Sydney Basin. Although more sample points would be needed in order to present a detailed picture the pattern obtained supports the previous suggestion of higher coalification rates near the coast south of Sydney. It is thus reasonable to assume that the high rank coals found in the central and southern portion of the Sydney Basin are the result of both high geothermal gradients, possibly due to igneous activity in that area, and, compared with the Hunter Valley, longer duration of coalification.

CONCLUSIONS

The results obtained support the notion that the tectonic history of the northern and southern portions of the Sydney Basin differed quite considerably. During the Permian a fore-deep was situated between the then rising New England Fold Belt to the northeast of the Hunter Valley and the broad shelf margin of the older central and southern fold belt to the southwest. The consequence of this geological setting was early differentiation into large anticlines and synclines near the orogenic margin, and most of the coalification took place after deformation. In the central and southern portion of the Sydney Basin coalification was influenced by higher heat flows, probably due to igneous activity, and it is also likely that coalification lasted longer.

These conclusions have been based on a geological interpretation of reflectance measurements of coals and phytoclasts, and it is thought that a wider application of the methods described, e.g. construction of regional phytoclast reflectance maps, would open new avenues for coal petrological activity that could elucidate a wide range of geological problems.
Fig. 21.10. Regional variation of Ro-depth factors down to 1 per cent Ro maximum (oil) in the Sydney Basin.
ACKNOWLEDGMENTS

The project was supported by a grant from the Australian Research Grants Committee. This is gratefully acknowledged. Some reflectance measurements were carried out by Miss K. Jackson while Mr E. Krupic undertook the not always easy task of preparing the samples in accordance with the high standards of quality required for photometric analyses. The manuscript was typed by Mrs J. P. Odgers. These three members of the Department of Geology, The University of Newcastle, are thanked for their work. Samples and some bore logs were made available by the New South Wales Geological Survey, Department of Mines, and Esso Australia Ltd.

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Section 4

Age and Stratigraphic Relations of Glacial Deposits
ABSTRACT

In Australia, late Palaeozoic mountain glaciation commenced in the Namurian or Westphalian, and was followed quickly by continental glaciation that culminated in the Stephanian or early Sakmarian. Ice caps occupied many parts of Australia, and contributed debris to inland basins, as well as to margins of both the Tethys Sea and the Tasman Geosyncline. In Antarctica, lobes of large continental glaciers, probably sited over the East Antarctic Shield, contributed debris to three basins along the present Transantarctic Mountains. Africa was widely glaciated from Gabon and the Congo Basin in the north to the tip of the continent in the south and lobes probably even reached eastward into then-attached Madagascar. Locally subglacial topographic relief in Africa was considerable, suggesting that ice caps grew on several upland regions and flowed outward as lobes, probably at different times at different places. In South America, ice caps developed east of the Paraná Basin and in Uruguay, less certainly over the Asunción Arch, to the south of the Chaco-Paraná Basin and west of the Sauce Grande Basin. Mountain glaciers along the western border of the continent waxed in mid-Carboniferous times, and in Bolivia, for example, disappeared before the Permian. In the Falkland Island region a subcontinental ice cap lay to the west, perhaps with an ice shelf extending eastward to open waters. In Peninsular India several elongated basins received glacial debris from ice centres between basins which on the west and northwest were possibly of subcontinental extent. Lobes of glaciers reached northward to dump sediment into the margin of the Tethys Sea in regions now in Pakistan and along the flanks of the Himalayan Mountains. Possible Laurasian glaciation in the Kazanian has recently been recognised in northeastern Siberia.

Reassembly of the Gondwana continents and comparison with interpretations from palaeomagnetic studies indicate that glaciation ensued on upland regions available to cold precipitation as the supercontinent moved across the south rotational pole. The glaciation waxed in the Carboniferous when the pole was over the Transvaal and waned in the mid-Permian when it moved from Antarctica into the palaeo-Pacific.

INTRODUCTION

No event in Earth history is more dramatic than the widespread glaciation in Gondwanaland during the late Palaeozoic. Many scientists have now added details to reconstructions of late Palaeozoic world geography and climatology during the past century and a quarter. Here this knowledge is reviewed briefly with an emphasis on problems still remaining.

Only by clarifying such geologic history can we approach understanding of several
fundamental matters. Among these are the causes of changing climates, the role of changing arrangements of land and sea on the directions and strengths of flow of ocean and air currents, and upon the pattern of continental drift. Perhaps by fitting all these interlocking aspects of moving continents and changing climates together we can also better understand evolutionary trends among plants and animals, and clarify matters of palaeobiogeography. However, study of these significant tasks has barely begun.

THE DISTRIBUTION OF CONTINENTAL GLACIERS IN THE LATE PALAEOZOIC

Evidence of continental ice sheets that waxed and waned during the Carboniferous and Permian periods is preserved on all southern continents and in Asia. The evidence consists mainly of remnants of erosional features such as striated glacial pavements and glacial valleys on the one hand, and of a sedimentary record on the other. Sediments laid down as till beneath melting glaciers or in zones near their margins often display features diagnostic of glaciation but those farther and farther away show their initial glacial origin less and less as the debris is reworked in the glacio-marine environment. In addition, large dropstones that were carried by icebergs are at places recognised with confidence but their sources are not necessarily close at hand; icebergs may float long distances in ocean currents before melting and dumping debris on the sea floor. In reconstructing ancient geography during glacial times, sedimentary facies are placed in a sensible arrangement and checked against all available directional-current properties in the sediments, including striae on the basement floor. In these studies palaeoecological interpretations from fossils and thickness changes are also most helpful.

The most serious difficulties in reconstructing ancient glacial geography arise, not only from the scattered distribution of outcrops which only locally are reasonably complete, but also from the difficulty of obtaining precise stratigraphic correlations both regional and intercontinental. Moreover, glaciation is primarily an erosional process, leaving but a scant record; sediments deposited nearby in bulk are on the whole laid down by other processes so that the trace of glaciation grades through stages of subtlety to non-existence. Where the stamp of glaciation upon the strata is weak it may not be possible to distinguish between distant mountain (alpine) glaciation and continental ice sheets. In addition tillite lying upon a scoured and striated basement floor may record a late stage of the glaciation as the glaciers retreated and there may be no way to determine when the glaciation began at the locality. Under these circumstances, debris from the region was eroded and perhaps transported far away to be deposited as beds with no imprint of glaciation whatsoever. In fact, a typical facies model includes a glaciated shield with only a patchy veneer of tillite. Related deposits peripheral to the region beneath the glacial centre may be devoid of tillite remnants. Away from the ice centre the sediments lying directly upon the striated floor can be expected to be progressively older and older. The oldest glacial sediments in such a reconstruction will lie in marine sections off the glaciated continent and interbedded with sediments of other origins. The recognition and documentation of the glaciation may be most difficult in such areas because ice occupied a region hundreds or thousands of kilometres distant, and transportation processes between the ice centre and the depositional centre may have obliterated nearly all vestiges of the glaciation. It is in this marginal prism, however, that the palaeontological control may be most complete and the dating of the strata most satisfactory. As the glaciation waned, deposits transgressed upon the shield centre, and those laid down in patches near the centre and most likely to be preserved may also record only the latest stages in the glacial episode. Whereas most of the deep erosion beneath the ice centres no doubt took place during the waxing and culmination of the glaciation, the striae and patches of tillites observable to us today may document only the last ice retreat.

During the past eight years we have examined representative stratigraphic sections of all of the major remnants of late Palaeozoic glacigene sediments except those in Si-
Late Palaeozoic Glaciation

Australia

The late Palaeozoic glaciation has left its mark on many parts of Australia, from the Bonaparte Gulf Basin to Tasmania and from central Queensland to the Perth Basin (Fig. 22.1). Striated floors are clearly preserved here and there, but especially in the Fleurieu Peninsula of South Australia, in the Bacchus Marsh area of Victoria, and in Tasmania. In these regions, the striae show incontrovertibly that glaciers lay upon hilly terrain and moved irregularly toward the north and northeast. Elsewhere in Australia the glacial impress is recorded in sediments—at places strongly recorded and at others very in-
definitely. As the data come in, however, facies arrangements are documenting the details of the glacial geography. Along the western margin of the Tasman Geosyncline in southern New South Wales, for example, tillite interpreted as ground moraine passes eastward into outwash debris coursed with fluviatile channels, and in turn into normal non-marine and marine sediments. Many of these latter sequences, and those in Tasmania as well, contain impressive dropstones; and in the region of marine strata, fossils afford the best palaeontological correlations available in any Gondwana continent. Most of the Australian locations, including many boreholes, where glacigene sediments and striated pavements have been reported, are shown on maps in Crowell and Frakes (1971a, 1971b).

The distribution of glacial sediments throughout much of inland Australia is less well known, but as boreholes are drilled, and geological and geophysical exploration continues, useful information is accumulating. In northern South Australia and adjoining regions, subsurface basins contain diamicrites occasionally with striated and facetted stones. Rare exposures of similar strata are scattered about in this region, and in the Bonaparte Gulf, Canning, Carnarvon, and Perth basins, mostly in Western Australia. The South Australian remnants are interpreted as consisting of glacial debris carried from glacial centres into intracratonic depositional sites. Although at some places the original glacial imprint has been almost obliterated by non-glacial transportation processes, the evidence is sufficient to require ice centres on terrains marginal to basins, but direct evidence of ancient ice caps upon these regions has long since been eroded away. On the south, however, this interpretation is powerfully supported by the convincing glacial topography and pavements, with remnants of tillite still preserved in the Fleurieu Peninsula, central Victoria, and Tasmania. Although many of the strata containing a glacial imprint are non-marine, a few marine intercalations show that a chain of sea connections crossed the continent (Garratt, 1969; Thomas, 1969; Ludbrook, 1969; Wopfner, 1970). In addition, these studies show that the glaciation was well-developed during the Sakmarian.

In Tasmania, tillite lies upon striated pavements within valleys primarily on the west and extends mainly southeastward to a fossiliferous marine section (Banks, 1962). Fossils in the latter section, and rare *Rhaeto*pterus* and microflora interbedded within tillite, afford an unusual opportunity for regional correlation. The data show that an ice sheet occupied western Tasmania during the Late Carboniferous, and tillite deposition ceased in the region in the very Early Permian (M. J. Clarke, pers. comm.). Dropstones in overlying beds, exposed mainly in eastern Tasmania, indicate that floating bergs reached the region until late Kungurian or early Kazanian time (middle Late Permian).

In southern New South Wales, fossiliferous marine beds with dropstones occur. In this region, coarse sedimentary breccia and deformed sandstone bodies within sandy diamicrite lie just east of an ancient shoreline. Inland of this depositional site, palaeodrainage patterns have recently been worked out by Herbert (1972), who described channels or valleys filled with up to 230 m of sediment, mainly conglomerate. These are convincingly interpreted as fluvial channels that carried glacially derived debris from western glaciers downslope and eastward to the sea. Inasmuch as these channels immediately underlie earliest Permian strata they are considered by Herbert to be latest Carboniferous in age. These deposits, along with remnants of glacial terrain and tillites in the Mudgee region, document ice centres to the west and northwest at this time. The coastline was apparently irregular with offshore bars and islands, and at places glacial valleys or fjords may have reached eastward. There is little convincing evidence bearing on the extent westward from the shoreline of this ice sheet, or on the height and ruggedness of the terrain. A large ice cap may have existed, however, since glacial debris also reached into basins of northeastern South Australia (Cooper Basin) and it seems unlikely that local alpine glaciers would have had sufficient size. Although the evidence is scant we tentatively favour an ice cap over northwestern New South Wales. north of a low swale with marine incursions through north-
Late Palaeozoic Glaciation

In the Hunter Valley region, Tamworth Trough, New England, and central Queensland, interpretation of the glacial record is especially difficult. In this part of the Tasman Geosyncline deformation and sedimentation went on together during much of the Carboniferous and Permian, so that the structure and stratigraphy are complicated. Since only very minute parts of the immense stratigraphic thicknesses preserve a glacial imprint, in a region of inferred high mountains not far away, we prefer a hypothesis involving local alpine and valley glaciers rather than continental ice sheets or ice caps. Beds containing this faint glacial imprint are as old as Namurian or Westphalian, based on stratigraphic position, the presence of *Lepidodendron* and *Rhacopteris*, and a radiometric age (298 m.y.) obtained from interbedded Paterson Toscanite (Evernden and Richards, 1962; but see Harland et al., 1964). In writing our papers a few years ago (Crowell and Frakes, 1971a, 1971b) we tentatively concluded that the inferred alpine glaciation of northern New South Wales was largely Stephanian and Westphalian in age, and perhaps extended backward in time into the Namurian. Only in central Queensland (Joe Joe Formation) did the alpine glaciation last into the Permian (Campbell and McKellar, 1969). At that time of writing, glacials west of the Sydney Basin were considered as only earliest Permian in age so that a pause and change in glaciation type was suggested, from a waning mountain glaciation in the Stephanian to a waxing continental glaciation in the Sakmarian, over some 5 to 8 m.y. Late Carboniferous mountains with alpine glaciers here and there were pictured as largely worn down in New South Wales before the onset of continental glaciation encroaching from the southwest. According to recent work (e.g., Herbert, 1972; Gostin and Herbert, 1973), the continental glaciation may be latest Carboniferous. This suggests that continental glaciation in the south was partly contemporaneous with alpine glaciation in the north. Nonetheless, data in hand suggest that the mountains may have been highest, and most subject to alpine glaciation, somewhat earlier than the culmination of continental glaciation. Solution to this difficult palaeogeographic problem must await additional tectonic and stratigraphic work.

For 2200 km along the Tasman Geosyncline from central Queensland to Tasmania, Permian strata at places contain large lone-stones interpreted as glacial dropstones. Locally they have fallen upon mats of fossils growing in place, and reach sizes up to 2 m. Many are isolated within laminated units and thus could not have been transported laterally by a swift current, nor have they tumbled from a steep cliff close at hand. They are therefore envisaged as having fallen from wasting icebergs, but it is unknown just how far the icebergs might have drifted from their place of calving before dropping their load. Nor is it known whether they were born at the snouts of alpine or valley glaciers, or at the seaward edge of a continental ice shelf. All that can be stated with confidence is that icebergs originated somewhere up-drift as late as the Kazanian (middle Late Permian). Although the Permian icebergs may have come from Antarctica, considered as then attached to Australia, they might also have been contributed from Australian sources where alpine, valley, piedmont, or ice cap glaciers reached to the sea locally. Perhaps in time geologists will be able to match some of the distinctive dropstones with their source terrain as knowledge progresses and careful palaeogeographic and palaeotectonic reconstructions are undertaken.

Antarctica

A piecemeal record of late Palaeozoic glaciation is scattered over about 2200 km along the Transantarctic Mountains, and in the northern Sentinel Range (Fig. 22.2). In such a vast region with so few outcrops pro-
Fig. 22.2. Late Palaeozoic glaciation in Antarctica. Modified from Frakes, Matthews and Crowell (1971). Locality ‘B’ from Barrett (1972) and locality ‘PMH’ from Pinet, Matz and Hayes (1971).
truding through the present ice cap it is difficult to reconstruct the glacial geography; nonetheless a picture is coming to light (Lindsay, 1970; Frakes, Matthews, and Crowell, 1971). In the Pensacola Mountains over 300 m of diamictite overlies Devonian (?) sandstone and underlies Glossopteris-bearing strata (Permian). The diamictite is replete with deformed sandstone bodies, suggesting mobility of the sediments when still soft, wet, and unfrozen. Several striated boulder pavements, soft-sediment grooves resembling drag marks, and palaeocurrent indicators in associated beds (such as ripple marks, sole marks, and cross stratification), suggest that ice flow was in general toward the south-southwest but palaeocurrent flow towards the northeast is indicated higher in the section.

Because the stratigraphic section is similar in the Pensacola Mountains and the Sentinel Range the two areas are considered as possibly parts of the same basin. The areas are, however, about 600 km apart, and perhaps lie in different tectonic provinces at present so that this generalisation may not be justified. Within the Sentinel Range are about 1100 m of cleaved metadiamictite containing a few stratified intercalations with dropstones, and lying within a section with palaeocurrent indicators showing flow from south to north. On this evidence alone it is questionably inferred that the glacial front lay to the south. The thick section of diamictite containing deformed sandstone bodies may have been deposited in a depression marginal to glaciers but no fossils have been recovered to show that the site was marine or non-marine, nor to give the age.

Many striated pavements, diamictites with faceted and striated stones, and interbedded laminated sequences with dropstones (varves?) have been recorded from within the Transantarctic Mountains from the Ohio Range northwestward as far as the Darwin Glacier area (Fig. 22.2). Small patches of tillite have been recently described at the north by Barrett (1972) and by Pinet, Matz and Hayes (1971). Differences in facies, thicknesses, and inferred flow directions of the tillite and associated strata from place to place throughout the Transantarctic Mountains suggest a varied geography. Broad divides capped by ice probably separated shallow basins or embayments in the continental margin in which the derived sediments accumulated. For example, an ice centre presumably lay south of the Sentinel Range, in the vicinity of the Thiel Mountains, and may have contributed debris westward to the Ohio Range and eastward toward the Pensacola Mountains. Early ice flow within the Pensacolas, however, seems to have come from still farther north, in the region of the present Weddell Sea.

Another sedimentation basin apparently lay in the vicinity of the Horlick Mountains; the Ohio and Wisconsin Ranges contain tillites derived from the Queen Maud Mountains, where tillites are exceptionally thin. Still another accumulation occupied the area of the Queen Alexandra Range, but there is no direct evidence that this depositional site was directly connected with the Horlick region. Still farther northwest, near the Darwin Glacier, are additional tillitic remnants. Since most of these glacial sequences thin toward the west, it is inferred that continental glacial centres lay still farther in that direction, over the East Antarctic Shield, but evidence is lacking to decide whether one immense ice sheet occupied the region as today, or whether there was a series of smaller centres.

The age and correlation of the Antarctic glacial deposits are not well known, primarily because of the lack of fossils and the sparsity of outcrops. Since in each region the glacial sequence is overlain conformably by plant-bearing Permian beds, and Early Permian spores have been recovered from glacial beds in the Ohio Range (Kemp, this volume) and the Darwin Mountains (Barrett and Kyle, this volume), it is concluded that glaciation lasted into the Permian. The age of the oldest glacial strata is much less well known, however, and rests mainly on stratigraphic considerations. The beds in each region lie disconformably upon Devonian rocks, and in the Pensacola Mountains the presumed Upper Devonian beds were apparently still soft so that they were easily scribed by overriding glaciers. Stratigraphic relations in the Pensacolas as well as in the Sentinel
Range therefore suggest a Carboniferous onset of glaciation, and permit a start in the Early or Middle Carboniferous. The thinner glacial sequences lying on consolidated rocks at other localities in the Transantarctic Mountains seem consistent with a somewhat younger onset of glaciation northwest of the Horlick Mountains, either Late Carboniferous or Early Permian. In conclusion, we find the record in Antarctica weakly suggestive of migration of glacial centres across the continent, beginning in the Pensacola-Sentinel region in the Middle Carboniferous and in the northwestern Transantarctic Mountains in the latest Carboniferous or earliest Permian. Better stratigraphic control is sorely needed, however.

Africa

In South Africa and South West Africa the record of glaciation is especially complete (Martin, 1961; Stratten, 1969; Frakes and Crowell, 1970a; Matthews, 1970; Crowell and Frakes, 1972). Ice carrying large amounts of debris entered the Karroo Basin from the northwest, north, northeast, and east (Fig. 22.3) as shown by the distribution of Dwyka Tillite, many striated pavements, and scoured glacial valleys. The Dwyka Series (including both tillite and Upper Dwyka Shales) reaches a maximum thickness of more than 3000 m in an east-west trending trough just north of the outcrop band in the southern Cape Ranges. Here the formation lies conformably upon shales and quartzites of the Witteberg Series, the uppermost subdivision of the Cape System. Plant remains from the lower Witteberg and underly the Bokkeveld Series are considered to be Devonian by Plumstead (1969), the Bokkeveld invertebrates are Early Devonian (Wolfart and Voges, 1968) and fish remains from the Upper Witteberg as Early Carboniferous (Viséan) by Gardiner (1969). Beds immediately and conformably underlying the Dwyka Tillite in this region are therefore tentatively assigned to the Lower Carboniferous, but they could well be latest Devonian in age. Similar plants have also been recovered from intercalations within tillites at the northern margin of the Karroo Basin (Plumstead, 1969: 33). Plants immediately overlying the glacial sequence are considered to be uppermost Carboniferous or lowermost Permian (Plumstead, 1970). Karroo microfloral studies place these latter beds a bit higher, at the beginning of the Permian (Hart, 1969).

In Botswana and Southwest Africa tillitic remnants and glacial striae outcrop, and boreholes suggest that glacial sediments underlie a large part of the Kalahari Desert. A glacial lobe is therefore inferred to have moved from the African interior toward the southwest, and to reach into a marine environment as shown by fossils such as Eurydesma and Peruvissipra. These fossils, along with Mesosaurus found stratigraphically about 200 m above the glacial strata, suggest that the Botswana lobe was of Early Permian age. Basal glacial beds, however, lie well below the fossils, as also does the basal tillite of the Namaland lobe, suggesting that glaciation in this region began in the Carboniferous. The Namaland lobe (du Toit, 1921) is the older, south-flowing lobe recorded mainly in the Orange River Basin and characterised by many red clasts of the Precambrian Fish River Series.

In northern South West Africa glacial valleys up to 1500 m deep contain remnants of tillites, striae, and roches moutonnées (Martin, 1961). The subhorizontal glacial beds abut the steep striated valley walls and both currents and ice flow were from east to west down the gradient of the valleys, now being exhumed. Ice, assigned to the Kaokoveld lobe, moved through a mountain range from an ice sheet centred farther east. It reached to the site of the present Atlantic Ocean and probably into Brazil before the fragmentation of Gondwanaland. The width of the Kaokoveld lobe is unknown and its northern limit is shown arbitrarily in southern Angola in Figure 22.3.

Far to the east, near the common borders of Zambia, Rhodesia, and Mozambique, Dwyka Beds are preserved in a long block, tipped northward against a system of normal faults, and in several other small patches (Bond, 1967). Along the southern boundary, strata locally contain glacial beds and lie depositionally upon Precambrian basement. These glacial beds occur only in the mid-Zambezi region and contain striated and
Fig. 22.3. Late Palaeozoic glaciation in Africa. Modified from Crowell and Frakes (1972), Frakes and Crowell (1970a), and with addition from Micholet, Molinas and Penet (1970).
faceted stones and possible varves; only one striated pavement has been reported. Tills in this wide region were apparently laid down discontinuously in hollows between hills carved in the basement either by valley glaciers or by small ice caps (Chappell and Humphreys, 1970). Their age is unknown except that they belong to the Dwyka Series of the Karroo System on the basis of regional stratigraphic relations. Overlying coal beds, correlated with the Middle Ecca Series of South Africa (Lower Permian), contain a Glossopteris flora (Gair, 1959).

Far to the north, along the eastern margin of the Congo Basin in Zaire, tillite is preserved within glacial valleys and lies at several places on a striated floor. The distribution of presumed glacial debris, which includes thick deposits interpreted as flow-tills recovered from cores within the basin to the west, suggests that a series of mountain glaciers coalesced into a piedmont ice sheet along the western base of a low mountain range. Glossopteris remains and spores and pollen suggest a Late Carboniferous-Early Permian age. Far to the west in Gabon, diamictites with large striated blocks and varved (?) sequences occur at the base of the Karroo System, and in the absence of fossils, are correlated with the Dwyka (Micholet, Molinas and Penet, 1970), on regional relations alone.

In southwestern Madagascar conglomerates and diamictites contain rare striated and faceted stones, and striated and polished floors with roches moutonnées have been reported (Besairie, 1961). Palaeocurrent structures in beds near the unconformity with the Precambrian basement indicate flow toward the northeast, presumably away from a distant area of glaciation. Beds about 1000 m higher stratigraphically contain brachiopods indicating a Middle Permian age, so the glacial beds are probably at least Early Permian in age, and may be Late Carboniferous.

In looking at Africa as a whole we know that ice sheets existed in the Permo-Carboniferous interval. Lobes extended peripherally from centres of ice accumulation and the record is most complete at the downflow ends of these lobes. The extent and limits of the main ice centres that provided the ice to flow into the lobes is quite obscure, however, and we may not ever be able to map their extents because the record has been eroded away. Enough patches of glacial debris remain in the Zambia-Rhodesia region to conclude that the whole of central Africa from the Transvaal to Tanzania was glaciated, but it is not known yet whether all ice action was simultaneous. Several centres of accumulation may have occupied this broad region, and may have occupied different places at different times. On the north it seems likely that mountain glaciers rather than continental ice sheets lay in the region bordering the Congo Basin, both on the east and on the west near Gabon, but whether these coalesced with each other or to the south with still other ice sheets is not known. The age of the glaciation in Africa may include most of the Carboniferous if we accept the sparse data from the Cape Ranges. Glaciation ceased early in the Permian throughout the whole region according to present palaeontological interpretations. Refined age studies and additional fossils may resolve some of these problems.

South America

Geomorphic evidence of glaciation is unequivocally preserved along the eastern margin of the Paraná Basin in Brazil and a glacial imprint is discernible in many sedimentary sequences of late Palaeozoic age elsewhere in South America and the Falkland Islands (Fig. 22.4). Intracratonic basins received glacial debris from ice caps near or beyond their margins and Carboniferous mountains, inferred to contain glaciers, shed large volumes of sediment to sites now situated along the eastern flanks of the Andes (Frakes and Crowell, 1969).

Glacial valleys with striated surfaces show that ice flowed westwards into the Paraná Basin (Martin, 1961; Rocha-Campos, 1967, 1972; Frakes and Crowell, 1969, in press; Landim, 1972). Sediments within the basin here include probable eskers, ice-deformed units, periglacial features, glacio-fluvial outwash, and laminated lacustrine units (varves?) with striated and faceted dropstones. Three major ice lobes reached into the Paraná Basin from the east and south
Late Palaeozoic Glaciation

and may have coalesced near the basin centre at times of maximum advance. These lobes are reconstructed on the basis of glacial striae, subglacial topography, stone provenance, and basin isopachs, i.e., greatest thicknesses mark the axes of the lobes (Landim, 1972; Frakes and Crowell, in press). The northern lobe (Fig. 22.4) is probably the Brazilian end of the Kaokoveld lobe of South West Africa (Martin, 1961). In fact, facies interpretations suggest as many as five major advances and retreats of the ice lobes. An intermittent seaway at times lay west of these ice fronts and extended southward to a proto-Atlantic Ocean through the Sauce Grande region to the junction between the Falkland Islands and South Africa, and thence to the palaeo-Pacific Ocean.
Along the western margin of the Paraná Basin diamictites contain stones derived from the Asunción Arch and facies and palaeocurrents suggest reworking of glacial deposits from a western centre. By analogy with sedimentary facies on the eastern margin it is reasonable to infer glaciation upon the Asunción Arch, but striae and glacial topography have not been noted. The presence of an ice cap there, however, is supported by interpretations in the Bolivian Basin, westward beyond the arch (Helwig, 1972). Tillites containing striated and faceted stones are coarsest in the southeastern part of the Bolivian Basin and were probably derived from an ice cap on the Asunción Arch. In addition, palaeocurrent investigations and petrographic and radiometric studies of granitic stones suggest that the Pampean Ranges to the southwest contained mountain glaciers and contributed significantly to the basin. The glacial sequence in the Bolivian region is Middle Carboniferous in age and lies stratigraphically well below fossiliferous marine strata of the earliest Permian. Glaciation in this region had ceased by the end of the Carboniferous.

Southwest of the Asunción Arch and directly connected with the Paraná Basin lies the Chaco-Paraná Basin, mainly a subsurface feature (Padula and Mingramm, 1969). Upper Carboniferous diamictites are correlated with similar beds in the adjoining Paraná Basin but have not yet been studied in detail. Padula and Mingramm (1969, fig. 4) show a concentration of glacial deposits at the southeast and lithofacies and thickesss suggest derivation of sediments primarily from the Uruguayan Shield and the Pampean Ranges. On the west the basin connects through a narrow passage to the Paganzo Basin in which fluvial deposition dominated (Azcuy and Morelli, 1970). Conglomerates with rare striations and diamictites may be reworked tills, but a glacial contribution to this basin is not well founded.

During the Carboniferous and Permian periods marine waters from the Pacific reached across the site of the present Andes and deposited sediments in western Argentina. In the Calingasta-Uspallata and central Patagonian basins, Carboniferous strata contain diamictites and laminated shales with isolated stones up to 40 cm in diameter. Mountains along the proto-Andean chain apparently contained glaciers, but the extent of the glaciers both temporally and spatially is unknown. In the Sauce Grande area (Fig. 22.4) about 1000 m of massive diamictite, now deformed and metamorphosed to the greenschist facies, contains faceted and striated stones and is interpreted as glacial debris that has moved downslope from a glaciated area, probably to the southwest (Coates, 1969). In the Falkland Islands, facies on the west contain massive diamictite with probable esker and subglacial fans and huge blocks up to 8 m in diameter (Frakes and Crowell, 1967). Toward the east these facies grade into diamictite with small stones and with many disrupted sandstone bodies, and still farther east, with interbeds that include dropstones in graded thin units. An ice centre lay to the west and apparently contributed debris to an eastern deepening basin, perhaps beneath a wasting and floating ice shelf. Unfortunately the stratigraphic control in the Falkland Islands is poor, and the age of the glacial sequence is known only as Carboniferous to Early Permian.

In summary, the glacial record in South America and the Falkland Islands allows us to conclude that ice caps of at least subcontinental dimensions occupied eastern Brazil, Uruguay, and southern Patagonia, and with somewhat less confidence, the Asunción Arch, areas south of the Chaco-Paraná Basin, and regions southwest of the Sauce Grande Basin. In addition, proto-Andean and Pampean Mountains contained glaciers of unknown size. The glaciation began in Middle Carboniferous times, and waned before the Permian in Bolivia and very early in the Permian elsewhere, but better chronostratigraphic control is needed.

Asia

In the northern hemisphere the record of glaciation during the late Palaeozoic is found in India, the Salt Range region of Pakistan, and in a belt along the Himalayan Mountains (Fig. 22.5). These are all parts of Gondwanaland, considered as now separated and displaced northward from other Gond-
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Fig. 22.5. Late Palaeozoic glaciation in India and countries adjacent to the north. From Frakes, Kemp and Crowell (in prep.), including data from Ghosh and Mitra (1969).
wana continents. Recently, in addition, glacio-marine deposits of Late Permian age have been described from Siberia (Mikhaylov et al., 1970), and are presumably the record in Laurasia of the same ice age which has not yet been recognised elsewhere on that supercontinent.

In Peninsular India glacial sediments are preserved in a series of elongate depressions and lie unconformably upon Precambrian basement at the base of the Gondwana sequence (Fig. 22.5; Frakes, Kemp and Crowell, in prep.). Basin-margin faults are interpreted by Ghosh and Mitra (1969) as having moved during deposition of at least the beds that immediately overlie the glacialis. Tectonic deformation probably occurred during glaciation, depressing the sedimentation sites, and elevating the marginal regions. The Gondwana beds contained within these depressions consist of glacial, fluvial, paludal, and lacustrine beds with a few marine intercalations (Robinson, 1967; Ghosh and Sastry, 1967). The glacial unit—the Talchir Formation—has a maximum reported thickness of 300 m and is composed of diamictite (both tillite and tilloid), conglomerate and sandstone (probably fluvial), varve-like beds with and without dropstones, and thin graded sandstone beds in shales (probably turbidites). In the Pranhita-Godavari region there is an undoubted striated floor (Smith, 1963). Some subrounded stones, with rare but definite striae and facets can be matched with local terrains but many have presumably come from distant unknown sources. Bedded diamictites, soft-sediment folds, intraformational breccias, and deformed sandstone bodies show that some facies were subaqueous in origin, and moved downslope into deeper water, and others were glacio-fluvial in origin (Ghosh and Mitra, 1969). In Figure 22.5 some of the directional data, primarily from till fabrics, erratic trains, and a few striated pavements, are simplified from Ghosh and Mitra (1969).

From the Salt Range of Pakistan to northern Assam diamictites and associated dropstone shales outcrop at many places along the southern belt of the Himalayan Mountains. Although the tectonics along the range are exceptionally complex, most interpretations place these exposures within the Gondwana sequence related to Peninsular India, and consider the beds as thrust beneath a Tethyan sequence originally at the north (Gansser, 1964). In the Salt Range diamictites contain boulders up to 1.2 m in length, many faceted and some striated. These are associated with pebbly shale units and with mudstones containing disrupted sandstone bodies, as well as with other facies suggesting a glacio-fluvial, glacio-lacustrine, or glacio-marine environment. An ice centre on the east probably contributed debris to a depositional environment on the west (Teichert, 1967; Frakes, Kemp and Crowell, in prep.). The Agglomeratic Slates of Kashmir, north of the Salt Range, are reported to have a glacial imprint but need much further work (Gansser, 1964: 58).

The Blăini Boulder Beds, southeast of the Salt Range within the complex Himalayas, contain diamictite with striated stones (Gansser, 1964) but their correlation with the Talchir Formation of Peninsular India is on the basis of lithologic similarity only (Pascoe, 1959). The interpretation of these deposits is still controversial (e.g., Rupke, 1968). Late Palaeozoic strata suggesting glaciation also occur along the Kosi River in Nepal and in the Rangit Valley in Sikkim, in Bhutan, and farther eastward (Jacob, 1952).

The age of the glacial beds in the Salt Range is probably Late Carboniferous because a marine Sakmarian fauna (perhaps also latest Carboniferous) occurs just above them (Teichert, 1970). The fauna is similar to those in both eastern and western Australia. Moreover, a still higher marine fauna in this section is also Artinskian. The oldest plant remains from Peninsular India consist of spores and pollen from shales near the base and from siltstone beds overlying basal Talchir glacial in the Damodar region (Lele, 1964; Lele and Karim, 1971). Megaplant fossils in the Talchir shales are dominated by Gangamopteris with also Noegegerathiospis, Cordiacarpus, and Samaropsis. Glossopteris comes in only at the uppermost part of the Talchir section. Palynological studies in progress suggest correlation with forms commonly considered Sakmarian in Australia.
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(Frakes, Kemp and Crowell, in prep.). In summary, data in hand indicate that glaciation from several subcontinental centres of unknown shape, size, and extent waxed in the Stephanian, and waned by the Sakmarian; this suggests that glaciation in southern Asia was perhaps somewhat later than in other Gondwana continents except Australia. The extent of the ice caps is reconstructed by many to spread over much of northeastern India on the basis of regional considerations, under the premise that most of the Indian record has long since been eroded away and preserved sections lie only around their periphery (e.g., Robinson, 1967). The questionable glacial deposits, which may be largely marine, strung out along the Himalayan chain, are pictured as being at the northern rim of large conjectural ice sheets, some of which had floating ice shelves. There is no evidence to support the hypothesis that a single large ice cap extended from then-attached Africa, across Madagascar, to southeastern India; it seems more reasonable to advocate a more local ice cap of subcontinental size entirely within Peninsular India. In addition, ice did not directly adjoin ice centres in western Australia as a Late Palaeozoic seaway probably lay between these subcontinents.

In the Verkoyansk region, east of the Lena River in northeastern Siberia, Mikhaylov and others (1970) have described up to 1000 m of pebbly siltstones and siltstones with limy layers of sparse pebbles. They attribute the deposits to clast rafting by shore ice in a marine environment. Brachiopod faunas both above and below are assigned to the Kazanian. If these beds document glaciation not far away, we may have a record of Late Permian glaciation in Laurasia. On the other hand, they may document rafting by shore ice only without glaciation, or deposition by some non-glacial process. This possible glaciation would be at a time somewhat later than any in Gondwanaland except for the evidence from dropstones in southeastern Australia.

**DISCUSSION**

The piecemeal glacial record for the Gondwana continents as described briefly above is placed on an assembly of the continents in Figure 22.6. The arrangement is modified somewhat from previous schemes in order to close in the gap between Africa, Antarctica and the Falkland Islands, particularly in accord with the reconstruction of Ford (1972: fig. 2). It is premature, however, to attempt a serious reconstruction of this tectonically complicated junction. India, with Ceylon unrealistically shown in its present relation to the peninsula, is located only approximately, but until more is known about the geology along the coast of East Antarctica, correlations are highly speculative. In Figure 22.6, note that glacial centres, as yet undocumented, are postulated during the late Palaeozoic along the Tethyan coast of East Antarctica. Orogenic belts line up reasonably satisfactorily and glacial centres are roughly compatible between continents. The magnetic polar-wander path is simplified from McElhinny (1973: fig. 136).

The pattern of glaciation is shown without regard to timing, which is known to range from the Middle Carboniferous into the Late Permian—say, from at least within the early Westphalian to the Kazanian—an interval of about 70 m.y. In Australia the meagre chronostratigraphic data suggest that it began early and lasted longer, perhaps because of proximity to the Tethyan seaway. Unfortunately the timing of the beginning and of the end over Antarctica and much of Africa is not yet adequately known. Nonetheless, data in hand support the hypothesis of du Toit and many others (du Toit, 1921; King, 1962; Ahmad, 1966; Crowell and Frakes, 1970a, 1970b) that major glacial centres migrated as the clustered Gondwana continents glided across the south rotational pole.

For continental glaciation to ensue, not only is a near-polar position of a continental mass required, but also suitable sources for evaporative moisture and properly positioned uplands. Both the palaeo-Pacific and Tethys oceans provided the prime source, but there was also open water, at times marine, from the Paraná Basin southward along the junction between Africa and South America, to the vicinity of the Falkland Islands, Pensacola Mountains, and Sentinel Range and thence to the palaeo-Pacific Ocean. As the
strung-out Gondwana continents glided across the pole, oceanic and air currents in the subtropical and temperate zones were probably deflected southward so that relatively warm water and moist air moved up across the land and across a southern polar front (Frakes and Crowell, 1970c: figs. 6, 7 and 8). Cloudiness and precipitation, mostly as snow on broad uplands or mountains beneath the polar front, probably caused ice centres to grow. Still farther south, polar easterlies and the corresponding oceanic drift may have carried icebergs northward along the Antarctic-Australian coast; the icebergs, upon melting, dumped their debris as dropstones as far north as central Queensland. A major palaeoclimatological problem, however, is to find moisture sources for the vast glaciation over much of Africa south of the present equator. The patchy record in this wide region suggests, however, that ice caps, albeit large ones, occupied different places at different times rather than one tremendous ice sheet covering the whole region in the
Late Carboniferous.

The late Palaeozoic glaciation apparently waned rather quickly in mid-Permian times when the Gondwana supercontinent moved away from the south rotational pole. Although palaeomagnetic data are still sparse, McElhinny (1973: fig. 136) shows Mesozooic poles to be well away from the continent. Perhaps the supercontinent had rotated so that it lay in a more latitudinal than meridional orientation, and as a consequence did not deflect ocean and air currents in the same way. The geographic causes of the beginning of the glaciation in the Carboniferous are far less clear, and in fact there may have been but a slight waning of glaciation after the Ordovician-Silurian episode so clearly recorded in the Sahara (Beuf et al., 1971; Crowell and Frakes, 1970a, 1970b).

Better documentation of the timing of the late Palaeozoic glaciation through all types of correlation studies is very much needed. In particular, special studies are required where nearly complete stratigraphic sections lie upon datable older beds without significant breaks. Such sections are likely to occur around the margins of late Palaeozoic continents such as in the Cape Ranges of South Africa where stratigraphic arguments as evaluated at present suggest the possibility that glaciation began in the Early Carboniferous. Other places where the onset of glaciation may have left a record are in basins of Western Australia, the Tethyan zone of northern India and Pakistan, in the Falklands-Pensacola-Sentinel sector, and in the Tasman orogenic belt. Regional correlation studies along with careful facies reconstructions of limited stratigraphic intervals will certainly clarify the glacial geography as well.

REFERENCES


The Early Permian Glacial Beds of South Victoria Land and the Darwin Mountains, Antarctica

PETER J. BARRETT AND ROSEMARY A. KYLE

ABSTRACT

The Late Palaeozoic glacial beds of the Transantarctic Mountains comprise the lowest formation of the Victoria Group (Permian-Triassic), and disconformably overlie the Taylor Group (Devonian and older(?)) of the Beacon Super-Group. In the Darwin Mountains they include tillite, sandstone and shale, some of which is locally slumped. In south Victoria Land the glacial beds are mainly tillite, though some sandstone and shale is preserved in two Late Palaeozoic glacial valleys. Thickness of the glacial strata ranges from 15 to 82 m and from 0 to 50 m in the two areas. Clasts in the tillite in both areas are mainly granitic, but form only one per cent of the rock. Between one- and two-thirds of the tillite has a matrix of chlorite and sericite, and the remainder is mainly quartz, much of which was derived from the quartzose sandstones of the underlying Taylor Group. Glacial striations, asymmetric ripples and a clast fabric indicate a palaeo-ice flow to the southeast in the Darwin Mountains. The only set of striations reported from south Victoria Land (Mount Metschel) also trends southeast.

The glacial beds are underlain by the Maya Erosion Surface, which is cut in Mid to Late Devonian strata in south Victoria Land, and in inferred Devonian strata in the Darwin Mountains. The glacial beds in both areas are overlain by the Pyramid Erosion Surface and Glossopteris-bearing Permian coal measures, but the change in the proportions of clast lithologies, from granite-dominated (like the tillite) at the base of the coal measures, to vein quartz-dominated a few metres higher, suggests that the Pyramid Erosion Surface does not represent a major time break. Direct evidence for the age of the glacial beds comes from a plant microfossil assemblage in a shale sample from Colosseum Ridge, Darwin Mountains. The assemblage is comparable with the Early Permian Stage 2 assemblage of Evans (1969).

A plant microfossil assemblage dominated by Gangamopteris in a shale that grades down into tillite at Mt Fleming, south Victoria Land, also indicates an Early Permian age for the youngest glacial beds. It is suggested that the glacial beds of south Victoria Land and the Darwin Mountains are a record of the final retreat of the late Palaeozoic ice sheet, and of deposition of the entrained glacial debris under wet base conditions in Early Permian times. This does not preclude the existence of an extensive dry base ice sheet in the Carboniferous, or of wet base glaciation for which the sedimentary record was destroyed by the last major advance of the ice.
INTRODUCTION

Our purpose is to review the recent work on the glacial beds at the base of the Victoria Group (Beacon Super-Group) of south Victoria Land and the Darwin Mountains, and to present new evidence on their age, character and distribution. Before 1960 the chief peculiarity of the Antarctic Gondwana sequence (Beacon Super-Group, Table 23.1) was the apparent absence of beds representing the Permo-Carboniferous glaciation. Then between 1960 and 1963 Upper Palaeozoic tillites were recognised by different expeditions at a number of localities along the Transantarctic Mountains and in the Ellsworth Mountains (Long, 1962, 1965; Grindley, 1963; Doumani and Minshew, 1965; Haskell et al., 1965; Schmidt et al., 1965; Craddock et al., 1964). Ironically the glacial beds were to be discovered last in the type area of the Beacon Super-Group of south Victoria Land (Fig. 23.1). Pebbly mudstone believed to be tillite was first reported there by Pinet, Matz and Hayes (1967) from Mt Feather and Mt Fleming, where it occurs as lenses on the Devonian-Permian contact. Subsequently it was found to be more extensive, and has been named the Metschel Tillite (McKelvey et al., 1970, 1972).

METSCHEL TILLITE

Lithology, Distribution and Thickness

The type section of the Metschel Tillite is at the east end of Mt Metschel, an isolated nunatak in the Skelton Neve (Figs. 23.1 and 23.2). The formation consists entirely of massive light greenish-grey, poorly sorted sandy siltstone with scattered clasts up to 0.6 m across. Most of the clasts, however, are less than 4 cm long, and they form only about one per cent of the rock. Table 23.2 shows that there is no great variation in clast lithology with size or with stratigraphic position. About 70 per cent are granite, and the rest include sandstone, quartzite, schist and acid volcanic rocks.

Thin section examination shows that from one-third to two-thirds of the rock consists of a matrix of sericite and chlorite, and the remainder is moderately sorted quartz sand (Fig. 23.3, Table 23.3). Pinet (1969: 87) had previously noted the lack (less than 3 per cent) of grains other than quartz. The distribution of the non-micaceous grains is approximately log-normal, though the histograms indicate two modes at about 1.8 and 2.8 phi. The coarser of the two represents mainly rounded grains believed to come from Taylor Group sediments. Mirsky et al.

Table 23.1. Stratigraphy of the Beacon Super-Group in south Victoria Land and the Darwin Mountains (after McElroy, 1969; McKelvey et al., 1965; Barrett et al., 1971). Equivalent formations and surfaces are connected by lines.

<table>
<thead>
<tr>
<th>Darwin Mountains</th>
<th>South Victoria Land</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ellis Formation</td>
<td>Lashly Formation</td>
<td>Triassic</td>
</tr>
<tr>
<td>Misthound Coal Measures</td>
<td>Feather Conglomerate</td>
<td>Permian</td>
</tr>
<tr>
<td>Pyramid Erosion Surface</td>
<td>Weller Coal Measures</td>
<td>Asphalt Erosion Surface</td>
</tr>
<tr>
<td>Darwin Tillite</td>
<td>Metschel Tillite</td>
<td>Early Permian</td>
</tr>
<tr>
<td>Maya Erosion Surface</td>
<td>Metschel Tillite</td>
<td>Mid-Late Devonian</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Devonian</td>
</tr>
<tr>
<td>Hatherton Sandstone</td>
<td>Beacon Heights Orthoquartzite</td>
<td>and</td>
</tr>
<tr>
<td>Unnamed sandstone</td>
<td>Altar Mountain Formation</td>
<td>Heimdall Erosion Surface</td>
</tr>
<tr>
<td>Brown Hills Conglomerate</td>
<td>New Mountain Sandstone</td>
<td>older (?)</td>
</tr>
<tr>
<td>Kukri Surface</td>
<td>Kukri Surface</td>
<td>Ordovician and older</td>
</tr>
<tr>
<td>BASEMENT COMPLEX</td>
<td>BASEMENT COMPLEX</td>
<td></td>
</tr>
</tbody>
</table>
Table 23.2. Numbers and Sizes of Clasts in the Metschel Tillite at Mount Metschel. Granitic clasts appear dominant irrespective of size or stratigraphic position.

<table>
<thead>
<tr>
<th>Sample area</th>
<th>$1,\text{m}^2$ 3 m above base</th>
<th>$1,\text{m}^2$ 2 m below top</th>
<th>4 m wide strip from bottom to top</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower size limit</td>
<td>1 cm</td>
<td>1 cm</td>
<td>10 cm</td>
</tr>
<tr>
<td>Maximum clast size (long axis)</td>
<td>4 cm</td>
<td>4 cm</td>
<td>62 cm</td>
</tr>
</tbody>
</table>

**Lithologies**

- Granite: 17
- Schist: 1
- Quartzite: 2
- Other: 3
- Total: 23

Granitic clasts appear dominant irrespective of size or stratigraphic position.
Fig. 23.2. Typical stratigraphic sections for the Metschel and Darwin Tillites. Numbers are samples in the Victoria University of Wellington rock collection referred to in this paper.

(1965) found that the quartzose sandstones of the Taylor Group in the Mt Gran area (Fig. 23.1) have a mean size of from 1.0 to 2.7 phi, and comprise almost entirely subround and round grains. Matz (1968) gives grain size data from upper Taylor Group strata at Mt Fleming that confirm this. The finer mode represents mainly very angular to subangular grains in their first cycle of erosion. Both populations seem to be represented in about the same proportions.

A most significant feature, discovered in the tillite at the west end of Mt Metschel, is a striated and finely grooved surface on a thin

Table 23.3. Grain Size Data for Tillite Samples in Figure 23.3. The proportion of micaceous matrix and the median size for the whole rocks vary considerably, but the rest of the rock (mainly quartz) is in each case a moderately sorted fine sand.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Median (phi)</th>
<th>Micaceous matrix %</th>
<th>Mean* (without micaceous matrix)</th>
<th>Standard deviation*</th>
</tr>
</thead>
<tbody>
<tr>
<td>23289</td>
<td>5.2</td>
<td>52</td>
<td>3.0</td>
<td>1.6</td>
</tr>
<tr>
<td>23287</td>
<td>more than 6</td>
<td>66</td>
<td>2.5</td>
<td>1.2</td>
</tr>
<tr>
<td>23143</td>
<td>3.5</td>
<td>36</td>
<td>2.6</td>
<td>1.2</td>
</tr>
</tbody>
</table>

* Graphic mean and standard deviation in phi units after Folk and Ward (1957).
tillite bed at the base of the formation (Pl. 23.1). The surface is exposed as a narrow ledge and extends for over 10 m perpendicular to the strike of the lineation, but the strike direction (120°-300°) varies only 3°. The striations disappear into the hillside beneath a film 1-2 mm thick of dark greenish-grey siltstone, like the soft sediment striations described by Lindsay (1970: 1163 and fig. 12A) from the Beardmore Glacier area. They indicate the presence of moving temperate ice. No features to indicate the sense of ice movement were evident.

The formation is quite variable in thickness throughout south Victoria Land due to erosion prior to deposition of the overlying coal measures and to irregularities in the subglacial surface (Fig. 23.2). On the south ridge of the big eastern cirque at Mt Fleming the thickness ranges from 0 to 10 m and on the north ridge we found a maximum thickness of 7 m. The lithology is similar to that at Mt Metschel, with scattered granitic clasts in a sandy silt matrix, but here the upper contact is gradational from tillite into carbonaceous shale over 30 cm. Gangamopteris leaves in the shale in effect date the youngest glacial beds (see p. 344).

The northernmost outcrops of tillite are in the floor of the northeast cirque of Shapeless

Fig. 23.3. Cumulative curves and histograms for the grain size distributions of the non-micaceous fraction (mainly quartz) of the Metschel Tillite at Mt Metschel (samples 23287 and 23289) and the Darwin Tillite at Misthound Cirque (sample 23143). The coarser of the two modes represents mainly rounded rederived grains; the finer represents very angular-subangular grains. Data are from long axis measurements off thin sections (about 400 grains for each), and they have been converted to equivalent sieve size by using the transformation of Friedman (1962).
Mountain and on the east face of Mt Bastion. At both localities the tillite is only 3 to 4 m thick, and the only clasts are rare vein quartz pebbles up to 8 cm long. Mudstone with scattered quartz pebbles was noted at Mt Bastion ten years ago (Allen, 1962: 288), but its significance was not realised at the time. Tillite appears to be absent at this stratigraphic level north of the Mackay Glacier (Mirskey et al., 1965).

Water-laid sediments are not common in the formation, and appear to be confined mainly to fill material of two valleys cut below the general level of the subglacial erosion surface. The Kennar Valley occurrence includes folded tillite and varvoid fine sandstone with delicately preserved load features, filling a steep-walled valley cut in Aztec Siltstone (B. C. McKelvey, pers. comm.). Barrett (1972) has described a similar steep-walled valley at Alligator Peak. It is filled in the lower part by 20 m of complexly folded, boudinaged and brecciated tillite, sandstone and conglomerate, with fold axes and lineations trending northeast, parallel to the valley wall. The upper part is a 30 m thick body of massive well-sorted fine sandstone deposited soon after the slumping. The sand body has subsided along the axis of the valley due to compaction of the glacial debris. Barrett (1972) suggested that to have such a contact preserved the valley could not have been ice-free for long or some downslope movement would be evident. Relief of lesser magnitude, 1-3 m, can be seen at the west end of Mt Metschel and at Aztec Mountain (McKelvey et al., 1972: fig. 6). At both places the relief is evident from a steep contact of tillite against a sandstone bed in the Aztec Siltstone. McKelvey et al. (1970) noted that the contact was not as clear to the south. According to Dr P. N. Webb (pers. comm.) this was a reference to the type section, where the tillite appears to grade down into the folded upper 3 m of the Aztec Siltstone. We re-examined the contact, and also found it difficult to locate, because the lower 1 m or 2 m of the tillite are greyish-purple and dusky brown, like the underlying 10 m of claystone. However, the tillite could be identified by the presence of scattered clasts. The contact was traced over about 5 m, and had a local relief of 1.3 m. Folding

Maya Erosion Surface was named by Harrington (1965: 22), before tillite had been found in south Victoria Land, for the surface that cut through the Devonian Aztec Siltstone near Maya Mountain. He believed that the surface resulted from glacial and periglacial erosion because of its stratigraphic position, separating Devonian siltstones from the Permian coal measures. McKelvey et al. (1970) discovered that tillite was quite widespread, though in places it had obviously been removed by post-glacial erosion, and concluded that both upper and lower boundaries of the formation were erosion surfaces. They proposed that the lower glacially formed surface take the name 'Maya', because of Harrington’s belief in the glacial origin of the surface that he named, and proposed the name 'Pyramid' for the post-glacial surface, which locally cuts into Aztec Siltstone, destroying both glacial beds and the surface on which they rest.

The Maya Erosion Surface clearly truncates the Aztec Siltstone at several localities, but at others the nature of the contact is not clear. The contact between horizontal Aztec strata and the slumped valley fill at Alligator Peak is sharp and dips at 40°, indicating some degree of lithification prior to valley cutting. Barrett (1972) suggested that to have such a contact preserved the valley could not have been ice-free for long or some downslope movement would be evident. Relief of lesser magnitude, 1-3 m, can be seen at the west end of Mt Metschel and at Aztec Mountain (McKelvey et al., 1972: fig. 6). At both places the relief is evident from a steep contact of tillite against a sandstone bed in the Aztec Siltstone. McKelvey et al. (1970) noted that the contact was not as clear to the south. According to Dr P. N. Webb (pers. comm.) this was a reference to the type section, where the tillite appears to grade down into the folded upper 3 m of the Aztec Siltstone. We re-examined the contact, and also found it difficult to locate, because the lower 1 m or 2 m of the tillite are greyish-purple and dusky brown, like the underlying 10 m of claystone. However, the tillite could be identified by the presence of scattered clasts. The contact was traced over about 5 m, and had a local relief of 1.3 m. Folding

Lower and Upper Boundaries

The lower and upper boundaries of the Metschel Tillite are the Maya and Pyramid Erosion Surfaces respectively (Harrington, 1965; McKelvey et al., 1970). The length of time that they represent is important in considering the age of the Metschel Tillite. The
also occurs in the upper few metres of the Aztec Siltstone at Mt Ritchie. Both instances may be due to shear stress exerted on the poorly lithified Aztec Siltstone by the movement of Permo-Carboniferous ice. At Shapeless Mountain the contact appears gradational over 1 cm from a light-coloured shaly siltstone into medium grey very fine sandstone with sparsely scattered quartz pebbles. The contact is exposed over only 3 or 4 m and the outcrop requires more careful study. At all other localities visited the contact is sharp and unweathered, and gives no clue as to the time interval that it represents.

The upper boundary of the Metschel Tillite is more obvious than the lower because of the lenses of pebbles and boulders that immediately overlie the Pyramid Erosion Surface. Pinet et al. (1971) interpreted the lenses of clasts as lag deposits from reworking of the glacial debris, and we concur with this view. Where the lenses rest on tillite they normally have a large proportion of granitic clasts. For example, at the summit of Mt Ritchie (Fig. 23.4) a thin basal conglomerate rests on tillite and is dominated by granitic clasts mostly 5-20 cm but up to 1.2 m long. Minor lithologies are vein quartz, schist and acid volcanics, all found in the Metschel Tillite. South of the summit the Pyramid Erosion Surface has cut through the tillite and deep into the underlying Aztec Siltstone. Here the thin basal conglomerate is also dominated by granitic pebbles and boulders. Clasts in the Weller Coal Measures above the basal conglomerate are mostly vein quartz, presumably concentrated by weathering away of less resistant lithologies and by sorting in outwash or post-glacial streams. However, a few granitic clasts can be found at most localities in the 10 m above the basal conglomerate. Where post-glacial erosion has entirely removed the glacial beds from a large area, as at Mt Crean and Portal Mountain, a few granitic clasts, up to 10 and 70 cm long respectively, are scattered through the lenses of quartz pebbles at the base of the Weller Coal Measures. That granitic clasts do survive to be deposited above the basal conglomerate at some localities suggests that the time represented by the Pyramid Erosion Surface is small compared with that represented by the Weller Coal Measures.

The observed local relief on both surfaces is quite small. That for the Maya Erosion Surface is 80 m at Alligator Peak, and for the Pyramid Erosion Surface is 100 m at Mt Ritchie. However, these are local aberrations and over most of the area of outcrop the surfaces are virtually parallel, showing that there was little areal variation in amount of erosion on the Pyramid Erosion Surface. Variation in erosion on the Maya Erosion Surface was also small for the thickness of the Aztec Siltstone has been reduced by only 100 m over an outcrop belt 150 km long.
DARWIN TILLITE
Lithology, Distribution and Thickness

The nearest glacial beds equivalent to the Metschel Tillite are in the Darwin Mountains 150 km south, though equivalent strata may yet be recognised in the intervening area when it is explored. The beds were discovered in the 1962-3 season by Haskell et al. (1965), who named them the Darwin Tillite. The type section at the south end of Colosseum Ridge is 27 m thick and comprises three members. The lowest is a mottled red and green sandstone with scattered clasts up to a metre across, the middle is 6 m of laminated to thin-bedded sandstone and siltstone with ripple marks and penecontemporaneous folds, and the upper is 7 m or more of greenish-grey sandstone like the lower member. The lower and upper members were interpreted as ice-contact deposits, and the folding of the middle member was attributed to overriding ice. Subsequent work has shown that these lithologies are typical of the glacial beds at other localities in the area, but that these members cannot be consistently recognised away from the type section.

Frakes et al. (1968) re-examined the glacial beds on Colosseum Ridge and described a 65 m thick section 6 km north of the type locality. The section is not only thicker, but also more complex than the type section. It includes four diamictite beds, though only the upper one, which is 17 m thick, is considered to be directly deposited from ice. The lower diamictites include bulbous masses of sandstone which are interpreted as load casts of sand that sank into water-saturated unconsolidated sediment. The interpretation requires that the lower diamictites were not compacted at the time of deposition, and hence are subaqueous, not ice-contact, deposits. Frakes et al. (1968) obtained a southeast palaeocurrent direction from asymmetrical ripple marks within the glacial beds, and deduced an east-southeast transport direction for a clast fabric, but found no direct evidence of ice flow direction.

The 1970-1 Victoria University expedition examined the Darwin Tillite at six localities, including the two described above (Barrett et al., 1971). Thicknesses of the Darwin Tillite at measured sections ranged from 15 m on the edge of the Hatherton Glacier (J4, Fig. 23.1) to 82 m at the north end of Colosseum Ridge (E2, Fig. 23.1). In thin section the tillite is similar in mineralogy and grain size distribution to the Metschel Tillite (Fig. 23.3). Tillite is the most common lithology at most places, but the north end of Colosseum Ridge is atypical in that the glacial beds there contain a large proportion of shale and the slumped intervals reported by Frakes et al. (1968). Despite the small proportion of ice-contact deposits two finely striated bedding surfaces like that at Mt Metschel were found near the base of the section at E2, and another set was found in the middle of the formation at Misthound Cirque (E7) where the striations are clearly within a tillite sequence. Measurements on the three sets of striations all indicate a northwest-southeast ice flow. Asymmetrical ripple marks indicate southeast palaeocurrent flow, confirming the southeast flow determined by Frakes et al. (1968) from ripple marks. It seems probable, then, in view of the coincidence of striations and palaeocurrent directions, that the ice also flowed to the southeast.

Haskell et al. (1965) and Frakes et al. (1968) concluded from clast composition

Table 23.4. Clast Lithologies in the Darwin Tillite; (1) Upper 2 m of Formation at E7 (Fig.23.2); (2) Upper Diamictite at North End of Colosseum Ridge (Frakes et al., 1968); (3) Lower Tillite at Type Section (Haskell et al., 1965)

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<tr>
<th></th>
<th>(1)</th>
<th>(2)</th>
<th>(3)</th>
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<tbody>
<tr>
<td>Granite</td>
<td>25%</td>
<td>50%</td>
<td>60%</td>
</tr>
<tr>
<td>Schist</td>
<td>10</td>
<td>5</td>
<td>10</td>
</tr>
<tr>
<td>Acid vole.</td>
<td>15</td>
<td>45</td>
<td>30</td>
</tr>
<tr>
<td>Quartz</td>
<td>35</td>
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</tr>
<tr>
<td>Quartzite</td>
<td>15</td>
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<tr>
<td>Sandstone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. of clasts</td>
<td>58</td>
<td>82</td>
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(Table 23.4) that the source area comprised older quartzose sediments and a chiefly granitic basement, like that now exposed in the Darwin Glacier area. At most localities we also noted that granitic clasts were dominant. However, a count of clasts in the upper 2 m of the tillite at Misthound Cirque showed that there is also a significant proportion of schist, volcanics, and quartz pebbles (Table 23.4) indicating a more varied source for the youngest glacial beds.

The north end of Colosseum Ridge is especially interesting not only because of the extensive slumping, but also because of the red-bed strata, and because one of us (RAK) has extracted Early Permian plant microfossil assemblages from the uppermost beds of the Darwin Tillite there. Greyish-red (5R 4/2) fine sandstone and siltstone beds are moderately common in the lower part of section E2, and some are involved in the slumping. It is most unlikely that these beds are equivalent to the Mid-Upper Devonian Aztec Siltstone 150 km north, despite the similarity in colour, for the tillite unit from 32 to 59 m above the base of the section has large patches of greyish-red also. The greyish-red fades irregularly upwards into the more common light greenish-grey. Red and green colours were noted by Haskell et al. (1965: 242) in the lowest tillite at the type section, but the red colour has not been seen at any other localities. The patchy distribution of the greyish-red in the tillite suggests that the colour was derived by glacial erosion of red-bed strata, probably equivalent to the Aztec Siltstone to the north, and has subsequently been partly reduced.

The strata with the plant microfossils are near the top of a section 27 m thick of medium grey shale (E1, Fig. 23.2). The shale is clearly equivalent to at least a part of the 82 m thick section of shale, sandstone and tillite 500 m west (E2, Fig. 23.2), for the interval is a slope-former between bluffs of Hatherton Sandstone (Devonian ?) and Misthound Coal Measures (Permian), and can be traced across a scree slope into the glacial beds at E2. Furthermore the pollen-bearing shale is clearly overlain by carbonaceous sandstone of the Misthound Coal Measures (Fig. 23.2), with a basal conglomerate resting on an erosion surface, as described below. It is suggested that the shales accumulated in a proglacial lake prior to the establishment of the post-glacial alluvial drainage system that deposited the overlying coal measures.

**Lower and Upper Boundaries**

The lower and upper boundaries of the Darwin Tillite are erosion surfaces equivalent to the Maya and Pyramid Erosion Surfaces of southern Victoria Land. The Maya Erosion Surface in the Darwin Mountains is best exposed at the north end of Colosseum Ridge, where it is gently undulating with a relief of about 50 m and wave lengths up to 500 m (Plate 23.II). The surface is everywhere cut in the Hatherton Sandstone, a well-sorted fine-grained quartzose sandstone at least 300 m thick. No datable fossils have been recovered, but it is the lithologic equivalent of the Arena Sandstone and Beacon Heights Orthoquartzite of south Victoria Land (Barrett et al., 1971), and is believed to be of Devonian age. The erosion surface at E1 weathers out as an extensive platform (Plate 23.II, 1). Elongate mounds 0.5 m high and 15 m long carved in Hatherton Sandstone (Plate 23.II, 2) strike north-south, and a groove 1 m deep strikes at 330°, sub-parallel to the strike of striated surfaces at E2 and E7. Therefore it is believed that the features at E1, and the erosion surface of which they are a part, were cut by Late Palaeozoic ice.

The Pyramid Erosion Surface in the Darwin Mountains is marked by lenses of clasts up to 60 cm long at the base of the sandstone bluffs of the Misthound Coal Measures. The basal conglomerate includes all lithologies found in the underlying tillite, though Frakes et al. (1968) noted that the proportion of granite was smaller than in the underlying tillite. Lithologies other than vein quartz are rare above the basal conglomerate, as is the case in south Victoria Land, suggesting that again there is no major time break between the Darwin Tillite and the basal conglomerate of the overlying coal measures.
Pl. 23.II. The Maya Erosion Surface at the north end of Colosseum Ridge, Darwin Mountains. 1. View west from E₁ to E₂. The wavy line at the top of the sandstone bluffs in the middle distance (heavy arrow) is the Maya Erosion Surface. The light arrow indicates the Pyramid Erosion Surface.

2. View northwest to the shales that represent the Darwin Tillite at E₁. The platform in the foreground was cut in Hatherton Sandstone by Late Palaeozoic ice. Site of the sample from which Early Permian plant microfossils were extracted is arrowed. Figure (circled) indicates scale.
PALAEOONTOLOGY AND AGE

A microfloral assemblage from shale in the upper Buckeye Tillite in the Ohio Range (Schopf, in Long, 1964; Rigby and Schopf, 1969) until now has provided only the palaeontological evidence for the age of the Late Palaeozoic glacial deposits in Antarctica. It included a species of *Nuskoisporites* (now *Parasaccites* in part) similar to *N. triangularis* (Mehta) Potonié & Lele, which is found in the Bacchus Marsh sandstone in Australia and just above the Talchir Boulder Bed in India. The age of the Buckeye Tillite was given by Rigby and Schopf as 'no older, and maybe somewhat younger', than the other Gondwana glacial deposits.

Three samples from the Metschel Tillite and five from the Darwin Tillite have been examined, but plant microfossils have been found in only one. It is a medium grey laminated shaly siltstone from the upper part of the Darwin Tillite from Colosseum Ridge in the Darwin Mountains (sample 23066, Fig. 23.2). The microfloral assemblage is listed in Table 23.5.

A large proportion of the spores were poorly preserved and many could not be identified specifically. Quantitative estimations of the relative abundance of each form species have not been made but it was possible to estimate the relative proportions of monosaccate pollen grains (12 per cent) and striate bisaccate grains (2 per cent).

The assemblage shows the characteristic features of Stage 2 of the Permian zonal system of Evans (1969) in the relative abundance of monosaccate grains and the paucity of striate bisaccate forms. There is a diversity of trilete spores, and the assemblage includes eonolette forms and a significant number of monocolpate grains.

Assemblages similar to that from the Darwin Tillite have been described from the Bacchus Marsh beds in Victoria (Virki, 1939, 1945; Pant and Mehra, 1963), in Western Australia in the Grant Formation of the Canning Basin (Balme, 1964) and in the Nangetty Formation and lower Holmwood Shale of the Perth Basin (Segroves, 1971), in India in the Talchir Series just above the Boulder Beds (Potonié and Lele, 1961), in the African lower Karroo beds (Hart, 1969), and in South America in the Bajo de Velez beds of Argentina (Menendez, 1969).

Evans (1969) assigned an Early Permian age to Stage 2, Stage 1 being Late Carboniferous. Helby (pers. comm.) considers these stages may be older than previously thought. The upper part of the Darwin Tillite contains a Stage 2 microflora and is therefore of earliest Permian age, or possibly older.

At Mt Fleming, in southern Victoria Land (Figs. 23.1 and 23.2) poorly preserved leaves were found immediately above the gradational contact between the Metschel Tillite and the overlying Weller Coal Measures. The leaves are mostly *Gangamopteris* spp. Macrofossil assemblages in the lower part of the Permian coal measures throughout south Victoria Land and in the Darwin Mountains (Kyle, in prep.) contain significant numbers of *Gangamopteris* leaves, and a *Glossopteris* form with an impersistent midrib is common in the Beardmore Glacier area (Grindley, 1963). Relatively common *Gangamopteris* associated with *Glossopteris* are characteristic of the lower coal measures sequence in Australia and of the Indian Talchir-Karharbari floras (Maheshwari, 1965). Predominantly *Glossopteris* flores are characteristic of Middle and Upper Permian sediments. The lower Weller Coal Measures and their equivalent in the Darwin Moun-
tains, the lower Misthound Coal Measures, are Early Permian in age. Grindley (1963) also considered the lower coal measures in the Beardmore Glacier area to be of the same age. Therefore, the macrofloral evidence indicates that the glacial beds from south Victoria Land to the Beardmore Glacier are no younger than Early Permian in age.

**DISCUSSION**

The glacial origin of Late Palaeozoic tillites in the Transantarctic Mountains has been well established (see Frakes et al., 1971), but until now no precise age has been obtained for the beds at any locality. The youngest in both south Victoria Land and the Darwin Mountains have been shown here to be of Early Permian age, and it seems probable that the entire thin sequence of glacial beds in these two areas is Permian also.

For a large part of the Carboniferous the Transantarctic Mountains may nevertheless have been the site of an extensive ice sheet that left little record in the area that it covered. Gow (1968), reporting on the 2164 m deep drill hole that penetrated the West Antarctic ice sheet at Byrd Station, revealed that the 'dirty' ice at the bottom of the hole extended only 4 or 5 m above bedrock. There existed no significant sedimentary record of ice sheet glaciation in that area, even though the base was at the pressure melting point, and though the area has experienced at least intermittent ice sheet glaciation over the last four million years (Mercer, 1972). It appears from this that most debris deposited within the maximum extent of a continental ice sheet is deposited during the final retreat phase of the glacial period. This occurred in Sakmarian or early Artinskian times for that part of the Antarctic ice sheet that extended southward from south Victoria Land to the Beardmore Glacier and beyond. Crowell and Frakes (1971) have suggested that this same ice sheet extended northwards into Tasmania, where the ice reached sea level as late as Kungurian times, supplying pebbles by rafting to sediments of the Ferntree Group (Banks, 1962; Brown et al., 1968: table 7.2). The southern margin of the ice in Victoria Land must have been well north of the Mawson Glacier (76° 20' S) by this time, for by then a northwest-sloping floodplain had been established in south Victoria Land and a southeast-sloping floodplain extended from 78° S to beyond the Ohio Range (Barrett and Kohn, this volume).

**ACKNOWLEDGMENTS**

We wish to thank fellow members of Victoria University of Wellington Antarctic Expeditions 15 and 16 (1970-71 and 1971-72) for willing assistance in the field and for subsequent discussions. We are grateful to Dr R. Helby, N.S.W. Geological Survey, for his comments on the age assignment. We also thank the University Grants Committee and the Victoria University Council for financial assistance, Antarctic Division, D.S.I.R. for logistic support, and the U.S. Navy for air transport to and within Antarctica. Mr E. Hardy drafted the diagrams.

**REFERENCES**


A Model for the Sedimentation of the Dwyka Glacials in the Southwestern Cape

J. N. THERON and H. J. BLIGNAULT

ABSTRACT

A detailed stratigraphic study, extending from Laingsburg to Loeriesfontein for a strike distance of 400 km, is in progress. The internal arrangement of the glacial sequence is remarkably consistent, both laterally and vertically. A marked characteristic is the cyclic repetition of a fundamental unit which typically commences with a significant thickness of massive tillite followed by stratified sediments. The latter include stratified diamictite, sheet and shoestring arenites, boulder beds and massive till-flow units.

The maximum development of the glacials resulted in the accumulation of four such cycles, each of which represents an advance and retreat of an ice sheet. The environment of deposition throughout was that of alternating grounded and floating ice shelves. The upper cycles are transgressive over the lower cycles.

New palaeo-iceflow data show that, contrary to present belief, the general ice advance was due westward. The spreading centre, situated closely to the east, probably coincided with the pole at that time.

The rock record indicates a deterioration in climate from lower to upper Dwyka times, which is related to the spreading of the polar environment towards lower latitudes. The final waning of glacial conditions is recorded within one recession.

Sediments of the Cape sequence behaved in a hydroplastic fashion during the first advance of a grounded ice sheet and it is shown that the accumulation of such a thick glacial sequence can be ascribed to the coincidence of high latitudes with the active Cape-Karroo Basin.

INTRODUCTION

At the previous Gondwana Symposium (Cape Town, 1970) it became apparent that very little indeed was known about the actual mode of deposition of the Dwyka glacial sediments. With this in mind a study of the glacials was undertaken, concentrating mainly on the north-south outcrop area (Fig. 24.1) where the tectonic effects of the Cape Orogeny were negligible. The initial aim was to investigate the internal arrangement of the glacial sequence and great care was taken to observe and document the many smaller distinctive features.

That this approach was demonstrably successful is being documented elsewhere (Blignault and Theron, in prep.), and the more salient aspects are discussed below.

STRATIGRAPHY

The Dwyka glacial sequence constitutes a lithostratigraphic unit with Group or Sub-Group status. Its lower limit is defined by the first appearance of glacigenic sediment and its upper limit by the final disappearance of such material. Near the top there are intercalations of shale of the same type as the overlying shales which should be grouped
Fig. 24.1. Outcrop map of the Dwyka glacials, showing the position of areas of detailed investigation.
Sedimentation of the Dwyka Glacials

with the Ecca sequence (see also Winter and Venter, 1972); both contacts are easily recognised in the field.

In the south the glacials overlie rocks of the Cape sequence conformably, but they overstep all the Cape units towards the north and finally cover the gneisses of the Nam-aqua Metamorphic Complex.

INTERNAL ARRANGEMENT AND REGIONAL ASPECTS

The present study revealed a remarkably consistent internal arrangement of the Dwyka over the entire area investigated. The various stratigraphic units persist along strike from Laingsburg to Langberg (Fig. 24.1), i.e. about 400 km; elaborate field checking between profiles, largely assisted by the recognition of such units on aerial photographs, are conclusive; even in areas of poor outcrop between profiles, recognisable units are found in their expected stratigraphic positions.

In vertical dimension an equally consistent pattern has been established; the glacial sequence consists of the repeated stacking of a fundamental unit which will be referred to as a cycle. These cycles are regionally persistent in a lateral sense, mappable on a small scale and should have Formation status. The upper cycles transgress the lower cycles towards the northern margin of the Dwyka basin. Four cycles have been recognised towards the south and will be referred to as cycles 1-4.

The type cycle consists of a basal and a terminal zone. The basal zone consists of a significant thickness of massive diamictite; laterally persistent stratification, or sorting features are essentially absent. The terminal zone, in contrast, typically has stratified and sorted units interbedded with massive diamictite beds which are generally a few orders of magnitude thinner than the basal massive diamictite zones. In general, the thickness of the terminal zone is subordinate to that of a basal zone. There is furthermore a distinctive difference between the terminal zones of the lower and upper Dwyka. The contacts between the terminal and basal zones, and for that matter between cycles, are normally abrupt, and ‘soft’ deformational features below the interface of cycles are conspicuously absent.

TERMINAL ZONES

In contrast to the massive and homogeneous nature of the basal zone diamictites, those of the terminal zones offer abundant evidence of the environment of their deposition. Two basic types, low and high energy, which reflect the condition of the subaqueous environment of deposition are recognised. In using these terms, we exclude high flow regimes (Elliot, 1965) like turbidity currents and mass flows which are not necessarily related to the energy conditions of the environment of deposition.

The low-energy type terminal zones consist predominantly of stratified diamictite which is locally interbedded with shale to form a rhythmite. Another common rhythmic variation is the intercalation of massive diamictite beds (varying in thickness from centimetres to metres) with stratified units. The stratified sequences typically contain outsize clasts.

In the high-energy type terminal zones, distinctly sorted and stratified units are commonly separated by massive diamictites to form mega-rhythmites. A specific mega-rhythmite, distinguished by its unique pattern of succession, may be regionally uniform or may acquire different characteristics along strike. The sorted and stratified units are of the following types.

1. Boulder beds form regional marker horizons and are of two types:
   - (a) imbricated single layer boulders embedded in massive diamictite and disposed along a stratigraphic horizon in a disrupted fashion; boulder pavements are always associated with this type;
   - (b) the single layer type which apparently grades into boulder rudite which has a disrupted framework and a matrix of diamictite.

2. Arenite bodies (1-3 m) are important marker horizons; they are discontinuous in outcrop but persist regionally along the same stratigraphic horizon. The shoestring type predominates, but lenticular bodies are common and sheet arenites are found locally. The
Fig. 24.2. Fence diagram depicting the internal arrangement of the Dwyka sequence
shoestrings have subvertical sides and are very characteristically deformed to anticlinal forms, their B-axes being parallel to the shoestring trends. The only 'soft' deformational features (not necessarily of the type described above) observed within the glacial sequence, were displayed by these arenite bodies.

3. Stratified diamictites (with outsize clasts) which are associated with pebble rudites (1-5 m).

4. Relatively thin (1-2 m) massive diamictite beds (not graded) are defined by persistent boundary surfaces. These massive units form sequences of two or three layers where definitely recognised; they are, however, believed to be more widely developed.

ICE-FLOW

It is claimed that field evidence shows conclusively that ice advance was from the east. It is perhaps necessary, owing to the controversial nature of this conclusion, to give a brief account of the criteria used. The following friction structures associated with east-west striae on hard pavements were found to be consistently orientated and indicate ice-flow westwards (Embleton and King, 1968):

1. Crescentic gouges, chattermarks and lunate fractures have their shallower dipping surfaces inclined to the west.

2. Crescentic fractures are concave westwards with their steep surfaces dipping in the same direction.

3. Nailhead striae point with their heads in a westerly direction.

The basal erosional forms on sediments of the Cape sequence differ markedly from those normally expected on 'hard' pavements. Long rectilinear 'groove-like' features not only abrade the upper beds but also deform lower-lying strata, effectively forming folds with wavelengths varying from centimetres to metres.

Unequivocal evidence of ice movement on such pavements where the basement acted in a hydroplastic fashion (Elliot, 1965) is very rare. Conclusive evidence of such a kind is preserved at Elandsdrif where scribing clasts are situated on the western ends of 'grooves', and crescentic fractures on embedded clasts indicate flow in the same direction.

Macrofabric analysis of massive diamictite substantiates the above evidence; the AB-surfaces of discoidal clasts (cycles 1 and 2; four sampling sites) are preferentially imbricated towards the east, pointing to a westerly flow of the transport medium.

The areal variation of ice-flow data (Fig. 24.3) associated with the deposition of cycles 1 and 2 shows:

1. a divergent pattern pointing to a spreading centre of the ice-sheet due east, but probably situated west of the east coast and

2. that there is no obvious relation between the flow pattern and the isopachs; the Augustfonteineberg 'trough', if it existed at this time, apparently had no effect on ice movement.

Ice-flow data from the upper cycles are sparse, but the palaeoflow/isopach relation is too tempting to ignore; apparently the ice-movement was influenced by the now prominently developed Augustfonteineberg 'trough' towards which the ice-flow was directed locally.

Ice-flow was previously believed to be in the opposite direction (i.e. from west to east), but if the evidence presented by Stratton (1968) is critically examined, it reveals two important points.

1. At nine stations the preferred orientation of clast long axes has an east-west trend; at three of these the long axes are preferentially disposed in the horizontal while at the remaining six stations the long axes are preferentially imbricated in an eastward direction. Ice-flow, however, was still considered to be from the west.

2. Stratton (1968: pl. XIXB) depicts as substantiating evidence an 'upflow' bevelled boulder; it is quite clear both from this photograph and a field inspection that this so-called 'bevelling feature' is due to the intersection of two planar surfaces. The horizontal surface forms part of the pavement, and the other could be a facet acquired at any time during the past glacial history of the boulder; in fact, well-developed facets are quite common on the boulders of boulder beds where pavements are absent. Erosion on the upflow side of an obstacle should not affect an inclined planar surface, but would
Fig. 24-3. Isopach and palaeoflow map
rather cause irregular stoss-like features like those described for roche moutonnées.

**ISOPACHS AND THE BASIN**

A new isopach map (Fig. 24.3) for the Dwyka glacials in the southwestern Cape is presented. The pattern shows a striking resemblance (Fig. 24.4) to that of the Cape sequence (Rust, 1967; Theron, 1972), with the axis of maximum negative movement displaced inwards with respect to the present-day continental outline. A comparable northward shift is apparent for the main southern east-west axis from Cape to Karroo times (cf. isopachs of Venter, 1969; Winter and Venter, 1972; Ryan, 1967). It is inferred that the tectonic framework of basin formation maintained its character from Cape to Karroo times and the inward shift of the negative axis probably was causally related to the effects of Cape Orogeny on the African Craton/Plate which had been stable since the Pan-African Orogeny.

The rock record conclusively provides evidence for both a floating and grounded ice shelf environment, which again suggests that the water was deep enough to accommodate floating ice shelves, but never so deep as to prevent the grounding of advancing ice sheets.

An accelerating rate of depression of the basin is postulated for the upper Dwyka times because

1. the anomalous thickening of the Dwyka lithosome in the south and in the Augustfonteinberg 'trough' is predominantly due to thickening of cycles 3 and 4 and
2. palaeoflow data show that the Augustfonteinberg 'trough' possibly did not exist in lower Dwyka times, or else did not have such significant dimensions as in upper Dwyka times.

**THE MODEL**

The formulation of this model is inhibited to some extent by the north-south linear pattern of the outcrop area studied (our 'reference line' with respect to the Dwyka basin) which trends approximately at right angles to the inferred ice advance and retreat.

The model proposed below is based largely on concepts put forward by Carey and Ahmad (1961), Frakes and Crowell (1967) and Boulton (1972). It is necessary, though, to discuss certain aspects of the dry-base advance. The classical concept involves a body of ice, with its base below freezing point, moving over bedrock. The ice/bedrock contact defines the surface across which the stress couple operates. Where a polar ice sheet advances over unconsolidated water-logged sediment, the upper portion of the sediment will become frozen and effectively form part of the advancing grounded ice shelf. The surface of differential movement is defined by the frozen/unfrozen sediment contact. Such an advancing ice sheet is then, in fact, wet base. It is concluded that the dry-base concept does not apply where polar ice sheets are grounded on unconsolidated sediment.

The regional persistence of the four cycles, the thickness of their basal diamictite zones and absence of stratification and sorting features therein, indicate four major grounded ice advances. The basal massive diamictite zones are considered melt-out tills from a grounded shelf.

The first major ice advance caused deformational features which indicate wet-base conditions. Although deformational features on sheet sediments at the base of each succeeding cycle are absent, it is concluded on the grounds discussed above that advances 2, 3 and 4 were wet-base. The spreading centre was situated due east and could very well have been close to the position of the pole at that time (see Creer, 1972, for the position of the palaeomagnetic pole in Dwyka time).

The four terminal zones mark recessive periods; it is quite possible that there were many more minor interglacial periods at lower latitudes (west of our reference line), but these four recessions mark major periods of amelioration as the buoyancy line retreated to rather high latitudes (east of our reference line); the fourth recession in fact was complete. The persistence of rafted clasts throughout the glacial sequence shows that the distal boundary of the floating shelf and iceberg tracks never retreated across our reference line except in the final waning
Fig. 24.4. Superimposed isopach map of the Cape, Bokkeveld and Dwyka sequences.
stage. This inference gives us some measure of the position of our reference line with respect to the pole. In view of the basal freezing concept formulated above, which implies basal 'fixing' of probably significant proportions of the underlying sediment, it is thought possible that the terminal zones do not represent the full sedimentary record of the interglacial period. The absence of part of the upper Witteberg sequence at Genadendal and Droëhofek can perhaps be ascribed to this effect. The interpretation of the depositional environment of the terminal zones is therefore probably restricted to the more general and regional characteristics of the environment.

The high-energy mega-rhythmites of cycles 1 and 2 are typical products of wet-base retreats as is envisaged by Carey and Ahmad (1961) and Frakes and Crowell (1967).

1. The deformed shoestring arenites are typical ice contact features (Boulton, 1972) and probably represent eskers that coalesced locally to form outwash sheet arenites close to the buoyancy line.

2. The boulder beds, commonly interpreted as lag deposits (Flint, 1971), could have been formed by the winnowing action of the high energy water currents along the buoyancy line; the sporadic development of pavements on these boulder beds, without the deposition of significant thickness of till, points to minor rejuvenations. Their wet-base nature is indicated by rotation of the boulders during the bevelling by the overriding ice. These minor interglacial advances might also account for the deposition of some of the thicker massive diamictite units which form an integral part of the mega-rhythmites.

3. The presence of intercalated stratified diamictite with rafted clasts precludes a supraglacial origin of the mega-rhythmites and they probably represent ice-rafted sediments reworked by bottom currents.

4. The well-defined, massively stratified diamictites are considered as till-flows down the foreset slope. Till-flows can perhaps result in the formation of quite thick diamictite units, but these cannot always be conclusively distinguished from tillites. It can be shown, however, that both genetic types occur in the high-energy mega-rhythmites.

The low-energy type terminal zones of the upper cycles are indicative of a markedly different depositional environment. Apart from the more local development of boulder beds at the base, these zones provide no other evidence of high-energy meltwater. Re-worked ice-rafterd sediment in the form of stratified diamictite predominates and the intercalations of thin massive diamictite units are interpreted as mass flows of the unstable foreset slope environment. The development of meltwater at the buoyancy line was probably restricted to a marked extent and this can be attributed to a more severe polar climate in contrast to the more temperate condition that prevailed during the first two interglacials.

It is reasoned above that the temperatures decreased from lower to upper Dwyka times. This means that the polar conditions spread across our reference line, a rather high palaeolatitude, towards lower latitudes from lower to upper Dwyka times. The rock record shows no gradual amelioration towards the closing of the glacial period, in fact the last two advances probably mark the peak of polar development and terminal zone 4 represent the final waning stages and termination of the glacial conditions.

The Cape-Karroo sedimentation cycle is, for the following reasons, considered to be a continuous event.

1. The absence of prominent pre-Dwyka erosional features.

2. The tectonic framework remained essentially the same throughout Cape-Karroo times.

3. The first glacial advance effected 'soft' deformational features on sediments of the Cape sequence. It can be argued that these features are developed on pro-glacial outwash or reworked Cape sediments. That this is at least not always the case is borne out by the following arguments:

(a) A typical fossil of the Witteberg sequence (Haplostigma) was found as part of a 'soft' pavement, and

(b) bioturbated arenites and intraformational conglomerates, typical of the Bokkeveld sequence, were observed to be associated with 'soft' deformational features.
It is claimed, therefore, that the Cape sediments reacted hydroplastically at the time of the first grounded shelf advance.

It follows that the coincidence of high latitudes with the active Cape-Karroo basin allowed for the accumulation of an exceptionally thick glacial sequence. Perhaps it is not fortuitous that the increase in rate of downwarping in upper Dwyka times coincided with the maximum development of polar conditions, suggesting that the increase in the polar ice mass had some influence on the tectonic development of the Dwyka basin.

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REFERENCES


Non-glacial Origin for Conglomerate Beds in the Wajid Sandstone of Saudi Arabia

DONALD G. HADLEY AND DWIGHT L. SCHMIDT

ABSTRACT

Conglomerate beds containing sparse boulders in the upper part of the Wajid Sandstone of Permian or older (?) age have been interpreted as Gondwana glacial deposits (Helal, 1965). These beds, exposed in the Bani Khatmah area near the southeastern edge of the Arabian Shield of Saudi Arabia, contain normally graded polymictic conglomerate, commonly in fluvial channels cut into mature quartz sandstone. None of the features observed suggest deposition in a glacial environment. The conglomerate beds constitute a laterally and vertically restricted facies about 150 km wide in the quartz-sand facies of the Wajid.

Recent palynological evidence suggests that the Wajid Sandstone is of Cambrian or Ordovician age. The conglomerate of the Wajid is probably derived from locally uplifted Precambrian rocks in the southern part of the Arabian Peninsula. This is in contrast with an otherwise extensive, stable, deeply weathered, and low-lying Precambrian Shield that was the source of the predominant quartz sandstone of the Wajid during early Palaeozoic time.

INTRODUCTION

Steineke and others (1958: 1301) reported boulder beds containing clasts of granitic and metamorphic basement rocks in the upper part of the Wajid Sandstone in the Bani Khatmah area of Saudi Arabia (Fig. 25.1). The beds were first described, however, in sections measured in 1950 by S. B. Henry and R. A. Bramkamp (Powers et al., 1966: D28). These were later interpreted by Helal (1963, 1965) to be Permian and Carboniferous tillites.

The literature of Gondwanaland contains few examples of upper Palaeozoic glacial deposits existing independently of the Glossopteris flora. The report of Permian and Carboniferous tillites in the upper Wajid has been questioned because the Glossopteris flora is not known from the Arabian Peninsula (J. M. Schopf, pers. comm.). These reported glacial deposits have been cited in recent literature on Gondwanaland (e.g., Hamilton and Krinsley, 1967). In any reasonable reconstruction of Gondwanaland, Arabia is far removed from the grouping of the Gondwana glacial deposits in Africa, South America, Australia, Antarctica, and India. Except for a relatively slight lateral movement of Arabia during the Tertiary opening of the Red Sea and Gulf of Aden, Arabia has been part of North Africa since the Precambrian and is more than 3000 km from the nearest Gondwana glacial deposits in Africa.

The age of the Wajid Sandstone is Permian or older (?) (U.S. Geol. Survey—Arabian American Oil Co., 1963) but younger than Precambrian; Helal (1965) suggested a late Palaeozoic age. More recent evidence, however, suggests an early Palaeozoic age (Brown, 1970). The specific age of the Wajid Sandstone remains unknown. In subsequent pages
we have briefly reviewed the Palaeozoic stratigraphy of Saudi Arabia in order to put the age problem in proper perspective.

Conglomerate beds in the upper part of the Wajid Sandstone are exposed only in the cuestas of Bani Khatmah and the southern part of Jabal Tuwayq (Fig. 25.2) between lat. 18° and 19°N and long. 45° and 46°E at the western edge of Ar Rub' Al Khali (Fig. 25.1). Fig. 25.2 shows the geology of the Bani Khatmah area, as taken from the 1:500,000-scale map by Bramkamp and others (1963); the regional geologic setting is given on the geologic map of the Arabian Peninsula (U.S. Geol. Survey—Arabian American Oil Co., 1963).

The Bani Khatmah area is largely covered by aeolian sand deposits, above which rises the Jabal Tuwayq escarpment and its southern extension, the Bani Khatmah escarpment (Powers et al., 1966: D99, D100; Helal, 1965: 195-6; Bramkamp et al., 1963). These sands constitute the western border of the great Ar Rub' Al Khali desert. The sand area is a complex combination of Quaternary aeolian and fluvial deposits, including dissected pediments near the escarpment outcrops. In a few places, in association with old pediments, many 'erratic boulders' have accumulated. These boulders are granitic and metamorphic rocks of Precambrian age and have a maximum diameter of about 2 m. Undoubtedly, it is this unusual abundance of lag boulders that made earlier workers think...
Fig. 25.2. Geologic map of the Jabal Tuwayq-Bani Khatmah area, western Ar Rub' Al Khali, Kingdom of Saudi Arabia. Geology from the Geologic Map of the western Ar Rub' Al Khali quadrangle, Kingdom of Saudi Arabia (Bramkamp et al., 1963)

that the deposits in the upper Wajid Sandstone were glacial.

The remoteness of the Bani Khatmah area has made it difficult to test the validity of the glacial interpretation. We briefly studied the critical outcrop area at Bani Khatmah during February 1973. The outcrop at the southern end of Jabal Tuwayq was not examined, but we did examine the lower Wajid farther west and north in southern Saudi Arabia.

REGIONAL STRATIGRAPHY

The regional stratigraphy from the Precambrian Shield rocks through the Upper Permian Khuff Formation is summarised in Table 25.1. The Wajid Sandstone is underlain by Precambrian rocks and overlain by the Khuff Formation. The thin Khuff Formation in the Bani Khatmah area is over-
lain directly by the Jurassic Tuwayq Mountain Limestone. No intervening Triassic and Jurassic formations are present in this area.

The Wajid Sandstone is the basal quartz sandstone formation that overlies the Precambrian basement of the southern and southeastern part of the Arabian Shield (Figs. 25.1 and 25.3) (Powers et al., 1966: D27; Alabouvette and Villemur, 1973). It is a Nubian-type (Whiteman, 1971), continental, flat-lying, homogeneous sandstone that has been slightly tilted (generally less than 3°). The Wajid may be as thick as 950 m (Powers et al., 1966: D28). In the absence of detailed study and because there are no obvious marker units, the formation is divided in general terms into lower, middle, and upper parts (Table 25.1).

The base of the Wajid is well exposed 100-150 km west of Bani Khatmah. In this area, examined by the authors, the basal beds dip less than 2° and are exposed in steep cliffs and as outlying buttes above the dissected Precambrian basement rocks. The contact has low relief, but in a few places hills of Precambrian granite rise well above the contact level. The basal part is conglomeratic (Table 25.1).

Cross-bedding is characteristic and con-

| Table 25.1. Summary of Regional Palaeozoic and Lower Mesozoic Stratigraphy of Central Saudi Arabia, modified from Powers et al. (1966: D7, D20-D32) |
|---|---|---|---|
| Age | Formation | Subdivision | Thickness (m) | Generalised lithology |
| **BANI KHAHMA AREA — SOUTHERN SAUDI ARABIA** |
| JURASSIC | Tuwayq Mountain Limestone | (specific) | 50 | Massive coral-bearing dense limestone |
| | Kluff Formation | (specific) | 0 to 154 | White, massive, fine- to coarse-grained sandstone and subordinate red and green shale |
| PERMIAN | (unconformity) | ~950 | Sandstone, conglomerate |
| | (general, South-central Saudi Arabia) | | | Lower 70 m: cross-bedded white to yellow, poorly cemented, coarse-grained quartz sandstone; upper 50 m: restricted facies of grey-green to red siltstone and subordinate interbedded tan to purple fine-grained sandstone; subordinate small plates and concretions of limonitic and hematitic cemented sandstone |
| | (upper) | ~120 | | Buff, cross-bedded, friable, poorly sorted, subangular, quartz sandstone; some thin quartzitic and conglomeratic beds; reddish-brown and purple hematitic layers common |
| ? | Wajid Sandstone | (middle, 200 km northwest of Bani Khatmah) | ~365 | Buff, cross-bedded, friable, poorly sorted, subangular, quartz sandstone; some thin quartzitic and conglomeratic beds; reddish-brown and purple hematitic layers common |
| | (lower, 125 km west of Bani Khatmah) | ~150 | Light grey, weathers buff to light reddish-brown, well-bedded, cross-bedded, moderately to well-sorted, moderately cemented, porous, coarse-grained quartz sandstone; contains vertical tubular structures of Tigillites (Alabouvette and Villemur, 1973); thin dark brown zones of limonitic and hematitic cemented sandstone along bedding and joint planes; basal 10 m in addition contains conspicuous beds of rounded quartz-pebble conglomerate and grit |
| | (unconformity) | | Precambrian basement complex |
Non-Glacial Origin for Conglomerate Beds, Saudi Arabia

### Generalised Lithology

<table>
<thead>
<tr>
<th>Age</th>
<th>Formation</th>
<th>Subdivision</th>
<th>Thickness (m)</th>
<th>Generalised lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>NORTH-CENTRAL SAUDI ARABIA</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PERMIAN</td>
<td>Khuff Formation</td>
<td>(general)</td>
<td>171</td>
<td>Lower limestone, middle shale, upper limestone; dominantly sandstone south of 21°N</td>
</tr>
<tr>
<td>Unconformity</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jauf Formation</td>
<td></td>
<td>(general)</td>
<td>299</td>
<td>Varicoloured silted shale containing beds of sandstone and limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(general)</td>
<td>1072</td>
<td>Interbedded sandstone, siltstone, and shale; base sharply marked by Lower Ordovician graptolites in shale</td>
</tr>
<tr>
<td>DEVONIAN</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tawil member</td>
<td></td>
<td></td>
<td></td>
<td>Medium- to coarse-grained sandstone with large-scale cross-bedding and common conglomerate</td>
</tr>
<tr>
<td></td>
<td>Tabuk Formation</td>
<td>upper cyclic deposit</td>
<td></td>
<td>Thick-bedded, variably cemented massive sandstone; locally cross-bedded and gradationally bedded; contains streaks of quartz pebble conglomerate; Scolithus less common than below</td>
</tr>
<tr>
<td>SILURIAN and ORDOVICIAN</td>
<td>Tabuk Formation</td>
<td>lower cyclic deposit</td>
<td></td>
<td>White to red-brown, thin-bedded sandstone, shaly sandstone, sandy shale, siltstone, and shale with hematitic cementation locally along bedding planes; vertical tubular structures, Scolithus, and worm trails (?) abundant</td>
</tr>
<tr>
<td>CAMBRIAN</td>
<td>Saq Sandstone</td>
<td>(general)</td>
<td>&gt;600</td>
<td>Laterally and vertically homogeneous unit of buff to grey, red, and white, friable, commonly cross-bedded, poorly- to well-sorted quartz sandstone; red-brown iron cementation common along joint planes; basal part is conglomeratic; marine Craziana tracks in upper part; Middle Cambrian trilobites in Jordan; correlative with the Quweira, Umm Sahm, and Ram Sandstones of Jordan where composite thickness is 850 m (Quennell, 1951)</td>
</tr>
</tbody>
</table>

spicuous in all parts of the Wajid. Foreset beds 1-10 cm thick are common in cross-bedded medium-grained sandstone to gritstone units 30 cm-3 m thick; they are abundant within conglomeratic beds and channels. In the middle Wajid, 50-75 km west of Bani Khatmah, well-developed trough cross-bedding indicates a consistent northwesterly transport direction; cross-bedding in the lower and upper parts of the Wajid likewise indicates a generally similar transport direction.

The Upper Permian Khuff Formation (Powers et al., 1966: D29) is the lowest marker formation that extends across most of Saudi Arabia. It disconformably overlies the Wajid Sandstone in the Bani Khatmah area. Farther north it rests directly on Precambrian basement rocks (Fig. 25.3). Still farther north, about the latitude of Ar Riyâd, it overlies older Palaeozoic sedimentary rocks, the Saq, Tabuk, and Jauf Formations (Figs. 25.1 and 25.3). The Khuff is 154 m thick in the southern Jabal Tuwayq area; it decreases to zero in the southern part of the Bani Khatmah escarpment because of erosional truncation prior to Jurassic deposition. North of the Bani Khatmah area, the
Age and Stratigraphic Relations of Glacial Deposits

Fig. 25.3. Regional stratigraphic section of lower Palaeozoic rocks of Saudi Arabia from Bani Khatmah north to the vicinity of Al Kitadah (Fig. 25.1), showing the unconformity at the base of the Khuff Formation and the possible time equivalence of the Wajid Sandstone to the Tabuk Formation and Saq Sandstone. The section is modified from Powers et al. (1966: pl.1).

Khuff is represented by a uniform outcrop belt 25 km wide that extends northward for 625 km.

Upper Wajid Sandstone at Bani Khatmah

Two sections of the upper Wajid Sandstone were measured at Bani Khatmah—at Khashm Khatmah (Station No. 4, Fig. 25.2) and 5 km east-southeast of Khashm Khatmah (Station No. 6). The sections are 46.8 and 134.2 m thick; stratigraphy and lithologic character of the measured intervals are shown in Fig. 25.4.

The upper Wajid is composed of conglomerate and quartzose sandstone. The sandstone is fine grained to pebbly, is light grey, cream, maroon, and green, strongly cross-bedded to parallel laminated, and friable. The conglomerate beds and conglomeratic cross-bedded units grade progressively upward into fine-grained, parallel-laminated sandstone. The units closely resemble 'fining upward cycles', features common to fluvial deposits (Allen, 1965; Visher, 1965). Cross-bedding is of the trough type (Potter and Pettijohn, 1963); individual units show abundant truncation (Pl. 25.1). Foresets of the cross-beds are commonly very pebbly.

Apart from the conglomeratic beds and cross-beds, the sandstone is exceptionally mature, texturally and compositionally. The framework constituents are subrounded to well rounded, and well sorted. A size-distribution analysis of a representative sandstone bed plotted as a histogram (Fig. 25.5A) consists of eight Udden grades that range from 8 mm (pebbles) to 1/32 mm (fine silt) and smaller. It is strongly unimodal, 96.3 per cent of the sample being in the chief and proximate classes (medium- to very coarse-grained sand, Wentworth, 1922). The Wajid size analysis is similar to that of normal fluvial sediments, as shown by the histogram in Fig. 26.5B. Helal (1965: fig. 2) described nine histograms of material collected from the lower 50 m of the upper Wajid; all are similar in character to the one shown in Fig. 25.5C. He concluded that his size-distribution data support a glacial origin for the Wajid. However, his samples consist chiefly of sand and gravel, and none have more than 25 per cent silt and clay. This is an uncommonly low proportion of fine-grained sediment for normal glacial till (Fig. 25.5D).

Table 25.2 shows three modal analyses of
the upper Wajid Sandstone from the Bani Khatmah area. The rock is mature subgreywacke and orthoquartzite (Fig. 25.6), which generally consists of more than 90 per cent quartz (of all species), less than 10 per cent fine-grained metamorphic rock fragments, and less than 2 per cent feldspar. The framework minerals are generally very well rounded. Original calcite cement has been dissolved in most samples, but can still be seen in embayments in detrital grains. Some beds are tightly cemented by hematite, which also actively etches and embays the detrital grains. The hematite is secondary and post-
Pl. 25.1. Cross-bedded and channelled upper Wajid Sandstone from the basal part of the measured section at Station No. 4, Bani Khatmah area. Channel outlined in ink.

dates the original calcite cement. Primary porosity of the sandstone was as much as 34 per cent.

Conglomerate zones are in the lower part of the measured section at Khashm Khatmah (Station No. 4) and at three levels in the section measured at Station No. 6. The conglomerate at Khashm Khatmah consists of polymictic, poorly sorted clasts in channels 1 m or more thick and non-channelled beds that are sometimes cross-bedded and sometimes not. More than one conglomerate zone probably exists at Khashm Khatmah. A float boulder 45 cm in diameter was found on talus in the covered interval and strongly indicates two or more conglomerate zones there. The three conglomerate zones measured at Station No. 6 contrast with sections measured by Henry and Bramkamp (Powers et al., 1966: D28) and Helal (1965), who reported two beds of 'boulder erratics'. The third bed could easily have been missed by them, as most of the Bani Khatmah escarpment is covered by talus debris. Moreover,

Table 25.2. Modal Analyses (in per cent) of upper Wajid Sandstone Samples from the Bani Khatmah Area

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Rock type</th>
<th>Subgreywacke</th>
<th>Orthoquartzite</th>
<th>Orthoquartzite</th>
</tr>
</thead>
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<tr>
<td>81546</td>
<td>Igneous quartz</td>
<td>61.7</td>
<td>62.0</td>
<td>65.5</td>
</tr>
<tr>
<td></td>
<td>Polycrystalline quartz</td>
<td>4.1</td>
<td>1.6</td>
<td>2.6</td>
</tr>
<tr>
<td></td>
<td>Microcrystalline quartz</td>
<td>—</td>
<td>0.2</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>Feldspar</td>
<td>0.1</td>
<td>0.7</td>
<td>0.9</td>
</tr>
<tr>
<td></td>
<td>Rock fragments</td>
<td>7.1</td>
<td>1.1</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>Cement</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Hematite</td>
<td>24.8</td>
<td>32.6</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Carbonate</td>
<td>Tr</td>
<td>Tr</td>
<td>23.4</td>
</tr>
<tr>
<td></td>
<td>Heavy minerals</td>
<td>—</td>
<td>—</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>Voids</td>
<td>2.2</td>
<td>1.8</td>
<td>6.4</td>
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<tr>
<td></td>
<td>Total</td>
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<td>100.0</td>
<td>100.0</td>
</tr>
<tr>
<td></td>
<td>No. of point counts</td>
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<td>671</td>
<td>235</td>
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</tbody>
</table>
the uppermost bed at Station No. 6 is only about 1 m thick and contains very sparse clasts.

The clasts are rounded to subrounded and range in size from pebbles to boulders. The largest observed in outcrop was 40 cm, but clasts as much as 2 m in diameter (Pl. 25.II) are found on an older high-pediment surface. Most of the clasts are composed of Precambrian plutonic and metamorphic rocks that include granite, syenite, diorite, several types of gneiss and schist, rhyolite porphyry, metabasalt, metatuff, and metasedimentary rocks. The variety of the clasts indicates widespread erosion of the basement complex.
Pl. 25.II. Slope below the high pediment is covered with coarse lag boulders of resistant Precambrian igneous and metamorphic rocks derived from the pediment near Station No. 5 sandstone (Pl. 25.III). Above the conglomerate bed several of the sandstone beds are strongly cross-bedded. Beds in the lower part are commonly mottled, and show load casts and flame structure.

**Age of the Wajid Sandstone**

A Cambrian-Ordovician age of the Wajid is suggested by recent palynological evidence from well cores studied by Hemer (pers. comm.; Brown, 1970: 80). Moreover, a correlation of the upper Wajid with part of the Tabuk Formation of Early Ordovician to Early Devonian age is considered feasible. The quartz-sand facies of the lower and middle parts of the Wajid is similar to that of the Saq, whereas the mixed quartz-sand facies and silty facies of the upper Wajid are similar to, though much less extensive than, the interbedded sand and shale facies of the Tabuk. Thick Nubian-type sandstones, such as the Saq-Tabuk and the Wajid, are widespread on large shield areas and furthermore, the cross-section and map patterns (Figs. 25.1 and 25.3) suggest previous continuity of these formations across the pre-Khuff erosional gap in central Arabia. The correlation has been supported by Brown (1970: 80). Reports of Late Permian spores from a drill core in possible upper Wajid and of Early Devonian to Carboniferous spores from another drill hole in possible Wajid (Powers et al., 1966: D29), have not been confirmed and are assumed unreliable (L. F. Ramirez, pers. comm.).

The age of the Khuff is well established as Late Permian, on the basis of microfossils of diagnostic spores, pollen, and algal assemblages (Rezak, 1959: 537). The Late Permian age is supported by less diagnostic megafossils also (Powers et al., 1966: D32). Helal (1965) assigned a Permian and Carboniferous age to the Wajid without conclusive age criteria other than that it is older.
Pl. 25.III. Inter-bedded maroon and cream, laminated to massive, fine- to medium-grained Wajid Sandstone, measured section at Station No. 6, Bani Khatmah area. Load casts and flame structure are prominent features in this part of the measured section.

than the overlying Upper Permian Khuff Formation. He concluded that the upper Wajid is of glacial origin and that the glaciation must be of Gondwana age. He further supported his case by correlating the upper Wajid with reported glacial or aqueoglacial deposits’ in southwestern Oman (Morton, 1959: 10). The Oman boulder beds are found in sandstone with associated sandy limestone containing Metalegoceras aff. clarkei Miller of Sakmarian-Artinskian or Early Permian age (Hudson, 1958). We know of no further
study of these deposits, and their glacial origin is not firmly established.

Fluvial Origin for the Upper Wajid Sandstone

Fluvial processes account for all Wajid Sandstone features observed in the Bani Khatmah area. We saw no evidence to indicate a glacial environment for the upper Wajid Sandstone. Further, in view of the probable early Palaeozoic rather than Permian and Carboniferous age of the Wajid Sandstone, a Gondwana glacial origin is not possible.

Helal (1965) suggested a glacial origin for the upper Wajid on the basis of age criteria and grain-size analyses from the lower 50 m of the section at Bani Khatmah. He showed histograms with flat bimodal character (Helal, 1965: fig. 2) indicative of tillite, and also presented a 50 m section of a massive till-like deposit. We observed no such massive deposit at Bani Khatmah. The entire section of upper Wajid Sandstone in the area is well bedded, commonly cross-bedded, and the coarse clasts are in channels and beds. Also, our grain-size analyses are sharply unimodal (Fig. 25.5-A) and characteristic of normal fluvial sedimentation (Fig. 25.5-B). Helal described rafted blocks. We observed rounded clasts of various sizes in sandy matrix, but always in conglomerate channels or in strongly cross-bedded units. Finally, Helal stated that some boulders have striations indicative of ice abrasion. We did not observe any striations on boulders, and if they occur, they are not abundant or obvious.

PEDIMENTS AND LAG BOULDERS

Two dissected pediments are exposed at the base of the Bani Khatmah escarpment: an older high pediment and a younger low pediment (Pls. 25.IV, 25.V). Each pediment is covered with a lag of large resistant clasts derived from the conglomerate beds of the

Pl. 25.IV. High pediment (HP), low pediment (LP), and modern dissection channels (DC) as seen from the top of the Bani Khatmah scarp. The low pediment is not well developed at this locality. The terrace marked HP in the centre of the photograph is about 100m across.
The high pediment is exposed only as isolated remnants (Pls. 25.IV, 25.V) but was initially graded to the uppermost part of the Wajid Sandstone, which was exposed to erosion at the time the pediment was formed. The high pediment has larger boulders in its gravel deposit than the low pediment, which is commensurate with its steeper gradient. The gravel deposit of the high pediment is saprolitic and is about 2 m thick; its surface is covered with a lag of resistant Precambrian igneous and metamorphic boulders and cobbles that are subrounded to rounded.

Boulders 30 cm in diameter are abundant, and boulders 1 m in diameter are conspicuous; the largest observed was 2 m in diameter (Pl. 25.II).

The low pediment formed much later than the high pediment. It lies about 3 m above the modern stream channel, is graded to the lower part of the Wajid scarp, and is only moderately dissected (Pls. 25.IV and 25.V). The gravel deposit of the low pediment, in general, contains finer debris and has a much less developed and relatively inconspicuous lag gravel surface than the high pediment.

Saprolitic weathering during multiple cycles of erosion, both during the retreat of the Wajid Sandstone scarp and during the dissection of the pediment gravel deposits, has locally resulted in a relatively large ac-
cumulation of resistant boulders of igneous and metamorphic rocks on the pediments and in the modern stream channels. These erosional boulders represent a very large volume of denudated Wajid Sandstone. The boulders are derived from normal fluvial conglomerates in the Wajid, and exotic sedimentary processes during Wajid deposition are not necessary to account for them.

BEDROCK SOURCE OF PRECAMBRIAN DETRITUS

The bedrock source of the coarse Precambrian clastic debris in the upper Wajid remains an unresolved problem. The quartz sand deposition of the Wajid and Saq Sandstones across a north-south distance of 1500 km in Arabia is characteristic of the erosion of a stable, deeply weathered, low relief crystalline terrane. The coarse debris in the upper Wajid, even as limited in distribution as it appears to be, suggests sharp uplift in the source area.

Final cratonisation of the Arabian Shield occurred during the latest Precambrian to Cambrian time, after the Najd orogeny (Brown, 1972). Since then, the Arabian craton has been stable, aside from epeirogenic movements such as, for example, those in the central Arabian arch (Powers et al., 1966: D102). Pre-Khuff erosion of the arch accounts for the discontinuity of the Wajid and Saq-Tabuk Formations beneath the continuous Permian Khuff Formation in central Arabia (Figs. 25.1 and 25.3). Such broad swells, however, can hardly account for a rough relief that would produce boulders like those observed in the upper Wajid.

The transport direction of the lower, middle, and upper Wajid in the latitude of Bani Khatmah is from the southeast. About 120 km south-southeast of Bani Khatmah, Precambrian rocks overlain by Cretaceous sandstone are exposed (U.S. Geol. Survey—Arabian American Oil Co., 1963). The intervening area is the sand desert of the western Ar Rub' Al Khali, and, unfortunately, little is known of the Palaeozoic history of this part of southern Arabia.

In the area between Aden and the Dhufar, Beydoun (1966) considered scattered outcrops of deformed immature clastic and carbonate rocks to be of early Palaeozoic age, but later (Beydoun, 1970) considered them to be of late Precambrian age. As a result, the only reported lower Palaeozoic sedimentary rocks in southern Arabia occur in the Dhufar, and they constitute a small outcrop of sandstone and micaceous shale. A thick section of Upper Cambrian and Lower Ordovician clastic and carbonate rocks in Oman (Beydoun, 1966: H15; Morton, 1959) was folded and partly metamorphosed in pre-Middle Permian times (Glennie et al., 1973), which suggests that a Palaeozoic orogeny may have occurred in at least eastern Arabia. Hence, the nature of the ultimate source of the Precambrian debris in the upper Wajid conglomerate beds remains a moot question until more is known of the early Palaeozoic palaeogeography of southern Arabia.

ACKNOWLEDGMENTS

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Characteristics of Modern Glacial Marine Sediments: Application to Gondwana Glacials

L. A. FRAKES and J. C. CROWELL

ABSTRACT

On the basis of structural, textural and geo-chemical studies of modern glacio-marine sediments from Antarctic waters and in the Beaufort Sea, it is apparent that basic differences exist between these sediments, and those laid down on land as till. Relative to the ‘crustal abundance’ value, glacial marine deposits are depleted in iron like most nearshore marine sediments, whereas tills are enriched—this probably results from transport of reduced iron in seawater beyond the continental shelf. Glacial marine sediments are also frequently laminated and many are slightly better sorted than tills. Using these criteria, many of the forty-six samples of Gondwana glacial rocks analysed appear to have been deposited subaqueously, and it is likely that glacial marine environments were more widespread than previously believed.

INTRODUCTION

Among diamictites of the Gondwana late Palaeozoic are many which display evidence of the presence of water in motion during deposition. This is usually seen in the development of stratification on the scale of a few centimetres, and in some instances it can be further supported by enclosed marine faunas. Also strata associated with diamictites include such clearly water-laid deposits as varve-like strata with dropstones and glacio-fluvial beds. This opens the possibility that many non-stratified diamictites originated through subaqueous deposition.

Unfortunately, criteria for distinguishing water-laid diamictites are scarce, in part because modern subaqueous environments of glacial deposition are poorly known. This paper examines some of the structural, textural and compositional characteristics of Recent glacial marine sediments from polar regions and attempts to contrast these with features of Pleistocene till. Sufficient differences exist to allow discrimination among late Palaeozoic ‘tillites’; many of these were probably deposited under subaqueous conditions.

STRATIFICATION

Stratified diamicts are known from glacial deposits of all ages including the Pleistocene. They were termed ‘stratified drift’ by Flint (1971). Such deposits can be separated into the easily recognised ice-contact types, which frequently display gross and irregular lithologic changes, and those originating through the more subtle imprint of the sorting action of water. Stratification in late Palaeozoic diamictites can be observed in many outcrops of all depositional basins, with bedding units ranging from a few millimetres to many metres in thickness. Some thin units are identical to bedded diamictites intercalated in varve-like sequences and these probably originate from increased abundance of dropstones in a subaqueous environment during single or multiple warm intervals. Emplacement by downslope movement of till or by
deposition directly from melting ice does not seem likely because the underlying layers are rarely eroded or disturbed.

In thirty-two piston cores containing 192 m of modern glacial marine sediments from the Ross Sea, Antarctica, stratified diamictons (1 mm-1 m thick) constitute approximately 30 per cent of the whole. Here, and in the Weddell Sea, bedding probably results from reworking of glacial debris by bottom currents (Chriss and Frakes, 1972; Anderson, 1972). Surface exposures of late Palaeozoic diamicrites frequently show only a small percentage to be stratified, but fresh cores from boreholes commonly show the figure to be much higher. For example, in seven boreholes through the Dwyka Formation of the Karroo Basin (total core about 2700 m), approximately 70 per cent of the diamicrite is stratified, most at the scale of a few centimetres or smaller. Though the precise mode of origin of the stratified Dwyka is problematic, it seems that flowing water must have played a large role in their deposition. The abundance of stratified diamicrite, here as in other central basin localities throughout Gondwanaland, leads to the speculation that many of these glacial deposits were laid down under subaqueous conditions.

**Texture**

The texture of a diamicrite can perhaps be useful in determining whether or not the sediment was deposited subaqueously. In the formation of subaerially deposited till, all material is dropped as the ice melts, and reworking by meltwater is only locally effective. This gives rise to the typically poorly sorted and non-stratified diamict called till. On the other hand in the formation of marine glacials fine debris below a certain size limit from floating icebergs is subject to current action and hence is widely dispersed during the fall to the bottom. Thus, we can expect that subaquously deposited glacial detritus should be somewhat better sorted than till.

Inclusive Graphic Standard Deviation, a measure of sorting, is plotted against Mean Size, all in phi units, in Figure 26.1. There

![Fig. 26.1. Mean size v. inclusive graphic standard deviation of tills (Landim and Frakes, 1968) and glacial marine sediments (Ross Sea, Chriss and Frakes, 1972; Beaufort Sea, Nayudu, 1970). All in phi units.](image)
are two sets of data for the glacial marine environment—sixty-three samples from the Ross Sea (Chris and Frakes, 1972) and twenty-eight samples from the Beaufort Sea (Nayudu, 1970)—and data for ninety-seven tills (summarised by Landim and Frakes, 1968). Beaufort Sea samples cover a size range comparable to that of the tills and the apparent differences when compared with tills may thus be real. The Ross Sea data and those for the tills are not statistically comparable in that they are calculated over different size ranges. However, this wide sampling of tills shows that they are more poorly sorted than glacial marine sediments of the Beaufort Sea and probably than Ross Sea materials as well. All but two till samples are very-poorly to extremely-poorly sorted, while most Ross Sea samples fall into the poorly sorted category. Additional analyses of tills and glacial marine sediments are needed to substantiate this.

Samples from diamictites can be compared with this foregoing data by grain-size measurement in thin section (300 grains in the range —2 to 6 phi, by point count and with application of a size correction factor for sliced spheres). In comparison (Fig. 26.2) note that the six samples thus far analysed fall within the glacial marine field, although two (Ohio Range and Hill View) fall near the till boundary. The indication is that subaqueous glacial deposits occur in the Pensacola and Sentinel Mountains of Antarctica, in the Wynyard Tillite from Zeehan, Tasmania, and possibly in the Wisconsin Range of Antarctica. A marine origin has been suggested for diamictites at the first two localities (Frakes, Matthews and Crowell, 1971).

**ELEMENTAL ABUNDANCES**

In view of the subaerial nature of till deposition from melting of continental ice it is likely that conditions are suitable for oxidation of till materials. The environment during diagenesis may be either oxidising or re-

![Fig. 26.2. Size and standard deviation values from point counts of thin sections of six late Palaeozoic diamictites, plotted in the till-glacial marine fields from Fig. 26.1.](image-url)
ducing, but the former seems more likely. For glacial sediments laid down under subaqueous conditions the incidence of reducing environments is probably greater, as it would be also during diagenesis. An analogy may be made with sediments deposited under reducing conditions in a fjord in British Columbia (Presley, Kolodny, Nissenbaum and Kaplan, 1972). Elements which precipitate during oxidation processes may be expected to be concentrated in tills and depleted in glacial marine sediments. Comparison of the types (Fig. 26.3) shows that twenty-one tills (Roaldset, 1972) display high abundances of total iron, relative both to the crustal abundance value and to those of glacial marine materials (eighty samples from West Antarctic waters (Angino, 1966); nine samples from around Antarctica analysed by atomic absorption in this study). Though the differences are marked more data are needed to delineate the fields suggested here.

Fig. 26.3. Fe v. Mn in tills (Roaldset, 1972), modern glacial marine sediments (Angino, 1966 and Table 26.1) and Cainozoic glacial marine sediments from Antarctica (this study). All values in weight per cent (±10 per cent).

Angino (1966) observed that glacial marine sediments from around Antarctica are notably deficient in Fe, Ti and Ni and enriched in V and Cu with respect to crustal abundances. It is likely that variations in provenance and/or biological activity strongly influence the abundance of Ni, V and Cu as the nine samples of the current study contain markedly variable amounts of these elements (Table 26.1). However, glacial marine sediments seem to be consistently deficient in Fe and further, Mn is definitely above crustal estimates (950 ppm) in the preponderance of samples. A chemical explanation for this
Table 26.1. Abundance of Selected Elements in Surface Sediment from Nine *Eltanin* Piston Cores near Antarctica, compared with Estimated Crustal Abundances

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude (S)</th>
<th>Longitude</th>
<th>Fe %</th>
<th>Mn ppm</th>
<th>Ni ppm</th>
<th>V ppm</th>
<th>Cu ppm</th>
</tr>
</thead>
<tbody>
<tr>
<td>E 7-12</td>
<td>66°33'</td>
<td>48°09'W</td>
<td>3.13</td>
<td>792</td>
<td>45</td>
<td>121</td>
<td>50</td>
</tr>
<tr>
<td>E12-8</td>
<td>64°05'</td>
<td>40°35'W</td>
<td>3.82</td>
<td>939</td>
<td>61</td>
<td>133</td>
<td>63</td>
</tr>
<tr>
<td>E22-23</td>
<td>62°26'</td>
<td>19°03'W</td>
<td>4.18</td>
<td>1434</td>
<td>75</td>
<td>147</td>
<td>98</td>
</tr>
<tr>
<td>E22-24</td>
<td>63°02'</td>
<td>14°58'W</td>
<td>4.61</td>
<td>1248</td>
<td>82</td>
<td>127</td>
<td>30</td>
</tr>
<tr>
<td>E27-4</td>
<td>68°03'</td>
<td>17°35'E</td>
<td>4.50</td>
<td>728</td>
<td>58</td>
<td>106</td>
<td>106</td>
</tr>
<tr>
<td>E27-21</td>
<td>69°02'</td>
<td>17°50'E</td>
<td>3.76</td>
<td>883</td>
<td>53</td>
<td>95</td>
<td>71</td>
</tr>
<tr>
<td>E33-2</td>
<td>77°00'</td>
<td>164°34'W</td>
<td>3.19</td>
<td>1251</td>
<td>81</td>
<td>53</td>
<td>172</td>
</tr>
<tr>
<td>E33-4</td>
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<td>2044</td>
<td>109</td>
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<td>127</td>
</tr>
<tr>
<td>E33-10</td>
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<td>125°17'W</td>
<td>3.18</td>
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<td>70</td>
<td>71</td>
<td>121</td>
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<tr>
<td>Crust</td>
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<td></td>
<td>5.63</td>
<td>990</td>
<td>75</td>
<td>135</td>
<td>55</td>
</tr>
</tbody>
</table>

awaits determination of the valence states of the two elements, but reduced Fe in solution in low pH seawater would not adsorb on clay particles as it does under oxidising conditions. Possibly low pH and negative Eh are suggested by the common presence of well-preserved microfloras both in modern Antarctic glacial marine sediments and in late Palaeozoic diamicrites (Kemp, 1972; this volume). On the other hand, Mn may be concentrated because of decreased detrital flux under the present very low accumulation rates in the Ross Sea.

In attempting to apply this method of separation of unlithified glacial materials to Palaeozoic diamicrites, one must have some knowledge of the diagenetic and weathering history of the rocks. An idea of how diagenesis has affected glacial sediments can be gained from our unpublished analyses of fossiliferous Tertiary glacial-marine sediments recently discovered in Antarctica. When compared with their modern analogues, samples from these strata are greatly depleted in both Fe and Mn (Fig. 26.3). This indicates that diagenesis in the Ross Sea has taken place under reducing conditions for otherwise we would expect a concentration of the oxides of these elements. On the other hand the effects of weathering in the current cycle cannot be completely known for any Palaeozoic diamicrite. However, in oxidation Fe and Mn should be concentrated, at least in part through massive removal of major ions such as Ca, Na and Mg, but also possibly by introduction of Fe and Mn from adjacent terranes.

Forty-six late Palaeozoic diamicrites were analysed for abundance of Fe, Mn, Ni, V and Cu by atomic absorption (Table 26.2). The results are plotted on Figure 26.4, on which three fields are delineated—till, modern glacial marine and 'old' glacial marine, from Figure 26.3. The majority of diamicrites plot within the area of old glacial marine sediments, suggesting that they may have been subjected to generally reducing conditions during deposition, diagenesis and weathering. These include diamicrites from three localities containing marine fossils (Tarija Formation, Bolivia; Lower Tepuel Group, Argentina; Itararé Sub-Group, Brazil). Another fossiliferous diamicrite from the Lower Tepuel Group at La Carlota is somewhat more enriched in Fe and Mn and falls in the field of modern glacial marine sediments. Seven samples have similar concentrations to those of modern glacial marine sediments, as does a diamicrite from the Gale Mudstone, Pensacola Mountains, Antarctica analysed by Schmidt and Williams (1969). Another seven are greatly deficient with respect to Fe and six of these are enriched in Mn, perhaps reflecting the differences in ionic potential between the two elements. A sample from immediately above a striated floor at Nooitgedacht in South Africa and thought to be located in the highland centre from which ice radiated, is the only one to fall within the till field. However, since obviously oxidised and highly permeable samples were not selected in this study, the relative scarcity of samples in the till field should not be taken to indicate a general lack of till deposition in Gondwanaland.
Table 26.2. Abundance of Selected Elements in Forty-six late Palaeozoic Diamictites of Gondwanaland

<table>
<thead>
<tr>
<th>Locality</th>
<th>Sample</th>
<th>Fe %</th>
<th>Mn ppm</th>
<th>Ni ppm</th>
<th>V ppm</th>
<th>Cu ppm</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>SP₄</td>
<td>2.64</td>
<td>141</td>
<td>37</td>
<td>23</td>
<td>14</td>
<td>marine fossils</td>
</tr>
<tr>
<td>Argentina</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tarifa Fm.</td>
<td>ZH₄</td>
<td>2.01</td>
<td>158</td>
<td>23</td>
<td>42</td>
<td>9</td>
<td></td>
</tr>
<tr>
<td>Jaguel Fm.</td>
<td>JAG</td>
<td>3.01</td>
<td>486</td>
<td>39</td>
<td>24</td>
<td>25</td>
<td></td>
</tr>
<tr>
<td>Yalguaraz Fm.</td>
<td>YA</td>
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<td>1305</td>
<td>47</td>
<td>65</td>
<td>32</td>
<td></td>
</tr>
<tr>
<td>Bajo De Veliz</td>
<td>BV₁</td>
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<td>204</td>
<td>9</td>
<td>18</td>
<td>4</td>
<td>microflora</td>
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<tr>
<td>La Carlota</td>
<td>LC₁</td>
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<td>28</td>
<td>15</td>
<td>11</td>
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<tr>
<td>Tepuel</td>
<td>ST₅</td>
<td>3.04</td>
<td>404</td>
<td>37</td>
<td>40</td>
<td>24</td>
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<td>Brazil</td>
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<td>Ponta Grossa</td>
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<td>2.47</td>
<td>127</td>
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<td>microflora</td>
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<td>SC₁</td>
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<td>'pelodite', microflora</td>
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<td>23</td>
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<td>140</td>
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<td>65</td>
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<tr>
<td>Abraamkraal BH</td>
<td>AB₂</td>
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<td>110</td>
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CONCLUSIONS

Three features which may be useful in determining mode of origin of diamictites appear to characterise modern glacial marine sediments: (1) fine bedding and lamination are common; (2) sorting is poor, but better than in tills; and (3) total Fe appears to be less abundant in glacial marine sediments than in tills. Application of these tentative criteria to fifty-two samples of late Palaeozoic diamictite suggests that deposition in subaqueous environments may have been more common than previously believed. On the basis of the very low abundance of Fe in these rocks it seems likely that few have been subjected to extensive oxidation during deposition, diagenesis or weathering, and hence that many diamictite accumulations such as in the Karroo Basin, the Paraná Basin, the eastern Falkland Islands, the Sentinel and Pensacola Mountains of Antarctica, and in the basins of the Andean trend in South America must have been deposited subaqueously. Tills are no doubt present in the late Palaeozoic sequences, even among the analysed samples, but the criteria utilised here suggest that glacial marine strata are abundant throughout Gondwanaland.

ACKNOWLEDGMENTS

We wish to take this opportunity to express our deep appreciation for the cooperation of the staff of the Antarctic Marine Geology Facility at Florida State University: D. S. Cassidy, Yang-Ja Park Chung, LaVerne Lamb and Rosemary Raymond, and to those who performed the analyses: R. Immel, A. Kaharoeddin and M. Miyajima.

REFERENCES


Late Palaeozoic Glacial Faunas of Australia: Distribution and Age

GRAHAM McCLUNG

ABSTRACT

Late Palaeozoic marine invertebrate faunas occur in glacial sediments in Australia, India, southern Africa and South America. Three faunal assemblages are recognisable. The oldest is confined to Argentina and may be of Early Carboniferous age. An intermediate fauna is of Late Carboniferous age and is recognisable in eastern Australia and Argentina. The youngest fauna is recognisable in all areas. It indicates that the most widespread period of glaciation began in the latest Carboniferous, and possibly continued into the earliest Permian.

INTRODUCTION

Glacial depositional features are recognisable in Late Palaeozoic marine sediments containing invertebrate faunas in Australia, southern Africa, South America and India. This paper summarises the composition and distribution of such faunas in Australia, where the biostratigraphy is best displayed. Comparable faunas from elsewhere in Gondwanaland are then considered in relation to Australian sequences, and the implications of spatial and temporal variations in these faunas with respect to Gondwana palaeogeography are evaluated.

Only marine invertebrates are discussed. The resulting correlations can be valid only for marine sediments, which in many areas make up a small proportion of the total glacial sequence. The conclusions of this paper may be regarded as a framework for comparison with other schemes based on microfloras, macrofloras and vertebrate faunas.

DISTRIBUTION OF AUSTRALIAN GLACIAL MARINE SEDIMENTS

The relationship of the distribution and development of glaciers to depositional and erosional features characteristic of glaciation is described by Crowell and Frakes (this volume). The most common manifestations of glaciation in marine sediments are of three types: near shore sublittoral deposits in which the effects of glaciation may be partly obscured by reworking; shelf sediments which contain dropstones; and mass movement deposits in which glacial sediments are transported downslope from the area of their initial deposition. Only the first type provides evidence of nearby ice sheets and usually is a result of deposition from retreating glaciers. Although many dropstones imply the presence of floating ice, the location of the source area and type of ice can rarely be determined in ancient sediments. The spatial relationships between mass movement deposits and the glaciers which contributed material to them may be equally difficult to interpret.

In this paper greatest consideration is given to sublittoral sediments which show glacial features. Such sediments are widespread in Australia (see Fig. 27.1) and occur in the earliest deposited units of a marine sequence in which overlying formations are of Permian age. In many cases the glacial marine sediments are the highest beds in a sequence deposited in terrestrial and transitional environments. Invertebrate faunas...
found in these marine sediments are usually of low specific diversity, although numerous individuals may be present. This lack of variety in indicative of a difficult environment and low water temperatures (Runnegar, in press).

The distribution and relationships of Australian late Palaeozoic glacial sediments deposited in a sublittoral environment are shown in Figure 27.1. Papers describing the sequences or associated faunas are cited in the explanation. In New South Wales and Tasmania marine shelf sediments overlying the glacials contain common dropstones and erratics, but it is only in the lowest beds of the sequence that the sediments indicate the existence of nearby glaciation. Dropstones are also known from Queensland, although not from the Yarrol sequence in which sublittoral sediments show a glacial imprint. Since these sediments contribute little information on the location of the ice source, they will not be considered further in this paper.

**INTRACONTINENTAL FAUNAL CORRELATIONS**

The general homogeneity of Australian invertebrate faunas characteristic of the glacial sequences has been recognised for some time (Thomas and Dickins, 1959; Dickins, 1964, 1970; Runnegar, 1969b; Waterhouse, 1970). More recent work has allowed finer faunal subdivision of this interval (Runnegar and McClung, this volume; Clarke and Banks, this volume).

The critical interval includes the earliest deposited units of the northern Sydney Basin, in which fluvial and lacustrine sediments of the Seaham Formation pass transitionally into marine sediments at the top of that unit and in the overlying Dalwood Group (Runnegar, 1969b; McClung et al., in prep.). Within this interval three invertebrate zones have been recognised—the *campbelli*, elon-

Time units are those described by Runnegar and McClung (this volume). Numbers on the map correspond to the following localities. Information on the sequences is contained in the reference listed below. Western Australia. 1. Perth Basin, Nangetty Formation. Diamictites contain striated and faceted clasts, dropstones occur in other sediments (Clarke et al., 1951; Playford, 1959). A small fauna consists of arenaceous forams (Crespin, 1958). 2. Carnarvon Basin, Lyons Group. A striated pavement and striated clasts in diamictites have been recorded by Condon (1967). Moderately diverse faunas of bivalves and brachiopods occur in the upper parts of the unit (Dickins, 1956, 1957, 1963; Thomas and Dickins, 1959; Thomas, 1958, 1967, 1971). 3. Canning Basin, Grant Formation. Striated clasts occur in diamictites and possible dropstones are found in other sediments (Vevec and Wells, 1961). A small fauna of arenaceous forams was described by Crespin (1958). South Australia. 4. Fleurieu Peninsula, Cape Jervis Beds. Striated pavements, *roches moutonées*, tillites, striated boulders and laminated shales with dispersed clasts are associated with a microfauna of arenaceous forams (Ludbrook, 1967). Other possible glacial deposits characterised by boulder clays and faceted or striated pebbles are associated with a similar microfauna in the subsurface of the Murray Basin (Ludbrook, 1967). Victoria. 5. Bacchus Marsh area, Bacchus Marsh Beds. Striated pavements, *roches moutonées*, tillites and possible eskers are widely distributed (Spencer-Jones, 1969). Conulariids and brachiopods occur near the base of the sequence at Coimadai (Garratt, 1969; Thomas, 1969). Tasmania. 6, 7, 8, 9. Central and Western Tasmania, Wynyard Tillite and equivalents. Scattered outcrops of tillites and laminated claystones fill basement lows and are occasionally associated with striated pavements (Banks, 1962; Clarke and Banks, this volume). Sparse faunas occur in the upper parts of the unit at some localities (Banks, 1962; Jago, 1972; Clarke and Banks, this volume; Clarke, pers. comm.). Overlying the tillites are the lowermost marine units of the Parmeener Super-Group which contain occasional dropstones and erratics and are associated with a diverse fauna of brachiopods and bivalves (Banks, 1962; Clarke and Banks, this volume). New South Wales. 10. Southern Sydney Basin, Wasp Head Formation. Gostin (1968) described a striated pavement, dropstones, and diamicticites which may in part represent solifluction lobes. Bivalves dominate the fauna (Runnegar, 1969a). 11, 12. Northern Sydney Basin, Seaham Formation. Glacial striae occur with rhythmically layered sediments and dropstones (Rattigan, 1967). A sparse fauna of brachiopods and bivalves occurs in the upper-
most parts of the formation (Runnegar, 1969b; McClung et al., in prep.). Conformably overlying the Seaham Formation are the Tamby Creek Formation, Beckers Formation and Cranky Corner Sandstone at Cranky Corner, and the Lochinvar and Allandale Formations near Lochinvar. Dropstones occur both in horizons and scattered throughout the sequence (Rattigan, 1969). Moderately diverse bivalve dominated faunas are associated (Runnegar, 1969b; McClung et al., in prep.). 13. Gloucester-Nowendoc area, Giro Diamicite and Glory Vale Conglomerate. Dropstones occur in thin intervals of finely laminated sediments (Mayer, 1972), while a sparse fauna of brachiopods and bivalves are found in interbedded conglomerates and diamicites (Mayer, 1969; Runnegar, 1970). 14. Macleay Valley, Kullatine Formation. Diamicites contain rare striated clasts, while a brachiopod dominated fauna was described by Campbell (1962). Lindsay (1966) believed glaciation had little effect on deposition of this formation. Queensland. 15, 16. Western Yarrol Basin, Youlambie Conglomerate. Rhythmically layered siltstones containing dropstones, pebbly mudstones, and boulder beds with faceted pebbles occur in the lower parts of this formation, while overlying beds contain a sparse bivalve dominated fauna (Dear, 1968; Dear et al., 1971). The more highly fossiliferous Burnett Formation occurs to the east (Maxwell, 1964), and is believed to be an offshore facies equivalent (Dear, 1968; Dear et al., 1971).
384 Age and Stratigraphic Relations of Glacial Deposits

gata and konincki zones of Runnegar and Mc
Clung (this volume). These zones are based on brachiopod-dominated shelf assemblages and at present are difficult to recognise in bivalve-dominated faunas characteristic of the sublittoral sediments that are associated with most Australian marine glacial deposits. Consequently only broad similarities are recognisable between the northern Sydney Basin faunas and those of most other glacial sediments elsewhere in Australia. Table 27.1 documents the faunas of the sequences shown in Figure 27.1, and shows that many invertebrate species are common to more than one formation. Species from the Lyons Group in Western Australia are similar to those from eastern Australia, although there are slight differences in nomenclature. The Grant and Nangetty Formations in Western Australia, and the Cape Jervis Beds in South Australia lack the brachiopods and bivalved molluscs used to compile Table 27.1, but can be correlated with other formations using foraminifera (Crespin, 1958; Ludbrook, 1967). The Grant and Nangetty Formations are similar in stratigraphic position to the Lyons Group.

The close faunal similarity evident in Table 27.1 suggests that most Australian sublittoral marine glacial sediments were deposited approximately contemporaneously, probably during the waning stages of large continental ice sheets. An earlier period of glaciation is suggested by an older fauna characteristic of the levis zone of Campbell and McKellar (1969), which occurs in the Kullatine Formation in central Northern New South Wales, associated with a Rhacopteris flora (Campbell, 1962). Although beyond the scope of this paper, the same period of glaciation may also be represented by glacial sediments in the non-marine Spion Kop Conglomerate (White, 1968). Rhacopteris has also been reported from the Wynyard Tillite and equivalents in Tasmania (Clarke and Banks, this volume) and parts of this sequence may be of the same age.

GLACIATION ELSEWHERE IN GONDWANALAND

Late Palaeozoic glacial sediments occur in all other parts of Gondwanaland (Frakes and Crowell, 1967, 1969, 1970; Frakes, Amos and Crowell, 1967; Frakes, Matthews and Crowell, 1971; Crowell and Frakes, 1972; Robinson, 1967; Ghosh and Mitra, 1970a, 1970b). Small faunas of marine invertebrates are known from India, southern Africa and South America, but none have been found in Antarctica and the Falkland Islands. India. Within the Talchir Formation and correlative units, marine faunas occur at widely separated localities in the Salt Range, Peninsular India and the eastern Himalayas (Robinson, 1967; Sastry and Shah, 1964). All faunas are broadly similar in species content, with differences in species composition and abundance probably reflecting environmental variation. Species of Eurydesma, Deltopecten, Peruvispira and spiriferid and productid brachiopods are either identical or very similar to those in Australian faunas, with closest similarity to the Lyons Group assemblage (Thomas and Dickins, 1959).

Southern Africa. Small faunas occur in beds overlying and interbedded with glacial sediments of the Dwyka Series in the Kalahari, Warmbad and Great Karroo Basins (McLachlan and Anderson, 1973; Martin et al., 1970; Martin and Wilczewski, 1970; Dickins, 1961). Faunas are often poorly preserved and of low specific diversity. The presence of Eurydesma and Peruvispira indicates affinities to Indian and Australian glacial faunas, while the goniatite Eoasianites (Glaphyrites) described by Martin et al. (1970) provides possibilities of international correlation.

South America. Numerous invertebrate fossil localities are now known from glacial sequences in Argentina, Brazil and Uruguay. The faunas have recently been reviewed by Amos (1964), Rocha-Campos (1967, 1970a, 1970b), Amos and Sabattini (1967) and Runnegar (in press). At least four faunal assemblages are recognisable, differing in composition in relation to their stratigraphic position.

The oldest fauna occurs in the lower beds of the sequence in the Rio Blanco and Patagonian Basins of Argentina. Amos (1964) reports the presence of the genera Productella and Geniculifera but these identifications may be in need of review.

A younger fauna occurs in the Patagonian and Calingasta-Uspallata Basins in Argentina. Levipustula levis and other brachio-
Late Palaeozoic Glacial Faunas 385

pods and bryozoans indicate close affinities to faunas of the Kullatine Formation and other units in eastern Australia (Amos and Sabattini, 1967). Miller and Garner (1953) described the associated goniatite genera *Anthracoceras* and *Eoasianites*.

Many localities higher in the sequence in Argentina and Brazil contain typical Gondwanan genera such as *Eurydesma, Deltotpecten, Pyramus, Myonia, Australomya, Keeneia* and *Peruvispira* (Rocha-Campos, 1967, 1970a; Runnegar, in press; Harrington, 1955). In Argentina the brachiopods *Cancriella* and *Martiniopsis* are associated with these molluscs, or occur at similar stratigraphic horizons (Harrington, 1955; Amos and Sabattini, 1967; Runnegar, in press). Species of these brachiopods closely resemble those from eastern Australia (Runnegar, in press). The goniatite *Eoasianites (Glabphyrites) rionegrensis* is found at a similar stratigraphic position in Uruguay (Gloss, 1967).

The youngest assemblage occurs in sediments of the Guata Sub-Group in Brazil (Rocha-Campos 1967, 1970a, 1970b). Runnegar (in press) recognised both Tethyan and Gondwanan affinities in the fauna, and remarked on similarities to Australian Artinskian species. The relationship of this fauna to glacial sediments is not clear. Glacial features have been recorded from the Guata Sub-Group (Rocha-Campos, 1967; Frakes and Crowell, 1969) but their position with respect to the marine units is uncertain.

THE AGE OF GONDWANALAND GLACIAL FAUNAS

Within the interval including glacial sediments, invertebrate faunas with general Carboniferous affinities are replaced by assemblages having more in common with Australian Permian species. Plants including *Glossopteris* replace *Rhacopteris*-dominated floras in this interval. In Australia these faunal and floral changes have been arbitrarily used as indicators of the Permian-Carboniferous boundary. South American workers believed similar changes took place during the Late Carboniferous. This disagreement in the age of the glacial sediments results from the endemic nature of Gondwanaland floras and faunas. Changes in microfloral composition also occur within this interval, but there appear to be greater possibilities of correlation with northern hemisphere sequences (Helby, pers. comm.). Despite these problems it is worthwhile reviewing the meagre evidence at present available for the age of the glacials and the position of the Permo-Carboniferous boundary.

Limits can be placed on the beginning and end of widespread glaciation in Australia. The goniatite *Cravenoceras kullatinense* Campbell occurs below sediments containing striated clasts in the Kullatine Formation and is closely comparable with European species of Namurian age (Campbell, 1962). The *Levipustula levis* fauna is probably of Westphalian age (McKellar, 1965), although *Levipustula* may have a longer range (Kotlyar and Popeko, 1967). *Levipustula levis*, one of the species used by McKellar (1965) to determine an age for this fauna, is long ranging in eastern Australia. The report by Lindsay (1969) of *L. levis* from beds underlying the *C. kullatinense* locality has recently been confirmed (Northcott, pers. comm.). Mayer (1972) also reported the species from an assemblage otherwise typical of the late Visean *Rhipidomella fortimucula* zone. Helby (1969) believed microfloral assemblages from the upper Isaacs Formation, laterally equivalent to the partly glacial Sea­ham Formation, were of late Stephanian age, but subsequent work indicates a significantly greater age (Helby, pers. comm.).

The Holmwood Shale overlies the Nag­getty Formation in the Perth Basin, and contains the goniatites *Juresanites* and *Uraloceras*, which resemble species from the Sak­marian Substage of Russia (Glenister and Furnish, 1961). Goniatites from other formations stratigraphically higher than the marine glacial sediments in Queensland, New South Wales and other parts of Western Australia include latest Sakmarian and Artinskian species (Glenister and Furnish, 1961; Armstrong et al., 1967). This limits the age of Australian marine glacial sediments discussed in this paper to within Namurian to Sak­marian time, while dropstones in many eastern Australian sequences are common throughout the Artinskian and Kungurian.

Until the fauna associated with the earliest glacial sediments in Argentina is reviewed it
is difficult to assess their age. Amos (1964) considered it to be Early Carboniferous. Younger glacial sediments contain *Levipus-tula* and the goniatites *Anthracoceras* and *Eoasianites*. Miller and Garner (1953) compared these goniatites with middle Pennsylvanian species from North America. The occurrence of species of *Eoasianites* (*Glaphyrites*) in South-West Africa and Uruguay cannot be directly related to other invertebrate assemblages. However, they are found in formations which elsewhere have a *Eurydesma* fauna. Both species appear to be most closely related to Late Carboniferous

### Table 27.1. The Distribution of Invertebrate Species in Australian late Palaeozoic Glacial Sediments

<table>
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<th>Species</th>
<th>Burnett Formation</th>
<th>Youallabe Conglomerate</th>
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<th>Tamny Creek Formation</th>
<th>Booroomin Formation</th>
<th>Leichhardt Formation</th>
<th>Woodhead Formation</th>
<th>Backus Marsh Beds</th>
<th>Wyndham Tillite</th>
<th>Tasmania, Faunizone 1</th>
<th>Tasmania, Faunizone 2</th>
<th>Tasmania, Faunizone 3</th>
<th>Cape Jervis Beds</th>
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Montospira spp.
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Strophalosia sp.
S. subcircularis
Wyndhamia sp.
Costalosia apicallosa
Levipustula levis
Cancrinella levis
C. lyoni*
Composita magnicarinia
Pseudosyrinx allandalensis*
Cyrtella magnumaniensis australis*
Trigonotrema campbelli*
T. stokesi*
T. narshensis cordialis*
Sulcoplepis stutchburi
S. sp. nov. A.*
S. sp. nov. B.
Spinuliplic spinnulosida
Spiriferellina australis
Martiniopsis elongata*
M. konincki*
M. sp. nov.
Fletcherihyris spp.
Gilledia spp.
Conularids
Foraminifera

| * Brachiopod and bivalved mollusc species with an asterisk are regarded as significant for correlation. In most cases illustrations or actual specimens have been examined. The original generic designations of some species have been changed to correspond to more recent studies. Information on species distribution was taken from the following sources: Campbell (1962); Clarke (1969, 1970); Clarke, pers. comm. (1973); Clarke and Banks (this volume); Coleman (1957); Crespin (1958); Dear (1968); Dear et al. (1971); Dickins (1957, 1958, 1963); Fletcher (1958); Ludbrook (1967); McClung et al. (in preparation); Maxwell (1964); Runnegar (1965, 1967a, 1967b, 1969a, 1969b, 1970a, 1970b, in press); Thomas (1958, 1967, 1971); Thomas and Dickins (1959).

forms from the northern hemisphere (Martin et al., 1970; Closs, 1967). In Bolivia glacial sediments are overlain by limestones of the Copacabana Group which contain Wolfcampian to early Leonardian fusulinids and brachiopods (Helwig, 1972). The widespread Eurydesma faunas of Gondwanaland contain sufficient species in
common to demonstrate broad contemporaneity of these sequences and their associated marine glacial sediments. The age of these sediments with respect to the Permian-Carboniferous boundary is not clear, but the traditional Australian view that the system boundary coincides with the first appearance of a *Eurydesma* fauna, or with the change from a *Rhacopteris* to a *Glossopteris* flora has little direct faunal evidence to support it. Some of the genera previously regarded as characteristic of the *Eurydesma* fauna are now known to occur in sediments as old as late Visean or early Namurian (Runnegar, in press). In fact the evidence available from invertebrate fossil distribution suggests that the most widespread period of glaciation in Gondwanaland occurred towards the end of the Carboniferous period, but was preceded by less widespread glaciation in eastern Australia and Argentina. The faunas are too meagre to allow confident correlation with northern hemisphere or Tethyan sequences, but direct faunal correlations slightly favour a Late Carboniferous age. Since these sediments are overlain by formations containing Sakmarian faunas in Western Australia, glaciation may have continued into the Asselian or Sakmarian in some areas.

**SYNTHESIS**

The earliest recognisable period of glaciation may have occurred in Argentina during the Early Carboniferous. No record of glaciation of this age exists elsewhere in Gondwanaland. In world maps for this period Smith *et al.* (1973) place Argentina (and southern Africa) in a polar position, whereas Australia is much closer to the palaeoequator. The widespread and diverse Early Carboniferous faunas of eastern Australia are quite different from Argentinian assemblages, and are interpreted by Campbell and McKellar (1969) as characteristic of warm waters. Alpine or piedmont glaciation began in eastern Australia and Argentina in post-Namurian time. This event may have been a precursor to the slightly younger period of continental glaciation which affected sedimentation in all parts of Gondwanaland. It is during this period that widespread faunal and floral changes took place. The age of this glaciation remains problematical, but it probably began late in the Carboniferous and may have continued into the earliest Permian. In India and much of Australia glaciation had ceased prior to Artinskian time. In southeastern Australia and possibly in Brazil dropstones and other glacial features in Permian sediments indicate the continuation of frigid or cool temperate conditions into Artinskian and later time.

**ACKNOWLEDGMENTS**

Stratigraphic information and photographs of spiriferids from glacial sediments at Bacchus Marsh were provided by Dr George Thomas. Mr M. J. Clarke made available an early draft of a paper on the Tasmanian sequence (Clarke and Banks, this volume) and also provided other information. I would like to thank Dr Bruce Runnegar for numerous helpful comments and criticisms during preparation of the manuscript.

**REFERENCES**


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Significance of Early Permian Marine Faunas of Peninsular India

S. C. SHAH and M. V. A. SASTRY

ABSTRACT

A study of the marine fauna found in the dominantly non-marine Permian Gondwana sequence of Peninsular India has enabled the Talchir and Karharbari Formations to be dated as Asselian and Sakmarian respectively. The base of the Gondwana sequence is probably of very Early Permian age. Additional data including the discovery of an *Eurydesma* fauna at two new localities and a tectonic interpretation of the Narmada-Son line, support two short-lived marine transgressions into Peninsular India. Both came from the north at slightly different times.

INTRODUCTION

Marine faunas have been found at four localities in the lower part of the Gondwana sequence in Peninsular India (Fig. 28.1). They are the only known marine faunas between the Cambrian (Vindhyan) and the Jurassic in the peninsula. The sequences at these localities, together with other important sequences, are shown in Figure 28.2. A study of the stratigraphic relationships and the faunal assemblages from these beds has led to the following conclusions. First, the marine beds at Manendragarh and Rajhara occur in the lower and middle parts of the Talchir Formation, while those at Umaria and Badhaura occur in the basal part of the overlying Karharbari Formation. Second, the faunal assemblages at Manendragarh and Rajhara are different from those at Umaria and Badhaura. The faunas from the first two localities show a close resemblance to those of Khemgaon-Wak in Sikkim, and the older faunas of Kashmir and the Salt Range, whereas those from the last two localities resemble younger faunas from Kashmir. Although the fauna from the Amb Formation in the Salt Range is younger than that from Manendragarh and Rajhara, its relationship with the faunas from Umaria and Badhaura is still not clear. References to published descriptions of these faunas are given in Waterhouse (1970) and Ahmad (1972).

AGE OF TALCHIR FORMATION AND LOWER AGE LIMIT OF GONDWANA SEQUENCE

In Peninsular India, the Talchir Formation is a continental deposit with marine intercalations at Manendragarh and Rajhara. In the extra-peninsular region at the Salt Range, Kashmir, and Sikkim, the equivalent beds are mainly marine with a few zones of continental deposits. The Talchir Formation has as its base a boulder bed that forms a relatively small part of the whole unit. The marine fossiliferous zones are not overlain by boulder beds. A glacial origin for at least some of the boulder beds at a few places in Peninsular India is now generally accepted (Ghosh and Mitra, 1972). The *Eurydesma* bearing rocks of Manendragarh and Rajhara, which occur just above the boulder beds, are considered by us to be of Asselian age by comparison with the similar faunas from Western Australia. A similar age can be assigned to the whole of the Talchir Formation because it is overlain by the Umaria marine beds of Sakmarian age. This would also imply that the lower limit of the Gond-
Fig. 28.1. Early Permian marine transgressions in Peninsular India

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Fig. 28.2. Correlation chart of Early Permian sediments of the Indian region. Marine units are underlined.
wana sequence may be of earliest Permian age, although the glaciation may have extended back into the Late Carboniferous. Sediments above the Talchir boulder bed or its equivalents would be younger than Carboniferous.

MARINE TRANSgressIONS IN PENINSULAR INDIA

In view of the isolated small occurrences of the marine beds in otherwise widely distributed continental deposits, various contradictory suggestions of the directions and times of marine transgressions into Peninsular India have been expressed since 1921. These views were summarised by Sastry and Shah (1964) and Ahmad (1972). But our present analysis, using the information from all known Permian marine beds, suggests that the opinion expressed by us in 1964 is still valid. It is supported by the discovery of *Eurydesma* faunas in the Talchir Formation or its equivalent at Rajhara (Shah, 1969; Dutta, 1971) and Darjeeling (Acharyya, 1972), and by the recognition of the difference between the marine faunas of Umaria and Manendragarh (Lele and Chandra, 1969; Waterhouse, 1972). The argument for a weak zone along which the transgressions entered the centre of Peninsular India is supported by the recent geomorphological studies of Tewari (1968) and Choubey (1969, 1971).

COURSE OF THE MARINE TRANSgressIONS IN PENINSULAR INDIA

**Asselian Transgression**

This transgression probably came from east of Manendragarh and Rajhara because the *Eurydesma* faunas from these areas show a close resemblance to those of Khemgaon-Wak in south Sikkim. This area of transgression generally marks the northern boundary of the outcrops of continental Gondwana sediments in this region. It is also known that Gondwana continental sediments continued below the Indo-Gangetic alluvium as far as the Main Central Thrust Zone of the Himalayas in Sikkim, where pebble beds at Lachi are considered to be equivalent to the Talchir boulder beds (Muir-Wood and Oakley, 1941).

The direction of this transgression was parallel to the present Son River drainage, except in its upper reaches. The ancient drainage in the Rajhara area during the Early Permian was towards the northeast or north-northeast (Israili, 1966). Further westwards, palaeocurrent and sedimentological data from west of Manendragarh indicate an Early Permian palaeoslope towards the east (Banerjee, 1964). The southern limit of the transgression probably coincided with a zone of weakness that is the eastern continuation of the Satpura trend (West, 1962).

**Sakmarian Transgression**

In Sakmarian time, the marine transgression apparently entered Umaria from the west and not from the east. This must have been along the Narmada rift. Auden (1949) has concluded that this rift is a major crustal feature of ancient origin that influences the deposition and folding of Vindhyan (Late Precambrian-Cambrian?) and Gondwana sediments. Along the existing Narmada River Valley, three post-Permian marine transgressions have occurred. The Narmada fracture zone, which forms the northern edge of the existing river valley, bends northwards at its western end as evidenced by the three plutonic centres of Kathiawar (Auden, 1949). The sea connection towards Umaria would have been along the northern boundary of the outcrops of continental Gondwana sediments in the area to the north of Mohpani (Crookshank, 1936).

Some support for a western source for this transgression may be given by Crookshank's (1936: 240) discovery of two imperfect casts of gastropods in rocks assigned, on lithological grounds, to the Bagra Beds, outcropping north of Budhimai and west of Umaria. Gastropods are unknown elsewhere in the Bagra Beds; and the lithologies in question are also found in the marine Permian Badhaura Formation of western Peninsular India. The only gastropods in the region come from the Early Permian Umaria Beds, 330 km east of Budhimai. On this evidence the gastropods at Budhimai are considered to be possibly of Early Permian age.

The Badhaura area is in western Peninsular India 500 km south of the Salt Range where the boulder beds have boulders and pebbles from this general region (Teichert,
Age and Stratigraphic Relations of Glacial Deposits

1966). This suggests that in Asselian time, the drainage was from the Badhaura area to the Salt Range. This palaeoslope might have been toward the north, approximately parallel to, and west of the Arvalli Mountains. This weaker zone of the crust may have been thrown down tectonically to produce the marine transgressions.

To the south, the Arvalli rocks probably continued as an elevated area during Permian time and so formed the eastern coastline of a sea extending from the Salt Range to the western end of the presumed Narmada rift valley. The coastline of this transgression has been shown in Figure 28.1. The two arms of sea which entered Peninsular India could not have been very narrow because the articulate brachiopods in the faunas could not have tolerated the brackish water environment that would be expected in narrow estuaries.

A GEOMORPHOLOGICAL FEATURE BETWEEN UMARIA AND MANENDRAGARH

As shown above, Peninsular India experienced two short-lived marine transgressions at slightly different periods. These two transgressions for most of their course followed the existing valleys of the Narmada and Son Rivers. This Narmada-Son line is a very significant feature in the tectonic history of India (West, 1962; Hazra and Roy, 1962; Tewari, 1968; Choubey, 1969, 1971; Ahmad, 1972). However, the following observations suggest that the Narmada-Son depression was interrupted near Rewa-Sohagpur.

The two marine transgressions coming from opposite directions in Peninsular India stopped at points about 150 km apart. Since both transgressions had followed valleys in which there were older non-marine deposits, it follows that there were two rivers flowing simultaneously from the area between Umaria and Manendragarh. This area acted as a watershed. The presence of such a highland is also supported by other evidence. The presence of two sets of faults in this central region, one extending eastwards and the other extending westwards (Krishnan, 1960: 331), indicates that this area between Umaria and Manendragarh was a stable one. Further, the drilling by the Geological Survey of India also indicates a basement high of metamorphics in this area (Joshi, pers. comm.). Even in recent times, the Amarkantak plateau in this area forms a watershed for the east flowing Son River and also for the west flowing Narmada River. As mentioned before, Banerjee (1964) has also suggested that there was a stream in this area in Early Permian time flowing from west to east.

This hidden highland, which was probably present somewhere between Umaria and Manendragarh during Early Permian time, is named here the 'Hughes Ridge' in honour of T. W. H. Hughes, of the Geological Survey of India, who conducted an extensive survey of Gondwana sediments of this region.

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REFERENCES


The Palynology of Late Palaeozoic Glacial Deposits of Gondwanaland

ELIZABETH M. KEMP

ABSTRACT

Analysis of the palynology of the late Palaeozoic glacial sequence of all Gondwana continents in terms of an Australian palynological zonal scheme has shown the following patterns: microfloras older than Stage 1 are known to occur with probable glacial sediments only in the Paganzo Basin of western Argentina. Stage 1 microfloras have been positively identified only in Australia, where they are preserved in basins which may have been marginal to the main continental glaciation; they have, however, tentatively been reported from the Paraná Basin of Brazil. Stage 2 assemblages are most widespread, occurring in the lower parts of the examined sequence of Itararé Sub-Group rocks in the Paraná Basin, in the Talchir Formation of the Salt Range and Peninsular India, in diamicites in Victoria Land, Antarctica, and possibly, according to sketchy data, in the basal Dwyka Tillite in South Africa. Stage 3 assemblages characterise much of the Itararé and the Congo Basin sequences. In Australia microfloras of this category generally occur in post-glacial sequences, although evidence of iceberg activity persists into even younger sediments in some Tasman Geosyncline sections.

Preliminary interpretation confirms evidence from other sources in indicating that the oldest glacial deposits lie in west Gondwanaland, but it also suggests that the bulk of the glacial sediments are synchronous within the limits of resolution afforded by palynological means, which may be broad. Palynological evidence lends no support to suggestions that glaciation ceased at a relatively early date in much of South America and Africa.

INTRODUCTION

The idea that the late Palaeozoic glacial deposits of different continental fragments of the Gondwana landmass might not be synchronous probably has its origins in the work of du Toit (1921). The idea was elaborated and further expressed by King (1962: 44) thus: 'The maximal centres of glaciation were, however, not contemporaneous in the several continents, but were sequential from west to east'. King visualised a west to east progression both in the commencement and in the termination of glaciation, such that the westernmost continents of the Gondwana complex, i.e. Africa and South America, were experiencing post-glacial warming at the same time that refrigeration was commencing in the east in Antarctica and Australia. Such a movement of climatic zones was understood to have been brought about either by polar wandering from west to east, or by drift from east to west of the whole landmass across a fixed geographic pole.

The proposal for such a mechanism has drawn support from palaeomagnetic studies, recently summarised by Creer (1970a,
Although the dating of many rock units on which measurement of pole positions are based is extremely broad, a general picture emerges which suggests wander of the south magnetic pole (and the south geographic pole, according to the axial dipole theory), from a broad scatter of positions in North Africa-northern South America in the early Palaeozoic, through a southern Africa pole position in the mid-Palaeozoic, to a Permo-Carboniferous grouping in the vicinity of the Ross Sea (Creer, 1970b: fig. 7).

The suggested movement of the Gondwana landmass relative to palaeolatitude during the late Palaeozoic has been invoked as a mechanism to explain the initiation of glaciation and its progress across the supercontinent, by considering the possible effects on global circulation patterns (Frakes and Crowell, 1970). The same authors have recently suggested details of such a late phase of ice cap migration (Crowell and Frakes, 1970). The evidence for an east-west migration in the commencement, culmination and cessation of glaciation is, however, very meagre. Faunas of possible Early Carboniferous age in glacial marine sequences in Andean Basins of South America have been cited as evidence supporting an early westerly beginning. These faunas, however, have subsequently been referred to the Late Carboniferous (Amos and Rocha-Campos, 1970). An Early Carboniferous age for the base of the Dwyka Tillite in South Africa has also been suggested, on the basis of palaeobotanical, stratigraphic and palaeomagnetic data (Plumstead, 1969; McElhinny and Opdyke, 1968). The timing of the suggested migration of ice centres across the Antarctic continent rests on very few data, and these primarily stratigraphic (Frakes, Matthews and Crowell, 1971). An early westerly cessation, with glaciation essentially over in South America, Africa and India, while still of continental proportions in Antarctica and southern Australia (King, 1962; Crowell and Frakes, 1970) is again based on interpretation of extremely meagre palaeontological data, although persistence of some ice into the Late Permian is suggested by large dropstones in some eastern Australian basins.

Efforts to establish a chronological framework into which late Palaeozoic events might be set are hampered by the ephemeral nature of glacial deposits themselves, and by the restricted biota which could survive under rigorous physical conditions. Marine faunas which are demonstrably contemporaneous with glacial deposits are relatively rare throughout Gondwanaland, and are of a restricted nature, factors which complicate their chronological interpretation. Quite commonly, however, glacial deposits contain plant microfossils, consisting of spores and pollen produced by vegetation which must have existed in periglacial environments, and which expanded during phases of climatic warming and ice retreat. There can be little doubt that such a flora was equally as specialised as any fauna, but it may offer some advantages as a correlative tool in that such remains are easily incorporated into both marine and non-marine sediments. The present study represents a preliminary attempt to see what, if any, light can be shed on the relative ages of glacial deposits from different continents by their contained microfloras.

Australia is to date the only one of the Gondwana continents for which a sequence of microfloral assemblages has been established within late Palaeozoic strata. For eastern Australia, Evans (1969) outlined a series of palynological 'Stages' for the probable Late Carboniferous and Permian, which were based on the observed ranges of suprageneric spore groups, and on the ranges of individual form-species. The units are thus, in a broad sense, assemblage zones. They are as yet very imperfectly defined, but nonetheless offer the most valuable chronologic framework yet proposed on a palynological basis. One advantage offered by the Australian zonal scheme is that it has been erected largely in basins that appear to have been marginal to the main continental glaciation; hence the sequences can be expected to be most complete. Parallel palynological subdivisions have been recognised in Western Australian basins by Balme (1964), and Segroves (1972). The oldest assemblages, designated Stage 1 and Stage 2 by Evans, occur in stratigraphic units which carry a recognisable, though variable glacial imprint in a
number of eastern Australian sedimentary basins. Stage 1 microfloras appear to be associated with plant megafossils which Rigby (1973) refers to the *Gondwanidium* flora, those of Stage 2 coincide with the advent of a *Glossopteris* dominated vegetation. Microfloras of Stage 3 occur, for the most part, in Australia in rock units which suggest that glaciation was waning or completely past, although there is evidence of ice-rafting in some Tasman geosyncline sequences bearing microfloras of this designation.

A further late Palaeozoic subdivision was identified by Helby (1970), who designated as the 'Grandispora' microflora an assemblage lying stratigraphically below Evans Stage 1 unit (referred to by Helby as the *Potonieisporites* microflora), and which occurs in pre-glacial strata in eastern Australia. Subsequently Helby (pers. comm.) has recognised an intermediate assemblage, informally designated the 'Anabaculites' assemblage, which occupies an intermediate position between the *Grandispora* microflora and Stage 1 microfloras, and which has been recognised in both eastern and western Australian basins.

The present study aims to classify palynological data from glacial deposits of the Gondwana continents in terms of the Australian framework. If we accept, for the moment, that the Australian scheme is truly a chronological one, reflecting evolutionary changes in late Palaeozoic floras, then this synthesis should give some guide to chrono-

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logical relationships between glacial deposits in different regions of the Gondwana landmass. The fact that the Gondwana continents make up a single late Palaeozoic floral province enables such comparisons to be made with relative ease. The age of the palynological units in Australia in terms of international time divisions remains a problem. The age of the base of Stage 1 in Australia has not been precisely determined but it may be as old as Westphalian on the basis of its relationships with faunas in underlying sequences.

Figure 29.1 shows the locality of data sources against a background of Gondwana-land as reconstructed by Smith and Hallam (1970). Only those localities are shown for which published data are available concerning the palynology of glacial sediments. This excludes the wealth of data which are to be found only in oil company files. The quality of the palynological data available for such a synthesis varies widely. The published information ranges from fully illustrated microfloral assemblages, as in the Congo Basin, to mere mentions of ages obtained on the basis of palynology, as in the Chaco-Paraná Basin. Additionally, I have examined assemblages from glacigenic sequences in the Paraná Basin, in the Salt Range of Pakistan, and in the Ohio and Wisconsin Ranges of Antarctica. These localities are also shown in Figure 29.1. The problems of comparing microfloras from widely separated areas are compounded by lack of a uniform taxonomy; to facilitate future comparisons the major forms discussed in this study are figured in Plates 29.1 and 29.2.

SOUTH AMERICA

In South America, the best known deposits of late Palaeozoic glacial rocks occur in the Paraná Basin, an intra-cratonic depression occupying large areas of southeastern Brazil, Uruguay and northeastern Argentina. Palaeozoic strata of the Paraná Basin extend westwards into the largely sub-surface Chaco-Paraná Basin (Padula and Mingramm, 1969), and thence even further westward to connect with the Paganzo Basin and the Andean Geosyncline. Glacigene strata are best known from the Paraná Basin proper, where they are exposed along both eastern and western basin margins. Rocks of the eastern outcrop belt are referred to the Itararé Sub-Group; this belt continues southward into Uruguay, where rocks of glacial origin are referred to the San Gregorio Formation. The distribution of outcrop areas and the localities mentioned in the text are shown in Figure 29.2. The distribution, nature and thickness of the Itararé in the Paraná Basin has been reviewed by Frakes and Crowell (1969).

In the Chaco-Paraná Basin, glacial sediments are reported to occur rarely within the Sachoyaj Formation (of Late Mississippian age, according to Padula and Mingramm, 1969), but are abundantly represented within the overlying Charata Formation, which is equated at least in part with the Itararé Sub-Group. In the Paganzo Basin (the Rio Blanco Basin of Frakes and Crowell), the imprint of glaciation is slight. Strata containing diamictites and conglomerates of possible fluvio-glacial origin occur in the lower section of the Paganzo Group, according to Azcuy

Pl. 29.1. Palynomorphs from the Itararé Subgroup, Paraná Basin, Brazil
All magnifications x 500. Localities of figure specimens are indicated.
Fig. 1. Breviradites sp. cf. B. jingurdahiensis/Sinha, Itu. Fig. 2. Microbaculispora tentula Tiwari. Itu. Figs. 3, 4. Lophotrioidites—Horridotrioidites sp. Itu. Fig. 5. Verrucosisporites sp. Itu. Fig. 6. Anapiculatisporites sp. cf. A. spinosus (Kosanke). Itu. Figs. 7, 8. Punctatosporites sp. cf. A. minutus Ibrahim. Sorocaba Rd. Fig. 9. Retusotrioidites diversiformis (Balme & Hennelly). Itu. Fig. 10. Granulatisporites sp. cf. G. quadriflex Segroves. Sorocaba Rd. Fig. 11. Verrucosisporites sp. cf. V. pseudoreticulatus Balme & Hennelly. Sorocaba Rd. Figs. 12, 13, 16. Indotriadites spp. Itu. Fig. 14. Punctatisporites gretensis Balme & Hennelly. Itu. Figs. 15, 17, 20. Dentatispora spp. Itu. Figs. 18, 19. Spongocystia sp. Gramadinho. Fig. 21. Densoisporites sp. Itu. Fig. 22. Vittatina costabilis Wilson. Sorocaba Rd. Fig. 23. Vittatina sp. cf. V. L. Jansoni. Agua Azul. Figs. 24, 25. Vittatina sp. Gramadinho. Figs. 26, 29. Vittatina sp. cf. V. subssacata Samoilovich. Gramadinho. Fig. 27. Taeniocystis sp. Mafra. Fig. 28. Protopalaeoxypinus sp. Gramadinho. Fig. 30. Parasaccites sp. cf. P. mehta (Lele). Itu. Fig. 31. Plicatipollenites sp. cf. P. indicus Lele. Itu. Fig. 32. deutisites sp. Sorocaba Rd.
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and Morelli (1972), who reviewed the sedimentary characteristics of this basin. However, few details are yet available concerning the glacial nature of these deposits.

Palynological investigations have been carried out on late Palaeozoic strata throughout all three sections of the Paraná-Chaco-Paganzo Basin complex, but most are unpublished. In the Paganzo Basin, Menendez (1965, 1969) described microfloras from La Rioja province; these are from non-glacial strata probably equivalent to the lower section of the Paganzo Group. Menendez assigned a probable Westphalian age to the microflora, which is dominated by variously sculptured, acavate, trilete spore types, and is certainly older than Stage 1 of the Australian sequence. Subsequently, Menendez and Azcuy (1969, 1971) reported microfloras from the Lagares Formation, equivalent to the partly glacigene lower section of the Paganzo Group. The spore-bearing formation is overlain by lavas dated radiometrically at 295 m.y., i.e. around the Westphalian-Stephanian boundary (Thompson and Mitchell, 1972). These microfloral assemblages are again older than Stage 1 types, and may be equated with the informal 'Anabaculites Assemblage'; they are probably not as old as the 'Grandispora' assemblage of the Sydney Basin.

A further locality within the general area of the Paganzo Basin is cited by Menendez (1969). This is at Bajo de Velis, where varve-like shales rest on crystalline basement. The shales have yielded a microflora which differs distinctly from those described above, containing monosaccate pollen types and rare disaccate striate forms, suggesting its equivalence with Stage 2 Australian microfloras, although data are at present somewhat sketchy. Within the Chaco-Paraná Basin, no details of palynological assemblages are available to date, although Padula and Mingram (1969) cite palynological evidence for the dating of both the Sachoyaj and Charata Formations.

In the Paraná Basin, published microfloral studies are concentrated in the Uruguayan sector of the eastern outcrop belt. Macchivello (1963) recorded spores from the base of the glacial sequence near Tacuarembó in northeastern Uruguay. Bharadwaj (1969), in a general review of lower Gondwana formations, figured a microfloral assemblage from concretions in the San Gregorio Formation at an unspecified locality; the assemblage is a diverse one, with upward of 20 per cent of disaccate, striate pollen types, a lower percentage of monosaccate types, abundant cavate, trilete spores, and, notably, a diversity and relatively high frequency of pollen of the Vittatina type. Such a composition suggests equivalence with Stage 3 microfloras in the Australian region. Detailed taxonomic treatment of individual form-species from the San Gregorio Formation has been undertaken by Marques-Toigo (1970) and Ybert and Marques-Toigo (1970).

For the present survey, productive samples were examined from seven localities in exposures of Itararé Sub-Group rocks in São Paulo, Paraná and Santa Catarina states of southeastern Brazil. Stratigraphically, the samples range from the Itu Formation at the base of the sequence, through to the Itapetinga Formation at the top, although the relative stratigraphic position of many of the sampled sections is difficult to determine because of low regional dip and lack of detailed mapping in most areas. Lithologically, the samples examined included silstones and diamictites, the latter frequently with sandy, recrystallised matrices. Sample lithologies,
Fig. 29.2. Sketch map of central South America showing approximate margins of Paganzo, Chaco-Paraná and Paraná Basins, glacial outcrop areas and palynological localities mentioned in text (after Frakes and Crowell, 1969; Padula and Mingramm, 1969; Azcuy and Morelli, 1970).

Microfloras from the samples examined fall into two groups, although the number of samples studied to date is too small for this division to be more than tentative. Stratigraphically, the lowest sample examined came from quarries at Itu, where fine sandstones and siltstones, which have been termed varves (see Frakes and Crowell, 1969), rest on basement rocks. The microflora from this locality is dominated by trilete, cavate spore types (referable to the genera Densoisporites, Dentatispora and Indotriradites); monosaccate pollens make up about 5 per cent of the assemblage, and striate disaccate pollens are present in very low frequencies. The presence of rare Striatitii suggests equivalence with Stage 2 of the eastern Australian sequence, or with its Western Australian correlate, the 'Microbaculispora' assemblage of Segroves (1972), which is known from the upper part of the glacial Nangetty Formation in the Perth Basin. Other localities where comparable assemblages occur are the Sorocaba-Itapetininga Road section (see Frakes and Figueiredo, 1967), where it occurs in a massive mudstone at the base of the sequence at 121 km, and from Gramadinho, where the assemblage is found in sediments of the Gra-
madinho Formation.

The remainder of the samples yielded a microflora which shows a considerable increase in diversity from that described above. This second assemblage was recovered from supposed mudflow units at 124 km and 165 km in the Sorocaba Road sequence (see Frakes and Figueiredo, 1967), from pebbly siltstone at Rio do Sul, from diamictites at Mafra, Agua Azul and Jurumirim, and in a poorly preserved state, from diamictites at Ponta Grossa. The microflora from all of these localities is characterised by a content of Striatiti ranging from 5 to 15 per cent; by comparable percentages of monosaccate pollen; and by a diversity of pollen of the Vittatina type. Cavate, trilete spore forms occur in quantities ranging from 12 to 25 per cent. This last group fluctuates widely in abundance throughout late Palaeozoic sediments in Australia, and its occurrence seems to be controlled by local factors. Hence it is not considered to be of great stratigraphic import.

The abundance and diversity of Vittatina type pollens in these glacial sediments is of some interest. Included in the group in Paraná Basin assemblages is a species which bears transverse thickened bands on the distal face, at right-angles to the prominent proximal ribs, and which is closely similar to the species V. costabilis Wilson. Pollen of this type has previously been recorded only from the northern hemisphere; it occurs in Upper and Lower Permian rocks in the U.S.A. (Wilson, 1962; Tschudy and Kosanke, 1966), in Upper Permian rocks of the U.S.S.R. (Samoilovitch, 1961), and has recently been reported from strata of possible Pennsylvanian age in Canada (Barss, 1972). Other forms of stratigraphic interest occurring in the Itararé rocks include forms close to, but probably not conspecific with, the species Verrucosisporites pseudoreticulatus Balme & Hennelly, and Marsupipollenites triradiatus Balme & Hennelly. In Australia, these species are characteristic of Stage 3 microfloras, as these are defined by Evans (1969), although M. triradiatus appears late in Stage 2. The presence of closely related species in the Itararé sections, together with the high relative frequency and diversity of Vittatina species, and of striate bisaccate pollen, is compelling evidence for equating these assemblages with Stage 3.

Preliminary studies, then, suggest that Stage 2 microfloras may be present in the lower parts of the Itararé Sub-Group, and Stage 3 assemblages occur in the higher parts of the sequence. A basic similarity with the Australian sequences is indicated, although the distribution of assemblages remains tentative in view of the poor stratigraphic control. Further, Helby and Runnegar (pers. comm.) have indicated that Stage 1 assemblages might also be present within exposed Paraná Basin sequences.

PAKISTAN AND INDIA

In the Indian subcontinent, Palaeozoic glacial rocks occur in two distinct areas—in elongate depressions on the Precambrian shield of Peninsular India, and lying unconformably on older sedimentary sequences in the Salt Range of Pakistan and in the Himalayas. The Talchir Formation includes basal boulder beds and overlying shale units. In Peninsular India, the Talchir Boulder Beds, which occur at the base of the sequence, are now recognised as being probably hetero-chronous in the different basins in which they are known to occur (Ghosh and Mitra, 1970). In the Salt Range, the basal boulder beds (the ‘Tobra Beds’ of earlier usage) show marked variation in thickness between different areas, but tend to thicken in an easterly direction. Teichert (1967) recognised three facies within the Salt Range Boulder Beds—an eastern tillitic facies grading upwards into marine sandstone, a freshwater silty facies with sparse boulders, characteristic of the central Salt Range, and, in the western Salt Range and Khisor Range, a mixed diamictite and sandstone facies, at least partly of marine origin.

Most of the spore and pollen assemblages described to date from the subcontinent have come from shale sequences overlying the boulder beds. In Peninsular India palynological assemblages have been described from the Talchir ‘Needle Shales’, by Surange and Lele (1956) and Surange (1966), and from siltstones above the boulder beds by Lele (1966). Again, in the Salt Range, the major work of Virkki (1946) was based on the paly-
nology of samples taken from above the boulder beds at Kathwai. However, Lele and Karim (1971) have recently described a diverse spore assemblage from the matrices of boulder beds in the Jayanti Coalfield, and Balme (in Teichert, 1967), briefly reported an assemblage from the boulder bed sequence at Zaluch Nala in the Salt Range.

In the present survey, two further samples from the Zaluch Nala sequence have been examined. One of these samples comes from the 'C member' of the section described by Teichert, the unit which also furnished the samples examined by Balme; the other comes from the basal or 'A unit' of the sequence. Sampled horizons are shown in Figure 29.3. Preservation of spores and pollen from the matrices of diamictites in this sequence is excellent, although only in the higher sample were they abundant enough to allow quantitative estimates of relative frequencies. The 'C unit' assemblage is dominated by trilete spores, with cavate, cingulate forms referable to the genus Dentatispora, and acavate ornamented forms belonging to the genera Hornotriiletes or Lophotriiletes each making up roughly 25 per cent of the total assemblage. Monosaccate pollens make up some 12 to 15 per cent of the total, and striate disaccate forms approximately 2 per cent. This frequency of disaccate Striatiti suggests correlation with Australian Stage 2 microfloras. The presence of acritarchs probably referable to the genus Cymatosphaera in the boulder bed samples may be an indication of at least near-marine depositional conditions, since this form does not appear to have been recorded to date from non-marine strata.

Virkki’s (1946) descriptions of microfloras from beds immediately overlying the boulder beds at Kathwai suggest rather poorly diversified assemblages from the 0.4 m and 1.2 m horizons above the boulder bed datum, although no quantitative data were given. Monosaccate pollens are common, and there is a suggestion of a slightly greater abundance and diversity of Striatiti. Probably the assemblages are Stage 2 equivalents, but there are insufficient data for a definite assessment.

In the peninsula, the microfloras described by Lele and Karim (1971) from Jayanti Coalfield came from boulder beds which the authors claim may have been deposited under fluviatile conditions. Further, on the basis of local stratigraphy, they indicate that the beds may lie well above the base of the Talchir Formation. The microflora from the boulder bed is dominated by monosaccate pollens, and by disaccate forms which show considerable diversity. The content of Striatiti, however, is low, again suggesting correlation with Stage 2. From the South Rewa coal basin, the assemblage described by Potonie and Lele (1961), which represents a still higher horizon in the Talchir Formation, is basically similar to that from the Jayanti area. Both of the peninsular Talchir assemblages differ little from those of Zaluch Nala; they do, however, lack the component of cavate, trilete spore types, which may be of local significance only.
ANTARCTICA

Late Palaeozoic rocks of glacial origin outcrop across a wide region of Antarctica, occurring in the Transantarctic Mountains from Victoria Land to the Pensacola Mountains. They are also known from the Ellsworth Mountains of West Antarctica. The distribution, nature and thickness of these deposits were reviewed by Frakes, Matthews and Crowell (1971). In the West Antarctic basin, comprising the Ellsworth and Pensacola Mountains, the sequences are thick, essentially unbedded diamicrites of the Whiteout Conglomerate and the Gale Mudstone. In the Ohio and Wisconsin Ranges, the glacial deposits are thinner, and are referred to the Buckeye Formation, which is made up of relatively thin diamicite units alternating with shale, sandstone and conglomerate. Striated boulder pavements and grooved sandstone beds occur at several levels within the glacial sequence at these localities, indicating interruption of deposition by periods of active erosion by ice.

Glacigene rocks also occur in southern Victoria Land (the Beardmore Basin of Frakes et al.), where sequences consisting of varying proportions of diamicite, sandstone, siltstone and shale have been referred to the Pagoda, Darwin and Metschel Tillites. Detailed descriptions of tillitic sequences in south Victoria Land and the Darwin Mountains are given by Barrett and Kyle (this volume).

For the most part, the Palaeozoic glacigene rocks of Antarctica are unfossiliferous. Poor preservation characterises microfloras from the entire region of the Transantarctic Mountains, due probably to organic metamorphism resulting from intrusion of Ferrar Dolerites. However, Schopf (in Long, 1965) and Rigby and Schopf (1969) recorded the presence of rare spores in the middle of the Buckeye Formation in the Ohio Range, an assemblage dominated by monosaccate pollens referable to Parasaccites triangularis (Mehta). Of greater significance in the inter-continental correlation of the Antarctic glacials is the assemblage reported by Barrett and Kyle from the upper part of the Darwin Tillite at Colosseum Ridge in the Darwin Mountains. This microfossil suite is the most diverse yet recovered from late Palaeozoic strata in the Transantarctic Mountains, and includes a small proportion of disaccate striatitid pollens, in addition to a relatively high frequency of monosaccate types, both features which distinguished Stage 2 microfloras. The microflora also contains Marsupipollenites triradiatus, which occurs in the younger parts of the Stage 2 interval in Australia.

In the present study, nineteen samples were processed from the Buckeye Formation in the Ohio Range, and seven from outcrops...
in the Wisconsin Range. The position of the samples examined is shown in Figure 29.4. Only four samples in the former section and five in the latter were productive, and in these, preservation was extremely poor and only a limited number of sporomorph species could be identified.

In the Ohio Range, the Buckeye Formation disconformably overlies the marine Early Devonian Horlick Formation, and is in turn overlain by the Discovery Ridge Formation. Palynologically, the most productive sample in the sequence came from a dark grey, recrystallised diamictite 40 m above the base. Monosaccate pollens dominate the assemblage recovered from this sample, mainly types referable to Parasaccites and Plicatipollenites, but with rare specimens of a Potonieisporites species; trilete spores, including Microbacularispora tentula Tiwari and Verrucosisporites sp. are less common. No disaccate Striatiti were observed, in this sample or in any throughout the sequence, but their absence may well be attributed to a preservation factor. The sample described above is notable for the abundance of recycled Early Devonian spores which it contains, deriving from erosion of the Florlick Formation or its equivalents, which contains a relatively diverse microflora (Kemp, 1972). Of particular interest in the diamictite is the presence of acanthomorph acritarchs, suggestive of a marine influence. It is not, however, possible at this stage to establish whether the acritarchs are in place or are recycled from older sediments. They do not, however, occur in the Horlick Formation. Most acritarchs in the Buckeye are referable to the genus Multiplicisphaeridium Staplin, which ranges from the Silurian to the Triassic. Within the Buckeye Formation, recognisable recycling is confined to the lower part of the unit. A further interesting aspect of the Ohio Range microfloras is that they reflect a parent vegetation which apparently flourished in short interglacial periods: the presence of striated pavements above the productive intervals confirms that the floras do not represent a postglacial vegetation.

The present study provides the first record of spores from Buckeye Formation tillites of the Wisconsin Range. All samples from the basal units of diamictite and conglomerate yielded spores, although these were intensely carbonised and identified only with difficulty. Forms identified include an apparent diversity of Parasaccites and Plicatipollenites types, Potonieisporites, Puctatisporites grentensis, Cycadopites, Verrucosisporites and rare non-striate bisaccates. No Striatiti were observed, nor is there any evidence of recycling.

In summary, microfloras from glacial deposits of the Ohio and Wisconsin Ranges are suggestive of Stage 2 of the Australian scheme. This designation is, however, inconclusive because of the poor preservation of the material recovered. The assemblage described from the Darwin Mountains by Barrett and Kyle appears to represent late Stage 2 microfloras. As yet, none of the available palynological data from the Antarctic glacial deposits confirms the suggestion made by Rigby and Schopf (1969) that these might be younger than similar deposits on other continents, a suggestion which was based on the identification of species of Glossopteris and Gangamopteris in units conformally overlying the glacials.

AFRICA

No samples from southern Africa were examined directly during the present investigation. Published information from that continent is meagre, with most coming to date from the Congo Basin, where Bose (1971) recently summarised the results of earlier investigations by Bose and Kar (1966), and Bose and Maheshwari (1966), concerning the palynology of the 'Assises glaciares et periglaciares' of Cahen (1954). The oldest assemblage from this series came from the Elila River area (point 8 on Fig. 29.1). This microflora has some features in common with the Paraná Basin assemblages, notably in its diversity of Vittatina type pollens, which include the northern V. costabilis forms. The frequency of striate saccate pollen types (probably greater than 6 per cent, according to Bose and Kar, 1966: 140) probably equates with lower Stage 3 assemblages in Australia, although most of the trilete spore species which distinguish that unit are missing.
Further south in Africa, Hart (1963, 1967, 1969) has reported the palynological results of examination of Dwyka samples, although only in the case of the Salisbury borehole (Hart, 1963) are location and stratigraphic horizons given. At this locality, in the Orange Free State, striate saccate pollens are present in very low frequencies to the base of the sampled sequence, i.e. to the base of shales immediately above a tillitic unit. The presence of *Marsupipollenites triradiatus*, *V. pseudoreticulatus* and *Granulatisporites trisinus* indicate equivalence with Stage 3 at this latter horizon. The precise relationship of the 'Camerati' florizone, to which Hart assigns most Dwyka sediments, with the Australian microfloral stages is not determinable; both Stages 1 and 2 could possibly be represented.

Recent unpublished studies by R. Falcon from a borehole in the mid-Zambezi Valley, show the presence of Stage 2 microfloras in Dwyka varves and conglomerates in a pre-Karroo valley (R. Falcon, pers. comm.).

### Falkland Islands

During the present investigation some fifteen samples from the Lafonian Diamictite and the overlying Port Sussex Formation of the late Palaeozoic Falkland Island sequence were examined palynologically. Lithologically, the samples included varve-like dark siltstones, diamictites, coaly shales and fine grey sandstones, but all were barren of spores and pollen, although fine woody fragments were common. The reason for the apparently natural loss of any palynomorphs from the sequence is unknown.

### Summary and Conclusions

The distribution of microfloras within the glaciated Gondwana area is summarised in Figure 29.5. Some of the most important features of this distribution are:

1. Only in one locality, that of the Paganzo Basin, are pre-Stage 1 microfloras recorded from the same stratigraphic horizon as glacially derived sediments. Broadly comparable microfloras in Australia are not

![Fig. 29.5. Gondwanaland reassembly showing localities of microfloras discussed](image-url)
associated with any traces of glacial conditions.

2. Stage 1 microfloras are at present known definitely only from Australia, although their possible presence in South America has been mentioned. Within Australia, they are recorded in the main from areas that were peripheral to the main continental glaciation. In at least three basins, relatively thick, essentially non-glacial, older Carboniferous sequences underlie strata bearing the Stage 1 assemblages.

3. Microfloras of Stage 2 are the most widespread in their geographic distribution, occurring on all continents.

4. Stage 3 microfloras have been recovered from tillitic matrices in the Paraná Basin, and comparable microfloras occur in glacially derived sediments in the Congo Basin. In Australia, similar microfloras occur most commonly in strata which were deposited in post-glacial environments (Evans, 1969). The presence of Stage 3 microfloras at several localities within Itararé Sub-Group rocks of the Paraná Basin is in agreement with other fossil evidence, i.e. the presence of *Glossopteris* (Rigby, 1970), and of marine invertebrates (Amos and Rocha-Campos, 1972), which suggests that at least part of the glacial sequence is relatively young.

Interpretation in terms of any model remains ambiguous, probably at least in part because of the sparseness of data points. However, some preliminary conclusions may be cautiously drawn. In terms of a glacial migration model, it might be expected that the microfloras would indicate a greater age for the glacials in the western Gondwanaland sector, i.e. in South America and southern Africa. To date there has been recorded one occurrence of pre-Stage 1 microfloras in this region, which is that in the Paganzo Basin. This remains the oldest 'glacial' microflora yet recorded, although the nature of the glacial sediments with which it appears to be contemporaneous is obscure. It is conceivable that microfloras of the Sachoyaj and Charata Formations of the Chaco-Paraná Basin are similar to these, and if so, they would reinforce the record of older microfloras in west Gondwanaland glacials, but as yet there is no information on their constitution.

Across the bulk of the Gondwanaland re-assembly, most glacial sediments yield microfloras which suggest that, within the limits of palynological resolution at least, they are essentially synchronous. The record to date indicates that tillitic sequences of the Talchirs in India, much of the Itararé in Brazil, the glacial sequences of the Transantarctic Mountains, and possibly much of the Dwyka Tillite yield microfloras which are essentially similar to those occurring in glacial and near-glacial sediments in eastern and western Australian basins. The earliest phase of these microfloras, referred to Stage 1 of the Australian sequence, appears to have been preserved in areas which were essentially peripheral to the main continental glaciation in Australia. There is a suggestion that these occur in similar stratigraphic position in the Paraná Basin, but further work is required to establish the distribution in this basin.

At the other end of the scale, it might be expected that glacial sediments in eastern Gondwanaland would show evidence of their comparative youth by their association with relatively young microfloras. This does not seem to be the case, since it is in the western sector, i.e. in the Paraná Basin, and in the Congo, that glacially derived sediments yield Stage 3 microfloras. In the Paraná Basin, these diverse assemblages have been extracted from the tillites themselves. In Australia, similar assemblages occur predominantly in sediments suggestive of deposition in post-glacial environments, although some glacial impress is present in Bowen Basin sequences of this age and younger. Tasmanian sequences containing dropstones have not been investigated palynologically. There does not, however, seem to be anything in the palynological record to support the contention that glaciation was past in west Gondwana regions at an earlier date than the bulk of the Australian glaciation.

The palynological evidence, however, cannot be entirely used to negate the possibility of some migration of ice centres, for the following reasons. First, there is no certainty that the oldest microfloras have yet been recovered from the glacial sequences. Most of the available data have come from the margins of basins, and it is possible that older microfloral material may yet be recovered.
from deeper basin centres. Second, the possibility that such movement has occurred, but at a rate which is too rapid to be detected by current palaeontological means, must be considered. The presence of probable Westphalian faunas below, and early Sakmarian faunas above the glacial sequences in Australia, brackets those deposits as having been laid down through an interval which may be as long as 40 m.y. At present, we can recognise two subdivisions within that interval on a palynological basis (Stages 1 and 2); we have no means of assessing the relative duration of each, but it could be as long as 30 m.y., which is ample time for quite significant continental motion to have occurred.

One final point that warrants some comment concerns the extent to which the microfloral assemblages defined in this late Palaeozoic interval are influenced by environmental factors. There seems little doubt that such extreme environmental conditions have influenced the composition of the floras to a marked degree—whether or not the distinction between the microfloral Stages 1 and 2 is purely one of a change from a very rigorous to a slightly milder regime is presently not clear. The change is marked by an increase in diversity from Stage 1 to Stage 2, but there are species changes as well, and the microfloral change is marked by a change of some magnitude in the associated plant macrofossil assemblage, from a *Rhacopteris* flora (or a *Gondwanidium* flora in Australia, according to Rigby, 1973) into a *Gangamopteris-Glossopteris* assemblage, so significant evolutionary changes may also be involved.

acknowledgments

I am indebted to Drs L. A. Frakes and J. C. Crowell for provision of the samples on which much of this study is based, and for many helpful discussions. Preliminary work was carried out at Florida State University, with financial support from the U.S. National Science Foundation, Office of Polar Programs. The survey was completed at the Bureau of Mineral Resources, Canberra, and is published with the permission of the Director.

APPENDIX

LOCALITY AND LITHOLOGY OF SAMPLES EXAMINED FROM PARANA BASIN, BRAZIL

Sample no.

205 Laminted light grey fine sandstone and interbedded dark grey siltstone. Quarry at Itu, near top of exposed section.

213 Massive dark grey mudstone. Sorocaba-Itapetininga Rd, 121.5 km, mudstone unit near base of sequence.

202 Greenish-grey massive siltstone. Sorocaba-Itapetininga Rd, 124 km.

212 Light brown sandy diamicite with sparse pebbles. Sorocaba-Itapetininga Rd, 155 km, mudflow near top of sequence.

204 Dark grey, massive, recrystallised mudstone 12 km E of Rio do Sul, within 50 m of base of Itarare section.

207 Dark grey calcareous siltstone with sparse pebbles. Ituporanga-Rio do Sul Rd, at 20 km.

203 Dark grey diamicite, sandy, recrystallised matrix. Highest diamicite in Mafra sequence; quarry 20 km west of Mafra on road to Coninhus.

201 Massive, dark grey recrystallised siltstone with rare small pebbles. Road metal quarry, slightly west of Agua Azul, on road to San Mateus do Sul.

215 Dark grey diamicite, some bedding. Gramadinho-Curitiba Rd, at 191 km.

210 Dark grey diamicite. 1 km SE Railway Stn at Jurumirim.

REFERENCES


30 The Age and Stratigraphic Relationship of the Glacial Sediments in Southern Africa

I. R. McLACHLAN and ANN M. ANDERSON

ABSTRACT

Since the last Gondwana Symposium, new fossil discoveries have shown that a marine incursion of Dwyka age extended well into the Great Karroo Basin in South Africa (McLachlan and Anderson, 1973). Marine fossils associated with this incursion occur sparsely within bedded diamictite and shales very near the top of the glacial deposits, and indicate an Early Permian age. The overlying Upper Dwyka shales do not contain any evidence of glacial activity, and the only animal fossils present, in the White Band, contribute little age information. In the southern Cape Province the Dwyka Tillite rests apparently conformably on the Upper Witteberg shales, considered to be latest Middle Devonian (plants) or Lower Carboniferous (fish).

INTRODUCTION

This article should be read in conjunction with our recently published review (McLachlan and Anderson, 1973), which contains charts, sections and maps. Although marine invertebrates have been known from the Dwyka glacial deposits in South West Africa in the Kalahari Karroo Basin since the beginning of this century (Schroeder, 1908), they were only discovered in the Great Karroo Basin after the last Gondwana Symposium (McLachlan and Anderson, 1973). The contribution of these new faunas to the dating of the glacial beds in South Africa has been limited but has provided significant environmental information. It should be noted that we use the word 'marine' in the biological sense: a depositional environment is not considered to be 'marine' unless there is some supporting fossil evidence, as we feel the word implies at least a connection with the oceanic seas and their faunas.

THE MARINE INCURSION INTO SOUTHERN AFRICA

The Kalahari Karroo and Great Karroo Basins lie parallel and have a similar distribution of sediments. In both cases there are deltaic coal-measure facies in the east and marine shale facies in the west. The coal facies are traditionally regarded as being younger (Ecca series) than the marine facies, which accumulated in the later stages of the Dwyka glacial period. These two basins were linked in the west, and probably formed the eastern part of a continental sea that also extended into the Paraná basin in South America. The connection (or lack of it) is discussed in detail by Martin (1961).

The Kalahari Karroo (including the Warmbad Basin)

In the Kalahari Karroo Basin the outcrop is not continuous from east to west, as the central portion is overlain by Kalahari sands and other, later deposits. Boocock and Van Straten (1962), on lithological grounds, suggest that the coal measures of eastern Botswana possibly correlate with the Auob Sandstone coals in the Dwyka of South West Africa (i.e. they lie stratigraphically below
Age and Stratigraphic Relations of Glacial Deposits
the 'White Band'). Heath (1966, 1972) follows this interpretation. Green (1969), argues on the evidence of additional boreholes that the sandy Nossob and Auob units extend eastwards into the basin beneath the Middle Ecca sands.

The marine beds outcrop on the western side of the basin, in semi-desert country. The exposures are, therefore, often good, being unobscured by soil or vegetation. The stratigraphy has not yet been fully elucidated. Our attempted synthesis is based mainly on information from Heath (1966, 1972), Martin (1953, 1961, 1968), Martin and Wilczewski (1970) and Frakes and Crowell (1970).

No fossils have yet been recorded from the four lowest glacial units in Heath's (1966) stratigraphic column. The sediments appear to have been laid down as ground moraine and partly redistributed by the slumping of unstable dumped material, some of it, at least, subaqueously. The overlying fossil-bearing beds also show signs of glacial activity. They are interbedded with unsorted diamictite with striated clasts, and often themselves contain dropstones. The areal and vertical distributions of the fossils are shown in McLachlan and Anderson (1973: figs. 1 and 3). As many were discovered near the turn of the century, information relating to the precise location of the finds is often lacking. The faunas are very sparse.

According to Martin, Walliser and Wilczewski (1970) the fossils occur in two different situations. In the shale and boulder-mudstone units they are present within slightly phosphatic calcareous concretions that are sometimes silicified. The fossils, which include both pelagic and benthonic forms, are generally well-preserved in an uncrushed state. Within the shale beds there are sandy and pebbly units, interpreted by Heath (1966) as turbidite deposits. Fossils have been found in sandy channels presumed to have been scoured by these turbidity currents. The contained fossils (Eurydesma, bryozoans and asteriodeans) appear to have been transported.

The known outcrop of the fossiliferous beds has been extended by our recent discovery of calcareous fossiliferous nodules in outcrops 32 km northwest of Upington in Gordonia. Fish and coprolites are present, and it is likely that the nodules will yield a microfauna.

The glaciomarine beds are disconformably overlain (according to Heath, 1966, 1972) by the Nossob Sandstone. The sediments above the disconformity show no signs of marine conditions, nor of any glacial activity.

The Great Karroo Basin

Information on the geology of the glacial beds in the Dwyka has been drawn mainly from Du Toit (1921), Stratten (1968), and Crowell and Frakes (1972), as well as from our own observations.

The lack of marine invertebrates prompted some previous authors to regard the Lower Karroo as non-marine, whereas others, on sedimentological grounds, postulated marine conditions. Recent fossil discoveries leave no doubt that there was a link between the Great Karroo Basin and the sea during the later stages of the Dwyka glacial episode.

The stratigraphy of the Dwyka sediments in the Great Karroo Basin is generally the same as in South West Africa. The massive, lower parts of the diamictite are barren of animal fossils, with the exception of arthropod and other trails, which occur on interbedded flagstones in the western Cape Province and Natal, and also in the Warmbad Basin. The upper portions of the Dwyka Tillite Stage are usually bedded and appear to have been laid down subaqueously. They contain thin cone-in-cone limestone bands, and locally show slumping, graded bedding and other massflow features. Arenaceous foraminifera, sponge spicules and radiolaria, as enclosed in calcareous nodules within interbedded flagstones at De Kalk, and diamictite at Riet River Bridge B, indicate that the body of water in which the glacial material was deposited was marine. The stratigraphically highest invertebrate faunas recorded (those at Blaauw Krantz) lie some distance (± 9 m) above the nearest rafted-pebble beds which mark the last signs of glacial activity. The fauna at Tankwa River seems to be from the same horizon, but there it lies.

Fig. 30.1. Evidence for a marine incursion into southern Africa during Dwyka times.
Age and Stratigraphic Relations of Glacial Deposits

Woods (1952) suggested, on the basis of White Band occurrences, that the Upper Dyke shales are of Carboniferous age.

Johnston (1951) reassessed the material, and concluded that they belong to the same family as the Carboniferous Dykes. It is not sufficiently exact to support the view that the Dykes are Carboniferous.


development of glacial deposits

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<th>Invertebrates</th>
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<th>Miospores</th>
<th>Megaplanthus</th>
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<td>Mesoiceras (lithic) - close relationship to forms from Gondwana, South America - possibly early Ammonite. Antitriceras (lithic) - similar forms occur in Lower and Upper Permian of E.E. Bolivia and W. Australia.</td>
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<td>Peltopleurites (lithic) - resembles the older (South American) species of Peltopleurites more closely than the younger Antitriceras in the South American Province, South America - possibly early Ammonite. Antitriceras (lithic) - similar forms occur in Lower and Upper Permian of E.E. Bolivia and W. Australia.</td>
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Fig. 30.2. Summary of age information.
about 7 m below the highest dropstones. The last appearance of dropstones is unlikely to be stratigraphically constant. The marine faunas seem to mark the fullest extent of the marine transgression connected with the terminal stages of the ice retreat, and may therefore be approximately contemporaneous. The overlying Dwyka shale contains no fossil evidence of marine conditions. Isostatic rebound or some other earth movement seems to have severed the connection with the oceans shortly after the withdrawal of the glaciers.

Fossiliferous concretion deposits in similar stratigraphic settings have been found at a number of other points around the Karroo Basin, indicating that the marine incursion reached at least as far as a line joining Pietermaritzburg and Kimberley (see Fig. 30.1). Some generalisations about the faunas are possible. Only microfaunas have been found in the bedded diamictite sections—arenaceous foraminifera, radiolarians and sponge spicules at Riet River Bridge B, De Kalk and at Ashburton near Pietermaritzburg. The highest fauna, as typified by the Blaauw Krantz deposit (but also found at Tankwa River), is dominated by palaeoniscoid fish and coprolites. Many of the coprolites are 'enterosporae', interpreted by Williams (1972) as the fossilised intestinal valves of sharks (Chondrichthians). Fossil wood is also common at these levels, suggesting well established vegetation, including large trees. The invertebrates are scarce, there being orthoceroid nautiloids (10 specimens), pelecypods (5 specimens, 2 species), and brachiopods (3 specimens, 1 species).

The occurrence of moulds of crystals of the mineral glauberite within concretions at the Blaauw Krantz deposit is of interest. Glendonites, which are calcitic pseudomorphs after glauberite, have also been described from much the same setting in Permo-Carboniferous sediments in New South Wales (David et al., 1905), Queensland and Tasmania (Raggatt, 1937).

AGE OF THE GLACIAL BEDS

The problems of dating the full history of a glacial period are discussed by Crowell and Frakes (1972: 2891). They suggest that the most complete record of glaciation is present only near the depositional centres, and that a progressively younger veneer of tillite is found up the depositional slopes, towards the ice centres.

The limited fossil information available on the age of the glacial beds in South West Africa and the Great Karroo Basin is summarised in Figure 30.2. There is insufficient evidence available at present to indicate any difference in age between the deposits in the two areas.

Pre-glacial Beds

In the southern Cape Province outcrops the Dwyka Tillite appears to lie conformably on the underlying Upper Witteberg shales (Loock, 1967; Rossouw, 1970; Theron and Blignault, this volume), but further north it is unconformable across older sediments (Winter and Venter, 1970). Opinions are divided on how much of a time break there is. Savage (1972) and Theron and Blignault (this volume) describe what appears to be a hydroplastic response of the Witteberg shales to grooving by the Dwyka ice during its first advance. This seems to indicate that the underlying Upper Witteberg shales were still unconsolidated at the time of deposition of the first tillites and that the time break is small.

From the Upper Witteberg Gardiner (1969) described eleven species of fish from about 200 m below the top contact as being of Early Carboniferous (Viséan) age on the basis of comparison with related Palaeonisciformes in Scotland. Plumstead (1967: Table 3) examined collections of lycopods from similar stratigraphic position as the fish, and tentatively derived a latest Middle Devonian age, which depends largely on Protolepidodendron eximium and its close relationship to Argentinian specimens.

Glacial Beds

The only age information at present available from the lower part of the tillite is weak. Poor impressions of Noeggerathioptis or Psygomo phyllium were found crushed under a quartzite boulder in the tillite near Matjiesfontein (Du Toit, 1930). Since these plants are associated with the 'mixed' Glossopteris
assemblage further north, Plumstead (1964) considered them to be of Late Carboniferous age.

Marine invertebrates provide what is at present the most reliable age for the upper part of the glacial beds in the southern Great Karroo and Kalahari Basins. Only six forms have been described from the sparse South West African faunas, and these indicate an Early Permian (Dickins, 1961) or a Late Carboniferous to Early Permian age (Martin et al., 1970). *Attenuatella* and *Phestia* from the Great Karroo Basin have been interpreted as indicating an Early Permian age (McLachlan and Anderson, 1973).

The palaeoniscoid fishes described by Gardiner (1962) from the Ganikobis Shale in South West Africa do not supply any age information.

**Post-glacial Beds**

In the southern Karroo Basin and in southern South West Africa the glacial beds are overlain by shales of the Upper Dwyka Stage which bear no signs of glacial activity. Leschik (1959) described a microflora from South West Africa from the Nossob Sandstone, just above the highest glacials, as being Early Permian. The reptile *Mesosaurus* and benthonic crustaceans found in the White Band at the top of the shales do not provide precise age information (see Von Huene, 1940; Broom, 1931; Fabre, 1967).

The northern part of the basin poses a separate problem. Unambiguous glacial sediments, including diamictites interbedded with dropstone-bearing varves, are preserved in deep valleys scoured in the basement rocks. However, these beds grade upwards (see Fig. 30.2), without any sharp change, into water-worked conglomerates with rounded pebbles, carbonaceous sandstones and shales, and thin coal bands. Similar conglomerates, often bleached, are encountered frequently over much of the area covered by the Middle Ecca Coal Measures. Thin coal bands sometimes lie below the pebble bands. These conglomerates have been interpreted as glacial by some authors (e.g. Plumstead, 1964, 1969) and as fluvial conglomerates resulting from the reworking of glacial material during the later coal measures cycle by others (e.g. Du Toit, 1921: 193, 217; Nel and Jansen, 1957). We support the latter view. Plumstead (1964, 1969) described lycopods from carbonaceous shales directly above diamictite near Welkom in the Orange Free State. She concluded that the sediments are of Late Devonian age and represent a northern extension of the Witteberg Series deposited after a very early onset of glaciation from the north. Hart (1963) studied the miospores from this lycopod zone in a borehole further north near Vierfontein. He concluded that they indicate a Late Carboniferous age (see Fig. 30.2). We consider these sediments to belong to the Ecca. Anderson (1973: chart 35), largely on the evidence of miospores, considered the whole Dwyka section to be of Early Permian age. We find it difficult to reconcile these widely differing age conclusions. On the basis of our own experience of the Dwyka geology of the Karroo Basin, and following the reasoning outlined in Crowell and Frakes (1972: 289), we would expect a younger age for the glacials in the northern part of the basin.

**DISCUSSION**

It is apparent that information relating to the age of the glacial beds is sparse, and studies of different fossil groups often produce conflicting datings. Judging from papers presented at the present Gondwana Symposium there is also considerable difficulty and difference of opinion in the interpretation of the boundaries of the international time units. For the relative dating of the Karroo sediments, palynology is likely to be the most useful tool, in spite of problems of miospore carbonisation. Miospores are certainly more abundant and widespread in the Karroo Dwyka sediments than any other group of fossils. Palynology can, therefore, be used as a control against which other fossil information can be assessed, as indicated by Anderson (1973). Before the age of the Karroo glacials can be satisfactorily compared with those of the other Gondwanaland areas, much more basic geological and palaeontological work must be done.

**ACKNOWLEDGMENTS**

We would like to thank Dr J. M. Dickins for his identifications of the Great Karroo
REFERENCES


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Section 5

Advances in Stratigraphy and Palaeontology
ABSTRACT

Provincial differences between the marine invertebrate faunas of the cool-temperate Gondwana successions and the warm-temperate or tropical Russian and North American successions, make it difficult to use the classical Permian time divisions in the Gondwana province. A new regional scale is needed. Since the new scheme should be characteristic of the province, it should be based on the most complete cool-temperate sequences, preferably those from eastern Australia.

A number of consecutive invertebrate assemblages have been recognised in the New Zealand and eastern Australian Permian, but a search for sequences displaying the boundaries between stratigraphically adjacent assemblages reveals that the constituent taxa do not usually appear or disappear simultaneously. Consequently, the stratigraphic boundaries of the assemblages cannot be precisely defined. It is suggested that an alternative scheme based on reference points (boundary stratotypes) located in sections containing evolving lineages of spiriferoid or productoid brachiopods would provide stable and recognisable time divisions. These should be named for some geographic feature coupled with an age term commensurate with their rank. It is recommended that the formalisation of such a set of time divisions for the Permian Period in the Gondwana province be supervised by a small international committee.

INTRODUCTION

Most internationally accepted divisions of Permian time are based on Russian and North American sequences of invertebrate fossils. The diversity of the marine faunas and the abundance of carbonates and evaporites in these sequences suggest that they were formed in the tropical to warm-temperate parts of the Permian Earth. Palaeomagnetic data and modern global tectonic theory support this interpretation.

Similar evidence indicates that the seas surrounding Australia, New Zealand, and parts of India, Pakistan, South America, and southern Africa were cool to cold-temperate throughout most of the Permian Period. For this reason the provincial differences between the Gondwana marine Permian faunas and those of the standard sequences make it difficult to recognise the North American and Russian time divisions in Australia and other parts of Gondwanaland. Since a precise and detailed time scale is a basic necessity for many different aspects of geology, some attempts have been made in Australia to devise scales based on the local sequences of invertebrates (Hill, 1952; Dickins, 1963; Dickins et al., 1964; Runnegar, 1969; Dear, 1972a; Clarke and Banks, this volume). Waterhouse (1967, 1969) has produced analogous though more formal subdivisions of the Permian sequence in New Zealand.

Sufficient information is now available from the most complete fossiliferous marine Permian sequences of the Gondwana province for us to ask several important questions. First, is
there a need for a separate set of Permian time divisions for the Gondwana faunal province? If so, how should these time divisions be defined and named? Is priority of definition and nomenclature important? Where are the best sequences of fossils characteristic of the province? And finally, will such time divisions be accepted and used by the national and international community?

IS A NEW REGIONAL TIME SCALE NEEDED?

We have already noted the difficulty of using terms like Sakmarian, Kungurian, Wolfcampian, Guadalupian, and Djulfian in Australia, and the difficulties are similar in other Gondwana countries. The meagre ammonoid faunas of western and eastern Australia (Glenister and Furnish, 1961; Armstrong et al., 1967) allow approximate correlations with northern hemisphere sequences, but it is difficult if not impossible to locate the boundaries of the classical Permian stages. Thus it is possible to argue that some of the Australian assemblages are Artinskian in age, but there are not enough fossil ammonoids to allow us to recognise where the base of the Artinskian occurs. Since the Artinskian probably represents many millions of years of time, the use of such stages is not good enough for local purposes. Moreover there is no reason for thinking that the base of the Artinskian, recognised primarily by events in the history of warm water nektic ammonoids or benthonic foraminifera, should be reflected in the history of the cool-temperate benthos of Australia.

Waterhouse (1967, 1970a; Bamber and Waterhouse, 1971) argued differently. He viewed the Permian as a time of alternating cold and warm periods reflecting the advance and retreat of polar ice. In his scheme the stage and zone boundaries are short transitions between cool and warm periods which can be correlated throughout the world by their distinctive brachiopod assemblages.

Our experience suggests a need for greater caution. In 1969 we began to look for the boundaries of the informal assemblage zones then used to zone the eastern Australian Permian. We searched for sequences of uniform lithology (hopefully reflecting a uniform depositional environment) where we could examine the changes that occurred as one assemblage zone succeeded another. In all cases where such sequences were available we found no evidence for a catastrophic or even accelerated revolution of the invertebrate assemblages. Instead we found slow evolutionary changes in the species of most genera. As could be expected, we extended the ranges of many species, and restricted others by narrowing the definition of the species. The only abrupt changes we observed occur where the lithology changes suddenly or where there is evidence of a hiatus. It follows that if our understanding of the situation is correct, there are no important, widespread natural events which will conveniently divide Permian time.

This is not to say that the evolution of species and communities progressed at a uniform rate. Nor does it imply that there were no continent-wide or larger climatic or tectonic events which may have adversely affected the shelf benthos. It means that these varying evolutionary rates and climatic or tectonic events are not yet obvious in the Australian Permian fossil record, whereas the continuity of life is. We are therefore faced with the problem of dividing a continuum, rather than identifying a series of international events.

The continuity of invertebrate lineages is not as well displayed in the Western Australian sequence (Dickins, 1963), but they have not been examined in the same way. An apparently similar situation existed in eastern Australia before the studies of the last decade, and it is reasonable to suppose that greater continuity will be found in Western Australia when the same technique of study is adopted.

In New Zealand, Waterhouse (1967, 1969) described a series of taxonomically distinct assemblages which he formalised as 'stages'. But apart from a few exceptions, these are small faunules, in many cases separated by great thicknesses of unfossiliferous sediment (Runnegar and Armstrong, 1969). Moreover we do not agree that stratigraphically adjacent assemblages are as different as Waterhouse suggests. Viewed in the context of the eastern Australian successions, the New Zealand faunas represent significant data points on a long continuum.
Fig. 31.1. Location of sections shown in Figure 31.3. Horizontally shaded areas contain both marine and non-marine Permian sediments; in the stippled areas the Permian sediments are dominantly non-marine. Faults separate western zone of relatively undeformed sediments from eastern zone of moderate to intense tectonic deformation. Localities 3, 4, 7, 8, 10 and 13 contain sequences in which reference points could be fixed to formalise a Permian time scale.
We conclude that while it is possible to correlate the Gondwana sequences with the classical Permian stages and zones approximately, the boundaries of these stages and zones are not reflected by natural events in the Gondwana province. A locally derived scheme would have immediate practical application and greater precision within the Gondwana area.

WHERE SHOULD THE TIME DIVISIONS BE DEFINED?

The boundaries between cool and warm-temperate Permian faunas will not be abrupt and there will be some areas where Gondwana and Tethyan elements occur together. Such areas include parts of Western Australia and the Salt Range in Pakistan. The best place to define provincial time divisions would seem to be near the centre of the province, or at least as close to the centre as well preserved sequences are known. The sequences which formed in transitional areas can then be used for inter-province correlation. The same philosophy is used in defining interdigitating rock units.

If we accept this approach, the eastern Australian sequences provide the best available standard. Within eastern Australia there are four areas suitable for closer study: the northern part of the Bowen Basin, north Queensland (Runnegar et al., in prep.), the Hunter River Valley and southern coastal sections of the Sydney Basin, New South Wales (McClung, in press; Runnegar, in press a), and the island of Tasmania (Clarke and Banks, this volume) (Figs. 31.1 and 31.3). Rich Permian faunas occur in many other areas (Runnegar, 1969), but the sequences are less complete, contain many non-marine intervals, or are tectonically deformed. They yield important supplementary data, but for the framework of the time scale we should concentrate on the four areas mentioned above, where both evolution and superposition are clearly visible.

HOW SHOULD THE TIME DIVISIONS BE RECOGNISED?

Shaw (1964) notes that 'each objectively definable fossil taxon divides geologic time into three segments—the time before it appeared, the time during which it existed, and the time since its disappearance'. He calls the total range of a species, from its first to last individual, the biozone of that species, and concludes that this is a quantity that can never be measured because of the impossibility of finding the first and last individuals. The best geologists can do is to estimate the biozone by measuring the total stratigraphic range of the species against other species of the assemblage. This is the 'range zone' of the American Code of Stratigraphic Nomenclature. We can assume that it is normally less than the biozone.

The best measure of total stratigraphic range, one that we feel most closely approximates the biozone, is one measured in an evolving lineage. For if in any stratigraphic section we can collect ancestral species below and descendant species above the species we are concerned with, we can measure the maximum limits of its biozone as well as its stratigraphic range. A range zone mapped in this way has greater temporal significance than another identified by a cryptogene, a taxon without obvious ancestors.

Another method of correlating sequences of fossiliferous rocks is to use an association of species or other taxa—the 'concurrent range zone' of the American Code. Such zones provide important first approximations, for every student of biostratigraphy is aware that successive communities of animals and plants have populated the earth for longer or shorter periods of time. But in the sublittoral marine environment where the community structure may be loosely defined in terms of the interdependence of organisms (Speden, 1966), it is unreasonable to suppose that constituent species of a community will have evolved at the same or even similar rates. Since a concurrent range zone, by definition, is composed of several or many range zones, its boundaries will be imprecise if the constituent taxa have slightly or markedly different ranges (Shaw, 1964). This will normally be true in sequences where the environment has remained relatively constant for long periods of time. Since these are the sequences that yield the best biostratigraphic data, concurrent range zones will be difficult to use in a precise sense. We are therefore
faced with the dilemma; range zones have precise boundaries that are difficult to identify because the organism may become rarer near the limits of its range; concurrent range zones have boundaries that are easier to identify approximately, but precision is impossible because the ranges of constituent taxa are not the same.

Again it is the evolving fossil lineage that provides the best data, since if one species is slowly evolving into the next, the populations may remain relatively large and our estimate of the stratigraphic range of the species will be more accurate. Ideally we should use the continuous evolution of the most common fossils for our biostratigraphic zonation.

Fig. 31.2. Frequency distributions of the number of growth ridges in successive populations of an inoceramid bivalve from a 7m thick section, Cretaceous Western Interior Basin, U.S.A. (Data from Kauffman, 1970: fig. 4.) A reference marker (arrow) could be used to define the difference between species A and B if the taxonomy had not been previously formalised. It would then identify the beginning of a time division which could be recognised elsewhere by the first appearance of species B.
HOW SHOULD THE TIME DIVISIONS BE DEFINED?

According to the new International Guide to Stratigraphic Classification, Terminology, and Usage (summarised in Hedberg, 1972), zones are biostratigraphic units. They are not automatically measures of time. In the view of the authors of this guide, time intervals should be defined by markers placed at appropriate places in well exposed fossiliferous or other sequences and then correlated with other places by biostratigraphic units which are thought to be approximately or very nearly coeval. The markers are called boundary-stratotypes (we prefer reference points), and they are correlated by assemblage-zones, taxon-range-zones, concurrent-range-zones, Oppel-zones, lineage-zones, or other subsidiary methods. Little attempt is made to evaluate the relative advantages of the different methods of correlation, but for the reasons discussed above it is clear that zones based on consecutive evolving species (lineage-zones) are best.

The advantage of this approach to the subdivision of time for geologic purposes is that the marker, the reference point, provides an objective standard like the type section of a formation or the type specimen of a species. It locates a source of additional data which can be assessed by workers throughout the world. The problem of recognising the junction between two periods of time is then resolved to a matter of correlation with the section containing the reference point.

It is of course desirable that at least some of the possible correlations with the section containing the reference point be visible before the reference point is chosen. Since fossils provide the only commonly available method of dating sedimentary rocks, it is logical to locate the reference point at the boundary of one or more fossil zones. In the kinds of sequences we have been discussing, where different species are evolving at different times and at different rates, the reference point is best located at an arbitrary boundary between two consecutive taxa. If the evolution is really slow and gradational, and if the taxonomy has not previously been formalised, it may even be convenient to use the reference point to define the boundary between stratigraphically adjacent taxa (Fig. 31.2). Ideally then the reference marker should be located within a fossiliferous bed containing a population of the taxon whose ancestors and descendants occur in under- and overlying beds of the sequence. This of course is not always practical, but inadequate sections should not divert us from our concept of the time divisions and their use.

Correlation with the section containing the reference point is accomplished by all means available. If the evolving lineage used to locate the reference point occurs elsewhere, it is reasonable to assume that the evolutionary changes were approximately contemporaneous. Other evolving lineages, and the appearance and disappearance of cryptogenetic species, will provide additional evidence, as could radiometric dates, palaeomagnetic reversals or other criteria. If the reference point is well chosen there should be no immediate or long-term problems of correlation with it.

HOW SHOULD THE TIME DIVISIONS BE NAMED?

The International Guide distinguishes between chronostratigraphic units which are the rocks deposited during a specific interval of 'geologic' time, and geochronologic units which are the corresponding intervals of 'geologic' time. For example, the Permian chronostratigraphic system was deposited during the Permian geochronologic period. To us this seems a valueless distinction. Since the rocks of the Permian System cannot be identified unless it is known they were formed in the Permian Period, it is impossible to recognise chronostratigraphic units without recourse to correlation. But according to the International Guide, the boundary stratotypes (reference points) define not the time periods (geochronologic units), but the extensive bodies of rock formed within the limits of time enclosed by adjacent boundary stratotypes (the chronostratigraphic units).

Our view of the procedure is different. Each reference point separates two segments of time and the name applied to each of these segments identifies each period of time. There is no need for a separate set of 'chronostratigraphic' units as the expression 'Permian granites' is easily understood to
mean either 'granitic rocks formed during the Permian Period' or 'granites of Permian age'.

Obviously geologists wish to distinguish longer and shorter periods of time, so a hierarchy of terms is needed. The terms 'Era, Period, Epoch, and Age' adequately describe the larger categories of time, and there is no need for the equivalent chronostratigraphic terms 'Erathem, System, Series, and Stage'. Thus the Artinskian Age should eventually be known as the period of time identified by two reference points in some Russian succession, and we can speak of Artinskian sandstones in the same way we referred to Permian granites.

For finer subdivisions the nomenclature becomes more complex as many geologists see no clear distinction between a fossil zone and its temporal significance. Again we should probably resort to reference points, though it is certain that some geologists will prefer to use unformalised fossil zones as local indicators of contemporaneity.

In summary, any time divisions we recognise should be defined by reference points in fossiliferous section and named after some geographic feature coupled with an age term commensurate with their rank. The smallest divisions of time need not be formalised and could be named in the same way as the fossil zones used to identify them.

THE PROBLEM OF FACIES

We have been speaking about cool-temperate sublittoral assemblages of shelf invertebrates. But the shelf is not homogeneous since it comprises a number of different environments each with its own characteristic community. Some of these communities will intergrade, but others will contain few or no species in common. We are not yet able to describe many of these communities in the eastern Australian Permian, but it is possible to distinguish mollusc-dominated sublittoral sand assemblages from brachiopod-dominated assemblages in the more off-shore silts (McClung, in press; Runnegar, in press a; Runnegar et al., in prep.). Since the fossils found in rocks formed in these two environments rarely occur together, it is impractical to have the primary time scale related to fossils from both environments.

Experience has shown that the most valuable biostratigraphic indices are the spiriferoid and productoid brachiopods, as these are common, and evolve in relatively constant and predictable ways (Plate 31.1). We therefore construct our primary time scale for the eastern Australian Permian from the brachiopod zones found in the low energy silts, and then relate it to a less well displayed zonation of the molluscs in the higher energy sands (Fig. 31.3). Both sets of zones can be named and have some temporal significance, but we suggest using the brachiopod sequence where possible to locate the reference points for the time scale. The age of the mollusc zones can then be estimated by interpolation or interdigitation.

The same procedure can be used to date palynological zones in both marine and non-marine sediments. Eventually we hope to recognise zones based on bryozoans, trace fossils, and possibly corals and crinoids. Each will have some temporal and some ecologic significance, but again we suggest that their age should be determined with respect to the brachiopod-based reference points.

BRACHIOPOD ZONES AND THE LOCATION OF REFERENCE POINTS

All recent biostratigraphic subdivisions of the eastern Australian marine Permian recognise a number of consecutive assemblages of brachiopods, molluscs, or other organisms (Hill, 1952; Dickins et al., 1964; Runnegar, 1967, 1969; Dickins et al., 1969; Clarke, 1969a, 1969b; Waterhouse, 1972; Dear, 1972a; Clarke and Banks, this volume). The initial study for the existing zonal scheme was that of Dickins in the Bowen Basin of Queensland, where he and his co-authors recognised four Permian assemblages which they informally named 'Faunas I, II, III, and IV'. They subdivided Fauna III into units IIIA, IIIB, and IIIC, and divided Fauna IV into three unnamed parts (Fig. 31.4).

Each of the larger and smaller units is an assemblage zone in the sense of the American Code.

In 1967 and 1969 Runnegar extended Dickins's informal assemblage zones to other parts of eastern Australia, added a new zone—the 'Allandale fauna'—at the base, and...
suggested that Fauna I was a local variant of Fauna II. He also found it difficult to recognise Fauna III outside the Bowen Basin and coined the term ‘Ulladulla fauna’ for an assemblage of approximately the same age in the Sydney Basin, New South Wales (Fig. 31.4). Clarke (1969a) subsequently recorded the ‘Allandale’ and ‘Ulladulla’ faunas from Tasmania, and Dear (1972a) showed that Fauna II in the Bowen Basin resembled those found in Fauna IIIA. He also introduced a number of informal terms for locally restricted brachiopod assemblages in the Bowen Basin. Finally, Waterhouse (1972) recognised six or possibly seven of the New Zealand faunal ‘stages’ in the marine Permian of eastern Australia. These partly correspond with the assemblage zones recognised by Australian authors.

Before 1970 few of the sequences outside the Bowen Basin had been collected in any detail. During 1970-72 Clarke assembled large collections from many sections in Tasmania (Clarke and Banks, this volume), and we have collected from the most important sequences in the Sydney Basin (McClung, in press; Runnegar, in press a). We have also used large collections from the Bowen Basin assembled by Dorothy Hill and her students at the University of Queensland, by J. M. Dickins at the Bureau of Mineral Resources, Canberra, and by J. F. Dear at the Geological Survey of Queensland. These were supplemented by additional collections from the northern Bowen Basin made in 1971 and 1972 (Runnegar et al., in prep.).

During this collecting program it became obvious to Clarke and us that the characteristic taxa of the recognised assemblage zones often occur in younger and older zones. As the work progressed it became increasingly difficult to maintain the old nomenclature without modifying the original concept substantially. Eventually we agreed that the evolution of the spiriferoid brachiopod Martiniopsis (also called Tomiopsis, Ambikella, or Ingelarella) and the productoid brachiopod Wyndhamia (also called Strophalosia, Branxtonia, Multispinula, or Echinalosia) provided a more precise method of subdividing the Permian faunal succession. Our zonal scheme is based primarily on species of these two genera.

**Pl. 31.1. Internal moulds of species of Martiniopsis Waagen. Arrows indicate evolutionary pathways; asterisks mark brachial valves of species used to define lineage zones. Figures 7-9 demonstrate variation within a single species; specimens resembling 9 constitute less than one per cent of the population. 1-2, M. elongata McClung, Herbert, and Helby; University of New England F13017. 3, M. koninkii Etheridge; UNEF12243. 4-5, M. branxtonensis Etheridge; 4, UNEF13117, 5, UNEF13113. 6-9, 24, M. ovata (Campbell); 6, UNEF13115, 7, UNEF13111, 8, UNEF1811, 9, 24, UNEF1813. 10-11, M. plana (Campbell); UNEF11802. 12-13, M. ingelarense (Campbell); UNEF12188. 14-16, M. mantuansensis (Campbell); 14, UNEF11788, 15, UNEF11789, 16, UNEF11862. 17-18, M. profunda (Campbell); 17, UNEF11812, 18, UNEF11810. 19-20, M. profunda alta Runnegar, McClung, and Javes; UNEF13123. 21-22, M. plicata (Campbell); UNEF11873. 23, M. angulata (Campbell); UNEF13119. 25, M. plana-M. brevis transition; UNEF1804, 26-27, M. brevis McClung, Armstrong, and Runnegar; 26, UNEF12219, 27, UNEF12220. 28-29, M. undulosa (Campbell); 28, UNEF5836, 29, UNEF5837. 30-31, M. isbelli (Campbell); 30, University of Queensland F16217, 31, UNEF11805.

**The Evolution of Martiniopsis**

Martiniopsis is a large smooth or coarsely plicate spiriferoid with an external microornament of shallow elongate grooves (Campbell, 1959, 1960, 1961; Armstrong, 1970; McClung et al., in prep.). Campbell (1961) observed that the dorsal and ventral edges of internal plates supporting the sides of the delthyrium and notothyrium generally become progressively longer in successively younger species. This occurs in both smooth and plicate species-groups. The inner parts of these plates are called dental and crural lamellae; the outer parts are called adrninicala (brachial or pedicle), and they are best observed on natural internal moulds of the shell (Plate 31.1). In general the brachial adminicala are more significant, and they can be of two types, straight and divergent (Pl. 31.1, figs. 7-14), or flexed (Pl. 31.1, figs. 29-31).

It is possible to resolve most of the morphology of the species of Martiniopsis into four variables: the shape of the commissure (which
generates the plicae, sulcal furrows or ribs etc.), the shape of the brachial admicula (straight or flexed), the ratio of the length of the brachial admicula to the length of the valve, and the degree of shell thickening.

Using these generalised criteria it is possible to distinguish several lineages of species; for example, a lineage of smooth shells with flexed brachial admicula exemplified by the series brevis-undulosa-isbelli (Pl. 31.1, figs. 26-31). All of these shells are more or less externally similar but the relative length of the brachial admicula and the degree of shell thickening in the pedicle valve increases along the series. Since they are found in successively younger deposits in many parts of eastern Australia, we believe they represent a slowly evolving series of populations. As the specific names are applied to populations sampled from the continuum (in the strictest sense the populations from the type localities), we should expect to find intermediates.

![Diagram of pelecypod zones](image)

Fig. 31.3. Summary of eastern Australian biostratigraphic data for the construction of a Permian time scale. Each column summarises the depositional environment and transgressive-regressive relationships of the major rock units. Fossiliferous intervals are shown in black (brachiopod-dominated) or by horizontal shading (mollusc-dominated). Brachiopod lineage-zones named for species of Martiniopsis Waagen or Wyndhamia Booker could be used to locate reference markers (arrows). Mollusc assemblage-zones named for species of the pelecypod Megadesmus Sowerby can be used to identify the time intervals in the nearshore sands.

The locations of the fossiliferous sections are: 1, Collinsville-Havilah area; 2, Gebbie Creek section; 3, Exmoo area; 4, Homevale-Carrinyah area; 5, north Lochinvar anticline; 6, Barrington district; 7, Pokolbin inlier; 8, Cranky Corner syncline; 9, coastal section, Ulladulla-Batemans Bay; 10, Montague Roadstead and N.S.W.D.M. Calalla No. 1; 11, Maria Island; 12, Hobart area; 13, Deep Bay section. The biostratigraphy of these areas is summarised in Runnegar et al. (in prep.), McClung (in press), Runnegar (in press a), Clarke and Banks (this volume), and Campbell and McKelvey (1972).
A small example will illustrate the point. The type horizons of *Martiniopsis undulosa* and *M. isbelli* are on opposite sides of a prominent sandstone ridge near Carrinyah homestead in the northern Bowen Basin (Runnegar et al., in prep.). They are separated by a stratigraphic distance of approximately 350 m. Each horizon contains a variable population, but only a small proportion of the specimens from the type horizon for *undulosa* could be referred to *isbelli* and vice versa. Both species occur in approximately equal numbers at a single locality in the Porcupine Formation at the extreme northwestern end of the Hunter Valley (Warner, 1972). We conclude that the Porcupine locality is intermediate in age between the type horizons of *isbelli* and *undulosa*.

A second example from the same lineage may be more convincing. Clarke (in Clarke and Banks, this volume) has large collections of smooth *Martiniopsis* with flexed admicula from eleven levels in an 85 m thick, perfectly exposed section at Deep Bay in the Cygnet area of Tasmania. The lower fossils resemble *Martiniopsis brevis* and these pass gradationally to forms resembling *M. undulosa* in successive beds in the section.

Using similar evidence from many parts of eastern Australia, we can show that the oldest species of *Martiniopsis*, *M. elongata*, produced the lineage *elongata-konincki-branxtonensis-ovata-plana-brevis-undulosa-isbelli* and also the lineages *elongata-konincki-branxtonensis-ovata-ingelarensis-mantuanensis* and *elongata-konincki-branxtonensis-profunda-profunda alta-plica-angulata* (Pl. 31.I; McClung, in press; Runnegar, in press a; Runnegar et al., in prep.). Some complications occur. The first lineage, represented by *elongata-konincki-branxtonensis-brevis-undulosa-isbelli* in the northern Sydney Basin.

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**Fig. 31.4.** Summary of zonal schemes for the eastern Australian Permian and later Carboniferous, and the New Zealand Permian.
### Advances in Stratigraphy and Palaeontology

<table>
<thead>
<tr>
<th>Zone</th>
<th>Possible location of reference point marking base of zone</th>
<th>Rock unit</th>
<th>Evaluation</th>
</tr>
</thead>
<tbody>
<tr>
<td>ovalis</td>
<td>Montague Roadstead, Jervis Bay, south Sydney Basin&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Nowra Sandstone</td>
<td>section perfectly exposed; first appearance of <em>ovalis</em> follows slight change in lithology; associated fauna not very diverse; possible</td>
</tr>
<tr>
<td>isbelli</td>
<td>West Carrinyah section, Nebo district, north Bowen Basin&lt;sup&gt;4&lt;/sup&gt;</td>
<td>upper part of Moonlight Sandstone</td>
<td>section well exposed; underlying beds mollusc dominated sands; diverse associated faunas; possible</td>
</tr>
<tr>
<td></td>
<td>Deep Bay section, Cygnet area, Tasmania&lt;sup&gt;3&lt;/sup&gt;</td>
<td>Malbina Formation</td>
<td>section well exposed; first appearance of <em>isbelli</em> follows change in lithology; diverse associated faunas; possible</td>
</tr>
<tr>
<td>undulosa</td>
<td>Deep Bay section, Cygnet area, Tasmania&lt;sup&gt;3&lt;/sup&gt;</td>
<td>unnamed alternating siltstone and fine sandstone</td>
<td>section perfectly exposed; uniform lithology; associated fauna diverse; ideal</td>
</tr>
<tr>
<td>brevis</td>
<td>Deep Bay section, Cygnet area, Tasmania&lt;sup&gt;3&lt;/sup&gt;</td>
<td>unnamed alternating siltstone and fine sandstone</td>
<td>section perfectly exposed; uniform lithology; lower beds not exposed, but have been drilled; possible</td>
</tr>
<tr>
<td>plana</td>
<td>Southeast of Exmoor homestead, northern Bowen Basin&lt;sup&gt;2&lt;/sup&gt;</td>
<td>upper part of Tiverton Formation</td>
<td>exposure poor but could be drilled; slight change in lithology; associated fauna not very diverse; not very good</td>
</tr>
<tr>
<td>ovata</td>
<td>Cedar Creek, Pokolbin inlier, north Sydney Basin</td>
<td>Farley Formation</td>
<td>exposure good; lithology uniform; few associated fossils; possible</td>
</tr>
<tr>
<td>branxtonensis</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>konincki</td>
<td>Cranky Corner syncline, north Sydney Basin&lt;sup&gt;4&lt;/sup&gt;</td>
<td>Cranky Corner Sandstone</td>
<td>exposure poor but has been drilled; first appearance of <em>konincki</em> follows change in lithology; associated faunas moderately diverse; possible</td>
</tr>
</tbody>
</table>

- *elongata* Zone is too poorly known to warrant formal definition at present time

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<sup>1</sup> Runnegar (in press a); <sup>2</sup> Runnegar et al. (in prep.); <sup>3</sup> Clarke and Banks (this volume); <sup>4</sup> McClung (in press).

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(McClung, in press), is reflected by the partial lineage *ovata-plana-undulosa-isbelli* in the northern Bowen Basin (Runnegar et al., in prep.), and by *ovata-plana-brevis-undulosa-isbelli* in southern Tasmania (data supplied by M. J. Clarke). These variations may result from local variations in the evolving lineages, but are probably due to sampling, which yields populations of slightly different ages in different areas. In each case an unfossiliferous interval occurs where a species is missing from the composite lineage (Fig. 31.3). These problems will probably never be totally resolved, and different workers will have different concepts of the biologic distribution of *undulosa* and its ancestors and descendants. As these concepts vary, so will the temporal significance of a zone named for one of the species. The only way to achieve stability is to fix the boundary between two zones in a section where the evolution of the species and the associated community can be observed and measured for comparison with other areas. We therefore recommend that where possible reference points be chosen to coincide with the boundaries between zones named for *elongata*, *konincki*, *branxtonensis*, *ovata*, *plana*, *brevis*, *undulosa*, and *isbelli* (Fig. 31.3; Table 31.1). Each reference point would then identify the beginning of a Chron and some of them would identify the beginning of an Age.

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**The evolution of Wyndhamia**

Many species of strophalosiid productoids have been described or identified from the Permian of eastern Australia. With the larger collections available to us we recognise relatively few species mostly confined to two different lineages (Runnegar et al., in prep.).
Almost all of the species have numerous fine spines on the exterior of the brachial valve, which is the reason they are called Wyndhamia and not Strophalosia (Clarke, 1969b).

Wyndhamia first becomes common in the ovata zone of all eastern states. The common species is W. preovalis, a small inflated shell with a deeply concave brachial valve and an unthickened shell (Pl. 31.II, figs. 5-7). It is the first representative of the lineage with which we are concerned.

Maxwell (1954) named the species preovalis because he believed it gave rise to a younger species he called ovalis. This species is similar in almost all respects, except that the insertion areas of the adductor muscles are located on a raised platform in the pedicle valve (Pl. 31.II, fig. 1). Another species, minima, described by Maxwell as a variety of W. clarkei, is in fact intermediate between preovalis and ovalis (Pl. 31.II; Dear, 1972b). These are the only names commonly applied to the lineage in eastern Australia, but W. maxwelli (Waterhouse, 1964) probably belongs here also.

The type locality of minima in the northern Bowen Basin yields specimens of both minima and ovalis (Dear, 1972b). Lower in the sequence only minima occurs, and lower still only preovalis. Overlying beds contain mainly ovalis (Runnegar et al., in prep.). It follows that preovalis evolves slowly into ovalis through minima. The same sequence of form is better displayed in the Wandawangan Siltstone of the southern Sydney Basin (Runnegar, in press a).

We regard the first appearance of ovalis as a useful zonal index, but again this is not
### Table 31.2. Ranges of Common Brachiopods

<table>
<thead>
<tr>
<th>BRACHIOPODA</th>
<th>levis</th>
<th>campbelli</th>
<th>elongata</th>
<th>koninki</th>
<th>transalpina</th>
<th>ovata</th>
<th>brevis</th>
<th>subalpina</th>
<th>tibelli</th>
<th>ovata</th>
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a concept that is easily transmitted to other workers. A reference point located in a sequence of beds containing *minima* and *ovalis* would provide a better standard.

**Other Fossils**

The primary zonation is of course supplemented by many other fossils. Zones based on locally abundant fossils have been described by Dear (1972a) in the Bowen Basin and by Clarke and Banks (this volume) from Tasmania. McClung (in press) recognises a zone earlier than the *elongata* zone in the northern Sydney Basin by the mutual occurrence of *Trigonotreta campbelli, Eurydesma cordatum*, and *Megadesmus pristinus*. This appears to succeed the Carboniferous assemblage zones named for the productoid brachiopods *Marginirugus bar ringt oneis* and *Levipustula levis* by Campbell and McKellar (1969). Eventually these and perhaps other yet unrecognised zones will require reference points. We are arguing here for the necessity for reference points rather than attempting to finalise a time scale.

**Are sections suitable for reference points available?**

Most of the zones we have outlined could be formalised by reference points in adequately exposed fossiliferous sequences containing both the ancestral species and the zonal index (Table 31.1; Fig. 31.3). Few of these sections are ideal, but most of these and several others could be improved by a modest diamond drilling program.

Do the zones cover the whole of the Permian Period?

The answer is almost certainly 'no', since very late Permian marine deposits are unknown from eastern Australia. Other sequences, possibly those of the Canning Basin in Western Australia, will be needed to complete the younger part of the time scale.

At the other end of the scale, the zones probably include all of the Early Permian, but may include parts of the Carboniferous as well (Fig. 31.4; Runnegar, in press b). The precise identification of the Permian-Carboniferous boundary in the Gondwana province requires additional study.

**APPLICATIONS FOR THE GONDWANA PROVINCE**

If a series of formally named Ages were based on reference points (boundary stratotypes) located in well-documented fossiliferous marine sections from eastern Australia and elsewhere, these would provide an international time framework for the Gondwana province. Palynological and other fossil zones could be related to the reference point-bounded Ages and the time scale could be applied to both marine and non-marine sediments. We ask whether the procedures we have outlined are acceptable, and whether the formalisation of such a time scale is desirable. If the answers to both questions are affirmative, we suggest a small international committee be formed to select appropriate reference point localities.

**ACKNOWLEDGMENTS**

This study was supported by grants from the Australian Research Grants Committee and the University of New England. We thank Mike Clarke, Tasmanian Geological Survey, for a great deal of information and spirited discussion of the procedures involved. The ideas presented have also been discussed with J. M. Dickins, Bureau of Mineral Resources, Robin Helby, New South Wales Geological Survey, N. F. Hughes and W. B. Harland, University of Cambridge, and Rodney Gould, University of New England; we thank them for their interest and advice.

**REFERENCES**

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---, Armstrong, J. D. and Runnegar, B., in prep. *Martiniopsis*-like spiriferids from the Permian of eastern Australia.


### Table 31.4. Ranges of Common New Zealand Brachiopods and Pelecypods

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<th>Brachiopod</th>
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--- in press b. Late Paleozoic Bivalvia from South America: provincial affinities and age. An. Acad. brasil Cienc.
---, McClung, G. and Javes, I. R., in prep. The evolutionary basis for Queensland divisions of Permian time.


Estheriids of the Indian Gondwanas: Significance for Continental Fit


ABSTRACT

Faunal studies on Late Permian to Early Jurassic estheriids of India are reported. The estheriid-Glossopteris association in the Kamthi Formation at Kawarsi and Punwat is here considered to be Late Permian. Red-banded shale of the Mangli Beds, with multiple estheriid zones, 91 ± m above the Glossopteris bed at Kawarsi, is here considered to be Early Triassic.

Restudy of the Panchet Formation (Raniganj Basin) revealed sixteen new estheriid-bearing units. In addition to an undetermined number of units at fifteen Lystrosaurus localities, estheriids have been found from 24.4 to 518.2 m above the Permo-Triassic transition (= Schizoneura-Glossopteris zone). The Mangli Beds with eleven estheriid-bearing units (and containing the genus Cornia), can now be stratigraphically placed between two of the above estheriid-bearing exposures 90 and 106.6 m above the transition. These clarifications allow comparisons with Triassic estheriids from western and eastern Australia; palaeolimnadiids occur in both, as do cyziciids.

The Lower Jurassic Kota argillaceous limestone biota (Pranhita-Godvari Basin) with seven new estheriid-bearing units, have numerous similarities with the Antarctic non-marine Jurassic biotas (Carapace Nunatak-CN, Storm Peak, Blizzard Heights, etc.). Both have abundant conchostracans (at least two common genera—Cyzicus and Palaeolimnadia), insects (beetle elytra—now known from the Kota Formation and two Antarctic localities—Southern Victoria Land (CN) and Hope Bay on the Antarctic Peninsula; and insect nymph abdomen), as well as fish, plants and ostracodes. These data support a previously postulated dispersal route across Antarctica and India for Mesozoic estheriids and insects.

INTRODUCTION

Scattered fossil estheriid occurrences from the Indian Gondwana formations have long been reported from reconnaissance studies (Jones, 1862). Two new species were described by Jones from Mangli and Kota. Feistmantel (1877) discussed the fossil biota (estheriids and plants) in an attempt to clarify the age and correlation of specific formations. However, a systematic, biostratigraphic evaluation of the conchostracan-bearing beds has not been available previously. A clear picture of conchostracan biostratigraphy in the Indian Mesozoic especially, is relevant to the problem of the relations of continental blocks in the Gondwana configuration. Consequently, a field program during 1971-72 was designed to obtain data from the following areas (teams involved are indicated): the Panchet Formation (Raniganj
Advances in Stratigraphy and Palaeontology

Fig. 32.1. Map showing localities from which material mentioned in the text has been recovered

Fig. 32.2. Conchostracan localities of the Panchet Series, Raniganj Coalfield
KOTA LIMESTONE

Fig. 32.3. Upper Gondwanas of Pranhita-Godavari Basin. Outcrop Belt of Kota Limestone Basin, Fig. 32.2; P. Tasch and S. C. Ghosh); the Kota Formation (Pranhita-Godavari Basin, Fig. 32.3; P. Tasch, C. N. Rao, and B. R. J. Rao, with S. C. Shah as consultant on previous collections of the Kota-Maleri biota); Mangli Beds (see Fig. 32.1) and various Glossopteris bed explorations (P. and conchostracan-insect beds (K-1 to K-7). Based on map by C. Nageswara and S. C. Shah.

Tasch and B. R. J. Rao). Technical advice has been given by M. V. A. Sastry.

PERMIAN AND PERMO-TRIASSIC TRANSITION

Permian outcrops were studied at four localities. Locality I: below station RN-1, lat. 23°41'55'', long. 86°57'30''. Locality II:

At Locality I, two beds with *Schizoneura* were found 65—158 m below the base of station RN-1. The upper one of these is equivalent to Locality III (see below), and is here designated the *Schizoneura-Glossopteris* zone in the Raniganj Basin. At Locality II beds with *Glossopteris* were found by channeling, 0.5 and 1.73 m below the soil base. A third bed with a carbonised flora occurred in a grey argillite an undetermined distance below the second. Fossil estheriids are known from the upper of these three beds.

At Locality III, below a capping sandstone, a carbonaceous shale contained *Schizoneura* and other plants. *Glossopteris* is known from these beds. This is the equivalent of the transition zone at Locality I.

At Locality IV, a massive sandstone overlain by a thin argillaceous sandstone contained fossil estheriids and abundant *Glossopteris*. The uppermost joint occurrence of *Schizoneura* and *Glossopteris* is taken as the Permo-Triassic transition. The possible extension of this datum outside the Raniganj Basin (see Fig. 32.1) is now under investigation by the Geological Survey of India. One extension has been inferred—in the Mangli Beds. Since these occur about 100 m above the *Glossopteris* bed at Kawarsi (Loc. III)

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![Diagram](image-url)
they are here regarded as Triassic in age, and placed in the Panchet Formation of the Raniganj Basin (Fig. 32.4).

TRIASSIC

In a traverse of the Panchet Formation, coals and plant-bearing carbonaceous shales \((Schizoneura-Glossopteris\) zone) pass stratigraphically up into khaki-green, highly micaceous siltstones (Lower Panchet) which often contain coquinas of small fossil estheriids. The overlying red shales (Upper Panchet) contain a similar conchostracan fauna, worm burrows, carbonised plants and other organisms, along with the \(Lystrosaurus\) fauna (Tripathi and Satsangi, 1963), and fish scales.

The Triassic estheriids occur at twenty-seven localities and range from the \(Schizoneura-Glossopteris\) zone throughout the Panchet Formation. At the localities sampled by Tasch and S. Ghosh, the number of estheriid-bearing units at each locality varied from four to ten (Fig. 32.4). Among the estheriids found are a few species of \(Cyzicus\) and a palaeolimnadinid species. At Mangli, which is 100 km south of Nagpur, Central India, four species including \(Cyzicus mangliensis\) (Jones), two other cyzicid species, and a species of the spined genus, \(Cornia\), are known.

The recurrent estheriid beds in the Raniganj coalfield are matched in the Panchet of the Bokaro coalfield according to S. Ghosh. This raises the possibility of a more precise interbasinal correlation. In addition conchostracan faunas continue into the formation next above the Panchet, the Mahadeva.

JURASSIC

The outcrop belt of the Kota Limestone is shown in Figure 32.3. A similar estheriid-
bearing ferruginous silt and shale to the one at Metapalli (K-4), occurred in southern outcrops as a capping bed at Chittur (K-5A) (Fig. 32.5), and also in the type section of the Kota Formation (K-7) (Fig. 32.3). Bed 9 at K-1 could also be an equivalent. Repetition of this bed in southern and northern outcrops may possibly be due to faulting.

In several respects the Kota Formation is remarkable for its fauna. It contains abundant insects, often excellently preserved. There are four to six genera of Coleoptera, and fossil elytra and isolated abdomen are common in several beds. As many as twenty species may be represented. There are several blattids, which are prominent at several Gondwana localities outside of India. There are at least eight distinct insect-bearing horizons in a vertical sequence (Fig. 32.5; for example K-1 and K-2). This is a minimum figure since the missing interval represents a thickness double the total for the measured sections, and might very well contain several more such horizons. Beetle elytra in bed 5 at K-5 should permit further correlation of southern and northern outcrop belts, presumably linking K-5 and K-1, and/or K-2, which also bear similar remains.

Fish scales and other skeletal fragments were noted in both northern and southern belts at Kota. Vertebrates have long been known, having been studied by Kutty and Roy Chowdury (1972), Robinson (1967) and associates. They include sauropod dinosaurs and crocodilians.

Species of cyziciids, palaeolimnidads and estheriinids recur through the sections (Fig. 32.5; for example K-1 and K-2). Other associates of the estheriids are fish, plants and abundant insects. The significance of these associations will be discussed below.

Two other items pertaining to the Kota merit comment here.

1. An indicator of the size of the mini-basins (ponds) bearing the estheriid fauna was recorded at the type locality (K-7) (Fig. 32.6). Here the estheriid-bearing bed extended over a distance of some 35 m.

2. Recurrent mud-cracks and ripplemarks (Loc. K-5, K-6) denote drying events and shallow water respectively.

Indian Triassic Gondwana Estheriids in a Gondwana Context

The species of *Cyzicus* are not sufficiently well known to permit a comparison of those from India with those of other Gondwana continents. Poor preservation in highly micaceous siltstones further complicates the problem. Nevertheless, cyziciid cosmopolitanism could be related to proximity of continents. There are two elements of the India fauna that can be so compared with faunas elsewhere: palaeolimnidads and a spined estheriid, *Cornia*.

Recent studies by Tasch on Bureau of Mineral Resources, Australian Museum, his own and other collections, have shown palaeolimnidads to be quite widespread in the Gondwana area. They occur in the Triassic rocks of the Bowen and Sydney Basins of eastern Australia, and the Bonaparte Gulf and Canning Basins of Western Australia. Further, they are also known from the Triassic of Brazil (Botucatu Sandstone). A palaeolimnidad from the Karroo sequence (Triassic) of Angola, Africa (see Fig. 32.7) is at present under study.

Of course palaeolimnidads range from the Permian through the Jurassic, and occur outside the Gondwana area as well. What is important is the fact that palaeolimnidads were widely distributed on the continents of the southern hemisphere (Tasch, 1970: fig. 2).
The genus *Cornia* is distinctive in that it bears a small umbonal spine. It ranges from the Carboniferous through the Triassic of the U.S.S.R., and it is now known also in Australia. Although not recognised as such by Mitchell (1927: pl. 2, fig. 4), a species of *Cornia* occurs in the Newcastle Coal Measures, a fact Tasch confirmed by study of the Australian Museum specimens (Number F.25474, labelled *Estheria coglanii* and consisting of three specimens, one of which bears a small umbonal spine). Aside from cyziciids which occur in both cited areas, these new data extend the picture of biological connections between southern continents.

The affinities of the Russian conchostracan faunas were discussed at the South African Symposium (Tasch, 1970), at a time when India's non-marine arthropod fauna was still poorly known. Indications based on fossil data were given of the dispersal of insects across India to the Russian area in Figure 1 of that paper. Now, the new evidence from Mangli provides more data on this dispersal route, which was also used by various estheriids including *Cornia*.

**COMPARISON OF THE INDIAN JURASSIC AND OTHER GONDWANA MESOZOIC ESTHERIID**

In the same Australian collections referred to in the previous section, species of *Estheriina* were found. These came from cores in the Lower Triassic of both eastern and Western Australia (Rewan Formation, Bowen Basin, and Blina Shale, Canning Basin respectively). Though the record is sparse, distribution of estheriinids in the Jurassic Kota Formation of India, the Lower Cretaceous of Brazil (Bahian), and in the Australian Triassic denotes that estheriinids were widely dispersed
throughout Gondwana continents during Mesozoic time. Data on other estheriid genera and insects, discussed below, sustain this interpretation (Figs. 32.7 and 32.9).

Likewise species of Palaeolimnadia in the Indian Jurassic can be compared with palaeolimnacid of the Transantarctic Mountains (Queen Alexandra Range) and Mauger Nunatak (Fig. 32.9), the Western and eastern Australian palaeolimnacid populations of Triassic age, as well as the Russian and South American forms of the same age (Botucatu Sandstone).

Data from both the Indian Triassic and Jurassic estheriids (and as considered subsequently Jurassic insects) lead to similar inferences on the proximity of continents at the cited time.

Fig. 32.8. The Indian Mesozoic non-marine fauna in its Gondwanaland context

As depicted in Figure 32.7, fossil beetles (Coleoptera) presently are distributed in the Triassic of Australia and Africa, Jurassic of Antarctica (Carapace Nunatak and Hope Bay), India (Kota Formation), and South America (Tasch, 1970).

This pattern of coleopteran zoogeography fits the evidence from other Gondwana insects (homopterans for example) (Tasch, 1971). It also adds new data pertaining to dispersal/migration routes through Jurassic time in the southern hemisphere (Tasch, 1970: fig. 1).

When Zeuner described the Jurassic beetle from Grahamland, Antarctica (1959) he was unaware of such a fauna in the Kota Formation (Rao and Shah, 1959), or of the Antarctic beetle from Carapace Nunatak (Tasch, 1973). Accordingly he compared his beetles only to the Australian Triassic and the Rus-
Estheriids of the Indian Gondwanas

Fig. 32.9. Gondwanaland distribution of two Mesozoic conchostracans

The new data discussed above suggest that the Indian subcontinent was positioned close to both Antarctica and Australia. Smith and Hallam (1969: fig. 1) positioned India close to Madagascar and Antarctica, while Veevers et al. (1971: fig. 4) placed India close to southwest Australia and not in contact with Antarctica. Jurassic beetles and palaolimnadinids from Antarctica and India and Triassic palaeolimnadinids, and estheriniids in India and Australia, required non-marine dispersal routes between these areas during the indicated geologic periods (Figs. 32.7, 32.8, 32.9). Neither of the cited reconstructions fully meet these specifications, although each meets them in part. Some further adjustment seems needed in fitting India into the Gondwana configuration (cf. Johnstone et al., 1973: fig. 2A).
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The Stratigraphy of the Lower (Permo-Carboniferous) Parts of the Parmeener Super-Group, Tasmania

M. J. CLARKE and M. R. BANKS

ABSTRACT
Detailed information obtained from the systematic 1:15,840 mapping program of the Geological Survey of Tasmania, fully cored diamond drill holes, and certain large-scale civil engineering projects is used to supplement and amend existing published accounts of the lithostratigraphy of the lower (Permo-Carboniferous) parts of the Parmeener Super-Group. Whereas macro- and microfloral criteria prove the lowermost parts of the Parmeener Super-Group to be of Late Carboniferous age, for most part the faunas and floras are of Permian age and display affinities which are almost exclusively with the Eurydesma-Glossopteris cold-water realm. Biostratigraphically, the Tasmanian sequence is as complete as any in eastern Australia, if not more so, and ten informal assemblage faunizones are recognised within the Gondwana Eurydesma fauna. The detailed distributions of these faunizones indicate considerable lateral variations in lithofacies from place to place, and demonstrate the presence of a depositional and/or erosional hiatus of variable duration over much of the Tasmania Basin. On the basis of Glossopteris and Dulhuntyispora microfloras, the uppermost Permian is represented in non-marine sequences which persist until the end of Triassic times. No satisfactory method of tracing the Permo-Triassic boundary in Tasmania is known.

INTRODUCTION
In Tasmania, the lower (Permo-Carboniferous) rocks of the Parmeener Super-Group (Banks, 1973) are almost everywhere sub-horizontal, and rest with pronounced landscape unconformity on a folded basement composed of Precambrian and early Palaeozoic strata intruded by Late Devonian to Early Carboniferous granites (McDougall and Leggo, 1965). Compared with other eastern Australian sequences in the Sydney and Bowen Basins, that of the Tasmania Basin is much thinner, and (excluding the basal tillite which is extremely variable in its development) rarely exceeds 500 m in thickness. However, despite this much reduced thickness, palaeontological evidence demonstrates that the Tasmanian sequence is as complete as any in eastern Australia, if not more so. Lateral variations in lithofacies are considerable, particularly in the vicinity of basement highs. As a consequence, rock unit terminology varies widely from place to place, and palaeontological criteria provide the sole means of detailed correlation. Nevertheless, rocks of the lower parts of the Parmeener Super-Group lend themselves to a broad quadripartite lithological or environmental subdivision. These divisions are: 'Lower Marine Sequence' (including the basal tillite); 'Lower Freshwater Sequence'; 'Upper Marine Sequence'; and 'Upper Freshwater Sequence' (c.f. Johnston, 1888; Twelvetrees, 1911; David, 1950). Such a
subdivision is applicable to most of Tasmania, and for the sake of simplicity is used herein. Important exceptions include the Cygnet area, southern Tasmania, where the Lower Freshwater Sequence is absent, and most of northeastern Tasmania where the Lower Marine Sequence is absent and the remainder of the succession is much attenuated. The use of this crude quadripartite subdivision does not imply that the lower and upper boundaries of any one unit are necessarily everywhere of the same age. On the contrary, evidence is presented which demonstrates that the boundaries of the Lower Marine Sequence, Lower Freshwater Sequence and Upper Marine Sequence vary in age from place to place.

The adoption of this simplified subdivision may appear to be a retrograde step in view of the more detailed lithostratigraphic scheme used in the last major compilation of Tasmanian Permo-Carboniferous stratigraphy (Banks, 1962). However, progress in several fields has been rapid, and it is now considered necessary to adopt a more rigorous separation of lithostratigraphic and biostratigraphic criteria. In particular, officers of the Geological Survey of Tasmania have completed the detailed 1:15,840 regional mapping of Parmeener Super-Group rocks in northern Tasmania, and since 1968, have extended this to southern Tasmania. This program in itself has solved many lithostratigraphic problems, but as is only to be expected, other

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**Fig. 33.1. Generalised lithological faunal characters, Parmeener Super-Group (lower part), and suggested external correlations**
Stratigraphy of the Parmeener Super-Group

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difficulties have been brought to light. Many of these have been resolved by fully cored stratigraphic bore holes. In addition, certain large-scale civil engineering projects such as the Hydro-Electric Commission Fisher Tunnel and the Associated Pulp and Paper Manufacturers Mersey Great Bend-Wesley Vale pipeline, have provided a wealth of detail in critical, and until recently, little-known sequences.

The last decade has also witnessed the welcome innovation of local biostratigraphic 'zonal' schemes (Dickins et al., 1964 et seq; Waterhouse, 1965 et seq; Runnegar, 1967 et seq. See Fig. 33.1). Whereas various tentative correlations of the Tasmanian sequence with these schemes have been attempted, most are incomplete, and only those of Runnegar (1967, 1969a) withstand critical examination, and then only in part. This is understandable, partly because most outside workers lack familiarity with Tasmanian lithostratigraphic details, but also because of the lack of modern biostratigraphic data for the Tasmania Basin. It is hoped that some of these deficiencies will be rectified herein. The new lithostratigraphic information is summarised where appropriate and a preliminary statement of Tasmanian faunal distributions is presented. Lithostratigraphic and biostratigraphic considerations are rigorously differentiated. References earlier than Banks (1962) are given only where certain lithostratigraphic names as used herein differ from that work. Details of the authorship and original definitions of many of these names are given in Smith (1957).

PARMEENER SUPER-GROUP (LOWER PART)

The oldest rocks within the lower parts of the Parmeener Super-Group which can be reliably dated comprise rhythmite clays which occur stratigraphically near the middle of the Wynyard Tillite in its type area in the Wynyard-Hellyer Gorge area, northwestern Tasmania. Near the summit of the rhythmite sequence the occurrence of *Rhacopteris ovata* (M'Coy) (Banks, 1967) and a Stage 1 microflora (Evans, pers. comm.) indicate a Late Carboniferous age. This horizon has also yielded the oldest known fossil insect in the southern hemisphere (Riek, pers. comm.). Below this and within the lower parts of the rhythmite sequence, the probable arthropod track *Tasmanadia twelvetreesi* Chapman occurs in profusion. *Tasmanadia* is similar to certain of the tracks described and figured as *Diplichnites* from the Dwyka Series in South Africa (Savage, 1970). The remainder of the Wynyard Tillite has so far proved unfossiliferous. At a much higher stratigraphic level in the Maydena area, *Deltopecten illawarensis* (Morris), *Eurydesma* and *Pyramus* have been recorded from conglomerate near the summit of a sequence which has been correlated with the Wynyard Tillite (Runnegar, 1969a; Jago, 1972). A fragmentary *Eurydesma* (Allandale) fauna also occurs towards the summit of a tillitic conglomerate in the Frankford area. It is therefore evident that tillite deposition occurred during the Late Carboniferous and persisted until the Early Permian. It is less evident that deposition began and ceased everywhere at the same time. Nevertheless, it is believed that the Permo-Carboniferous boundary occurs within the upper parts of the Wynyard Tillite, although in its type area, the first appearance of the *Eurydesma* fauna is about 100 m above its summit. It must also be admitted that outside Tasmania, considerable doubts still remain about the relative ranges of the *Rhacopteris* and *Glossopteris* floras, about the coincidence of the incoming of the *Eurydesma* fauna with that of the *Glossopteris* flora, and about the exact relationship of these changes with respect to the Permo-Carboniferous boundary (Helby, 1969; Black, Morgan and White, 1972).

Basal tillite occurs in many places west of the meridian of Hobart. The thickest proved development is in the type area of the Wynyard Tillite where a thickness of about 600 m includes glaciolacustrine rhythmite clays as well as tillite (Gulline, 1967; Gee and Gulline, in press). In the Woodbridge-Cygenet area diamond drilling has so far proved a minimum thickness of 300 m for the Woodbridge Tillite, and it may be far thicker. At Maydena, the tillite is 175 m thick (Jago, 1972), and at Poatina the Stockers Tillite is at least 110 m thick (McKellar, 1957). Generally, these thick developments of tillite occur in basement troughs. However, in the
Hobart area the Glenorchy Borehole showed an absence of tillite. This, coupled with a much increased thickness of the pyritic and glendonitic Woody Island siltstone compared with surrounding areas, may indicate that the Hobart area formed part of a deeper water zone beyond the limits of tillite deposition. In the region of basement highs such as Golden Valley, Western Bluff, Fisher River, Beaconsfield, Frankford, Cradle Mountain and Maria Island, the tillite is either very thin or absent, and is generally replaced by much thinner developments of conglomerate.

Above the poorly fossiliferous Wynyard Tillite and its lithological correlates, rich marine faunas become established. They are almost wholly of the Gondwanan *Eurydesma* realm. Ten successive informal assemblage faunizones are recognised (Fig. 33.1). The palaeontological basis of these assemblages is shown diagrammatically (Fig. 33.2) and summarised in the following paragraphs.

**Faunizones 1-3**

Collectively, these assemblages are characterised by the presence of *Megadesmus globosus* (J. Sowerby) var. nov., *Myonia morrisi* Etheridge, *Neoschizodus australis* Runnegar, *Pyramus laevis* (J. Sowerby) and an enormous profusion of *Deltotpecten (illawarenis-waterfordi)* Dickins group), *Eurydesma* (mainly *hobartenses* (Johnston)) and *Keeneia* (*twevtreessi* Dum—*oca!* (J. Sowerby)—*platschismoides* Etheridge). In addition, brachiopods are well represented and diagnostic forms include a characteristic neospiriferid which is almost certainly *Trigonotreta stokesi* Koenig sensu stricto non Armstrong 1968, *Martiniopsis ovulum* (Waterhouse)—*koninci* Etheridge group, and *Pseudosyrinx allandalesis* Armstrong. Faunizone 1 is characterised by the additional occurrence of *Srophalosia* sp. nov. (generally similar to *Wynhamia ovalis* (Maxwell), but smaller, no dorsal valve spines, and brachial ridges strongly developed), *Streptorhynchus* sp. nov. (aff. *S. pelicanensis* Fletcher—dorsal valve more inflated, ears more differentiated), *Cyrtella nagmargensis australis* Thomas, *Phestia darwini* (de Koninck) and *Promytilus cancellatus* Maxwell. *Costalosia apecallosa* Clarke, *Srophalosia subcircularis* Clarke and *Eurydesma cordatum* Morris are confined to Faunizone 2. *Srophalosids* and *E. cordatum* are not present in Faunizone 3, but more importantly, *Sulciplica stutchburii* Auctt, *Sulciplica* sp. nov., a new notospiriferid genus and *Rhabdocantha* appear for the first time.

On the basis of the bivalves and gastropods, Faunizones 1-3 can be confidently correlated with the Sydney Basin Allandale Fauna. This correlation is also supported by the comparatively meagre brachiopod evidence from the Sydney Basin (Runnegar, 1969b; McClung, pers. comm.), and the presence of late Stage 2 microfloras throughout the Golden Valley Group and most of the Quamby Mudstone at Golden Valley. A similar association of an Allandale macrofauna and late Stage 2 microfloras occurs in the Cranky Corner Basin, New South Wales (Helby, pers. comm.; Runnegar, pers. comm.).

Whereas the comparative rarity of brachiopods in the Sydney Basin Allandale Fauna, and developments of poorly fossiliferous volcanic sequences at this level in the Bowen Basin may prevent a more widespread individual recognition of Faunizones 1-3, their value in Tasmania is considerable. Thus the establishment of conditions suitable for the proliferation of shallow water benthonic faunas in erratic-rich siltstone, sandstone, calcareous siltstone and limestone of 'Golden Valley Group' facies (Golden Valley Group proper (Clarke, 1968); Massey Creek Group (Gee, 1971), in part; Kansas Creek Beds, in part; Spreyton Beds (Burns, 1964), in part; Inglis Siltstone (Gee, Gulline and Bravo, 1968), in part; Darlington Limestone; Bundella Mudstone) varied in time from place to place (Fig. 33.4). The dark, massive-bedded, pyritic and glendonitic Quamby Mudstone, which is a litho-stratigraphic unit, is more or less confined to the Golden Valley- Poatina area and lacks diagnostic macrofaunas (Wells, 1957; Clarke, 1968). However, Tasmanite Shale occurs near the base of the formation at Quamby Brook, near the base of the Spreyton Beds at Latrobe, and
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near the base of the Inglis Siltstone in the Hellyer Gorge, as well as several other localities in northern and northwestern Tasmania (Fig. 33.4). At this latter locality an *Ammodiscus oonahensis* Crespin formainferal assemblage occurs. Usually the Tasmanite Shale lacks macrofossils, but at Latrobe it (together with the Spreyton Beds immediately above) yields rich macrofaunas which can be confidently assigned to Faunizone 1 (Clarke, in Jennings, in press). On this basis, the Quamby Mudstone in its type area may be indirectly inferred to approximate in age to Faunizone 1. The Tasmanite Shale probably formed marginal to basement ridges in a restricted basinal environment.

In southern Tasmania diagnostic macrofaunas commence with the Bundella Mudstone. In the Hobart area all assemblages belong to Faunizone 2, but further south in the Cygnet district, the lowermost parts of the formation are significantly older and yield assemblages typical of Faunizone 1.

On Maria Island the Lower Marine Sequence is much reduced in thickness and rocks of Faunizone 1 age are absent. The lowermost beds comprising the ‘Erratic Zone’ and the Darlington Limestone both belong in Faunizone 2. Above these units the Lower Freshwater Sequence rests directly on a few metres of thin-bedded calcareous siltstone rich in bryozoans and spiriferids and belonging to Faunizone 3. Elsewhere in most of eastern and northeastern Tasmania the Lower Marine Sequence is absent (Fig. 33.4). The sole exception is at Musselroe Bay in the extreme northeastern corner of the island, where rich Faunizone 2 assemblages are developed immediately above granite basement. This occurrence delimits a northern or eastern margin to an elongate peninsula-like land area which occupied most of northeastern Tasmania during Lower Marine Sequence times (Banks, 1962).

The recognition of a tripartite subdivision of a more broadly conceived Allandale Fauna within the Lower Marine Sequence is also important with respect to the base of the Lower Freshwater Sequence. In the Hellyer Gorge-Prestolenna area, the Fisher River area, and the Mersey Coal Basin, only Faunizone 1 can be recognised within the Lower Marine Sequence. To the east and northeast, through the Frankford, Quamby and Beaconsfield Quadrangles the Lower Freshwater Sequence rests successively on Faunizone 2 and Faunizone 3 assemblages (Fig. 33.4). Usually, as at Poatina and Golden Valley, the passage from the Lower Marine Sequence into the Lower Freshwater Sequence is gradational, but in the Fisher Tunnel it is abrupt (Clarke and Farmer, in press). It is possible, therefore, that a disconformity may be locally developed at the base of the Lower Freshwater Sequence and/or Faunizones 2 and 3 may be represented in unfossiliferous beds such as the upper parts of the Kansas Creek Beds in the Fisher River area, and the Macrae Mudstone at Poatina and Golden Valley. Overall, however, this seems to be unlikely and the simplest interpretation of the available evidence suggests that the base of the Lower Freshwater Sequence is diachronous, with the onset of non-marine conditions occurring first in the north and northwest and only later spreading to the east, northeast and southeast.

Faunizones 4-5

Together, Faunizones 4 and 5 witness a profound change in faunal composition. Part of this abruptness is due to the fact that no continuously marine sequence is developed through the Faunizone 3-Faunizone 4 interval because of the intervention of the Lower Freshwater Sequence. The most important faunal changes include the appearance of true productids and aulostegids (*Anidanthus*, *Cancrinella*, *Terrakea*, *Taeniothaerus*), the appearance of *Martiniopsis* of the *ovata* (Campbell), *profunda* (Campbell) and *valida* (Campbell) species groups, the appearance of costate spiriferids such as *Grantonia hOBartensis* Brown, *G. cracovensis* Wass and *Sulciplica tasmaniensis* (Morris), and the appearance of *Punctospirifer*. Syringothyroids become very rare but strophalosids continue in profusion and include members of the *Wyndhamia preeovalis* (Maxwell), *W. enorme* Clarke and *W. jukesii* (Etheridge)—
Stratigraphy of the Parmeener Super-Group

dalwoodensis Booker species groups. Faunal diversity increases with the appearance of simple rugose and tabulate corals such as Euryphyllum, Cladochonus and Thamnophora, blastoids such as Thaumatoblastus, conulariids such as the giant Paraconularia, certain bivalves such as Conocardium and Atomodesma ( Aphanaia), rare occurrences of Spirigerella and trilobites (Wass and Banks, 1971), and an increase in the variety and abundance of Fletcherithyris and Gilledia, together with crinoids such as Calceolospongia. Part of the increased faunal diversity of Faunizone 5 and the predominance of carbonate rocks at this level may indicate slightly higher water temperatures. The restricted vertical distribution of Taeniothaerus and Terrakea appears to be more fundamental since both occur in a variety of rock types. Despite the pronounced faunal changes which take place between Faunizone 3 and Faunizone 4, several Lower Marine Sequence species groups persist. In particular, Deltopecten, Eurydesma and Keeneia continue in abundance, although never in quite the same extraordinary profusion as in the Lower Marine Sequence.

The occurrence of Anidanthus springsur- ensis (Booker), Cancrinella farleyensis (Etheridge and Dun), Taeniothaerus subquadratus (Morris), Terrakea pollex Hill group, Wyndhamia preovalis, W. enorme, W. jksesidalwoodensis, Martinipis ovata, M. profunda, M. valida, Gilledia homevalensis Campbell and many other forms indicates that collectively, Faunizones 4 and 5 can be confidently equated with Fauna II in the Bowen and Sydney Basins. Faunizone 4 may approximate the New Zealand Telfordian Stage, and Faunizone 5 appears to include at least part of the succeeding Mangapirian Stage. Since the Lower Freshwater Sequence (Faulkner Group in the Hobart area and Maria Island; the Aberfoyle and Elephant Pass Formations in northeastern Tasmania (McNeil, 1965); and the Liffey Sandstone (=Mersey Coal Measures (Stephens, 1873), in northern Tasmania) is directly overlain by

Faunizones 4-5 (=Fauna II), it is evident that it is significantly older than the Greta Coal Measures in New South Wales (Runnegar, 1969a). The Mersey Coal Measures cannot be proved to be 'a lateral facies of the Berriedale Limestone' (Waterhouse, 1970), but the Greta Coal Measures and Berriedale Limestone may well be of Mangapirian age.

Faunizones 4-5 (Cascades Group and correlates in Banks, 1962) appear to have a rather more restricted distribution than hitherto thought. They are certainly present in the Hobart area (Rayner Sandstone and Cascades Group), Maria Island (‘Productus’ and ‘Crinoidal Zones’), Friendly Beaches (Peter Limestone), Elephant Pass (Gray Siltstone (McNeil, 1965)) and various other localities in northeastern Tasmania, at Maydena (Jago, 1972) and Frankford. At the last two localities the faunas are associated with developments of sandstone and conglomeratic sandstone, but elsewhere the predominant rock types are calcareous siltstone and limestone. Faunizones 4-5 have not been recognised in the extensive Parmeener Super-Group outcrop of the Great Western Tiers nor elsewhere in northern and northwestern Tasmania other than at Frankford. In the Fisher River area, and inferentially over much of northern and northwestern Tasmania, a pronounced hiatus or paraconformity separates the Lower Freshwater Sequence and the Upper Marine Sequence. In the Cygnet area, southern Tasmania, this hiatus additionally embraces the Lower Freshwater Sequence, so that an incomplete Upper Marine Sequence rests on an incomplete Lower Marine Sequence. In detail the situation is complex because at Cygnet the base of the Upper Marine Sequence is younger than in the Hobart area, but older than in the Fisher River area (Fig. 33.4 and see later).

Faunizones 6-10

Assemblages younger than Fauna II are well developed in Tasmania. Rich Fauna IV assemblages with Wyndhamia ovalis are known from Malbina Member E of the Hobart area and its correlates at Eaglehawk Neck and elsewhere in southern Tasmania (Banks and Read, 1962). Below this level,
rich post-Fauna II—pre-Fauna IV assemblages are developed. Basically these faunas are not like the distinctive Queensland Fauna III, but like the mixed Sydney Basin Ulladulla Fauna. These faunas, despite an overall similarity of character, show disconcerting variations from place to place. They are also rather sporadically developed in otherwise unfossiliferous rocks of the Upper Marine Sequence, a sequence which is much affected by rapid lateral variations in lithofacies, particularly in its lower parts. As a consequence a more detailed evaluation of these faunas has proved difficult. However, the recent discovery of a continuously exposed and continuously fossiliferous sequence through this critical interval at Deep Bay, Cygnet, now permits a clarification of most of these difficulties. The Deep Bay section realises many of the attributes of the ideal stratigraphic section. It is continuously exposed and unfaulted, continuously fossiliferous, the fossils are well-preserved, and for most part, there are no abrupt changes in lithology. The details of the Deep Bay section are shown diagrammatically (Fig. 33.3) and summarised below.

Faunizone 6 This assemblage is marked by the incoming of Martinioptis ingelarensis (Campbell), M. sp. nov.—M. undulosa (Campbell) group, M. angulata (Campbell) —M. globosa (Campbell) group, M. magna (Campbell), Fletcherithyris parkesi Campbell and Myonia corruagata Fletcher. Sulciplica stutchburi occurs in profusion. Aperispirifer wairakiensis (Waterhouse) is the dominant neospiriferid. Deltopecten and Eurydesma are still abundant. Characteristic of this faunizone at Deep Bay is an enormous profusion of smooth ostracods, Peruwispira-Ptychomphalin, and Megadesmus gryphoides (de Koninck).

Faunizones 7-8 Species confined to these faunizones include Terrakea concava Waterhouse, Fusispirifer sp. nov. (a very transverse form with obsolete ornament) and Martinioptis sp. nov. (a medium-sized, strongly plicate, transverse form with a deep groove on the dorsal fold). Important new species which enter near the base of Faunizone 7 include Aperispirifer lethamensis Waterhouse, Sulciplica phalaena (Dana)—transversa Waterhouse group, Terrakea brachythaera (Morris), Wallichollis subcancellata (Morris), Astartila intertrepid (Dana), Vacunella curvata (Morris), Myonia carinata (Morris), and Gilledia ulla-dulensis Campbell. New entrants in Faunizone 8 include Martinioptis strzeleckis (de Koninck), Warthia microphalma (Morris), Volsella mytiliformis (Etheridge) and possibly Etheripecten leniusculus (Dana). Eurydesma hobotense and Sulciplica stutchburi are abundant at the base of Faunizone 7 but rapidly decrease in numbers and disappear below its summit, whereas Deltopecten, Wyndhamia preovalis and W. jukesi-dalwoodensis remain abundant throughout both faunizones. Many species characteristic of Faunizones 4-5 (==Fauna II) or even older horizons, such as Schuchertella, Thanthropora, Sulciplica tasmaniensis, Grantonia cracovensis, Conocardium and Paraconularia derwentensis (Johnston), finally disappear at the top of Faunizone 8.

Faunizone 9 This faunizone marks the entrance of Martinioptis isbelli (Campbell), Fusispirifer avicula (Morris) and 'Notospiri­fer' duodecemcostatus (M'Coy), the acme of Martinioptis magna, together with the continued persistence but eventual extinction of Deltopecten, Etheripecten fittoni (Morris), Aperispirifer lethamensis and Wyndhamia jukesi-dalwoodensis. At Deep Bay Atomodesma (Aphanaia) occurs in profusion near the summit of the zone and Punctospirifer is abundant towards its base.

Faunizone 10 Wyndhamia ovalis and Megadesmus grandis (Dana) enter for the first time and distinguish this faunizone. No overlap of these species with Deltopecten and Wyndhamia jukesi-dalwoodensis occurs in Tasmania.

Many lineages, but particularly the martiniopsids, neospiriferids, Sulciplica and strophalosiids show gradual upward change. Speciation within these continuously variable lineages is therefore somewhat arbitrary. Doubtless, a more refined taxonomy beyond that currently available is necessary and would give the definition of each faunizone a greater distinctiveness. Such a process should be approached with caution since the proliferation of new taxa for minor morpho-
logical variations is to be deplored.

A detailed external correlation of Faunizones 6-10 poses several problems. Faunizone 10 with *Wyndhamia ovalis, Fusispirifer avicula* and *Megadesmus grandis* appears to offer a reasonably sure correlation with Fauna IVB-IVC in Queensland. Below this, firm correlations are less evident. Stratigraphic position and the occurrence of *Martiniopsis isbelli* with the peak development of *Martiniopsis magna* suggests a broad equivalence of Faunizone 9 with Fauna IVA. Together, Faunizones 9-10 offer a reasonably sound correlation with the New Zealand Flettian Stage. On stratigraphic position, general faunal characters, but particularly on the basis of the very distinctive *Terrakea concava*, Faunizones 7-8 may approximate in age to the New Zealand Barrettian Stage. Faunizone 6 is most certainly post-Fauna II on the basis of the martiniopsids, *Fletcherithrys parkesi* and *Myonia corrugata*, and may therefore approximate with Fauna IIIA. This conclusion is based largely on stratigraphic position since faunal resemblances are slight. Indeed, Fauna IIIA is largely distinguished by characteristic bivalves and some gastropods, most of which do not occur in Tasmania (*Veteranella, Atomodesma* cf. *mytiloides* Beyrich, *Wilkingia*, *Pseudomonotis* and *Platyteichum*). The long-ranging *Martiniopsis ingelarensis, Stutchburia* cf. *costata* (Morris) and *Streblopteria* are the only forms common to both assemblages. Equally Faunizone 6 is almost certainly pre-Barrettian and therefore presumably Mangapiran, assuming the completeness of the New Zealand sequence. However, once again faunal resemblances are slight; indeed, Faunizone 5 appears to approximate more closely with the Mangapiran.

At Deep Bay the base of Faunizone 6 is not exposed. However, the Deep Bay Borehole proved this faunizone to rest directly on the Lower Marine Sequence (Bundella Mudstone) about 35 m below the lowest exposed horizons. Faunal assemblages from the Bundella Mudstone unequivocally belong to Faunizone 2. Consequently, a pronounced hiatus or paraconformity involving Faunizone 3, the Lower Freshwater Sequence and Faunizones 4-5, separates the Upper and Lower Marine Sequences. In the borehole, and in other sections of the Cygnet area, there is little physical evidence of this hiatus apart from a few thin bands of conglomerate, flaser-bedded siltstone and some lenses of coarse gritty sandstone crowded with *Peruvispira-Ptychomphalina* immediately above the Bundella Mudstone. This is most surprising in view of the Hobart sequence. Whether Faunizone 3, the Lower Freshwater Sequence and Faunizones 4-5 were deposited, and then removed by subsequent erosion prior to Faunizone 6, or never deposited is unknown. To date, the Lower Freshwater Sequence is unknown south of Taroona where it is immediately followed by the Grange Mudstone (Banks, 1952) which yields rich Faunizone 4-5 assemblages. South of Taroona large-scale block faulting has caused this critical interval to be obscured by higher Permian and Triassic rocks intruded by thick sheets of dolerite. However, in the Whitewater Creek area, south of Kingston, *Cancriella farleyensis* occurs abundantly in gritty sandstone and siltstone. Elsewhere in Tasmania, this species is a reliable index for Faunizones 4-5, but the lithofacies at Whitewater Creek is quite foreign to the Cascades Group. This may indicate a facies change causing a southward thinning and disappearance of the Cascades Group. However, faulting obscures the relationship of the Whitewater Creek *Cancriella* horizon to the beds above and below, so it is impossible to be sure. It is to be hoped that as detailed mapping proceeds north from Cygnet, more definite evidence will come to light. Ultimately, stratigraphic drilling may have to be utilised. Away from the Cygnet area Faunizone 6 has so far only been recognised in sections in the Beaconsfield Quadrangle. As noted by Green (1959), the West Arm Group (Gee, 1971) is much reduced in thickness, contains several prominent bands of conglomerate, and may display several internal disconformities. At Middle Arm Faunizone 6 is developed in sandy limestone about 20 m above the Liffey Sandstone. Even allowing further possible disconformities within the West Arm Group, in all probability it rests with dis- or paraconformity on the Lower Freshwater Sequence. Faunizones 7-9 are more widely distribu-
In the Hobart area Malbina Member A (Banks and Read, 1962) is not richly fossiliferous, but persistent collecting of the outcrop at Mt Nassau has yielded assemblages diagnostic of Faunizone 8 (Clarke, 1971). It is therefore necessary to postulate a dis- or paraconformity between Malbina Member A and the Cascades Group (Fig. 33.4). The lithologic break is certainly an abrupt one.

It is interesting to note that coarse sandstone of Malbina Member A type is of Faunizone 7 age at Arcadian Siding in the Maydena area and at Bronte Park, whereas similar lithologies are associated with Faunizone 9 assemblages in the Cygnet area. Thus coarse sandstone of Malbina A type becomes progressively younger away from the Bronte Park-Maydena area towards the east and southeast.

On Maria Island, at Friendly Beaches and elsewhere in eastern and northeastern Tasmania, Faunizones 7-9 are often present in thin, much attenuated sequences of arkosic and glauconitic sandstone. At Beaconsfield, Faunizone 9 is developed in reasonably pure limestone and rests directly and with inferred paraconformity on Faunizone 6. In the Fisher Tunnel, Faunizone 8 rests with proven paraconformity on the Lower Freshwater Sequence (Clarke and Farmer, in press). This paraconformity is probably developed over the entire extensive Parmeener Super-Group outcrop of the Great Western Tiers and western and northwestern Tasmania, since Faunizone 8 is the oldest detectable horizon within the Upper Marine Sequence at Poatina, Golden Valley, Western Bluff and the Cradle Mountain area (Fig. 33.4). However, lack of outcrop or paucity of diagnostic faunas immediately above the Lower Freshwater Sequence precludes certainty.

Faunizone 10 has so far been proved only on Maria Island and in southern Tasmania. Assemblages diagnostic of this faunizone are first encountered immediately below the Risdon Sandstone in Malbina Member E at Mt Nassau (Banks and Read, 1962) and at a similar stratigraphic level at countless other localities in the Brighton, Hobart, Kingborough and Sorell Quadrangles. One of these is at Eaglehawk Neck which is the type locality for Fusispirifer avicula (Banks, 1971). Rich Faunizone 10 assemblages (Clarke, in press) are also known from over a dozen localities within the Ferntree Mudstone of southern Tasmania. Previously this formation was thought to be poorly fossiliferous and no external correlations were possible (Banks, 1962). However, the presence of Astrartila intrepida, Megadesmus grandis and Vacunella curvata at a horizon about 40 m below the summit of the formation at Blackmans Bay, suggested to Runnegar (1967) that it was not significantly younger than the Peawaddy Formation in Queensland (= Fauna IVB-IVC). However, Waterhouse (1969, 1970) recorded Martinops ansusulcata (Waterhouse), Dellopecten and Eurydesma from 'tillite' in the Ferntree Mudstone at Grasstree Hill, and favoured a correlation with the New Zealand Waititian Stage which he equates with the Tatarian Stage of the world standard. More recent work (Clarke, in press) supports the original conclusion of Runnegar. In essence, diagnostic faunas occur at three separate and restricted horizons within the Ferntree Mudstone. One horizon (about 2 m thick) occurs in the middle of the formation, a second (6-8 m thick) occurs about 40 m below its summit, and a third (about 3 m thick) occurs no more than 15 m below its summit. Both the Blackmans Bay locality and the Grasstree Hill locality belong to the second horizon, which is the most widespread. Individually the three horizons within the Ferntree Mudstone show some differences in faunal composition, but these differences can be easily explained by slight differences in depositional environment or bottom conditions in so far as these are reflected in the lithology. Collectively over thirty species are represented within the faunas and most, including Wyndhamia ovalis, Fusispirifer avicula and Megadesmus grandis, are known from Malbina Member E. The remaining species, which include Martinops magna, M. isbelli, M. globosa and Terrakea brachythaera, are known from even older horizons as well as Malbina Member E (Fig. 33.2). Tillite is not developed at Grasstree Hill, or anywhere else in the Ferntree Mudstone. Indeed, tillite is unknown in Tasmania above the basal Wyndyard Tillite and its correlates. Martinops ansusulcata,
the critical species of the Waiitian Stage, is based on inadequate material so that its recognition is problematical. The vertical distribution of *Deltopecten* and *Eurydesma* everywhere in Tasmania strongly suggests that neither is likely to be found in faunas from the Ferntree Mudstone.

The Upper Freshwater Sequence marks the re-establishment of non-marine sedimentation and marine conditions are never again established in Tasmania during the Permian Period. Lithologically the Upper Freshwater Sequence is more closely related to the upper (Triassic) parts of the Parmeener Super-Group, but since it is of Permian age it is briefly mentioned here. It has long been known that the Cygnet Coal Measures at Cygnet and correlative horizons such as the Jackey Formation at Poatina, contain a *Glossopteris* flora and *Dulhuntyispora* microflora (Balme, 1962, 1967) and are therefore of Permian age. However, recent drilling of the once productive Cygnet Coal Measures at Mt Cygnet demonstrates that the Cygnet Coal Measures and Barnets Member (Banks and Naqvi, 1967) of the Springs Sandstone (‘Triassic’) are the same lithostratigraphic unit. Clearly, much further work is required before any detailed understanding of the lithostratigraphic and biostratigraphic relationships of units close to the Permo-Triassic boundary in Tasmania can be expected.

**GENERAL FAUNAL OBSERVATIONS**

The marine Gondwanan *Eurydesma* Fauna in Tasmania is extremely rich in terms of numbers of individuals, but diversity is low. There are no warmer-water forms such as reef-building corals, fusulinids and goniatites. Diversity increases above the Lower Freshwater Sequence but even so, several genera of brachiopods, bivalves and gastropods, which occur in varying degrees of abundance in the Sydney and Bowen Basins, are rare (*Anidanthus*, *Cleiothyridina*, *Spirigerella*, *Punctospirifer*, *Plekonella*), or absent (*Lissochonetes*, *Neochoonetes*, *Filoconcha*, *Attenuatella*, *Martinia*, *Stenocisma*, *Psilocamarra*, terebratuloids other than *Flectherithyris* and *Gilledia*; *Australomya*, *Cypricardinia*, *Glendella*, *Veteranella*; and *Platyteichum*). Rather surprisingly and unlike Early Permian (Allandale) faunas in the Sydney Basin, brachiopods (particularly strophalosiids, neospiriferids and martiniopsids) dominate Tasmanian faunas from the first appearance of *Eurydesma*. Bryozoa occur in profusion at many levels. The sudden appearance and abundance of true productids in Faun zones 4-5 is noteworthy, the more so in view of the abundance of strophalosiids throughout the Lower Marine Sequence. This again contrasts with the Sydney Basin where true productids such as *Anidanthus* and *Cancrinella* are present in the Allandale Fauna, yet strophalosiids are unknown (Runnegar, pers. comm.). The persistence of *Eurydesma*, *Deltopecten* and *Keeneia* at much younger horizons than in Queensland and Western Australia, probably indicates the continued influence of cold-water conditions throughout most of the Permian in Tasmania.

**SUMMARY AND CONCLUSIONS**

Macro- and microfloral evidence proves the lowermost parts of the Parmeener Super-Group to be of Late Carboniferous age. Above these horizons rich Permian marine faunas separated by one thin horizon of coal measures and other non-marine rocks are developed. These faunas prove that the Tasmanian sequence, though much reduced in thickness, much affected by lateral changes in lithofacies, and with breaks of variable duration from place to place, is as complete as any in eastern Australia. In detail, these faunas belong almost exclusively to the Gondwana cold-water realm and probably represent the most extreme development of the marine *Eurydesma* fauna anywhere in the world. According to generally accepted world correlations (Dickins, 1968; Runnegar, 1969a) marine sedimentation ceased in Late Kungurian or Early Kazanian times, and thereafter non-marine conditions persisted until the end of the Triassic (Townrow, 1962). On the basis of the *Glossopteris* flora and *Dulhuntyispora* microfloras, the lowest parts of this non-marine sequence are of Permian age. However, no satisfactory method of recognising the Permo-Triassic boundary in Tasmania is known. This is equally true of the Permo-Carboniferous boundary.
ACKNOWLEDGMENTS

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REFERENCES


Correlations between Triassic continental vertebrate faunal assemblages are of
importance in the history of Gondwanaland but are difficult to make. Until recent
decades our knowledge was mainly confined to Early Triassic faunas dominated by
therapsids, and Late Triassic faunas in which dinosaurs were most prominent. We
now know, particularly in South America, a number of faunas of intermediate age,
which are not simply transitional in nature but have positive distinctive features, such
as an abundance of gomphodont cynodonts and rhynchosaurs, together with a
variety of thecodonts and a few early dinosaurs. Faunas of this type are presumably
of Anisian, Ladinian and very probably Carnian age.

Comparisons of reptilian faunas are potentially of great value in determining the
relationships of various continental areas during the assumed Mesozoic 'disintegration' of
Gondwanaland. Unfortunately vertebrate faunas are too few in number and too 'spotty'
in distribution to make them of value for much of the Jurassic and Cretaceous. However,
recent discoveries in the Triassic of the Gondwana continents give us a reasonable
sequence of events in the history of vertebrates of that period.

We have long known, mainly from the Beaufort Beds of the Karroo, Early Triassic
faunas dominated by therapsids, and Late Triassic faunas from the red-beds of (for
example) the upper Keuper and equivalents in North America and South Africa in which
therapsids had practically disappeared and archosauromorph reptiles, particularly prosauro-
pod dinosaurs, were dominant. Recent finds, particularly in South America and east
Africa, have furnished much data on intermediate stages.

Voluminous references are not needed for the discussion below, since much of the data
referred to was covered in a plenary paper by me (1972) for the last Gondwana symposium.
Bonaparte (1970) has given a succinct description of all South American Triassic
tetrapods, and the older materials from Africa were summarised by Haughton and
Brink (1954).

In general all adequately known Triassic faunas can be arranged in three faunal types,
which I characterised some years ago as successive A, B and C groups (Romer, 1966).
The A-type faunas are characteristically Scythian in age and are best known from
the Early Triassic Lystrosaurus and Cynognathus Zones of the Beaufort Beds of the
South African Karroo. The dicynodont Lystrosaurus is the most characteristic reptile
of the Lystrosaurus Zone; other therapsids, mainly primitive carnivorous cynodonts such
as Thrinaxodon (Galesaurus), early and primitive thecodonts such as Proterosuchus
(Chasmatosaurus) and early lepidosauromorphs, are also present. The Lystrosaurus Zone in
South Africa is succeeded by the terminal Cynognathus Zone of the Beaufort, which
with the Lystrosaurus Zone, is considered to be of Scythian age. Here the typical cynodont
is a large form, Kannemeyeria; of somewhat
advanced and specialised nature among carnivorous cynodonts is *Cynognathus*; and a notable advance is the presence of 'gomphodons'—cynodonts which have shifted to a herbivorous mode of life, with the development of a grinding 'molar' dentition. *Diademodon* and *Trirachodon* are typical *Cynognathus* Zone representatives of this group. Other reptile types are sparsely represented, but we may note a primitive thecodont type in *Erythrosuchus*, and somewhat of an advance over the primitive proterosuchian type in *Euparkeria*.

Faunas of A-type are also widespread in other Gondwana continental regions. Of especial interest is the presence of the *Lystrosaurus* fauna in Antarctica. The fossils brought back as a result of two expeditions include not only *Lystrosaurus* but also other materials now being studied by Colbert, which could be readily believed to have been collected from the Karroo deposits of South Africa. It seems beyond question that definite land connections existed in some fashion in the Early Triassic between Antarctica and South Africa. No *Lystrosaurus* Zone materials are known from South America, but from Argentina Bonaparte has described skulls which unquestionably belong to the two most familiar genera of the *Cynognathus* Zone, *Cynognathus* itself and the dicynodont *Kannemeyeria*. There is thus strong evidence indicating that in the Early Triassic South Africa was connected with South America as well as Antarctica. In India, too, an A-type fauna of South African type was present, for in the Panchet Formation there are numerous skulls of *Lystrosaurus* as well as remains of *Chasmatosaurus* (*Proterosuchus*). Higher in the sequence in the Godavari Valley of Peninsular India, there is the incompletely known fauna of the Yerrapalli Formation. It includes a dicynodont similar to *Kannemeyeria* and may be a *Cynognathus* Zone equivalent. It is possible that, with further exploration, the Rewan Formation of Queensland, Australia, may prove to be of the familiar A-type, but as yet the reptilian fauna of this Triassic formation is very incompletely known.

As regards the northern, 'Laurasian', continents, we know little of the reptilian faunas of North American and western Europe in the Early Triassic, since the North American Moenkopi and the European Bunter are essentially 'mud flat' deposits in which amphibians are abundant, but reptiles very rare. In Russia and eastern Asia, however, there are finds which tend to oppose any theory that the A-type fauna characterised above was confined to Gondwanaland. In the Early Triassic of Russia there are incompletely preserved remains of forms similar to the South African thecodonts *Proterosuchus* and *Erythrosuchus*. Young, and more recently Sun, have described reptiles from Shansi and Sinkiang provinces of China which include therapsids comparable to *Kannemeyeria*, to South African cynodonts, and to *Erythrosuchus*, but in all cases they are apparently generically distinct. It appears reasonable to believe that, as in the Permian, interchange between northern and southern supercontinents was possible. But since in most recent continental reconstructions eastern Asia is widely separated from Gondwanaland by a broad Tethys Sea, travel over a long migration route might be expected to result in generic changes. But in addition we find in China forms identified generically with *Lystrosaurus* Zone reptiles—*Proterosuchus* and most especially *Lystrosaurus*. The latter is also known from Indochina. Perhaps the A-type fauna was ubiquitous, north as well as south?

Until recently we knew little of B-type faunas, intermediate between Early and Late Triassic assemblages. This was because northern Middle Triassic deposits are mainly marine in origin, and although the South African Molteno Beds are rich in footprints, they have yielded little bone. Recent discoveries, however, have now provided excellent sequences of B-type faunas in South America. In western Argentina there is a conformable sequence of formations, Chañares-Los Rastros-Ischigualasto, the last followed by red-beds of Late Triassic age. The Los Rastros beds of Argentina contain almost no fossils, but the Santa Maria Formation of southern Brazil has furnished an abundant fauna intermediate in type between those of the Chañares and Ischigualasto beds. Hence a B-type sequence of faunas of Chañares-
Santa Maria-Ischigualasto seems well established.

The A-type fauna is mainly one of therapsids and the C-type of the Late Triassic is mainly one of archosaurs, including members of the basic archosaur order Thecodontia, and especially dinosaurs of thecodont descent. Hence, a priori, one would expect transitional B-type faunas to show a diminution of therapsids, an increase in archosaurs (including a variety of thecodons), and the beginnings of the rise of dinosaurs. This is true in considerable measure of these South American B-type assemblages. Carnivorous therapsids are distinctly fewer in this sequence; thecodons are throughout abundant and varied; and in the Santa Maria and Ischigualasto Formations dinosaurs appear, although few in numbers and limited in variety. But, interestingly, these B-type faunas are not merely negatively transitional, but show positive characters of their own—in a wealth of gomphodont cynodonts and of rhynchosaurs.

It was noted above that in the A-type faunas there had begun the development of a herbivorous 'gomphodont' side-line of cynodonts. In A-type faunas the gomphodonts present are of primitive types; in the South American B-type faunas they are members of a more advanced family, the Traversonodontidae. These gomphodonts are present in modest numbers in the Santa Maria beds, constitute a considerable fraction of the finds at Ischigualasto, and more than half of all specimens collected in the Chañares Formation.

Equally surprising is the abundance of rhynchosaurs. These are large herbivorous relatives of Sphenodon of New Zealand, and notable for a peculiar skull, a powerful parrot-like beak and a shearing type of dentition. Two rare primitive forms were present in the Karroo A-type fauna, and a few later forms have long been known from England and India, but it has been assumed that they amounted to little in the fossil record. They do not appear in the Chañares Formation, probably due to ecological factors (their diet appears to have been a specialised one). But in the Santa Maria beds they are abundant (nearly all specimens collected in the exposures near Santa Maria city are rhynchosaurs) and at Ischigualasto almost all specimens collected from the lower half of these very productive beds are rhynchosaurs. The time of deposition of B-type beds seems to have witnessed an explosive evolution of gomphodonts and rhynchosaurs, which were taking the place of the dicynodonts, the dominant herbivores of Late Permian times.

When did B-type faunas flourish? We have no direct means of correlation between these continental deposits and the marine beds on which the standard Triassic sequence is based. It is generally agreed that the Cynognathus beds of A-type are to be equated with the upper part of the marine Scythian Stage. The B-type faunas certainly cover Anisian and Ladinian time, and the sequence Chañares-Santa Maria-Ischigualasto has led a majority of workers to equate this series with the Anisian-Ladinian-Carnian marine stages. I have been inclined to question inclusion of the Carnian as part of the B-stage, since (among other reasons) this would result in the faunal 'Middle' Triassic B-stage including part of the stratigraphically Upper Triassic sequences. However, it will be recognised that in the Upper Triassic Keuper beds of Europe the characteristic C-type reptile fauna is found only in the upper, Norian, portion and reptiles are absent in the lower, Carnian, portion. I now incline to the view that the majority opinion is correct, and that faunally the 'Middle' Triassic includes Carnian as well as Anisian and Ladinian, and that faunally the C-type 'Upper' Triassic is confined to the Norian and Rhaetic.

If we compare this now well-established B-type sequence of South America with other regions, we find that very similar faunas are present in Africa. The Ntaware Beds of Zambia perhaps represent an early B-stage. The Manda Beds of East Africa are conspicuous. In these beds, as the result of the work of Huene and of Parrington and his students, a reptile fauna of nearly a score of genera, which in its general character is quite certainly of a B-type, has been found. Unfortunately many of the forms present are known only from incomplete material and many have been described only in brief pre-
liminary fashion. It is probable, however, that comparison should be made with Chañares fauna rather than those of Santa Maria and Ischigualasto. The rhynchosaur present in the Manda, *Stenaulorhynchus*, is more primitive than *Scaphonyx* of Santa Maria and Ischigualasto, and the cynodonts and dicynodonts present include 'holdovers' of types present in the A-type faunas and not present in the South American B-type. I strongly suspect that when the Manda materials are fully described, certain of the forms present will prove to be generically identical with Chañares reptiles.

No B-type faunas are as yet known from Australia or Antarctica. In the Triassic of India the Maleri Beds follow the Panchet and Yanapalli Beds. Apart from some thecodont scutes, only four tetrapods are known from them—a metoposaurid amphibian, a phytosaur, an unnamed coelurosaurian dinosaur and a rhynchosaur, *Paradapedon*. The Maleri is generally considered to be an Upper Triassic formation. Metoposaurs are Carnian and Norian. Phytosaurs, as far as known, are Late Triassic in some sense of that term, but obviously this specialised type must have had its beginnings earlier than the Norian (even if the possible Early Triassic date of *Mesorhinus* be left out of consideration). The Maleri rhynchosaur is comparable in degree of advancement to *Scaphonyx* of Ischigualasto and Santa Maria. While dinosaurs become abundant only in the C-type faunas, they appeared in B-type faunas at Santa Maria and Ischigualasto. All in all, an interpretation of the Maleri as an advanced B-type fauna of possibly Carnian age is reasonable.

The presence of *Metoposaurus* and a phytosaur in the Maleri is of geographic interest. These are typically northern continent forms; a metoposaur is reported from Morocco but there appears to be no trace of these readily identifiable forms in any other southern continental area at any Triassic stage. Scutes possibly (but far from certainly) identifiable as phytosaurian are reported from Madagascar, but of these crocodile-like thecodonts, very abundant in Europe and North America, there are no other traces in the southern continents. It seems obvious that there was some readily traversable connection between India and Laurasia at the time of deposition of the Maleri Beds.

As we have noted, there is little evidence of terrestrial reptiles in northern continental areas identifiable as Middle Triassic on stratigraphic grounds. From Middle Triassic beds in southern Switzerland, studied by Peyer and by Kuhn-Schnyder, a few shore dwellers have been reported, together with numerous aquatic types. In England, sandstones formerly assigned to the Upper Triassic Keuper now appear to be Middle Triassic, and the presence in them of a relatively primitive rhynchosaur suggests that if a representative faunal assemblage should ever be discovered it would prove to be of the B-type. Two other northern localities may well represent a terminal B-stage of probable Carnian age.

The interesting Elgin deposits of Scotland are being thoroughly re-studied by Walker. The fauna is sparse, but includes the aetosaurid thecodont *Stagonolepsis*, an advanced rhynchosaur *Hyperodapedon*, comparable in general to *Parasuchus* of the Maleri and *Scaphonyx* of South America, and *Ornithosuchus*, which has been variously interpreted as a primitive saurischian or as a thecodont paralleling early saurischians in various regards. Walker (1961) believes the Elgin beds to be comparable to the Norian, C-type, Keuper of the continent. But no rhynchosaur has ever been found in the continental Keuper; the aetosaurid pattern was already established in the B-type fauna of Ischigualasto, and (perhaps due to the limited range of the fauna) the Elgin beds show none of the most characteristic reptiles of the C-type beds of the northern hemisphere, such as prosauropod dinosaurs and phytosaurs. It seems reasonable to believe that here, as in the Maleri, we have a fauna of late B-type, probably Carnian in age.

In Nova Scotia there are Triassic red-beds generally considered to be equivalent to the Newark Series of the Atlantic coast of the United States farther south. Much, if not all, of the true Newark sequence is obviously Late Triassic, with a C-type fauna. But in the lowest part of the Nova Scotia beds there have been found (Baird, 1962 and pers.
comm.) numerous, if incomplete, specimens of a rhynchosaur, and a single gomphodont jaw. Are we dealing here with an exceptionally late survival of rhynchosaurs and gomphodonts? More probably, I think, the deposition of these Nova Scotia beds may have begun in Carnian times, and that here, as I believe true of Elgin and Maleri, we have a final appearance of B-type faunas.

Typical C-type faunas of my classification are to be found, for example, in the upper, Norian, Keuper of continental Europe and in equivalent Late Triassic red-beds of such other areas as the Red Beds and Cave Sandstone of South Africa, the Los Colorados Formation of Argentina, the Newark Series of the eastern United States, and the Lufeng Series of China. Most characteristic is the abundance of prosauropod saurischian dinosaurs, such as Plateosaurus. Thecdonts persist and primitive crocodilians have made their appearance. Therapsids are almost extinct, except for the specialised little tritylodonts, but small early mammals are making their appearance. Varied long-snouted phytosaurs are abundant in the north, but not in South Africa or Argentina.

A feature of the various C-type faunas is that there appear not to be any marked regional distinctions between those of the varied areas concerned. The abundant prosauropods, present in every area, belong to a number of genera, but the general pattern, from South America and South Africa to Europe, China and eastern North America, is essentially the same; probably careful comparative study would result in the identification of the same genera in areas widely separated geographically. Even the tritylodonts, once thought to be very rare, are now known from Europe, western North America, China and Argentina, as well as in South Africa, the original 'home' of Tritylodon. One gains the impression that connections between continents were readily available throughout the world. We know, however, nothing of this stage in Antarctica or Australia. Except for the non-appearance of phytosaurs in South Africa and South America, there is no marked contrast between the reptile faunas of Gondwanaland and Laurasia.

A minor problem concerns the age of the red-beds of the western United States—Chinle, Dockum, Popo Agie. These formations are surely of rather Late Triassic age and the wealth of phytosaurs (and of metoposaurs) tends to tie them in with faunas which are definitely of C-type. But we do not find the abundant dinosaurs, notably prosauropods, found in typical C-type assemblages. Of dinosaurs there are only a few scraps, apart from a fortunate find in a quarry of skeletons of the coelurosaur Coelophysis. Can these formations be of slightly earlier age than the typical C-type beds; or, equally probably, is this dearth of dinosaurs due to the fact that we are dealing with swampy areas unsuitable for them?

In summary, the attempt to work out Triassic reptile faunas on the A, B, C basis appears to give a reasonable overall solution to problems of the faunal sequence, but the Middle Triassic fauna extends onward into deposits of Carnian age which are by definition Upper Triassic. A certain amount of differentiation can be seen between Laurasian and Gondwanaland regions, but it seems clear that (as was true of the preceding Permian) a fairly free interchange took place between north and south.

REFERENCES
Affinities and Systematic Position of the South African Lower Middle Permian Dicynodonts (Therapsida : Dicynodontidae)

T. H. BARRY

ABSTRACT

Comparative osteological studies of a new ancestral dicynodont, *Eodicynodon oosthuizeni*, found in layers which predate the Middle Permian *Tapinocephalus* zone of the South African Beaufort Series, show the retention in this form of a series of pelycosaurian features which have disappeared in its *Tapinocephalus* zone descendants. This paper deals with the affinities and systematic position of this form.

INTRODUCTION

The therapsid fauna of the Upper Middle Permian *Tapinocephalus* Zone of the Beaufort Series of South Africa is generally considered to be the oldest terrestrial reptile fauna of all the Gondwana continents. To date comparable therapsid faunas have only been found outside Gondwanaland. The most comprehensive of these is from the Russian Middle Permian Zones I and II (see Table 35.1). Zone II is considered to be approximately equivalent to the South African *Tapinocephalus* Zone. Boonstra (1963) was of the opinion that four of the therapsid families from these deposits are on the same morphological level as their *Tapinocephalus* Zone counterparts but that five therapsid families are more primitive. The discovery of therapsids in the lower Middle Permian Ecca Series of South Africa (Barry, 1970) and the subsequent description of *Eodicynodon oosthuizeni* (Barry, 1973) have provided evidence of terrestrial forms which in time and evolutionary development predate the *Tapinocephalus* Zone fauna.

*Eodicynodon* is the only dicynodont that has been discovered in sediments of Ecca age. It provides the basis for the deductions made in this paper. As this form is considered to be directly ancestral to the *Tapinocephalus* Zone dicynodonts it will be compared with those forms in the Russian deposits which are also claimed to lie in the ancestry of the dicynodonts, viz. *Otsheria* (Chudinov, 1960) and *Venjukovia* (Efremov, 1940a). *Dimacrodon* from the Middle Permian San Angelo deposits of North America, which Olson (1971) believes to be vaguely related to venjukovoids, has not been included in this study as it is only known from poor material.

AFFINITIES AND SYSTEMATIC POSITION

Apart from the well-known features such as the tusk-like canines, the beak-like upper and lower jaws, the absence of post-canine and dentary teeth and the horizontally turned squamosal bar which characterise the more than one hundred species constituting the genus *Dicynodon*, all known dicynodonts display single, or fused premaxillaries and fused vomers. These features set dicynodonts apart from all other therapsids with the exception of the Gorgonopsia in which the vomers have become fused also. Amongst *Tapinocephalus* Zone therapsids too the development of a secondary palate is found only in the dicynodonts. However, in contrast with the later developing cynodont type of secondary palate, which is predominantly formed by the maxillaries and palatines, the dicynodont one features the extensive involvement of the premaxillaries only. *Eodicynodon* shows an in-
Table 35.1. Approximate Correlations of the most important Permian Continental Vertebrate Beds (adapted from Romer, 1969)

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<th>North America</th>
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<td>Endothiodon Zone</td>
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<td>Middle Permian</td>
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recipient secondary palate. *Otsheria* and *Venjukovia* do not seem to show any significant development along this line, which could indicate that the secondary palate developed fairly late in dicynodont history.

Some of the primitive features found in *Eodicyodon*, such as the retention of paired vomers, the retention of the lateral process of the pterygoid and the reduction of the anterior process of the pterygoid have proved to be the most significant indicators of the evolutionary level this form occupies. It also afforded valuable supporting evidence for the view that dicynodonts and dinocephalians were closely related ancestrally.

The paired vomers of the ancestral Pelycosauria are retained in all *Tapinocephalus* Zone therapsids with the exception of the Dicynodontia and the Gorgonopsia. In these groups the bones have become fused to form a single median bone. In the Dicynodontia, however, the form of the posterior end of the vomer indicates its originally paired condition. As the vomers are paired in *Eodicyodon* and in *Venjukovia* the fusion of the vomers would seem to have occurred fairly late in the evolutionary development of the dicynodonts. If Chudinov’s (1960) illustration, showing a single vomer in *Otsheria*, is correct, it would show a far earlier development of this feature but there is no evidence to support this assumption.

The dicynodont pterygoid would seem to be the palatal bone most altered in its evolution from the pelycosaurian stage of development. In the Pelycosauria the pterygoid has a well-developed anterior process which extends far forward in the palate and meets its mate over a considerable distance in the midline. Anteriorly this process is in contact with the vomer. Functionally it must have played an important role in mastication as it is studded with small teeth. Posteriorly, and roughly in line with the lateral processes of the pterygoids, the pterygoids separate in the midline to enclose the interpterygoidal vacuity (see Fig. 35.1).

Comparison between the Pelycosauria and the *Tapinocephalus* Zone therapsids shows that while the anterior process shows slight reductions in length in the Therocephalia, it has become significantly reduced in the Dinocephalia, Dicynodontia and the Gorgonopsia. Amongst Dinocephalia the anterior process has been so much reduced in the genus *Tapinocephalus* that the process hardly projects beyond the anterior limits of the lateral processes. In the majority of Dicynodontia the anterior processes are reduced to two thin projections which do not meet in the midline.

As could be expected, the drastic reduction of the anterior process led to considerable changes in the shape and relationship of surrounding palatal bones. In lines leading towards the dicynodonts the median position vacated by the shrinking anterior process was taken up by the posteriorly extending vomer, so that the vomer-ptyerygoid contact was retained during these changes. This would seem to be the case also in the majority of dinocephalians but in this group there is a general tendency towards a pincer-like ex-
pansion of the posterior portion of the palatine resulting in the vomer being cut off from its contact with the anterior process of the pterygoid (see Fig. 35.1). Culmination is seen in the Russian forms *Eotitanosuchus* and *Syodon*, the palatines meeting in the midline. Development towards this condition is well advanced in forms such as *Tapinocephalus* and *Anteosaurus*. Gorgonopsians show the same type of development, the palatines taking up the position vacated by the receding anterior processes, but here it would seem to be related to a general median expansion of the palatines and not only the

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Fig. 35.1. Morphological series of skulls in ventral view, brought to about the same basal length A *Dimetrodon*. B *Anteosaurus*. C *Tapinocephalus*. D *Titanophoneus*. E *Ulemosaurus*. F *Dicynodon*. G *Eodicynodon*. H *Venjukovia*. I *Otsheria*. ap anterior process of the pterygoid; ipv interpterygoidal vacuity; lp lateral process of the pterygoid; pal palatine; vo vomer.
Amongst _Tapinocephalus_ Zone therapsids the Therocephalians show the least changes from the pelycosaurian condition. Only moderate reduction of the anterior processes of the pterygoid has occurred.

One of the most significant end results of the reduction of the anterior process is in the changes brought about in the interpterygoidal vacuity. In the pelycosaurs the interpterygoidal vacuity is, as its name implies, situated between the pterygoids and its boundaries are formed by pterygoidal elements. In the dicynodonts, however, the reduction of the anterior process has resulted in the vomers moving back right to the rim of the interpterygoidal vacuity where their posterior ends now form the anterior and antero-lateral borders of the vacuity. As vomeral elements participate in the formation of its boundaries the dicynodont interpterygoidal vacuity technically becomes an intervomero-interpterygoidal vacuity. In many _Tapinocephalus_ Zone dicynodonts the vomers form more than one half of the borders of the intervomero-interpterygoidal vacuity.

In _Eodicynodon_ the posterior portions of the vomers are already flared and an intervomero-interpterygoidal vacuity is present. This form would seem to represent an intermediate stage in that the thin anterior processes of the pterygoid extend slightly beyond the anterior limits of the lateral processes and the vomers form only the anterior third of the vacuity. In both _Othishria_ and _Venjukovia_ the vomers are still well removed from the interpterygoidal vacuity.

In the majority of Dinoccephalia the vomers end very close to the interpterygoidal vacuity but the vacuity remains intact. In _Jonkeria haughtoni_ (Boonstra, 1962), however, the posterior portions of the vomers actually form the anterior border of the interpterygoidal vacuity but the vomers are not forked. In _Tapinocephalus atherstoni_, Boonstra (1956: 145) found that 'the interpterygoidal vacuity is a fairly narrow slit just extending anteriorly to between the posterior ends of the prevomers'.

The Russian dinoccephalian _Titanophoneus_ shows an interesting condition in that an intervomero-interpterygoidal as well as an interpterygoidal vacuity is present, the latter posterior to the former. As the posterior portion of the intervomero-interpterygoidal vacuity lies between the anterior processes of the pterygoid the condition here would seem to represent a stage before the final link-up with the interpterygoidal vacuity. The final stage would seem to be represented in _Ulemosaurus_ (Efremov, 1940b) where a fully developed intervomero-interpterygoidal vacuity is represented. It would seem, therefore, that in this respect the Dinoccephalia and Dicynodontia follow a closely linked evolutionary pattern.

**Conclusions**

_Eodicynodon_ displays many general dicynodont features found in _Tapinocephalus_ Zone dicynodonts, which would seem to place it in the direct ancestry of at least some of the _Tapinocephalus_ Zone dicynodonts. However, some of these features indicate that this form occupies a position further back in the dicynodont history than its stratigraphical position would suggest. It is probable that forms transitional between _Eodicynodon_ and the _Tapinocephalus_ Zone dicynodonts must have existed and that these may yet be discovered in the Upper Ecca-Lower Beaufort contact zone, the lithology of which indicates a slow rather than a dramatic change in climatic conditions.

There still remains a far bigger gap in our knowledge of the evolutionary record between _Eodicynodon_ and the pelycosaur ancestors, but the evidence given above would seem to indicate that the development of certain specifically dicynodont features occurred further back in the ancestry of the dicynodonts than has been generally thought. One of the most surprising is that the typical two-tusked dicynodont condition is already developed in _Eodicynodon_, which still shows many primitive features.

The absence of post canine teeth in _Eodicynodon_ shows too that this form cannot be on the direct line leading to the _Tapinocephalus_ Zone endothiodonts. Thus Toerien's (1953) view that the _Tapinocephalus_ Zone endothiodont _Broilius_, which still displays post canine teeth, could be ancestral to dicynodonts is not acceptable.
ACKNOWLEDGMENTS

I am indebted to Mr Roy Oosthuizen of Zwartskraal, Prince Albert for the loan of this and other specimens, and for his continuing efforts to find additional material. I also wish to thank the South African Council for Scientific and Industrial Research and the Board of Trustees of the South African Museum for financial assistance to attend the 3rd Gondwana Symposium.

REFERENCES


ABSTRACT
Conodont faunas from the shallow marine shelf along the northern edge of Gondwanaland are shown to range through the whole Triassic. The status of previously defined Triassic biostratigraphic units based on conodonts is reviewed, and the conclusion is reached that the establishment of a zonal scheme for this interval on the Gondwanaland shelf would be premature because of the inadequate stratigraphical control on available samples. Comparison of microfloral and conodont biostratigraphy is shown to give added precision to the dating of sequences in terms of the international time scale.

INTRODUCTION
The partial reconstruction of Gondwanaland given in Figure 36.1 shows the known positions of Triassic conodont localities. Most of them are on its northern margin. The reconstruction is based principally on that of Smith and Hallam (1971). Additions rely mainly on information from Griffiths (1971) and Griffiths and Varne (1972) for reconstruction of the Tasman region, and palaeogeographical evidence for western Tethys (Hallam, 1971) and Timor (Audley-Charles, 1968).

Southeastern Spain has been included because of its close palaeogeographical links with the western part of the North African margin of Gondwanaland during the Triassic (see Hallam, 1971). The position of the boundary between the Palaeozoic and Mesozoic geosynclines along the southern margin of Gondwanaland has been added as a rough guide to prospective areas for recovery of additional Triassic conodont faunas.

DISTRIBUTION OF CONODONT FAUNAS
Most conodont faunas of Triassic age found to date occur along the southern Tethyan part of the Gondwanaland Shelf, as shown in Figure 36.1. A summary of the ranges of these faunas in terms of Triassic stage and substage nomenclature defined by Silberling and Tozer (1968) is given in Figure 36.2. It does not include a recently discovered Upper Triassic fauna from New Guinea.

Lower Triassic Faunas
The most important conodont successions for Lower Triassic conodont stratigraphy are those in the Salt Range and Trans-Indus Ranges of Pakistan, where Sweet (1970b) has been able to distinguish nine conodont zones of Early Triassic age (Fig. 36.3). These sequences are well-dated by ammonoids and represent the entire Early Triassic.

The zonation consists of acrozones, teilzones, and acmezones of differing biostratigraphical value, but it seems more readily applicable to Gondwanan faunas than the more detailed composite zonation of Sweet et al. (1971), which is based on the worldwide distribution of Lower Triassic conodonts.

Preliminary conodont evidence (Taraz, pers. comm.) from the apparently continuous Lower Triassic section at Abadeh suggests that the Iranian faunas, when described in detail, could be as important to Lower Tri-
Fig. 36.1. Generalised partial reconstruction of Gondwanaland for the Triassic showing the position of the Triassic conodont localities on the Gondwanaland Shelf.

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Fig. 36.2. Ranges of key Triassic conodont species on the Gondwanaland Shelf.
Fig. 36.3. Correlation of Lower Triassic conodont faunas of the Gondwanaland Shelf with Sweet's (1970b) zones of the Lower Triassic conodont succession of Pakistan.

assic conodont stratigraphy as those of the Pakistani succession.

The Pakistani zones were described adequately when proposed by Sweet (1970b) and Sweet et al. (1971) repeated the definitions of those zones which they included in their more expansive Triassic conodont zonation. The following comments express my views in the light of subsequent studies.

The Anchignathodus typicalis Zone is a valuable, widespread acrozone that spans the Permian-Triassic boundary. It has been recorded on the Gondwanaland Shelf from West Pakistan (Sweet, 1970b), Kashmir (Sweet, 1970a), and Iran (Sweet et al., 1971, Taraz, pers. comm.). Comparison of the ranges of Anchignathodus typicalis and Neogondolella carinata in occurrences of the Anchignathodus typicalis Zone in these areas indicates that the Abadeh section may be continuous from the Permian into the Triassic, whereas the Pakistan and Kashmir sequences are interrupted at the Permian/Triassic boundary as suggested earlier on macrofaunal evidence by Kummel and Teichert (1970) and McTavish and Dickins (in press) respectively.

Sweet et al. (1971) implied that the N. carinata Zone was represented in the ‘Claraia’ beds of northwestern Iran. However, evidence from the Abadeh section of central Iran (Taraz, 1969, 1971, pers. comm.) for the age of the ‘Claraia’ beds and the range of N. carinata, suggests that the presence of the N. carinata Zone in Iran has not yet been established.

The absence of the Neospathodus kummeli Zone in the closely sampled Guryul Ravine section, Kashmir (Sweet, 1970a) suggests that acceptance of this zone should be deferred until its biostratigraphical value is established by further records.

At first sight, the Neospathodus dieneri Zone seems to be a rather unsatisfactory biostratigraphic unit which Sweet (1970b) suggested might be of use only in Pakistan. However, recognition of the zone in Western Australia (McTavish, 1973) in close association with ammonoids helps to demonstrate its value for intercontinental correlation.

The Neospathodus cristagalli Zone (Sweet, 1970a, 1970b) is an unsatisfactory unit, largely because of confusion in the identification of N. cristagalli (Huckriede) and uncertainty about the relationships of this species to N. discretus (Müller) and N. conservatius (Müller). Unillustrated reports of N. cristagalli from Guryul Ravine, Kashmir (Nakazawa et al., 1970) and Nepal (Fuchs and Mostler, 1969), although well-dated, are anomalous and require re-examination.

The Neospathodus pakistanensis Zone is an extremely valuable acrozone. It is well-dated, and has a wide extent and short range. In addition to occurrences in Pakistan and Indonesian Timor (Sweet, 1970b; Sweet et al., 1971), the zone is represented also in Western Australia (McTavish, 1973). Israeli faunas of Dienerian (?) age (Hirsch, 1973) cannot be related to the Pakistani zonation.

Most difficulties of Lower Triassic conodont correlation are found with Smithian faunas. These are well-dated, widespread, and more variable than those of other Lower Triassic substages on the Gondwanaland Shelf. Sweet (1970b) defined the Neospathodus waageni Zone for Smithian conodont faunas of Pakistan and Indonesian Timor. It is also present in Western Australia (McTavish, 1973).

In a more ambitious worldwide zonation of conodont faunas dated as Smithian, three
zones of questionable value were proposed by Sweet et al. (1971). Faunas on the Gondwanaland Shelf representative of these zones were reported by these authors. They recognised the lowest zone, the *Furnishius-Parachirognathus* Zone in Kashmir and Malaya and suggested (p. 452) that, 'there may be a facies relationship between *Furnishius-* and *Parachirognathus*-bearing strata'. Also, they admitted that they could not be certain that the *Neospathodus pakistanensis* Zone was directly succeeded by the *Furnishius-Parachirognathus* Zone.

The *Neogondolella milleri* Zone was shown to succeed the *Neospathodus conservatius* Zone, but both guide-species have been found together in several well-dated Smithian samples. The zone assignment of a particular sample seems to depend, simply, on which of these guide-species is predominant. Furthermore, both species have been found with *Neospathodus waageni* in Western Australia (McTavish, 1973). Clearly, more detailed study of the stratigraphical relationships of *Neospathodus conservatius-* and *Neogondolella milleri*-dominated faunas to one another, and to faunas bearing mostly *Neospathodus waageni*, might show that Smithian zones named after these species are in fact coetaneous biofacies. Smithian conodont faunas, additional to those recorded above, have been reported from the Gondwanaland Shelf of Malaya (Igo et al., 1965; Nogami, 1968; Sweet, et al., 1971), Nepal (Fuchs and Mostler, 1969), Portuguese Timor (Nogami, 1968; Sweet et al., 1971), Afghanistan (Sweet, 1970b; Sweet et al., 1971), and Israel (Hirsch, 1973).

Spathian conodont faunas of the Gondwanaland Shelf represent two zones which were first described from Pakistan by Sweet (1970b). Israeli faunas of Spathian (?) age (Hirsch, 1973) has not been referred to zones.

The older, *Neogondolella jubata* Zone (Sweet, 1970b), is not synonymous with the *N. jubata* Zone of North America (Sweet et al., 1971), which is a teiulzone at the uppermost part of the range of *N. jubata*. The former zone was first reported from Pakistan, and it is represented in Western Australia (McTavish, 1973). The precise stratigraphical relationship of the *Platyvillosus* Zone (Sweet et al., 1971) to Smithian faunas has not been determined and its correlation with the lowermost Spathian fauna of Pakistan has not been established. Müller (pers. comm.) has recognised *Platyvillosus* in material from Afghanistan.

Nogami (1968) was the first to recognise the biostratigraphical significance of *Neospathodus timorensis*, in a collection from Portuguese Timor. However, the *N. timorensis* Zone was proposed by Sweet (1970b) for the occurrence of *N. timorensis* in Pakistan. Additional occurrences of this zone are from Nepal (Fuchs and Mostler, 1969; McTavish, 1973) and Western Australia (McTavish, 1973).

Dating of the *N. timorensis* Zone has been inconsistent (Nogami, 1968; Nakazawa and Bando, 1968; Fuchs and Mostler, 1969; Sweet, 1970b; Sweet et al., 1971; and McTavish, 1973). However, review of ammonoid, stratigraphical and conodont evidence for the age of the *N. timorensis* Zone (McTavish 1973) has indicated that this zone is probably of late Spathian age where *N. timorensis* and *N. homeri* are found together, but *N. timorensis* itself may persist into the early Anisian, as at Chios (Bender, 1967).

### Middle Triassic Faunas

Few conodont faunas of definite Middle Triassic age have been reported from the Gondwanaland Shelf. The one from Western Australia, dated tentatively as Middle Triassic, is now regarded as late Scythian in age (McTavish, 1970, 1973). Those from the Niti Pass area, India (Misra, Sahni and Chhabra, 1972) characterised by *Neogondolella sp. cf. N. constricta, N. mombergensis* and *Gondolella navicula*, may belong to the Middle Triassic *Neogondolella mombergensis* association of Nogami (1968), but precise dating is not possible.

Conodonts of the *Neogondolella mombergensis* association (Nogami, 1968) were found with an ammonoid fauna in a river boulder from Pualaca, Portuguese Timor, dated as lowest Anisian (Nakazawa and Bando, 1968). Both Anisian and upper Scythian ammonoids and conodonts were recorded from this sample, so the boulder may...
contain a mixed or condensed fauna.

The Gladigondolella tethydis association of Nogami (1968) is a widespread conodont fauna on the eastern part of the Gondwanaland Shelf. This association, recognised by the predominance of elements of *Gondolella navicula* Huckriede and *Gladigondolella tethydis* (Huckriede), has been generally regarded as younger than the Alpine *Gladigondolella tethydis* fauna of Mosher (1968a). However, because both guide species have long ranges, dating of this association on its own evidence cannot be precise. Sweet et al. (1971) proposed an Anisian to Ladinian, and tentative early Carnian, range for *G. tethydis*. *G. navicula* ranges from the upper part of the Anisian into the Norian (Nogami, 1968; Mosher, 1968a).

The Ladinian age proposed by Nogami (1968) for the occurrence of the *G. tethydis* association at Bakulnassi, Indonesian Timor, seems acceptable on the macrofaunal evidence of Welter (1922: 158). A similar age seems probable for the Malaysian occurrences of the association in the Kodiang Limestone of northwestern Malaya. These had been dated, with reservation, as Late Triassic by Nogami (1968). In the Bukit Kalong section, the *G. tethydis* association is followed upwards by an association characterised by *Gladigondolella malayensis* Nogami. Nearby, in the Bukit Kechil section, *G. tethydis* is found in the lower samples with *Pseudofurnishius murcianus* (lowest sample) and *G. malayensis* (about 5 m higher). The presence of *P. murcianus* indicates that the occurrence of *G. tethydis* in the Bukit Kechil section is of late Ladinian age.

The occurrence of *G. tethydis* with *Neo­spathodus timorensis* (Nogami, 1968) and upper Spathian ammonoids (Nakazawa and Bando, 1968) requires additional study before the range of *G. tethydis* can be extended. Other occurrences of *G. tethydis* on the Gondwanaland Shelf point to a Ladinian, and possibly early Carnian, age for the *Gladigondolella tethydis* association. The stratigraphically younger *G. malayensis* association may be of late Ladinian to early Carnian age.

The presence of *Pseudofurnishius murcianus* in Malaya (Nogami, 1968) marks the easternmost occurrence of the Tethyan *Epigondolella mungoensis-P. murcianus* fauna. The occurrence of *P. murcianus* in southeastern Spain (van den Boogaard, 1966; Simon, 1966; van den Boogaard and Simon, 1973) accurately dates the beginning of the late Middle Triassic transgression recorded by Hallam (1971) at the western end of Tethys. In Israel, *P. murcianus* is found in association with *Epigondolella mungoensis* (Diebel) (Huddle, 1970; Hirsch, 1973). A late Ladinian age has been interpreted for *E. mungoensis* (Mosher, 1968a; Sweet et al., 1971). The first Triassic conodonts discovered on the Sinai Peninsula (Eicher, 1946, 1947), about 30 km southwest of Huddle’s locality, were not figured or described; however, an Upper Muschelkalk dating of the associated macrofauna suggests that conodonts of the *Epigondolella mungoensis-Pseudofurnishius murcianus* fauna may have been the first from the Gondwanaland Shelf to be recognised. The *E. mungoensis* fauna was first recovered from a well-dated, ammonoid-bearing sample of Turonian (Late Cretaceous) age from the Cameroons (Diebel, 1956). This occurrence, which has stimulated much interest and posed many problems, will be considered more fully in the discussion of post-Norian conodont faunas.

### Upper Triassic Faunas

Most Upper Triassic conodont faunas from Malaysia and Portuguese Timor were referred to an *Epigondolella abneptis* association by Nogami (1968). Upper Triassic conodonts have been recognised also in collections from Indonesian Timor (van den Boogaard, pers. comm.). The youngest Triassic faunas have been recovered from India, New Guinea, and New Zealand.

The latest Carnian to middle and possibly late Norian range of *Epigondolella abneptis* (Sweet et al., 1971: fig. 3) is consistent with that of the *E. abneptis* association of Nogami (1968), as recognised herein. However, Nogami’s (1968) identification of *E. abneptis* (Huckriede) seems to be broader than that generally accepted (Mosher, 1968a, b; Sweet et al., 1971, this paper) so the number and range of records of the *E. ab-
neptis association in the Timor-Malaysia region must be reduced.

Among specimens from the uppermost Carnian to Norian of Portuguese Timor that Nogami (1968: pl. 8, figs. 1-7) referred to *E. abneptis*, one (Nogami, 1968: pl. 8, fig. 8), dated questionably as Norian, seems closer to *Paragondolella polygnathiformis*, a species suggestive of a Carnian, probably late Carnian, age. A similar age seems probable for the uppermost sample from the Bukit Kalong section, Malaya, which contains specimens (Nogami, 1968: 122, pl. 8, figs. 9-11) intermediate in form between late Ladinian *Epigondolella mungoensis* (Diebel) and *E. abneptis*, but closer to *E. abneptis*. A sample from Manatuto, Portuguese Timor, which contained *Epigondolella abneptis* and *Daonella indica*, was dated as Ladinian-early Carnian (Nogami, 1968). This dating seems acceptable on macrofaunal evidence, but it casts doubt on the validity of the identification of *E. abneptis*.

The youngest Triassic conodont faunas from the Gondwanaland Shelf are of late Norian age. In New Zealand, conodonts have been found with the late Norian *Monotis richmondiana* Zittel in the Mount Mason and Okuku Limestones of the Warepan Stage (Jenkins, 1968; Jenkins and Jenkins, 1971; Campbell, 1959; Tozer, 1967). *Paragondolella steinbergensis*, present in the Okuku Limestone and known outside New Zealand only from the late Norian Alpine faunas of Austria (Mosher, 1968a) is the only conodont in the fauna that is of stratigraphic importance. Hence the Carnian-early Norian age given by Jenkins and Jenkins is too old.

*Neospathodus hernsteini* (Mosler) indicates a late Norian age for recently recovered faunas from New Guinea and India. In Austria, the Upper Norian *Epigondolella bidentata* Zone is distinguished by the presence of *N. hernsteini* (Sweet et al., 1971).

The small collection of conodonts from the Kuta Limestone, Gurai area, Chimbu District, New Guinea, shown to me by Dr R. Nicoll, Bureau of Mineral Resources, Canberra, contains *N. hernsteini* closely associated with ammonites, identified as a Late Triassic species of *Joannites* (Kummel, pers. comm.). Another occurrence of *N. hernsteini* (= *N. lanceolatus* Mosher) in a small conodont fauna from the Niti Pass area of the Kumaun Himalaya suggests that the Rhaetic age originally indicated by Sahni and Prakash (1973), is erroneous.

**Post-Norian Faunas**

Records of conodonts in post-Norian rocks are rare (Sweet et al., 1971), and they are generally considered to have become extinct in the latest Triassic (e.g. Sweet et al., 1971; Clark, 1972b). The dating of an Indian fauna as Rhaetic (Sahni and Prakash, 1973) has just been shown to be incorrect. The well-preserved *Epigondolella mungoensis* fauna in well-dated Upper Cretaceous sediments of the Cameroons (Diebel, 1956) may be a Late Cretaceous relict (Diebel, 1956; Lindström, 1964), or a reworked Triassic fauna (Mosher, 1967; Müller and Mosher, 1971). Recent well-dated records of the *E. mungoensis* fauna (see Sweet et al., 1971) show that it is essentially of late Ladinian age. Thus, reworking may seem the obvious answer to the problem of the Cameroons fauna, though the source area remains a vexed question.

**BIOSTRATIGRAPHICAL APPRAISAL**

**Distribution**

The restriction of the Triassic conodont faunas of the Gondwanaland Shelf to the southern margin of Tethys may be more apparent than real. No marine Triassic sediments in facies favourable for conodonts are known to me from Antarctica. However, ammonoid-bearing Triassic sediments of Anisian to Norian age in Chile (Harrington, 1962; Stipanicic, 1969) could yield conodont faunas of considerable biogeographical interest.

Provincialism among Triassic conodont faunas has been reported (Huckriede, 1958; Mosher, 1968a), but evidence of provincialism among the biostratigraphically important gondolellids and neospathodids is hard to find. On the other hand, subtle environmental differences may be recognised when comparing shallow shelf and deeper water conodont faunas from the Triassic of the Gondwanaland Shelf.

Little is known of the ecology of Triassic conodonts. McTavish (1973) inferred that
TRIASSIC TIME SCALE
(Silberling & Tozer, 1968)

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PERMIAN (P)

Fig. 36.4. Geographical and chronostratigraphical distribution of Triassic conodont faunas along

the Gondwanaland Shelf

species of *Neospathodus* favoured shallower water than species of *Neogondolella*, which Clark (1972a) and McTavish (1973) consider to be more common in deeper water environments.

**Zonation**

Although conodont species are apparently sufficiently widespread through the Triassic on the Gondwanaland Shelf to suggest the possibility of presenting a zonal scheme for the region, it would be premature to do so at present. Except in Kashmir (Sweet 1970a) and parts of some sections from the Salt Range and Trans-Indus Ranges (Sweet, 1970b), where sampling is at close enough intervals for detailed conodont stratigraphy, most Triassic conodonts are from spot samples collected originally for other purposes. Other samples come from allochthonous blocks in Timor (Nakazawa and Bando, 1968; Nogami, 1968; van den Boogaard, pers. comm.) and possibly New Zealand (the Okuku Limestone of Jenkins and Jenkins, 1971).

Few well-dated, continuous, and tectonically undisturbed sections of Lower to Upper Triassic marine sediments similar to that of Spiti are known to me from the Gondwanaland Shelf. As a result, any conodont faunal sequence established for this region may have to be composite. Hence, future detailed study should be directed to development of a zonation based on detailed morphogenetic analysis of conodonts, especially of gondolel-
lids and neospathodids, collected in closely sampled key sections rich in calcareous sediments. More information on phylogeny, such as that of Clark and Mosher (1966), Mosher (1968b), and Clark (1972b) is needed to provide a foundation for future research.

**COMPARISON WITH TRIASSIC PALYNOLOGICAL DATA**

It has proved difficult in the past to relate Triassic palynological zones to marine invertebrate zones, and to the world time scale. Preliminary observations suggest that the most fruitful way to approach this problem is through the study of conodonts and microfloras from associated samples. A good example is to be found in the Lower Triassic Mianwali Formation of the Salt Range, Pakistan, in which conodonts (Sweet, 1970b) and microfloras (Balme, 1970) were recovered from the same sections (Landa, Nammal, Zaluch, and Wargal) and in some cases the same samples. This formation has been divided, from the base upwards, into the (i) Kathwai Member, (ii) Mittiwali Member; and (iii) Narmia Member (Kümmel and Teichert, 1970). By relating the microfloras (Balme 1970: 421-3, figs. 2, 3) to the conodont faunas in close association, it can be shown that the microfloras of (i) the Kathwai Member are Griesbachian, (ii) the Mittiwali Member are Dienerian and possibly Smithian, and (iii) the Narmia Member are Spathian in age. Microfloras representative of the lower Dienerian, most if not all of the Smithian, and the uppermost Spathian were not described.

A less direct but equally interesting comparison, based mainly on subsurface material, can be made for the Lower Triassic in Western Australia. Balme (1969b) recognised close similarity between microfloras of the Kockatea Shale and the Kathwai and Mittiwali Members of the Mianwali Formation. Macrofaunal and conodont evidence (McTavish and Dickins, in press) establishes an Early Triassic (Griesbachian to Smithian) age for the Kockatea Shale of the Perth Basin. Balme stated also (1969a, 1969b) that the early Scythian was not represented in the subsurface Locker Shale of the Carnarvon Basin. Conodont evidence (McTavish, 1973) indicates that the basal part of the Locker Shale sampled to date is not older than middle Dienerian. Finally, in suggesting that the lowest, conodont-bearing part of the Locker Shale ranges in age from Dienerian to latest Spathian, the conodont evidence confirms and adds precision to Balme’s (1969a: 70) interpretation that most of the Locker Shale was younger than the Kockatea Shale.

**ACKNOWLEDGMENTS**

Professor K. J. Müller, Bonn, Dr A. van den Boogaard, Amsterdam, and Dr H. Taraz, Teheran, kindly provided unpublished information on Gondwanan Triassic conodont faunas. Dr E. C. Druce, Canberra, discussed aspects of Triassic conodont stratigraphy. Dr R. Nicoll, Canberra, kindly showed me the New Guinea conodont fauna. Professor B. Kummel identified ammonites found with the New Guinean conodonts. Mr M. H. Johnstone, Perth, has assisted greatly in attacking problems of palaeogeography and plate tectonics. The paper is presented with the permission of the Management of West Australian Petroleum Pty Limited.

**REFERENCES**

Triassic Conodonts and Gondwana Stratigraphy


Section 6

Tectonics, Igneous Activity, Geochronology, Structural Geology and Nature of Continental Margins
Deposition of Permian-Triassic Gondwana sequences in East Antarctica was preceded by development of extensive erosion surfaces cut across Devonian strata or Precambrian and Lower Palaeozoic basement. The best developed Gondwana sequences outcrop in the Transantarctic Mountains; the depositional basins were confined laterally by the East Antarctic craton and in West Antarctica by repeated orogenic uplift which was the principal factor in their evolution. Permian post-glacial dark shales were deposited in a basin deepening southward from Victoria Land. Permian coal measures fill two basins separated by a palaeogeographic high; one basin of limited extent occurs in south Victoria Land, the other, comprising a prograding delta and floodplain sequence, extends from south of the high to at least the Ohio Range. Regional palaeoslope reversal preceded Triassic fluvial deposition which in the Beardmore area terminated with eruption of acidic pyroclastics; sequences are, however, incomplete except in widely separated localities where they are capped by Jurassic basaltic rocks. Other Gondwana sequences are Permian in age, of limited areal extent, widely separated, and cannot as yet be tied to those elsewhere. Important exceptions are in the Pensacola and Ellsworth Mountains where Gondwana-related rocks are the upper parts of deformed sequences as old as Precambrian. One phase of the deformation is correlative with the Gondwanian Orogeny, which is better documented on other continents, and also in West Antarctica and the Antarctic Peninsula where the late Palaeozoic Trinity Peninsula Series may be partly contemporaneous with the early stages of deposition of Gondwana strata and may partly represent turbidites derived from glaciomarine deposits. Evolution of those basins with undeformed Gondwana sequences cannot as yet be related satisfactorily to those of other continents except in the broadest terms. Relations in the South Atlantic sector are the most difficult to interpret, and there is a striking lack of correlation between South Africa and the usually juxtaposed part of Dronning Maud Land.
constitute the Victoria Group of the Beacon Super-Group. The Beacon Super-Group (Barrett et al., 1972) consists largely of non-marine strata ranging in age from Devonian or older to Jurassic (Table 1). It unconformably overlies a Precambrian to Ordovician basement of metasedimentary and metavolcanic strata intruded by granitic rocks. Within the Beacon a marked disconformity, the Maya Erosion Surface (Harrington, 1965), separates the Devonian and older strata of the Taylor Group from the Permian to Jurassic Victoria Group. Other regional disconformities have been recognised in the Beacon, but none are as widespread. Where the strata are flat-lying, they have been extensively intruded by dolerite sills of Jurassic age, and in several places the sequence is capped by comagmatic basalt lavas. These igneous rocks form the Ferrar Group.

Beacon strata in the Transantarctic Mountains were deposited in two recognisable basins during the Permian but only one during the Triassic. The data and outcrop distribution elsewhere are insufficient for delineating the depositional basins even though thicknesses in certain cases might suggest that the basins were extensive.

**PRE-GONDWANA ROCKS**

The Gondwana strata of the Victoria Group lie with disconformity on the Taylor Group or with nonconformity or angular unconformity on Precambrian and Lower Palaeozoic basement. The Taylor Group, which overlies the Kukri Erosion Surface (Kukri Peneplain of Gunn and Warren, 1962), is best developed in south Victoria Land where it is about 1400 m thick (Table
Fig. 37.2. Geographic locations in the Transantarctic Mountains
### Table 37.1. Stratigraphic Correlation of the Victoria Group of the Beacon Super-Group

<table>
<thead>
<tr>
<th>Mt Mutschel area</th>
<th>Taylor Glacier area</th>
<th>North of Mackay Gl.</th>
<th>Prince Albert Mts</th>
<th>Priestley Glacier</th>
<th>Head of Rennick Gl.</th>
<th>Horn Bluff</th>
<th>Erosion Surface</th>
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**Legend**

- JURASSIC
- TRIASSIC
- ~ Pyramid ~
- ~ Maya ~
- ~ Heimdall ~
- ~ Kukri ~
- ~ Devonian and older ~
- ~ Victoria Group ~
- ~ Beacon strata 150 m ~
- ~ Beacon strata contact 60+ m not exposed ~
### Table 37.1. Stratigraphic Correlation of the Victoria Group of the Beacon Super-Group—Continued

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<th></th>
<th>Barrett, 1972a;</th>
<th>Barrett, 1972a;</th>
<th>Laird et al., 1971;</th>
<th>Skinner, 1964, 1965</th>
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Beacon strata 235 m
age unknown
Ellis Fm. 177+ m
(age uncertain)
Mithound Coal Measures 150 m

**DISCONFORMITY**

Darwin Tillite 27-82 m

Hatherton Sst.
Junction Sst.
Brown Hills Conglomerate

**UNCONFORMITY**
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Table 37.1. Stratigraphic Correlation of the Victoria Group of the Beacon Super-Group—Continued

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37.1). It consists predominantly of quartzose sandstone with conglomerate and feldspathic and lithic sandstone at the base, and red and green siltstone locally at the top (Table 37.1). The lower strata are unfossiliferous. A single fossil plant and rare acanthodian spines indicate a Devonian age for the Beacon Heights Orthoquartzite (McKelvey et al., 1972; Plumstead, 1962). It is locally overlain by the interbedded sandstone and variegated siltstone of the Aztec Siltstone. This formation has yielded conchostracans and freshwater fish remains which have been assigned a late Middle Devonian age by White (1968), but more recent collections led Ritchie (1969; see also McKelvey et al., 1972) to favour a Late Devonian age. Fossil spores from the Aztec are considered to indicate an early Late Devonian age (Helby and McElroy, 1969).

Unfossiliferous cross-bedded quartzose sandstones (Hatherton and Junction Sandstones, Alexandra and Castle Crags Formations) assignable to the Taylor Group, outcrop as far south as the Ramsey Glacier (Fig. 37.2; Table 37.1). The Horlick Formation of the Ohio Range (Long, 1964, 1965a) is a thin marine clastic sequence containing an abundant invertebrate fauna. The brachiopods indicate close affinities with the Malvinokaffric fauna of South America, the Falkland Islands and South Africa, and the Horlick Formation has been assigned an Early Emsian age (Boucot et al., 1967; Doumani et al., 1965).

The stratigraphic section in the Ellsworth Mountains is apparently conformable throughout (Craddock et al., 1964; Craddock, 1969). The Crashsite Quartzite overlies fossiliferous Upper Cambrian rocks and is overlain by the Whiteout Conglomerate, a glacially related diamicite. Brachiopods from the uppermost part of the Crashsite Quartzite indicate an early Emsian age (Boucot et al., 1967).

In the Pensacola Mountains a thick section of predominantly sandy strata occurs above a lower Palaeozoic angular unconformity and below the upper Palaeozoic glacially

Fig. 37.3. Diagrammatic representation of the Victoria Group of the Beacon Super-Group in the Transantarctic Mountains
derived Gale Mudstone (Schmidt and Ford, 1969; Schmidt et al., 1965; Williams, 1969). The sandy strata have been divided into the Neptune Group and the disconformably overlying Dover Sandstone. The Neptune Group is non-marine, except for the Elbow Formation which consists of marine clastics and minor phosphate beds. The Dover Sandstone is predominantly cross-bedded quartz sandstone, and is overlain, probably disconformably, by the Gale Mudstone. An isolated outcrop of lithologically similar sandstone includes a few carbonaceous interbeds that have yielded plant microfossils of probable Late Devonian age (J. M. Schopf, in Schmidt and Ford, 1969). Barrett et al. (1972) comment on apparent similarities between the Taylor Group and the Dover Sandstone and parts of the Neptune Group.

Elsewhere, the base of the Gondwana sequence is either not exposed, overlies metamorphic and granitic rocks of Precambrian and early Palaeozoic age, or overlies deformed sediments of probable early Palaeozoic age (Table 37.1).

**Victoria Group**

The Victoria Group constitutes the Gondwana sequence of the Transantarctic Mountains and ranges from (?) Carboniferous and Permian glacial strata to Triassic-Jurassic volcanoclastic rocks (Table 37.1). The thickness of the Victoria Group varies from a few meters in north Victoria Land to about 2300 m in the Queen Alexandra Range. The most complete stratigraphic sections are in the Queen Alexandra Range and south Victoria Land (Fig. 37.3). Early investigators of south Victoria Land strata now assigned to the Victoria Group include Allen (1962), Gunn and Warren (1962), Harrington and Speden (1962), Matz et al. (1972), McElroy (1969), McKelvey and Webb (1959, 1961), Mirsky et al. (1965), and Webb (1963). Using evidence from sedimentary petrology, Mirsky (1964) suggested a major disconformity in the Beacon of south Victoria Land, and Harrington (1965) proposed the terms Taylor and Victoria Groups with the Maya Erosion Surface between them. Recent work (McKelvey et al., 1970, 1972; Barrett et al., 1971, 1972) has clarified the somewhat confused nomenclature and correlation which had existed up to the late 1960s. In contrast, the nomenclature and lithostratigraphic correlation elsewhere has caused few problems.

The Maya Erosion Surface beneath the glacial beds of the Victoria Group has been examined in numerous places between south Victoria Land and the Ohio Range. It is generally planar but locally shows relief which attains a maximum of 80 m at Alligator Peak in south Victoria Land (Barrett, 1972b; see also Barrett et al., 1972) where the relief is due to local valley cutting. The anomalously thick sections of the Pagoda Formation in the Queen Elizabeth Range have been attributed to deposition in glacially modified pre-existing valleys (Lindsay, 1969).

In most areas the glacial beds rest with sharp contact on the basement or Devonian strata, but in a few localities the contact relations suggest that the underlying sediment was unconsolidated at the time of glaciation. In the Pensacola Mountains grooves attributed to soft-sediment deformation are inscribed on the upper surface of the Dover Sandstone (Frakes et al., 1971), and observations in south Victoria Land suggest the Maya Erosion Surface was cut across locally unconsolidated sediment (Barrett and Kyle, this volume; McKelvey et al., 1972).

In contrast to the Kukri Erosion Surface, the Maya Erosion Surface exhibits little, if any, sign of weathering except in the Geologists Range (Gregory, in prep.) where the underlying granite is deeply oxidised and weathered. This general lack of weathering can be attributed to late Palaeozoic glacial erosion that exposed fresh rock surfaces. The Kukri Erosion Surface has, no doubt, been trimmed by the Maya where the Taylor Group is absent.

The erosion surface at the base of the Gondwana sequence is equivalent to the Maya only where the glacial beds are present, and even this is questionable in the Ellsworth and Pensacola Mountains. It is probably equivalent to the Pyramid Erosion Surface where plant-bearing carbonaceous strata of Permian age rest on Devonian or older rocks. Where Triassic or Jurassic strata overlie the basement, the erosion surface may be partly
Fig. 37.4A. Distribution of late Palaeozoic glacial strata (Pagoda Formation and equivalents). The three basins of depositions in the Transantarctic Mountains identified by Frakes et al. (1971) have not been distinguished because the glacial basins do not necessarily tie in with the basins identified in later strata. Palaeo-iceflow data sources: a. Frakes et al., 1971; b. Long, 1965a; c. Minshew, 1967; d. Long, unpubl.; e. Coates, 1972; f. Lindsay, 1970; g. Barrett and Kyle, this volume.

Fig. 37.4B. Distribution of the post-glacial shales (Mackellar Formation and equivalents). The area outlined from the Nimrod Glacier to the Scott Glacier and beyond represents the Nimrod-Ohio Basin (see p. 504). See also Fig. 37.3 for the relation between the Nimrod-Ohio Basin and the South Victoria Basin (see p. 504). Palaeocurrent data sources: a. Frakes et al., 1971; b. Minshew, 1967; c. Barrett, 1965; d. Elliot, unpubl.; e. Barrett, 1970; f. Lindsay, 1969; g. Laird et al., 1971.
equivalent to the sub-Mawson surface (Barrett et al., 1972).

**Late Palaeozoic Glacial Strata**

Glacially related sediments overlying the Maya Erosion Surface outcrop from Mount Bastion in south Victoria Land, to the Pensacola Mountains, and also have been identified tentatively in Heimefrontfjella. Summaries and reviews of the glacial strata have been given by Coates (1972), Frakes et al. (1971) and Lindsay (1970) and therefore are discussed only very briefly. The most critical studies are those of Frakes et al. (1966, 1968, 1971) and Lindsay (1969, 1970). The glacial beds thicken southward from Mount Bastion and attain 1122 m in the Ellsworth Mountains (see Table 37.1). They have a variety of local formational names (Pagoda Formation and equivalents), but there is little doubt about their lithostratigraphic correlation. The lithologies include tillite, diamictite, conglomerate, sandstone, siltstone, shale and limestone. The term tillite has been used extensively but it is likely that some of the rocks so named should be described as diamictite because they are deposits formed through reworking of glacial debris by mass movements.

Several basins of deposition for the glacial strata have been identified (Frakes et al., 1971), and the strata in the Darwin Mountains and south Victoria Land are discussed in detail in this volume by Barrett and Kyle. Palaeo-iceflow was consistently along the length of the Transantarctic Mountains (Fig. 37.4A); the direction was away from Victoria Land toward the Ohio Range, except locally near the Shackleton Glacier (D. A. Coates, pers. comm.). Iceflow vectors in the Ohio Range are somewhat uncertain but perhaps the most important because of the regional interpretations of glaciation which have been developed on flow to the west. Long (1965a) determined west to east flow but Frakes et al. (1971) dispute this and maintain that flow was in the opposite direction. This difference of opinion has not yet been resolved. It is possible that both directions are represented.

Frakes et al. (1971) have summarised the data suggesting that the glacial strata other than those in the Ellsworth and Pensacola Mountains were deposited in a stable continental environment.

It should be noted that the upper part of the glacial beds are approximately the same age in the Ohio Range and south Victoria Land (see below), so that ice retreat in the Transantarctic Mountains was approximately contemporaneous wherever the ice centres were. The glacial strata are largely deposits from wet-based glaciers and as Barrett and Kyle (this volume) note, glacially related strata are commonly deposited in the final retreat stages (see also Mickelson, 1973). Little, if any, record remains of the onset of glaciation which may well have varied with location.

**Palaeontology and Age**

Plant material has been recovered from the glacial beds at several localities, and indicates that the ice advanced in those over vegetated areas (Schopf, 1973). However, determinable fossil remains have been found only in the Buckeye Formation and the Darwin Tillite. Plant microfossils extracted from shale interbedded in the upper part of the Buckeye Formation (Schopf, in Long, 1964; Rigby and Schopf, 1969) mention a species of *Nuskiosporites* very similar to one found both in the Bacchus Marsh Sandstone of Australia and just above the Talchir Boulder Bed in India. More recently, Schopf (1971b) has suggested a Sakmarian age for this spore assemblage. Plant microfossils recovered from the upper part of the Darwin Tillite are approximately the same age (Barrett and Kyle, this volume). At Mount Fleming, the Metschel Tillite is gradational over 30 cm into carbonaceous shales containing *Gangamopteris*. No precise age can be attached to *Gangamopteris* and the nature of this contact is uncertain, because elsewhere a disconformity, the Pyramid Erosion Surface, separates the glacial beds and overlying coal measures. The two plant microfossil assemblages suggest that from south Victoria Land to the Ohio Range the upper part of the glacial beds is Early Permian, and at least in part Sakmarian. However, the age of the lowest glacial beds is still open to question. The ages of the Gale Mudstone and
Whiteout Conglomerate are also uncertain. On regional lithologic and stratigraphic similarities, Schmidt and Williams (1969) assign a probable Early Permian age to the Gale Mudstone. However, Frakes et al. (1971) consider the Gale possibly as old as Early Carboniferous; this is based on a regional consideration of Gondwana glaciation and the unconsolidated nature of the Dover Sandstone at the time of deposition of the oldest diamictite. Frakes et al. (1971) also postulate a Carboniferous age for the Whiteout Conglomerate.

The Aztec Siltstone also displays evidence of soft sediment deformation at the time of till deposition (Barrett et al., 1971) and there is adequate reason to believe that the glacial beds there are Lower Permian and not Carboniferous. Situations analogous to that in the Pensacola Mountains are known elsewhere, for instance at Oorlogskloof in the western Cape Province where Dwyka glaciation cut soft-sediment grooves is the Nardouw Sandstone of the Table Mountain Group (Guidebook No. 2, Second Gondwana Symposium, South Africa, 1970). Nevertheless, if the migration with time of the ice centres across Gondwanaland is accepted (Crowell and Frakes, 1970), there is reason for assigning a slightly greater age to the Gale Mudstone and Whiteout Conglomerate.

Permian Post-glacial Strata

The Pagoda Formation in the central Transantarctic Mountains is conformably overlain by laminated shale and fine sandstone of the Mackellar Formation, sandstone of the Fairchild Formation and by Permian coal measures, the Buckley Formation (Table 37.1). Similar conformable sequences are known through to the Ohio Range, but toward the Byrd Glacier these formations are either absent or represented by non-carbonaceous strata unconformable on basement rocks (Fig. 37.3) (Skinner, 1964, 1965). Coal measures are disconformable on glacial strata in the Darwin Mountains and south Victoria Land (Table 37.1). Immediately north of the Mackay Glacier the carbonaceous strata are disconformable on the Taylor Group (Mirskey et al., 1965), but have not been found between there and the Freyberg Mountains (Fig. 37.2) where they rest on Ordovician granite (Dow and Neall, 1972).

Glossopteris-bearing beds in the Ellsworth Mountains are conformable on Mackellar-like strata which in turn are conformable on the Whiteout Conglomerate. Equivalent carbonaceous beds in the Pensacola Mountains, the Pecora Formation, are not in contact with the Gale Mudstone. Glossopteris-bearing strata also outcrop in the Whichaway Nunataks, Theron Mountains, western Dronning Maud Land, the foothills of the Prince Charles Mountains, and at Horn Bluff (Table 37.1).

The disconformity at the base of the coal measures in south Victoria Land has been named the Pyramid Erosion Surface (McKelvey et al., 1970). It is recognised throughout that area and the Darwin Mountains. The unconformity in the Freyberg Mountains can perhaps be referred to that erosion surface. Elsewhere in north Victoria Land, the surface below the Beacon is younger than the Pyramid. The basal erosion surfaces in western Dronning Maud Land and the Prince Charles Mountains may be approximately contemporaneous with the Pyramid.

Queen Alexandra Range to Nimrod Glacier

The Pagoda Formation in the central Transantarctic Mountains is conformably overlain by laminated shale and fine sandstone of the Mackellar Formation (Table 37.1). The Mackellar (Barrett, 1969, 1972a; Lindsay, 1969, 1970) consists of rhythmically laminated shale and fine sandstone together with interbedded black shale, thinly-beded or massive lens-like sandstone, and occasional thin limestone. There is little variation in lithology except in the Moore Mountains where numerous sandstone beds occur, in particular toward the top of the section. The sandstone has abundant bedding surface structures such as animal trails, mud-cracks and possible rain prints. Micro-cross-lamination is common in the thinner sandstone beds. The thick (3-12 m) fissile shale beds are carbonaceous but are calcareous where interbedded in laminated shale and sandstone.

The Fairchild Formation and the overlying Buckley Formation were first described as the Buckley Coal Measures (Grindley,
North of the Nimrod Glacier (Gregory, in prep.; Laird et al., 1971) the Mackellar is predominantly laminated, fissile, micaceous or carbonaceous dark shale interbedded with massive, cross-bedded medium sandstone. It is overlain conformably by sandstone, siltstone and shale assigned to the Buckley Formation or Buckley Coal Measures. The lower part consists of thick, massive cross-bedded coarse sandstone, locally containing mud-flake breccias and pebble beds, together with subordinate grey-green sandstone, carbonaceous siltstone and shale; it is probably in part equivalent to the Fairchild. The upper part consists of cyclic units similar to those recorded by Barrett, and also includes coal and Glossopteris.

Palaeocurrent vectors (Figs. 37.4B and 37.5) indicate flow to the south or southeast over most of the area from north of the Nimrod to the head of the Ramsey Glacier (northwest of the Shackleton Glacier); however, there is evidence for sediment influx from the west (Barrett, 1970; Laird et al., 1970).

South of the Byrd Glacier (Figs. 37.2 and 37.3) arkosic sandstone rests on an erosion surface which has as much as 60 m relief (Skinner, 1964, 1965). A basal conglomerate of quartz pebbles, and locally of Cambrian limestone clasts, fills lows in the erosion surface. The overlying cross-bedded arkosic sandstone is interbedded with fissile grey-green micaceous sandstone and black shale. These rocks are not fossiliferous and their lithostratigraphic correlation is uncertain.

**Queen Maud Mountains** In the Buckley Island area at the head of the Beardmore Glacier (Fig. 37.2) the tillite is absent and the Mackellar Formation directly overlies a grooved surface cut in Alexandria Formation sandstone (Coates, 1972). Beds, in part equivalent to the Fairchild Formation are described as the Lower Buckley Coal Measures (Young and Ryburn, 1968). The Upper Buckley Coal Measures are equivalent to part of the Buckley Formation. The Buckley also outcrops in the Dominion Range, Supporters Range and Graphite Peak area (Barrett, 1969; Collinson and Elliot, in prep.; McGregor, 1965). All these sections are compar-
able to the upper Buckley north of the Beardmore. Carbonaceous matter and fossil leaves are abundant at Graphite Peak.

The Pagoda Formation in the Shackleton Glacier area passes up into interbedded laminated to platy shale and fine sandstone of the Mackellar Formation. Dropstones are present in the lowest 2 m of the formation. Fine-grained sandstone with micro-cross-lamination is more abundant up section, becomes coarser grained and occurs in thicker units. Rhythmic alternations of shale and sandstone are common, and concretionary horizons are also present. The Mackellar (Barrett, 1965: Unit A) is absent at Cape Surprise. The massive sandstones above the Mackellar have been designated the Mount Butters Formation (La Prade, 1970) and Unit B (Barrett, 1965); however, there is no question of their lithostratigraphic correlation with the Fairchild. The formation is again divisible into a lower cliff-forming part of massive sandstone with minor interbedded shale and siltstone, and an upper slope-forming part which has a greater proportion of fine sediment. The quartz-pebble horizon defining the base of the Buckley Formation in the Queen Alexandra Range is recognised locally. The overlying cyclic beds are similar to those already described for the Buckley Formation, and contain an abundant Glossopteris flora. La Prade (1970) records a composite thickness of 467 m, considerably less than that north of the Beardmore Glacier.

At Mount Fridtjof Nansen the Mackellar equivalent (Barrett, 1965: Unit A) overlies residual pockets of glacial or fluvo-glacial sediment with scattered large clasts (Barrett, 1966). It consists of shale overlain by interbedded micro-cross-laminated fine sandstone and shale, together with a few thin lenticular limestone beds. Unit B (Barrett, 1965), the Fairchild equivalent, is largely medium to fine sandstone with abundant micro-cross-lamination and occasional trough cross-bedding. The lower part of Unit C (Buckley Formation) comprises about 50 m of cross-bedded sandstone with subordinate finer-grained clastics, and the lower contact is again defined on the lowest occurrence of quartz pebbles which are locally abundant. Several thin limestone beds occur in this lower member. Higher in the sequence the beds become finer-grained and carbonaceous.

The post-glacial strata of the Nilsen Plateau consist of laminated micaceous and carbonaceous shale interbedded with minor siltstone, mudstone, and very fine sandstone; they constitute the Roaring Formation (Long, 1965b, unpubl. ms.). It is locally absent due to erosion before deposition of the Amundsen Formation which is, therefore, locally disconformable. Cross-bedded and current-rippled medium to fine arkosic sandstone, together with minor carbonaceous and micaceous siltstone and mudstone make up the Amundsen. The upper contact is defined as the base of the Glossopteris-bearing shale and mudstone sequence of the Queen Maud Formation, and therefore the Amundsen is probably equivalent to the Fairchild plus some of the lower Buckley. At the type section of the Queen Maud Formation at the head of the Scott Glacier, the lower contact is placed at the base of the first quartz-pebble conglomerate, an horizon well below the carbonaceous beds. On the Nilsen Plateau the thickness of the Queen Maud Formation varies from 95 m at one complete section to at least 200 m. The variation in thickness is attributed to erosion prior to deposition of the overlying Nilsen Formation. The Queen Maud Formation, like the other coal measures, consists largely of cyclic beds of sandstone, siltstone, carbonaceous shale and mudstone, and coal. The Glossopteris flora is common in the carbonaceous beds.

The Weaver and Queen Maud Formations, which are similar to the Mackellar, Fairchild and Buckley Formations, outcrop at the head of the Scott Glacier (Doumani and Minshew, 1965; Minshew, 1967). The Weaver Formation has a gradational lower contact with the Scott Glacier Formation, but at the type locality it is disconformable and locally it directly overlies basement rocks. Here the Weaver consists of a basal conglomerate and overlying medium to coarse sandstone which grades both laterally and vertically into dark silty shale. The shale makes up the entire lower member of the Weaver in some localities but elsewhere is interbedded with sandstone and minor limestone lenses. Shale decreases upwards and the
upper part of the middle member consists of fine to medium sandstone interbedded with siltstone. The lower and middle members of the Weaver are equivalent to the Mackellar. North of Mount Blackburn, the middle member is formed of interbedded sandstone and siltstone, and interfingers northwards with coarse sandstone which represents the middle and upper members of the Weaver and possibly the lowest part of the overlying Queen Maud Formation. Katz and Waterhouse (1970) measured a section north of Mount Blackburn but did not record lithologies comparable with the conglomeratic sandstones just described. The upper member of the Weaver Formation, which consists of feldspathic sandstone, with subordinate shale units particularly prominent in the upper part, is equivalent to the Fairchild. The sandstone is cross-bedded, and rippled marked and intra-formational conglomerate consisting of black shale clasts is widespread. The shale in the upper part is locally carbonaceous and may include a few thin, low-grade coal seams. A few thin limestone beds are also present. The basal quartz-pebble conglomerate beds of the Queen Maud Formation are succeeded by poorly developed cycles, with a total thickness of about 75 m of calcareous sandstone or arenaceous limestone, siltstone, shale and sparse lenticular shaley coals. The upper part comprises typical cyclical units of sandstone, siltstone, shale, mudstone, carbonaceous shale and coal with which sideritic concretions and lenses may be associated. Shale is the most abundant lithology in the upper part of the formation. Tuffs occur in the upper 200 m at Mount Weaver and a single bentonite bed has also been recorded. Composite stratigraphic sections suggest a thickness of 525± m for the carbonaceous upper part of the formation. Glossopteris is abundant.

Palaeocurrent vectors (Figs. 37.4B and 37.5) indicate flow to the southeast and east for the Queen Maud Mountains.

Horlick Mountains Glacial strata of the Buckeye Formation (Table 37.1) are disconformably overlain by the Discovery Ridge Formation in the Ohio Range (Long, 1965a) and conformably by the Weaver Formation in the Wisconsin Range (Minshew, 1966, 1967). The Discovery Ridge Formation is a thinly-bedded grey and black shale unit which can be divided into a lower member of platy silty shale up to 45 m thick and an upper member of carbonaceous and fissile shale up to 150 m thick. Locally, there are lenticular limestone beds up to 65 cm thick. The two members have an interbedded gradational contact. The lower and middle members of the Weaver Formation in the Wisconsin Range are equivalent to the Discovery Ridge Formation. The lower member is a black shale with scattered dropstones, and includes interbedded sandstone in its middle part. Sideritic concretions are concentrated in two horizons but also occur scattered throughout. The middle member consists of rhythmically interbedded sandstone and shale in units a few centimetres thick. Carbonaceous matter is common as also are sole markings and trace fossils.

The upper member of the Weaver Formation in the Wisconsin Range is largely cross-bedded sandstone but finer-grained beds including carbonaceous shale and thin coal beds occur near the top. It is overlain by 25 m of conglomeratic sandstone which forms the base of the Queen Maud Formation.

In the Ohio Range (Table 37.1) (Long, 1964, 1965a) the black shales of the Discovery Ridge Formation have a gradational contact with the overlying Mount Glossopteris Formation. No continuous complete section is exposed, but it is clear that the formation consists of a lower sandy part equivalent to the Fairchild, and an upper carbonaceous part equivalent to the Buckley. The sandstone in the lower part forms cliffs up to 30 m high and has a total thickness of 100-150 m. A prominent quartz-pebble horizon (see Minshew, 1967; Mirsky, 1969) occurs about 150 m above the base of the formation and this can be considered the base of the upper part, the Buckley equivalent. The remainder of the formation, which may be as much as 600 m thick, consists of a cyclic sequence of sandstone, siltstone, carbonaceous shale and coal similar to those already described. Coal is more abundant in the lower half of the upper unit. A total of 23 m of coal in beds 1-3 m thick occurs in one section. There is an
Fig. 37.5. Distribution of Middle and Upper Permian strata (Fairchild, Buckley, lower part of the Feather Conglomerate, and equivalents). The distribution of the Fairchild, overlying Buckley, and equivalents corresponds to the Nimrod-Ohio Basin (see p. 517), whereas the South Victoria Basin (see p. 517) is represented by outcrops between the Darwin and Mackay Glaciers. See also Fig. 37.3 for the relation between the Nimrod-Ohio Basin and the South Victoria Basin, and the palaeotopographic high separating them. Palaeocurrent data sources: a. Long, 1965a; b. Minshew, 1967; c. Long, 1965b, unpubl.; d. Barrett, 1965; e. Barrett, 1970; f. Laird et al., 1971; g. Haskell et al., 1965; h. Barrett and Kohn, this volume; i. Brook, 1972; j. Hjelle and Winsnes, 1972; k. Aucamp et al., 1972; l. Mond, 1972.
abundant *Glossopteris* flora and, locally, carbonaceous shale has yielded conchostracans (see p. 512). Palaeocurrent measurements all indicate flow to the east (Fig. 37.5).

**Ellsworth Mountains** The Polarstar Formation in the Ellsworth Mountains (Craddock, 1969; Craddock et al., 1964) consists of interbedded argillite and sandstone in beds 1 mm to 1 m thick. The lower 600 m is strikingly similar to the black shales of the Mackellar (J. M. Schopf, 1969 and pers. comm.). The sandstone, which is more abundant higher in the section, consists of quartz, feldspar, volcanic and other rock fragments in an argillaceous matrix. A few coal beds as much as 30 cm thick, and a few possible Bentonitic beds up to 75 cm thick occur in the upper part of the formation. Many of the argillites are carbonaceous and six from the upper part have yielded *Glossopteris* (Craddock et al., 1965). Graded bedding, cross-bedding, current ripple marks and slump structures have been recorded in the arenaceous beds.

Palaeocurrent measurements from the base of the formation indicate derivation from the south-southeast (Matthews et al., 1967).

**East Antarctica** The Pecora Formation (Schmidt and Ford, 1969; Williams, 1969) in the Pensacola Mountains structurally overlies the Gale Mudstone but outcrops only in isolated nunataks in the southern Forrestal Range and in the Pecora Escarpment where it is intruded by dolerite sills. The Pecora Formation consists of thin-bedded fine sandstone with thin interbeds of siltstone and shale. The sandstone is relatively impure, feldspathic, and has chert as the only significant rock fragment. Many of the interbeds are carbonaceous or coaly and locally contain *Glossopteris* leaves.

About 250 m of sandstone with minor intraformational conglomerate and shale, and intruded by dolerite, is exposed in the Whichaway Nunataks (Table 37.1) (Stephenson, 1966). The petrography of the sandstone suggests a granitic and older sedimentary rock source, but one sample had abundant rock fragments of probable intermediate and acid composition. *Glossopteris* has been recovered from the shales.

A coal-bearing sequence intruded by dolerite sills is exposed in the Theron Mountains (Brook, 1972; Stephenson, 1966). The strata comprise varying amounts of medium and fine sandstone, laminated siltstone, shale and mudstone. The dominant lithologies grade from one to another both laterally and vertically. Coal and coaly shale are ubiquitous and may grade into mudstone and shale with plant remains. Composition of the sandstone suggests a metamorphic provenance with possibly some acid volcanics; however, volcanic sandstones such as in the Buckley Formation have not been recorded. A south-westerly source has been inferred for these sediments (Fig. 37.5).

There are scattered outcrops of Permian strata in western Dronning Maud Land. A thin section of sandstone with subordinate fine sandstone, siltstone and dark grey shale is exposed in an isolated nunatak southeast of Vestfjella (Hjelle and Winsnes, 1972; Winsnes, 1969). The shales have yielded *Glossopteris*, *Gangamopteris* and *Vertebraria*. Palaeocurrents indicate derivation from the east or northeast (Fig. 37.5).

A possible Permo-Carboniferous tillite overlain by Permian sediments which include coal and carry *Glossopteris* are reported from northeastern Heimefrontfjella (Adie, 1970: table VII). Plumstead (1971) also refers to this area.

Further east, on the Kirwan Escarpment, the basement is cut by a gently undulating erosion surface which is weathered to a depth of 5 m, and on which the late Palaeozoic Amelang Formation was deposited (Aucamp et al., 1972). The section comprises two members. The 25 to 70 m-thick lower member consists of sandstone with intercalated conglomerate, shale and mudstone. A basal poorly sorted conglomerate up to 3 m thick is not considered to be a ‘tillite’ and the glacial striae on the exhumed erosion surface are believed to be the result of recent ice movement. A thin carbonaceous shale near the base of the member has yielded woody remains tentatively assigned a Permian age. The sandstones are characterised by ferruginous concretions. The 10 to 50 m-thick upper member of poorly sorted fine sandstone also contains ferruginous concretions.
The Permian Amery Group (Mond, 1972), forms the only Beacon Super-Group strata outcropping between 10°W and 150°E around the periphery of East Antarctica (Fig. 37.1). *Glossopteris*-bearing siltstone has also been recorded in moraines in the upper part of the Lambert Glacier 250 km to the south (Trail, 1963; Trail and McLeod, 1969). The Permian strata were first described as the Amery Formation (Crohn, 1959) but on the basis of more detailed studies (Mond, 1972) the strata have been subdivided into three formations (Table 37.1) which constitute the Amery Group. The Radok Conglomerate consists of pebble and cobble conglomerate interbedded with medium to coarse sandstone and thin-bedded and fissile carbonaceous siltstone and shale. The pebbles and cobbles are largely of high-grade metamorphic rocks similar to the adjacent basement complex. The lower contact is obscure but it is believed to be a nonconformity (Mond, 1972) and not faulted as postulated by Crohn (1959). The overlying interbedded arkose, feldspathic sandstone, siltstone, shale and coal constitute the Bainmedart Coal Measures. Cross-bedded feldspathic sandstone forms as much as 50 per cent of the section. Much of the siltstone and shale is carbonaceous and grades into coal, of which sixty seams ranging from 8 to 350 cm thick have been recorded. *Glossopteris* leaf impressions are abundant. The lower contact of the Flagstone Bench Formation is obscured by moraine, and it may be faulted. This formation consists of very coarse, locally pebbly, feldspathic sandstone with iron concretions, together with interbedded grit and sandy siltstone. Carbonaceous matter has not been recorded.

The Radok Conglomerate and Bainmedart Coal Measures dip about 110° to the east, whereas the Flagstone Bench Formation dips about 5° to the northeast. This suggests a possible discordance; however, since the lower contact of the Flagstone Bench Formation is obscure, the significance of the change in attitude is uncertain. It is possible that the attitudes could be ascribed to tilting of fault blocks.

South Victoria Land and Darwin Mountains The Pyramid Erosion Surface (Barrett and Kyle, this volume; McKelvey et al., 1970, 1972), which separates the Darwin and Metschel Tillites from the Misthound and Weller Coal Measures, seems to be nearly planar. However, locally at Mount Ritchie there is as much as 100 m of relief where the Pyramid has cut through the Metschel Tillite and Maya Erosion Surface and the Weller Coal Measures rest on the Aztec Siltstone.

The Misthound Coal Measures (Haskell et al., 1964, 1965; Barrett et al., 1971) comprise a basal conglomeratic sandstone with clasts up to 60 cm across of all lithologies found in the tillite, and overlying massive fine sandstone that higher in the section is interbedded with thin coal and carbonaceous siltstone from which poorly-preserved *Ganapopteris* and *Glossopteris* have been recovered. Palaeocurrent data indicate flow to the southeast. The coal measures are stratigraphically overlain by the Ellis Formation, but a dolerite sill has been intruded along the contact. The Ellis consists of sandstone with scattered quartz pebble horizons and carbonaceous laminae, and subordinate siltstone. Stem impressions are found locally. The Ellis Formation may be either Permian or Triassic.

Between the Mulock and Mackay Glaciers, the Pyramid Erosion Surface is overlain by the Weller Coal Measures which consist of coarse to fine sandstone with minor very fine sandstone, siltstone, shale and mudstone (Barrett et al., 1971; McKelvey et al., 1970, 1972). A thin conglomerate occurs at the base and contains granitic clasts which persist for a few metres upwards. Many of the clasts above the base are strongly weathered, indicating post-glacial weathering of glacial debris prior to redeposition in the coal measures (Barrett and Kohn, this volume). The bulk of the formation is cross-bedded carbonaceous sandstone with thin beds and lenses of carbonaceous siltstone. Fining-upwards cycles have been recognised. Shale and coal are minor constituents in the upper third. The coals are up to 7 m thick. Logs, stems and abundant *Glossopteris* leaves are present. Equivalent units are the Mount Bastion Formation (Mirskey et al., 1965), the lower part of the Mount Bastion Coal Measures (Allen, 1962) and the Mount Fleming Formation (Matz et al., 1972).
The coal measures are overlain by the Feather Conglomerate (Barrett et al., 1971; Barrett and Kohn, this volume), a unit of conglomeratic sandstone and grit which becomes finer-grained and more feldspathic northward. Quartz pebbles are abundant south of upper Wright Valley. The Robison Peak Formation (Matz et al., 1972) is equivalent to the Feather.

Barrett and Kohn (this volume) have located a watershed in the Mount Fleming area which separates flow to the south from flow to the northwest for the lower part of the Weller. The upper part of the Weller is absent south of the watershed. Palaeocurrents flowed to the northwest for both the upper part of the Weller and the Feather.

North Victoria Land No Permian strata are known between a point immediately north of the Mackay Glacier and the Freyberg Mountains where about 60 m of quartzose and feldspathic sandstone, and minor carbonaceous shale and coal outcrop (Dow and Neall, 1972; Sturm and Carreyer, 1970). The carbonaceous beds have yielded a Permian plant assemblage.

At Horn Bluff, 700 km to the northwest, cross-bedded feldspathic sandstone and minor shale and some coal outcrops (Mawson, 1940). The shales are carbonaceous but have yielded only plant microfossils. The coal occurs in laminae and lenses in the sandstone. Some sandstone is reported to contain zeolite.

Palaeontology and Age No precise age can be given for the Mackellar Formation and its correlatives because although carbonaceous matter is abundant, determinable plant remains have yet to be found. However, plant remains from concretions in the lower member of the Weaver Formation are reported (J. M. Schopf in Minshew, 1967) to be similar to the Glossopteris flora which is abundant in the overlying coal measures here and elsewhere in the Beacon strata.

The abundant Glossopteris flora in the coal measures has been the principal basis for their assignment to the Permian. The sparse representation of plant microfossils in the Transantarctic Mountains is attributable to baking by the ubiquitous Jurassic dolerite sills. Pavlov (1958) reported Permian or possibly Triassic microfossils at Horn Bluff; J. M. Schopf (pers. comm.) believes a Triassic age is more likely. Schopf (1962) and Long (1965a) described plant microfossils from the Ohio Range which were assigned an age based on the much better preserved megafossils. The Amery Group, which is not intruded by dolerite, has yielded relatively abundant plant microfossils for which a late Permian age has been deduced (Balme and Playford, 1967; Kemp, 1973).

Permian megafossils are dominated by species of Glossopteris, though Gangamopteris is a common associate, particularly in the lower part of the section. Noeggerathioipsis, a cordaitean, is less common. Sphenopсидs such as Paracalamites, the equisitalean Phyllotheca, and coniferous remains are rare. Vertebrazier roots and stems, and fossil wood (Antarcticoxylon and Dadoxylon) are quite common. The most definitive studies are by Plumstead (1962), Gridland (1963), Rigby and Schopf (1969), and White (1970). One noticeable feature of the flora is the paucity of extra-Gondwana elements (Rigby and Schopf, 1969) which is in contrast to Glossopteris floras from other Gondwana continents. Assemblages with Gangamopteris are generally considered to be older than those with Glossopteris alone.

Rigby and Schopf (1969) regard the Antarctic flora as representative of the Middle and Late Permian; however, Plumstead (1962) considers it to be of Early and Middle Permian age and has recently reported on a possible Late Carboniferous flora from Heimefrontfjella (Plumstead, 1970).

The stratigraphically important plant localities are those near the top and either near the base of the coal measures or within the Fairchild and equivalents. Between the Ohio Range and south Victoria Land, a lower age limit is given by the Sakmarian microfossils from the Buckeye Tillite (Schopf, 1971) and the Darwin Tillite (Barrett and Kyle, this volume). A Gangamopteris-dominated assemblage has been found just above the gradational contact of the Metschel Tillite with the Weller Coal Measures, and in the lower part of the Misthound Coal Measures (Barrett and Kyle, this volume; Haskell et
al., 1965). The Fairchild Formation in the Queen Elizabeth Range has yielded a few Gangamopteris leaves from near the lower contact (Barrett, 1969). Equivalent fossiliferous strata are found only in the Wisconsin Range, where shales in the uppermost part of the Weaver Formation have yielded a Glossopteris assemblage (Minshew, 1967). The coal measures flora is dominated by Glossopteris and has been assigned a middle and late Permian age (Rigby and Schopf, 1969). The strata overlying the coal measures in the central Transantarctic Mountains have recently been dated as Lower Triassic on the contained vertebrate fossils (see p. 516). Glossopteris is found to within a few metres of the contact at a number of localities (Barrett, 1969; Elliot et al., 1972). Plant evidence of a Triassic age is based elsewhere on either the association of Dicroidium and Glossopteris or the occurrence of Dicroidium alone. Glossopteris alone occurs in the Weller Coal Measures within 12 m of the overlying Feather Conglomerate, the lower part of which is also assigned to the Permian (Barrett and Kohn, this volume).

A comprehensive account of the coals from the Mount Glossopteris Formation and a review of other occurrences has been given by Schopf and Long (1966) (see also Schopf, 1962; Long, 1965a). Coals have also been described in detail from near the Mackay Glacier (Mulligan et al., 1963a, 1963b), the Theron Mountains (Brown and Taylor, 1960) and the Prince Charles Mountains (Bennett and Taylor, 1972). Brief descriptions are also given for coals from the head of the Victoria Valley (Allen, 1962), the Queen Alexandra Range (Barrett, 1969) and the Nilsen Plateau (Long, unpubl. ms.).

The coal seams, wherever they occur, are lenticular and cannot be traced for more than a kilometre or so. The maximum recorded thickness of any one seam is 11 m, but generally they are much less. Shaley coal or coaly shale is also relatively abundant. Seat earths have not been found, with one possible exception at Mount Weaver (Minshew, 1967). Most of the coal, which is high in ash and low in sulphur, is high rank, anthracitic or semi-anthracitic (Schopf and Long, 1966). Devolatilisation is attributable to the presence of dolerite sills that have affected most of the coal. Coal in the Bainmedart Coal Measures, Prince Charles Mountains, has not been affected by dolerite intrusions and is considered to be high volatile-bituminous (Bennett and Taylor, 1972).

Permineralised peat in the Buckley Formation near Mount Augusta at the southern end of the Queen Alexandra Range (Schopf, 1970, 1971a) is under continuing study and should provide invaluable information concerning the Glossopteris flora. On the basis of this peat Vertebraria is considered to be the root material of the Glossopteris flora.

Invertebrate fossils have been found in exotic blocks and in situ at a few localities. Indeterminable pelecypods were discovered in a debris slope of material probably derived from the upper member of the Weaver Formation in the Wisconsin Range (Minshew, 1967). Fossil insect wings have been reported from the Ellsworth Mountains (Tasch and Riek, 1969) and Theron Mountains (Plumstead, 1962). The most important invertebrate fossils are the conchostracans found in carbonaceous shale about 160 m below the top of the Mount Glossopteris Formation (Doumani and Tasch, 1965; Tasch, 1969). By comparison with Leaia in the lower Beaufort of South Africa, the dominant species, Leaia gondwanella (Tasch) and Cyzicus (Lioestheria) doumanii, have been assigned a Middle to Late Permian age.

**Triassic Strata**

Triassic rocks outcrop in the Transantarctic Mountains between the Nilsen Plateau and north Victoria Land (Fig. 37.6A). In the Beardmore area a disconformity between Permian coal measures and Triassic strata marks a significant palaeocurrent reversal. No comparable disconformity is present in south Victoria Land. In north Victoria Land Triassic strata form a thin veneer on an erosion surface cut across deeply weathered basement rocks (Skinner and Ricker, 1968; Gair, 1967). This surface, which is exposed from the Freyberg Mountains to the Priestley Glacier (Gair et al., 1969), testifies to the continued exposure of this area as a sediment source through much of Beacon time.
The Triassic strata pass up into volcanoclastic beds in the Beardmore area (Barrett and Elliot, 1972). In south Victoria Land a marked disconformity is developed locally on the Triassic beds and it is overlain by a laharic unit of basaltic composition which is related to the succeeding basalts. The veneer
of Triassic beds in north Victoria Land is overlain by the thickest succession (1370 m) of Jurassic basalts in Antarctica.

Central Transantarctic Mountains

The Triassic Fremouw Formation (Barrett, 1969, 1972a) is readily divisible into three parts. The lower part, 75 to 125 m thick, consists, in most places, of three or more fining-upwards cycles, each up to 40 m thick. A typical cycle is bounded by erosion surfaces and consists of cross-bedded medium to coarse sub-arkosic sandstone which passes up into micro-cross-laminated very fine sandstone, green siltstone and finally green mudstone. The basal sandstone may be conglomeratic, containing quartz pebbles, mudstone and siltstone intraformational clasts, phosphatic pebbles and fossil bones of the *Lystrosaurus* fauna (Elliot et al., 1970; Kitching et al., 1972). Locally the lower Fremouw is a massive quartz sandstone (Barrett, 1969). The lower Fremouw passes up into 300 m of alternating mudstone and fine volcanic sandstone in which root horizons are locally abundant. Air-fall vitric tuffs first occur in this part of the Fremouw. The upper Fremouw consists of about 300 m of slope-forming, greenish-grey volcanic sandstone with minor carbonaceous siltstone and mudstone. The sandstone is mainly parallel-bedded or has low-angle discordant bedding. The level at which carbonaceous matter comes in is extremely variable. Locally there are coal seams, root horizons, stems, and logs up to a metre in diameter and 22 m long. *Neocalamites* is prominent in some sections in the middle upper Fremouw, and *Dicroidium* has been found in the upper Fremouw.

The contact with the overlying Falla Formation is disconformable at the few observed localities. The thickness of the Falla varies widely (Table 37.1). At the type section (Barrett, 1969, 1972a) the lower 270 m consists of a cyclic sequence of sandstone and shale, the cycles being from 5 to 54 m thick. The sandstone rests on local erosion surfaces and passes upward into greenish-grey fine sandstone, thin carbonaceous shale with, locally, *Dicroidium*. Above the cyclic beds, tuff is increasingly important and forms the entire upper 110 m.

The Prebble Formation (Barrett and Elliot, 1972), which is probably disconformable on the Falla, outcrops only in the Queen Alexandra Range and the Otway Massif area. It has varied lithology, consisting of laharc debris, pyroclastic breccia, tuff and tuffaceous sandstone. Lahar deposits predominate over other lithologies and are poorly sorted rocks with slightly rounded clasts up to 80 cm across set in a tuffaceous matrix. Acidic volcanic clasts are more common than sedimentary and basaltic rock fragments, although any one clast type may be dominant. The sedimentary clasts consist of Beacon lithologies, including sparse carbonaceous shale and coal. Basaltic clasts are like the Kirkpatrick Basalt and Ferrar Dolerite. Accretionary lapilli are common in some sections and indicate close proximity to the volcanic source. One vent-like feature outcrops in the Queen Alexandra Range. The Prebble is included in the Beacon Super-Group because the composition of the volcanic material is similar to that in the upper Falla. Nevertheless, there is evidence of local disconformity at the base of the Prebble.

North of the Queen Alexandra Range Triassic strata are represented only by lower Fremouw sandstone bluffs which outcrop in the southern half of the Queen Elizabeth Range.

Triassic strata are well developed in the Queen Maud Mountains. It should be noted that the stratigraphy described by Grindley (1963) and McGregor (1965) has been revised by Barrett (1969, 1972a). The upper part of the Falla Formation of Grindley is equivalent to Barrett's Falla Formation. The Dominion Coal Measures appear to be part or all of the upper Fremouw and lower Falla of Barrett. La Prade (1970) described the Triassic strata in the Shackleton Glacier area as the Mount Kenyon Formation, but a revision of the stratigraphy (Collinson and Elliott, in prep.) will propose introducing Barrett's nomenclature. The threefold division of the Fremouw is recognisable in the Shackleton Glacier area. The lower part has an abundant *Lystrosaurus* Zone fauna, and *Dicroidium* has been recovered from the upper part of the formation. Similar rocks outcrop on Mount Fridtjof Nansen, and...
probably form most of the high peaks in the Queen Maud Mountains. Correlative strata, the Nilsen Formation (Long, 1965b, unpubl. ms.), consisting of sandstone with lesser amounts of conglomerate, greenish-grey mudstone, siltstone and shale outcrop in the Nilsen Plateau. The conglomeratic beds contain a wide variety of pebbles and cobbles of igneous, metamorphic and sedimentary rock types.

The Falla Formation has been recognised around the upper part of the Shackleton Glacier (Collinson and Elliot, in prep.) including Misery Peak where *Dicroidium* is present (Townrow, 1967). It may also be present in the highest peaks such as Mount Fridtjof Nansen.

In both formations palaeocurrent vectors indicate flow from the southeast (Fig. 37.6A), in marked contrast to that in the Permian Buckley Formation (Fig. 37.5). This change in palaeocurrent direction and the renewed sediment influx in the upper Fremouw are related to orogenic activity which is discussed later.

**Victoria Land** A thick Triassic section also outcrops in south Victoria Land (Table 37.1). The Permian-Triassic boundary has been tentatively placed between the lower and upper members of the Feather Conglomerate (Barrett and Kohn, this volume), in the belief that the 90° change in palaeocurrent direction at this level correlates with the reversal of palaeoslope at the Permian-Triassic boundary in the Beardmore Glacier area. The upper part of the Feather Conglomerate (Fleming Member or Formation of Barrett et al., 1971, and McKeelvey et al., 1970) consists of fining-upwards cycles of cross-bedded coarse to fine quartzose sandstone and green siltstone. The Fleming is equivalent to the upper part of the Robison Peak Formation of Matz et al. (1972). The Fleming is overlain by more than 500 m of the Lashly Formation, which Barrett and Kohn (this volume) have divided into four members. The lowest, Member A, consists of fining-upwards cycles of thin fine-grained sandstone grading up into much thicker carbonaceous siltstone and claystone. Root horizons and calamitid stems are common. This is equivalent to the Mount Bastion Formation of Matz et al. (1972). *Dicroidium* (Barrett et al., 1971), *Glossopteris*, *Noeggerathopsis* and *Paracalamites* (Matz et al., 1972) have been recovered from this unit. Member B consists mainly of massive, cross-bedded feldspathic and lithic sandstone, but includes a few thin coals and sparse *Dicroidium*. Member C comprises fine sandstone, siltstone and thin coals, and has yielded abundant *Dicroidium*, and Member D is similar to Member B. Lashly B, C and D are equivalent to the Allan Nunatak Formation of Matz et al. (1972). The Lashly Formation is overlain disconformably by the Mawson Formation.

In the Priestley Glacier area (Skinner and Ricker, 1968), Triassic strata consist of cross-bedded arkosic sandstone, locally conglomeratic at the base of the section, together with finer-grained micaceous sandstone, siltstone, mudstone, carbonaceous siltstone and shale, and sparse thin coal. At Timber Peak there are silicified logs and *Vertebraria*, and a thin coaly shale has yielded late Triassic microfossils (Norris, 1965). Further north, the Beacon is reduced to some 15 m of cross-bedded arkosic sandstone with pockets of well-rounded quartz pebbles at the base (Gair, 1967). Similar sediments are interbedded with the overlying basalt (see p. 517). Nathan and Schulte (1968) recorded as much as 80 m of quartzite and feldspathic sandstone with a prominent basal conglomerate at the head of the Campbell Glacier. Mudstone, shale and minor coal are interbedded with the sandstone. Clasts of palagonitised scoria and basalt occur in some sandstones. The most northerly outcrops of Triassic strata are in the Freyberg Mountains and Morozumi Range (Sturm and Carryer, 1970), where they occur as rafts up to 20 m thick in Ferrar Dolerite, and in sections resting on basement rocks. The sediments are quartzose and arkosic sandstone with minor laminated siltstone and mudstone, some of which is carbonaceous. One 15 m section resting on basement granite has yielded late Triassic plant microfossils (G. Norris, in Gair et al., 1969).

Palaeocurrent indicators demonstrate a marked swing in palaeoslope between the lower and upper parts of the Feather Con-
glomerate (Figs. 37.5 and 37.6A), from west-northwest to near north (Barrett and Kohn, this volume). The northerly palaeoslope con­tinued during Lashly time but gradually swung back to northwest. Palaeocurrent data are not available for the thin Triassic se­quence in north Victoria Land.

**Palaeontology and Age** The lower Fremouw Formation is notable for its vertebrate fauna. Part of a labyrinthodont jaw was found at Graphite Peak (Barrett et al., 1968). Subsequently more material was discovered at Coalsack Bluff (Elliot et al., 1970), principally in the conglomeratic lenses in the lower parts of the fining-upwards cycles. The only specifically identified tetrapod so far from Coalsack Bluff is *Lystrosaurus murrayi* (Col­bert, in press), but there are also recognisable amphibian, thecodont and other therapsid fossil bones. A more abundant fauna was recovered in the McGregor Glacier area (Kit­ching et al., 1972). Here, examples of *Lystrosaurus* are again abundant, the identi­fied species being *L. curvatus* (Col­bert, in press). In addition, the small theriodont, *Thrinaxodon cf. liorhinus* and the cotylosaur *Procolophon cf. trigoniceps* are present. Other material includes both large and small labyrinthodont amphibians (Colbert and Cosgriff, in prep.) and numerous small prola­certid eouschians. This fauna has consider­able similarity with the *Lystrosaurus* Zone of the lower Beaufort of South Africa, and hence is assigned an Early Triassic age.

*Dicroidium*, of which *D. odontopteroides* appears to be the most abundant species, is found in the upper Fremouw and lower Falla of the Queen Alexandra Range (Bar­rett, 1969, 1972a). It is also present in the Falla Formation at Misery Peak, and an ex­cellent flora has been found in erratic boul­ders in a recent moraine at Mount Burnstead (Townrow, 1967). Abundant *Dicroidium* is also present in Triassic sections in the Nilson Plateau area (H. E. Ehrenspeck, pers. comm.). A peat deposit from the upper Fre­mouw is under study by J. M. Schopf. In south Victoria Land, *Dicroidium*, of which *D. odontopteroides* is again the commonest species, is present in the Lashly Formation (Members A, B and C) (Barrett et al., 1971; Barrett and Kohn, this volume). There are earlier records of *Dicroidium* from near Mac­kay Glacier in beds now assigned to the Lashly Formation—Allan Nunatak Forma­tion of Matz et al. (1972); Triassic plant beds described by Gunn and Warren (1962); see also Plumstead (1962), and Tow­nrow (1967). The *Dicroidium* floras recovered so far have been assigned a Middle and Late Triassic age by Rigby and Schopf (1969) and Tow­nrow (1967), though an Early or Middle Triassic age was favoured by Plum­stead (1962). Thus, the lower Fremouw is Lower Triassic, whereas the macroflora­bearing upper Fremouw and Falla are Middle and Upper Triassic, as is the Lashly Formation.

Plant microfossils have been reported from a few localities. Those from the Fremouw of the Queen Alexandra Range are being studied currently (Fasola, pers. comm.). The microflora from Member C of the Lashly at Mount Feather has a late Middle Triassic age (Helby and McElroy, 1969). In northern Victoria Land, palynomorphs from the Tri­assic at Timber Peak have been assigned a Middle to Late Triassic age (Norris, 1965) and a similar age is reported for plant micro­fossils from the Freyberg Mountains (Nor­ris, in Gair et al., 1969). It is possible that the microfossils from Horn Bluff are Triassic (see p. 530).

**FERRAR GROUP**

The Beacon Super-Group is intruded by dolerite sills and dykes and is capped by a variety of rocks, chiefly basaltic lavas, as­signed to the Ferrar Group (Grindley, 1963). The diabase sills are ubiquitous in the Transantarctic Mountains and have been recorded from Horn Bluff to Vestfjella (Fig. 37.6B). The dolerites range from 191 m.y. to as young as 147 m.y. (Compston et al., 1968; Elliot, 1970; Ford, 1972; Rex, 1967, 1972; Wade et al., 1965; Webb and Warren, 1965). The lavas have as wide a distribution as the sills (Table 37.1; Fig. 37.6B) and nearly as great an age range, 179-148 m.y. (Borns et al., 1972; Elliot, 1970; Rex, 1972; Wade et al., 1965). Excluding a 220 m.y. old dolerite cut­ting Permian sediment in the Pecora Escarp­ment (Ford, 1972), the spread of dates may
reflect, to some extent, events subsequent to intrusion.

From a stratigraphic viewpoint the most interesting aspect of the basalts in the Beardmore area is the presence of thin interbedded sediments which have yielded conchostracans (Elliot and Tasch, 1967; Tasch, 1969, 1970a, 1970b) and holostean fish (Schaeffer, 1972). In south Victoria Land basalts and related rocks were first described by Gunn and Warren (1962) and because of the report of a Jurassic tillite have been the subject of considerable re-investigation (Borns and Hall, 1969; Hall and Borns, 1972; Ballance and Watters, 1971). Along the edge of the polar plateau, the Beacon is capped locally by a diamicite which overlies strata as old as Permian, indicating considerable erosion subsequent to deposition of the Triassic sediments. The diamicite (Mawson Diamicite of Ballance and Watters, 1971; Mawson Formation of Barrett et al., 1972, and Borns and Hall, 1969) is largely a laharic deposit of basaltic composition, in contrast to the Prebble Formation lahars which are acidic (Barrett and Elliot, 1972). The section at Carapace Nunatak is somewhat different; sedimentary beds (Carapace Sandstone of Ballance and Watters, 1971) are overlain by basalt lavas that have an intercalated unit of pillow basalt and a unit of Mawson lithology. Plants, ostracods and conchostracans have been found at the top of the sedimentary section. Thin lacustrine sediments beneath the pillow basalt unit have yielded a variety of crustacean and insect fossils (Borns and Hall, 1969; Borns et al., 1972). Plumstead (1962) and Townrow (1967) reported on the plants and the latter suggested a Middle Jurassic age.

Further north the basalts are either not observed in contact with the Beacon or conformably overlie thin Upper Triassic strata (Skinner and Ricker, 1968; Gair, 1964, 1967; Gair et al., 1969). A 20 m-thick sedimentary interbed about 180 m above the base of the basalts at Section Peak consists of cross-bedded sandstone with disseminated carbonaceous matter and thin coaly lenses from which Early Jurassic plant microfossils have been recovered (Norris, 1965).

Basins of Deposition and Relations with Other Gondwana Continents

The distribution and thickness of the outcrops of Gondwana strata are such that only the evolution of basins in the Transantarctic Mountains between south Victoria Land and the Pensacola Mountains can be discussed meaningfully. Further it is recognised that the data on which this interpretation is based are far from adequate.

The glacial strata will be discussed because of possible relations with the Antarctic Peninsula and other Gondwana continents, but the deposition of the glacial beds is not necessarily related to the basins of deposition recognisable in later Permian time. The post-glacial Gondwana strata of the Transantarctic Mountains will be discussed in terms of a model in which the stratigraphic record is visualised as largely controlled by orogenic uplift in west Antarctica. Two Permian basins are recognised, tentatively named the Nimrod-Ohio Basin and the South Victoria Basin, and one Triassic, tentatively named the Nilsen-Mackay Basin. The Jurassic igneous rocks will also be discussed briefly because of intercontinental relations.

Late Carboniferous and Early Permian Glacial Strata

Frakes et al. (1971) and Crowell and Frakes (1970) postulate a migration with time of the ice centres across Gondwanaland. Glaciation in Antarctica is thought to have started during the Early Carboniferous with an ice centre in the Weddell Sea which deposited much if not all of the glacial strata of the Pensacolas. The same ice centre presumably gave rise to the glacial strata along the Natal Coast which indicate a source lying to the east and therefore in Antarctica (Crowell and Frakes, 1972). This ice centre has left no other clear evidence of its existence in Antarctica except, possibly, for the tillite reported from Heimefrontfjella which is considered to be Upper Carboniferous (Adie, 1970; Plumstead, 1970). During the Late Carboniferous glacial beds were deposited in a marine environment in the Ellsworths, but in a non-marine environment in the Ohio Range, where westward flow of ice is postulated. Finally ice flowed along the
length of the Transantarctic Mountains away from Victoria Land in the Early Permian.

Crowell and Frakes (1971a, 1971b) present evidence for ice centred to the south and west of Tasmania in the Late Carboniferous, with a considerable expansion into Australia during the Early Permian. Ice may have persisted till Late Permian time. The glacial centre for all this time would have been in north Victoria Land and strongly suggests that in the Transantarctic Mountains only the waning stages are recorded.

Lindsay (1970), on the other hand, has suggested that all the glacial strata in Antarctica can be related to one ice centre in Victoria Land, assuming Long's (1965a) data are valid rather than those of Frakes et al. (1966), and that the direction of flow in the Pensacolas is uncertain (Schmidt and Williams, 1969).

Only in the Ellsworth Mountains and possibly the Pensacola Mountains is a marine environment envisaged. Whatever the true relative position of the Ellsworths at that time, the marine environment would have been an arm of the sea that is documented by marine fossils in the Paraná Basin (Crowell and Frakes, 1970) and South West Africa (Martin and Wilczewski, 1970) (Fig. 37.7). The thickness of the Dwyka Tillite in the southern Karroo (Crowell and Frakes, 1972) suggests a subsiding basin, which may also have been marine. The marine environment in the South Atlantic sector of Gondwanaland was probably an epicontinental sea. Its extent across Antarctica is uncertain because of the absence, except in the Antarctic Peninsula, of Carboniferous strata, but consideration of the Devonian geology in the Transantarctic and Ellsworth Mountains and Marie Byrd Land suggests it was possibly confined to the Weddell Sea sector. Devonian marine beds are known only from the Ohio Range and from the Ellsworth Mountains which have a long history of Palaeozoic subidence.

The Trinity Peninsula Series and possibly correlative strata in the Antarctic Peninsula (Dalziel and Elliot, 1973) are in part Carboniferous. They are predominantly interbedded shale and quartzose greywacke of turbidite aspect, with other lithologies very subordinate. A few pebbly mudstones have been recorded (Elliot, 1965, 1966; see also Halpern, 1965) but there is no evidence as yet in favour of a glacial rather than a mudflow origin. The provenance of the strata is largely granitic but with some volcanic and metamorphic rocks. The thickness has been estimated at more than $13,500$ m but tight folding precludes an accurate determination. The thickness, lithology and the lack of basement to the sequence favour a rapidly subsiding basin or a deep oceanic environment, and it seems more likely that the sediments were deposited along the 'Pacific' margin of Gondwanaland rather than in an inland sea. It is possible that the turbidites of the Trinity Peninsula Series were derived in part from the late Palaeozoic glacio-marine strata;
the petrology of the diamicites in the Ellsworth and Pensacola Mountains (Frakes et al., 1971; Williams, 1969) and the turbidites (Elliot, 1965, 1966; Aitkenhead, 1965) is not inconsistent with a common source area.

By Early Permian time, the main ice centre lay in north Victoria Land although it may have been initiated in the Late Carboniferous. The upper parts of the glacial strata are probably Sakmarian, but whether all the glacial beds are so young is open to question, even though the bulk of glacial strata are deposited during the waning stages of glaciation. Further, there is no way yet of demonstrating whether, for instance, the thirteen separate tillites recorded by Lindsay (1970) represent fluctuation in the ice front in that area during the final retreat or whether they represent significant readvances of the ice spread over tens of millions of years. Frakes et al. (1971) suggest at least four cycles of glaciation.

Radiometric dating (Craddock, 1970a) suggests metamorphism in the Thurston Island area (Figs. 37.1 and 37.7) and this area through Carboniferous and Early Permian time may also have been above sea level. There is no other direct evidence of events in West Antarctica at this time but the depositional basin inferred for the Mackellar would suggest a topographically elevated area confining it. The geanticline may have petered out towards Africa and South America or it may have continued through, with the Trinity Peninsula Series being deposited on its 'Pacific' side in an unrelated basin, and by Permian time with the Ecca turbidites of the southern facies being shed into the Karroo Basin on the continental side (Ryan, 1969). Further data on this point are included in the caption to Figure 37.7. There are no recorded strata comparable with the southern Ecca in Antarctica and the continuation of the Ecca Basin southeastward from Africa remains uncertain. Although the Trinity Peninsula Series is considered to have been deposited marginal to Gondwanaland (Dalziel and Elliot, 1971, 1973), the relation to the cratonic areas is uncertain. Arguments have been advanced (Dalziel and Elliot, 1971, 1973; Frakes and Crowell, 1968) in favour of some strike slip displacement between East and West Antarctica, and this would bring the northern part of the Antarctic Peninsula towards the base of the Weddell Sea. Again, if the Ellsworth Mountains should be rotated in order to rationalise the structural trends with those of other elements of the Gondwanian Orogen (Elliot, 1972; Schopf, 1969) it would seem most logical to place them further toward (or into) the Weddell Sea so that the stratigraphic section in the Pensacola Mountains would represent an active but continental environment, except possibly for the Gale Mudstone, whereas the Ellsworth section would be the equivalent wholly marine sequence. Any simple explanation like this fails to account for the fact that the Pensacola Mountains were involved in late Precambrian and early Palaeozoic orogenic episodes, neither of which are recorded in the Ellsworth Mountains. This geanticline may also have continued through to the foreland to the New Zealand Geosyncline (Barrett et al., 1972; Brown et al., 1968). Considerable volcanism is recorded in this geosyncline during the Early Permian. A marine environment persisted on the continental side of this foreland as shown by the marine sediments in Tasmania (see Brown et al., 1968); how far this sea extended towards Antarctica is uncertain, but the interpretation of the data for the Permian of the Transantarctic Mountains would suggest that it was not far.

There is no evidence from Antarctica bearing on the ice centres for glaciation of western Australia or India, even though ice movement vectors suggest Antarctica was a source area.

Post-glacial Permian Strata

Nimrod-Ohio Basin The black shale unit (Mackellar Formation and equivalents) overlying the glacial beds in the central Transantarctic Mountains can be traced for a distance of 1000 km from the Nimrod Glacier to the Ohio Range. The lithology, abundant rhythmically alternating black shale and fine sandstone or siltstone with the coarser sediment becoming more prominent up-section is consistent with a pro-delta and delta slope environment (Selley, 1970). However, it should be noted that this ideal sequence is
Gondwana Basins of Antarctica

not developed everywhere. The abundance of carbonaceous material suggests a body of water with restricted circulation. The remarkable extent of the beds clearly precludes a lacustrine environment and has led to an analogy with the Gulf of Bothnia and the Baltic Sea, an analogy recognised by Du Toit (1937) for the Upper Dwyka Shales. Strong support for such an environment is given by the Sr-isotopic composition of thin limestone lenses (Faure and Barrett, 1973) in the Mackellar Formation of the Queen Alexandra Range. The isotopic ratios are high (0.719-0.723) and indicate that there was no free exchange with open oceanic waters which would have reduced the ratios to that of Permian sea water (0.707). The palaeocurrent vectors are summarised in Figure 37.4B, and although locally variable, indicate flow towards the Ohio Range.

The lower 600 m of the Polarstar Formation in the Ellsworth Mountains is formed of Mackellar-like strata, and may therefore have been deposited at the open-water end of this restricted embayment. The immensely thick marine succession in the Ellsworths shows that it was an area of continued subsidence and concomitant deposition through most of the Palaeozoic.

Lindsay (1970) discussed the post-glacial shales in terms of isostatic rebound following deglaciation. The post-glacial strand line migrated to just north of the Nimrod Glacier on glacial retreat and then migrated back due to isostatic rebound. The interval between deglaciation and return of the strand line with deposition of the Fairchild sandstone is that represented by the Mackellar Formation and its equivalents. The post-glacial shale and fine sandstone is interbedded with thicker bedded and coarser sandstone in the Nimrod area and Queen Elizabeth Range, and these areas were close to the strand line. Elsewhere, parts of the ideal sequence, which represent pro-delta and delta slope environments, may be missing. The palaeotopographic high in the Queen Maud Mountains is reflected in marked thinning in the Nilsen Plateau where these beds are also locally absent. The Weaver Formation east of the Scott Glacier includes considerable sandstone (however, see Katz and Waterhouse, 1970) and Minshew (1967) has postulated a beach and barrier island complex for the depositional environment.

The postulated depositional environment, an elongate body of water extending along the Transantarctic Mountains from the Nimrod Glacier to the Ohio Range, must be confined topographically. This requires rising land on the cratonic side, East Antarctica, sparse evidence for which is found only in the lithology and palaeocurrent vectors for the Mackellar of the Queen Elizabeth Range (Lindsay, 1969), and in West Antarctica through which a geanticline may have passed (Barrett et al., 1972; Elliot, 1972) (Figs. 37.7 and 37.8). Such a situation seems to be required also by the Permian coal measures and Triassic fluvial sediments.

The cliff-forming lower part of the Fairchild Formation is mostly parallel-bedded sandstone with some parting lineation, but there are also cross-bedded intervals. Low-angle discordant bedding is locally developed and is often associated with broad scouring and intraformational conglomerates which suggest strong currents. The upper slope-forming part includes micro-cross-laminated sandstone and the sequence as a whole is lithologically more variable. Carbonaceous laminae and coal streaks occur locally. Barrett (1970a, 1970b) suggested deposition from braided distributaries on an advancing and aggrading deltaic plain. More specifically it seems possible that the lower part of the formation represents a delta platform environment in which the sediment was reworked (see Selley, 1970). The possibility that lithostratigraphic equivalents of the Fairchild also include barrier sands or offshore bars (Minshew, 1967) seems unlikely because the palaeocurrent vectors are oriented similarly to those of the coal measures and have only one mode (Klein, 1967). The lithologically more variable upper part of the formation was deposited from distributaries higher on the delta plain, with the fine sediment and carbonaceous matter deposited in local ephemeral interdistributary lagoons. The Fairchild grades up into the floodplain environment of the Buckley Formation without any marked break. The lower part of the Buckley differs little from the upper part of
Fig. 37.8. Palaeogeographic map of part of Gondwanaland for the Late Permian. The Nimrod-Ohio Basin adjacent to the South Pole was separated from the South Victoria Basin by a palaeo-topographic high which was the site of deposition by Late Permian time, and hence the continuous outcrop distribution in the Transantarctic Mountains (see p. 525). See also caption to Fig. 37.7.
the Fairchild, but cyclic deposition is better developed higher in the formation. The cycles consist of cross-bedded coarse to conglomeratic sandstone overlying local erosional surfaces and grading up into micro-cross-laminated fine sandstone, siltstone, carbonaceous shale and coal, a sequence typical of coal-bearing fluvial strata. The variability of current direction higher in the Buckley points to a decreasing palaeoslope, higher sinuosity streams, and a floodplain with meandering streams (Barrett, 1970a).

This vertical sequence of black shale with interbedded fine sandstone overlain by massive sandstone of probable deltaic origin, which grades up into a well-developed floodplain environment of varied lithology but with abundant carbonaceous material and plant remains, can be traced from north of the Nimrod Glacier to the Ohio Range. In the Nimrod area the Mackellar is locally absent and the Fairchild equivalent apparently overlies the Pagoda Tillite. Strata further north cannot be identified with any particular formation, and the head of the basin for at least the Mackellar and Fairchild was, therefore, in this area. During coal measures time the topographic high persisted but it was probably buried by later Permian time. Almost all palaeocurrent vectors indicate flow towards the Ohio Range and this suggests that the known exposures of the Fairchild and equivalents are diachronous. If this is so, the lower carbonaceous beds in the Nimrod area are probably temporal equivalents of the Fairchild and Mackellar further down the basin.

Although there is no direct evidence of Beacon strata adjacent to much of the Transantarctic Mountains, recent radio echo sounding (Drewry, 1972) has demonstrated the existence of mesa topography in the area between the South Pole, the Ohio Range and the Queen Maud Mountains (Fig. 37.5). Beacon strata are likely to be the local bedrock for this topography.

The extension of the Nimrod-Ohio Basin to the Pensacola or Ellsworth Mountains is uncertain. Erratics in the Thiel Mountains suggest that Beacon rocks and Ferrar Dolerite occur sub-glacially to the south (Schmidt and Ford, 1969). In the Ellsworth Mountains, the lower 600 m of the Polarstar Formation is lithostratigraphically similar to the Mackellar and passes up into more sandy strata in which thin coals and Glossopteris have been found. A near-shore marine, possibly pro-delta, environment has been postulated (Craddock et al., 1965). This is consistent with the great thickness of the glacial beds of the Whiteout, the continuity of the section and the lithology, sedimentary structures and thickness (1500 m) of the Polarstar. Palaeocurrent measurements from the base of the formation indicate derivation from the south-southeast (Matthews et al., 1967) and this may have continued into later Permian time. Argument can be made for the Polarstar being the marine edge of the basin stretching from the Nimrod Glacier to the Ohio Range and beyond, but there is little biostratigraphic control for assessment of the age relations, and lithostratigraphic correlation is valid only for the glacial strata, and possibly for the Mackellar-like strata in the lower part of the Polarstar. Further, there is some question, based on structural trends, whether the Ellsworth Mountains are now in the same position relative to the Transantarctic Mountains as they were during Gondwana sedimentation (see p. 530). Nevertheless, such a marine basin could be the final remnants of the marine embayment in this sector during Late Carboniferous and Early Permian time. It may have been connected to the open ocean by a seaway crossing the West Antarctic geanticline for which continuing thermal events are indicated in the Thurston Island area (Craddock, 1970a). The depositional basin of the Transantarctic Mountains also supports the presence of elevated land through West Antarctica. It should be noted that volcanic detritus is common in the upper part of the Buckley although recent work (Faure and Barrett, 1973) suggests the source area might be Precambrian. Volcanic detritus and bentonite beds are also recorded in the Polarstar and Queen Maud Formations. The bentonite beds suggest contemporaneous volcanism.

The Nimrod-Ohio Basin can be visualised as being flanked by a cratonic area in East Antarctica and a geanticline through West Antarctica. Retreat of the ice allowed marine
waters to enter the isostatically depressed area southward from the Nimrod Glacier and establish a brackish water environment, here tentatively named the Mackellar Sea. Streams emptied into this embayment; fine sediment was deposited away from the shoreline and also out in the probable marine part in the Ellsworth Mountains. The basin was much wider than the present outcrop distribution indicates because enough time elapsed for the deposition of considerable thicknesses of sediment before the delta fronts of the adjacent rivers encroached into the present area of the Transantarctic Mountains. These rivers transported and reworked the debris left by the glaciers and the products of continued subaerial erosion of adjacent land. Gradually the basin was restricted by deltas advancing from the flanks, and finally the central part was filled by advance along the Transantarctic Mountains of the delta plain and distributary channels now represented by the Fairchild. The deltas advanced across the earlier pro-delta and delta slope deposits, and were themselves covered by the alluvial plain deposits that followed in their wake.

The source areas remained far distant or were relatively low because coarse sediment is decreasingly important with time. Streams may have deposited braided alluvial fans near their sources but the only possible evidence of such an environment is over the palaeotopographic high south of the Byrd Glacier where the unfossiliferous coarse arkosic sandstone sequence may be the temporal equivalent of part of the coal measures.

The coal measures developed on a broad floodplain with streams close to base level, and with low relief in adjacent areas. Vegetation accumulated in marshy areas and abandoned channels eventually to form coal. The streams flowed along the length of the Transantarctic Mountains, out into deltas and then the open waters of the Mackellar Sea. The deltaic deposits built out rapidly across the stable continental part of the basin but in the subsiding marine part were probably reworked by oceanic and turbidity currents into the coarser part of the Polarstar Formation that overlies the Mackellar-like strata. Despite the aggradation of more than 700 m of floodplain deposits, the delta front never reached the Ellsworth Mountains, because there the Glossopteris-bearing beds are considered to be pro-delta deposits.

The Mackellar Sea, probably an epicontinental sea, was isolated from the oceans some time in Late Permian or earliest Triassic time by the rising Gondwanian Orogen. Probably the rivers continued to flow but constituted part of an internal drainage system that still deposited floodplain sediments along its meandering courses.

South Victoria Basin The Weller and Misthound Coal Measures have a high proportion of sandstone and only subordinate amounts of finer beds including carbonaceous shale and coal. The floodplain cycles that characterise the Buckley Formation are apparently better developed northwards to Allan Nunatak (Matz et al., 1972; Mirsky et al., 1965) whereas in the south, sandstone predominates. Barrett and Kohn (this volume) postulate that the sediment was largely derived from glacial debris and transported from the Ross Ice Shelf area. The high variability of the palaeocurrent direction is similar to that of well-developed floodplains, but there is a remarkable lack of fine sediment. The interpretation of this is uncertain but it may be related to low post-glacial topography interfering with stream flow. The Feather Conglomerate which overlies the Weller is conglomeratic in the south but becomes finer to the north where it is almost entirely sandstone. It would appear to be a pediment deposit.

North of the watershed identified by Barrett and Kohn (this volume) the palaeocurrents indicate derivation from the Ross Ice Shelf area, with high sinuosity streams depositing the Weller and low sinuosity streams giving the braided stream deposits of the Feather. South of the watershed, stream flow was again highly variable as it was for the lower part of the Weller to the north. Flow was to the south and in the Darwin Mountains 150 km further south the flow was to the southeast. Regional interpretation for the area south to the Nimrod is hampered by lack of data. Variable directions are likely because that area was a per-
sistent topographic high throughout the Permian and not part of a well-defined basin.

The sequence of events that can be established for the Nimrod-Ohio Basin has no counterpart in the much less extensive South Victoria Basin. Neither the Mackellar nor the Fairchild can be recognised, and the coal measures differ in lacking the fine sediment that characterises the upper Buckley. The time represented by the Pyramid Erosion Surface is uncertain but may coincide with deposition of all or part of the Mackellar and Fairchild Formations. On the other hand, the lower parts of the Weller and Mithound may be temporal equivalents of either or both of those formations.

The South Victoria Basin probably was confined by the low-lying or distant West Antarctic geanticline, the topographic high of north Victoria Land, and by rising land in East Antarctica. Streams flowed generally away from the geanticlinal area carrying coarse sediment that was deposited in extensive meanders. Coal swamps developed locally but were never as extensive as in the floodplain of the Nimrod-Ohio Basin. The streams passed west of the north Victoria Land high, which must also have been a local sediment and stream source. An extension of this basin along the coast of Antarctica cannot be demonstrated. Rising mountains of the Gondwanian Orogen then shed coarse sediment which was spread by braided streams across the earlier floodplain of the Weller Coal Measures. These braided streams built up alluvial deposits of conglomerate which passed northward into sandstone. This basin ceased to exist as a separate entity by late in the Permian or early in the Triassic, and merged into the single Triassic basin which is described later.

Other Basins In the Theron Mountains an alluvial plain was built up by streams flowing from the southwest, possibly from the rising land that also restricted the epicontinental Mackellar Sea in which the marine beds of the Ellsworths were deposited. Relations, if any, with the Pensacolas and the Shackleton Range are uncertain.

Across the continent a large basin developed in the Prince Charles Mountains area. Coarse conglomerates were deposited close to a mountain front consisting of high-grade metamorphic rocks, but with continued erosion the mountain front retreated and a floodplain with coal swamps developed. Later the coal swamp environment was replaced by an alluvial plain in which conditions were not favourable for plant preservation.

Elsewhere, in western Dronning Maud Land, at Horn Bluff and in north Victoria Land the Permian outcrops all represent parts of non-marine basins of uncertain dimensions, although possibly extensive.

Intercontinental Relations The geanticline which passed through West Antarctica and confined the basins of deposition of the Transantarctic Mountains also passed off southern Africa where it exerted a major influence on the development of the Karroo Basin. During Middle and Late Permian time sediment was shed from the rising mountain front and deposited in marine and non-marine environments in the Karroo Basin. The Permian strata outcropping in western Dronning Maud Land are thin cratonic sequences and there is, as yet, little basis for relating them to events recorded in southern Africa.

The Amery Group, which unconformably overlies Precambrian rocks, constitutes the remnant of a formerly more extensive Permian basin. The tightest bathymetric fit of India to Antarctica would place the Amery Ice Shelf area near the Mahanadi Valley (Dietz et al., 1972; Smith and Hallam, 1970). However, it is likely that the Indian Gondwana strata were not deposited in fault-bounded valleys (Robinson, 1967) and were much more extensive originally. Palaeocurrent vectors for the Raniganj Formation in the Godavari Valley indicate northeasterly flow swinging to north-northwesterly (Robinson, 1967) and possibly derived in part from a source area in Antarctica. The data available for the Amery Group are not inconsistent with those from India; however, there is no clear identification of the Glossopteris-bearing Bainmedart Coal Measures and overlying non-carbonaceous sandy beds, the Flagstone Bench Formation, with any strata in India.

North Victoria Land remained a source
area throughout this time but whether it also was connected to the West Antarctic geanticline is uncertain. Also uncertain is the extension of the Weller Coal Measures basin down palaeoslope toward the coast of Antarctica. However, Horn Bluff strata may be related to the Murray Basin (Brown et al., 1968), but their age is uncertain. Also the thin sequence with Glossopteris in the Freyberg Mountains might be related to the Murray Basin or to the Tasmanian sequence, but again there are insufficient data for correlation. Antarctica provides no data bearing on other Permian basins in Australia.

By latest Permian time the deformation accompanying the Hunter-Bowen Orogeny of eastern Australia had ceased and also the intense volcanism of the Early Permian in the New Zealand Geosyncline had terminated. However, palaeogeography supports the idea that the West Antarctic geanticline continued through to the New Zealand area (Brown et al., 1968).

**Triassic Strata**

*Nilsen-Mackay Basin* Triassic exposures are confined to the Ross Sea sector of the Transantarctic Mountains (see Fig. 37.6A). The Lower Fremouw fining-upwards cycles were deposited by low sinuosity, possibly braided, streams (Barrett, 1970a). The high proportion of fine sediment and the root horizons in the middle Fremouw suggest a floodplain with abundant vegetation, but one crossed by low sinuosity streams (Barrett, 1970a). There was a renewed influx of sediment and a return to a possible braided stream environment for the upper Fremouw, but the alluvial plain was still stable enough for accumulation of carbonaceous sediment and coal. The lower Falla represents a continuation of the upper Fremouw environment and although carbonaceous matter is common, coal was not formed. The petrology of the sediments shows that volcanic detritus, including air-fall vitric tuffs, becomes important in the middle Fremouw. The lower Falla sandstone is quartzose and feldspathic like that of the lower Fremouw, but upper section volcanic detritus and air-fall material become abundant, and accretionary lapilli are recorded, indicating a volcanic source within 100 km and probably as little as 10 km distant. The volcanic regime persisted through Prebble time, and a volcanic neck has been located cutting the Prebble Formation (Barrett and Elliot, 1972). The contemporaneous nature of the source of the volcanic detritus in upper Fremouw sandstones is suggested by the strontium isotopic ratios in interbedded carbonate lenses (Faure and Barrett, 1973).

In south Victoria Land, the upper member (Fleming) of the Feather Conglomerate is considered to be the lowest unit in the Triassic. Like the lower Fremouw it consists of fining-upwards cycles of an alluvial environment. Barrett (pers. comm.) considers it the lithostratigraphic correlative of the lower Fremouw; the lowest member of the Lashly which has abundant root horizons, and Lashly B, with massive sandstone and logs, bear the same relation to the middle and upper Fremouw. A similar depositional environment, an alluvial plain, with periods of renewed influx of coarser sediment is indicated for the Fleming and Lashly Formations. The sediments are predominantly sandy but as yet volcanic detritus has not been recorded in them. In north Victoria Land, the thin Triassic sequences are arkosic and where volcanic detritus is present, it is basaltic and related to the overlying lavas (Gair, 1967; Nathan and Schulte, 1968).

Palaeocurrent data indicate a dramatic palaeoslope reversal in the central Transantarctic Mountains, a reversal that has been related to early Mesozoic orogeny in the Pensacola Mountains (Barrett et al., 1972; Elliot, 1972). The possible significance of phosphate pebbles in the lower Fremouw, the only known source being the Elbow Formation of the Neptune Group in the Pensacola Mountains, has already been discussed (Barrett, 1969, 1972a). The change in palaeoslope in the lower Fremouw suggests a source not previously represented in the strata, and the incoming of volcanic detritus suggests renewed erosion of the area forming the source for the volcanic sediment in the Buckley, or a new volcanic source which might have been contemporaneous (Faure and Barrett, 1973). Increasing grain size in the upper Fremouw reflects renewed uplift,
Fig. 37.9. Palaeogeographic map of part of Gondwanaland for the Triassic. The Nilsen-Mackay Basin coincides with the Transantarctic Mountains. (See also the caption to Fig. 37.7).

and the lower Falla petrology indicates a return to the lower Fremouw source or a new one of similar composition. The palaeocurrent data do not indicate sediment influx
from the flanks of the present mountain range, but this is the case in south Victoria Land where a source to the southeast is indicated for the Lashly Formation. This source was repeatedly uplifted to give the coarser sediment of Members B and D of the Lashly Formation.

The Nilsen-Mackay Basin, formed by the uplift of the Gondwanian Orogen and the reversal of palaeoslope in the Nimrod-Ohio Basin, may be visualised as an alluvial plain bounded by a cratonic area on one side and a rising mountain front on the other (Fig. 37.9). Braided streams flowed away from the mountains depositing coarse alluvial fans, crossed pediment surfaces and then flowed along the length of the basin. In this central part conditions changed repeatedly because of episodic uplift of the mountains, and no floodplain close to base level and comparable to the Permian coal basins ever developed. After the initial uplift and deposition of the coarser sediment, the alluvial plain probably developed a gentler gradient, while along the mountain front volcanoes erupted and the extruded material was reworked into the basin. Conditions changed again with renewed uplift, more coarse alluvial fans were built out from the mountain front, and coarser sediment was deposited by braided streams in the centre of the basin; away from the streams vegetation flourished and coal swamps developed locally. Volcanic activity was renewed, encroached on the centre of the basin and almost completely overwhelmed fluvial processes, blanketing the landscape with pyroclastic debris. Further down the palaeoslope, away from the area of volcanic activity, fluvial conditions continued, with braided streams depositing coarse sediment, and with vegetation still flourishing. This basin was terminated by regional uplift and was then subjected to erosion. Subsequently the whole area was overwhelmed by basaltic rocks.

**Intercontinental Relations** The uplift of the Gondwanian Orogen, which shed the sediments of the upper Ecca and lower Beaufort into the Karroo Basin, intensified in Triassic time. Coarse sediment was poured into the Karroo Basin, and formed wedge-shaped rock units such as the Molteno and Red Beds (Botha, 1969). The Trinity Peninsula Series was probably deformed at this time on the oceanic flank of the orogen (Dalziel and Elliot, 1973). The apparent lack of deformation in the Falkland Islands Gondwana sequence (Adie, 1952a; Frakes and Crowell, 1967) remains an anomaly and suggests that the Falkland Islands are a displaced block.

Early Mesozoic deformation is recorded both in the Pensacola and Ellsworth Mountains (Ford, 1972; Craddock, 1972), and early Mesozoic intrusions outcrop in the Thurston Island area (Craddock, 1970a). Thus there is considerable evidence for the Gondwanian Orogen in parts of West Antarctica, though no evidence has yet been found for it in Marie Byrd Land. It is the anomalous early Mesozoic structural trends of the Ellsworth Mountains which have led to suggestions that that range has been rotated subsequent to deformation. Orogenic activity in West Antarctica is also reflected in the palaeocurrent patterns and lithologies of the Triassic strata.

Deformation in Australia slightly predated that through West Antarctica, the Hunter-Bowen Orogeny being latest Permian (Brown et al., 1968). Triassic strata occur in Tasmania where the sediment source lay to the north, in the Sydney Basin, and in others far removed from any Antarctic connection. The limited outcrops in north Victoria Land, the Freyberg Mountains and at Horn Bluff, have not yielded information allowing any correlation. Nevertheless, north Victoria Land probably remained a regional high until late in Triassic time and formed a source for strata in adjacent areas.

The foreland to the New Zealand Geosyncline was now the site of Permian volcanic activity, while the axial region to the east continued to receive sediment and contemporaneous volcanic material (Brown et al., 1968). The intermediate volcanism on the foreland and possible correlative intrusive rocks are consistent with linking this area with West Antarctica (see also Griffiths, 1971). Deposition in eastern and central Australia was mainly terrestrial, and it appears that, apart from a small area to the
north of Brisbane, marine strata were largely confined to Western Australia and to New Zealand (Brown et al., 1968), the 'Pacific' edge of Gondwanaland.
Jurassic Igneous Rocks

Jurassic tholeiites are widespread from Horn Bluff to western Dronning Maud Land (Fig. 37.10). Most are basaltic sills, but lavas were also erupted and accumulated to great thicknesses. At least locally in south Victoria Land there was considerable erosion before deposition of the Mawson Formation which underlies the basalts.

Correlative dolerite with geochemical peculiarities similar to the tholeiites in the Ross Sea sector (Faure et al., 1972) outcrops in Tasmania (Compston et al., 1968; Heier et al., 1965). There is no such clear correlation between the tholeiites of western Dronning Maud Land and southern Africa, although they are approximately contemporaneous (Fitch and Miller, 1971; Rex, 1972; G. Faure, in Aucamp et al., 1972).

It should be noted that this basaltic igneous activity is in marked contrast to the acidic extrusive and intrusive rocks of Jurassic age, not all of which are necessarily contemporaneous with the basaltic rocks, through the Antarctic Peninsula (Dalziel and Elliot, 1973; Rex, 1972) and West Antarctica (Craddock, 1970a, 1970b). However, these acidic rocks are consistent with the evidence of acidic volcanism in the Triassic rocks, including the Triassic-Jurassic Prebble Formation (Barrett and Elliot, 1972).

SUMMARY

The evolution of the principal depositional basins in Antarctica can be visualised in terms of an elongate basin confined by a cratonic area in East Antarctica and a geanticline in West Antarctica which constituted the Gondwanian Orogen. These are the principal factors in the evolution of the basins.

The basins were, except for the early stages, entirely non-marine. Glacio-marine beds were deposited in the Ellsworth Mountains in an arm of the embayment that penetrated also into Brazil and parts of southern Africa. Post-glacial shales were deposited in brackish water which formed a constricted arm of the earlier marine embayment. By Middle and Late Permian time, the marine embayment was reduced to the Ellsworth Mountains area by uplift off Africa during the early stages of the Gondwanian Orogen. The Beacon strata of the central Transantarctic Mountains represent a regressive sequence of pro-delta, delta plain and alluvial floodplain environments during Permian time. Orogenic activity intensified during the Triassic, and episodic uplift is recorded in the coarse sediment influx in various Triassic formations. Continued high gradients are suggested by the low sinuosity of the streams, and this may reflect continued slow uplift between periods of more intense movement. Contemporaneous volcanic activity is recorded in the Triassic strata and encroached within the Transantarctic Mountains by late Falla and Prebble time, and also reflects the orogenic activity.

Evidence for latest Palaeozoic and early Mesozoic orogenic activity is recorded throughout the Pacific margin of Gondwana-land, though it is not everywhere synchronous. Only in southern Africa and the Ross Sea sector is the control on sedimentation by this orogen so clear. However, the marked disparity between the Gondwana record of western Dronning Maud Land and southern Africa implies that the orogen must have passed well to the west of Dronning Maud Land, the strata there being thin cratonic sequences analogous perhaps to some of the thin sequences in Rhodesia.

The relation between the various fragments surrounding the South Atlantic are still obscure except in the general terms considered here. Questions remaining include the relations of the basement massifs of Patagonia and the Falkland Islands to the Gondwanian Orogen; the location of the missing fragments of deformed strata that occupy the gap between southeastern South Africa, where the fold belt is truncated, and the Pensacola Mountains; the original location of the Ellsworth Mountains; and the relations between the Scotia Arc and Antarctic Peninsula, and Gondwanaland.

There are similar problems with New Zealand and West Antarctica. The truncated early and middle Palaeozoic orogens have been used to locate Australia and Antarctica. The Gondwana strata of north Victoria Land are too thin and limited for meaningful correlation. Similar limitations
apply to the Amery Group and possible cor-
relative strata in India.

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Gondwana Basins of Antarctica


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Tectonic and Sedimentary Framework of Gondwana Basins in Southern Africa

IZAK C. RUST

ABSTRACT
The African Gondwana basins are cratonic depositories which occur in two major tectono-sedimentary terrains.

The Karroo Basin is the type for all African Gondwana basins. Together with some satellite basins it forms one major tectono-sedimentary terrain. It is mainly a cratonic marginal shelf terrain which merged southwards into a tectonically much more active miogeosynclinal or even foredeep environment.

The main sediment source for this basin was the epeirogenically elevated Transvaal craton, but vague and as yet unconfirmed indications of an early southerly sediment source exist. Marine encroachment was from the south or the southeast and at times perhaps two-thirds of the present Karroo outcrop area was a shallow marine environment. Otherwise the sedimentary milieu ranged between lacustrine, swamp, deltaic, floodplain and dunefields.

Towards the later stages of the Karroo Basin the southern zone was severely folded, overthrust from the south, and uplifted. This tectonic event is commonly ignored in Gondwanaland reconstructions. The resultant fold range developed into a prominent southerly source of sediment.

The tectonic activity along the southern margin of the basin produced remarkably few structural effects in the immediate Karroo Basin, but resultant regional disconformities in the Gondwana basins indicate that widespread effects occurred on the African crustal plate.

The other main tectono-sedimentary unit, here referred to as the Zambezian terrain, is a collection of several graben-type yoked depositories in which non-marine sedimentation took place. This terrain probably merged southwestward into the now displaced Madagascan block, but the exact relationship is still obscure. Although the tectonic character of the two terrains differed markedly, the climatic overprint on the sediments is so strong that broad regional correlation of the sedimentary units is fairly easily accomplished. This sort of correlation is assisted by fossil evidence, which likewise accentuates the role of climate, and by structural evidence, which in its basic simplicity highlights the remarkable stability of the African plate over great areas during deposition of the Gondwana sediments.

The depositional cycle was terminated by the development of a complex basaltic shield volcano with late rhyolitic effusives. The basalt apparently flooded the greater part of the sedimentary basins. Possibly the Lesotho-Lebombo-Nuanetsi-Sinjal line represents a major sub-crustal lesion which initiated and localised volcanic activity at a time when Madagascar drifted away from a mainland join opposite Mocambique. The older radiometric ages in the east, and the younger ones in the west are suggestive of a systematic shift in the volcanic activity across the continental plate in the terminal stages of the African Gondwana basins.
INTRODUCTION

This review of the Gondwana basins on the African continent attempts to reconstruct the sedimentary and tectonic framework of basins mainly south of about lat. 10°S. The complex facies changes and correlation problems are the main reasons for excluding the northerly basins. In certain places the time-span of African Gondwana sediments exceeds that of the classical Karroo sequence but in this paper the treatment of the history of the Gondwana basins is confined to the interval between the onset of Dwyka glaciation and the termination of the main basaltic volcanism of the Stormberg. There is a great need for an in-depth analysis of the Gondwana basins of Africa. A vast amount of information is available, much of this in the form of unpublished or restricted reports of the Geological Surveys of African states, unpublished university theses and data accumulated by numerous mining companies.

In compiling this paper I have made use mainly of readily available recently published data, and though I believe the model so constructed is reasonably correct, I am acutely aware of irregular treatment and blank spots. Like Geoffrey Bond (1967) I also want to say that this paper involved a great deal of extraction and paraphrase. Hopefully I have not quoted out of context or without proper acknowledgment, nor paraphrased inaccurately.

TECTONO-SEDIMENTARY TERRAINS

The tectonic framework of any particular basin of deposition and its source areas has a direct and major bearing on the quantity and quality of the sediments which accumulate in it (Krumbein and Sloss, 1963). Each tectono-sedimentary end-product, a ‘basin’, has certain distinctive characteristics that serve to identify it from other ‘basins’. The interpretation of observed characteristics in terms of the original tectonic pattern and depositional history might not always be easy, due to various factors, but the distinctive character of a basin is usually noticeable.

Occasionally it happens, as it seems to have happened in Gondwanaland, that broadly similar tectonic and sedimentary conditions operate essentially simultaneously over large tracts, producing several large or partly disconnected basins which are practically identical in character.

Such a terrain, which may measure several hundred thousand square kilometres, containing a cluster of several near-identical sedimentary basins, is termed a tectono-sedimentary terrain.

It seems as if two such tectono-sedimentary terrains can be recognised in the African fragment of the Gondwanaland plate.

Outline of the Karroo Terrain

The Karroo tectono-sedimentary terrain is a cratonic area of basins and swells, bounded partly along the south by a compressive weal, and along the east by a monoclinal downwarp (Fig. 38.1).

The type Karroo Basin of South Africa is the best known in this terrain. Others are the Botswana and Congo Basins.

The type Karroo Basin is a marginal cratonic shelf which merged southwards into a tectonically much more active miogeosynclinal trough. Several tectonically defined sub-units can be recognised in it. On the shelf area, the Lesotho Rise was for a long time a fairly strong positive element. The transition between the stable shelf area and the tectonically active miogeosynclinal trough in the south is ill-defined; the same applies to the boundary of the so-called Natal Trough. The southern trough received turbidites during Ecca time but eventually it was elevated and deformed, and itself supplied sediments during Stormberg time. The nature of the boundary between the Karroo and Botswana Basins is in some doubt except that it was an upwarp in existence as early as Dwyka time, and still quite conspicuous during Ecca time; but whether the two basins were actually disconnected, especially during Ecca and Beaufort time, is not clear. During Dwyka time the marine incursion (McLachlan and Anderson, 1973) probably flooded the area between the two major basins.

The Botswana Basin, though mostly covered by Kalahari sand, seems to be a miniature of the main Karroo Basin except that the Beaufort sequence is attenuated or apparently missing. The basin merges eastward and northeastward into the Waterberg and Rhodesian Basins, thus providing valuable
stratigraphic links with these areas. To the west the three main occurrences of Karroo rocks in South West Africa form part of the Botswana Basin.

Northwards of the Botswana Basin the Karroo tectono-sedimentary terrain can be extended into the Congo Basin in the Republic of Zaïre, and may include even the small occurrence in Gabon, but these areas are not discussed in detail. Like the type Karroo Basin the Congo Basin contains a thickness of several hundred metres of Karroo sediments, but the tectono-sedimentary environment may have been more complex (Furon, 1963; Veatch, 1935). The Gabon Basin is the only one which contains Karroo-age evaporites (Micholet et al., 1970).

Outline of the Zambezian Terrain

The boundaries of the Zambezian tectono-sedimentary terrain are rather ill-defined, but the terrain includes the major part of Africa east of longitude 25°E. (Fig. 38.1). Structurally the area is quite complex, but it is characterised by a series of disconnected yoked fault-basins which apparently started their existence as partly interconnected downwarp basins. The Madagascan block is considered part of the Zambezian tectono-sedimentary terrain.

The type basin in the Zambezian terrain is crudely T-shaped, the bar being the Mid-Zambezi-Luangwa Basin complex, the leg being the Lower-Zambezi Basin in Mocambique. The Mid-Zambezi-Luangwa structure

Fig. 38.1. Karroo Basins and tectono-sedimentary terrain. The basic basin character in the Karroo tectono-sedimentary terrain. K is an open basin and swell structure. 1 is the type Karroo Basin; 2 is the Botswana Basin; 3 is the Etjo Basin; and 4 is the Congo Basin with the Kafue embayment. The Gabon Basin (not shown, but see Fig. 38.3 for location) probably also belongs to the Karroo terrain. Z is the Zambezian tectono-sedimentary terrain in which fault tectonics played a major role in the development of the basins. 5 is the mid-Zambezi Basin; 6 is the Luano-Luangwa Basin complex; 7 is the lower Zambezi Basin; 8 is the East African Basin complex; and 9 is the Madagascan Basin.
Gondwana Basins and Continental Margins

is not quite continuous, and in addition it changes in aspect from a fairly simple trapdoor-type fault-basin in the southwest to a regular graben in the northeast, but this may be due to post-Karroo tectonics. A major source of sediment was the Choma-Kaluma ridge and its extensions; this ridge delivered sediment not only to the southeast, but also into the Barotse Basin, which may have been part of the Congo Basin. The Choma-Kaluma ridge could therefore mark the boundary between Karroo-type and Zambezian tectono-sedimentary terrains.

The Bulawayan block was a fairly stable granitic craton which was covered with a thin mantle of sediment only towards the very end of Karroo sedimentation, having previously always been slightly uplifted. The Zoutpansberg-Tuli Basins lie along the southern margin of the Bulawayan block and although these basins are completely separated from other Zambezian basins, they seem to have a similar tectono-sedimentary aspect.

The narrow Nyasa rise separates the Luangwa Basin from the Tanzanian Basin complex comprising the Rukwa, Ruhuhu, Ruaha and Tanga Basins. The tectonics and sedimentology of these basins are not well known but it seems as if situations comparable to those of the type Zambezian existed also in the Tanzanian and related basins.

It seems as if the sedimentary basins in the Zambezian terrain were initially asymmetrical downwarps much larger than the parts now preserved, and that late Karroo faulting produced the elongated graben basins, the structure of which was further accentuated by post-Karroo faulting.

The Madagascan Basin is at present only half a basin, the opposite half being practically unknown. The preserved basin consists of a series of recurrent grabens more or less parallel to the structural and topographic backbone of Madagascar (Cliquet, 1957). This eastern ridge was clearly a major source of sediment which was dispersed to the west. The Karroo sediments of the Madagascan basins are distinguished by several marine interbeds; these play an important rôle in the reconstruction of Madagascar’s position in Gondwanaland as well as the subsequent movement of Madagascar. The Lebombo monoclinal downwarp on mainland Africa may mark the surface expression of one of the major faults in the Zambezian terrain, namely the rift structure along which Madagascar separated from Africa.

The contrast between the Karroo tectono-sedimentary terrain and the Zambezian terrain is therefore distinct. The Karroo terrain is characterised by broad, open basins with intervening swells. The Zambian terrain is distinguished by its tensional fault tectonics. With the single exception of the Cape fold belt, the entire African continent with its Karroo cover has not been subject to any large-scale deformation or metamorphism.

The Dwyka Tectono-sedimentary Regime

Indications are that Dwyka glaciation started during Early Carboniferous or even earlier, and ended more or less during Early Permian (Crowell and Frakes, 1972). The movement of the magnetic pole during this interval was from northwest swinging to southeast across southern Africa. There seems to be no question that this pole movement should be interpreted in terms of the drift of the African continent past the rotational pole. The Dwyka glaciation is therefore a climatic response to tectonic displacement of the continental plate.

A major ice sheet centred over Zambia, Rhodesia and Transvaal, radiated several lobes (Frakes and Crowell, 1970a; Crowell and Frakes, 1972) (Fig. 38.2). Although this area was probably elevated somewhat, the topography was apparently fairly mature, most workers reporting topography of the order of 100 m to 150 m or less (Drysdall and Weller, 1966; Watson, 1958; Bond, 1967; Tavener-Smith, 1962; Frakes and Crowell, 1970a) but others refer to ‘rugged topography’ (Denman and Money, 1970), ‘very uneven’ (Haughton, 1963), and even ‘mountainous’ (Dixey, 1937), though Drysdall and Weller (1966) pertinently refute this last statement with reference to the Luangwa area. The eastern periphery of the Congo Basin was a glaciated mountainous area (Frakes and Crowell, 1970a) situated more or less in the region now occupied by the western rift lakes. Veatch (1935) advanced argu-
Fig. 38.2. Dwyka glaciation. The central core of the Dwyka ice cap may have been continuous from the Congo to Rhodesia as shown here. The radial movement of the ice is well documented and took place during several advances and retreats. Different lobes can be recognised. A marine shelf environment existed to the south.

ments that the Dwyka topography in the Congo Basin sloped gently upwards to the north. This corresponds with data from Gabon (Micholet et al., 1970) indicating ice movement from the east.

Along the eastern boundary of the main Karroo Basin an ice centre lay off the present coast of Natal (Matthews, 1970). Likewise an extra-continental centre was developed on some presumably elevated land to the west of the Karroo Basin (Stratten, 1968). In the case of the western source there is strong evidence that elevated land existed in that area from at least Ordovician to Early Devonian times.

The actual centres of deposition of glacial debris in central Africa were most probably determined mainly by pre-existing drainage and by the location of lowlands produced by ice scouring. Several glacial advances and retreats have been recorded in various places and (Frakes and Crowell, 1970a) and no doubt the isostatic rebound of the crust following the gradual reduction of the ice load caused extensive denudation and redeposition. The Dwyka sedimentary record in the Zambezian terrain is therefore rather confusing, and simultaneous events are difficult to recognise and correlate. It may be significant to note the progressive importance of ordinary coarse clastics at the base of the Karroo sequence towards Kenya in the extreme northeast. The absence of glacigene sediments indicates absence of glaciation in the extreme northeast. The coarse clastics may represent fluviatile outwash debris.

In the Karroo tectono-sedimentary terrain major downwarps developed quite early in Dwyka time and paved the way for the accumulation of glacigene sediments up to 1000 m thick. These sediments accumulated
by various processes, such as ground moraines, end moraines, mud slumps, gravity sliding, rafting, fluvioglacial processes and so on. A similar tectono-sedimentary condition might have existed in the Congo Basin, where almost 600 m of glacigene sediments representing at least five glacial pulses have been recorded (Furon, 1963). These high thickness values contrast sharply with the thicknesses of a few tens of metres usually measured in the Zambezian terrain. The great thicknesses are entirely due to the development of downwarp basins; in these basins isostatic rebound never operated to reduce the original thickness through erosion.

The interpretation of the Madagascan glacigene sediments (lower Sakoa sediments) in terms of source area and dispersal pattern seems fraught with ambiguities (Frakes and Crowell, 1970a). However, if a Gondwanan position of Madagascar is assumed opposite Mocambique, and not opposite Tanzania, the ice flow pattern for Natal (Matthews, 1970) becomes logical, and the observed non-glacial northeasterly dispersal trend in the Madagascan Basin (Frakes and Crowell, 1970a) can be interpreted as being due to the effects of the elevated area north of Swaziland (Hunter, 1969; Matthews, 1970).

During and after the final retreat of the continental ice sheets over Africa an undetermined amount of isostatic rebound occurred. Several workers have briefly referred to this process and invoke this uplift as the operative mechanism responsible for the major denudation of the original ground moraine and other related deposits on the Zambezian terrain. Obviously, however, the effects of isostatic uplift were relatively soon cancelled by the tectonically induced downwarping of the Ecca time and it must be presumed, therefore, that this transition some time during Late Carboniferous and Early Permian marks the beginning of negative tectonism in the Zambezian tectono-sedimentary terrain.

In areas where tectonic basins formed even before the glaciation, especially in the area occupied by the type Karroo Basin (Rust, in press) and possibly elsewhere in the Karroo tectono-sedimentary terrain, and in areas where little ice actually rested on the crust, isostatic rebound was negligible or absent. In the type Karroo Basin the continuation of tectonically controlled downwarping obviously proceeded unchecked, as recorded in the distribution of the late Dwyka shale sequence (Crowell and Frakes, 1972; McLachlan and Anderson, 1973; Winter and Venter, 1970).

It is significant to note that the loci of maximum thickness of the shale unit are northward of those of the Dwyka Tillite. This is in keeping with the regional tectonic situation which had operated since the Ordovician along the southern boundary of the African continent, namely a progressive northward shift of the axis of maximum downwarp (Truswell, 1970).

Tectonic activity was not confined to the southern belt, but operated also in the west. De Villiers (1944) has suggested that the western tectonic episode in fact slightly predates the main southern event.

The regional situation of the marine incursion into the Dwyka Basin (McLachlan and Anderson, 1973) is obscure, but the main oceanic connection was probably to the southeast.

The Ecca Tectono-sedimentary Regime

The Lower to Middle Permian Ecca sequence consists of water-laid clastic sediments part of which contain vast reserves of coal (Fig. 38.3). Excellent reviews of the geology and stratigraphy of the Ecca are given by Haughton (1963, 1969) and Truswell (1970).

The Ecca sequence of the type Karroo Basin is traditionally subdivided vertically into lower, middle and upper units but actually the situation is much more complex due to vertical and horizontal dovetailing of sediments from various environments (see Ryan in Truswell, 1970, fig. 56; Ryan, 1968).

The correlation of stratigraphic units in the Ecca sequence in various basins throughout southern Africa is based primarily on lithology but it is well known that this method is unreliable. In places even over short distances the lenticularity of the beds may prevent correlation except in broad terms. Lithological units, especially in a large basin, are notoriously diachronous, and Truswell (1970) has pointed out that this fact has been ignored in analyses of the Karroo
Fig. 38.3. Ecca Basins. The tectonic situation of the Ecca Basins is dominated by major upwarps in the extreme south, in the Transvaal highland, the Choma-Kaloma ridge and eastern Madagascar. A flysch environment developed in the southern marine embayment but elsewhere coal swamps were widespread. The Gabon Basin (G) is unique among the Ecca basins in that it contains an evaporite sequence.

Basin. To illustrate such a possible discrepancy: McLachlan and Anderson (1973) consider the upper Dwyka shale and the middle Ecca coal beds to be isochronous; this conflicts drastically with traditional correlations. The fact of the matter is, however, that until synchronous planes can be recognised in the Ecca and other Karroo sequences the regional correlations accepted at present (based on general stratigraphic position, lithological character, relationship to unconformities, broad fossil evidence, etc.) can hardly be improved upon.

Ryan (1968) recognises three major facies in the Ecca of the type Karroo Basin and each of these facies can be subdivided vertically into three broadly similar lithologies: shale in the lower and upper units and sandstone in the middle unit. It seems likely that this regional correspondence relates to a tectonic overprint on the sedimentary pattern. Although some regard the lower, middle and upper units as approximately isochronous no reliable time-planes have been demonstrated so far.

The southern Ecca facies is a flysch-type
sequence which accumulated in the deep Karroo trough (Truswell and Ryan, 1969; Theron, 1967). The interbedded sandstones pass laterally into shale to the north and primary structures indicate sediment transport from the south northwards then turning eastward. The fossil remnants are mainly fragmentary glossopterid plant debris, obviously drifted in, as well as a limited trace fossil assemblage. Most of the sediments seem to have been deposited by turbidity currents in fairly deep water but towards late Ecca time shallow water conditions prevailed; just how shallow is indicated by the presence of terrestrial vertebrate fossils in these beds (Barry, 1970).

Originally the geosynclinal trough was probably partly invaded by an ocean which may have been situated to the east. The majority of the sediments were derived from a geanticlinal ridge along the southern side of the trough. The petrology of the sediments indicates acid plutonic rocks in the provenance area, as well as sedimentary rocks, low-grade metamorphic rocks and possibly basic igneous rocks. Unconfirmed reports of volcanoclastic debris in the Ecca (M. R. Johnson, pers. comm.) suggest that active volcanism may have operated during the elevation of the sediment source area.

The western Ecca facies is notably arenaceous and contains the conspicuous Tanqua sandstone, which makes a pronounced salient into the Ecca trough (Winter and Venter, 1970; see also their fig. 10, a facies map of the Ecca). The western Ecca was clearly derived from a granitic source area to the west and southwest of the major trough and, as elsewhere, the sedimentary environment changed from deep water at the beginning to shallow water at the end. The structural environment suggests a somewhat unstable shelf marginal to a miogeosynclinal trough.

The northern Ecca facies is distinguished by an abundant development of coal and sandstone in the middle unit. Both coal and sandstone pass laterally into shale to the south and there are numerous lines of evidence indicating a sediment source situated in the central Transvaal. Numerous disconformities indicate small-scale tectonic oscillations. Deep water conditions never developed and the environment was mainly lacustrine, swampy and deltaic. The only major tectonic downwarp developed in Natal but both in size, degree of downwarp and type of sedimentation it was quite unlike the southern geosynclinal trough.

The shape of the middle Ecca coalfield is like a horseshoe, one tip lying near Ladysmith in Natal, curving northwards through Witbank at the apex, and swinging back through Vereeniging to reach the other tip of the horseshoe near the Orange Free State goldfields. The isopach map of the Ecca indicates that the location of the Lesotho rise was a controlling feature in determining the gross shape of the original coal swamp area.

Up to twelve coal seams are developed in Natal (Geological Survey, 1959), indicating appreciable variation in the conditions of sedimentation, but only about four seams towards the middle of the succession are regionally of economic importance. In places these more persistent seams may be up to 2.5 m thick, but values of a metre or less are much more common.

In the Witbank field five important seams occur (Smith, 1970). Apparently they were deposited in protected swamps in hilly country. Glaucerate horizons suggest some marine flooding at a late stage.

The coalfield near Vereeniging and Clydesdale is interesting because it overlies Precambrian Malmani dolomite which produced a rugged pre-Karroo karst topography. In some of the sinkholes up to 40 m of coal accumulated by a process of progressive slumping. Elsewhere in this area the coal seams are split by a conglomerate horizon which represents reworked glacial debris washed in from nearby hills.

More coal occurs in the small basins at Springbok Flats and Waterberg. The Waterberg 'basin' is actually an embayment of the much larger Botswana Basin. The middle Ecca was deposited under shallow-water fluviatile conditions with an average sediment transport direction westwards (Ryan in Haughton, 1969). The coal seams accordingly diminish in quality and quantity towards the deeper parts of the Botswana Basin. 'Wash-outs' in the coal seams suggest that some differential tectonic movements
affected the Botswana Basin during coal formation.

The sedimentology and tectonism of the Botswana Basin and its extension into South West Africa are imperfectly known, owing to inadequate exposures and numerous uncertainties in stratigraphic correlation (Green, 1969). The Ecca sediments are mainly fine-grained elastics and the major sediment sources were probably to the east, near the Waterberg embayment, and to the north. Conditions favourable for the deposition of coal existed only in the eastern portion of the Botswana Basin but somewhat carbonaceous sediments do occur to the west.

There seems to be no record in the Ecca rocks of a marine incursion in the west, similar to that during Dwyka time (Martin and Wilczewski, 1970; McLachlan and Anderson, 1973). Later history suggests that already in Ecca time the western sea retreated, probably because of mild upward epeirogenic movement in the area west of the Botswana Basin. Hart (1964) and Rillett (1965) suggest that one or more marine incursions penetrated into the main Ecca Basin as far northward as the Free State, Lesotho and Natal, but this is based on insufficient information.

Northwards into the rest of the Karroo tectono-sedimentary terrain only the Congo Basin is known to contain Early Permian Ecca-type sediments. After the Dwyka glaciation numerous enclosed basins or marshes developed in the Congo downwarp, and in a manner probably similar to the Transvaal area, carbonaceous shale, some sandstone and coal were deposited (Furon, 1963). Veatch (1935) has suggested that the land was higher in the west and northwest.

The tectonic framework of the Karroo tectono-sedimentary terrain during Ecca time was essentially uncomplicated and closely similar to the situation which developed during Dwyka time. Over most of the area mild tectonic flexuring took place, the only marked tectonic movement being in the southern geosynclinal trough and to a much lesser extent the Natal downwarp. The sedimentary environment varied appreciably, ranging from a typical turbidite environment to a coal swamp deltaic complex, but most of the Ecca sediments were probably deposited in a lake environment.

Ecca sedimentation in the Zambezian tectono-sedimentary terrain took place in several downwarp basins which were eventually partly bounded by fault scarps. During Ecca times the sedimentary pattern was mainly of a delta-floodplain-lake type with important coal formation in the Wankie-Gwembe area. The best known basin, which also serves as a model for all the other basins in the Zambezian terrain, is the Mid-Zambezian coal basin. The Karroo Basin in the lower Zambezi Valley probably originally linked up with the main Wankie-Luangwa depository and it is possible that the lower Zambezi Basin split northwards along the present Lake Malawi (Nyasa) to connect with the Malawi-Tanzania Basins. Northeast of the Luangwa Basin more Ecca-age deposition took place in a vaguely L-shaped series of basins, of which the Rukwe, Ruhruru, Ruaha and Tanga Basins are probably best documented.

The discussion which follows is based mainly on work by Bond (1956, 1967), Tavener-Smith (1958, 1962), Drysdall and Weller (1966), and Watson (1958). There was during Ecca times a certain continued, albeit restricted, change in the configuration of the basins and source areas in the Zambezian terrain, the trend being towards an increase in the size of the Ecca lakes with time. These lakes served as depositories for the Wankie and Gwembe Formations. The major provenance areas were the extensive Bulawayo-Salisbury granite block (also known as the Lumagundi highlands) to the southeast of the Wankie Basin, and the Kalomo-Choma ridge along the northwestern boundary of the Wankie-Luangwa Basin complex. It appears highly probable that the triangular basement block east of the Luangwa Basin was also an important supplier of debris.

The topographic relief in these hilly tracts was low, 150 m being the average of the quoted figures, but the Mbanga hill, about 45 km east of Wankie, may have had a relief of some 600 m. To the northwest of Wankie high points occurred in the Kalomo-Choma ridge, some of the hills protruding even
through the late Karroo Batoka basalts, but towards the middle of the ridge, near Tara, the topography was probably much more subdued because it seems as if the ridge was partly flooded at times. Within the central part of the Mid-Zambezi Basin the topography was reduced to very low relief, partly because of the denudational effect of the Dwyka ice sheets and partly because of the long-continued differential negative movement in that area.

The shoreline during deposition of the main coal seam was not far northwest of the Wankie-Siankondobo-Nkandabwe line, swinging northwards towards Choma, and possibly northeast from there. The nature of the opposite southeastern shoreline of the Wankie-Gwembe basin has not been documented well but the topography on the Bulawayo-Salis­bury block must have been quite low and the actual shoreline could have been quite irregular.

Northeast of the Mid-Zambezi Basin lies the linear Luano-Lukusashi-Luangwa Basin. In its extreme southern and northern areas it contains poor quality coal seams with much interbedded coarse grit derived from nearby sources. These sequences indicate much more tectonic unrest in the Luano-Luangwa complex than in the Mid-Zambezi Basin.

The Luangwa Basin abuts the North Nyasa ridge at its northern extremity (Haughton, 1963). This vaguely defined structure separates the Luangwa Basin complex from several small down-faulted fragments of coal-bearing Karroo rocks towards Tanzania. These scattered outcrops in all probability belong to one previously connected basin, part of which may have stretched along the east of the North Nyasa ridge all the way from Lake Rukwa past Zomba to the present confluence of the Shire and Zambezi rivers, and another part opened to northeast, either as a single basin or a series of smaller basins from Lake Malawi to beyond Dar Es Salaam. Unfortunately very little is known of the facies variations and sediment dispersal of the Ecca-age sedimentary rocks in eastern Africa. However, it is clear from a study of the known geology (see, for instance, an excellent resumé by Haughton (1963) of the geology of all the areas under discussion) that the basic sedimentary pattern which operated in the mid-Zambezi Basin also controlled sedimentation in the northeastern areas.

The blanket of Ecca-age sediments is nowhere very thick in the Zambebian tectono-sedimentary terrain. In the type Mid-Zambebian Basin, values of less than 100 m are common; the thickest sequence of some 245 m occurs at Siankondobo. In the Luano-Luangwa Basin complex similar thicknesses occur but in the Ruhuhu Basin the thickness is about 400 m.

In the lower Zambezi Basin, between Chicoa and Port Herald, some 600 m to 2400 m of Ecca-age sediments occur in a basin with a highly irregular floor (Haughton, 1963). The sandstone-shale-coal sequence seems to have been deposited as a deltaic complex. The coal contains the same floral assemblage as at Wankie.

The distribution of the Ecca basins in the Zambebian tectono-sedimentary terrain suggests a network of weak crustal zones which caused localised sagging under tensional stress. Some of these basins seem to be confined to the periphery of ancient granitic cratons, notably the Bulawayo-Salisbury block, but the structural control on other basins is not clear.

The sedimentary environment everywhere was deltaic-lacustrine, and the prevalence of coarse clastics, and abundant plants at certain levels, indicate a mild rainy climate.

The Beaufort Tectono-sedimentary Regime

Almost throughout southern Africa Beaufort sedimentation was marked by low energy conditions; this is mostly interpreted in terms of a low flux of tectonic energy (Fig. 38.4). Broadly speaking, there seems to have been no significant variation from place to place in the sort of sediment being deposited during Beaufort time, but the western half of southern Africa has no Beaufort cover. It seems reasonable to suppose that this vast tract was at the time just sufficiently elevated to prevent both significant sediment accumulation and vigorous denudation. The Beaufort fossil flora and fauna are also uniform over the entire area; what zoning there is (Kitching, 1970), is relatively unimportant regionally.
Fig. 38.4. Beaufort Basins. The tectono-sedimentary situation during Beaufort time indicates overall low tectonic energy flux. However, along the southern margin of the type Karroo Basin a major upwarp shed coarse debris into it. The eastern margin of the Zambezian terrain may have merged into a marine environment. Madagascar was in a true position southwest of the one shown here, and moving actively northeast, producing a horst and graben basin structure. Marine encroachment took place from both the northern and southern ends of Madagascar.

Obviously, during Beaufort time no sharp changes in climate and tectonism affected southern Africa.

In the southern part of the type Karroo Basin, Beaufort sedimentation proceeded without a break from Ecca time and in fact the location of the Ecca-Beaufort boundary by means of lithological markers is well-nigh impossible. This problem accounts for the appreciable difference in the maximum thickness allocated to the Ecca by Ryan (1968)—about 3600 m—and by Winter and Venter (1970)—about 2000 m. Fossils are too scarce to be of much use in fixing this boundary.

However, the upper boundary of the Beaufort is well defined. Over vast areas, especially in the Zambezian tectono-sedimentary terrain, an appreciable hiatus marks the contact between the lower Beaufort correlates and Stormberg rocks, and even though the sequence is conformable the change in lithology is distinctive.

In the type Karroo Basin the Beaufort sequence is subdivided lithologically into three (Haughton, 1969) and palaeontologically...
into five units (Kitching, 1970). The lower lithological unit accumulated on vast mud- and sandflats crossed by meandering river channels (Theron, 1967). Near the centre of the Karroo Basin, a feature which was still active during Beaufort times, about 4000 m of lower Beaufort sediment accumulated; northwards this unit thins rapidly. The fossil content of the lower Beaufort (Tapinocephalus, Cistecephalus and Daptocephalus reptiles, freshwater molluscs, some fish and relatively little Glossopterus and Dadoxylon) indicates that the environment was mainly lacustrine marshy and fluviatile. The climate was evidently not favourable for luxuriant plant growth because only some insignificant coal seams developed towards the extreme northeast in the Beaufort Basin.

The middle Beaufort sediments are about 1000 m thick, thinning northward, and are distinguished by a prominent sandstone sequence, the Katberg Sandstone, at the top of the middle Beaufort. The Katberg Sandstone, which also occurs as downfaulted outliers near East London (M. R. Johnson, pers. comm.), contains pebbles of various igneous and metamorphic rocks, derived apparently from a source towards the southeast (Mountain, 1946). This unusual influx of coarse clasts into the basin is certainly connected with a tectonic pulse in the southern geanticlinal area. Possibly this tectonism related to, or was responsible for, the closure of the seaway which in Ecca times connected the Karroo Basin with open sea to the southeast. Radiometric dating of some of the granitic pebbles, to detect a possible syntectonic granite complex in the geanticlinal ridge, yielded ambiguous results (D. H. Elliot, pers. comm.).

Towards the northern half of the Beaufort Basin Theron (1970) has indicated at least two sediment sources, one apparently far to the southeast and another a nearby granitic source. This nearby source, which was responsible for the coarse-grained arkoses in the middle Beaufort, may have been a locally uplifted part of the Lesotho rise. According to latest estimates the age of the middle Beaufort, which corresponds with the Lystrosaurus Zone (Kitching, 1970), is Early Triassic, but no finer time divisions have yet been made.

The upper Beaufort is a reddish, varicoloured mudstone-sandstone sequence, thinning northwards from a maximum thickness of some 700 m near Queenstown. The distinctive reptiles define the Cynognathus Zone. The complete faunal and floral assemblage suggests that during this last stage of the Beaufort Basin the environment remained swampy with much deposition from rivers. The palaeocurrent patterns are commonly highly dispersed because of meandering river channels and also because of peripheral drainage in the Free State and Transvaal areas. It is suggested that the upper Beaufort may represent sediment stripped from lower and middle Beaufort outcrops in the steadily rising southern source area.

No Beaufort sediments have been recognised in the extreme western part of the Botswana Basin. This means that the western part of the basin was possibly uplifted slightly.

In northern Angola the fish beds in the Cassanje Series correspond in age to lower or middle Beaufort (Haughton, 1969). But these beds are only about 10 m thick (Furon, 1963) and indicate very slight downwarping in that area. The situation in the rest of the Congo Basin during Beaufort time is not clear but it seems as if there was a long period of post-Permian erosion in that area. Sedimentation resumed in the Jurassic (Furon, 1963).

In general the western and northern part of the Karroo tectono-sedimentary terrain was an area of non-deposition and some erosion during Beaufort time. In the main Karroo Basin appreciable downwarp of the trough zone was accompanied by some tectonic action in the southern source area, but sedimentation took place throughout under terrestrial and swamp conditions.

In the Zambezian tectono-sedimentary terrain Beaufort sedimentation is represented mainly by the Madumabisa Shale in Rhodesia and Zambia and its correlates elsewhere: the Mwanza sequence in the Shire area (lower Zambezi Basin), the Chiweta Beds in Malawi and the Ruhuhu Beds in Tanzania (Drysdall and Weller, 1966).

The type Madumabisa Basin is once again
the Wankie-Gwembe area. In the Gwembe district, Zambia, Tavener-Smith (1962) has demonstrated that the lower Madumabisa Shale is in fact isochronous with much of the upper Gwembe Coal Measures so that the lithological contact between the two formations represents a distinctly diachronous plane; this relationship is probably typical of much of the Ecca-Beaufort transition elsewhere.

The Madumabisa Mudstone is about 500-700 m thick. The lower unit, a grey carbonaceous mudstone, increases in thickness away from the shoreline. The fine-grained sediments and carbonates indicate general low topographic relief in the source areas. The poor bedding implies that the sedimentation was slow and continuous while downwarp proceeded gently. The general homogeneity of the lower Madumabisa sediments suggests gentle uplift in the source areas. The main source area was the Choma-Kaloma ridge.

The middle Madumabisa is noticeably better bedded and has a more varied lithology. Evidently intermittent small-scale earth movements were active during this time but the limey mudstone implies low topographic relief in the provenance area. On the other hand the development of some grits and feldspar-rich sandstone in the sequence may be interpreted as indicating some local tectonic elevation of a granitic source area. The situation is suggestive of block faulting and this activity may have been a forerunner of the later extensive graben faulting. The non-marine fossils of the middle Madumabisa—freshwater lamellibranchs, ostracods, fish scales and *Glossopteris*—signify a lacustrine assemblage. The reptiles are comparable with those of the *Tapinocephalus* and *Endothiodon* (now *Cistecephalus*) Zones, and the middle Madumabisa therefore correlates with the lower Beaufort.

The uppermost Madumabisa of the Gwembe district consists almost entirely of a featureless calcareous mudstone representing extremely low energy sedimentation. The vertical progression in the Madumabisa sequence from non-calcareous mudstone plus siderite near the base to calcareous mudstone without siderite near the top, suggests a concomitant change in water chemistry from mildly acid and reducing at the beginning to oxygenated and mildly alkaline conditions towards the end.

Bond (1956), comparing two localities about 130 km apart in the Wankie basin, indicated a removal by erosion of some 650 m of Madumabisa Mudstone towards the edge of the basin, while at the same time sedimentation proceeded without disruption in the basin centre. This erosional hiatus implies uplift of the southeastern shore of the Madumabisa lake towards the end of its existence. No such tectonic event has been recognised in the Gwembe area which was probably much nearer the centre of the Madumabisa Basin. Tavener-Smith (1962) suggested that the Madumabisa lake eventually dried up because of increased evaporation in a dryer and warmer climate. Throughout its history, this lake underwent mild epeiric warping with minor fault tectonics along its periphery.

In the upper Luangwa Basin the Madumabisa history is much as recounted above except that initially much more vigorous conditions of sedimentation prevailed, resulting in rapid accumulation of arkoses, alternating with mudstones (Drysdall and Kitching, 1962). This situation may have been due to a more rugged local topography in the source area, but towards middle Madumabisa time shallow water lacustrine conditions were established in the basin. A swampy mudflat environment with possibly some red dust dunes (see also Bond, 1956) characterised the northern part of the Luangwa Basin towards late Madumabisa time. The reptiles of the Madumabisa Mudstone of the Luangwa Basin occur mainly towards the top of the approximately 700 m sequence and correlate with the *Endothiodon* and *Cistecephalus* Zones of the type Beaufort Basin.

The lower Zambezi Basin in Mocambique contains about 1000 m of lower Beaufort sediments, mainly sandstone, grit, arkosic grit and calcareous sandstone in the lower half of the succession, and reddish mudstone, marl and siltstone with some dolo-limestone in the upper half. The sequence suggests terrestrial alluvial plain sedimentation so typical of Beaufort time and indicates energetic conditions at the start of sedimentation, tapering
off to low energy, very shallow water conditions towards the end. Here, too, the entire middle and upper Beaufort sequence is missing below the succeeding Stormberg sediments. The regional character of this hiatus proves that non-deposition operated over the entire Zambezian tectono-sedimentary terrain during Late Permian to Middle Triassic or possibly even later.

The lower Beaufort sediments in northern Malawi also display this fining-upwards feature. The lower 250 m consist of sandstone and grit, and are overlain by the Mount Waller Mudstone (200 m) and the Chiweta Mudstone (280 m) (Haughton, 1963). The relationship between the various scattered Karroo outcrops in northern Malawi and the Luangwa Basin is unknown.

Towards Tanzania and Kenya it becomes increasingly difficult to correlate events with known history in the rest of the Zambezian terrain. But it seems clear that the sedimentary and tectonic character became more and more atypical towards the northeast. Tectonic activity was more irregular, and one of the resultant unconformities (near Nyakatitu in Tanzania) is in fact associated with a marine incursion (Spence, 1957). This is the earliest known marine incursion associated with the Karroo Basins on the African mainland and suggests that fracturing of Gondwanaland facilitated marine flooding of parts of the continent. Such tectonic activity, as well as the resultant effect on the sedimentary environment, explains the difficulties experienced at present in correlating formations and events. On the other hand, this is obviously a critical area that needs to be studied in greater detail.

**The Stormberg Tectono-sedimentary Regime**

Traditionally the Stormberg sequence includes the basalts and other igneous rocks; in this section the sediments will be discussed, and in the next section aspects related to the lavas.

The tectonic activity of Stormberg time was distinctly different in the Karroo and Zambezian tectono-sedimentary terrains (Fig. 38.5). In the Karroo terrain very strong pulsatory uplift developed along the southern margin of the main Karroo Basin. At the same time in most of the western and northern part of the Karroo terrain there was little or no deposition. In the Zambezian terrain Stormberg tectonism produced extensive faulting, which tended to follow pre-existing downwarps; in this manner the existing basins changed into graben basins which were yoked to the adjacent rising blocks. The major monoclinal downwarp structure of the Lebombo Mountains is probably a drape-over structure located on a deep-seated fault and seen in this light it may be the only really well-defined junction between the Karroo and Zambezian terrains.

In the Karroo tectono-sedimentary terrain Stormberg sedimentation was initially restricted to the rather small Molteno Basin. This basin covered no more than about 200,000 km² (compared with about five times that area for the main Karroo Basin) and a maximum thickness of some 700 m accumulated towards the south, wedging to zero northwards.

A very much smaller, probably intermontane, valley in the Waterberg area accumulated some Molteno-type sediments, but their sedimentology is largely unknown. No comparable sediments have been reported from the rest of the Botswana Basin, nor are any known in Angola and the Congo Basin (see comment by Furon, 1963, on Congo Basin correlations by Veatch, 1935).

The Molteno molasse-type clastic wedge represents the syntectonic sedimentary response of the renewed tectonic movement along the southern boundary of the Karroo tectono-sedimentary terrain. The Molteno sequence consists of a cyclic alternation of four basic facies types; each cycle contains a basal conglomerate facies, followed by a sandstone, siltstone-shale, and shale-coal facies (Rust, 1959; Turner, 1970). The conglomerate facies mantles a continuous erosion surface with low relief, and it merges upwards into the sandstone facies. The clasts are mainly extra-basinal types (lower Palaeozoic quartzites) but intrabasinal shale pebbles also occur. Most of the quartzite pebbles (one quartzite boulder 75 cm in diameter lies near the town of Molteno) are derived from the Devonian Witteberg Group; it is significant to note that...
Fig. 38.5. Stormberg Basins. The Stormberg Basins as shown here actually incorporate several distinct stages. The main tectonic activity was severe compressive folding in the south, involving Karroo sediments up to lower Beaufort, and semigraben faulting in the Zambezian terrain. The Madagascan block was drifting actively northeastward and marine encroachment of the western side of Madagascar was almost complete. The overall climate was dry and the palaeowind direction indicates westerly to southwesterly winds.

The sandstone facies consists of zones up to 50 m thick of relatively coarse-grained sandstone fining upwards. Abundant trough and planar cross-bedding and other vectorial primary structures indicate sediment transport to the north and northeast (Smits, 1966). The Indwe Sandstone, a prominent unit in the Molteno sequence, contains some very large clastic feldspar grains, and this may be indicative of a nearby granitic source area. The nearest known granite area lies towards the east. This suggests renewed elevation of the eastern highland known to have existed during Dwyka and Ecca time.
Turner (1970) interprets the environment of the Molteno sandstone facies as a typical fluviatile pointbar complex in a braided river system with low sinuosity.

The siltstone-shale facies passes gradually upward from the sandstone, and reaches thickness values of 5-7 m. This material was deposited almost exclusively from suspension in permanently inundated lakes, ponds and abandoned channels. The bulk of the floodplain deposits is probably in-channel and not overbank sediment.

The uppermost facies of each cyclothem consists of carbonaceous shale and/or coal. The coal was probably derived mainly by in situ growth in back-swamps with some shale deposition taking place in overbank areas. Tectonic activity during the deposition of this and the previous facies must have been quite slight.

Molteno sedimentation was clearly governed to a large degree by tectonic pulses.

In the Karroo Basin the energetic fluviatile Molteno conditions graded into the more subdued situation which applied during the deposition of the Red Beds, a sequence of reddish mottled mudstone and fine-grained sandstone (Botha, 1967, 1968). Like the Molteno the Red Beds in the type area accumulated in a relatively small elongate basin with the deepest downwarp, about 500 m, in the south but wedging rapidly northwards. In the extreme north the Red Beds in places overlap the Molteno sandstone to rest with no apparent structural break directly on upper Beaufort sediments. The sediment dispersal pattern indicates inflow from the south and southeast. Clearly, the palaeogeographic situation was not changed from Molteno time but obviously an appreciable climatic change had taken place. Calcareous concretions, limey zones, clay-pellet conglomerates, mudcracks and the scarcity of plant remains all indicate terrestrial deposition in an arid climate. Even the dinosaur population was affected by this adverse climatic development because smaller, presumably more mobile forms dominate the vertebrate record in late Red Beds time (Haughton, 1924).

Botha (1967) suggests that the trend of the downwarp axis of the basin at this time was towards the northeast; this may reflect the effect of crustal flexuring prior to the extrusion of the late Karroo basalts along the eastern margin of the Karroo tectono-sedimentary terrain.

In the Karroo terrain the Red Beds have a wider distribution outside the main basin than the Molteno beds. Correlates of the Red Beds are known in the central Transvaal (the so-called Bushveld Mudstones), as well as in the Botswana Basin and along the Lebombo escarpment, extending northwards into the Soutpansberg-Tuli area. It seems that a veneer of Red Beds (up to 115 m thick in the Waterberg area) was deposited over most of the Transvaal block. There must have been tectonic activity but the main imprint on the sediments seems to have been due to the influence of an arid climate.

The Cave Sandstone is the top unit of the Stormberg. In the type area around Lesotho three facies can be recognised (Beukes, 1970). The Cave Sandstone consists mainly of very fine-grained sandstone. Only the middle member displays undoubted dune structures; in the lower and upper members the massive bedding is ascribed to the effects of percolating rainwater which destroyed primary aeolian structures in the unconsolidated dune sand. The thickness variations of the Cave Sandstone are fairly irregular (100 m or less, to about 240 m) and in part this may be because in places the contact between the sandstone and the basalt follows a dunefield topography, whereas in other places there is evidence of river channel erosion prior to the extrusion of the lavas.

The primary control on the deposition of the Cave Sandstone was undoubtedly climatic. The palaeowind pattern shows wind consistently blowing from the west. The provenance material was probably mostly Beaufort sediments, but heavy minerals prove that the granite-gneiss basement complex as well as Transvaal metamorphics and the Ventersdorp lavas also supplied material.

The Cave Sandstone was deposited mainly as dunes, playa lake deposits and some river channel deposits in a generally arid climate with seasonal rainstorms. There was an almost unnoticeable gradation from the Red Beds to Cave Sandstone deposition due to the
change in climate. The cessation of sedimentation was brought about by the steadily increasing lava cover which eventually covered the source areas. In some places early lava flows followed the depressions in between dunes, later to be covered by sand, and yet more lava. Kingsley (1964) found fossil dinosaur tracks on sandstone interbedded in this way between lava flows.

There is little information on the tectonic environment to be obtained from the Cave Sandstone. However, the regional trend of the type Cave Sandstone basin (Beukes, 1970) is almost identical to that of the preceding Red Beds, Molteno and upper Beaufort Basins. This stresses the regularity of the Karroo tectonics in this area throughout the Triassic. In a relatively small area the cumulative downwarp during the Triassic was about 2000 m; this indicates a special localisation of negative movement which is all the more remarkable when it is recalled that practically the same area was the site of the Lesotho rise during early Karroo time.

In the rest of the Karroo tectono-sedimentary terrain the Cave Sandstone is usually absent. Some occurs in the eastern and central parts of the Botswana Basin (Green, 1969), and the small outlier known as the Etjo Sandstone (Hodgson, 1970) in northern South West Africa is commonly correlated with the Cave Sandstone.

In the Zambezian tectono-sedimentary terrain a completely different tectonic pattern which operated during the Triassic produced a sequence of sediments quite unlike the type Stormberg succession in the Karroo Basin. And yet the climatic overprint is so strong that certain parts of the sequences in the Karroo and Zambezian terrains are almost identical.

The tectonic activity in the Zambezian terrain took the form of extensive faulting with the formation of elongated yoked basins. These grabens were confined to pre-existing basins. Initially their size was reduced drastically, but eventually the spreading sedimentation overstepped the lower Karroo rocks almost everywhere.

The Escarpment Grit and its correlates represent the initial syntectonic sedimentation. Drysdall and Kitching (1962) suggest that the Escarpment Grit in the Luangwa Basin is of early Stormberg age; this dating is generally accepted (Bond, 1967; Tavener-Smith, 1962) although Anderson and Anderson (1970) suggest that the Escarpment Grit is of pre-Molteno age.

Bond (1967) recognised two tectono-sedimentary cycles in the upper Karroo in the Zambezian terrain. A typical succession in the Mid-Zambezi Basin is:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nyamandluvo Sandstone</td>
<td>c. 20 m</td>
</tr>
<tr>
<td>Forest Sandstone</td>
<td>c. 65 m</td>
</tr>
<tr>
<td>Pebby arkose</td>
<td>c. 85 m</td>
</tr>
<tr>
<td>Tectonism 2</td>
<td></td>
</tr>
<tr>
<td>Fine red marly sandstone</td>
<td>c. 72 m</td>
</tr>
<tr>
<td>Escarpment Grit</td>
<td>c. 20 m</td>
</tr>
<tr>
<td>Tectonism 1</td>
<td></td>
</tr>
</tbody>
</table>

Towards the northwest, in the Gwembe district, Tavener-Smith (1962) found basically the same sequence, but much thicker:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Millet seed sandstone</td>
<td>200 m+</td>
</tr>
<tr>
<td>Pebby arkose with basal conglomerate</td>
<td>c. 450 m</td>
</tr>
<tr>
<td>Tectonism 2</td>
<td></td>
</tr>
<tr>
<td>(No disconformity observed)</td>
<td></td>
</tr>
<tr>
<td>Sandstone and mudstone</td>
<td>c. 1430 m</td>
</tr>
<tr>
<td>Escarpment Grit</td>
<td>c. 400 m</td>
</tr>
<tr>
<td>Tectonism 1</td>
<td></td>
</tr>
</tbody>
</table>

The above relationship is typical of the Mid-Zambezi Basin: the upper Karroo sequence is extremely thick towards the northwest and relatively thin towards the southeast (Gair, 1959; Dixey, 1935; Molyneux, 1903). Primary structures and petrology conclusively prove that the provenance of the upper Karroo sequence was the Choma-Kalomo granitic ridge (Tavener-Smith, 1962). Textural data indicate that the grit and sandstone were rapidly deposited not far from the provenance areas. On the other hand, the millet seed sandstone is a well-sorted mature deposit and was probably accumulated by wind action in the same way as the Cave Sandstone.

The general aspect of the upper Karroo sequence in the Zambezian terrain is that of a terrestrial red-bed facies (Tavener-Smith, 1962). The climate was semi-arid. A high rate of erosion and a high relief in the source
areas as well as appreciable subsidence in the basin areas are suggested by the great volume of the upper Karroo sediments, their coarse grain and textural immaturity.

Two tectonic events are recorded in the upper Karroo sediments of the Zambezian terrain. Each event was preserved as a cycle consisting of coarse-grained feldspathic clastics which fined upwards as the effects of the original tectonic pulse waned. The first cycle in the upper Karroo sequence was interrupted at a fairly advanced stage, but the second cycle proceeded much further towards stable low energy conditions. The second cycle was in fact interrupted by the tectonic pulse which heralded the Karroo volcanism.

The original Karroo faulting has been exploited and obscured by later movements and detailed descriptions of the original Karroo fault systems do not exist. Tavener-Smith (1962) states:

At an early stage in the pre-Stormberg movements overstrained warping gave way to fractures of considerable magnitude: the present fault system on the northwest side of the [mid Zambezi] Valley was initiated at this time . . . At the close of Karroo times further warping and faulting resulted in another phase of uplift northwest of the basin . . . Renewed movement on the fault planes during Pleistocene-Recent times is indicated by coarse, arkosic sands . . . .

The geological history of the Luangwa Basin during Stormberg time (Drysdall and Kitching, 1962) is essentially as for the Mid-Zambezi Basin. The Escarpment Grit lies with an abrupt, slightly diachronous contact on the Madumabisa Mudstone. The grit is about 200 m thick and is overlain by 300 m of the Ntawere Sandstone and Mudstone; fossil evidence suggests a Molteno age for the Ntawere Sandstone—this would possibly make the Escarpment Grit in the Luangwa Basin older than elsewhere. The rest of the upper Karroo sequence in the Luangwa Basin contains a varied succession of marl, sandstone and grit—with a possible aeolian unit near the top of the 1700 m sequence.

The original structure of the Luangwa Basin may have been an asymmetrical basin similar to the Mid-Zambezi Basin, but it eventually developed into a graben structure. The prevalence of grit in the upper Karroo succession may point towards an increased degree of tectonic activity in the Luangwa Basin.

Notwithstanding Drysdall and Kitching’s contrary views (1962: Table I) it is almost certain that Bond’s (1967) sedimentary cycles can be recognised in the rocks of the Luangwa Basin:

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Age</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aeolian sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper grits</td>
<td>c. 1000 m</td>
<td></td>
</tr>
<tr>
<td>Tectonism 2</td>
<td>?</td>
<td>c. 170 m</td>
</tr>
<tr>
<td>Red Marl</td>
<td>?</td>
<td></td>
</tr>
<tr>
<td>Ntawere Formation</td>
<td>c. 300 m</td>
<td></td>
</tr>
<tr>
<td>Escarpment Grit</td>
<td>c. 200 m</td>
<td></td>
</tr>
<tr>
<td>Tectonism 1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Madumabisa Shale</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The strong tectonic overprint on the sediment of the upper Karroo in the Zambezian tectono-sedimentary terrain simplifies the correlation of the upper Karroo sequence towards the northeast. The correspondence of the Escarpment Grit with the Chiweta Grits (Dixey, 1927) and the Kingori Sandstone (Drysdall and Kitching, 1962) seems straightforward. There is a close lithological and faunal similarity between the Manda Beds and the Ntawere sediments, and the Mazeras sequence in Kenya is almost identical with upper Karroo sequences to the south (Haughton, 1963).

The Stormberg Volcanism and Tectonics

The Karroo sedimentation, which operated over a period of some 100 m.y., was terminated by ‘what must have been one of the most spectacular volcanic episodes the earth has ever seen’ (Cox, 1970). The resultant basaltic province is a classic continental tholeiitic sequence of Jurassic flood-basalts and dolerites, and it constitutes a continental feature of the first magnitude (Rhodes and Krohn, 1972) (Fig. 38.6). The development of this vast igneous province is intimately associated with the disruption of Gondwanaland. Numerous workers have investigated the Stormberg Volcanics but in this review reference will be made mainly to aspects related to the tectonic pattern; the discussion that follows is based for the greater part on the work of Cox (1970), Woolley and Garson (1970), Vail (1970) and Rhodes and Krohn 1972).
Radiometric age determinations indicate that the Karroo tholeiitic episode spans the Early and Middle Jurassic, but cumulating evidence suggests that long-continued magmatism followed on this first phase. Certainly it seems more and more inescapable to regard the Karroo volcanism and the rift tectonics in Africa as essentially continuous. But with reference to the Karroo sedimentary basins it is probably not particularly meaningful to pursue the tectonic and igneous history beyond the Middle Jurassic.

Because of the major structural overprint of the Gondwanaland disruption tectonics on the African continent, certain structural elements gain prominence during the Stormberg volcanism. Regardless of this shift in accent, the basic distinction between the Karroo and Zambezian tectonic terrains remained throughout the Jurassic.

In the Karroo tectonic terrain Jurassic volcanism was confined to the main Karroo Basin and the Botswana Basin. Radiometric data on the South West African and Kao-
koveld igneous suite (Siedner and Miller, 1968) suggest a Cretaceous age for these basalts, granites and alkaline intrusives. Elsewhere in the Karroo terrain, Karroo volcanics and intrusives are absent.

Naturally the present distribution of the Karroo lavas in the Karroo tectonic terrain represents but a fraction of their original distribution but it is significant that the location of the main outcrops in both the Karroo and Botswana Basins corresponds to areas of downwarp during sedimentation. It seems as if crustal downwarp proceeded even during extrusion of the lavas.

The conduits of the Karroo sheet lavas are considered to be the dolerite dykes which abound in the main Karroo Basin, and are particularly common around Lesotho (Vail, 1970). This distribution of tensional structures in the Karroo Basin is a reflection of the stress pattern in the crust at the time of igneous intrusion. The Lesotho area had been a place of localised downwarp since at least late Beaufort time and obviously this was a favourable situation for tensional fractures to develop at a time when the regional tectonic state in the crust promoted a breakthrough to the mantle. Woodward (1966) has pointed out that the compressive stress along the southern periphery of the basin prevented the intrusion of dykes.

Rhodes and Krohn (1972) found a close relationship between the regional geochemical variations of the Karroo basalts and dolerites and the tectonic configuration of the sedimentary basin. They suggested that the observed patterns of geochemical variation seem to be independent of pre-Karroo structures, just as the Karroo sedimentary basin seems to have been mainly insensitive to older structures in the crust. This suggestion is in some conflict with the viewpoint of Saggerson and Logan (1970), who maintained that pre-Karroo structures, particularly the downwarped margins of stable continental cratons, strongly control the location of major Karroo igneous activity. The separation of the African continent into two different tectonic terrains may reconcile these ideas. In the Karroo tectonic terrain older structures seem to have been ignored in the development of the tectonic fabric but in the Zambezean terrain the converse is true in many instances.

The fulcrum in the distribution maps in various chemical elements in the Karroo igneous rocks is always in the vicinity of Lesotho (Rhodes and Krohn, 1972). The chemistry of this zone (lower Ti, Fe, Na, K and P, and higher Si, Al and Mg) indicates low-pressure, shallow depth fractionation of the parent magma.

Feeder dykes (Vail, 1970) are relatively scarce in the Zambezean tectonic terrain, being confined mainly to the southern and eastern boundaries of the Bulawayo-Salisbury granitic block (also known as the Rhodesian craton or the Lumagundi block). In fact, it becomes somewhat embarrassing to explain the presence of Karroo basalt at Featherstone in the very centre of this area devoid of known feeder dykes. The Batoka basalts near Salisbury may have their feeder dykes located to the west and obscured below the Kalahari sand cover; this area has a basement structural setting similar to that at Nuanetsi, where a major igneous complex developed (Cox et al., 1965).

The geochemical pattern of the Rhodesian basalts is distinctly different from that of the main Karroo Basin (Rhodes and Krohn, 1972) and lends further support to the view that the structural stress environment controls the fractionation chemistry of magmas (Yoder and Tilley, 1962).

By far the most impressive structure of the Karroo volcanism is the cuspate-shaped Lebombo-Nuanetsi-Lupata monoclinal downwarp along which a thickness of several kilometres (estimates vary from 9 km to 13 km) of basaltic and rhyolitic lava accumulated. This lineament is moulded along the eastern boundaries of the ancient Transvaal and Rhodesian cratons and there seems to be little doubt that it represents a deep-seated crustal lesion which developed in response to large-scale tensional stresses generated during the break-up of Gondwanaland, specifically the spreading of the Mocambique channel (Flores, 1970). The geochemistry of the Lebombo basalts is significantly different from the rest of the Karroo lavas (Rhodes and Krohn, 1972). Their chemistry is consistent with high-
pressure fractionation of a parent-magma intersected at great depth by a deep-seated fracture system.

The systematic change in the petrology of the Karroo lavas from south to north is a direct response to the regional change in tectonic environment (Woolley and Garson, 1970). The change from dominantly tholeiitic basalts in the Lesotho area, through the development of rhyolites in the Lebombo monocline, to the alkaline intrusions and effusions in the Lupata area and northwards into Malawi, is matched by a systematic change in tectonic environment from a region of little deformation in the south (mainly tensional fractures), through a zone of monoclinal flexuring, to an area of major block faulting in the north. The southern volcanism was associated with magma generation at a relatively shallow depth, whereas northwards the magmas were produced at progressively deeper levels. However, several have pointed out (Cox et al., 1965; Manton, 1968; Stratten, 1970; Woolley and Garson, 1970) that the volume of rhyolite and other acid as well as alkaline differentiates associated with the Lebombo-Nuanetsi-Lupata line demands some degree of wall-rock assimilation, and that a simple stress-controlled model of magma genesis should not be presumed.

The general tectonic picture which develops from a study of the Karroo volcanism highlights the concentration of tectonic activity along the eastern margin of the African continent. Without doubt this is a reflection of the active disruption zone in Gondwanaland. The lava petrology suggests that shearing proceeded from the north to the south. A concomitant easterly rotation of the Antarctic (?) block may explain why the normal tensional stress was less in the south.

The situation along the western part of the African continent is less easily evaluated. The Cretaceous ages of the Etendeka lavas and other intrusives, not to mention similar ages for parts of the lava sequences at Lupata (Flores, 1964) and southern Lebombo (Stratten, 1970), indicate the need for further study of this aspect of the Mesozoic igneous suite in Africa.

The Tectono-sedimentary Situation in Madagascar

Madagascar has always been, and still is, somewhat of an enigma in terms of its place in Gondwanaland. One of two positions is commonly favoured: either opposite Kenya, or opposite Mocambique (Kenya: du Toit, 1937; King, 1965; Mocambique: Wegener, 1924; Wellington, 1954, 1955). Teichert's (1970) complaint that Madagascar 'is usually pushed around quite a bit on continental drift maps' is more than justified. More work is urgently needed in order to resolve this problem.

The geology and structure of Gondwana formations in Madagascar are discussed by Besairie (1960, 1961), Cliquet (1957), Furon (1963) and Haughton (1963, 1969).

Briefly, the Gondwana succession consists of three sedimentary sequences: the Sakoa at the base, followed successively by the Sakamena and the Isalo. These groups are separated by regional unconformities and each sequence is well dated by marine beds. The Madagascan Gondwana succession does not contain any volcanic rocks.

The Sakoa is correlated with the Dwyka-Ecca sequence, the Sakamena corresponds to the Daptocephalus to Cynognathus Zones (Late Permian to Early Triassic) of the upper Beaufort, and the Isalo sequence correlates with the Stormberg sequence (Late Triassic to Early Jurassic) (Anderson and Anderson, 1970).

The Sakoa sequence lies with strong unconformity in fault-bound basins on the Precambrian basement rocks. These faults predate the Sakamena. The basal Sakoa Formation is a tillite sequence like the Dwyka. These glacigene sediments occur only in the extreme southwest of the Madagascar Basin. Bedrock topography suggests that ice flow may have been towards the northwest (Besairie, 1961), and Frakes and Crowell (1970a) measured vectorial structures produced by water flow which indicate flow towards the northeast. Coal measures follow conformably on the glacigene sequence. In turn the coals are overlain, in places progressively, by a red-bed sequence which becomes conglomeratic towards the top. Overlying the conglomerate horizon is the marine
Vohitolia Limestone, evidence of a first, Early Permian marine incursion in the southernmost part of the Madagascan Basin.

The Sakoa sequence is overlain unconformably and transgressively by the Sakamena sequence. The Sakamena consists mainly of continental sedimentary rocks with a total thickness of some 3500 m. At least three marine horizons occur in the Sakamena. At the base is a complex series of conglomerate wedges which abut a horst-and-graben topography, but towards the west, away from the main sediment source area, shale, mudstone and sandstone predominate. The basal marine limestone has some faunal affinities with the Ruhembe Beds of Tanzania while the plant assemblage higher up in the sequence has a Glossopteris/Dicroidium aspect. Several reptiles, amphibians, ammonites and fish occur in the uppermost Lower Sakamena and Middle Sakamena, including the reptile Tangasaurus, also found in the Tanga Beds in Kenya. The age suggested for this assemblage is Early Triassic (Anderson and Anderson, 1970). These fossil affinities are some of the reasons why Madagascar is frequently shown adjoining Kenya in palaeogeographical reconstructions.

The distribution of the marine horizons in the Sakamena Basin indicates that a major marine incursion occurred in the southwest during the Late Permian. A lesser marine flooding took place somewhat later while late in the history of the Sakamena Basin marine conditions developed in the extreme north of the basin (the Barabanja embayment). These marine transgressions probably represent extremely shallow flooding of an essentially continental environment. What precisely the position was, offshore, towards the rapidly widening crustal rift, is difficult to gauge, but presumably typical ocean floor was being developed in the eo-Mocambique Strait (Flores, 1970).

The deposition of the Isalo was preceded by renewed graben faulting. The sequence can be subdivided into three units the lower of which (Isalo I) is mainly conglomeratic and arenaceous. This is usually correlated with the Molteno/Escarpment Grit sequences of mainland Africa. Isalo II displays two distinctly contrasted facies, being of a marine/lagoonal aspect in the north, and continental in the southwest. The Isalo II marine fauna indicates a Liassic age. Isalo III represents a major marine transgression during the Mid-Jurassic when marls and diverse limestones were deposited along the entire western coastal zone of the Madagascan block. No graben faulting affects the Isalo. This reflects a distinct change in the tectonic environment towards the Mid-Jurassic, namely a reduction in the tensional stress situation.

Recently Flores (1970) listed some strong arguments in favour of the hypothesis that Madagascar originally adjoined Africa opposite Mocambique. His arguments are based on the structural conformity of pre-Karroo lineaments and the comparative late Palaeozoic to early Mesozoic history of both terrains, allowing for the cumulative displacement of Madagascar away from Mocambique.

Rifting may actually have commenced in pre-Karroo time. The major fracture zone was somewhat dog-leg shaped and consisted of two dissimilar parts. A tensional fracture zone developed approximately between Durban and Soutpansberg; the movement in this zone was at right angles to the strike of the lineament. The second part was a wrench fault zone which trended about northeast from the Soutpansberg area; the movement in this zone was a left-lateral wrench type.

During Sakoa time the tensional fracture zone had begun to open, allowing moderate and temporary marine flooding in the south. According to the palaeogeographic maps of Flores (1970), the estimated drift rate in the period from Carboniferous to Early Triassic was roughly 1 mm per year. The increased degree of marine flooding during Sakamena time is a reflection of the steadily widening strait. During early Isalo time continental deposits may have mantled the margins of the rift but gradually a fairly permanent seaway was being established. The onset of large-scale volcanism on the African mainland is represented by two features in Madagascar. Firstly, the Isalo II and especially Isalo III marine beds indicate widespread marine flooding on the margin of the Madagascan block. Secondly, the palaeogeographic reconstructions of Flores (1970) suggest a sharp
increase in the drift rate, namely about 15 mm per year, during the interval from Early Triassic to about Early Cretaceous. Obviously this was a tectonically highly active period in Gondwanaland history. Flores (1970) suggests that the Madagascan-Indian block (which moved as a single unit until about Mid-Cretaceous) was under some tectonic compression at the time of the Karroo floodbasalt eruptions, and that the compressive stress situation changed to tensional only during the Mid-Cretaceous, when India was severed from Madagascar. The estimated drift rate of Madagascar in the interval Early Cretaceous to Early Triassic is about 8 mm per year; thereafter the rate rapidly decreased.

Seen in the above context the Madagascan block becomes tectonically part of the Zambezian terrain. While the major part of the Madagascan sedimentation under review compares closely with the terrestrial Karroo sequence, the significant marine intercalations place Madagascar in a different sedimentological environment.

DISCUSSION

The reasons for the different tectonic expression of the Karroo and Zambezian terrains are not clear. Generally the Gondwana basins in Africa are not, except in rare instances, moulded on pre-Karroo lineaments. Throughout the Karroo tensional stress was dominant except for the compressional welt along the southern boundary of the type Karroo Basin. Cox (1970) pointed out that the tensional stress was directed more or less northwest to southeast in the area referred to in this paper as the Zambezian terrain. Flores (1970), in explaining the development of the Mocambique Strait, suggested a tensional stress field directed roughly east-northeast to west-southwest. Clearly more work needs to be done before a coherent picture of the regional stress pattern through time can be assembled.

The compressional welt along the southern boundary of the type Karroo Basin may have resulted from an abutment collision between part of eastern Gondwanaland and Africa because of a rotation movement developed during the shearing away of the two newly developed plates.

The regional structural controls for the localisation of the Karroo volcanism, especially in the Zambezian terrain, seem to be pre-Karroo lineaments. This is an instance where pre-Karroo structures have played a major role in localising Karroo tectonism. On the other hand, similar structural control seems to be absent for the volcanic activity on the Karroo tectonic terrain; admittedly, the basement is covered over most of this terrain and evaluation of basement structure effects on the magmatic activity is difficult.

The petrography of the Karroo volcanic rocks seems to suggest that the crustal lesions reached progressively deeper in the northern regions of Africa plate; this supports the hypothesis that the shearing of Gondwanaland was initiated in the north.

The apparent absence of magmatism in the southern fold zone is reasonably easily explained in the light of available information. It appears that the compressive phase was unfavourable for the intrusion of basaltic magma. On the other hand, the fold deformation may have been a high-level decollement type unsuited for the generation of granitic magma. However, the igneous activity, if any at all, presumed to be associated with the southern sediment source is still shrouded in some mystery.

For a period of some 150 m.y. Africa was largely an area in which 'closed-system' erosion and sedimentation took place. Seen regionally, very little sediment left the boundaries of the African continent, and relatively little material was introduced from outside it. The only major source of sediment which is now no longer completely visible on the African continent is the source area which curved round the southern margin of the Karroo Basin. During this 150 m.y. period, Africa changed from being a land-locked part of the Gondwanaland supercontinent to being a separate and ocean-encircled continent. During this transition Africa suffered no more than slight regional tilting, minimal folding, some faulting and practically no metamorphism.

Nevertheless several regional breaks in the
sedimentary record attest to tectonic effects which affected the entire part of Africa under discussion. The basal contact of the Ecca sequence is such a disconformity. In the extreme south the sedimentary record is completely continuous from Dwyka to Ecca time, but towards the north the combined effects of isostatic readjustment coupled with the onset of localised downwarping produced a complex relationship with Dwyka and pre-Karroo rocks and the basal contact surface of the Ecca sequence is markedly diachronous.

A second break in the sedimentary record almost throughout Africa is the unconformity which defines the base of the Stormberg sequence. This break is so well-defined that the deduction is inescapable that the syntectonic sediments represent a short-lived but regional tectonic event. The tectonic pattern associated with this unconformity is significant: in the Zambian terrain it is block faulting, in the Karroo terrain it is strong folding in the south. This tectonic activity heralded the final disruptive phase of Gondwanaland. Eastern Gondwanaland started to rotate away from Africa; shearing forces in the north produced the tensional fault system in the Zambezian terrain, at the same time the rotating plate crushed the southern tip of Africa, crumpling the erstwhile Karroo trough to transform it into a prominent mountain range. Prior to this spasm of activity the northern terrain was for a very long time tectonically dormant with the result that a major part of the Beaufort succession was never deposited there.

The unconformities in the Madagascan sequence cannot be directly linked with African tectonics. Even though the Madagascan block started off as part of the Zambezian tectonic terrain its disruption from Africa effectively severed tectono-sedimentary connections. This is reflected in the difficulty of correlating the Madagascan marine transgressions with similar events on the African mainland, not to mention the absence of Karroo volcanism in Madagascar.

The base of the Karroo basalts marks another important regional disconformity. In several places Karroo lavas lie on Karroo sediments and it can be demonstrated that no time break is contained in the junction, but elsewhere the lava can overstep the entire Karroo sequence. It is remarkable that in the Lebombo monocline the lava conformably overlies the Stormberg sediments for hundreds of kilometres along strike: this could only have happened if the development of the downwarp proceeded with extreme regularity, or if the downwarp developed only after a certain thickness of lava had flowed over the as yet horizontal sediments.

Seen in regional context the disconformities of the Gondwana basins in Africa are relatively minor structures because of the absence of severe tectonic disturbance. However, the regional distribution and persistence of some of these planes make them valuable in mapping, and to a certain extent they are useful time markers.

Proven marine incursions into the Karroo Basins are restricted to marginal areas of southern Africa and somehow these marine transgressions are not associated with notable structural, sedimentological or faunal polarisation. This problem needs to be investigated further. Another problem that has received insufficient attention is the shape and lateral connections of the ocean which presumably lay to the southeast of the Karroo Basin. The development of the Mocambique Strait initiated a notable series of marine floodings. It may be significant to note that towards the Middle Jurassic, vast areas of northeast Africa were inundated, indicating exposure to the eo-Indian Ocean.

The climatic changes seem to follow a logical pattern in the light of displacement of southern Africa through the various global climatological zones. The glacial episode of Dwyka time and later was produced by the movement of the Gondwanaland plate across the southern rotational pole, but Frakes and Crowell (1970b) rightly stressed that the development of the Gondwanaland ice caps was abetted by some critical deflections of moisture-bringing warm equatorial currents by projections of the supercontinent into the Tethys-Indian Ocean. It seems as if the westerlies were the main winds involved in the glacial environment of Gondwanaland. The models proposed by Frakes and Crowell
(1970b) are extremely useful in understanding the palaeogeography of the glaciated Gondwanaland.

Conditions favourable for the development of coal swamps operated at various times and places during the existence of the African Gondwana basins. The most important and widespread swamp environment followed on the regression of the ice sheets at a time when the climate must still have been quite cool and wet. The vegetation which flourished in the Ecca swamps was *Glossopteris-Ganagamopteris* as well as horse-tails, lycopods, ferns, gymnosperms, a conifer and large *Dadoxylon* trees, whose distinct growth-rings have been interpreted as indicative of marked annual (seasonal) climatic variations. This flora is a typical woodland assemblage and differs sharply from the mud-flat swamp environment of the more arid Beaufort climate.

The Ecca woodland flora inhabited the rolling hills, the shores of the numerous lakes and invaded the deltas and swamps. The wooded area was widespread in a belt northwards of the present Transvaal-Natal area. Much of the original coal peat was no doubt washed into the lakes but in many places, especially the smaller, partly land-locked swamp basins, the vegetal debris accumulated by *in situ* growth. The balance between tectonic downwarp and clastic and biogenic sedimentation was at all times a critical factor in the quality of the resultant coal. In areas of tectonic unrest conditions suitable for the accumulation of coal were never stable long enough to produce thick coal seams. The facies relationship between coal, shale and carbonaceous shale is a useful indicator of basin geography, and interbed relationships are clues to the tectonic fabric during deposition.

During Beaufort time the generally dry climate supported an entirely different swamp flora, mainly an equisetalean assemblage, and only very insignificant coal swamps developed; one is near Swaziland.

The Triassic Molteno coal swamps were almost identical to the Permian Transvaal/Natal coal swamps: a delta environment draining a wooded mountainland. The Molteno floral assemblage was a *Dicroidium* flora, which included cycads, ginkgoales, pteridosperms, ferns, *Rhexoxylon* trees and other plants. The regional setting of the swamps suggests that the climate may have been typically continental semi-arid. Once again, as in the case of the Permian coal swamps, the tectonic pattern of the Molteno Basin and its source area determined the formation and temporal sequence of the coal swamps, and as before the coal swamps were located in an area proximal to the clastic sediment source.

In the Madagascan block, swamp conditions developed at least twice. It is possible that the Sakoa swamps were part of the Transvaal-Natal deltaic complex. Towards mid-Sakamena time, carbonaceous and plant-bearing shales were deposited in what must have been a near-shore or estuarine environment. This happened at a time (Late Permian to Early Triassic) when mainland Africa was being increasingly desiccated.

The first clear signs of an increasingly arid climate occur in the Beaufort rocks (Keyser, 1966). The Madumabisa lake sediments indicate a shrinking lacustrine environment everywhere in the central Zambebian terrain. A detailed sedimentological study of the upper Beaufort sequence may well indicate that this red-bed formation marks the start of arid conditions in the Karroo tectono-sedimentary terrain.

The Molteno and Escarpment Grit rocks are products of an active tectonic regime. This tectonic overprint is very strong, but it is conspicuous that coal deposition took place only in the south where a major mountain range situated in relatively high latitudes supplied sufficient moisture to sustain the necessary plant growth.

The Red Beds and their correlates in the rest of Africa are generally accepted as having been deposited under arid conditions. Some geologists have suggested that the aeolian Cave Sandstone and its correlates are further proof of an arid climate but this is not necessarily so; in fact, Beukes (1970) frequently refers to the effects percolating rainwater had on the destruction of primary dune structure. However, there is no denying that wind was responsible for the accumulation of the Cave Sandstone and this process is more effective in dry climates. Palaeowind
directions in the Cave Sandstone indicate that at the time the southern part of Africa was still in the prevailing westerlies wind belt. Considering that the continent of South America formed the major windward area west of Africa it is reasonable to deduce that the climate in Africa during the deposition of the Cave Sandstone was typically mid-continental semi-arid.

The Karroo cycle was terminated by spectacular volcanism. The African landscape during the Middle Jurassic must have been awesome indeed. By the Late Jurassic Africa was ripped free and slowly moved into its present position.

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Evolution of the Southwestern Continental Margin of South America

JOSE FRUTOS AND ALVARO TOBAR

ABSTRACT

The palaeogeographic and structural evolution of the Andes of Bolivia, Chile and Argentina is analysed, and an interpretation in terms of palaeosubduction zones and sea-floor spreading centres is offered.

The continental margin of southwestern South America evolved in two Phanerozoic tectonic cycles. The Precambrian history of the region is little known. The Hercynian (Cambrian to Early Triassic) and the Andean (Late Triassic to Recent) cycles show the characteristics of an ideal tectonic cycle.

Successive geosynclinal basins evolved in the region, superimposing their respective directions on the preceding systems. These directions have changed clockwise from E-W for the Precambrian; WNW-ESE to NW-SE for the Palaeozoic; NNW-SSE to N-S for the Mesozoic-Cainozoic.

In this paper we assume the tectonic axes analysed for each system are produced normal to the direction of greater regional compression. These correspond to the collision component between an oceanic and a continental plate for the region under consideration.

In early Palaeozoic times a subduction zone oriented WNW-ESE was active in the vicinity of the Arauco Peninsula (37°S), changing its orientation to NW-SE in late Palaeozoic times; it is well represented in the Pichilemu area (34°30'S, 72°00'W). The palaeogeographic and structural reconstructions of pre-drift Gondwanaland suggest that the Chile Rise might have been the spreading centre associated with the subduction zones in the Palaeozoic.

The evolution of the Andean geosyncline is associated with the Chile-Peru trench and the East Pacific Rise; the rise was of prime importance as a spreading centre for the Andean orogen in the transition period between the Hercynian and Andean tectonic cycles (Permian to Middle Triassic).

The plate tectonic interpretation is a generalised one, generated from several oversimplified assumptions; it is useful as a first working hypothesis to be developed in the future.

INTRODUCTION

The geological evidence obtained to date through the regional mapping and structural analysis of the meridional Andes in Bolivia, Chile and Argentine has enabled us to elaborate a tectonic, stratigraphic and palaeogeographic history from the Precambrian to the Recent. The Andean Belt is the product of the superimposition of the structural directions of successive geosynclinal systems on...
Fig. 39.1. Palaeozoic (Hercynian tectonic cycle) palaeogeographic-tectonic scheme of southern South America
Evolution of the Southwestern Margin of South America

HERCYNIAN TECTONIC CYCLE

- Palaeozoic oceanic sediments (non-geosynclinal)
- Palaeozoic palaeosubduction zone
- Maximum extension of Palaeozoic marine basin
- Approx. distribution of Precambrian igneous rocks
- Approx. distribution of Palaeozoic igneous rocks (Hercynian cycle)
- Lower Palaeozoic granitoids (Eo-Hercynian stage)

ANDEAN TECTONIC CYCLE

- Mesozoic-Cenozoic intrusive rocks (Andean Cycle)
- Generalized profile fig. 3
- Marine Palaeozoic
- Palaeozoic Pillow Lava
- Approx. location of 2000-400 m.y. rocks (after Halpern, 1972)

LEGEND

Fig. 39.2. Palaeogeographic-tectonic scheme and distribution of plutonism in southern South America, from Precambrian to Recent
previously formed magmatic and sedimentary rocks. These structural directions have changed from E-W for the Precambrian; WNW-ENE to NW-NE for the Palaeozoic; NNW-SSE to N-S for the Mesozoic-Cainozoic (Figs. 39.2, 39.5, 39.6, 39.7).

Subsequent to a Precambrian history that is little known, the continental margin of southwestern South America evolved in two Phanerozoic tectonic cycles, whose history has been compiled with comparative ease. The two Phanerozoic tectonic cycles are the Hercynian (Cambrian to Early Triassic) and the Andean (Late Triassic to Recent), both showing eugeosynclinal-miogeosynclinal, anatactic, orogenic and taphrogenic phases and the development of successive structural stages that are characteristic in the evolution of an ideal tectonic cycle.

The present Andean cordillera were then built from a mixture of rocks of diverse ages, outcropping in belts trending NNE-SSW, and showing the effects of the Pliocene-Quaternary tectonism (Andean uplift). In no way do these belts represent the structural orientations of the former episodes. The preceding characteristics made the study of the Precambrian and Palaeozoic rocks in the Andean domain rather difficult. Knowledge of rocks of these ages is further advanced in the Bolivian and Argentinian Andean foreland, from where we have obtained the necessary information to build up a palaeo-geographic scheme capable of explaining the stratigraphic-structural characteristics of each geologic system represented in the present Andean belt.

Several authors have contributed to the geologic-palaeogeographic knowledge of southern South America, but for the present work, Harrington (1962), Borrello (1967, 1972), Ahlfeld and Branisa (1960), Cecioni (1970), Muñoz-Cristi (1968), Corvalán (1965), Ruiz et al. (1965) and more recently Halpern (1972), Mégard et al. (1971), James (1971), Vilas and Valencio (1969) have been the more important ones. In addition to these works we have made use of abundant unpublished material from our own studies along most of the length of Chile, and material from regional mapping done by several Instituto de Investigaciones Geológicas (I.I.G.) geologists to whom we are indebted.

Much of the information is presented in the form of figures to which the reader is referred, especially those dealing with the movements of South America, the areas of sea-floor spreading, and structural analyses. The conclusions regarding the orientation of structural axes are supported by over 2000 measurements done mainly by the senior author.

Tectonic and Palaeogeographic Development During the Precambrian—Pre-Hercynian Tectonic Cycles

In Chile, the westernmost region of the southwestern continental margin of South America, a vast number of rocks have been improperly attributed to the Precambrian because of their advanced metamorphism, presence of pytgmatic veins, micro-folding, successive recrystallisation and development of two or three schistosity planes (Fig. 39.9); but these structures are thought to be normal in a region that has been mobile since Precambrian times. Of all the outcrops described as Precambrian (Ruiz et al., 1960, Corvalán, 1965) we believe that only those like the Belen Schists (18°30'S, 69°30'W) that might represent part of the Arequipa Massif (Mégard et al., 1971); the El Limon Verde mica-schists (22°40'S, 69°00'W) that might represent the westward continuation of the Pampean Ranges Massif (Harrington, 1962); and the Cordillera de Nahuelbuta schists and gneiss (37°00'S to 38°30'S, 70°00'W) that might represent the westward continuation of the Patagonian 'mesocraton' (Harrington, 1962), are regarded as probably Precambrian if proper structural criteria are used. More than forty isotopic age determinations have been carried out on supposed Precambrian rocks by Pb/α, Rb/Sr and K/Ar methods (unpublished report, Laboratorio de Geocronologia, I.I.G., Chile), but no age greater than c. 400 m.y. has been obtained, probably because of metamorphism by later orogenic phases.

The rocks we attribute to the Precambrian in Chile outcrop in belts oriented N-S to NNE-SSW. These orientations reflect the late Pliocene block tectonism, but the blocks show, without exception, structures with dif-
Evolution of the Southwestern Margin of South America

HERCYNIAN GEOSYNCLINE

Fig. 39.3. Generalised tectonic-stratigraphic profile across the Hercynian Geosynclinal System
different orientations produced by all the several diastrophisms which have occurred since Precambrian times. The structural analyses done on these rocks (Figs. 39.1, 39.2, 39.6, 39.7) demonstrate that the oldest recognised structural directions seems to correspond to foldings in the Late Precambrian with axes oriented E-W with respect to the present geographic co-ordinates (Figs. 39.2, 39.7, 39.8). In the eastern Andean domain the superposition of tectonic effects is not so strong and it is even less in the cratonic domain of central-eastern South America, a result of the progressive migration to the west and clockwise rotation of the successive mobile belts (Borrello, 1972; Halpern, 1972).

The Precambrian terrain in Bolivia consists mainly of a thick detritic sequence with some calcareous intercalations, and a volcanoclastic sequence, both affected by several tectonic phases and metamorphism varying from greenschist to deep amphibolite facies (Mégard et al., 1971).

In northwestern Argentina extensive outcrops of Precambrian rocks show a much lower degree of metamorphism and tectonism than at the continental margin. On the Humahuaca to Salta Road (northwest Argentina), unconformably below a Cambrian detritic sequence marking the beginning of the Hercynian deposition, there are well-stratified, flyschoid sediments with low grade metamorphism, in which synsedimentary structures of marine type can be easily recognised. These rocks have fold axes oriented ENE and a superimposed NW set representing a later folding episode. They are intruded by granitic rocks with ages c. 500 and 600 m.y. (Halpern, 1972). Similar characteristics are shown by the Precambrian in northeastern Argentina and in Uruguay.

Considering in general the disposition and structural directions of these Precambrian rocks, we agree with Engel and Kelm (1972) in thinking they were part of an orogenic belt, probably an intracraticone one, that ex-
Fig. 39.4. Generalised tectonic-stratigraphic profile across the Andean Geosynclinal System
tended south of the central South American and African (Congo) cratons. In Africa it is represented by the Damara fold belt. The formations of the southwestern part of this fold belt are folded and metamorphosed (Salem granite, 550 m.y.). The orientation of structures in the Damara Basin is NE-SW (Fig. 39.10) (Choubert and Faure-Muret, 1971). The Kalahari craton (2500 to 2650 m.y. and 1850 to 2000 m.y.) bounds the Damara Basin on the south in Africa, and the Pampean Ranges Massif (2000 ± 100 m.y., Halpern, 1972), considered here to be its westward projection, has an analogous position in South America.

The structural orientation NE-SW in the fold belt is the response to a stress field whose main component was normal to that direction, produced by movements of the bounding cratonic masses. This late Precambrian diastrophism, named Asyntian (Aubouin, 1972) and Katangian (Choubert and Faure-Muret, 1971), has an age of about 650 to 620 m.y. The isotopic ages 2000 ± 100 m.y. (Halpern, 1972) from the Pampean Ranges Massif might represent the older African Eburnéenne diastrophism (2200-1850 m.y.).

TECTONIC AND PALAEOGEOGRAPHIC DEVELOPMENT DURING THE PALAEOZOIC–HERCYNIAN TECTONIC CYCLE

After the Late Precambrian orogenic movements a period of distension marked the initiation of the Hercynian tectonic cycle. This event is documented by the presence of extensive post-orogenic molasse deposits that, coupled with deep erosion, transformed the domain of the former orogen into a platform. Conglomerates and red sandstones demonstrating this stage of evolution are found in Valle de Humahuaca, northwestern Argentina and conglomerates and evaporites are found at Tarija, southern Bolivia. Below the
Chile Altiplano, it is probable that some of the coarse sandstones near the Cambro-Ordovician boundary represent this stage. After this platform stage rapid subsidence began in a basin elongated WNW-ESE, thus making an angle with the preceding Pre-cambrian orientations (Figs. 39.2, 39.3, 39.10). The eu- and miogeosynclinal phases of the tectonic cycle are well documented by widely distributed Ordovician, Silurian and
Devonian rocks (Figs. 39.3, 39-5). In the southwestern margin of the basin here considered to be eugeosynclinal, lavas of probable submarine origin are found inter-fingered with greywacke-type deposits that outcrop in the Coast Ranges between Arauco (37°S) and Valdivia (40°S). The rocks are metamorphosed from medium quartz-albite-muscovite schists to high grade glaucophane-lawsonite schists, with serpentinite (Hervé, pers. comm.), in parallel bands oriented WNW-ESE. The metamorphism decreases in intensity to the northeast. We have interpreted these features, following the idea of Ernst (1971), as produced by the proximity of an active subduction zone migrating slowly with time towards the northeast (Figs. 39.2, 39.3, 39.10). The preceding relations are well shown along the road from Toltén to Gorbea (39°10’S), Cañón Province, Chile.

A belt of Ordovician and Silurian flysch deposits, with structural orientations WNW-ESE, outcrops in the Coast Ranges of Aconcagua and Coquimbo Provinces, Central Chile, and in the pre-Cordillera of Mendoza Province, Argentina. This belt is interpreted here as placed between the eugeosynclinal and miogeosynclinal areas of the basin (Figs. 39.2, 39.3). In the miogeosyncline, lutites, limestones and sandstones were well represented all over the Peru-Bolivia basin (Figs. 39.1, 39.2, 39.3). During the same period clastic sediments were deposited in limited amounts on the foreland area.

After this first development of the geosyncline, an important compressive diastrophism took place in the Middle to Late Devonian, producing a widespread unconformity that can be recognised in the whole region. The tectonic axes oriented WNW-ESE, represented in Figures 39.5, 39.7 and 39.9, were produced by this event, and at the same time the sub-positive westward prolongation of the Pampean Ranges Massif emerged, becoming a definite divide between the Cuyo and Bolivia Basins. In Antofagasta Province, Chile, the extreme westward projection of the Pampean Ranges Massif shows Ordovician, Silurian, and probably Lower Devonian marine fossiliferous rocks that

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Fig. 39.6. Equal-area (Schmidt net) structural analysis and representation of principal compressional-extensional tectonic movements in different stratigraphic units and places. Geometric analysis of superimposed fold systems (lower hemisphere projection).
Fig. 39.7. Equal-area (Schmidt net) structural analysis and representation of principal compressional-extensional tectonic movements, in different stratigraphic units and places. Geometric analysis of superimposed fold systems (lower hemisphere projection).

Fig. 39.8. Orientation of B axes of folding related to the diastrophic phases recognised in the Chilean Andes. The progressive clockwise reorientation of the principal axes of compression is documented by 6,120 measurements.
emerged in the Late Devonian and were again covered by marine rocks during the Hercynian cycle.

The Middle-Late Devonian diastrophism terminates the first structural stage in the Hercynian cycle, here named Eo-Hercynian (Fig. 39.5), following the nomenclature of Mégard et al. (1971).

Marine deposition evolved independently in the Cuyo and Bolivian Basins between the Devonian and Late Carboniferous. Cuyo Basin rocks of this age are represented in the Coast Ranges of the Coquimbo and Aconcagua Provinces, Chile, by the Estratos de Huentelauquén y Puerto Manso (31°30'S, 71°30'W) (Muñoz-Cristi, 1968). The Bolivian Basin has extensive outcrops described elsewhere by Schlatter and Nederlof (1966). The Carboniferous sandstone and fossiliferous slate outcropping at Juan de Morales, Tarapacá, Chile (20°10'S, 69°20'W), represent the southwestern border of the miogeosynclinal Bolivia Basin.

Orogenic movements brought to an end marine sedimentation in the Hercynian geosyncline, and produced extensive Late Carboniferous to Early Permian folding which can be traced throughout the whole region. This diastrophism defines the second structural stage of the Hercynian tectonic cycle named Late-Hercynian (Mégard et al., 1971) (Fig. 39.5).

The structural orientations produced by the Late Hercynian diastrophism are oriented WNW-ENE to NW-SE at an angle with respect to the preceding Eo-Hercynian structures (Figs. 39.1, 39.2, 39.7). We have interpreted the new structures as parallel to a subduction zone with a new orientation and a new direction of movement (not represented in the figures). The existence of this new or re-oriented ancient subduction zone making an angle with the Eo-Hercynian one is documented by the presence at Pichilemu, Chile (34°30'S, 72°00'W) of metamorphic rocks in bands oriented NW-SE juxtaposed with the progressively metamorphosed belts...
Fig. 39.10 Migration of Pangaea-Gondwana Blocks related to ancient sea-floor spreading centres and palaeotrenches.
of the Eo-Hercynian structural stage described above (Del Campo, pers. comm.).

During the Hercynian cycle broad bands of granitic intrusives with ages decreasing progressively to the northeast were emplaced. In the Arauco region (37°S) they give isotopic ages of c. 450 m.y., in O'Higgins and Valparaiso Provinces (33°S to 34°S) ages of c. 350 m.y., and in Coquimbo and Atacama Provinces (30°S to 32°S) ages of c. 270 m.y. This zonation and polarity is repeated in the Bolivian Basin domain, Ordovician granites occurring in the Pampean Ranges Massif and Carboniferous-Permian on the northeast inside the basin. Whether this is the product of another northeastward-dipping subduction zone active in the vicinity of the Pampean Ranges Massif, remains an open question.

Following closely after the Late Hercynian orogeny, there was a period of distension accompanied by deep and widespread erosion and deposition of detritic continental sediments or post-orogenic molasse and keratophyric and rhyolitic volcanism. This period represents the Taphrogenic Phase which brought to an end the Hercynian tectonic cycle (Fig. 39.5). The volcanic centres active during this phase had an orientation close to that of the subsequent Andean chain.

With the development of this stage, named the Fini-Hercynian structural stage, the Hercynian orogen became an extensive peneplain (Corvalan, 1965).

The palaeogeographic and structural reconstruction of the pre-drift Gondwanaland suggest that the Chile Rise (Herron and Hayes, 1969), might have been the spreading centre associated with the evolution of the southwestern continental margin of South America during the Palaeozoic. If we suppose the Chile Rise fixed in its position through time, the assumed spreading directions match adequately with the compression directions reconstructed for the continent, and with the inferred palaeo-subduction zone for the late Palaeozoic (Fig. 39.10). The spreading rates have evidently been quite considerable. We suppose that, during the transition period between the Hercynian and Andean tectonic cycles (Permian to Middle Triassic), the East Pacific Rise began its activity as a spreading centre in an incipient way before taking control of the tectonic and palaeogeographic evolution of the Andean Cycle which was expressed in the development of the Andean Geosyncline (Figs. 39.10, 39.3, 39.4).

TECNIC AND PALAEOGEOGRAPHIC DEVELOPMENT DURING THE MESOZOIC-ANDEAN TECTONIC CYCLE

The Andean tectonic cycle began with a Middle to Late Triassic distension stage producing the initial subsidence in the basin that was to show later eu- and miogeosynclinal phases. The basin was oriented NNW-SSE (Fig. 39.4), at an angle to the former Palaeozoic basin orientations, and placed on the southwestern continental margin of South America, but at an angle with the present coastline. During the initial and intermediate phases of evolution of the tectonic cycle (Jurassic to Early Cretaceous) the Andean Geosyncline consisted of a eugeosyncline with greywacke and important intercalated volcanics well represented in the present Coast Ranges of northern Chile by La Negra Formation and its equivalents (Tobar et al., 1968; Thomas, 1970; Cecioni and Garcia, 1960). Toward the foreland a miogeosyncline received sediments of sandstone, limestone and lutite but with almost no volcanics. Both domains were separated, along the whole length of the basin, by an indefinite partially submerged ridge, termed 'dorsal mesoliminar' (Frutos, 1973). This ‘wrinkle’ emerged and migrated with time towards the foreland and finally overrode the miogeosynclinal area during the orogenic phase of the Andean cycle, especially during the compressive diastrophism in the Late Cretaceous-Early Tertiary (Laramide).

By the end of the Jurassic the Nevadian or Araucano diastrophic movements made the marine domain recede, leaving a narrow Titonian-Neocomian basin that finally disappeared as a consequence of the mid-Cretaceous sub-Hercynian diastrophism. This compressive phase produced fold axes and other meridionally oriented structures, which make an angle with the orientation of the long axis of the basin.

In the late Miocene the structural axes in most of the basin were still meridionally oriented, but a transverse compression in
Evolution of the Southwestern Margin of South America

southern Peru, northern Chile, and Bolivia produced NE axes due to movements in the Santa Cruz Elbow.

The taphrogenic phase of the Andean cycle is manifested by acidic volcanism which produced the extensive ignimbritic cover in the Altiplano beginning in the Pliocene. The fractures associated with the volcanism and the ones giving rise to the principal structural blocks of the continental margin the Coast Ranges, Central Valley, and Andes Cordillera are oriented NNE, parallel to the present coast, which is also a reflection of these structures.

During the whole development of the Andean tectonic cycle active volcanism and plutonism, associated with an evolving subduction zone, took place in bands parallel to the structural axes in each period, migrating steadily eastward (Ruiz et al., 1965; Farrar et al., 1970). In Jurassic times the centre of the volcanic belt was located a little to the west of the present coastline in northern Chile, and from there migrated inland approximately 200 km to its present position at the crest of the Andes, making an angle with its initial orientation.

In the southernmost part of South America, a partially independent Magaellanian geosynclinal basin, not discussed here in detail, evolved at the same time as the Andean geosyncline, and included comparable morphotectonic elements. Its principal difference is the oroclinal bending of the southernmost tip of South America beginning in the Late Cretaceous, maintaining the angularity of the preceding structural elements, but rotated about 90° from their original position (Frutos et al., 1972).

At the end of the Hercynian tectonic cycle and the beginning of fragmentation and drift of Gondwanaland, the Pichilemu-Arauco subduction zone and its associated Chile Rise spreading centre began to decline and were replaced by the East Pacific Rise and the Peru-Chile trench. This pair developed gradually, reaching in the Jurassic an importance maintained up to the present (Fig. 39.10). The observed orientation of the Jurassic axes is supposed to be the response to a situation where both the ancient Chile Rise and the new East Pacific Rise were active, producing a component directed towards the NE to ENE.

Polar wandering and sea-floor spreading data show that when South America drifted apart from Gondwanaland (assuming a static Africa) it followed an arcuate path starting towards the SW in the Jurassic and changing progressively to the W and WNW at present. We suppose that at the same time the active spreading centre changed progressively from the Chile Rise to the East Pacific Rise. This model fits the observed compression directions and magmatic features of the Andean Geosyncline. The curved path might be due to the reorientation of the convective cell that was rafting South America.

The several compressive pulses and associated magmatism observed in the different tectonic phases are supposed here to be related to episodes of greater spreading rate or plate convergence, and the distension events to lower rates. This enables us to define the various structural stages (Fig. 39.5) for each tectonic cycle.

The Andean cordillera are oroclinally bent and transversally slip faulted at several places besides the Santa Cruz Elbow and Patagonia. These features are supposedly produced by differential drift rates of continental blocks of various sizes. The Santa Cruz Elbow represents the extreme of this situation.

REFERENCES


Abstract

Australia's development during the past 300 m.y. has been dominated by three events: rift-drift on the western and southern margins; subduction and marginal sea spreading on the eastern margin; and collision in the north (Timor, Papua).

On the western margin, rifting started in the Early Permian, and continued in the northwest until sea-floor spreading started in the Late Jurassic, and in the southwest until the Early Cretaceous. On the southern margin, rifting started in the Late Jurassic, and continued to the onset of spreading in the Palaeocene. Along these margins, nearly all the igneous activity (mainly basalt flows and intrusives) is explicable in terms of the rift-drift history. Another connection is that between drift and marginal subsidence, as shown by marginal marine transgressions.

The subduction events of the eastern margin are manifested by plutonic and volcanic activity, and by marine transgressions over the stable platform of the continental interior.

Up to the climactic collision with volcanic arcs in the Middle Miocene, the Timor margin faced Tethys, and accumulated a predominantly marine succession.

The individual tectonic timetable of each of the four Australian margins was overshadowed by a continent-wide submergence in the Aptian.

Introduction

Our aim in this paper is to trace the Late Palaeozoic and Mesozoic (300 to 65 m.y.) history of Australia. The Cainozoic (65 m.y. to present) is dealt with only to round off the description of phases of the continent's geological development that had earlier beginnings. If anything is known in geology, it is the existing state of the world, and so, we start with the end-product of geological development, the present configuration of Australia and adjacent areas, for it is to this point that our history leads.

In terms of plate tectonics, Australia lies in the eastern part of the Indian Plate (Fig. 40.1), and is bounded by the convergent volcanic and seismic Benioff zones of the Indonesian Arcs, northern New Guinea, New Britain, the Solomon Islands, the New Hebrides, and the Tonga-Kermadec Arc; a seismic zone marking a transform fault through New Zealand and the Macquarie Ridge; and the divergent ridge that lies half-way between Australia and Antarctica. The Ninety East Ridge provides a convenient boundary between this eastern region and the rest of the Indian Plate.

The continental block of Australia occupies a position some five degrees of latitude north of the centre of this region and its northern limits, represented by Timor and southern New Guinea, coincide with the northern convergent plate boundary. Its other sides are limited by oceans.

During 1972 and 1973, the Deep Sea Drilling Project drilled at thirty-three sites.
Fig. 40.1. Eastern part of the Indian Plate. Bathymetry: Physico-Geographical Atlas of the World, 1964: Moscow, U.S.S.R. Acad. Sci.; Seismicity: Baraganzi and Dorman (1969); Cleary and Simpson (1971); Doyle et al. (1968). Volcanicity: Holmes (1965); Kuenen (1950). Sea-floor spreading: Magnetic anomalies: Christoffel and Falconer (1972); McKenzie and Sclater (1971); Weisest and Hayes (1972). Ages of oceanic basement: Andrews et al. (1973); Burns et al. (1972); Hayes et al. (1973); Kennett et al. (1973); Luyendyk et al. (1973); Vevers et al. (1973); von der Borch et al. (1972); Winterer et al. (1972).
The deep sea drilling recovered oceanic basalt overlain by basal sediment, whose age west of Australia is Late Jurassic (150 m.y.) to Palaeocene (60 m.y.), south of Australia Late Eocene (45 m.y.), in the Tasman Sea latest Cretaceous (65 m.y.), and in the Coral Sea Early Eocene (50 m.y.). With other evidence, notably the sequence of sea-floor spreading, magnetic anomalies south of Australia and the less easily recognised anomalies in the Tasman Sea (not shown in Fig. 40.1) given by Hayes and Ringis (1973), these ages are interpreted as dating the generation of the ocean west and south of Australia by sea-floor spreading, and east of Australia by the obscure process of marginal sea formation. According to this view, the Australian continental block separated from its Gondwanaland neighbours during the late Mesozoic and Palaeocene; from India on the west during the Late Jurassic and Early Cretaceous; and from Antarctica on the south during the Palaeocene and Eocene. The fragmentation of its eastern edge started in the latest Cretaceous. Since continental breakup was but the last in a long series of geological events, much of what follows will be devoted to an examination of these preliminary events in the earlier history of Australia.

HISTORICAL OUTLINE

We present the data that we judge to be significant in a series of simplified palaeogeographic maps (Figs. 40.3-13) and diagrammatic stratigraphical columns (Figs. 40.14-17).

Fig. 40.2. Gondwana basins of Australia. Basins formed during the Late Palaeozoic, Mesozoic, and Early Tertiary evolution of the Australian continental fragment of the Indian Plate. They are divided between the West, South, East, and North Sectors discussed in the text. The stippled area represents water depths greater than 2000m. Closed circles locate selected deep test wells on the continental shelf, open circles identify sites drilled in the Deep Sea Drilling Project Legs 21, 22, 27, 28 and 29. Late Carboniferous-Triassic basins underlying the Great Artesian Basin are: A—Arckaringa, P—Pedirka, C—Cooper, G—Galilee. S is the Jurassic-Cretaceous Surat Basin.
The maps acknowledge the reconstruction of Gondwanaland with the Lord Howe Rise/New Zealand segments against the eastern margin of Australia, and Antarctica against the southern margin. A continental mass, once including the blocks now comprising India, Broken Ridge, Kerguelen Plateau, and possibly a portion of Southeast Asia, lay against the western margin. The relative fit of these fragments against Australia is open to discussion beyond the scope of this paper: that they were present is apparent from the known geology of the basins that lie along the western and northwestern margins of the present Australian continent.

Considerable licence has been taken in the construction of the palaeogeographic maps. Apart from simplifications demanded by the limitations of displaying maps of so large an area in this printed format, they have been grouped according to irregular units of geological time to emphasise the major historical changes. In the process, however, some minor but no less distinctive events have been combined, and some major system boundaries have been relegated in significance to time lines within a major sequence.

The bases for much of the grouping are:
1. The palynological scales now applied throughout Australia for much of the subsurface sections of Devonian to Early Tertiary age.
2. Changes in depositional regimes and tectonic events which may be related to broad climatic changes or to major episodes in the fragmentations of Gondwanaland.

We find it expedient in the following discussion to consider Australia in terms of four sections (Fig. 40.2). Some overlap of presentation is unavoidable, such as in the grouping of various parts of the Murray and Tasmania Basins in the East and South sectors (Fig. 40.15). The maps that follow (Figs. 40.3–13), with their expanded captions, provide a cinematographic view of the geological evolution of Australia, and are accordingly presented here before the analysis of the sequence of events in each sector.

**West Sector**

The West Sector (Fig. 40.14) includes the Canning, Carnarvon, and Perth Basins and their offshore extensions, and the offshore Browse and Dampier Basins. They formed in response to a major tensional stress between Australia and the 'Indian' segment of Gondwanaland over a long period of time, at least from the Late Carboniferous until the Aptian, when oceanic crust finally appeared between the drifting continental segments.

**Sedimentary History**

Except in the southernmost part of the sector, which was non-marine until the Aptian, non-marine glacigene deposits in the Late Carboniferous and at the base of the Permian were succeeded by a Sakmarian/Artinskian marine sequence (Holmwood Shale, Fossil Cliff and Callytharra Formations, Nura Nura Member) and, after a regression, by Artinskian/Kungurian marine sediments. The final Permian marine event is registered by the Tatarian Hardman Member of the northern Canning Basin. The mixed marine/non-marine basal Triassic Blina Shale of the northern Canning Basin and the marine Kockatea Shale of the northern Perth Basin were succeeded in the latest Scythian by non-marine sandstone. The Dampier Basin and the marginal Carnarvon Basin, in contrast, remained marine up to the earliest Late Triassic. Except for the Middle Jurassic Cadda Formation, the Perth Basin was non-marine during the Jurassic. On the western margin of the Carnarvon Basin, the Jurassic Dingo Claystone accumulated in a deep marine trough. A marine transgression swept across the northern part of the sector in the Oxfordian/Kimmeridgian. A transgression in the Aptian left the South Perth Formation and Dandaragan Sandstone in the Perth Basin, the Birdrong Sandstone in the Carnarvon Basin, and in the Canning Basin at the peak of the transgression, the Bejah Beds. Deposition persisted in most areas into the Cenomanian, and after a hiatus in the Coniacian and locally also in the Turonian, it returned in the Santonian. The Santonian marked a radical change in the source of the sediments: from almost wholly detrital before the Santonian to almost wholly calcareous during the Santonian and later. On the Exmouth Plateau, a velocity change at a relatable horizon presumably reflects the change from neritic sediments on the shelf...
Late Palaeozoic and Mesozoic History of Australia

Fig. 40.3. Late Carboniferous—Early Permian. This sequence encompasses palynological stages 1 and 2 (Evans, 1969). Black, White and Morgan (1972) showed that in the volcanic provenance of northeastern Queensland, the *Glossopteris* flora was present 314 m.y. ago, i.e. in the Late Carboniferous, and this map probably represents a Late Carboniferous time only. It certainly represents the period when most of the so-called glacigene sediments were laid down.

Four provinces are recognisable:

1. The western sag of the Perth and Carnarvon basins, regarded as forming along what later became a fundamental fracture between Australia and 'India'. 2. A series of non-marine basins forming an arc from the Canning Basin in the west, through the Pedirka and Cooper Basins to the Galilee Basin in the east. 3. A southern region of now isolated basins, but assumed to have been once interconnected, characterised by the presence of marginal marine sediments. 4. An eastern, volcanogenic province which extended north-south along the present Australian margin.

Most of these provinces were cross-connected. Thus the southern, marginal marine province extended as far north as the Arckaringa Basin and the shape of the Galilee Basin appears to have been affected by structuring of the eastern volcanic province. Furthermore, the marine influence which affected the south also intruded into the southern margin of the eastern volcanic belt.

Identification of these provinces allows us to draw certain generalisations about Gondwanaland and the early Gondwana glaciations.

The arcuate ring of non-marine basins stands around the presumed high ground of the Kimberley and Arunta/Amadeus blocks. The Amadeus in particular was a region of folding of Alpine proportions during the Alice Springs orogeny of mid-Carboniferous age and may be assumed to have had greater elevation than at the present day. The Pedirka, Arckaringa and southern Canning regions were well located to receive outward sediment from glaciations on that high ground.

The southern province includes the region where more distinct signs of glaciation in the form of basement striae are evident. Although sedimentation could not proceed until the erosive ice had retreated from the area, it is possible to imagine that the southern region was depressed by a north moving ice-sheet, to be later infilled by a shallow sea with sediments deposited during retreat of the ice. Glacigene complexities of the western and eastern provinces would have been more local in origin.

This was the time when Gondwanaland 'stirred'. The initial western rift, the trends formed by wrenching which became the Fitzroy Trough of the Canning Basin and the basement structuring of the Cooper Basin each reflect the onset of tectonic events which eventually resulted in the fracture and dispersal of Gondwanaland.
Fig. 40.4. Early Permian. This period includes palynological Stages 3 and 4 of Evans, i.e. the Sakmarian and Artinskian.

The Australian portion of Gondwanaland became relatively quiet at this time. The western rift continued to develop. The areas of the central basins receiving sediment shrank in size, possibly due to a combination of maturing topography and reduction in the rate of erosion—in turn a function of the amelioration of the overall climate. The marginal seas of the southern province retreated, yet the sea managed to enter the eastern province across the foundered remnants of the Late Carboniferous andesite belt. Climatic changes permitted freshwater coal swamp conditions to develop in the central areas.
Late Permian

Fig. 40.5. Late Permian. This interval represents palynological Stage 5 of Evans and is labelled 'Late' Permian for convenience only. Strictly speaking it commences in the late Artinskian, i.e. Early Permian. Furthermore, there is the probability that on a worldwide scale the end of Stage 5, i.e. the last of the Glossopteris flora in Australia, may have occurred before the end of the Permian.

It is marked in the west by continuation of the rift system into which periodic marine incursions made their way. Deposition was notably more marine at the northern end of the rift. The Cooper and Galilee Basins in the central area continued to infill with freshwater coal swamps, and the sea entered the Sydney and Bowen Basins in the last major pre-Cretaceous marine transgression.

The sequence ended in the eastern region with extremely widespread coal swamp deposition and a relatively mature topography.
Fig. 40.6. Early Triassic. Palynological units Tr 1 and 2 (Evans, 1966) are represented by this interval. The precise definition of the actual time span covered, and details of the palaeontological sequence await further work. Throughout Australia the period differs from its predecessor by the disappearance of coal-forming climatic conditions. The sea made its way into the western rift system, even as far south as the Perth Basin, where the ammonite-bearing Kockatea Shale was deposited. The crustal movements causing this transgression, either relative rise in sea level or a lowering of the rift-valley floor, could be further reflected by the change in rift position westwards in the Carnarvon Basin, perhaps reflecting the first major exten­sional movement that culminated in the separation of the 'Indian' segment from Australia.

All the basins in the east which received sediment in Permian time continued to founder in the Early Triassic. However, the eastern margin of the Bowen and Sydney Basins was modified by uplift and volcanic extrusion. As presented on the map, the marine Early Triassic of the Maryborough Basin was isolated by these movements from the inland basins. There is the enigma, on the other hand, of an occurrence in the southern Galilee Basin, of acritarchs which are normally thought to represent marginal marine conditions of deposition, and it is conceivable that very shallow and ephemeral sea-ways extended further across the continent than might be generally supposed.

The Lower Triassic of the Cooper, Galilee, Bowen, Sydney and Tasmanian Basins character­istically includes red-beds which represent the acme of a climatic trend towards a warmer regime from that experienced by the region at the beginning of Gondwana time.
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**CARBONATES**

**MARINE MARG. MAR.**

**FLUVIATILE COAL MEAS.**

**EXTRUSIVES**

**Fig. 40.7.** Mid-Late Triassic. The trends initiated in the Early Triassic were maintained during the remainder of Triassic time. The crustal stresses lessened and in both western and eastern areas there was a net infilling of the basins formed with largely non-marine sediments. A maturing of the land forms occurred so that in the east at least extensive coal swamps were able to develop.
Fig. 40.8. Rhaetic-Early Jurassic. The Rhaetic is classed here as part of the Triassic, but is grouped with the Early Jurassic palynological units J1-3 in order to express the major change in depositional sequences in eastern Australia which took place in the Triassic, pre-Rhaetic time. The long existing depression of the Cooper Basin became the site of a southwest-northwest flowing river system which cut across the north-south grain of the older Bowen-Sydney trend to extend eastwards towards the Lord Howe Rise. This new trend controlled sedimentation throughout the Jurassic.

Briefly, during the earliest Jurassic J2 time there was an extension of marginal marine conditions across the Australian block. For the rest of the time deposition was largely fluviatile.

The western rift again experienced the ingress of the sea at its northern end. The existence of marginal marine conditions across the northern segment of the continent is speculative, but is expected to have foreshadowed the distinctly marine conditions of the succeeding Mid-Late Jurassic time.
Fig. 40.9. Middle Jurassic-Early Neocomian. This period covers palynological unit J4 to mid K1a (i.e. to mid-Crybelosporites stylosus Zone time). The position of the Jurassic-Cretaceous boundary is not clearly defined, but is not critical to understanding the important events which took place within this interval.

In the east the general region of the Great Artesian Basin continued to sag. The river system which developed early in Jurassic time expanded, but continued to find outlets to the east. Basement movements in the Late Jurassic created an arch, the Nebine Ridge, between the Surat and Eromanga Basins. The continental masses which in the region of the present western margin of Australia had for over 100 m.y. been the site of a complex rift system, in Late Jurassic time began to drift apart. The opening began in the north where oceanic crust made its first appearance in the rift floor. Marginal and open marine conditions extended around the northern sector of the continent. During this time the first rifting phases of the southern margin, between Australia and Antarctica, developed. These appear as east-west graben-like structures of the Polda, Robe, Otway and Gippsland Troughs. Injection of the Tasmanian dolerites and extrusion of the Late Jurassic volcanics of the Robe Trough are thought to be related to this rifting period.
Fig. 40.10. Late Neocomian-Early Albian. During this interval much of the present Australian continent was inundated by shallow seas. Open ocean lay to the west and north of the continent and normal shelf sediments were deposited around that margin. Uplift of the eastern margin cut the connection between the Surat Basin and an ocean further east. Ingress of the sea over continental Australia was only from the north. Nevertheless, during the Aptian, the time of maximum transgression, the sea extended for over 2000 km to the south of the Papuan shelf. Marine beds are recorded as far south as Renmark in South Australia.

In contrast, the rift-fill sediments of the Otway region are almost entirely non-marine. Only at the most northerly of sampled locations in the Otway Basin is there evidence of an ephemeral spill southwards from the Renmark region across the intervening high. This situation fits the classic rift model in which arching and keystone collapse at the early stages of the process leaves a marginal high separating the rift from the hinterland. With this model in mind, the anomalously marine Aptian of the Eucla Basin may have been connected northeastwards to the marine embayment of the Great Artesian Basin, rather than southwards into the Antarctica-Australia rift system.
Late Albian - Cenomanian. During Albian times the seas over eastern Australia retreated northwards. To the west and north, outward building of the continental shelf onto the Mesozoic oceanic crust continued apace. Sedimentation continued in the centre of the Eromanga Basin, but was non-marine.

To the south the separation of Australia from Antarctica began to gather momentum. Shallow but distinctly marine conditions prevailed for the first time, particularly westwards in the Eucla Basin; only marginal marine conditions existed in the east of the rift, in the Otway Basin.
Fig. 40.12. Turonian-Maastrichtian. During this, the remainder of Cretaceous time, the western and northern borders of the continent continued to develop as normal ocean margins. The separation of Australia from Antarctica progressed steadily, but still was insufficient for oceanic crust to be emplaced in the rift centre and for other than very restricted marine conditions to influence the eastern end of the rift. A new fracture began to develop for the first time during this period—the break which separated the Lord Howe Rise from eastern Australia and which marked commencement of formation of the Tasman Sea. Marginal marine sediments were deposited in the Gippsland Basin and shallow water sediments were laid down on the foundering Lord Howe Rise.
Fig. 40.13. Early Tertiary. During the Early Tertiary, Australia took on the basic shape of the present continent when the Coral Sea opened up and the Queensland Plateau began to founder. A narrow Southern Ocean and a Tasman Sea came into being. The Indo-Pacific Ocean lapped around the continent's western and northern margins.

Australia was finally separated from all other major components of the original Gondwana supercontinent.

Subsequent interactions of the northern margin of the continental mass with the Pacific plate in the region of Papua-New Guinea and with the subduction zone of the Indonesian Arc in Timor during the Neogene are events which affected Australia only and are not thought of as part of the 'Gondwana' story.
Fig. 40.14. Timetable of sedimentary, volcanic and structural events in the West Sector. Time scale from Harland et al. (1964, 1971). Localities shown in Fig. 40.2. Sedimentary events: five environments are indicated—U (uplift), L (land), M (mixed marine/non-marine), S (shallow marine), and D (deep marine). Glaciogenic sediments (solid triangles). Volcanics: V's. Structural events: unconformity ~
to deeper-water sediments of this marginal plateau (Veevers et al., in press a). There is a notable hiatus in the stratigraphical record in the Palaeocene.

Igneous History In the West Sector, the only known outcrops of igneous rocks in the period 0-300 m.y. are the Bunbury Basalt of the south Perth Basin, of latest Neocomian age, dated by reference to the enclosing sediments; and the Fitzroy Lamproites of the north Canning Basin, which were radiometrically dated as 17 to 21 m.y. (Early Miocene) by Wellman (1973). Other igneous rocks are known from drilling. By superposition, basalt at the DSDP Site 259 is basal Aptian (112 m.y.), and at Site 261 latest Oxfordian (150 m.y.). Late Jurassic basalt occurs in BOC Scott Reef-1, and Ashmore Reef-1, 450 km east of Site 261. Dolerite that intrudes Carboniferous sediments in Wapet Barlee-1 in the Canning Basin is radiometrically dated as 196 m.y. (Harding, 1967). Rhyolite in BOC Enderby-1, in the Dampier Basin, is dated by marine microplankton from intercalated sediment as earliest Triassic or possibly latest Permian. In Ocean Ventures Edel-1 in the Carnarvon Basin, a carbonatite-like association of phonolites, lamprophyres, and trachytes is dated radiometrically as 240 m.y. and 260 m.y. On the time scale adopted here, the 240 m.y. date corresponds with the Permian/Triassic boundary, and hence with the age of the rhyolite in Enderby-1.

Structural History Rapid subsidence of fault-bounded blocks, due to incipient rifting in the Carnarvon and Canning Basins and gentle downwarping elsewhere, accompanied the Late Carboniferous-Sakmarian phase of glacigenic deposition. Rifting started in the Perth Basin in the Late Permian and persisted into the Early Cretaceous. The initial deposition of the Dingo Claystone in the Carnarvon Basin, presumably into a graben, dates from the Early Jurassic.

Sea-floor spreading began off the northern part of the sector in the Oxfordian and off the southern part in the late Neocomian, where an intense phase of block faulting and erosion preceded continental breakup. The Dampier and Carnarvon Basins collapsed seaward in the Cenomanian, some 15 to 20 m.y. after continental breakup, and this persisted to the end of the Coniacian. Thereafter, the offshore parts of the basins of the west sector have remained covered by a shallow sea, except for the Exmouth and Naturaliste Plateaus, which have remained deeply submerged since the Late Cretaceous.

South Sector

The South Sector (Fig. 40.15) includes the Eucla, Duntroon, St Vincent, Otway, Bass, and Gippsland Basins. They were largely rift fill basins during the process of separation of Antarctica from Australia during the Late Mesozoic and Early Tertiary. The Murray Basin in Tertiary time is thought of as part of this sector, but in the Cretaceous and older periods it is better regarded as part of the East Sector.

Note that the Cretaceous and older rocks of the Murray Basin and the Triassic and Permian rocks of the Tasmania Basin are regarded as part of the East Sector.

Sedimentary History Deposition started with glacigenic sediments (non-marine to mixed marine/non-marine) in the Late Carboniferous-Sakmarian lying unconformably on Lower Palaeozoic to Precambrian rocks. The next record is of a ?Late Triassic non-marine sandstone at Bacchus Marsh, although Late Permian spores, presumably derived from sediments of that age that once existed in west Victoria, have been recovered from the Mesozoic of the Otway Basin. A second phase of deposition started in the Late Jurassic. An eastern re-entrant of the Eucla Basin (called the Polda Trough) accumulated non-marine sediments (Parkin, 1969). The western Otway Basin contains basalt dated radiometrically at 153 m.y. (Harding, 1967: 91) interbedded with Late Jurassic sediments overlain by the Early Cretaceous Otway Group. In the Gippsland Basin, basalt is overlain by the Early Cretaceous Strzelecki Group. No distinctly marine sediments in the Mesozoic appear in the Otway Basin until the late Albian-Cenomanian, but in the Eucla Basin, a marine transgression occurred in the Barremian and Aptian. Until the Eocene, the Otway Basin was crossed only intermittently by
the sea, but during most if not all of the Late Cretaceous, the Eucla Basin was marine. With the Middle/Late Eocene marine transgression, the South Sector, except Tasmania, acted in concert, and thereafter remained essentially marine.

Deep Sea Drilling Project Site 282, off the west coast of Tasmania, recovered Late Eocene clay on pillow basalt.

**Igneous History** No igneous rocks are known in the Eucla, St Vincent, and Murray Basins.

The intrusion of the Tasmanian Dolerite 167 m.y. ago (McDougall, 1961), and the extrusion of the Casterton basalt 153 m.y. ago and the basalt at the base of the Strzelecki Group, presumably in the earliest Cretaceous, mark the beginning of the main phase of deposition of the South Sector. Except for basalt extrusion in western Victoria during the Albian, and the intrusion of syenite in Tasmania 100 m.y. ago (the latter probably connected with events concentrated in the East Sector, no primary igneous rocks are known in the rest of the Cretaceous. Nevertheless, most of the Lower Cretaceous of the Otway Basin contain large proportions of detrital acid to intermediate volcanic material, sufficient in places to be described as a 'volcanic sandstone'. During the Cainozoic, in contrast, the eastern part has an almost continuous record of primary volcanism that continues almost to the present day.

**Structural History** The South Sector was firmly welded to Antarctica during most of the Phanerozoic until the Late Jurassic. Widespread subsidence and localised faulting induced ingress of a marginal sea during the Late Carboniferous and Early Permian. Thereafter the region remained relatively stable until initial rifting that resulted in the formation of the east-west oriented graben of the Polda, Duntroon, and Western Otway Troughs, commenced in the Late Jurassic. Separation of the continental masses of Australia and Antarctica was completed by the Eocene. In the intervening period, numerous down-to-basin faults developed in all the rift basins. The greater continuity of Late Cretaceous marine sedimentation in the Eucla Basin compared with the Otway Basin, implies that the seaway over what later became the boundary between Australia and Antarctica, entered from the west. The Bass and Gippsland Basins followed a different course because of the continued presence of the Tasmanian block between Australia and Antarctica. The Bass Basin formed as an extensional sag. Unlike the others, the Gippsland Basin was influenced by the east-west extensional stress associated with the separation of the Lord Howe/New Zealand mass in Late Cretaceous/Early Tertiary time. As Antarctica drifted away during the Tertiary, the southern Australian margin progressively sagged and subsided, resulting in the formation of the shallow Eucla, St Vincent, and Murray Basins, and the shelf margin Duntroon and Otway Basins. Movements between Tasmania and Australia induced further faulting and folding in the Gippsland Basin.

**East Sector**

The East Sector (Fig. 40.16), with its mobile belts, plutonic and metamorphic activity, and growth of marginal seas, reflects convergent plate activity. Its history is of a greater complexity than that of the sectors already described, and we restrict our attention here to aspects of its sedimentary and igneous history only. Note that the Cretaceous and older rocks of the Murray Basin and the Permian and Triassic rocks of the Tasmania Basin, which are shown in Figure 40.15, are grouped in the East Sector.

**Sedimentary History** The western and southern part of the sector (Arckaringa, Pedirka, and Cooper Basins) was a site of non-marine glacigene deposition in the Late Carboniferous-early Sakmarian (the Arckaringa Basin also registered the Sakmarian transgression which affected the South Sector), and intermittent non-marine deposition during the rest of the Permian, Triassic and Jurassic. The eastern part of the sector (Gippsland Basin and eastwards) had a similar Late Carboniferous to Jurassic sedimentary history except for marine incursions during parts of the Permian, Triassic, and Early Jurassic; it also has a generally thicker record, particularly in the Permian of the Sydney and Bowen Basins.

All the basins of the East Sector, except
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Fig. 40.15. See caption to Fig. 40.14.
Fig. 40.16. See caption to Fig. 40.14.
the Clarence-Moreton and Sydney Basins, were covered by the Aptian-Albian sea, which retreated in the late Albian. Thereafter, except possibly for a brief marine incursion from the north in the Miocene (Lloyd, 1968), the onshore history is of intermittent volcanism and non-marine deposition.

Eastward, parts of the Cretaceous to Cainozoic history have been revealed by deep-sea drilling. Site 283 in the central Tasman Sea has latest Cretaceous/Palaeocene claystone on basalt. The southern part of the Lord Howe Rise (Site 207) was shallow in the Late Cretaceous, and thereafter to the Middle Eocene subsided to its present depth; the northern Lord Howe Rise (Site 208) seems to have been at or near its present depth since at least the late Maastrichtian. Burns et al. (1972: 16) suggest that 'the Lord Howe Rise (and by implication the Tasman Basin, which was not sampled) may have been in existence by the Late Cretaceous'. At Site 287, in the Coral Sea Basin, Early Eocene chalk lying on basalt indicates that this basin came into existence in the Early Eocene. At Site 209, on the adjacent Queensland Plateau, the deepest recovered sediment is Middle Eocene; the section at this site indicates progressive deepening of the Queensland Plateau since at least the Middle Eocene. Two deep exploration wells in the Capricorn area of the Great Barrier Reef (Gulf Oil Aquarius-1 and Capricorn-1) reveal Palaeocene-Oligocene non-marine sediments, including red-beds, evaporites, lignite, and conglomerate, with a shallow marine intercalation in the early Late Oligocene, all overlain by a shallow marine Neogene sequence (Hekel, 1972). Dating of the non-marine sequence is not sufficiently precise to show if it contains the Late Eocene and Early Oligocene hiatus that has been found at all the deep sea sites, and is interpreted as indicating a change in oceanic circulation.

**Igneous History** No Permian and younger igneous rocks are known in the Eromanga Basin and underlying basins. Parts of the Jurassic and Cretaceous Great Artesian Basin sequence contain fresh andesitic volcanic detritus, interpreted as indicating contemporaneous andesitic volcanism (Senior, 1971). In addition, this sequence in the Surat Basin contains tuffs which have broken down to bentonite (Exon, 1971). The Bowen Basin and basins eastward contain abundant Permian and younger volcanics.

The isotopic ages of Permian and younger plutonic and metamorphic rocks of Australia and New Zealand are shown in Figure 40.18. The Australian dates fall into a Carboniferous to Early Triassic group and, reinforced by the New Zealand dates, into an Early Cretaceous group (see also Kemezys, 1971).

**North Sector**

The North Sector (Fig. 40.17) comprises on its eastern side the Laura and Carpentaria Basins, and South New Guinea; and on its western side, facing the eastern end of the Indonesian arcs, the Arafura Basin, onshore Bonaparte Gulf Basin and its offshore parts (known from the exploration wells Heron-1, Petrel-1, Ashmore-1, and Sahul Shoals-1), and the island of Timor, which we regard as Australia's northwest margin.

Two events are registered throughout the sector: the Aptian marine transgression, seen also in all the other sectors, and an Early Palaeocene hiatus, seen also in the West Sector.

**Sedimentary History** In the eastern part, the Permian is represented by coal measures in the Laura Basin. Otherwise the pre-Jurassic consists largely of Permian and Triassic granites, which intrude the Devonian-Carboniferous Hodgkinson Basin that underlies part of the Laura Basin, and Cambrian and Precambrian elsewhere. The Laura and Carpentaria Basins contain part of the Great Artesian Basin sequence of Jurassic and Neocomian non-marine to paralic sediments succeeded by Aptian and Albian, marine sediments. The South New Guinea sequence reflects outward growth of the continental margin by the deposition of prograding sedimentary wedges, with an extensive marine transgression in the Aptian and less extensive ones in the Eocene and Miocene.

In the western part of the sector, the Late Carboniferous-Permian and younger sediments rest on Early Carboniferous and older Palaeozoic rocks in the onshore Bonaparte
Fig. 40.17. See caption to Fig. 40.14.
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Carboniferous |  Permian | Triassic Jurassic Cretaceous

Fig. 40.18. Frequency histogram of isotopic dates of plutonic igneous and metamorphic rocks of eastern Australia (clear) and New Zealand (oblique lines). Time scale from Harland et al. (1964, 1971). The dates were determined by the potassium-argon method, supplemented by the rubidium-strontium method. Sources: (a) Australia: Cooper et al. (1963), Evernden and Richards (1962), Richards et al. (1966), Webb et al. (1963), Webb and McDougall (1964, 1968). (b) New Zealand: Hulston and McCabe (1972), Landis and Coombs (1967), Wellman and Cooper (1971).

Gulf Basin, and on Silurian in the Arafura Basin (Balke et al., 1973). The Late Carboniferous-Sakmarian sediments of the onshore Bonaparte Gulf Basin are glacigene, and the subsequent Permian and Early Triassic sediments are non-marine to marine. The oldest dated rocks on Timor are Permian paralic sediments, apparently without any glacigene influence. Marine Triassic sediments are recorded in Ashmore-1 and Sahul Shoals-1, and on Timor where they extend upwards through the Early and Middle Jurassic to non-marine sediments in the earliest Late Jurassic. In Heron-1 and Petrel-1, a marine transgression probably commencing in the Late Jurassic has persisted since that time. A marine transgression in the Aptian covered the Arafura Basin and Timor, and except for the notable break in the Early Palaeocene, persisted to the uplift of Timor in the Late Pliocene.

Igneous History Besides the Late Carboniferous, Permian and Triassic granites and Late Carboniferous volcanics, the known igneous rocks of the eastern part of the sector are restricted to the Neogene. Page and McDougall (1970a, 1970b) found a peak of volcanism between 15 and 12 m.y. ago within a general range of 18 to 7 m.y. ago.

Except for the Late Jurassic basalt of Ashmore-1, the known autochthonous igneous rocks of the western part of the sector are confined to Timor. These consist of Permian basalt, Early Eocene lava and tuff, Oligocene acid to basic tuffs with minor basalt, and Late Miocene pumiceous tuff.

Structural History Subsidence of fault-bounded platforms started in the Early Permian of the onshore Bonaparte Gulf Basin, and persisted as the characteristic structural style of the region to the end of the Cretaceous. In the Palaeocene, Timor was the site of overthrusting from the north, and South New Guinea was uplifted and eroded. In the Early Eocene, a slice of oceanic mantle and crust was thrust over the southeast New Guinea margin. The next major events were uplift of South New Guinea in the Oligocene, climactic diastrophism in both Timor and New Guinea in the Middle Miocene, and folding and uplift of Timor and uplift of much of New Guinea in the Pliocene with concomitant sinking of the Timor Trough.

CONCLUSIONS

Figure 40.19 summarises and interprets the geological events of the four sectors of Australia. Though on independent time-tables, the West and South Sectors follow a similar sequence of rifting, rupture, and adjustment, whereas the East Sector reflects convergent plate activity. The correspondence of a peak in plutonic activity in eastern Australia and New Zealand with the Australia- (but not world-) wide Aptian transgression is attributed to the Haug effect (Veevers and Evans, 1973), which states that 'times of orogeny are times of transgression of epicontinental seas on the continental interiors' (Johnson, 1971, 1972). This Aptian transgression is the most obvious geological event in the later history of Australia. Other events common to two or more sectors are: (1) the Late Carboniferous-earliest Sakmarian transgression due possibly to glacio-eustasy (McGowran, 1973); (2) the Early Palaeocene hiatus in the west and north—attributed to a change in oceanic circulation in the west, to collision in Timor, and to uplift causing regression in Papua (Veevers and Evans, 1973); and (3) the Late Eocene (except in the east) and Early Miocene...
Fig. 40.19. Summarised observations and interpretations of principal geological events in the four sectors. Same time scale and symbols as in Figure 40.14-17, except additionally: rifting (open triangles) and inception of sea-floor spreading (— — ). Closely broken line in West Sector refers to southwest part, and full line to northwest part. Closely broken line in North Sector refers to northeast part, and full line to northwest part.
transgressions, which may reflect eustatic changes in sea level.

Until the Late Jurassic to Early Cretaceous, Australia and New Zealand constituted a 'northeastern' part of Gondwanaland. They were bounded on the east by an orogen on the landward side of a convergent plate boundary. Sediment deposited over Australia during the Permian and Mesozoic was derived from this orogen as well as from the interior of Australia and East Antarctica, and was deposited in basins of interior drainage. Only during short periods of general subsidence of the Australian block, due to eustatic sea level rise or to a low stand of the continent due to the Haug effect, did the sea transgress the land. It is this dominant pattern of non-marine deposition in temporary basins of internal drainage that gives the Permian and Mesozoic sequences of Australia and contemporaneous sequences elsewhere in Gondwanaland, due probably to a circum-Gondwanaland orogen, their Gondwanan cast. Secondary features are the periglacial environment of the Late Carboniferous and Early Permian, and the thick sediment fills along lines of subsequent inter-continental rupture that were developed along what later became the western and southern margins of the continent.

Rupture of the Australian part of Gondwanaland by sea-floor spreading from 'India' in the Late Jurassic and Early Cretaceous, and from Antarctica in the Palaeocene, and by the formation of marginal seas from the Lord Howe Rise and New Zealand in the Cretaceous, led to its modern form, which was completed by the opening of the Coral Sea in the Early Eocene, and the collision of its northern margin with volcanic arcs during the Cainozoic.

The ancient orogen of the Great Dividing Range along the eastern Australian margin, and the development through collision of a more recent orogen along the Australian northern margin, extend the history of this pattern of internal drainage to the present. The asymmetrical drainage on the east and an arid climate in the west have led to only slight seaward progradation of these margins. Thus, as during much of Gondwana times, so now, though in a changed configuration, Australia has an asymmetrical drainage pattern (Sprigg, 1961), and the Cainozoic deposits of this drainage in the foredeep between Australia and New Guinea on the one side and in the Diamantina River, Cooper Creek, and Lake Eyre on the other may provide an actuogeological example of the Gondwana Series (Brown et al., 1968: 302).

Sources of data for the West Sector.
General: Balme (1969); Brown et al. (1968); McWhae et al. (1958); Veevers (1971).
Canning: Challinor (1970); Harding (1967); Veevers and Wells (1961); Wellman (1973).
Dampier: B.O.C. of Australia Ltd (in press c.) Enderby-1; Kaye et al. (1972); Veevers et al. (in press).
Carnarvon: Condon (1965-8); Ocean Ventures (1972) Edel-1.
Perth: Johnstone et al. (1973); Jones and Pearson (1972).
Deep Sea: Heirtzler et al. (1973); Luyendyk et al. (1973); Veevers et al. (1973); Veevers et al. (in press a, b).
Sources of data for the South Sector.
Eucla: Cockbain (1968); Lowry (1970); Ludbrook (1958); Parkin (1969); Playford and Cope (1971).
Otway: Bock and Glenie (1965); Glenie et al. (1968); Harding (1967); Parkin (1969); Singleton (1967); Wopfner and Douglas (1971).
Bass and Gippsland: Elliott (1972); Hocking (1972); James and Evans (1971); Richards and Hopkins (1969).
Tasmania: McDougall (1961); Spry and Banks (1962).
Sources of data for the East Sector.
Arckaringa: Parkin (1969); Wopfner et al. (1970).
Maryborough: Ellis (1968); Hekel (1972).
Gympie: Runnegar and Ferguson (1969); Andrews et al. (1973).
Queensland Plateau: Burns et al. (1972).
Coral Sea, Tasman Sea: Kennett et al. (1973).
Igneous rocks: See the references in caption of Figure 18, and additionally Dulhunty and McDougall (1966); McDougall and Wilkinson (1967); Webb et al. (1967); Wellman et al. (1969, 1970); and Wyatt and Webb (1970).
Sources of data for the North Sector.
Laura: de Keyser and Lucas (1968).
South New Guinea: Davies and Smith (1971); Page and McDougall (1970a, 1970b); Rickwood (1968); Visser and Hermès (1962).
NT Shelf: Skwarko (1966).
Petrel-1: Arco Ltd (in press a).
Laura: de Keyser and Lucas (1968).
Petrel-1: Arco Ltd (in press b).
Timor Trough: Veevers et al. (1973); Veevers et al. (in press b).

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Heitzler, J. R., Bolli, H. M., Carter, A. N., Cook, P. J., Krasheninnikov, V. A., McKnight, B. K., Proto-Decima, F., Renz, G. W., Robinson, P.


ABSTRACT

The Pacific border of Gondwanaland, along the edges of the Brazilian, African, East Antarctic, and Australia-New Zealand shields, was an active continental margin from Late Precambrian into Late Cretaceous time. Four major orogenic episodes (early Palaeozoic, middle Palaeozoic, Permo-Triassic, and Cretaceous) are defined by folding and faulting, igneous activity, and regional metamorphism. These episodes formed orogenic belts which can be traced intermittently along this restored continental margin. These belts have indistinct temporal and spatial boundaries, but they can be viewed as records of deformational peaks during a long interval of tectonic activity. In some places the sedimentary-volcanic sequences accumulated on a foundation of continental rocks, in other places the basement may have been oceanic crust. In general the orogens are younger toward the Pacific, but the existence of older igneous-metamorphic terrains seaward of some orogens indicates a more complex history than simple accretion of successive belts to the continental margin.

Three of these orogenic belts formed before the main fragmentation of Gondwanaland and are unrelated to the break-up. These early belts provide important clues for the proper reconstruction of Gondwanaland. The fourth belt developed during late Mesozoic and early Cainozoic time and is contemporaneous with a major fragmentation phase. Part of this youngest belt remains active today (South America), part appears to have been inactive since the Late Cretaceous except for volcanism (Antarctica), and part underwent a marked change in tectonic pattern between the Cretaceous and the late Cainozoic (New Zealand).

A model for the Mesozoic-Cainozoic fragmentation of Gondwanaland is proposed which may explain (1) these differences in the Cretaceous-Cainozoic history of the several segments of the Pacific margin, and (2) the complex arrangement of tectonic elements in West Antarctica.

INTRODUCTION

Because of rapid advances in the understanding of the geology of the southern lands and oceans, the concept of Gondwanaland has gained widespread if not universal acceptance during the past decade. This ancient landmass is considered to have existed during late Palaeozoic time, probably in close proximity to a northern landmass to form a single enormous protocontinent, Pangaea. Rifting and fragmentation appears to have begun in the early Mesozoic, and this initial break-up was followed by gradual dispersal of the southern lands into their present positions.

Although there is much evidence to support this grand geotectonic model, many details are lacking and important problems re-
main unanswered. Of particular significance and interest are the regions which form the Pacific margin of Gondwanaland, and the Antarctic continent has a special importance because of its central position in this marginal zone. What is the regional tectonic pattern of Antarctica? Do the varied orientations of mountain chains in West Antarctica represent tectonic trends, and if so how and when did they form? Does the modern circum-Pacific mobile belt, so well defined in South America and in the southwest Pacific, continue through West Antarctica? If this belt can be traced from Chile through West Antarctica to New Zealand, why has the Cainozoic behaviour of the several segments of the belt in this region been so dissimilar?

Attracted by these problems and their relevance to the central question of Gondwanaland and its fragmentation, the writer has worked on the geology of certain areas in West Antarctica since 1959. Some preliminary accounts of the areal geology of West Antarctica have been published, along with an interpretation of Antarctic tectonic patterns (Craddock, 1972a). In addition, with splendid international co-operation, a series of 1:1,000,000 geologic maps of areas of extensive rock exposure and 1:10,000,000 maps of the entire continent have been assembled (Bushnell and Craddock, 1970), and the available information on the geology, structure, fossil localities, and radiometric age localities was combined into a geologic map of Antarctica on 1:5,000,000 scale (Craddock, 1972b).

The present essay is an attempt to interpret Antarctic tectonic patterns (Fig. 41.1) in the broader context of the evolving Pacific margin of Gondwanaland. In addition, some ideas are offered about the origin of tectonic trends in West Antarctica and about the Mesozoic-Cainozoic tectonic history of that area.

Fig. 41.1. Tectonic provinces of Antarctica (after Craddock, 1972a)
THE PACIFIC MARGIN OF GONDWANALAND

Many attempts at the reconstruction of Gondwanaland have been published, and they will not be reviewed in detail here. In his masterly synthesis of the evidence for Gondwanaland, Du Toit (1937) offered a reassembly which, in the present writer's opinion, still requires no major modification. On the basis of much more information from Antarctica, the writer suggested a reconstruction (Bushnell and Craddock, 1970) which is quite similar to that of Du Toit. A modified version of this reconstruction is included here as Figure 41.2.

The Pacific margin of Gondwanaland, as it existed prior to fragmentation, consists of segments now located in Australia-New Zealand, Antarctica, the southern tip of Africa, and South America. The distribution of sedimentary and volcanic rocks, granitic intrusives, regionally metamorphosed terrains, and belts of folded rocks, shows that most or all of this Gondwanaland border existed as an active continental margin during much of
Phanerozoic time, at least until the Late Cretaceous or early Tertiary.

The Precambrian shields of the present Gondwanaland fragments are bounded by one or more Phanerozoic orogens. These orogens have indistinct temporal and spatial boundaries, but a general pattern of younging toward the Pacific can be recognized. An orogen commonly reveals an early history of marine sedimentation and volcanic activity, one or more deformational episodes, regional metamorphism, and granite emplacement; and a later history of stabilization, cratonization, and non-marine sedimentation, in some cases accompanied by volcanic activity. These patterns are well displayed in Australia-New Zealand, and they can also be shown in Antarctica and South America although the documentation is less complete.

Within this broad pattern, however, there remain many uncertainties and unsolved problems. What was the nature of the crust upon which these orogens developed? The description of ophiolites from Australia-New Zealand and South America suggests that at least some of these orogens formed on true oceanic crust.

What were the sources and dispersal patterns for the clastic sediments in each orogen? Were borderlands present? If so, of what kinds of rocks were they composed? The existence of older igneous-metamorphic terrains seaward of some orogens indicates a more complex history than the simple accretion of successive oceanic belts to the continental margin.

**PRECAMBRIAN SHIELDS**

The distribution of the Precambrian shields that formed the nucleus of Gondwanaland is shown in Figure 41.2; those which are important in this discussion are the Australian, East Antarctic, African, and Brazilian Shields. The geology of these Gondwana platforms has been reviewed by Ravich (1968) and available radiometric ages were compiled and discussed by Hurley and Rand (1969). Structural, lithologic, and age provinces recognized on opposing coasts are compatible with the proposed reconstruction, but Precambrian exposures are limited in some coastal areas, especially in Antarctica.

A boundary between the first Phanerozoic orogen and the older rock units of the shield can be drawn readily in most places, but a widespread early Palaeozoic thermal event that impressed apparent ages of about 500 m.y. on many older shield rocks makes it difficult to draw this line in some areas, particularly where outcrops are scarce.

South America poses a special problem of great importance in reassembling Gondwanaland and in understanding its development and eventual fragmentation. Many age determinations show that the greater Brazilian Shield is undoubtedly Precambrian in age, and that it extends southward at least into northern Argentina. In southern Argentina basement rocks outcrop in the Patagonian and Deseado massifs where the overlying rocks are mainly Mesozoic and Cainozoic in age. These isolated basement rocks may represent a southward continuation of the Brazilian Shield, but radiometric ages reported by Halpern (1969) and Stipanicic et al. (1971) suggest these rocks could be younger than Precambrian.

**EARLY PALAEZOIC OROGENS**

In late Precambrian and Cambrian time the Pacific border of Gondwanaland received thick deposits of sedimentary and volcanic rocks, and these geosynclinal strata—at least along most of the border—were deformed, metamorphosed, and intruded by granitic plutons during the early Palaeozoic. This orogenic event began in the Cambrian, and appears to have culminated in the Ordovician. Evidence for this early Palaeozoic orogeny is well displayed in Australia and Antarctica, and the orogen is shown in those continents in Figure 41.2. Although some evidence for this event is reported from southern Africa and South America, geologic relationships there are still too obscure to allow portrayal of its areal extent on a map.

In South Australia a thick sedimentary sequence was deposited in the Adelaide Geosyncline during late Precambrian and Cambrian time. Many of these sediments are of shallow-water marine origin, and in the limestone beds oolites, algae, and archaeocyathids are common (Brown, Campbell, and Crook,
1968). These rocks were folded, metamorphosed, and intruded by granites between the Middle Cambrian and Middle Ordovician (Geological Society of Australia, 1971). The correlation of these rocks and this orogeny in Australia with those in Antarctica has been discussed recently by Oliver (1972).

The Ross Orogen of early Palaeozoic age can be traced almost across the Antarctic continent, roughly parallel to the Transantarctic Mountains (Bushnell and Craddock, 1970). The presence of some rocks in this belt with apparent ages greater than 1000 m.y. suggests that this geosyncline formed, at least in part, on the subsiding edge of the East Antarctic shield. The geology of this orogen is discussed by Craddock (1972a) and is shown on the Geologic Map of Antarctica (Craddock, 1972b).

The extent of the early Palaeozoic orogen in southern Africa is difficult to establish because more than half of the Republic of South Africa is underlain by younger strata of the Cape and Karroo Systems (Truswell, 1970). Because of the limited exposures and the widespread development of a younger orogen, no attempt has been made to plot the early Palaeozoic orogen in Africa in Figure 41.2. Nevertheless, in the region around Cape Town, and within inliers in the Cape Fold Belt to the east, strongly folded strata of the Malmesbury and Cango Formations outcrop. These beds are intruded by the Cape granites of about 550 m.y. age (Coertze, Schifano, and Van Eeden, 1970), and they testify to the existence of an early Palaeozoic, pre-Cape System orogeny at the southern end of Africa.

The tectonic history of southern South America is poorly known, but the early Palaeozoic orogeny may have reached this part of Gondwanaland as well. Borrello (1969) recognised four main geotectonic cycles in Argentina, and his second cycle ended with the emplacement of upper Precambrian-Ordovician granitic plutons. Mon (1972) reported Cambro-Ordovician sedimentary rocks and lower Palaeozoic granites (probably Lower Ordovician) in northwestern Argentina.

MIDDLE PALAEozoIC OROGENS

Evidence for middle Palaeozoic orogeny along the Pacific border of Gondwanaland can be found from Australia to South America. The known geologic record of this event is extensive in Australia and New Zealand, more modest in Antarctica and South America, and quite limited in southern Africa. Manifestations of this orogeny began in the Silurian and continued into the Carboniferous, but it seems to have reached a peak during the Devonian in most places.

In southeastern Australia there was widespread deformation, metamorphism, and granitic intrusion during the middle Palaeozoic in the western Tasman Geosyncline (Geological Society of Australia, 1971). This activity has been well documented by Packham (1969) in New South Wales, where four orogenic pulses occurred between the Early Silurian and the Early Carboniferous. In addition, regional metamorphism and the emplacement of granitic plutons in western New Zealand took place during the Devonian (Suggate and Grindley, 1972a).

Evidence for this cycle occurs in two regions in Antarctica, as discussed by Craddock (1972a). A triangular area in northern Victoria Land, at the Australian end of the Transantarctic Mountains, contains strongly folded metamorphic rocks cut by Devonian-Carboniferous intrusives and overlain by undisturbed strata probably no older than Triassic. The coastal belt of West Antarctica, moreover, reveals a few exposures of igneous and metamorphic basement rocks that may have formed during the middle Palaeozoic orogeny (Bushnell and Craddock, 1970; Craddock, 1972b).

Most of the deformation that affects the Cape and Karroo Systems in the Cape Fold Belt of southern Africa appears to have taken place during Permo-Triassic time. However, Haughton (1969) suggested that the Cedarberg folds in the Cape System are pre-Karroo System, and he described one locality where apparently undisturbed Dwyka tillite rests upon folded Witteberg strata. These relationships seem to indicate that a mild deformation affected the Cape System in some localities before the deposition of the Karroo System began in the Carboniferous.
The ages of intrusive and metamorphic rocks in the southern Andes are largely unknown, but some evidence for a middle Palaeozoic orogeny has been reported. Harrington (1962) inferred a geosynclinal axis in the Precordillera during the early Palaeozoic, and these rocks were deformed during the 'Acadian' orogeny in the middle Palaeozoic. The tightly folded and metamorphosed Cortaderas Formation of northwestern Argentina has been described by Cucchi (1972), who considered the age of the deformation and metamorphism to be Devonian. Caminos (1972) and Munizaga, Aguirre, and Herve (1973) report Carboniferous apparent radiometric ages on respectively, granitic intrusives in northwestern Argentina and metamorphic basement in central Chile.

**PERMO-TRIASSIC OROGENS**

Indications of Permo-Triassic orogeny along the edge of Gondwanaland can be found from Australia to South America. The results of this tectonic cycle are especially clear in Antarctica, Africa, and South America. Du Toit (1937), with great insight for his time, recognised the importance of this event to the Gondwanaland hypothesis and named it the Gondwanide orogeny. Although it is not shown separately on Figure 41.2, the Hunter-Bowen orogeny in the eastern Tasman Geosyncline of Australia may be reasonably correlated with the Gondwanide event. Furthermore, some granitic intrusives in New Zealand have given Permo-Triassic ages (Suggate and Grindley, 1972a), but the evidence for Gondwanide orogeny there is too sparse to be more than suggestive at present.

In Antarctica the Ellsworth orogen can be traced from the northern Ellsworth Mountains southeastward to the Pensacola Mountains. This Permo-Triassic orogen is marked by intensive folding, low-grade regional metamorphism, and a number of granitic intrusives. This fold belt was described by Craddock (1964), and its importance has been emphasised by Craddock (in Bushnell and Craddock, 1970, 1972a) and by Ford (1972).

The Cape Fold Belt is a prominent tectonic feature at the southern tip of Africa, developed in the strata of the Cape and Karroo Systems. The writer's personal observations suggest that strong stratigraphic and structural similarities exist between this belt and the Ellsworth orogen in Antarctica. The exact time of the deformation in the Cape Fold Belt is difficult to establish, but according to Bishopp and Van Eeden (1971) the folding peak was probably reached during the Middle Triassic.

Southern South America provides evidence for Permo-Triassic orogeny in two separate areas. Du Toit (1937) stressed the importance of the northwest-trending folds in rocks as young as Permian in the Southern Hills of Buenos Aires province in northern Argentina, and he considered them the continuation of the Cape Fold Belt. It is interesting to note that Harrington (1970) interprets both these areas as 'folded aulacogenic ranges', rather than as miogeosynclinal belts. This thought is obviously very important for Gondwanaland reconstructions, but it cannot be tested until much more is learned about the geology of the area south of Buenos Aires province.

Permo-Triassic orogeny seems also to have affected the main Andean belt in South America. Harrington (1962) inferred the formation of the principal Cordillera belt during 'Hercynian' deformation, and Gansser (1973) recognises a strong pre-Mesozoic orogeny. Borrello (1969) considers that the third major geotectonic cycle ended during the Triassic with the consolidation of the Cuyan chain. Permo-Triassic apparent ages of granites in northwestern Argentina have been reported by Caminos (1972). The importance of early Mesozoic orogeny in southern South America, the Scotia Arc, and the Antarctic Peninsula has been emphasised by Dalziel (1972).

**CRETACEOUS OROGENY**

Unlike the preceding three orogenies, this event post-dates the beginning of the main fragmentation of Gondwanaland and must be related to the dispersal of the several parts. This geotectonic cycle begins with widespread Jurassic sedimentation and volcanism and continues locally into the Caino-
zoic, but the peak in most areas seems to have come during the Cretaceous. This event is clearly one of the most important in the history of the entire circum-Pacific belt of the present.

This orogeny did not significantly affect Australia proper, where the Tasman Geosyncline had been immobilised, but it was strong in New Zealand. The Rangitata orogeny began in the Jurassic and peaked in the Middle Cretaceous (Fleming, 1970; Suggate, 1972), terminating the New Zealand Geosyncline that had existed since the Carboniferous. The Rangitata orogeny caused widespread folding, regional metamorphism, and emplacement of granitic intrusives in eastern New Zealand.

A similar Cretaceous orogenic belt can be traced along coastal West Antarctica eastward to the Antarctic Peninsula and the Scotia Arc (Craddock, 1972a). In Marie Byrd Land sedimentary and volcanic rocks considered to be probably Carboniferous to Jurassic in age have been folded, regionally metamorphosed, and intruded by granitic plutons of Cretaceous age. Eastward into Ellsworth Land occur scattered exposures of Cretaceous intrusive rocks, mainly granitic in composition. All of these rock units are overlain by undisturbed Tertiary volcanic rocks. The Antarctic Peninsula contains many plutons which yield Cretaceous or early Tertiary apparent ages, but the late deformational history of the Antarctic Peninsula is not clear. In the northern part of the peninsula, stratified rocks as old as Jurassic appear unfolded and only gently tilted. In the southern part of the peninsula, however, Jurassic strata are strongly folded and Cretaceous strata are moderately deformed.

Evidence for this Cretaceous orogeny is not found in southern Africa, where an extensional stress regime at that time led to the creation of a number of large dip-slip normal faults (Bishopp and Van Eeden, 1971).

The west coast of southern South America has undergone a complicated but still poorly understood geologic history from the Jurassic to the present, the fourth and final geotectonic cycle of Borrello (1969). The Andean orogeny is distinguished by the emplacement of numerous batholiths of mainly felsic composition, ranging in age from Jurassic to early Tertiary. Cretaceous strata have been tightly folded in Tierra del Fuego, but northward compressive deformation is modest (Dalziel and Cortes, 1972). Indeed, Gansser (1973) has emphasised the importance of block faulting since the Mesozoic, as well as the lack of folding in the sediments of the Peru-Chile trenches. Many early workers have drawn comparisons between the Andean orogens in South America and the Antarctic Peninsula, and have considered these areas as segments of a once continuous belt which was disrupted in post-Cretaceous time to form the Scotia Arc. Some recent workers (Dalziel and Cortes, 1972; Katz, 1972), however, have emphasised important differences in the recent tectonic histories of these two areas.

**The Fragmentation of Gondwanaland**

Gondwanaland may well have existed intact throughout the Palaeozoic, probably in close proximity to Laurasia to form a single supercontinent. From the late Precambrian until the time of fragmentation, the Pacific border of Gondwanaland has functioned nearly continuously as an active continental margin. Dietz and Holden (1970) have offered an interpretation of the general history of break-up for Pangaea.

On geologic grounds it seems probable that Gondwanaland remained essentially intact well into Triassic time. However, the relative movements of the individual fragments between the Triassic and the recent is poorly understood. New information becoming available from the lands and the oceans is helping to clarify the situation, but much remains to be learned.

The different tectonic behaviour of different segments of the circum-Pacific belt since the Cretaceous, and the peculiar structural trends in the provinces of West Antarctica have proved especially puzzling. This chapter concludes with a few speculations on those problems.

Comparison of the geologic history of Australia and New Zealand suggests that these two land masses were once contiguous as part of Gondwanaland. The time of separation is unknown, but the presence of Cretaceous
strata may indicate the first opening of the Tasman Sea (Fleming, 1970). Indeed, it seems reasonable to relate the Rangitata orogeny to the opening of a marginal basin, with attendant eastward migration and deformation of New Zealand. Upon termination of the Rangitata orogeny in the Late Cretaceous, New Zealand enjoyed relative tectonic stability until the Kaikoura orogeny commenced in the Miocene.

The geologic pattern in Antarctica bears many similarities to that of New Zealand-Australia. Coastal West Antarctica is analogous to New Zealand, and East Antarctica to Australia. Accordingly, it also seems probable that all of coastal West Antarctica, including the Antarctic Peninsula, was originally closer to the edge of East Antarctica than at present. Cretaceous deformation in West Antarctica may also be related to seaward movement of the coastal belt, and the opening of a narrow marginal basin behind it. The widespread Jurassic mafic igneous rocks in Tasmania and the Transantarctic Mountains may be related to crustal extension associated with this basin opening. Much of the belt between the Ross and Weddell Seas has a bedrock surface at or below sea level, even after isostatic correction; because of the thick ice cover, the bedrock geology of this morphologically low zone is unknown. It is of interest here, however, to note the recent discovery of marine Cretaceous fossils in Quaternary deposits along the edge of the Ross Sea (Webb and Neall, 1972).

Thus in the Late Cretaceous East Antarctica and Australia remained connected, but they were probably separated by a marginal basin from West Antarctica and New Zealand, which also remained connected. However, about 81 m.y. ago (Christoffel and Falconer, 1972) New Zealand and West Antarctica began to separate, but the spreading ridge did not penetrate between Australia and East Antarctica at that time. Differential movement between East and West Antarctica at this time is likely to have occurred, probably along a transform fault. This movement of West Antarctica away from the new spreading ridge transverse to its grain may have caused (1) the bending of the Ellsworth orogen outward from the edge of East Antarctica, (2) the bending and extensive fracturing of the Andean orogen in West Antarctica, and (3) the intensive deformation of Mesozoic strata at the base of the Antarctic Peninsula.

About 55 m.y. ago (Weissel and Hayes, 1972) the spreading ridge penetrated between East Antarctica and Australia, terminating the differential movement between East and West Antarctica. Antarctica became aseismic at that time and developed stable margins. Moderate volcanism continues even today in West Antarctica, probably owing to extensive fracturing during oroclinal bending in the early Tertiary. The interesting results of LeMasurier (1972) show the possibility of significant vertical movements of crustal blocks during the Cainozoic, along with the volcanic activity.

The Tertiary tectonic peace of New Zealand was ended during the Miocene by the Kaikoura orogeny. This event may record the inception of the Alpine fault and the imposition of the modern strike-slip tectonic regime. To date, however, there is no evidence for such a structural feature active at this time in West Antarctica.

Thus during the late Cainozoic three quite dissimilar segments have developed in the circum-Pacific rim facing the Southern Ocean. Western South America, at least north of the Chile Ridge, is an active continental margin marked by abundant volcanism and numerous earthquakes along a deep dip-slip zone. The Antarctic continent has developed mainly passive margins, except perhaps for a transform fault boundary in the Scotia Arc; the Pacific border has no earthquakes and very slight volcanism. New Zealand is an active zone of many earthquakes and moderate volcanism, caused by the development of a major strike-slip fault between the Indian and Pacific plates.

Although this model suggests solutions to some problems, it does leave two important questions unanswered. If the Antarctic Peninsula has migrated toward the Pacific Ocean to reach its present isolated position, why are strata as old as Jurassic unfolded and only gently tilted? And why does there remain such a large gap in the Gondwana-
land reassembly between the Weddell Sea coast and the southeastern tip of Africa? Perhaps the answer to the latter question lies in (1) moving Madagascar further southward, as suggested by Flores (1970); or (2) moving the Antarctic Peninsula toward Africa and East Antarctica, as the present model suggests is probable; or (3) moving part of southern South America into the gap, especially if Precambrian rocks are eventually shown to be lacking in southern Argentina.

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REFERENCES


New Zealand and the Southwest Pacific Margin of Gondwanaland

J. R. GRIFFITHS

ABSTRACT

Reassembly of continental fragments in the southwest Pacific region establishes New Zealand as an integral part of Gondwanaland. Major elements of the Gondwana geology of New Zealand are reviewed and palaeogeographic maps constructed. These data are incorporated in a regional tectonic synthesis of the Gondwana margin, including eastern Australia and New Caledonia.

An episode of subduction occurred during the Late Carboniferous and Early Permian. Associated post-subduction granites and volcanics occur in the Permo-Triassic of eastern Australia. Following subduction a thick, predominantly clastic, sedimentary pile accumulated along the margin from mid-Permian to mid-Jurassic. These sediments show an easterly transition from continental shelf to slope environments, typical of aseismic continental margins. The Gondwana basins of eastern Australia developed on a relatively stable continental platform with extensive river and delta systems feeding the marginal ‘geosyncline’ to the east. A new episode of subduction began in the mid-Jurassic, and the thick sedimentary pile rapidly underwent polyphase deformation and metamorphism during the Rangitata Orogeny. Post-orogenic flysch-type sediments were subsequently deposited flanking the mountain belt, and indicate that subduction ceased in the Early Cretaceous, well before the end of related volcanism. The beginning of fragmentation of the Gondwana continent in the region was marked by the development of a system of rift valleys in the Late Jurassic and Cretaceous, in which thick clastic sediments accumulated.

INTRODUCTION

Although New Zealand has generally received little mention in the Gondwana context, except to be placed closer to the east of Australia on reconstructions, it must be considered as a part of Gondwanaland and has some of the thickest Permian to Cretaceous sequences known. The aim of this paper is to review briefly the relevant geology and palaeogeography of New Zealand, and then to consider the broader tectonic history of the Gondwanaland margin bordering the southwest Pacific Ocean, including New Caledonia and parts of Australia and Antarctica.

The basis for this discussion is a reconstruction of the continental fragments of the region, based on morphological, geophysical and geological considerations (Figs. 42.1, 42.3a). In this reassembly, Australia and Antarctica are fitted together by matching the 2000 m Antarctic isobath with the 3000 m Mesozoic/Cainozoic sediment isopach for the southeastern margin of Australia. New Zealand is then incorporated by closure of the Tasman Sea, bringing the Lord Howe
Fig. 42.1. Reconstruction of the southwest Pacific margin of Gondwanaland. Five degree latitudes and longitudes indicated. (Modified from Griffiths, 1971a: fig. 3.)

Rise and Campbell Plateau against Australia/Antarctica. During this operation, the Alpine Fault and associated Cainozoic deformation in New Zealand is 'reversed', so
Fig. 42.2 Major geological elements of New Zealand (see text)

establishing the pre-drift outline shown in Figure 42.1.

The oceanic crust between Australia/Antarctica and New Zealand is all less than 85 m.y. old (Weissel and Hayes, 1971; Hayes and Ringis, 1973; Falconer, in press), and hence this reconstruction is adopted as the basis for subsequent discussion of the Gondwana (Carboniferous to Cretaceous) geology in this paper.
GONDWANA PALAEOGEOGRAPHY OF NEW ZEALAND

The geology of New Zealand is described in detail in maps and publications of the New Zealand Geological Survey. Many of the data used in this paper to construct the palaeogeographic maps are taken from the 1:250,000 geological map series (N.Z. Geological Survey, 1959-68). Other more recent or comprehensive references also are given below.

The present structural framework of New Zealand is shown in Figure 42.2. The major units recognised are:

1. The Precambrian and Palaeozoic 'foreland' in South Island, comprising various schists, gneisses, intrusives and sediments.
2. The Carboniferous and Permian Te Anau Group (not shown) which outcrops east of the foreland and under the Hokonui Facies in South Island (Grindley, 1958).
3. The 'New Zealand Geosyncline' which includes the Carboniferous to Jurassic sediments of the Hokonui and Torlesse Facies.
   - The Hokonui Facies (Wellman, 1952; Fleming, 1970) consist of predominantly shallow-water clastic sediments, with volcanic detritus; the Torlesse Facies (Suggate, 1961; Fleming, 1970) is a sequence of argillites and 'greywackes', in part now metamorphosed to low grade schists. The boundary between these two units is gradational, and probably fluctuated markedly during their deposition.
4. The East Coast/Marlborough Basin, a moderately thick sequence of Cretaceous sediments (Wellman, 1959), which passes upwards into the Tertiary and westwards. Several other scattered occurrences (not shown) of Cretaceous sediments also occur.
6. The Taupo Volcanic Zone, a rift filled with Quaternary volcanics (Grindley, 1960; Healy et al., 1964).
7. The Alpine Fault, which is a Late Cainozoic dextral transcurrent fault with a recorded offset of 480 km (Wellman in Benson, 1952; Wellman, 1956).

Using the continental reconstruction (Fig. 42.1) as a basis, the Gondwana geology of New Zealand is plotted on four palaeogeographic maps (Figs. 42.3-6) which are discussed below.

The New Zealand Geosyncline was established along the Gondwana margin flanking the Pacific Ocean by Permian times (Fig. 42.3). The depositional history of the 'geosyncline' is considered in terms of two units. The Hokonui Facies is a fossiliferous sequence of shallow water clastics, derived from the west, and including fairly abundant volcanic detritus believed to reflect volcanic activity on the western foreland (Wellman, 1952; Fleming, 1970; Dickinson, 1971). The sequence is best exposed in the Southland Syncline, where it thickens eastwards (cf. Mutch, 1957: fig. 2). The Hokonui Facies was probably deposited largely on older continental crust, including the Te Anau Group. In contrast, the much thicker Torlesse Facies is a monotonous sequence of graded 'greywackes' and argillites, probably deposited mainly on oceanic crust (Landis and Coombs, 1967; Fleming, 1970; Landis and Bishop, 1972). It seems most likely that a western landmass was the source of the clastic material in the New Zealand Geosyncline. However, it has also been suggested, largely on the basis of sedimentological studies in the Torlesse Facies, that a substantial continental mass lay to the east (e.g. Bradshaw and Andrews, 1973). The existence of such a mass cannot be entirely discounted, but reconstructions of Gondwana-land (e.g. Smith and Hallam, 1970) appear to eliminate the possibility of the other Gondwana fragments being directly east of New Zealand, implying that any postulated mass has now disappeared without trace. The problem needs further study, particularly with regard to the effects of Cainozoic deformation, but at present there is no unequivocal evidence for a landmass east of New Zealand.

By the Late Jurassic (Fig. 42.4) the first signs of orogeny are evident, with uplift and progressive cessation of deposition (Wellman, 1956; Fleming, 1970) followed by a
Fig. 42.3 to 6. Palaeogeographic reconstructions of New Zealand during Gondwana times. Geological data largely from 1:250,000 Series maps (NZ Geological Survey, 1959-68).
more widespread marine regression. Griffiths (1973a, 1973b) has interpreted these movements as indicating the onset of subduction along the Pacific margin (see also Figs. 42.11, 42.12). A major orogenic event (the Rangitata Orogeny) occurred in the
Early Cretaceous. Polyphase deformation and metamorphism characterise the orogenic belt. Miyashiro (1961) and Landis and Coombs (1967) recognise contrasting high T/P and high P/T metamorphic facies on the foreland and ocean sides of the orogen. Landis and Bishop (1972) have recently reviewed the sedimentary, structural and metamorphic aspects of this orogen in more detail. As deformation progressed, erosion of uplifted areas provided coarse conglomerates and sandstones redeposited on the flanks.
of the orogen (Fig. 42.5). Along the eastern side, the Marlborough/East Coast Basin contains sediments derived from both west and east (Kingma, 1965, 1967), the latter inter-
interpreted here as indicating the presence of uplifted 'islands' above the leading edge of the Gondwana plate. The oldest sediments of this sequence are the Taitai Series (Lower Cretaceous), largely a mixture of reworked Torlesse Facies material and igneous debris (Wellman, 1959; Kingman, 1965). Similar sediments are also preserved in other localities (see Fig. 42.5). The Upper Cretaceous Clarence, Mata and Raukumara Series (Fig. 42.6) begin with similar conglomerates and sandstones, but pass upwards generally into finer sands and silts, which also transgress westwards indicating progressive erosion of the Rangitata Mountains. Small coalfields developed at Greymouth (Gage, 1952) and Kaitangata (Harrington, 1958) and elsewhere during the Cretaceous.

New Zealand Gondwana geology is compiled in Figure 42.7 as a time-space cross-section, which shows the relative positions of various sedimentary sequences and structural, metamorphic and igneous events. Cainozoic deformation has exposed varying levels of New Zealand (see inset, Fig. 42.7), and thus a fairly complete geological history can be reconstructed. The time-space plot should be read in conjunction with the palaeogeographic maps. It illustrates the general eastward accretion of sediment, followed by the Rangitata Orogeny, further erosion and deposition, and transgression of the sea across New Zealand in the Early Cainozoic.

A SYNTHESIS OF THE REGIONAL TECTONIC EVOLUTION

The geology of New Zealand, as outlined above, provides an important addition to discussions of Gondwanaland, as it gives a complete section across the continental margin from the Palaeozoic foreland to the Pacific Ocean. These data can now be integrated with those available from neighbouring parts of Gondwanaland, as a basis for palaeogeographic and palaeotectonic reconstructions. In this study, data for Australia is taken largely from Brown et al. (1968); for Antarctica from Bushnell and Craddock (1970); and for New Caledonia from Lillie and Brothers (1970). Other papers in this volume update these references (see Veevers and Evans (Australia); Barrett and Kohn; Craddock; Elliot; Grikurov and Lopatin (Antarctica)).

A more detailed version of the reconstruction, with Palaeozoic elements superimposed, is shown in Figure 42.8. Solomon and Griffiths (1972, in press) have presented a tectonic synthesis of the Tasman Orogenic Zone, and argue in favour of one or more Precambrian 'islands' which were incorporated within this zone during the Palaeozoic. Tentative limits of these blocks are now indicated. It is also suggested here that correlation of the Heathcote Axis (Cambrian 'greenstones') of Victoria, the Dundas Trough (Cambrian sediments, ophiolites and calcalkaline volcanics) of Tasmania, and the Bowers Group (Cambrian to Early Ordovician sediments) of Antarctica is possible (Fig. 42.8). This is speculative, but warrants detailed evaluation as it may offer a fairly precise geological match. During a complex Palaeozoic history of accretion, subduction and collision, the Tasman Orogenic Zone was gradually stabilised. By the Permian, the edge of Gondwanaland was well to the east of Australia, with thick sequences accumulating in the New Zealand Geosyncline and its extension northwards through New Caledonia, and probably southwards into West Antarctica.

The Te Anau Group of New Zealand is interpreted here as an Andean-type volcanic arc, related to an episode of subduction of the Pacific plate during the Late Carboniferous and Early Permian (Fig. 42.9). Lithologies in the Te Anau Group include volcanics (spilites, keratophyres, gabbroic intrusives, etc.), ophiolites (norites, dunites, peridotites, pyroxenites, serpentinites, etc.) and associated sediments (argillites, sandstones, spilitic tuffs, volcanic breccias, etc.) (Grindley, 1958). Challis (1968) indicates that the K₂O content increases from east to west across the Te Anau Group, a similar trend to those across other volcanic arcs related to subduction (e.g. Hatherton and Dickinson, 1969). The main zone of volcanism would have extended northwards to the east of Australia, but may have passed into West Antarctica. Griffiths (1973a) has argued that calcalkaline volcanism at a con-
continental margin may be initiated by subduction, but post-dates cessation of subduction by up to 50 m.y. or more. Thus it is suggested that the volcaniclastic fraction in the Hoko-nui Facies (and equivalents in New Caledonia) in the Permo-Triassic may be derived from continuing post-subduction volcanism. The Permo-Triassic granites and volcanics in New England (eastern Australia) are also interpreted as post-subduction ig-
neous rocks, in part derived from a possible Precambrian basement (Figs. 42.8-10). Elsewhere during the Late Palaeozoic, quiet platform conditions prevailed, with an extensive Late Carboniferous/Early Permian glaciation of Gondwanaland (see Crowell and Frakes, this volume). Probable evidence of this glaciation also occurs in New Zealand (cf. Brown et al., 1968: 212).

Following cessation of subduction in the Early Permian, the New Zealand Geosyncline was established as sediments were deposited along the Pacific margin (Fig. 42.10). The Hokonui and Torlesse Facies of New Zealand, and the Permo-Jurassic of New Caledonia, show eastward transition from continent to ocean, and indicate the meridional continuity of the continental margin. To the west, extensive river and delta systems provide a provenance for the marginal accumulations. These systems developed across the older Palaeozoic belt, and are represented in the eastern Australian basins which have coal measures and typical Gondwana faunas and floras. Epeirogenic movements, related to subduction in the east, may have led to local 'collapse' of the basement, with associated volcanism, to form the deeper intra-cratonic basins with basal volcanic facies (e.g. Sydney and Bowen Basins). Granite intrusion and acid volcanism continued in New England, and hence probably along a broad belt extending southwards to the west of New Zealand.

Orogenic activity commenced again in the mid-Jurassic in both New Zealand and New Caledonia. Griffiths (1973a; 1973b) has presented a more detailed model for the Rangitata Orogeny, in which a short (less than 50 m.y.) episode of subduction of the Pacific plate was accompanied by polyphase deformation and metamorphism, and largely postdated by granite intrusions and calcalkaline volcanism. Regional palaeotectonic interpretations of these events are shown in Figures 42.11 and 42.12. The detailed sequence of events in New Zealand is discussed elsewhere (e.g. Landis and Coombs, 1967; Fleming, 1970; Landis and Bishop, 1972; Griffiths, 1973a; 1973b). Granites and calcalkaline volcanics occur in New Zealand, and also on the coast of Queensland, where the postulated subduction zone comes closest to Australia (Fig. 42.11). Erosion of the marginal mountain belt occurred rapidly after uplift, providing the previously mentioned Cretaceous sequences of New Zealand, and of New Caledonia (Lillie and Brothers, 1970). In Australia, shallow non-marine platform sequences were deposited across large areas, with a brief Cretaceous marine transgression.

In summary, it is concluded that the southwest Pacific margin of Gondwanaland shows evidence of two episodes of subduction, each followed by a period of 'geosynclinal' deposition. Subduction is seen as a short-lived phenomenon (occurring for less than 50 m.y.), but related geological events occur over a period of up to 100 m.y., and across a zone up to 1000 km wide.

In the Late Jurassic, the first signs of rifting between Australia and Antarctica are also evident (Fig. 42.11). Extensive oil exploration activity has led to a very detailed knowledge of the southeast Australian continental margin (cf. Richards and Hopkins, 1969), which has been incorporated into a regional tectonic synthesis by Griffiths (1971a, 1971b, 1973b). A deepening rift system, including the Otway and Strzelecki Basins, was filled with coarse clastic sediments derived from its flanks. Forking of the rift, and some small relative movement of
Tasmania and mainland Australia, gave rise to a series of *en echelon* structures across the region (Fig. 42.12). At that time the Tasman Sea was also beginning to open, and marine conditions are recorded in western New Zealand (Wilson, in Fleming, 1970: 157). Subsequent sea-floor spreading in the Tasman Sea (80-60 m.y., Hayes and Ringis, 1973) and southeast Indian Ocean (55 m.y. to present, Weisell and Hayes, 1971) have brought the southwest Pacific Gondwanaland fragments to their present positions.

REFERENCES


approx. limit marine deposition

Fig. 42.9. Carboniferous to Permian. 'G' indicates general occurrence of glacial deposits.
Fig. 42.10. Permian to Lower Jurassic. BB = Bowen Basin; CMB = Clarence-Moreton Basin; MB = Maryborough Basin; SB = Sydney Basin.
Fig. 42.11. Upper Jurassic. D=dolerites; GAB=Great Artesian Basin.
Fig. 42.12. Lower Cretaceous. GAB = Great Artesian Basin (marine transgression).
Fig. 42.13. Upper Cretaceous
Structure and Evolution of the West Antarctic Part of the Circum-Pacific Mobile Belt

G. E. GRIKUROV AND B. G. LOPATIN

ABSTRACT

The three largest exposed mountainous regions of West Antarctica are described separately—Antarctic Peninsula, Marie Byrd Land with Eights Coast, and Ellsworth Mountains. All of them are underlain by the Riphean to early Palaeozoic geosynclinal folded (Ross) basement, above which the younger geosynclinal, orogenic and rift-volcanic assemblages outcrop along the Pacific coast. A folded Palaeozoic quasi-platform cover outcrops in the West Antarctic interior. The subglacial areas are mainly ancient plates with the Ross and/or pre-Ross metamorphic basement and the epi-Ross platform mantle.

Four major cycles in the tectonic evolution of West Antarctica are recognised: (1) the Riphean to early Palaeozoic (Ross), indicative of broad development of geosynclinal crust-forming processes on a pre-Riphean continental basement; (2) the middle Palaeozoic to early Mesozoic (Beacon), characterised by predominance of platform and quasiplatform regimes and by migration of geosynclinal and orogenic activities towards the Pacific margin; (3) the Jurassic to Cretaceous (early Andean), marked by complete extinction of geosynclinal processes succeeded by an intense orogenesis on the Pacific Coast and by a submergence of the ancient plates in the West Antarctic interior; and (4) the Cainozoic (late Andean), when the processes of crust fragmentation, differential block movements and riftogenesis resulted in partial basification of the continental crust.

INTRODUCTION

A number of tectonic schemes for Antarctica has already been published (Voronoï, 1964; Klimov, 1964, 1967; Ravich and Grikurov, 1970; Adie, 1962; Craddock, 1972a; Grikurov, 1972; Grikurov et al., 1972; Ford, 1972a, 1972b). Many of them favour a classical outlook on West Antarctic evolution suggesting a continuous migration of a geosynclinal crust-forming process towards the Pacific throughout the latest Precambrian, Palaeozoic and Meso-Cainozoic, and thus advocating a progressive lateral accretion on the East Antarctic ancient land mass. In our opinion West Antarctica is an ancient continental terrain, the structure of which has, since the Riphean, been complicated and modified by various epicratonic tectonic phenomena originating in the Pacific segment of the Earth.

In the vast ice-covered territory of Antarctica the bedrock subglacial relief provides the best natural basis for tectonic mapping. Figure 43.1 summarises the scarce topographic and geophysical data on the bedrock elevations. It shows distinctly the predominance of the ice-concealed shelf areas, submerged to a different degree below sea level and, in fact, continuous with well-known subglacial plains under the Ross and Filchner...
Ice Shelves. In sharp contrast to this general low-land terrain are three isolated mountain massifs of Antarctic Peninsula, Marie Byrd Land and Ellsworth Mountains; each of these is discussed below (Fig. 43.2).

**Antarctic Peninsula**

*The Riphean to early Palaeozoic geosynclinal folded complex.* The oldest rocks of the Antarctic Peninsula ('Basement Complex' of British authors) are correlated with the Ross Complex of the Transantarctic Mountains, though some of these rocks may belong to the pre-Ross (early-middle Riphean or older) metamorphic basement. The complex is 5000-6000 m thick; it consists of
variable metasedimentary and meta-volcanic rocks that had been, prior to the greenschist facies metamorphism, a geosynclinal assemblage of carbonate-terrigenous and slate-greywacke deposits with some basic volcanics. The Late Riphean age of the rocks and, consequently, their correlation with the Ross Complex, was suggested by acritarch studies (Iltchenko, 1972), but has not yet been confirmed by direct stratigraphic observations or isotopic age determinations; the latter show 'rejuvenated' ages in the range of 230-30 m.y. (Grikurov et al., 1967, 1971; Miller, 1960; Dalziel, 1972). Structure of the Ross geosynclinal unit is dominated by tight linear folds which often strike across the present elongation of the Peninsula and display features typical of polyphase deformation.

Various meta-intrusive rocks of gabbro-granite composition were described mostly from the Marguerite Bay area (Adie, 1954; Hoskins, 1963), but are also known from the central and southern parts of the Antarctic Peninsula. The contacts between them and stratigraphically dated units have never been observed. The isotopic ages of those rocks do not exceed 250 m.y. and their correlation with the Ross (or pre-Ross?) intrusions (Grikurov, 1972, 1973) is based mainly on degree of metamorphic alteration, and therefore remains tentative.

The middle-late Palaeozoic geosynclinal folded complex consists of two major assemblages which are closely related in age and space and may even be lateral equivalents. The most widespread and best studied is a thick slate-greywacke sequence, the Trinity Peninsula Series (Adie, 1957, 1964), the measured sections of which reach 3000-5000 m, but much greater estimates have also been claimed. The probable age of the uneroded part of the Trinity Peninsula Series is Carboniferous (Adie, 1957; Grikurov and Dibner, 1968), though it may range into the Permian (Grikurov, 1973).

Another assemblage is siliceous-volcanic in composition and variable in thickness. It comprises basic to acid meta-volcanic rocks, some quartzites and jasper-like rocks and volcanomictic sandstones and slates. Structural and petrographic speculations suggest that this sequence is a stratigraphic equivalent of the Trinity Peninsula Series.

Both units are closely related structurally and display tight linear folding with abundant isoclinal, overturned and even recumbent folds. The usual dip of axial planes is towards the Pacific, and the axial trace is parallel to the Antarctic Peninsula elongation. Only on Alexander Island is the orientation of the folds north-northwest and transverse with respect to the Pacific continental margin. The overturning of these folds is to the west.

*The Mesozoic (Jurassic) geosynclinal complex* has so far been reported only from southeastern Palmer Land (Laudon et al., 1969-70; Williams et al., 1972). It seems to include lower slate-greywacke and upper volcanic units, the former being palaeontologically dated as Jurassic. Both units are intensely folded into linear tight folds, either vertical or slightly overturned towards the Weddell Sea, and with axial traces parallel to the southern bend of Palmer Land. The age of the folding is Early Cretaceous, as intrusions dated as 90-100 m.y. (Halpern, 1967) are sharply discordant with respect to the folds.

*The Mesozoic orogenic complex* is represented by a wide suite of rocks, variable in composition but similar in structure, forming almost unfolded cover on the intensely deformed pre-Mesozoic rocks; between the latter and the Mesozoic orogenic complex there is a prominent regional unconformity. The lowest horizons of the orogenic complex seem to be of Early to Middle Triassic age (Orlando, 1968). The most widespread is a sedimentary-volcanic assemblage of Middle Jurassic to Early Cretaceous age, the base of which is locally conglomerates or sediments and tuffs, but the bulk of the section is composed of subaerial dacite-liparite lavas and coarse tuffs up to 3000 m thick. The Lower Cretaceous terrigenous flysch-like molasse is also about 3000 m thick, though it occurs only locally in the deepest intermontane troughs along the western coast of the Antarctic Peninsula. The Upper Cretaceous molasse has the same thickness and forms a rather widespread cover of conglomerates, slates, and loose sands off the eastern coast of the Peninsula.
Among the intrusives, gabbro and granite of 'Middle' Cretaceous (100 ± 15 m.y.) age are the most abundant. Emplacement took place during the climax of orogenic uplift before the Senonian. In some localities the older (Rhaetic to Liassic?) granites can be distinguished by petrographic and isotopic data. These are correlated with the early stages of an epigeosynclinal orogenesis, probably synchronous with the end of the folding in the Palaeozoic geosynclinal formations.

The Cainozoic rift-volcanic complex is represented mainly by high-aluminous lavas and tuffs of andesitic-basaltic composition 1500-2000 m thick. The complex is structurally undeformed, except for dislocations due to the Recent block movements. Its age is Miocene to Holocene, judging from geological and palaeontological data, but isotopic dates rarely exceed 4-5 m.y. The early Cainozoic (50-55 m.y.) intrusives of high-aluminous gabbro-diorites forming small sub-volcanic stocks along the rift zones are apparently genetically related to this episode.
The Antarctic Peninsula is a Mesozoic orogen developed from the middle-late Palaeozoic geosynclinal system. The latter, in turn, was underlain by a still older (Riphean) continental basement. The most conspicuous tectonic feature over the greater part of this region is a sharp structural unconformity at the base of the Mesozoic strata. The predominance of calcalkaline magmas and continuous increase in basicity and alumina content during the Mesozoic and Cainozoic, are typical.

Marie Byrd Land and Eights Coast

The Riphean-Early Palaeozoic geosynclinal folded complex and its basement projections. It is believed that the pre-Ross metamorphic basement of Marie Byrd Land is represented by a metamorphic complex of the Fosdick Mountains. Its thickness is estimated as 5000 m, and it consists largely of paragneisses and schists (with highly aluminous rocks at the base of the section) that had been intensely, though not uniformly, migmatised and granitised under high-grade amphibolite facies (Grikurov and Lopatin, in press). The folding is very complicated with numerous small plications striking east-northeast and west-northwest. No direct contacts of the Fosdick Mountains metamorphics with other rocks have been observed, and all isotopic age determinations of schists and migmatites fall within the range of 100-200 m.y. (Lopatin et al., in press; Halpern, 1972); therefore our assignment of this complex to the pre-Ross (early-middle Riphean or pre-Riphean) basement is tentative, and other ages (Halpern, 1972; Wilbanks, 1972; Craddock, 1972b) remain possible.

In eastern Marie Byrd Land and Thurston Island the essentially meta-intrusive rocks have been dated isotopically at 200-500 m.y. These metamorphics may belong to the pre-Ross basement (Lopatin and Orlenko, 1972) or represent the equivalents of the Ross (?) meta-intrusives of the Antarctic Peninsula. On the Hobbs Coast there are also some gneissose gabbros, diorites and granodiorites of unknown structural position with isotopic ages of 320-470 m.y.

A geosynclinal assemblage apparently of Ross age is best represented in western Marie Byrd Land (Ford Mountains), where its thickness reaches 10,000 m. It is composed
of phyllites, altered siltstones and sandstones with minor quartzites and quartz-albitophyres. The folds in the assemblage are broad and symmetrical, with a constant northwest axial strike. Latest Precambrian acritarchs (Il'tchenko, 1972) and the earliest Cambrian microphytoliths have been recovered from the metasediments, while the isotopic age of the phyllites is 450-470 m.y. The metasedimentary sequence is cut by middle Palaeozoic (320-350 m.y.) granodiorites and adamellites of orogenic type, as well as by younger intrusions.

The middle-late Palaeozoic (?) geosynclinal complex in Marie Byrd Land is probably represented by various meta-volcanic rocks which are a few kilometres thick and which were subjected to folding about northwest axes. The isotopic ages of the metavolcanics range from 115 to 370 m.y. According to Klimov (unpubl.) the metavolcanic unit is more likely to represent a eugeosynclinal facies of the Ross complex, whereas American geologists would rather assign it to the Mesozoic (Jurassic?) on the basis of isotopic ages.

Much-deformed gneissose rocks of gabbro-granodiorite composition dated as 200-300 m.y. may be related to the middle-late Palaeozoic geosynclinal complex. These rocks form large plutons in eastern Marie Byrd Land and particularly on Thurston Island.

The Mesozoic orogenic complex. The early Mesozoic orogenic sequences are exposed in eastern Marie Byrd Land and Thurston Island where a few fragmentary sections of undeformed andesite-rhyolite volcanics have been observed and isotopically dated as 150-200 m.y. Similar ages (160-230 m.y.) were obtained for adamellites and subalkaline granites from Thurston Island and Jones Mountains. Ages of 140-200 m.y. have also been obtained for high-aluminous olivine gabbros from the Thurston Island area, but the petrochemical characteristics of the latter, and their apparent distribution along a major fracture zone, suggest the possibility that they are younger and are related to the Early Cainozoic rift-volcanics of the Antarctic Peninsula.

The late Mesozoic orogenic intrusives are the most widespread. They penetrate all the older rocks in the area and isotopically 'rejuvenate' them to about 100 m.y. In western Marie Byrd Land the Cretaceous intrusions are of alkaline type and usually consist of subalkaline biotite granites and quartz syenites with isotopic ages ranging between 90 and 150 m.y. (mainly 100 ± 10 m.y.). In eastern Marie Byrd Land and Thurston Island, however, the Cretaceous (90-100 m.y.) migmatic rocks are much closer petrochemically to the normal calcalkaline varieties of the synchronous orogenic intrusives of the Antarctic Peninsula.

The Cainozoic rift-volcanic complex consists of three units. The oldest, trachybasaltic, forms the Miocene volcanic plateau a few kilometres thick. The Pliocene trachyte-phonolite-kenyte association forms large isolated volcanoes. The youngest unit is trachybasaltic, and is localised in parasitic cones on the slopes of major volcanoes. The plateau-basalts usually yield isotopic ages of 24-18 m.y., and the younger volcanics 12-0 m.y. Throughout the volcanic section there are features of subglacial eruption. In eastern Marie Byrd Land and Thurston Island the alkalinity of the volcanic rocks is somewhat lower and the proportion of high-aluminous calcalkaline varieties grows significantly.

Thus, Marie Byrd Land and Eights Coast are an orogenic area formed by the end of the Mesozoic as a result of recurrent orogenesis on a heterogeneous geosynclinal folded basement. The latter consisted of extensive Ross massifs and their continental basement, and younger fold zones developed from Palaeozoic (?) geosynclinal troughs that were underlain by a continental crust of Ross age or older. However, at present it is impossible to delineate the boundary of these two main structural types. It must also be emphasised that, though the Ross massifs preserved their internal structure both there and in northern Victoria Land, they are not to be considered as full equivalents of the typical Ross structures of the Transantarctic Mountains, since they were subject to recurrent orogenesis throughout the Phanerozoic. Because of this they are devoid of a platform cover.

Ellsworth and Pensacola Mountains

This region is shown in Figure 43.2 as a
separate tectonic province, marked by the presence of structural features typical both of mobile and stable middle Palaeozoic to early Mesozoic areas. Their geology and structure have been described earlier (Schmidt et al., 1964; Schmidt and Ford, 1969-70; Craddock, 1970, 1972a, 1972b; Ford, 1972a, 1972b; Ford and Boyd, 1968; Ford and Nelson, 1972; Craddock et al., 1964; Webers, 1972). Therefore we shall confine the discussion of the area to palaeotectonic interpretation.

The Ross Complex in the Pensacola Mountains is separated by a structural unconformity into a lower (Riphean) intensely folded geosynclinal meta-volcanic-sedimentary sequence about 10 km thick, and an upper (Cambrian to Ordovician?) folded oro-

Fig. 43.3. Main structures of the Late Andean cycle of West Antarctic evolution. 1. the areas of negligible submergence not followed by crust transformation; 2. the areas of moderate submergence with intermediate type of crust, a) within the rift zones; 3. the areas of intense submergence with suboceanic and oceanic crust; 4. the areas of uprising (orogens), not differentiated by the amplitude of uplift. Pre-Late Andean tectonic substratum: 5. the areas of pre-Ross consolidation; 6. the areas of Ross consolidation, a) refolded in the Early Mesozoic; 7. the areas of Mesozoic consolidation and/or recurrent Middle Palaeozoic-Mesozoic orogenesis; 8. boundaries of major structural zones; 9. sea level.
genic sequence of variable thickness and composition. In the Ellsworth Mountains the Ross Complex is similar in thickness and structure, though it contains no igneous rocks and no internal unconformity. Thus in this area it cannot be subdivided into geosynclinal and orogenic parts.

The next structural stage in both areas under discussion is very peculiar: it is practically identical in stratification and composition to the pre-Mesozoic part of the Beacon complex, but differs from the typical Beacon platform strata in its much greater thickness and a degree of structural deformation which apparently occurred in the early Mesozoic (Triassic or earliest Jurassic). In the Pensacola Mountains the platform-like assemblage of Devonian to Permian age overlies the eroded surface of the Ross Complex with marked unconformity and, though folded, is not dislocated as intensely as the Ross basement; the magmatic activity that succeeded the early Mesozoic deformation is represented by a unique differentiated trap intrusion of the Dufek Massif.

In the Ellsworth Mountains, however, there is no visible structural unconformity at the base of the platform-like Beacon Complex. Moreover, the latter seems to be dislocated by the early Mesozoic diastrophism intensely enough to form a fold pattern conformable with that in the underlying Ross basement; the synchronous magmatic activity in the Ellsworth Mountains is represented by Early Jurassic granites.

It should be noted that the Ellsworth Mountains are characterised by a highly dissected bedrock relief. A typical Alpine landscape, almost uneroded, includes the highest Antarctic peak (over 5000 m), while in direct juxtaposition to the mountain massif there are the deepest subglacial depressions in which the bedrock surface is submerged 2500 m below sea level.

The brief comparative outline of the structure of the Ellsworth and Pensacola Mountains given above makes it possible to consider the former as a present-day orogen, growing on the site of the early Mesozoic orogen. The latter was, in turn, related to the folding of the Devonian-Permian formations and was synchronous with an epigeosynclinal orogenesis in the West Antarctic fold system at the Pacific margin of the continent. However, in the Ellsworth Mountains the early Mesozoic orogeny and associated minor granite intrusions succeeded, not a geosynclinal, but a platform-like regime of the middle-late Palaeozoic. That is why this area can be regarded as a structure intermediate in type between the Transantarctic Mountains and the Pacific margin—a structure which had, after the Ross tectogenesis, become neither an epi-Ross platform nor a middle-late Palaeozoic geosynclinal zone. In the Pensacola Mountains the deviations from a platform evolution in middle Palaeozoic-early Mesozoic time are not so prominent, but still conspicuous enough to prevent direct correlation with typical areas of Ross solidation.

The Structure of Subglacial Areas

There is little evidence available on the upper crustal structure of the vast subglacial terrain of West Antarctica; the information known to the authors was obtained by seismic refraction measurements carried out under the USARP program and summarised elsewhere (Bentley and Clough, 1972; Bentley and Chang, 1971). These data seem to agree quite satisfactorily with geological observations on localities adjacent to the exposed mountain massifs where the bedrock surface is at or close to the sea level. In the areas where the bedrock surface is deeply submerged there is an apparent absence of a seismic layer comparable to the geosynclinal units of Ross age or younger. Such a structure is compatible with platform plates or median masses of the pre-Riphean or early-middle Riphean cratonisation.

Major Cycles of Evolution

The pre-Riphean history of West Antarctica is obscure. In our opinion the most likely alternative is the existence of an ancient continental crust comparable to the pre-Riphean crystalline basement of East Antarctica. However, this assumption at present remains unproved. On the other hand, there are no grounds to suggest the development of West Antarctic Pacific-type structural elements on a primitive oceanic crust.

Since the Riphean four major cycles of tectonic evolution can be distinguished.
The Riphean to early Palaeozoic cycle has already been named 'Ross'. The major tectonic process during that cycle was the formation of continental crust in an extensive epi-cratonic geosynclinal system on the present site of the Transantarctic Mountains and the West Antarctic mountain massifs. We believe that this crust-forming function consisted mainly of vertical accretion of the pre-existing supra-crustal complexes. Complexes of the latter type that lie within the present low-land areas were not involved in the Ross geosynclinal subsidence, and probably acted as extensive rigid masses of a median type, separated and surrounded by the Ross fold zones.

Cratonisation resulting from the Ross tectogenesis apparently decreased towards the Pacific, and the Ross folding was final in the strict sense only within the greater part of the Transantarctic Mountains with the exception of the Pensacola Mountains and northernmost Victoria Land. The apparent absence of the orogenic assemblages of the Ross Complex on the Pacific coast indicates the reduction of the degree of orogenic development of Ross geosynclines in the same direction.

Thus, we consider West Antarctica as a part of the pre-Riphean Pan-antarctic craton. This western part of the craton had split into several fragments by the beginning of the Ross cycle, but by the end of it the fragments were again welded together by thickened seams of the continental crust, i.e. by the Ross geosynclinal fold systems.

For the middle Palaeozoic to early Mesozoic cycle the term 'Beacon' is introduced in a chronological sense, as during that time interval one of the most characteristic structural complexes of Antarctica was formed, namely, the epi-Ross platform cover known to Antarctic geologists as the Beacon Super-Group. In the areas of the epi-Ross consolidation the Beacon time-span can be defined fairly precisely from the stratigraphic relations of the platform formations ranging in age from Devonian to Liassic. In West Antarctica no typical platform sections are exposed, though they are probably present over the Ross median masses now buried under the shelf and continental ice. Nevertheless, the Beacon time-span can be easily distinguished even in the Pacific-type mountainous regions.

At this time in the Antarctic Peninsula and, presumably in the coastal zone of Marie Byrd Land, geosynclinal subsidence and the early stage of epigeosynclinal orogenesis occurred, forming early Mesozoic fold zones. Though the Ross folded complex in western Marie Byrd Land and northernmost Victoria Land was not directly involved in the Palaeozoic geosynclinal subsidence, it still was structurally modified during the Beacon cycle by a recurrent orogenesis with associated granite activity. Finally, in the Ellsworth Mountains area the Beacon cycle is displayed in the accumulation (on the Ross basement) of the platform-like Devonian-Permian strata, in their Triassic folding, and in Liassic intrusive activity.

The major tectonic process of the Beacon cycle was the cessation of growth of new portions of continental crust over the greater part of West Antarctica, and the localisation of a geosynclinal crust-forming activity within a relatively narrow zone along the Pacific coast of the continent. The pre-Beacon continental blocks directly adjacent to this zone were subject to orogenic transformation, but on the whole a stable continental platform regime was dominant almost everywhere in West Antarctica. The Ellsworth Mountains, which failed to become cratonised during the Ross tectogenesis, were involved in a rather intense Beacon subsidence and subsequent crustal deformation, but due to the non-geosynclinal nature of the latter, no significant addition to the pre-Beacon continental crust occurred.

Thus, by the end of the Beacon cycle the present-day West Antarctic Mountains and the Transantarctic Mountains had been the Early Mesozoic orogens separated by ancient plates that later became subject to intense crustal submergence.

The Late Mesozoic cycle spans most of the Jurassic and all of the Cretaceous. The palaeotectonic events during that cycle can only be reconstructed for the Pacific-type areas of West Antarctica, and therefore it is reasonable to name this cycle 'Early Andean' in order to emphasise the relations between
the late Mesozoic West Antarctic structures and the synchronous tectonic elements of the Andes.

During the Early Andean cycle a geosynclinal crust-forming process was finally terminated within the Antarctic Peninsula by the late Mesozoic folding that led to the closure of the Jurassic geosynclinal trough in southeastern Palmer Land; in the zone of the early Mesozoic folding, substantial orogenic plutonic activity occurred within the anticlinoria which thus became saturated with sialic material. In Marie Byrd Land recurrent orogenesis, initiated during the previous cycle, continued; the non-geosynclinal nature of this orogenesis now became quite apparent, but its role in the transformation of the pre-Mesozoic continental basement was nevertheless very considerable as judged from the scope and intensity of the late Mesozoic plutonic activity.

It seems likely that the Ross-Beacon plates of ancient-Riphean or pre-Riphean origin continued their slow submergence. Locally subsidence must have been more intense to give rise to extensive tectonic depressions that would later become the major ice-accumulating basins. For the Ellsworth and Transantarctic Mountains a regime of moderate orogenesis and/or denudation rather than subsidence may be assumed.

Thus, during the Early Andean cycle the formation of new continental crust ceased over all of West Antarctica. However, a new form of tectonic activity developed, namely transformation of the previously accreted continental crust. This transformation continued as an orogenic uplift in the areas of Ross and Beacon crustal movements, but it also occurred simultaneously with intense submergence in the areas of earlier consolidation. In the Early Andean cycle, therefore, there was a change from a period of construction to one of destruction of continental crust.

The Cainozoic cycle will be termed 'Late Andean' to stress the time relations between the neotectonic activisation of West Antarctica and major mountain-building in the Andes. The structural scheme for this cycle is shown in Figure 43.3.

During the Palaeogene, the Pacific moun-

tainous margin of the continent must have undergone a considerable erosion reflected in formation of the pre-Miocene peneplain. Consequently, the general tectonic environment must have been fairly passive, but locally the rift zones with associated volcanism were already active. Somewhere in the West Antarctic interior, however, high mountains must have existed to initiate the continental glaciation; one of these could be represented by the Ellsworth Mountains.

In the Miocene the splitting of the continental crust along the Pacific margin caused intense block movements, the formation of rifts, and surface eruptions of great amounts of basaltic magmas. The eruptions were subaerial, submarine or subglacial, and varied in composition from calc-alkaline to alkaline lavas, but on the whole they covered large areas. Within the rift zones the structure of the crust was changed—an upheaval of low-density mantle material took place there, accompanied by disappearance of the granitic layer, and overall crust-thinning. Similar phenomena could have taken place in the deepest parts of tectonic depressions in the West Antarctic interior. On the other hand, the Ellsworth Mountains underwent a considerable emergence; they only recently pierced through the ice dome that was simultaneously somewhat lowered as the result of partial deglaciation.

It can thus be seen that in the Late Andean cycle of West Antarctica there was a noticeable revival of tectonic activity, causing extension and splitting of the continental crust; large-scale rift-bounded volcanism, block movements, and destruction of the continental lithosphere within zones of maximal submergence. All these processes took place simultaneously with a fast growth of the ice cover. Later there was a slow pulsing recession of the ice, the tectonic function of which will be given proper attention in future studies.

CONCLUSIONS

1. West Antarctica consists of large, ancient, cratonised blocks (plates), separated and surrounded by epicratonic mobile belts of Riphean age and younger. During the Riphean and Phanerozoic, the ancient plates
experienced mainly irregular subsidence, while the mobile belts went through various stages of geosynclinal, orogenic and rift-volcanic evolution. The final outcome of this complicated history was the change, in Cainozoic time, from a crust-forming (constructive) tectonic evolution to a process of partial destruction of the pre-Cainozoic continental basement.

2. Specific features of West Antarctic evolution, contrary to that in the eastern part of the continent, are caused by the tectonic processes which were generated within the Pacific segment of the Earth and exerted their influence upon the pre-Riphean continental basement of the surrounding landmasses. As a result of this influence, the West Antarctic continental lithosphere has long been subject to substantial transformation, and is still being modified. In East Antarctica (beyond the Transantarctic Mountains) such transformations had practically ceased to occur after the earliest Palaeozoic, while recent destructive processes are either limited or absent.

3. The West/East Antarctic boundary is, in terms of tectonics, a major structural element in the Earth's crust that separates Pacific and Gondwana segments in their south polar part. With respect to this principal separation the contour of the Antarctic platform is a subordinate tectonic feature which changed its position throughout geologic history.

4. If our concept of a single pre-Ross Panantarctic continental craton is correct, then an epicratonic regeneration of the Ross geosynclinal systems can be regarded as the first impulse in the expansion of Pacific segment onto Gondwana segment. This impulse appeared 'successful' for the Pacific margin of Antarctica only, where a high crust-forming and/or crust-transforming tectonic mobility was maintained after the Ross cycle and throughout the Palaeozoic and the Mesozoic, while in the rest of West Antarctica the Gondwana trend of platform (crust-conserving) evolution dominated almost all the Phanerozoic supracrustal history. However, the profound influence of the Pacific segment of West Antarctica has never ceased; it turned this vast territory into a 'mobile platform' by the end of the Mesozoic, and then began its destruction. In our opinion this process of destruction of the previously formed continental crust can be regarded as a new impulse of the Pacific expansion that differs basically from that of the Ross cycle.

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Contrasting Structures and Origins of the Western and Southeastern Continental Margins of Southern Africa

R. A. SCRUTTON, A. du PLESSIS, A. M. BARNABY AND E. S. W. SIMPSON

ABSTRACT

The contrasting structures and origins of the Atlantic and Indian Ocean continental margins of southern Africa are ascribed to two coeval and interrelated types of lithospheric plate interaction.

Along the Atlantic Ocean margin, between the Agulhas Bank and the Walvis Ridge abutment, the tensional regime associated with the rifting and opening of the South Atlantic has produced relatively simple basement structures. Seismic, magnetic and gravity data are consistent with the presence of a thick prograded sediment wedge, which is interrupted by basement rises near Cape Town, possibly near Luderitz and at the Walvis Ridge abutment. At the Walvis Ridge abutment the simple basement structures and subdued sea-floor relief characteristic of the sediment wedge are replaced by steep-sided basement ridges and associated steep bathymetric scarps.

The Indian Ocean continental margin is also believed to be related to the opening of the South Atlantic, having developed along a line of transcurrent motion between the South American plate (Falkland Plateau) and the African plate during the early stages of opening. A marginal fracture ridge, probably formed at the same time as the margin itself, exists beneath the continental slope. Together with older continental basement structures it has controlled the size and shape of the Mesozoic-Cainozoic sedimentary basin on the continental shelf.

INTRODUCTION

Plate tectonics theory postulates that the different types of boundary between continental and oceanic crust (continental margins) have formed as a result of different types of lithospheric plate interaction (Morgan, 1968). For example, most of the boundaries within the Red Sea (the best example of an ocean in the making) have been formed by thinning of the continental crust under tension, normal faulting and rifting, with the upwelling of basaltic material into the gap produced by the separation of the Arabian and Nubian plates (Lowell and Genik, 1972). In the adjacent Gulf of Aden, however, in addition to rifted boundaries, large sections of the continental margin have been formed by strike-slip motion between the Arabian and Somali plates. At the time of break-up the rift zone was offset in a number of places, so that as the two plates moved apart shear occurred between continents along the fracture zones connecting the offset rift segments (Laughton et al., 1970). At the other end of the Red Sea this type of shear motion is taking place at the present day in
the Dead Sea rift (Freund, 1965). The motion occurs along small circles centred on the pole of rotation for the two plates.

The same two types of lithospheric plate interaction were probably responsible for the formation of the continental margins around the South Atlantic during the initial opening of that ocean in Early Cretaceous times (Le Pichon and Hayes, 1971; Francheteau and Le Pichon, 1972). Those sections of margin parallel to the rift zone were formed by tensoidal faulting and rifting, while those sections aligned along small circles of opening, which meet the rifted sections at right angles, were formed by strike-slip faulting of one continental block against the other. The offsets created in the margins are called 'marginal offsets' (Le Pichon and Hayes, 1971).

Of the continent-ocean boundaries around southern Africa the one facing the Atlantic Ocean clearly seems to be of the rifted type. We may suppose that the one facing the Indian Ocean is similar, since that ocean is also an expanding ocean formed by sea-floor spreading. However, a closer look at the structure of the southeastern margin shows

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**Fig. 44.1.** Bathymetric map of the continental margins around southern Africa, redrawn from a chart compiled by E. S. W. Simpson and E. Forder at the University of Cape Town. Isobaths are at 500m intervals with the addition of the 200m isobath (broken contour). Depths given in km. The inset shows the location of our study area with respect to Africa.
that this is not so; it almost certainly originated as a gigantic marginal offset belonging to the South Atlantic rift system.

ATLANTIC OCEAN (WESTERN) CONTINENTAL MARGIN

The part of the Atlantic continental margin of southern Africa that concerns us here extends from the southern tip of Agulhas Bank (37°S) to 18°S, the point at which the Walvis Ridge abuts the continent (Fig. 44.1). This so-called Walvis Ridge abutment forms a natural northern limit to our study area, because the strong WSW-ENE trend of its north-facing scarp coincides with a striking change in sea-floor morphology (Simpson, 1971). One of the most important features of the western margin between these geographical limits is its continuous and more or less straight SSE-NNW trend. There is no evidence in the bathymetry for large offsets of the type mentioned in the Introduction.

In general, the sea floor exhibits low relief, except near, and to the south of, Cape Town where numerous submarine canyons cut into the continental slope. The gentle gradient of the slope merges smoothly with the subdued topography of the broad, well-developed continental rise, which in turn merges westwards with the floor of the Cape Basin. The impression gained from studying the bathymetry is that the rifting process responsible for the formation of the margin was not accompanied by major distortion or fragmentation of the continental edge.

Much of the information on the structure of this margin has been derived from marine geophysical studies. Seismic reflection data (Fig. 44.2) show that throughout its length continental basement forms a submerged ledge adjacent to the shoreline. The width of this ledge reaches 50 km in the vicinity of Cape Town, but elsewhere it is only a few kilometres wide (Dingle, 1971 and in press a; van Andel and Calvert, 1971). Seaward, basement drops rapidly and, apart from near the southern tip of Agulhas Bank, is not seen again on seismic reflection records until oceanic Layer 2 appears beneath the base of the slope (Uchupi and Emery, 1972).

Beneath the middle and outer shelf and the slope the topography and composition of basement must be inferred from seismic refraction, gravity and magnetic data. This has been done for the region between 28°S and 32°S by Scrutton (in press a). The crustal model he obtained indicates that seawards from the inshore ledge, basement drops rapidly to about 3 km below sea level (see profile III, Fig. 44.2). It then continues as a 150 km wide platform at about 4 km depth, underlain by high density rocks. Beyond this platform basement drops again, to a depth of 6 km, to meet oceanic Layer 2 beneath the base of the slope. Magnetic data from the nearby shelf and slope (Simpson and du Plessis, 1968) show a sequence of linear anomalies striking parallel to the continental margin. It seems most likely that these are related to the high density rocks beneath the wide basement platform. According to the gravity interpretation the total thickness of the crust decreases gradually from 30 km at the coastline to 15 km beneath the continental rise; this is typical of rifted margins.

The above model of crustal structure applies only to the region between 28°S and 32°S. Seismic reflection, gravity and magnetic data (du Plessis et al., 1972; Uchupi and Emery, 1972; Talwani and Eldholm, 1973) suggest that outside this region structures are different. From just north of Cape Town to the tip of Agulhas Bank continental basement does not form a wide platform at depth beneath the shelf. Instead, there is a basement high termed the Agulhas Arch (Fig. 44.5), the western flank of which dips seawards at about 5°. Du Plessis et al. (1972) have proposed that another, but smaller, high occurs in the Luderitz to Walvis Bay area, where there is a change in shallow structure and magnetic anomaly pattern. At the Walvis Ridge abutment basement rises to 2.5 km below sea level to form a plateau beneath the continental slope (Barnaby, in prep.). Extending west-southwestwards from this plateau are two steep-sided ridges (Fig. 44.5). The trend of these ridges perpendicular to that of the margin, and the fact that the northern scarp of the abutment coincides with an apparent right-lateral offset in the margin, has led to the suggestion that the easternmost section of the Walvis Ridge is associated with a major
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Fig. 44.2. Selected seismic reflection and total intensity magnetic profiles over the western continental margin of southern Africa. For their locations, see inset. Heavy stippling on the seismic profiles indicates basement rocks. The dashed line labelled B on profile III is the basement deduced from gravity measurements (Scrutton, in press a) and described in the text; it has been drawn assuming 2 km/s$^{-1}$ for the velocity of sound in the sediments.

South Atlantic fracture zone (Francheteau and Le Pichon, 1972; Barnaby, in prep.).

Since the formation of the western continental margin, sedimentation and crustal subsidence have been active along its entire length, but they have been greatest in those areas away from the basement highs (Fig. 44.5). (Highs occur at the Walvis Ridge abutment, perhaps between Walvis Bay and Luderitz, and near Cape Town). Seismic (Bryan and Simpson, 1971) and gravity (Scrutton, in press a) results show that the sediment wedge that has developed thickens seawards to reach maximum dimensions beneath the outer shelf (Dingle, in press b). During the early history of the margin, upbuilding of the wedge appears to have been dominant (Dingle, in press b), but since then outbuilding has become progressively more important (du Plessis et al., 1972).

INDIAN OCEAN (SOUTHEASTERN) CONTINENTAL MARGIN

The continental margin facing the Indian Ocean between Durban and the southern tip of Agulhas Bank is characterised by a steep continental slope with a persistent northeast-southwest trend (Fig. 44.1). The slope lies parallel and very close to the coastline in the northeast, but southwest of East London it diverges from the coast as the shelf broadens to form the Agulhas Bank. At a point 250 km south of Cape Agulhas it turns through 90° to face the Atlantic Ocean. Morphologically, the slope is a single steep scarp with a gradient of up to 7°, except on the east flank of the Agulhas Bank where it is divisible into a gentle upper slope and a steep lower slope (Fig. 44.3). Throughout the margin the continental rise is poorly developed or absent.

Early opinions on the structure and origin of this margin were influenced by investigations made on land. The uninterrupted trend and steepness of the slope led Kent (1938) to postulate a fault-controlled origin for it, using the pattern of faults he had mapped in coastal Natal (Fig. 44.5) as evidence for his hypothesis. In 1960, following more extensive mapping of the Natal faults, Beater and Maud (1960) and Maud (1961) proposed that northwest-southeast tension was the dominant stress regime, originating in the removal of lateral support when a continental mass broke away from southeastern Africa. They thought of the continental slope south of Durban as a series of step faults. Simpson (1966), on the other hand, considered the slope to be a large hinge fault, with the hinge in the vicinity of Durban and displacement increasing southwestwards. Between Durban and East London (Fig. 44.3) the faulted nature of the continental edge is further attested by the abrupt truncation at the slope of a set of easterly trending magnetic anomalies (Beattie, 1909). Du Plessis and Simpson (in prep.) have recently calculated from marine magnetic data that the causative magnetic bodies, which are probably related to Karroo magmatism, locally lie in the upper 4 km of the continental crust.

During the past few years evidence on the structure of the margin has been derived mainly from marine geophysical studies, chiefly magnetometric and seismic. Between Durban and the southern tip of Agulhas Bank (Fig. 44.4) a prominent positive magnetic anomaly overlies the base of the continental slope (du Plessis and Simpson, in prep.). It divides a region of relatively quiet magnetic field over the continental shelf and slope from a less quiet field over the adjacent oceanic areas. Talwani and Eldholm (1973) have interpreted the anomaly as being caused by the truncated edge of a magnetised, horizontal plate extending beneath the Transkei and Agulhas Basins seawards of the anomaly. Because seismic refraction experiments (Green and Hales, 1966; Ludwig et al., 1968) have discovered oceanic crust beneath these basins, it is logical to correlate the magnetised plate with oceanic basement. Du Plessis and Simpson (in prep.) have shown
that this correlation is correct and that the truncated, probably faulted, edge of the plate marks the boundary between oceanic and continental crust. Using the positive mag-
netic anomaly as a guide (Fig. 44.3, inset), the boundary is determined to be at the base of the continental slope east of 23°E, but to the west it diverges from the slope and occurs on the continental rise.

There are now several lines of evidence favouring a fault-controlled origin for the continental slope southeast of southern Africa. In addition to those mentioned above, seismic refraction results (Ludwig et al., 1968; Hales and Nation, 1972) and a preliminary interpretation of marine gravity data indicate a relatively rapid transition from continental to oceanic crust beneath the margin. It is therefore significant that Francheteau and Le Pichon (1972) have shown the trend of the slope to be approximately coincident with the 68° small circle about the pole of rotation for the early phase of South Atlantic opening. Furthermore, they have noted that the Falkland Plateau, as a continental feature belonging to the South American lithospheric plate (Ewing et al., 1971), fits around southern Africa in the reconstruction of West Gondwanaland, reaching almost as far as Durban (Fig. 44.3). When the South Atlantic opened, it would appear that the Falkland Plateau moved with right lateral strike-slip motion past southern Africa, producing the steep, fault-controlled continental slope along the line of the 68° small circle (Fig. 44.5). If this is so, the southeastern continental margin represents a gigantic marginal offset, some 1200 km in length, related to the South Atlantic rift system. The oceanic crust now underlyiing the Natal Valley, Transkei Basin, Agulhas Plateau and Agulhas Basin would have been generated at the offset mid-ocean ridge segment lying northeast of Falkland Plateau.

The theory of Le Pichon and Hayes (1971) concerning marginal offsets predicts the formation of a ‘marginal fracture ridge’ within, and parallel to, the continent-continent shear zone during the time the zone was active. Scrutton and du Plessis (1973) have reported that a non-magnetic basement ridge, which may be of the marginal fracture type, is present beneath the continental slope southwest of Port Elizabeth (Figs. 44.3 and 44.5). It lies within the postulated shear zone but can only be traced with certainty over a distance of 350 km. South-westwards it appears to give way to a narrow trough which lies at the base of the slope (Fig. 44.5). This trough is also magnetically quiet and may be underlain by material similar in composition to that of the ridge. Further to the southwest the trough continues beyond the tip of Agulhas Bank, following the line of the fracture zone responsible for the marginal offset. This fracture zone has been named the Agulhas Fracture Zone (Emery, 1972; Talwani and Eldholm, 1973).

The size and shape of the sedimentary basins that have developed on the southeastern margin during the break-up and dispersal of West Gondwanaland were controlled by old as well as newly created basement structures. One of the old structures is the Agulhas Arch, a pre-Mesozoic basement uplift (Dingle et al., 1971) that trends southwest from Cape Agulhas to the tip of Agulhas Bank. Together with the marginal fracture ridge (Fig. 44.5) it formed a large confining basin in which up to 5.7 km of Upper Mesozoic and Cainozoic sediments were deposited (Dingle, in press b). The basin is divided by a number of smaller basement highs (Leyden et al., 1971), which appear to be the seaward extensions of major anticlinal arches of the Cape Fold Belt, and is closed at its eastern end near Port Elizabeth by the Recife Arch (Dingle, in press b), a feature similar to the Agulhas Arch. Northeast of Port Elizabeth the Upper Mesozoic-Cainozoic sediment cover is thin on the shelf and often absent on the continental slope. It seems that the narrow, steep margin in this region has not been able to retain sediments. The latter have largely been swept beyond the continental margin into the Natal Valley. A second sedimentary basin lies within the narrow trough that appears to be the south-westward continuation of the proposed marginal fracture ridge. At least 2 km of sediment of unknown age have accumulated here.

SUMMARY AND DISCUSSION

The contrasting morphologies, structures and origins of the two continental margins of
southern Africa may be summarised as follows:

1. Bathymetry. Each margin possesses its own characteristic bathymetry and uninterrupted trend. The western margin exhibits low sea-floor relief and gentle gradients, and its slope strikes SSE-NNW. The southeastern margin is characterised by higher relief and steeper gradients, and the slope has an overall northeast-southwest trend, becoming ENE-WSW at the Agulhas Bank. The two margins meet at right angles near the tip of the Bank.

2. Deep structure. The change in crustal thickness between continental and oceanic areas takes place gradually off the west coast but rapidly off the southeastern coast. Beneath the western margin the boundary between continental and oceanic basement is not well defined, but beneath the southeastern margin a large barrier ridge occurs adjacent to the boundary and the boundary itself is sharp and apparently faulted.

3. Shallow structure. A prograding sediment wedge of variable thickness has developed along the full length of the Atlantic continental margin. On the Indian Ocean margin confining basement structures have restricted the accumulation of thick sediments to one part of the shelf.

4. Origins. The Atlantic continental margin originated under tension by rifting of the crust and the separation of South America from Africa. The Indian Ocean margin originated as a shear zone between the South American and African lithospheric plates, which was active during the early stages of opening of the South Atlantic. It may therefore be regarded as a gigantic marginal offset in the Atlantic continental margin.

If the two continental margins originated at the time the South Atlantic opened, geological phenomena related to the opening should give similar ages on both margins. Along the western margin, and on the South American side of the Atlantic, volcanic activity (Amarell et al., 1966; Siedner and Miller, 1968), the earliest post-rifting marine sediments (Zambrano and Urien, 1970; Dingle, in press a) and the oldest sea floor (Maxwell et al., 1970) are all Late Jurassic or Early Cretaceous in age. On our southeastern margin, however, the data are not as good because of the masking effect of the Indian Ocean opening, which occurred in mid-Jurassic times (Scrutton, in press b). Nevertheless, a widespread marine transgression at the end of the Jurassic (Haughton, 1969) and Late Jurassic-Early Cretaceous faulting in the Natal region (Beater and Maud, 1960; Maud, 1961) may both be indications of break-up in West Gondwanaland. The Natal faulting is worthy of further, brief comment in view of the new ideas on the mode of formation of the southeastern margin.

Three stress systems may have been active in the Natal region as the Falkland Plateau slipped southwestwards along the margin: (a) a shearing couple caused by the motion of the Plateau, followed by (b) tension parallel to the margin associated with the spreading mid-ocean ridge segment northeast of the Plateau and (c) tension normal to the margin caused by the removal of lateral support as the Plateau moved away. Maud (1961) has shown that once the faults of coastal Natal were in existence tension normal to the margin could produce the observed structure of 'tilted step-faulted blocks and related horsts and grabens'. Knowing now that a complex stress system existed in the area, it seems likely that the faults themselves were formed by an interaction of stresses rather than simple tension. In the coastal areas of the Transkei, southwest of Natal, the pattern of short west-east trending normal faults can be readily explained in terms of a shearing couple (Freund, 1965), like (a) above. In making any reappraisal of the post-Karoo structural evolution of eastern South Africa the proposed mode of for-
Fig. 44.5. Structural features associated with the evolution of the two continental margins of southern Africa. Apart from the Agulhas Arch, the outcrop of which is shown, and the anticlinal axes of the Cape Fold Belt, all the features are considered to be of post-Middle Jurassic age.

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Tectonic Framework of Sedimentation of the Gondwana of the Eastern Himalayas, India

S. K. ACHARYYA

ABSTRACT

The Gondwana sequence of the Eastern Himalayas occurs mainly as allochthonous sheets along the foothills overriding the Siwalik (Mio-Pliocene) foreland, and also as windows and discontinuous inner thrust sheets within the older metamorphics.

At its base is a diamictite recognisable along nearly all the Himalayan chain. It usually overlies the metamorphics unconformably. Marine (Lower Permian) fossils from the diamictites of the Lesser Himalayas are closely similar to those of the marine ingressions within the Talchir Formation of the Peninsula. There is no conclusive evidence favouring a glacigenic origin for the extra-Peninsular diamictites. They appear to have been deposited as submarine mud-flows in a tectonically active basin, frequently associated with intermediate to basic subaerial volcanism.

The overlying terrestrial sequence of the coal-bearing Damuda Group is well developed, mainly in the Eastern Himalayas. In the window zone of Sikkim the diamictite is conformably overlain by the Damuda. Elsewhere, especially along the foothills, the diamictite invariably overrides the immediately younger Damuda, although contrary to earlier belief the succession within the Damuda is normal.

The extra-Peninsular Damuda of the Darjeeling-Duars area appears to be quite different from the intra-cratonic Gondwana of Peninsular India. On the other hand, the elongated basin shape, immature and arenaceous lithic-fill of local provenance, and the sedimentary organisation, are remarkably similar to those of the Siwalik Group. The extra-Peninsular Damuda possibly represents a piedmont facies, deposited by the fast-aggrading braided streams close to the raised mountains to the north. Close to the eastern end, however, marine or deltaic facies are represented in the truncated sections.

Thus the diamictite with pyroclastics and the overlying Damuda of the Eastern Himalayas possibly represent late-orogenic isopic zones (of intra- to back-deep type) of an earlier geosyncline. Regional metamorphism and tectonism of the geosyncline possibly culminated during the Triassic Period in the Darjeeling-Sikkim Himalayas. These earlier molasse troughs have been subsequently deformed, mobilised and moulded during the Himalayan orogeny (Tertiary phase).

INTRODUCTION

Although the extra-Peninsular Gondwana sequences are extensive and geologically important, they are economically less attractive, more inaccessible, structurally disturbed, and less well known than those of the Peninsula (Fig. 45.1).

The late Palaeozoic (usually Lower Per-
mian) diamictite, representing the base of the Himalayan Gondwana sequence, is a remarkably uniform feature along nearly all the Himalayan chain; but the typical continental facies with Glossopteris is mainly distributed in the window zone of the Eastern Himalayan foothills, in Western Sikkim, and in Kashmir. Along the Eastern Himalayan foothills the Gondwana belt can be traced, with some structural discontinuity, from eastern Nepal (Jacob, 1952), eastwards for more than 800 km to the Brahmaputra near Sadiya (Goggin Brown, 1912).

REGIONAL GEOLOGIC SETTING

The Gondwana sequence of the Eastern Himalayas occurs as superficial nappes overriding the autochthonous Siwalik Group (Mio-Pliocene) along the so-called Main Boundary Fault. It is overridden in turn by the weakly metamorphosed older rocks, namely the pelitic Daling Formation (Late Precambrian or Early Palaeozoic?), or the calc-psammo-pelitic Buxa Group (Palaeozoic?) (Acharyya, 1971a). The Buxa Group usually occurs at a lower tectonic level than the Daling Formation. The low grade metamorphics are overridden in turn by more intensely metamorphosed and migmatised gneissic nappes. These have been differentiated into several laterally persistent interformational nappes, namely the Lingtse Gneiss, Paro Formation and Darjeeling Gneiss in the Darjeeling Sikkim-Bhutan Himalayas (Acharyya, 1971c; Ray, 1971; cf. Nandy et al., 1971; Das et al., 1971). The

Fig. 45.1. Distribution of the extra-Peninsular and Peninsular Gondwana
Paro and Darjeeling are lithologically closely similar to the Buxa and Daling respectively of the outer zone, while the Lingtse Gneiss, at the sole of the gneissic nappe, is essentially a sheared biotite-granite, intersheeted with epidiorites and phyllonites of Daling and Paro type. The gneisses continue to the Great Himalayas, being further intercalated with several types of granitoid rocks and migmatites.

All along the frontal zone the older diamictite overrides the relatively younger continental Gondwana (Fox, 1934; Ghosh, 1956). This, together with the spectacular inverted metamorphism of the overriding rocks of the Lesser Himalayas, simulates a regionally inverted succession (Heim and Gansser, 1939; Gansser, 1964). This popular belief, however, has not been confirmed by study of the sedimentary structures within the continental Gondwana and the Buxa Group from the Darjeeling-Duars area (Acharyya, 1968), and the Kameng area, Arunachal Pradesh (Das et al., 1971; Nandy et al., 1971). Gondwana sediments intersliced and interfolded with the Buxa Group have also been located in the Rangit Valley under the upwarped tectonic cover of the Daling Formation and the gneisses (Ghosh, 1952). The window zone reveals a more complete stratigraphic development than does the intensely dislocated frontal zone.

The Tibetan sedimentary zone, to the north of the ‘Central Crystallines’, overlies the gneisses of the Higher Himalayas with intrusion of sheet-like high level tourmaline granites obliterating the basal portion of the sedimentary succession. The Everest Pelite and Limestone of the Tibetan zone have been correlated with the Buxa and Daling Formations respectively of the foothills, while the base of the Everest Pelite, migmatized by granite intrusion, has been compared with the Darjeeling gneiss (sensu lato) (Wager, 1934, 1939). The Lachi ‘Series’, with a diamictite horizon similar to those of the foothills, conformably overlies the Everest Limestone. Late Permian marine fossils occur both above and at the base of the ‘pebble-bed’ (Muir Wood and Oakley, 1941; Raina and Bhattacharya, 1961, 1962), suggesting that they are younger than those of the outer zone. Recently Lower Permian and Upper Carboniferous marine fossils have also been located below the Lachi ‘pebble-bed’ (P. R.

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**GEOLOGICAL SETTING OF THE AREA OF STUDY**

- **Siwalik Group**
- **Extra Peninsular Gondwana**
- **Extra Peninsular Gondwana Undifferentiated**
- **Buxa & Daling Metamorphites**
- **Gneisses**
- **Tourmaline Granite**
- **Pre-Permian Sediments**
- **Permian & Post-Permian Sediments**
- **Area of Study in Details**

**SCALE**

- 50 km
- 100 km

**LEGEND**

- **Thrust**
- **Antiform**
- **Regional Strike**
- **Synform**

**Fig. 45.2. Geological setting of the area of study (modified after Gansser, 1964)**
Chandra and C. R. Sen, pers. comm., N. Sikkim Expedition). The Tibetan sequence is thus essentially marine, and continental Gondwanaland is not represented in the northern face of the Himalayas. However, indefinite plant remains were reported from the gritty sandstone overlying the Lachi and underlying the Tso-Lhamo 'Series' with a Middle Triassic marine fauna (Auden, 1935).

STRATIGRAPHY AND SEDIMENTATION

The diamictite formation of the Lesser Eastern Himalayas has been formally named the Rangit Pebble-Slate by Acharyya (1971b) after the Rangit Valley, Sikkim, where it was first identified and mapped (Ghosh, 1952). The coal-bearing terrestrial succession with Lower Gondwana plant fossils has been correlated with the Damuda Group of the peninsula. In the window zone of Sikkim, the Damuda conformably overlies the Rangit Pebble-Slate which in turn overlies the Buxa and Daling Formations unconformably. In some rare cases, however, as in the Rangit River section near Tatapani, the uppermost unit of the stromatolitic Buxa dolostone appears to be conformably overlain by the Rangit with a zone of slate-limestone intercalations (Acharyya, 1971b). The upper and lower contacts of the Rangit and Damuda are, however, always tectonic in the frontal zone. A prominent unconformity has also been reported at the base of Blaini Formation (diamictite) of the Western Himalayas (Bhargava, 1971, 1972; Srikantia and Sharma, 1971).

The mode of deposition of the various Gondwana units is analysed below.

Rangit Pebble-Slate

The Rangit Pebble-Slate, like other diamictites of the Himalayas, is correlated with the Talchir Boulder Bed, and has been ascribed a glacial origin without much hesitation. These correlations of the Himalayan diamictites are not only based on lithology but also on occasional marine fossils (Reed, 1932; Jacob and Banerjee, 1954; Sahni and Srivastava, 1956; Raina et al., 1971; Ganeshan, 1971, 1972; Acharyya, 1972a; Chandra, 1972).

Most of the Himalayan diamictites are intercalated with pyroclastics (Acharyya, 1972b), which are especially well developed with agglomeratic traps and flows around Kashmir and Arunachal Pradesh (N.E.F.A.), being linked with the Panjal and the Abor Volcanics. Agglomeratic traps and tuffs are intercalated with, or immediately overlie, the diamictites of the Kameng and Subansiri districts, Arunachal Pradesh (Banerjee, 1956; Laskar, 1956). Volcanic clasts, rhyolitic tuffs and glassy breccia are present at places within the Rangit Pebble-Slate of the Darjeeling Hills (Acharyya, 1971c, 1972a), and intraformational amygdaloidal basaltic flows have been reported from the same formation from Sikkim (Sinha Roy, 1973). Rhyolite or tuff pebbles were also collected by Wager from the 'pebble-bed' of Lachi in the Tibetan zone (Auden, 1935).

The popular belief that the Himalayan diamictites are of glacial origin requires a very careful reassessment. The problem is especially complicated because of their location in an orogenic belt (Frakes and Crowell, 1969), and they are deformed and metamorphosed to some extent. Non-glacial tills are known to be associated with volcanic activity at various localities in other parts of the world.

The diamictites of the Himalayas appear to be essentially marine as is shown by (i) their remarkable lateral persistence; (ii) thickness of the order of hundreds of metres; (iii) lithological homogeneity; (iv) presence of greenish-grey, carbonaceous siltstone-slate and impure limestone, with pyritic laminations and pockets; (v) presence of marine fossils, and absence of plant fossils. Glacio-marine and submarine density flow deposits are difficult to distinguish (Schmerhorn and Stanton, 1963; Harland et al., 1966; Frakes and Crowell, 1969), but the following observations from the Rangit Pebble-Slate of Sikkim and the Darjeeling area do not support the glacial hypothesis. Similar information is as yet not available from Eastern Bhutan and Arunachal Pradesh.

1. The clasts of the Rangit Pebble-Slate usually vary in size from 2 mm-10 cm, the maximum being 50 cm in Darjeeling and about 1 m in Sikkim. These are usually
rounded, roundness being crudely related to their size and competence. Some of the clasts appear to be faceted, but the shape is usually foliation and joint controlled. Tectonic striations on the faceted clasts have not yet been reported.

2. The clasts are usually fresh, but weathered and altered feldspathic and meta-argillaceous clasts are also present.

3. Pebbles and grits are locally absent within beds inter-stratified with typical pebbly or gritty slates.

4. The clasts are all local, derived from the underlying and adjacent exposures or intraformational sedimentary or volcanic clasts. Clasts of distant origin are absent.

The great lateral extent of the Himalayan diamictites indirectly supports glacial action (Harland et al., 1966; Casshyap, 1969), but in the light of the associated volcanism and nature of the clasts described above, it cannot be considered unequivocal evidence (Acharyya, 1972b).

Rare occurrences of slump structures and soft sediment deformation, the presence of intraformational clasts of various sizes, poorly sorted clasts of local derivation, and the poor stratification within the diamictites of the Darjeeling and Sikkim areas, favour a subaqueous mudflow mechanism of deposition (Crowell, 1957, 1964). The finely-bedded, graded, gritty, calcareous, carbonaceous siltstone, rhythmites and impure limestone, interbedded with massive pebbly or gritty slates, were possibly deposited from small diluted mudflows. Relatively clean feldspathic quartzites and conglomerates possibly indicate reworked and cleaned sediments or shallow marine deposits. The Eurydesma specimens of Sikkim and Darjeeling are located in such quartzose units.

All these features are normally expected in submarine mudflow deposits of a tectonically active trough. The lateral extent of the diamictites requires similar tectonic conditions over the elongated trough at about the same time (Smith, in Schermerhorn et al., 1963). The extensive contemporaneous volcanic activity would not only supply suitable material, but is also indicative of tectonic instability.

**Damuda Group**

The Damuda Group of the Eastern Himalayas consists essentially of coarse-to-medium grained, occasionally pebbly, feldspathic sandstones, with minor finer clastics and impure, persistent coal seams. Lithologically these have usually been compared with the basal unit of the Damuda Group (Barakar Formation) of Peninsular India (Fox, 1934; Krishnan, 1968), or their correlations have been kept open (Dutt, 1971). The floras, though sporadic and scanty, are similar to those of both the Barakar (Lower Permian) and Raniganj (Upper Permian) Formations of Peninsular India.

The truncated section of the Damuda in the main belt of the Darjeeling foothills contains younger forms of *Glossopteris* like *G. conspicua*; *Gangamopteris* is typically absent. Earlier identified forms of *Gangamopteris* (Acharyya, 1969) are now considered to be *Glossopteris communis* var. *stenoneura*. A few miosporal analyses indicate a broad correspondence with the Raniganj flora.

*Glossopteris* is usually the dominant form within the Damuda of the window zone, Sikkim, but an older *Glossopteris-Gangamopteris* assemblage is also developed at places (Dutt, 1953-4; Nautiyal et al., 1960-1). *Glossopteris, Noeggerathiopsis* and *Gangamopteris* have also been reported from the Damuda of Kameng district, Arunachal Pradesh (Das et al., 1971; Chandra, 1972). The lower coal-bearing sandstone unit of Kameng is overlain by a non-coal-bearing psammo-pelitic association similar to the middle Damuda of Peninsular India, and the *Glossopteris*-dominant upper unit is missing (Chandra, 1972). In the Subansiri section and farther east, in Arunachal Pradesh, the terrestrial succession gives place to marine rocks as is shown by frequent occurrences of pyritic black shale/slates with minor carbonaceous and coaly streaks, phosphatic and calcareous nodules, calcareous concretions, impure limestone and well-preserved Permian brachiopods, and the absence of plants (Dinner, 1905; Laskar, 1956; Sahni and Srivastava, 1956; Chandra, 1972).

Thus, except for Arunachal Pradesh, where finer clastics are relatively better represented within the Damuda, and pass
Table 45.1. Statistical Parameters of Damuda and Siwalik Sandstone, Darjeeling-Duars Area, Eastern Himalayas

<table>
<thead>
<tr>
<th>Stratigraphic Unit</th>
<th>Mean size</th>
<th>Standard deviation $\sigma_1$ &amp; $\sigma_g$</th>
<th>Skewness $Sk_1$ &amp; $Sk_g$</th>
<th>Kurtosis (KG)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Total variation</td>
<td>Mean size $M_2$</td>
<td>Standard deviation of mean size $s(M_2)$</td>
<td>Total variation</td>
</tr>
<tr>
<td>Damuda Group (Upper Permian)</td>
<td>0.94–3.49 $\phi$</td>
<td>1.88 $\phi$</td>
<td>0.823 $\phi$</td>
<td>0.98–2.2 $\phi$</td>
</tr>
<tr>
<td>Main belt</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Damuda Group, Outer belt</td>
<td>-0.08–3.26 $\phi$</td>
<td>2.06 $\phi$</td>
<td>0.901 $\phi$</td>
<td>0.09–2.00 $\phi$</td>
</tr>
<tr>
<td>emplaced tectonically within</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Siwalik (Upper Miocene?)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Middle Siwalik (Mio-Pliocene)</td>
<td>1.77–3.10 $\phi$</td>
<td>2.67 $\phi$</td>
<td>0.47 $\phi$</td>
<td>1.11 $\phi$</td>
</tr>
<tr>
<td>Lower Sandstone unit</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Middle Siwalik Upper pebbly</td>
<td>0.94–2.12 $\phi$</td>
<td>1.53 $\phi$</td>
<td>0.39 $\phi$</td>
<td>1.50–2.40 $\phi$</td>
</tr>
<tr>
<td>sandstone unit</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Note: Statistical parameters mainly after Folk and Ward (1957) and Inman (1952).*
Turbidites (6i2)

LEGEND

- DAMUDA MAIN BELT
- DAMUDA TECTONICALLY EMPLACED WITHIN LOWER SIWALIK
- MIDDLE SIWALIK (Upper beds)
- MIDDLE SIWALIK (Lower beds)

Fig. 45.3. Distribution of sediments from the Damuda and Siwalik Groups in relation to the environmental fields of Sahu (1964)

As usual, finer clastics are very poorly represented within the Damuda of this sector; and further these are frequently channelled by lenticular and bedded units of coarse sandstone with reworked clasts of finer clastics. The sandstones are invariably poorly to very poorly sorted. The other statistical parameters and statistical variations compare closely with those of the Siwalik (Table 45.1) and fall well within the fluvial field of Friedman (1967) and Moiola and Weiser (1969). In Sahu’s (1964) diagram these plot close to the ‘fluvial-turbidite boundary’ (Fig. 45.3), and thus compare well with the alluvial fan and braided river deposits as observed by Landim and Frakes (1968). Their CM pattern (Fig. 45.4) also compares remarkably well with those of the Siwalik and of alluvial fan and braided river deposits (Bull, 1962, 1964; Williams and
Rust, 1969). These represent essentially rolling bed-load and bottom suspension of channel proximal deposits. The RS segment of Passega’s (1964) master diagram, representing protected channels and overbank deposits, are typically absent.

In modal composition the Damuda sandstones are usually feldspathic greywacke, subarkose and lithic-greywacke (Pettijohn, Potter and Siever, 1972). However, their high matrix content is partly due to squashing of diagenetically altered feldspar and soft lithic clasts. Their rock fragments and heavy mineral assemblages compare well with those of the Siwalik, and match with the Daling and Buxa Formations and their metamorphic counterparts exposed to the north of these belts. Generalised palaeocurrent dispersal also corroborates the dominance of a northern source, and suggests oblique to longitudinal flow towards the southeast to east in the Darjeeling hills. The immature texture, the presence of soft and weathered lithic clasts and intraclasts, sheafs of mica, unstable heavy minerals like kyanites, sillimanite, hornblende, chloritoid etc., and frequent occurrence of quartz grains of low elongation ratio, further prove the nearness of the source areas.

Sedimentary structures of the Damuda and the subjacent Siwalik are also closely similar, being dominated by cross-bedding and irregular bedding contacts. Dominance of medium scale cross-strata of both planar and trough type (Conybeare and Crook, 1968); rapid lateral and vertical variation in flow conditions, as revealed by varying grain size; the type and scale of cross-bedding and bedding; the presence of pseudo-herringbone cross-bedding, cut-and-fill structures and shallow channel units and other associated sedimentary structures; are closely similar to those of braided streams and/or alluvial fan deposits (Doeglas, 1962; Williams and Rust, 1969; Smith, 1970, 1971).

Piedmont and alluvial fan deposits are not normally coal bearing. However, carbonaceous shale and sandstone, and lenticular impure coal seams in the Damuda Group in this sector appear to have been deposited in abandoned channels, interchannel areas and foothill marshes. In their impersistent and variable character these compare very well with the Wealden coal of Lower Saxony and
the Oligocene sub-bituminous coals of the Alps (Teichmuller and Teichmuller, 1968).

**GEOTECTONIC AND PALAEOGEOGRAPHIC SIGNIFICANCE**

Thus the extra-Peninsular Gondwana sequence, especially in the Lesser Himalayas, overlies the metamorphics unconformably. Clastic detritus is exclusively derived from the subjacent and adjacent older rocks. The earliest recognised isoclinal folding and bedding-controlled shears within the metamorphics, responsible for their principal foliation, are also typically absent within the Gondwana sequence.

The diamictite formation of the basal Gondwana sequence appears to be a late-orogenic submarine mud-flow deposited in a laterally persistent and tectonically active trough close to the rising mountain chain of the older metamorphics. Intermediate to basic, essentially subaerial volcanism was active along nearly all the chain especially around Kashmir to the west and Abor to the east. The overlying Damuda Group of the Eastern Himalayas possibly represents fast terrestrial sedimentation on a piedmont, with impersistent coal formation, close to a significantly raised mountain front.

The Lesser Himalayan sequence, being an allochthonous deltaic-to-marine facies of the Damuda equivalents in eastern Arunachal Pradesh, may not indicate an eastward regional palaeoslope, though this cannot be proved. Some land bridges in the eastern Himalayas during the Late Permian possibly favoured migration of some elements of the *Glossopteris* flora across Tethys to China (Kon’no, 1965).

Thus the Lesser Eastern Himalayan Gondwana sequences, with typical volcanic association and late-orogenic sedimentation, possibly represent intra- to back-deep molasse type sediments of an earlier geosyncline (Aubouin, 1965). The pre-Permian metamorphics of the Darjeeling-Sikkim area have also been relegated to several eugeosynclinal isopic zones of the parent geosyncline (Acharyya, 1971a, 1972c; Ray, 1971).

The Permian orogenesis, however, may not be equally recorded everywhere. Locally, post-orogenic emergence of the metamorphics may be followed by subsidence, and thus the diamictite may be overlain by marine shelf sediments. The succession at Krol in the Western Himalayas is of this type.

It is of interest to note that the Lesser Eastern Himalayan Gondwana sequence, and in rare cases the Daling and Buxa rocks, have been intruded by several types of mica-lamprophyres nearly identical to those of the eastern Gondwana basins of Peninsular India (Acharyya, 1969). On the other hand there is apparently no contemporaneous volcanism in Peninsular India. The mica-lamprophyres are believed to be genetically linked to, and broadly coeval with or slightly older than, the Rajmahal Traps of Late Jurassic to Early Cretaceous age (dated at 100-105 m.y. by McDougall et al., 1970). Thus the lamprophyres are post-tectonic and are more closely related to the peninsular shield. They possibly post-dated cratonisation of the extra-Peninsular Gondwana to some extent.

In the Darjeeling-Duars foothills, the outer autochthonous belt of the homoclinal Siwalik (Mio-Pliocene) succession is overridden by the frontal nappe of the main belt of the Damuda Group, backed by the overriding sheets of Rangit Pebble-Slate. These are in turn overridden by the piles of nappes of the older metamorphics, i.e. the Buxa, Daling and the gneisses (Acharyya, 1971c). In the Darjeeling Sub-Himalayas, several outer slices of unmetamorphosed Damuda have been tectonically emplaced within the dicotyledon-leaf-bearing para-autochthonous frontal wedge of the Lower Siwalik (Upper Miocene?). Both lithologically and in microspore assemblage, the outer slices of Damuda are closely comparable to those of the main belt. Thus the dynamic metamorphism of the main belt of Damuda must be mainly the result of the Tertiary orogeny. Regional metamorphism of the nappes of the inner belt is continuous with that of the extra-Peninsular Gondwana. Metamorphism and granitisation in the upper tectonic levels are broadly synkinematic with respect to the major shearing (Himalayan phase) and ultimately even outlast post-thrust east-west folding at places. Late-stage movement and metamorphism of the Lesser Himalayan nappes...
have been reflected in the Siwalik sedimentation, which becomes progressively coarser, texturally and compositionally immature, and records increasing influence of high grade metamorphic and granitoid rocks in the provenance, possibly due to advancing fronts of the gneissic nappes (Acharyya, 1971d). Late- to post-tectonic tourmaline granite pebbles are also present within the upper beds of the Siwalik. Thus the Siwalik represents fore-deep molasse of the Himalayan orogeny proper.

The effects of pre-Himalayan deformation and metamorphism have largely been obscured by later deformation and metamorphism. However, metamorphosed clasts of the Lesser Himalayan rocks and various granitoid and gneissic rocks within the extra-Peninsular Gondwana, and the pre-Gondwana tectonic grain within the older metamorphics, corroborate an earlier pre-Permian phase of regional metamorphism. Presence of non-foliated porphyritic-biotite-granite clasts, similar to the Lingtse Gneiss, within the Rangit Pebble-Slate, is especially noteworthy as these might be linked with the earlier phase metamorphism.

The closing of this earlier metamorphism has been further corroborated by four consistent K/Ar total rock ages from the augen gneisses of Paro Formation from the inner zone of Sikkim (178-213 m.y., Hamrabaev, pers. comm., minimum age due to possible argon loss). On the other hand, all K/Ar ages done so far from the Eastern Himalayas only record the last phase of Himalayan metamorphism closing around Early Oligocene to Early Pliocene (9-4 m.y., Krummenacher and Siegner, in Gansser, 1964; Wager, 1965; Eremenko and Datta, 1969; Acharyya, 1972c). Thus the earlier metamorphism is broadly contemporaneous with the molasse-type extra-Peninsular Gondwana of Permian age, whereas the later phase is broadly contemporaneous with the Siwalik of Mio-Pliocene age.

Exposed sections of the Lesser and the Great Eastern Himalayas are essentially nappes of the earlier geosyncline, overriding the Siwalik autochthon along their frontal termination. However, younger post-Permian geosynclinal sediments are well represented again in the Tibetan Himalayas. Thus it is conceivable that many of the sediments of the Mesozoic-Cainozoic Himalayan geosyncline of the Lesser and the Great Himalayas might be concealed below the telescoped nappes of the earlier geosynclines. Consideration should thus be given to the palinspastic reconstruction of the extra-Peninsular Gondwana sequence. Finally, the metamorphics of the Himalayas, though traditionally linked with Peninsular India, are possibly derived from the far north. Their relation to those of the trans-Himalayas merits critical analysis.

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Tectonic Control of Lower Gondwana Sedimentation in Peninsular India

M. K. ROY CHOWDHURY, B. LASKAR and N. D. MITRA

ABSTRACT

The Gondwana basins of Peninsular India have in general a morphotectonic 'half graben' configuration. An analysis of the tectonic framework of a few basins at the Auranga, Tatapani and Hasdo-Arand Coalfields located in the Koel, Son and Mahanadi Valleys, demonstrates that movement along the boundary faults has induced rapid accumulation of fanglomerates along the flank of rising uplands at the beginning of Damuda sedimentation. An epeirogenic pulse initiated the deposition of coal measures in these basins. As sedimentation progressed recurrent uplift along the boundary fault created periodically high relief in the source area and the repetitive units of pebbly sandstones were deposited.

On the other hand, sediment in the Ramgarh Basin of the Damodar Valley belt indicates that the boundary fault exerted little or no control on the pattern of clastic dispersal in the initial stage of Damuda sedimentation. However, during the later phase the loci of sedimentation were partly controlled by uplift along the boundary fault, which is manifested by the pattern of facies variation in Ramgarh and Raniganj Coalfields. Thus the Lower Gondwana coal measures were deposited under varying tectono-sedimentologic regimes in different basin belts.

STRUCTURAL ELEMENTS OF THE GONDWANA BASINS

The Lower Gondwana basins of Peninsular India (Ghosh and Mitra, 1970) have in general a morphotectonic 'half graben' configuration, one side of which is delineated by a pronounced boundary fault running parallel to the structural trend of the Precambrian rocks. The other margins of the basins are defined by a normal sedimentary contact. The boundary faults sometimes follow old shear zones within the Precambrian basement and thus faulting is assumed to be due to rejuvenation of movement along pre-existing zones of weakness (Chaterji and Ghosh, 1967). The dip of the boundary faults measured in numerous outcrops and in some boreholes varies between 50°-80°, with a mode of about 60°. Although the boundary faults are the most important structural elements of the Gondwana basins, inside the basin numerous tilted segments were formed by a network of intrabasinal faults. Tilting and differential subsidence of the blocks produced by interbasinal faults have resulted in variations in regional dip and strike of the coal measures. The present structural pattern is partly a result of original depositional configuration and partly a result of subsequent deformation, though the post-depositional faulting has, to a considerable degree, erased the original imprint of synsedimentational tectonism.

Bouguer anomaly maps indicate that grav-
'lows' are associated with the Gondwana basin belts. A gravity study by Qureshy et al. (1968) has revealed 'lows' of about 50 mgals associated with the Godavari and Satpura basin belts. It was further indicated that in the Godavari Valley, the eastern margin of the anomaly coincides with the boundary fault and the anomaly pattern is suggestive of a rift structure in this basin belt. The faults result from crustal tension prior to the deposition of the Gondwana sediments.

INITIATION OF DAMUDA SEDIMENTATION AND TECTONIC RELATIONSHIP

The onset of Damuda sedimentation in the different peninsular basin belts is marked by widespread deposition of coarse clastics. However, marked regional variation in their gross lithology reflects varying tectonic control on the different domains of sedimentation. In the Raniganj, Jharia, Bokaro, Ramgarh and Karanpura Basins in the Damodar Valley, coarse sandstones with rounded to subrounded quartz pebbles form the bulk of the lower member of the coal measures of the Damuda Group. The widespread blanket-form of the basal pebbly sandstone in the Damodar Valley basins can be ascribed to sedimentation from braided streams. The uniform coarseness of this basal unit indicates high relief of the source area with erosion and deposition proceeding rapidly. The palaeocurrent and facies organisation in these basins demonstrates further that the sediments were primarily derived from the northerly uplands; only in the Karanpura Basin was a southern source significant. Hence, the southern boundary faults exercised little control on the early Damuda sedimentation.

This situation contrasts with the deposition of basal Damuda polymictic conglomerates so characteristically developed in coalfields such as Auganga in Koel Valley, Tatapani in Son Valley and Hasdo-Arand in Mahanadi Valley belts. In the Auranga coalfield, a wedge-shaped unit of sandstone, containing abundant clasts of sharply angular local bedrock, is found to be developed along the northern boundary fault, in the Auranga River section. The current vectors in this area also show a fanning out of the current in the southwesterly and southeasterly directions from a northerly source area. Uplift along the east-west boundary fault has induced rapid accumulation of fanglomerate deposits at the foot of rising northerly uplands.

A similar pattern of tectonic control on sedimentation was imposed in the vicinity of the southern boundary fault in the Hasdo-Arand Coalfield of the Mahanadi Valley belt. Fanglomerates with sharp clasts of local country rocks have been noted along the southern boundary fault, suggesting contemporaneous deposition of the sediments derived from the upfaulted area to the south (Choudhury, 1973).

The tectonic control on the initial phase of Damuda sedimentation is well documented in the Tatapani Coalfield where the southern boundary of the basin is delineated by a prominent east-west boundary fault. A wedge-shaped block of fanglomerates occurs in an east-west trending depression along the boundary fault. These fanglomerates grade northerly and northwesterly into relatively finer and more regularly bedded sandstones with pebbly interbeds. Thus the size gradients parallel the direction of transport. It has been inferred from these data that the southern boundary fault was active during the onset of coal measure sedimentation in this basin, and the uplift of the southerly source area and subsidence of the Tatapani Basin were accomplished by the movement along this fault.

These examples of the fanglomeratic deposits developed at the base of the Damuda have been described in some detail to demonstrate that in certain basin belts boundary faults exercised a dominant control on sedimentation even at the embryonic stage of basin evolution. The geometry of the sedimentary bodies and the loci of deposition were controlled by the rate of uplift of the positive area, and the optimum condition for accumulation of thick fanglomerates occurred as the rate of uplift exceeded the rate of down cutting of the trunk stream channel at the mountain front (Bull, 1972). With relative tectonic stability the locus of principal deposition had been shifted from the fan apex to further down slope where coal forming conditions occurred.
TECTONIC CONTROL DURING LATER PHASES OF COAL MEASURE SEDIMENTATION

Direct evidence of tectonic control on sedimentation is not available from most of the peninsular basins. However, an analysis of the pattern of clastic infilling gives indirect evidence of the varying control of tectonism on the different phases of sedimentation. In this regard the Ramgarh Basin of the Damodar Valley coalfields is of special significance (Mitra, 1973), as this is one of those few Gondwana basins which has been surveyed by systematic drilling.

The most important structural feature of the Ramgarh Basin is the southern boundary fault. The half graben structure is similar to that of such other adjoining Damodar Valley basins as Jharia, Raniganj, etc. The coal measures of this basin exhibit a crudely cyclic sequence of sandstones, shales and coal seams. Basinal variation in gross lithofacies is exhibited by the sandstone-shale ratio and isopach maps of the different members of the coal measures.

The basal unit of the Barakar Coal Measures of the basin is a 90-140 m thick very coarse-grained sandstone with interbeds of subrounded pebbly units. Similar types of pebbly sandstones also characterise the basal unit of the coal measures in the adjoining Karanpura, Bokaro and Jharia Basins. The isopach map of this unit in Ramgarh (Fig. 46.1) shows that it forms a blanket type of deposit without appreciable regional varia-

Fig. 46.1. Isopach map of Lower Barakar pebbly sandstone unit of Ramgarh Basin
tion in thickness. Both the geometry of the pebbly sandstone unit and its lithology present a sharp contrast to that of the fanglomerase units described earlier. In fact the pebbly sandstone is relatively thinner in the vicinity of the boundary fault and reaches a maximum thickness in the central part of the basin. Evidently the boundary fault exerted little or no control on sedimentation at this phase of basin infilling.

As sedimentation progressed, two persistent seams, VII top and VII bottom, were deposited. These seams serve as key horizons because of their significant thickness and regional persistence. Therefore, the pattern of facies variation between these two persistent seams is of considerable importance in the interpretation of the tectonic events which were operative during the deposition of the upper members of the coal measures. The isopach (Fig. 46.2) and sandstone-shale ratio maps of the interseam sediments (Fig. 46.3) demonstrate that in the northern part of the basin either the VII top and VII bottom seams merge into a composite seam, or a very thin parting of 2-10 m thickness intervenes. The thickness of the parting increases towards the boundary fault where it reaches a maximum of 40-50 m. The sandstone percentage of the interseam sediments registers a sharp increase near the boundary fault, and an argillaceous facies is developed in the northern part of the basin away from the boundary fault. This pattern results from the

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**Fig. 46.2. Isopach map of VII top—VII bottom interseam sediments of Ramgarh Basin**
Fig. 46.3. Sandstone-shale ratio map of VII top—VII bottom interseam sediments of Ramgarh Basin

deposition of a wedge-shaped arkosic sandstone body between the seams close to the boundary fault in response to the movement along that fault, and the rejuvenation of faulting during the later phase of coal measure sedimentation exercised a distinctive control on the pattern of lithotypes. This further indicates that the basin axis progressively shifted southward with time as the southern boundary fault in this basin belt imposed a tectonic control on sedimentation. In the Raniganj Coalfield of this basin belt, a similar shift of the depositional axis towards the southern boundary fault with time is well documented (Ghosh and Mitra, 1970).

CONCLUSION

The deposition of a thick pile of fluvial sediments in the different Lower Gondwana basins implies regional subsidence with sedimentation. The Auranga, Tatapani and Hasdo-Arand Coalfields, which have been studied in greater detail than the others, had rapid accumulations of fanglomerate close to the contemporaneously upfaulted borderlands during the initiation of Damuda sedimentation. Regional subsidence in these basins was controlled by the movement along the boundary faults, and this subsidence initiated the Damuda sedimentation. As Barakar sedimentation progressed in these basins, recurrent uplift along the boundary faults produced repetitive units of coarse sandstones and conglomerates in their vicinities.

In other peninsular basins, such as those
located in the Damodar Valley, boundary faults imposed little or no control during the initial phase of Barakar sedimentation when a blanket of pebbly sandstone was deposited. These sediments were derived from somewhat distant northerly sources which were progressively uplifted in response to regional epeiric movement, and thus the pebbly sandstones acquired a uniform textural character during transport. During the later phase of Barakar sedimentation the loci of deposition were, however, partly controlled by uplift along the boundary faults as is shown by the pattern of facies in the Ramgarh and Raniganj Basins.

It should be emphasised, therefore, that the Lower Gondwana coal measures, developed in widely separated peninsular basin belts, were deposited under a variety of tectono-sedimentologic regimes. The observations made in a few of these provide a basis for, and show the need of, further analysis before the Gondwana tectonics of the Peninsula are understood.

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The eastern margin of Gondwanaland was established by Early Triassic time when the Australian part was stabilised and injected with granite intrusions in the New England and Kubor orogenic domains. After this, until the Late Triassic, magmatism was mainly restricted to areas of tectonism along the borderlands of the New Zealand 'Geosyncline' and was dominantly calcalkaline.

In the Late Triassic/Early Jurassic in eastern Australia, basic continental magmas began to increase relative to calcalkaline activity. The basic magmas initially were undersaturated to alkaline types and probably marked sporadic subcrustal melting associated with increasing tensional fracturing along the Pacific edge, before impending marginal faulting and orogenic uplift in the New Zealand Orogen. The basic magmatism reached a peak by the Middle Jurassic with widespread sub-crustal melting and intrusion of tholeiitic dykes and sills in Tasmania and lesser tholeiitic activity on the Australian mainland. Overall, the basic magmatism in the Australian Mesozoic shows a regional distribution of basalt type, resembling patterns found in the later Cainozoic basic volcanism, but on a continental scale related to the complete Gondwana structure.

In the late Mesozoic some basic magmatism continued, largely along new lines of rifting that split the Australian block from eastern Gondwanaland. Contemporary shoshonitic and non-orogenic calcalkaline activity along the eastern Australian edge represented magma generation related to stabilisation and/or rifting of the margin of the New Zealand Orogen, which was folded during the Rangitata Orogeny. In Queensland, the shoshonitic activity graded into more widespread granitic intrusion and calcalkaline volcanism, which probably reflected closer proximity to the New Zealand Orogen, where calcalkaline and basic activity has continued sporadically.

With the onset of significant opening of eastern Gondwanaland in the Late Cretaceous-Early Eocene the shoshonitic/calcalkaline magmatism ceased along the Australian margin and basic volcanism became prevalent. Some of this volcanism accompanied the initial Tasman Sea and Australia-Antarctic opening, but the bulk of it took place after the later spreading was well established. Two main peaks appear in the volcanism (Oligocene-Miocene and Pliocene-Recent) and tholeiitic volcanism is mainly confined to these.

**INTRODUCTION**

The continental blocks in the Australasian region largely represent separated fragments of Gondwanaland. The Australian block had become stabilised by the early Mesozoic, following folding and granitic injection into the New England Orogen, but its Pacific boundaries were sites of marine sedimentation.
along the New Zealand-Papuan ‘Geosyncline’ (reconstruction of Griffiths, 1971a, 1971b, 1972a, 1972b; and others).

After the Cambrian and prior to the Mesozoic break-up, magmatism in that part of Gondwanaland was concentrated along the Pacific margin during several cycles of stabilisation (Brown, et al., 1968). Calcalkaline magmas typical of unstable to stabilising tectonic environments were prominent. With break-up and separation of the individual blocks, magmas were generated along new divergent zones as well as under long active zones. In this broader tectonic environment stable continental (basic and alkali) and oceanic (basic to ultramafic) magmas assumed greater prominence. In the Australian block, change from dominantly calcalkaline to continental basic magmatism reflected a long-term transition into an essentially tensional tectonic environment, after the block had been removed from the Pacific edge by sea-floor spreading.

The Pacific margin, main lines of rifting, and sea-floor spreading have dominated post-Palaeozoic magmatism in the Australasian region. As a result, igneous rocks of this age in Australia occupy marginal positions, concentrating largely along the east coast, and generally extend less than 250 km and not more than 500 km inland.

This paper reviews and synthesises the magmatic evolution in that part of Gondwanaland (loosely termed easternmost Gondwanaland in the text), now dispersed in the Australasian region. Within this area, emphasis is placed on the details of igneous events in the Australian block, which are related in more general terms to the magmatism developed in the surrounding areas. The

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Fig. 47.1. Distribution of dated post-orogenic Middle-Uppper Triassic igneous rocks in eastern Australia (the dominant calcalkaline phase; 225-210 m.y.). Calcalkaline suites (c) and basaltic rocks (b). The stippled areas represent slightly older granitic batholiths (Early Triassic) injected into the New England (bottom) and Kubor (New Guinea, top) orogenic domains. The broadly hatched line represents the approximate position of the Pacific continental margin of Mesozoic Gondwanaland.
igneous distribution and magma types developed in the Australian block, during its late-Gondwana history are summarised in Figures 47-1-3.

**Late Gondwana Magmatism**

Three phases of magmatic activity can be discerned in Australian Gondwanaland from the time of its stabilisation to break-up, each phase becoming successively but transitionally more basic in character.

**The Dominant Calcalkaline Phase**

Stabilisation of eastern Australia continued after the Hunter-Bowen Orogeny (Late Permian-Early Triassic), with injection of post-orogenic calcalkaline magma into the New England/Yarrol and New Guinea Kubor orogenic domains (Geol. Soc. Aust., 1971), until at least Middle Triassic times (215-220 m.y.; Webb and McDougall, 1967, 1968; Harding, 1969). In Queensland late-stage granites associated with block faulting are flanked by volcanic sequences of similar age (Paine, 1969; Stevens, 1969) in the Esk Graben (andesites with minor trachytes and rhyolites), Maryborough Basin (trachyte-rhyolite-andesite), Nambour Basin (andesite-rhyolite-trachytes), Clarence-Morton Basin (basalt-andesite-rhyolite) and around the Auburn Arch (andesite and minor trachyte near Rockhampton, acidic and basic volcanics in the Monto-Biloela district, minor tuffs in the Dawson Range).

This largely calcalkaline magmatism was restricted to areas of tectonism along the continental margin. In Australia it was associated with local epeirogeny and some weak folding confined to the southern part of the Clarence-Morton Basin, while in the New Zealand depositional belt many of the volcanic rocks were submarine (Brown et al., 1968; Fleming, 1970).

**The Increasing Basic Phase**

In latest Triassic to Early Jurassic times, the frequency of basic continental magmas increased in eastern Australia. This probably reflects increasing tensional stresses at the continental edge prior to more major uplifts associated with marginal faulting of eastern Gondwanaland and impending orogenesis in the New Zealand Geosyncline (Vogt and Conolly, 1971; Griffiths, 1971a, 1971b, 1972a).

The basic magmas ranged from strongly undersaturated alkaline to saturated types with felsic associates. The rocks include the Somerset Dam layered aluminous tholeiitic intrusion (c. 210 m.y., Webb and McDougall, 1967; Mathieson, 1967), basic plugs around Sydney (Late Triassic, Hamilton, 1971), Murrumburrah feldspathoidal monchiquite (194 m.y., Wellman, et al., 1970), Delegate basic pipes (194 m.y., Lovering and Richards, 1964); Garrawilla alkali basaltic and alkaline volcanics and intrusives (181-193 m.y., Wellman and White, 1969), Mt Gibraltar syenite (178 m.y., Evernden and Richards, 1962), Prospect teschenite (168 m.y., Evernden and Richards, 1962); Tasmanian tholeiitic dolerites (167 m.y., McDougall, 1961); Gingenbullen quartz dolerite (c. 160 m.y., Boesen et al., 1961; Early Jurassic, Branagan, 1969) and Towallum basalt (Early Jurassic, Brown et al., 1968). Apart from eastern Australia, isolated dolerite intrusions in the Canning Basin are also dated to this period (196 m.y., Harding, 1969) and may relate to tholeiitic dolerite sills in other Western Australian sequences (Prider, 1969). This phase is transitional from subordinate basic activity during the earlier Triassic calcalkaline activity [e.g. the Undola basanite, interpreted as an interbedded flow in the Lower Triassic Sydney Basin sediments (Vallance et al., 1969) and two gabbroic stocks in the Connors Arch, Queensland, dated at approximately 220 m.y. (Webb and McDougall, 1968)]. Many of the more undersaturated magmas of the main basic phase were generated in the mantle (experimental duplication of Delegate pipe inclusions suggest accumulates from the magma at pressures up to 14-16 Kb (Irving and Green, 1970)), but some of the more saturated magmas were evolved or fractionated at higher crustal levels (Mathieson, 1967).

Some contemporaneous calcalkaline and shoshonitic igneous activity included the Brisbane Valley porphyrites and equivalent southeastern Queensland minor granites and
tonalites (c. 208 m.y., Webb and McDougall, 1967); the Benambra granite porphyry and quartz syenite and associated trachytic volcanics in Victoria (202-207 m.y., Singleton, 1968); the Cooma shoshonitic syenite in New South Wales (174 m.y., Veijayaratnam in Etheridge and Irving, 1972); and fresh to intermediate volcanics and detritus in the Eromanga, Surat and Maryborough Basins (mid-Jurassic, Paine, 1969; Stevens, 1969). Volcanic activity has also been postulated in Tasmania to account for intermediate (?) fragments in the volcanic wackes in the Upper Triassic-Lower Jurassic (?) beds, but detailed supporting evidence has not been produced (Hale in Spry and Banks, 1962). The calcalkaline-shoshonitic component is related to magmatism along the foreland margin of the New Zealand 'Geosyncline' which contributed andesitic to dacitic tuffaceous material and some granitic intrusions (c. 190 m.y.) in contrast to spilitic and basaltic lavas (largely Upper Triassic) in the deeper marine beds (Brown et al., 1966; Fleming, 1970).

**Tasmanian-Antarctic Tholeiites**

Continental basic magmatism reached a peak in southeastern Australia by Middle Jurassic time, with widespread tholeiitic injection in Tasmania (over 5000 km$^3$ of magma, Spry and Banks, 1962). Strong tensional forces at this time were related to incipient rifting of Australia and Antarctica and initial uplifts of the Rangitata Orogeny in New Zealand. The Tasmanian dolerites represent a larger episode and must be viewed against the grand-scale impulses of

![Fig. 47.2. Distribution of dated Upper Triassic-Middle Jurassic igneous rocks in eastern Australia (the increasing basic phase, 210-155 m.y.). Tholeiitic suite (t), alkali basalt-alkaline suite (a), calcalkaline suite (c) and shoshonitic suite (s). The area enclosed by the finely hatchured line represents the approximate limits of the earlier Middle-Upper Triassic magmatism shown in Fig. 47.1.](image)
basic magma generation in Gondwanaland in the Mesozoic, heralding its fragmentation.

The Tasmanian intrusions rose from numerous basement feeders and spread out as transgressive sheets and dykes (chonoliths) through a Carboniferous-Early Jurassic sedimentary blanket, accompanied by considerable block faulting (some tectonic features are summarised in Sutherland, 1966). Recent detailed geophysical work on the dolerite structures at Great Lake (Jones et al., 1966), northern Tasmania (Longman and Leaman, 1971) and Hobart (Leaman, 1972) has located many of the feeders. Around Hobart, Leaman’s work has revealed some thirty dolerite centres spread regularly over the district (2300 km²), with intrusive forms suggesting at least three and possibly more major pulses of intrusion, as well as some minor late-stage pulses. Thus, some spread in the Tasmanian dolerite intrusive ages is possible (McDougall’s 167 m.y. date was made on dolerite assigned by Leaman to one of the later pulses) and may compare with part of the spread observed in eastern Antarctica (155-185 m.y. averages; Adie, 1963; Webb and Warren, 1965; Anderson, 1965; Compston et al., 1968). Whether the Tasmanian ages would span the full Antarctic dolerite spread is uncertain, as superficially the Antarctic data suggest a general decrease in intrusive age with increasing latitude, a scheme into which Tasmania, restored to its Gondwana position, fits reasonably well (Sutherland, 1966). Clearly, much more dating of these dolerites is needed to test the possibility of a southward progression of magmatism. The main Tasmanian tholeiites appear to terminate on the Bass Strait margin to mainland Australia, which may mark an important structural feature at the time of intrusion (Sutherland, 1973a).

Differences between Tasmanian and Antarctic tholeiites are relatively minor. No volcanic ‘facies’ such as that described from Antarctica (e.g. Gunn and Warren, 1962) has yet been recognised in Tasmania, but may well have existed as the Tasmanian Jurassic surface is now stripped. Chemically, some of the Antarctic rocks are olivine tholeiites, types more magnesian than any yet identified in the Tasmanian dolerites. Otherwise, the close match, particularly in high initial Sr²⁸/Sr²⁶ (0.710-0.713) and U/K, Th/U, K/Rb ratios, leave little doubt that the magmas of the two areas were genetically related (Heier et al. 1965; Compston et al. 1968). They form a distinctive province from the tholeitic magma intruded further west in the South African-South American sector of Gondwanaland and were produced either from an unusual mantle or by a process involving widespread crustal contamination.

GONDWANA RIFTING MAGMATISM

The rifting is thought to date from the Late Jurassic into the Middle Cretaceous, when eastern Gondwanaland split into several segments and allowed accumulation of significant sedimentation within the rifts, prior to sea-floor spreading movements (Fig. 47.3). Basic magmatism was associated with those rifts, particularly the strong rift that propagated west to east and formed the Otway Rift Valley between Australia and Antarctica (Griffiths, 1971a; Johnstone et al., 1973).

Continental Basaltic Magmas

In the eastern Otway and Strzelecki Riffs, rocks interpreted as basaltic volcanics have been penetrated in the Upper Jurassic-Lower Cretaceous sequences (Robertson 1, Casterton 1, Moyne Falls 1, Woolsthorpe 1, Pretty Hill 1, Hawkesdale 1 and Duck Bay 1 Wells), reaching a maximum thickness of 600 m in the Hawkesdale Well, but some may be doleritic intrusives (well completion subsidy reports, Griffiths, 1971a; Hocking, 1972). Isotopic dating of rocks in the Casterton Well (c. 153 m.y.; Cundill, in Parking et al., 1969) suggests that this episode is slightly younger than all or most of the Tasmanian dolerite intrusions, but is contemporaneous with some of the Antarctic Victoria Land dolerites and basalts (cf. Compston et al., 1968). Some petrography on the Otway Rift rocks (Hocking, 1970) shows tholeiitic and olivine basaltic magma types comparable with those in Antarctica, but detailed studies will be required to demonstrate that they are an extension of the Tasmanian-Antarctic province. Further west and adjacent to the rift, a tholeiitic flow on Kangaroo Island,
South Australia, is now assigned to the mid-Mesozoic (Wellman, 1971) and may be related to the rift magmatism. Chemically, however, this rock appears distinct from known Tasmanian-Antarctic tholeiites, being generally more aluminous and less magnesian (Joplin, 1963: 198, 200-1; cf. Compston et al., 1968).

Tholeiitic basalts interbedded with Lower Cretaceous sediments are also known in the Bunbury Trough, Western Australia (Prider, 1969; Johnstone et al., 1973), lying in a rift analogous to the Otway structure, but striking north-south and related to rifting of the Indian-South African sectors of Gondwanaland. In bulk chemistry (Joplin, 1963: 230), these lavas fall within the range of both the Tasmanian-Antarctic and the more western Gondwana tholeiitic provinces, but they require detailed work to determine their precise affinities. Basaltic rocks of similar age are also known elsewhere in the Western Australian region in the Ashmore Reef sequence and possibly as tuff in the Broome Beds (Vevers, 1971). The unusual lamproites of the Fitzroy Basin have also been placed in this period on Rb-Sr dating (Prider, 1969), but are now considered to represent a much younger event and are discussed later.

The rifting formed a locus for the late tholeiitic activity, but if the overall pattern of Mesozoic basic continental magmatism in eastern Gondwanaland is examined, dominant regional magma associations can be distinguished (e.g. the southeastern Gondwana tholeiitic association and the eastern Australian alkali basalt association). Similar basic magma associations can be recognised in the Cainozoic of Australia (Sutherland, 1969), but the Mesozoic pattern is contin-
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ental in magnitude and related to the complete Gondwana structure, rather than micro-continental and related to a Gondwana fragment.

Mesozoic magmatism in easternmost Gondwanaland may represent a long-term thermo-tectonic event, such as the Karroo Volcanic Cycle postulated by Cox (1972) as initiating disruption of that part of Gondwanaland. This applies particularly to the tholeiitic rocks of the steady state phase, but there are significant differences between the two cycles, which probably reflect the influence of the Pacific margin.

**Shoshonitic-Calcalkaline Magmas**

In contrast to the basic magmatism associated with new rift zones, calcalkaline magmas continued to generate along the Pacific margin. In Australia these group into two associations, a more sporadic association of largely shoshonitic affinities extending from Tasmania into Queensland, and a more voluminous but regionally localised calcalkaline association in Queensland (Sutherland, 1973b). Rocks of the shoshonitic association form dyke swarms and small intrusive complexes, some containing volcanics, and range from absarokitic to acid potassic compositions. Isotopically dated examples (Ewerden and Richards, 1962; McDougall and Leggo, 1965; Sutherland and Corbett, 1973) include the Noosa Heads potassic diorites (137-144 m.y.), King Island biotite lamprophyre (137 m.y.), Cape Portland potassic appinitic complex (c. 91-103 m.y.), Cygnet syenitic complex (98 m.y.) and Mt Dromedary monzonitic complex (94 m.y.). These magmas probably evolved at depth under the relatively stabilised margin to the New Zealand orogenic belt. Their chemistry and distribution suggest that they originated as separate pockets of magma unrelated to differentiation from any basaltic parent (Joquin et al., 1972).

The shoshonitic association passes transitionally into the more strongly calcalkaline association north from the trachyandesitic volcanics in the Maryborough Basin (Lower Cretaceous, Grahams Creek Formation, Paine, 1969; Stevens, 1969). The calcalkaline rocks lie near a region of reactivated Jurassic downwarping that formed the Eromanga Basin (Paine, 1972) and may reflect a closer proximity to the New Zealand Orogen (Sutherland, 1973b). They occupy two main belts: granitic batholithic bodies (dated about 125 m.y.) intruded in the north of the Connors Arch structure; and a suite of rhyolitic, dacitic and andesitic volcanics and high-level granitic and quartz syenitic intrusives (dated between 110-116 m.y.) around Proserpine-Mackay and west of Bowen at Mt Abbott (Webb and McDougall, 1968; Paine, 1969; Paine et al., 1970). Small hypabyssal intrusives of similar age also lie to the west of the main calcalkaline belts in the north Bowen Basin, but range from basic to acidic compositions and are commonly porphyritic types, with some rocks resembling members of the shoshonitic association. Bentonic clays in the Eromanga Basin suggest that this calcalkaline activity may have continued from the Late Jurassic (Paine, 1959).

The fringe of shoshonitic to calcalkaline activity in eastern Australia reflected adjacent uplifts, orogenic movements and magmatic developments in the New Zealand 'Geosyncline' during the Rangitata Orogeny. Griffiths (1972a) would equate the New Zealand events with subduction at the Gondwana margin, beginning in the Middle Jurassic and continuing with westward under-thrusting and consumption of the Pacific plate until the Early Cretaceous. New Zealand's igneous record for this period (Grindley et al., 1959; Thompson and Kermode, 1965a, b, c), shows a large proportion of submarine intermediate, basic and ultramafic rocks (basalts and pillow lavas of the Matakoia and Tangihua Volcanics; keratophyres and spilites of the Mt Camel Volcanics; North Cape ultramafics), as well as calcalkaline volcanism (andesites, rhyolites and dacites of the Mt Somers volcanics; possibly ranging as young as 80 m.y., Hulston and McCabe, 1971), lamprophyric and minor syenitic activity (minimum 107-136 m.y.; Wellman and Cooper, 1971) and batholithic granite injection (mostly 100-120 m.y.; Aronson, 1968). Related activity would also be expected along the now submerged crustal areas lying between New Zealand and Australia (an example may be the rhyolite

**GONDWANA SPREADING MAGMATISM**

Separation of eastern Gondwanaland had commenced with sea-floor spreading off Western Australia by late Neocomian time (about 120 m.y., Johnstone et al., 1973). Further spreading, detaching segments along the Pacific margin, then took place between 60-80 m.y. ago, parting the Campbell Plateau from Antarctica (Pitman et al., 1968) and opening the Tasman Sea (Griffiths and Varne, 1972; Hayes and Ringis, 1973). Finally, significant separation of Australia and Antarctica started about 55 m.y. ago (Weissel and Hayes, 1972). These events altered the established Gondwana magmatism and magmatic patterns within the separated blocks became more individualistic in character. The post-spreading magmatism, although not strictly a Gondwana feature, is considered to be worth outlining briefly to illustrate the transitions and contrasts in subsequent magmatic history of the dispersed blocks inherited from the Gondwana structure.

**Australian Block**

With dissociation from the Pacific margin, calcalkaline magmatism in the Australian block declined, eventually leaving post-Palaeocene continental basic magmatism to continue in the dominantly epeirogenic tectonic setting. Broad patterns of eastern Australian magmatism recurring since the sea-floor spreading began, and based on known isotopic and other dating have been discussed by Sutherland et al. (1973c). Only limited activity is known during the period of Tasman Sea opening, and includes plugs and flows in the Rockhampton, Baralaba and Brisbane areas (60-70 m.y.; Webb et al., 1967; Harding, 1969) and some alkali basaltic rocks in the Gippsland Basin (Hocking, 1972). Some basic activity may also have continued in parts of New South Wales where both Jurassic and Tertiary igneous rocks occur (Vallance et al., 1969). Minor crustal subduction at the Australian margin during the Tasman Sea opening has been postulated (Hayes and Ringis, 1973); this may explain the limited belt of trachytic, andesitic, dacitic and rhyolitic rocks extending from Brisbane to St Lawrence (Malone et al., 1967; Paine, 1969).

Some alkali basalts associated with the Older Basalts of Victoria (Hocking, 1972) and the Barrington Volcano of New South Wales (McDougall and Wilkinson, 1967; Wellman et al., 1969) correspond with the initial Australian-Antarctic opening, possible rifting along the Australian margin of the Tasman Sea (Symonds, 1973), and opening of the Coral Sea (Mutton, 1973) from Palaeocene to early Eocene times, but the bulk of the basic volcanism has occurred since these events. Basic magmas have arisen along the length of eastern Australia, erupting a basaltic cover over 72,000 km². They represent four main magma types: alkali basaltic, more evolved alkaline, tholeiitic and restricted K-rich leucititic associations (Wass and Sutherland, 1973). Strontium isotope measurements on the basalts mostly show Sr⁸⁷/Sr⁸⁶ ratios (0.7034-0.7050) typical of mantle derivation, but some of the tholeiitic and felsic types show higher ratios (up to 0.7103) suggesting some crustal contamination (Compston et al., 1968; Webb and McDougall, 1968; Dasch et al., 1972).

In the most widespread volcanism, which was of Oligocene-Miocene age (Sutherland et al., 1973c), the main shield volcanoes were built and tholeiitic magmas first became prominent among the alkali basaltic continuum (Wass and Sutherland, 1973). The tholeiitic magmas tended to avoid the more massive structural blocks and rose on sedimentary basin margin structures (e.g. Bowen and Clarence-Morton Basins), suggesting that such structural environments provided more favourable sites for tholeiitic genesis. A revival of volcanism over the last five million years has produced the young basaltic fields of Queensland and Victoria, but its cause is uncertain. The complete pattern of Cainozoic volcanism does not directly correlate with any northward migration of the Australian plate, from a sea-floor spreading zone to the south, over any single simple mantle source or plume, but migration over complex or multiple plumes in association with other
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Magmatic mechanisms is feasible (Sutherland et al., 1973c).

The only other magmatic activity ascribed to the Australian Cainozoic is lamproitic activity in the Fitzroy Basin, Western Australia, now considered Miocene (17-21 m.y., Wellman, 1973). These highly potassic magmas were generated in a restricted area during, but remote from, the most widespread basic magmatism, and their distinctive composition is in keeping with their isolated position in an old shield area.

Sea-floor Areas

With separation of the Gondwana elements, magmatism created new oceanic crust between Australia, Antarctica and New Zealand. Sampling of this new crust has mostly been confined to dredge hauls (e.g. Watkins and Gunn, 1971), except for Macquarie Island, which is interpreted by Varne and Rubenach (1972) as fault bounded blocks of an uplifted example of oceanic crust. They suggest that rocks of the island originated from some spreading zone in the Miocene and now represent uplifted oceanic crust. The island consists of lower level amphibolite facies, dolerite dyke swarms, gabbros and serpentinised peridotites, and upper level zeolitic to green schist facies metamorphosed basaltic extrusives; the chemistry of the dolerites and basalts most closely resembles that of oceanic floor rocks.

East Antarctic Block

Magmatism in easternmost Antarctica since detachment by sea-floor spreading has been entirely continental and basic in nature. It is largely confined to the Ross Sea-Belleny Island region and is dated from late Miocene to Recent (Adie, 1963: chap. X; Hamilton, 1972). The suites are alkali basaltic to sodic felsic volcanics.

New Zealand Block

Magmatism in New Zealand and surrounding continental crustal areas, since they separated from Australia and Antarctica, has been mixed in character, with both calcalkaline (acid to basic) and continental basalts erupting (Grindley et al., 1959; Thompson and Kermode, 1965a, b, c). Present calc-alkaline magmatism, confined to the Taupo region, lies on the end of the Tonga-Kermadec trench (Karig, 1970), and possibly on a subduction zone which is probably being prevented from extending southwards by juxtaposition of the New Zealand crustal blocks across the Australian-Pacific plate boundary (Griffiths, in press). Recently extinct and late Cainozoic calcalkaline activity was more widespread and extended south of New Zealand to Solander Island, and thus was probably related to more extensive trench activity than that of the present zone.

The basic continental activity in New Zealand is commonly alkali basaltic or alkali in nature, but tholeiitic rocks are known. Volcanic islands rising from the submerged crustal areas and ocean floors to the south and east are generally alkali basaltic and alkali felsic in nature (e.g. Chatham Islands, Hay et al., 1970; Auckland Islands, Wright 1966, 1967, 1968, 1970; Lord Howe Island, Game, 1970; Tasman Sea guyots, Slater and Goodwin, 1973), but some include tholeiitic rocks (Norfolk Island, Green, 1973).

CONCLUSIONS

The magmatic evolution in the Australasian part of Gondwanaland, since the Middle Triassic, has been dominated by the Pacific margin tectonics and the main lines of subsequent rifting and separation of continental segments. It shows the following phases of development.

1. Calcalkaline activity was predominant along the Pacific margin until the Late Triassic.

2. Ingress of basic continental activity amongst the calcalkaline activity increased, culminating in overwhelming tholeiitic magmatism in the southern region in the Jurassic.

3. There were developments of basic continental activity associated with major rifting delineating the embryonic Australian, Antarctica and New Zealand blocks in the Late Jurassic to Early Cretaceous. These occurred contemporaneously with shoshonitic to calcalkaline magmatism along eastern Australia, which was, however, related to orogenic uplifts along the New Zealand 'Geosyncline'.

...
4. Basic continental activity, predominant in the Australian and eastern Antarctic segments since initiation of sea-floor spreading in the Late Cretaceous, with intervention of new basaltic oceanic crust between the continental blocks, and restriction of calcalkaline activity to the Pacific edge in New Zealand and Melanesia continued.

5. Following separation of the last Gondwana Australian Antarctica remnant, widespread basic magmatism occurred in eastern Australia, probably reflecting contemporaneity of plate boundary adjustments along a considerable length of the southwest Pacific rim from Eocene to mid-Miocene times (Brothers and Blake, 1973). New Guinea, New Caledonia and New Zealand probably lay along the former Tertiary plate boundary which has since migrated north and east by a complex process of sea-floor spreading behind active island arcs extending from New Britain, Solomon Islands, New Hebrides to New Zealand. Since this episode, basic magmatism in the remaining Australian and eastern Antarctic blocks has been restricted to the Queensland, Victorian and Ross Sea areas, but may pass transitionally into the island arc activity (e.g. Torres Strait basalts, Willmott, 1972).

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