## Sample numbers for photographs

<table>
<thead>
<tr>
<th>Sample</th>
<th>FJ 9</th>
<th>FJ 9</th>
<th>FJ 16A-11</th>
<th>FK998-994</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.104</td>
<td>4.4</td>
<td>BA70</td>
<td>FK16A-11</td>
<td></td>
</tr>
<tr>
<td>3.105</td>
<td>4.5</td>
<td>CH 88</td>
<td>FK998-994</td>
<td></td>
</tr>
<tr>
<td>3.118</td>
<td>4.6</td>
<td>CH 88</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.1</td>
<td>4.7</td>
<td>AJ 58</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.2</td>
<td>4.8</td>
<td>AJ 32</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.3</td>
<td>4.9</td>
<td>AO 3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

No samples were retained for other photographed specimens. Note figures 4.7 & 4.8 should read "Harajica Sst Member at section AJ".

## Relationship of sample numbers to stratigraphic units

### DEERING
- AC 23
- DF 59
- DH 2-14
- DJ541

### HERMANNSBURG
- AA 11
- AC 41
- AJ 107-131
- AK 23-39
- AM 1-4

### LJILTERA
- AA 17
- AC 49-60
- AF 32-68
- AJ 137-182
- AK 41

### CORAMINNA
- AQ828
- AR 2
- BP 2-41
- BR620
- BS 11-15
- BU 33-42
- ER 2
- ET 3-10

### BREVER
- AF 73
- AI 2
- AJ 186-187
- AO 3
- AP 10
- BJ280
- GF 12

### UNDANDITA
- AF 83-87
- AJ 196
- BA970
- BF981
- BG986

### Owen Springs
- CP 63
- CR 78

### Polly
- FG 11-23
- FJ 10-9
- FK346-304
- FI 29

### Langra
- FG 2-7
- FJ 21-17

### Horseshoe
- FG 26-28
- FJ 11-23

### Idracowra
- FG 1-FF 26

All sequences listed above apply to sample number order given in appendix 1a.
STRATIGRAPHY AND SEDIMENTOLOGY OF THE UPPER DEVONIAN
PERTNJARA AND FINKE GROUPS, AMADEUS BASIN,
NORTHERN TERRITORY

by

Brian Gordon Jones

A thesis presented to the
Australian National University in
partial fulfilment of the requirements
for the degree of Doctor of Philosophy

Department of Geology,
August 1970
Except where otherwise acknowledged in the text, all observational, experimental and interpretive work summarized in this thesis is solely that of the author.

Brian G. Jones
ABSTRACT

The stratigraphy, sedimentology and petrology of the Upper Devonian to Lower Carboniferous sediments of the Pertnjara and Finke Groups in the Amadeus Basin, Northern Territory, have been studied in an attempt to reconstruct the paleo-environment.

The sediments range from boulder conglomerates to lithic sandstones and calcareous siltstones. The conglomerates are usually massive but may exhibit flat bedding and low angle cross-bedding. The deposits are lensoid and lateral facies changes to pebbly sandstone are common. They were probably deposited by high energy, upper flow regime fluvial systems as a series of coalescing alluvial fans relatively close to an uplifted source area. The sandstone units vary from fine to coarse grained and pebbly. They exhibit cyclic sequences of sedimentary structures which can be attributed to deposition in a fluvial environment, probably from a low sinuosity or braided stream system. The siltstones also show cyclic bedding together with ripple marks, halite pseudomorphs, gypsum, calcite and dolomite. They were probably deposited in a semi-arid, or seasonally arid, playa-like environment. Lateral facies changes between all lithologies have been recorded and the environment probably consisted of alluvial fans near the source area giving way down the paleoslope to progressively finer grained deposits ending up with playa lake siltstones and evaporites.
ACNOWLEDGEMENTS

The author is most grateful to Dr K.A.W. Crook for the suggestion and supervision of this study. His criticisms and discussions have been most helpful in illuminating various problems as they arose.

I would like to thank the many people who have offered advice, valuable discussion and criticism during various stages of this study: special thanks are due to Dr V.A. Gostin and Messrs I.K. Crain, A.R. Jensen and A.T. Wells. Thanks are also due to Miss J. Gilbert-Tomlinson for discussions on paleontology, and to Drs E.M. Kempe and D. Berger for aid in the palynological investigation. I gratefully acknowledge the advice and assistance given by Mr P.J. Thomson on all the statistical problems dealt with in this thesis. Messrs R.M. Hopkins and L.G.G. Pearce of Magellan Petroleum (N.T.) Pty Ltd gave advice and made available company reports which were extremely useful and informative. Special thanks are due to Mr Pearce for very valuable discussions on problems encountered during the field work in 1967.

I would like to thank Professor D.A. Brown of the Geology Department, A.N.U., for the use of all departmental facilities during the three year period of study while I held an Australian National University Research Scholarship. Thanks must also go to the technical staff of the Geology Department, A.N.U., for assistance in many ways. Special
thanks are due to Mrs H. Drury for typing this thesis, and to Mrs D.A. Maxwell and Mrs W.A. Bell for typing tables, appendices, references and for preliminary typing.

I am also indebted to Mr J.N. Casey for the use of facilities at the Bureau of Mineral Resources, Geology and Geophysics, Canberra, during the final stages of preparation and writing of this thesis.

Finally, I take great pleasure in acknowledging the forbearance, encouragement, drafting and secretarial help given by my wife.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Chapter 1 Introduction</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geomorphology</td>
<td>1</td>
</tr>
<tr>
<td>Previous Geological Investigations</td>
<td>4</td>
</tr>
<tr>
<td>Pre-Pertnjara Stratigraphy in the Amadeus Basin</td>
<td>6</td>
</tr>
<tr>
<td>Lithology of the Pre-Pertnjara Sequence</td>
<td>9</td>
</tr>
<tr>
<td>Regional Mapping</td>
<td>11</td>
</tr>
<tr>
<td>Outcrops of the Pertnjara and Finke Groups</td>
<td>20</td>
</tr>
<tr>
<td>Sections Measured</td>
<td>21</td>
</tr>
<tr>
<td></td>
<td>23</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Chapter 2 Stratigraphy</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pertnjara Group</td>
<td>25</td>
</tr>
<tr>
<td>Parke Siltstone</td>
<td>28</td>
</tr>
<tr>
<td>Deering Siltstone Member</td>
<td>30</td>
</tr>
<tr>
<td>Harajica Sandstone Member</td>
<td>33</td>
</tr>
<tr>
<td>Dare Siltstone Member</td>
<td>36</td>
</tr>
<tr>
<td>Amulda Member</td>
<td>39</td>
</tr>
<tr>
<td>Hermannsburg Sandstone</td>
<td>41</td>
</tr>
<tr>
<td>Ooraminna Sandstone Member</td>
<td>45</td>
</tr>
<tr>
<td>Owen Springs Sandstone Member</td>
<td>47</td>
</tr>
<tr>
<td>Ljiltera Member</td>
<td>49</td>
</tr>
<tr>
<td>Brewer Conglomerate</td>
<td>51</td>
</tr>
<tr>
<td>Undandita Member</td>
<td>56</td>
</tr>
<tr>
<td>Finke Group</td>
<td>58</td>
</tr>
<tr>
<td>Polly Conglomerate</td>
<td>60</td>
</tr>
<tr>
<td>Langra Formation</td>
<td>62</td>
</tr>
<tr>
<td>Engoordina Siltstone Member</td>
<td>65</td>
</tr>
<tr>
<td>Horseshoe Bend Shale</td>
<td>66</td>
</tr>
<tr>
<td>Idracowra Sandstone</td>
<td>69</td>
</tr>
<tr>
<td>Hakea &quot;Formation&quot;</td>
<td>71</td>
</tr>
<tr>
<td>Discussion on Finke Group Nomenclature and Distribution</td>
<td>72</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Chapter 3 Sedimentary Features</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentary Features of the Siltstone Units</td>
<td>75</td>
</tr>
<tr>
<td>Bedding Characteristics</td>
<td>76</td>
</tr>
<tr>
<td>Ripple Marks</td>
<td>77</td>
</tr>
<tr>
<td>Section</td>
<td>Page</td>
</tr>
<tr>
<td>-------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>Hydrodynamic Sole Markings</td>
<td>83</td>
</tr>
<tr>
<td>Rheotropic Structures</td>
<td>84</td>
</tr>
<tr>
<td>Depositional Structures</td>
<td>88</td>
</tr>
<tr>
<td><strong>SEDIMENTARY FEATURES OF THE SANDSTONE UNITS</strong></td>
<td></td>
</tr>
<tr>
<td>Cross-stratification</td>
<td>91</td>
</tr>
<tr>
<td>Plane Bedding</td>
<td>104</td>
</tr>
<tr>
<td>Massive Bedding</td>
<td>105</td>
</tr>
<tr>
<td>Erosional Surfaces and Channels</td>
<td>106</td>
</tr>
<tr>
<td>Ripple Marks</td>
<td>108</td>
</tr>
<tr>
<td>Internal Hydrodynamic Structures</td>
<td>113</td>
</tr>
<tr>
<td>Hydrodynamic Sole Markings</td>
<td>115</td>
</tr>
<tr>
<td>Rheotropic Structures</td>
<td>117</td>
</tr>
<tr>
<td>Included Clasts</td>
<td>120</td>
</tr>
<tr>
<td>Tectonic Structures</td>
<td>125</td>
</tr>
<tr>
<td>Diagenetic Features</td>
<td>127</td>
</tr>
<tr>
<td><strong>SEDIMENTARY FEATURES OF THE BREWER AND POLLY CONGLOMERATES</strong></td>
<td></td>
</tr>
<tr>
<td>Bedding Features</td>
<td>128</td>
</tr>
<tr>
<td>Sedimentary Breccia</td>
<td>129</td>
</tr>
<tr>
<td>Internal Fabric</td>
<td>130</td>
</tr>
<tr>
<td><strong>ASSOCIATION AND VERTICAL PROFILE ANALYSIS</strong></td>
<td></td>
</tr>
<tr>
<td>Association Analysis</td>
<td>135</td>
</tr>
<tr>
<td>Vertical Profile Analysis</td>
<td>141</td>
</tr>
<tr>
<td><strong>FLUVIAL ENVIRONMENTS</strong></td>
<td></td>
</tr>
<tr>
<td>Fluvial Models</td>
<td>167</td>
</tr>
<tr>
<td>Proximal Analysis of Fluvial Deposits</td>
<td>179</td>
</tr>
<tr>
<td>Lacustrine Model</td>
<td>198</td>
</tr>
<tr>
<td>Deltaic Model</td>
<td>203</td>
</tr>
<tr>
<td><strong>PALEOCURRENT ANALYSIS</strong></td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER 4</strong></td>
<td></td>
</tr>
<tr>
<td>4.1 PETROLOGY OF THE SANDSTONES</td>
<td></td>
</tr>
<tr>
<td>Quartz</td>
<td>213</td>
</tr>
<tr>
<td>Rock Fragments</td>
<td>215</td>
</tr>
<tr>
<td>Feldspar</td>
<td>225</td>
</tr>
<tr>
<td>Micas</td>
<td>232</td>
</tr>
<tr>
<td>Opaque and Accessory Minerals</td>
<td>235</td>
</tr>
<tr>
<td>Authigenic Quartz</td>
<td>237</td>
</tr>
</tbody>
</table>
(Chapter 4 cont.)

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate Cement</td>
<td>239</td>
</tr>
<tr>
<td>Iron and Silica Matrices</td>
<td>241</td>
</tr>
<tr>
<td>Detrital Matrix</td>
<td>241</td>
</tr>
<tr>
<td>Void</td>
<td>242</td>
</tr>
<tr>
<td>Texture</td>
<td>243</td>
</tr>
<tr>
<td>Iron Oxide Staining</td>
<td>246</td>
</tr>
<tr>
<td>Variability of Mineral Suites</td>
<td>252</td>
</tr>
</tbody>
</table>

4.2 PETROLOGY OF THE FINE GRAINED ROCKS       | 263  |
| Analytical Procedures                        | 264  |
| Semi Quantitative Clay Mineral Analysis      | 270  |
| Mineral Identification                       | 271  |
| Non-clay Minerals                            | 275  |
| Interpretation                               | 277  |

4.3 HEAVY MINERALS                            | 287  |
| Heavy Mineral Analysis                       | 288  |
| X-Ray Analysis                               | 289  |
| Results                                      | 290  |

4.4 GRAIN SIZE ANALYSIS                      | 307  |
| Theoretical Considerations                   | 313  |
| Determination of Sieve Size Parameters       | 315  |
| from Thin Section Data                       |      |
| Sieve Size Analysis                          | 316  |
| Thin Section Size Analysis                   | 317  |
| Regression Equations                         | 320  |
| Relationship between Length and Breadth of  | 325  |
| Grains                                       |      |
| Comparison with Friedman's Equation          | 327  |
| Grain Size Correlation                       | 331  |
| Tests for Goodness of Fit                    | 331  |
| Discussion                                   | 333  |
| Grain Size Parameters                        | 337  |

4.5 CLASSIFICATORY ANALYSIS OF PETROGRAPHIC  | 348  |
| DATA                                         |      |
| R-Mode Factor Analysis                       | 350  |
| Q-Mode Factor Analysis                       | 359  |
| Cluster Analysis                             | 366  |
CHAPTER 5

STRUCTURE

The Rodingan Movement
The Pertnjara Movement
The Henbury Movement
Unnamed Movements
The Alice Springs Orogeny

THE RELATIONSHIP OF SEDIMENTATION TO INDIVIDUAL STRUCTURES

Western Amadeus Basin
The Carmichael-Deering Creek Structures
Tyler Pass-Carmichael Area
Goyder Pass Diapir
The MacDonnell Range Homocline
Waterhouse Range
Ooraminna Anticline
Ross River Syncline
The Gardiner-Tyler Uplift
Gosses Bluff
Gardiner Range
The Illamurta Structure
The Seymour Range Structure
Larrier Bore
Black Hill Range

RELATIONSHIP BETWEEN THE SEDIMENTARY UNITS

VOLUME CONSIDERATIONS

PALEONTOLOGY

Placoderms
Age of the Placoderms
Correlations Based on Placoderms
Palynology
Paleoecology

PALEOLATITUDE

ENVIRONMENT OF DEPOSITION

COMPARISON WITH OTHER AREAS OF DEVONIAN DEPOSITION IN AUSTRALIA

REFERENCES

APPENDICES, TABLES AND FIGURES ARE CONTAINED IN VOLUME II.
CHAPTER I

INTRODUCTION

The Amadeus Basin occupies a large area in the southern part of the Northern Territory and adjacent Western Australia (Fig. 1.1) between latitudes 23°30'S and 26°00'S and longitudes 127°30'E and 136°00'E. Eastwards beneath the Great Artesian Basin its extent has not been delineated. Thus the Amadeus Basin is elongated in an east-west direction for about 725 km (450 miles) and has a maximum north-south extension of 280 km (160 miles). The sedimentary succession occupies an area of approximately 155,000 km² (60,000 sq. miles) while the combined outcrop and subsurface area of the Pertnjara and Finke Groups, excluding the area under the Great Artesian Basin, is only about 40,000 km² (15,200 sq. miles).

The locations of Alice Springs, homesteads and important geographic features are given in Figure 1, 2. Access to most of the area studied is reasonably easy on graded roads and station tracks. In the northern and western areas access to sections in the Pertnjara Group was facilitated by recently cut seismic lines. Even where tracks do not exist, the upper portions of most Pertnjara and Finke outcrops can be reached with reasonable ease over the alluvial and sand plains.
The Amadeus Basin lies within the arid zone of central Australia. The mean annual rainfall varies from 255 mm in the north-west, to less than 125 mm in the south-east, with most of the area receiving between 150 and 250 mm. The average monthly rainfall is greater in summer than in winter and it usually occurs in thunderstorms or heavy downpours. It is very irregular and because of the high potential free-water evaporation (about 230 cm at Alice Springs - Slatyer, 1962) there are frequent periods of drought. There is a marked seasonal (17-30°C) and diurnal (23-28°C) temperature range and ground temperatures may reach 50°C in summer. At Alice Springs the mean monthly maximum temperature in summer is 35°C compared with a mean minimum of 22°C. In winter the means are 19.5°C and 4.5°C respectively and frosts are fairly common especially in sheltered areas.

With the exception of river banks where Eucalypts are quite common, the vegetative cover in the area largely consists of sparse grasslands and shrublands with spinifex and acacias being the most prominent forms. However, quite large tracts are hilly areas with abundant rock outcrops and very sparse vegetation. These, together with the spinifex covered sand dune fields and the salt pans, are virtually useless and undeveloped as pastoral areas. The main pastoral areas consist of saltbush and bluebush pastures on undulating to hilly terrain, and the short grass and forb pastures on alluvial pans, flats and parts of the sandy plains. Pastoral leases are large (average 2600 km²) and carry an average of 2 to 3 head of cattle per square kilometre of usable pasture land,
or 7 head per square kilometre of usable country within 5 kilometres of water (Perry et al., 1962).

An unusual characteristic for an arid region is the wide extent of bleached, sandy and red-earth soils, and the relatively minor extent of calcareous, gypseous, saline and alkaline soils that are normally expected in such a zone.

The physiography of the Amadeus Basin area has been studied by Perry et al. (1962), who subdivided the area into a large number of landform systems. Six generalized physiographic divisions can be recognised:— (1) Parallel strike ridges and valleys of the hilly areas, such as the MacDonnell, Gardiner and Waterhouse Ranges. The valley floors are covered by alluvium and the ridges are cut by transverse drainage. (2) Rounded hills with a dendritic drainage pattern, characteristic of areas underlain by the Brewer Conglomerate. (3) Mesas and low hills surrounded by alluvial and sand plains, common in the west and south. (4) Low relief sandy plains with occasional low rocky outcrops. Most of the plain is covered by irregular, or occasionally linear, sand dunes while the perimeter consists of alluvial sediments. (5) Flat areas such as the Simpson Desert with north-north-west aligned straight parallel sief dunes. (6) Flat salt lake playas which are the focus of internal drainage systems.
Geomorphological studies of the Alice Springs district have been carried out in detail by Mabbutt (1962, 1965, 1966 and 1967) and Bremer (1967).

An outline of the geomorphic features relevant to this study is given below.

Evidence of the earliest recognisable land surface is provided by the crest levels in the MacDonnell and Gardiner Ranges. A reconstruction of such a land surface would show sloping plains in the south passing north-westwards into an area of subdued ridges and vales, and then into the high rounded summits of the Arunta Block (Mabbutt, 1967). This phase is largely pre-Tertiary and may have extended back into the Paleozoic. The north-south drainage pattern of the Finke and other major rivers is probably due to a combination of inheritance and limited superimposition on this ancient surface (Mabbutt, 1966). Downwarping in the south-east probably instigated a further long cycle of erosion which preceded and accompanied the period of Tertiary planation and deposition. The upland areas of the Arunta block and western MacDonnell Ranges were weathered and eroded. Weathering produced secondary silicification of surficial horizons, overall iron staining and deposition of iron oxides in joint crevices, and the kaolinization of arkosic sediments. Erosion produced the essential features of the present ridge and vale tract of the western MacDonnell
Range, although with a lower relative relief.

These upland areas are surrounded by plains of lower relief on which silcrete (dominant in the south and south-east) and laterite (dominant to the north on Burt Plain) duricrusts developed during this cycle of erosion. The laterites are pisolitic or vesicular, generally 1.5 to 7.5 m thick, and are typically underlain by mottled and pallid zones. The silcrete is 1.5 to 9 m thick, is usually massive and grey, is underlain by a kaolinized pallid zone up to 30 m thick, and frequently can be seen to transgress local structures i.e. it is related directly to the old land surface. Around Alice Springs both types occur together, although they rarely intermingle and their occurrence may depend upon lithology and relief. Both silcrete and laterite duricrusts are assigned an early Tertiary age (Bremer, 1967) and although no unconformity has been seen between them, their apparent coevality may be a result of the long period of tectonic quiescence.

Later Tertiary downwarping to the south-east in the Lake Eyre Basin caused asymmetrical rejuvenation of erosion and produced the present land surface, where duricrust remnants are frequently seen to form mesas. Increasing aridity at this time truncated and partly disorganized out-going drainage with the consequent development of internal drainage systems such as Lake Amadeus.
After the early explorers had opened up the country the first geologist to visit the Amadeus Basin was Chewings. In 1886 he investigated the source of the Finke River and made some general comment on the geology. A few years later East (1889) commented on the sediments at Horseshoe Bend.

The first important observations on the Pertnjara Group were made by Chewings in 1891. He noted that, in the syncline at old Tempe Downs Station (approximately 20 km west of its present position), there was 600 m of "Devonian" mudstone and "Ooraminna Sandstone" conformably overlain by at least 600 m of sandstone. He noted (1894) that the mudstones consisted of alternating red mudstones and red and green shales. He described the lower sandstone as highly ferruginous and fine-grained, and the upper sandstone as red, friable and coarse grained, sometimes white and feldspathic. He considered that these sandstones and mudstones were not conformable with the Mereenie Sandstone, and that the "Ooraminna Sandstone" was massive and forms steep scarps. Chewings considered this to be an extensive formation in the James, Gardiner and MacDonnell Ranges and further afield. He noted that not far from the MacDonnell Ranges pebbles are found in the sandstone, and that the "puddingstone" conglomerate on the south flank of the MacDonnell is conformable with the red sandstone and they are one and the same formation. In 1894 Chewings named the "Devonian(?)" mudstones and sandstones the "Walker Creek
Series" and stated they are unconformable on the Larapinta Group. He distinguished the "Ooraminna Sandstone" from the "Walker Creek Series" because he was uncertain of the stratigraphic position of the sandstone at Ooraminna. In 1894 Chewings disagreed with Brown (1890) and stated that, even though no unconformity was seen in the Finke Gorge, MacDonnell Range, part of the Walker Creek Series is present there and should be separated from the fossiliferous Larapinta Group.

Tate and Watt (1896) measured 2130 m of post-Ordovician conglomerate at Ellery Creek Gorge and noted that further east the conglomerate rested on conglomeratic sandstones. They also noted that pebbles occurred in the "upper" red sandstone beds of the Krichauff Range and correlated them with the conglomerate across the Missionary Plain. Chewings (1914, 1928) upheld this view and also correlated them with similar sandstones in the ranges to the south.

In 1895 Brown correlated the sandstones at Ooraminna and Horseshoe Bend. Chewings (1914) disagreed with this and considered that the Finke River beds only come as far north as Maryvale. Chewings noted that the Finke River sandstones overlie red and purple shales and pale coloured micaceous sandstones and shales, and these in turn are unconformable on older rocks at Horseshoe Bend. Ward (1925) elaborated on this sequence and noted that the basal beds are coarse, pebbly grits with sandy, argillaceous interbeds. These are
overlain by red, white and greenish grey shales and then sandstone (now Langra Sandstone).

Chewings (1931) named the Pertnjara Series (post Ordovician conglomerate and sandstone) and Madigan (1932b) correlated the "Permo-Carboniferous" Pertnjara Series with the Finke River Series. Madigan (1932a) confined the name Pertnjara Series to the conglomerate, and considered the sandstones in the Krichauff Range belong to the Mereenie Sandstone, which unconformably underlies the conglomerate. Chewings (1935) disregarded his earlier hypothesis and on his map shows the Pertnjara confined to the northern Missionary Plain.

This concludes the early phase of exploratory geology, and it can be seen that the relationships between the rock units proposed by the various writers depends to a large extent on where they looked at sections. Hence Madigan's (1932a) conclusions about the sandstones on Krichauff Range are erroneous because he did not see the Parke Siltstone.

Later geological work on detailed aspects of the Pertnjara and Finke Groups has not been very extensive except for that of Leslie (1965), Pearce (1968) and various reports of the Bureau of Mineral Resources. Oil company geological and geophysical reports have frequently mentioned various aspects of the Pertnjara
Group and their findings will be dealt with in subsequent chapters. Regional mapping of the Amadeus Basin, including the Pertnjara and Finke Groups, by the Bureau of Mineral Resources was completed in 1964. The results of these regional surveys have been published: — Joklik (1955), Prichard and Quinlan (1962), Wells, Forman and Ranford (1964, 1965), Forman (1966), Ranford, Cook and Wells (1966), Wells, Stewart and Skwarko (1966), Forman, Milligan and McCarthy (1967) and Wells, Ranford, Stewart, Cook and Shaw (1967). A regional synthesis summarizing the results from the whole Amadeus Basin has been compiled by Wells, Ranford, Cook and Forman (in press). Apart from these regional surveys several more specialized articles have been compiled by Officers of the Bureau of Mineral Resources and are referred to in later chapters.

**PRE-PERTNJARA STRATIGRAPHY IN THE AMADEUS BASIN**

An outline of the Paleozoic stratigraphic succession in six subregions of Amadeus Basin is presented in Table 1.1 (after Wells et al., in press). Deposition occurred almost continuously from the Precambrian to the Upper Devonian — Lower Carboniferous, with only minor periods of widespread non-deposition and erosion during and following orogenies and tectonic movements. Localised unconformities and disconformities are common especially on the southern and western margins where the Paleozoic units are generally thinner and several were probably not deposited.
The metamorphosed basement rocks of the Arunta and Musgrave-Mann complexes are of Archaean age and unconformably underlie the Amadeus Basin sedimentary sequence.

The Precambrian formations generally have a large areal extent, a moderately uniform distribution of lithologies, and are coeval and may interdigitate with similar formations to the south and west. The Cambrian Pertaoorrrta Group is lithologically more variable and has been subdivided into 15 formations each of a more localized extent. The Group is largely confined to the northern and central part of the Amadeus Basin although its eastward extent has been curtailed by middle and upper Paleozoic erosion. The Ordovician Larapinta Group is composed of five formations of large areal extent and uniform lithology - either sandstone or siltstone. Likewise the ?Silurian-Devonian Mereenie Sandstone underlies a large area and is lithologically uniform.

A brief resume of the lithologies present in the pre-Pertnjara sedimentary sequence and in the underlying basement rocks in the Amadeus Basin area is presented to enable the reader to become acquainted with the lithologies controlling the provenance of the Pertnjara and Finke Groups. Special reference has been made to the occurrence of diagnostic minerals and rock types which may enable their reworked counterparts to be recognized in the Pertnjara and Finke sediments.
LITHOLOGY OF THE PRE-PERTNJARA SEQUENCE

The basement rocks along the northern margin consist of low and moderate grade metaquartzites, schists, gneisses and granites. Low, moderate and high grade metamorphics and granites also occur to the south and these are overlain by the Mt Harris Basalt, Bloods Range Beds and Dixon Range Beds, all of which were also metamorphosed to some extent in the Petermann Ranges Orogeny (Proterozoic - 1200M years; Forman, 1966). There are, of course, a great variety of rock types of varying mineralogy, with granite and gneiss being the most abundant.

The feldspar content of both the gneisses and granites is dominated by alkali feldspars (especially microcline) and lesser quantities of plagioclase (albite to oligoclase, but dominantly in the range An 27-33). In the Arunta Complex the ratio of alkali feldspar to plagioclase is usually about 2:1 (Forman et al., 1967), whereas in the Musgrave-Mann complex the ratio is higher reading 3-10:1 in the Olia Gneiss (Forman, 1966). In a granite near Mt Cavenagh homestead microcline has replaced a good deal of the oligoclase metasomatically (Wells, et al., 1966). Up to 15 percent of apatite has been recorded in isolated magnetite-apatite metaquartzites in the Arunta complex (McCarthy; in Forman et al., 1967). Apart from micas, pyroxenes, amphiboles, tourmaline and zircon which have a fairly widespread occurrence in the basement rocks, the Arunta complex also contains localized areas with relatively common beryl, garnet and kyanite.
The Heavitree and Dean Quartzites overlie the basement unconformably; they are very mature silicified orthoquartzite with minor conglomerate, arkose and siltstone in localized pockets near the base of the formation. The average size of the quartz grains is 1 mm and they are typically well rounded and show strained undulose extinction.

The Bitter Springs Formation and the Pinnyinna Beds consist of fine and some medium grained, tough, interbedded crystalline, dolomitic limestone, crystalline dolomite, dolutite, dolarenite, calcarenite, shale, siltstone and gypsiferous siltstone and sandstone. Minor spilitic basic volcanics occur near the base in the Loves Creek Member of the Bitter Springs Formation. These beds often show incompetent folding due to thick evaporites in the sequence.

The Areyonga Formation contains an immature sequence of boulder clay, conglomerate, arkosic sandstone, poorly sorted siltstones, and dolomite with abundant chert masses and infrequent thin bands of grey crypto-crystalline apatite. The Inindia Beds are similar with siltstones, sandstones, arkosic sandstones, chert, thin dolomites and limestones, and fine angular conglomerate. The sandstones are dominantly medium grained and contain a few phosphate pellets and rare glauconite. The Boord Formation consists of siltstone, calcilutite, calcarenite, dolomitic limestone, sandstone and boulder beds. The quartz sandstones of the Carnegie Formation are usually moderately sorted with 0.1 to 0.5 mm subrounded grains. The arenites contain
between 2 and 20 percent of quartzite and chert fragments, and the formation also contains quartz siltstone and minor shale. The beds of glacial origin in the Areyonga Formation and Inindia Beds indicate that Arunta metamorphics, Heavitree Quartzite and Bitter Springs Limestone, were being eroded from a landmass to the north.

The Pertatataka Formation consists dominantly of green and red-brown siltstone with minor limestones and sandstone. In the sandstone bands, authigenic glauconite occurs in various forms associated with phosphatic grains and opaque minerals (Schmerber, 1966). The siltstone, sandstone and pebbly sandstones of the Winnall Beds have been studied petrologically by Schmerber (1967) from the Erldunda No.1 Oil Well. The siltstones and shales are very chloritic and some bands are micaceous. The sandstones are well to moderately sorted, submature to mature orthoquartzites with angular to subangular grains. They contain rare feldspar and mica, and minor chert, glauconite and phosphate. The Ellis Sandstone is a medium grained, moderately to well sorted, well rounded, kaolinitic quartz sandstone with 15 percent of metaquartzite and chert. The Sir Frederick Conglomerate is composed dominantly of Dean Quartzite and rarer quartz mica schist pebbles. The Maurice Formation consists of medium grained quartz sandstones, quartz greywackes and fine micaceous sandstones and siltstones.
The Mount Currie Conglomerate and the arkose at Ayers Rock are considered to be synorogenic deposits associated with the Petermann Ranges Orogeny (Wells et al., in press). The Mount Currie Conglomerate contains boulders of sandstone at the base, fine grained acid and basic igneous rocks in the middle, and granite and gneiss in the upper parts, i.e., the clasts are occurring in reverse stratigraphic order. The arkose at Ayers Rock contains abundant feldspar and lithic fragments, is very coarse grained and poorly sorted and contains abundant epidote in the matrix (see appendix 2).

The Cambrian Pertaoorrta Group sediments are very varied. The Arumbera Sandstone is largely composed of red-brown, medium to coarse grained, moderately sorted, micaceous, feldspatholithic quartz sandstone with frequent epidote in the matrix. It also contains micaceous siltstone and shale, minor dolomite and is glauconitic in places. The poorly sorted Quandong Conglomerate contains some sandy horizons and numerous pebbles derived from the Areyonga Formation. The Eninta Sandstone is a red-brown, fine to medium grained, moderate to poorly sorted, feldspathic micaceous sandstone with minor siltstone and conglomerate horizons. The partly glauconitic, pink-brown and grey, crystalline dolomites and rare dolomitic fine sandstones and siltstones of the Todd River Dolomite grade laterally westwards into the limestones and dolomites of the Chandler Limestone. The latter is characterized by chert laminae and a foetid odour, and contains some evaporites in the sequence. The Tempe Formation is composed of siltstones,
dolomite and fine to medium grained, glauconitic sandstones associated with a few, thin, pelletal phosphorite horizons. The Cleland Sandstone consists of medium to coarse grained, subrounded to angular, ferruginous feldspathic sandstones, and fewer pebbly sandstones with pebbles of vein quartz, quartzite and chert. The Illara and Petermann Sandstones are fine to medium grained, micaceous sandstone with minor siltstone, and they are separated by the micaceous siltstones, shales, and infrequent sandstones of the Deception Formation. In the north the Hugh River Shale is composed of micaceous siltstones and shales with some chert and dolomite bands, and minor sandstones, while further south and west it becomes more sandy with calcareous, silty, fine grained, quartz greywackes. The Jay Creek Limestone contains 30 to 40 percent limestone, 50 to 60 percent micaceous and calcareous shale, and a few thin calcareous sandstone interbeds. The Giles Creek Dolomite is composed of dolomite interbedded with limestone, siltstone and shale. The overlying Shannon Formation consists of siltstone with interbeds of limestone, dolomite and lesser cherts. The Goyder Formation is composed of fine to medium grained, moderately sorted, micaceous, partly calcareous or silty sandstones, micaceous, green and red brown siltstones, dolomite, and limestone which is most abundant near the base. In Alice No.1 Well the fine sandstones contain quite abundant glauconite and collophane. The presence of metamorphic fragments and garnet in the upper part of the Pertaorrrta Group, indicates that basement rocks were being eroded in the source area. Orthoclase and microcline are the dominant
feldspars in Pertaoorrrta Group.

A summary of clay mineral data for the Proterozoic and Cambrian formations was given by Schmerber (1967). The Proterozoic units have a very monotonous clay mineral assemblage dominated by illite and chlorite, with only rare and localized occurrences of montmorillonite, kaolinite and mixed layer clays. A large proportion of the chlorite in the samples is probably diagenetic and the kaolinite, which is usually confined to sandstones near the surface, is probably secondary. The clay mineral assemblage in the Cambrian formations is slightly more varied but is still dominated by illite and chlorite. Montmorillonitic clays are more abundant, especially in the siltstones and shales eg. Goyder Formation. Mixed layer clays and kaolinite still only occur in very minor quantities.

The largely Ordovician Larapinta Group is composed of five alternating formations of marine sandstones and siltstones, some of which contain well preserved fossils.

The fine to coarse grained, well rounded and moderately well sorted, quartzose Pacoota Sandstone contains frequent pebbly sandstones and a few interbeds of micaceous and possibly kaolinitic siltstone. The well rounded pebbles (up to 10 cm) are predominantly vein quartz and siliceous sandstone. Most of the sandstones are supermature ortho-quartzite with 30 to 60 percent straight extinction quartz, 30 to 60 percent undulose quartz, 10 to 30 percent composite quartz, and rare chert. Occasionally microcline is present in very small amounts, and accessory minerals are well
rounded tourmaline and zircon. The cement is usually siliceous, and clays are probably kaolinite and illite with minor chlorite and montmorillonite. One notable 6 m thick bed in the Gardiner and Idirriki Ranges contains up to 50 percent glauconite in a granular or intergranular form. The formation also contains rare, thin bands of pelletal and nodular phosphorite.

The Horn Valley Siltstone is composed predominantly of siltstones with occasional limestones and rare sandstones. Calcareous sediments become more abundant to the south and consist largely of fossil fragments with rare sandy-biosparites or biosparudites. The siltstones are grey green to pale brown (black in subsurface) and are occasionally calcareous or sandy. The clay minerals are probably illite and kaolinite with minor chlorite. The fine grained sandstones are moderately well sorted, poorly rounded and contain dominantly straight extinction quartz, with minor undulose quartz, and rare composite quartz. The formation is slightly phosphatic throughout and contains minor quantities of glauconite.

The Stairway Sandstone was studied in detail by Cook (1966) and he subdivided it into three units. The Lower Stairway is a fine to very coarse grained sandstone, pebbly in places, with well rounded and sorted quartz grains. In places near the base there are concentrations of up to 20 percent pyrite ooliths. The Middle Stairway contains a varied lithology predominantly of sandy and
micaceous lutites interbedded with thin, very fine grained sandstones and pelletal and nodular phosphorites. The latter are in bands 3 to 20 cm thick, have a varied form but are typically in "structureless" or "sandy" pellets and generally consist of crypto-crystalline apatite. Carbonates (micrites) and calcareous siltstones become more common to the south-east. The Upper Stairway is composed dominantly of fine to very fine grained arenites with minor lutites and phosphorites. Most of the sandstones are super-mature orthoquartzite with 95 to 99 percent quartz. The quartz is typically straight extinction, with only small percentages of undulose, and rare composite, quartz grains. The lower Stairway sandstones are often bimodal with both modes well sorted while the upper unit is dominantly fine grained and contains slightly more feldspar and chert. In the Middle Stairway the infrequent sandstones are very fine grained, submature to immature orthoquartzite. Feldspars, where present in the sandstones, are dominantly microcline. Heavy minerals are limited to well rounded opaques, tourmaline and zircon. The lutites of the Middle and Upper Stairway are mainly orthoquartzitic siltstones with slightly more feldspar than in the associated sandstones. Clay mineral suites are predominated by illite, with minor kaolinite and chlorite.

The Stokes Siltstone is a poorly exposed, fairly uniform sequence of grey and green siltstones and claystones, with minor thin biomicritic limestones near the base and a few thin, fine grained, silty and calcareous
sandstone interbeds near the top. The siltstones contain a high proportion of very fine grained quartz and the clay minerals are dominantly illite with minor chlorite. The straight extinction, and rare undulose, quartz grains in the sandstones are usually moderately rounded and sorted. Thin pelletal phosphorites occur near the base of the formation and they are similar to those of the Stairway Sandstone.

The Carmichael Sandstone consists of red-brown, silty sandstones and siltstones. The sandstones are moderately to poorly sorted and rounded; becoming more poorly sorted and pebbly to the south. They are generally very fine to medium grained, immature or submature orthoquartzites with simple straight extinction quartz dominant over undulose quartz. The sandstones also contain minor feldspar (microcline) and composite quartz, and rare metaquartzite and chert. The dominant clay mineral in the siltstone is kaolinite.

The ?Silurian to ?Middle Devonian Mereenie Sandstone is generally a fine grained, mature or supermature orthoquartzite with well rounded and well sorted grains. Many of the quartz grains are frosted and pitted and they fall into the following categories:— straight extinction (40-50%), undulose extinction (40-50%), composite (5-20%), metaquartzite (1-2%), and chert (0-2%). The sandstones very rarely have up to one percent feldspar (mainly microcline) and the only heavy minerals seen were
well rounded grains of tourmaline, zircon and opaque minerals. The matrix is largely composed of secondary silica. Lutites, where present, are probably composed of kaolinite.

REGIONAL MAPPING

Regional mapping of the Amadeus Basin by Officers of the Bureau of Mineral Resources was completed in 1964. The following 1:250,000 Sheets contain Pertnjara or Finke Group outcrops or subcrops and have been used in the regional synthesis:- SF52/15, Mt Rennie; SF52/16, Mt Liebig; SF53/13, Hermannsburg; SF53/14, Alice Springs; SG52/3, Bloods Range; SG52/4, Lake Amadeus; SG52/8, Ayers Rock; SG53/1, Henbury; SG53/2, Rodinga; SG53/3, Hale River; SG53/5, Kulgera; SG53/6, Finke; and SG53/7, McDills.

In the regional mapping the Pertnjara Group outcrops were delineated and the Group was subdivided into three regionally mappable units. Likewise, the Finke Group outcrops were mapped but the fourfold subdivision of this Group on the latest maps (2nd Preliminary Edition, 1964; 1st Edition, 1968) is questionable. The author considers that the 1st Preliminary Edition (1964) of the Finke Sheet gives an interpretation which is much more closely aligned to the subdivision of the Finke Group based on the type sections of the constituent formations.
No mapping on a regional scale was undertaken by the author who used the approximately 1:46,500 scale aerial photographs on which the Bureau of Mineral Resources did the original field mapping. Formational boundaries were checked at each section studied by the author and these agreed with the Bureau of Mineral Resources' subdivisions in almost all cases. Exceptions were noted only on a few local structural anomalies studied by the author.

OUTCROPS OF THE PERTNJARA AND FINKE GROUPS

The Pertnjara Group outcrops on the outer flanks of almost all the dominantly east-west trending anticlinal ranges south of Alice Springs from the Mt Rennie Sheet in the west to the Hale River Sheet in the east and the Kulgera Sheet in the south (Fig.1.3).

The Parke Siltstone outcrops only south and west of Glen Helen where it forms strike valleys with generally very poor exposures. South of the James Range the Parke Siltstone is more extensive and appears to extend as far east as Maryvale. Its westward extension beyond the Cleland Hills is doubtful.

The Hermannsburg Sandstone is the most extensively outcropping unit of the Pertnjara Group. In the west it conformably overlies the Park Siltstone whereas in the east is conformably or disconformably overlies the Mereenie Sandstone. On the Henbury and Hermannsburg Sheets it reaches its maximum exposed thickness of approximately
The Hermannsburg Sandstone forms extensive, low (200 m) ranges with a characteristic steeply dissected topography. Outcrops are generally good in river valleys but tend to be poor and very weathered elsewhere.

The Brewer Conglomerate is confined to the MacDonnell Range on the northern margin of the Amadeus Basin from Mereenie Bluff to Alice Springs with the exception of smaller, locally derived occurrences exposed at Williams Bore, Larrier Bore and in northwest Hale River Sheet area. Except in major river valleys outcrops are generally poor and consist of weathered lag gravels which may have suffered reworking. A typical dendritic drainage pattern is developed on the conglomerate due to its massive nature and lack of planes of weakness.

Outcrops of the Finke Group are generally poorer than those of the Pernjara Group because of the generally low relief, low amplitude folding, and alluvial and sand cover. The sandstone units are frequently displayed in weathered mesas but the best outcrops occur on the flanks of the Black Hill Range. The Polly Conglomerate and Langra Formation crop out in the south-western portion of the Finke Sheet whereas the Horseshoe Bend Shale and Idracowra Sandstone are confined to the north and east of the sheet.
SECTIONS MEASURED

Figure 1.4 gives the location of all measured sections and logs studied by the author and includes the sections measured by B.M.R. and various oil companies, especially French Petroleum Company of Australia and Magellan Petroleum (N.T.) Pty Ltd. The alphabetic labels for each section are arranged in sequence from west to east (i.e., AA to AU) and from north to south (i.e., AA to GA). Localities do not conform strictly to this system and represent all localities from the Finke Group.

In this study sections were measured at approximately 16 km to 20 km intervals over most of the area of Pertnjara Group outcrop and from selected localities in the Finke Group. Their location depended upon conditions of access and suitability of river valley exposures. Most sections were measured in 1.5 m increments using a Jacob's rule (abney level and measuring rod). This is the most practical method of section measuring, when alone, where the dip of the strata does not exceed 25° to 30°. Where steeper dips were encountered sections were generally measured vertically (using a Jacob's rule) on the side of the valley and then projecting the section down dip, or by measuring horizontally or at low angles using a tape and abney. These methods of section measuring were also used by previous workers in the area and were considered to be the most satisfactory. A total of 83 sections were measured to provide formational characteristics and thicknesses, and from 25 of these sections sedimentological logs were prepared.
Representative samples of all rock types were collected at ¼ to 50 m intervals in the siltstone and sandstone sections, depending upon variability of the rock types present, and the conglomerate samples were generally collected at approximately 300 m intervals depending upon availability of exposures.

SEISMIC LINES

The Brewer and Missionary plains were extensively investigated by geophysical methods between 1965 and 1967. Numerous north-south seismic traverses cross these plains and drill hole cuttings could still be collected from most of the shot points. The shot holes are located at 500 m intervals along the seismic lines and the holes are usually 25 to 35 m deep. Samples were collected from 10 seismic lines at approximately 16 km east-west intervals across the plains to provide a stratigraphical and petrological link between the ranges to the north and south.
Stratigraphic nomenclature in the Amadeus Basin is relatively simple and as yet uncomplicated by many problems of synonymy. This is because relatively few people have worked in the area and most of the names were proposed for distinctive formations or groups and not for their subdivisions in specific localities.

This chapter deals with the nomenclature and definition of the stratigraphic units used throughout this study, together with a general description of each unit. The sedimentology and petrology are discussed in chapters 3 and 4 respectively, while correlations between the various units are not discussed until chapter 5.

PERTNJARA GROUP

The Pertnjara Group sediments consist of siltstone, sandstone and conglomerate and have a maximum preserved thickness of about 3600 m. They are the youngest known sediments of the Precambrian to Paleozoic Amadeus Basin succession.
Early references to the Pertnjara Group have been given in Chapter 1 and are summarized here together with later synonyms:

Chewings, 1891: Red sandstone and laterally equivalent conglomerate; Ooraminna (?Devonian) Sandstone; Devonian mudstone.

Chewings, 1894: Walker Creek Series; Ooraminna Sandstone - at Ooraminna only.

Chewings, 1931: Pertnjara Series.

Madigan, 1932a & Chewings, 1935: Pertnjara Series - conglomerate only; Mareenie Sandstone.


Wells et al., in press: Pertnjara Group.

The name Pertnjara (Chewings, 1931) is derived from the Aranda word for many stones and it refers to the conglomerate forming the undulating foothills of the southern MacDonnell Ranges on the Missionary Plain Syncline. According to the Australian Code of Stratigraphic Nomenclature (1964, section 14) the name Pertnjara is invalid and Chewings' earlier name "Walker Creek Series" should have been retained. The latter name published in 1894 gave the thicknesses and a brief description of all the units in the Series at the type locality in Walker Creek, west of Tempe Downs homestead, together with their extent and the lateral equivalence to the sandstone-
conglomerate sequence on the southern MacDonnells. However, Chewings himself renamed at least part of the series "Pertnjara Series" and this name has become established in all the subsequent literature. Thus to reinstate the name "Walker Creek Series" on the basis of priority would cause great confusion in the literature and a departure from the Australian Code must be accepted here.

The Pertnjara Formation was first defined by Prichard and Quinlan (1962) as the sequence of sandstone, quartz greywacke and conglomerate that overlies the Mereenie Sandstone with a regional unconformity in the type area on the southern flanks of the western MacDonnell Ranges.

The Pertnjara was elevated to group status by Wells et al., in press, when he defined its three constituent formations. They are the basal Parke Siltstone, the Hermannsburg Sandstone and the Brewer Conglomerate. The lateral extent of the Pertnjara Group and the distribution of rock types is given in Figure 2.1. All formations are present in the central part of the area, whereas the north-eastern and north-western parts contain only sandstone and conglomerate and the southern areas contain only siltstone and sandstone. The thickness of the preserved Pertnjara sediments (Fig. 2.2) is controlled by structure and gradually thickens from the margins towards the centre of the Missionary Plain Syncline. From seismic evidence the deepest parts of the syncline south of Glen Helen contain approximately 3600 m of Pertnjara sediments. The relationship of the Pertnjara
Group to the underlying Mereenie Sandstone varies from conformable to disconformable over most of the area, with localized areas showing unconformable relationships. The top of the Pertnjara Group is not present. The age of most of the Pertnjara Group is not precisely known although the lower Parke Siltstone has been dated as Late Devonian, probably Famennian. The upper units of the Pertnjara Group are thought to be Carboniferous.

**PARKE SILTSTONE**

The Parke Siltstone was defined by Wells et al., (in press) as the basal formation of the Pertnjara Group. The formation was named from Parke Creek on Dare Plain in the north-east part of the Lake Amadeus Sheet area (Fig. 2.3). The type locality is on the south-west flank of the Mereenie Anticline (section DF, 24°07'S, 131°42'E) and in this area the formation consists of siltstone with minor sandstone interbeds.

The area of Parke Siltstone outcrop and subcrop is shown in Figure 2.4 and the areas of maximum deposition occurred in the central part of the basin. Isopachs of the preserved Parke Siltstone are shown in Figure 2.5, and a map indicating interpreted original thicknesses is given in Figure 2.6. The westernmost outcrops are
relatively thin and occur at Cleland Hills, while the north­
eastward extent of the Parke Siltstone in the Missionary
and Brewer Plains has been tentatively established from
subsurface oil well and water bore data. The formation
extends at least as far south as the Mount Charlotte
Range in the east, and the north-west corner of the
Ayers Rock Sheet in the west.

In outcrop the Parke Siltstone forms a valley
with very poor exposures between the Mereenie Sandstone
and the overlying Hermannsburg Sandstone. In most areas
the Parke Siltstone appears to lie conformably on the
Mereenie Sandstone. However, in the southern part of
the basin the Parke Siltstone transgresses the depositional
edge of the Mereenie Sandstone and locally it unconformably
or disconformably overlies the Carmichael and Stairway
Sandstones of the Larapinta Group.

The contact of the Parke Siltstone and
Hermannsburg Sandstone appears to be conformable except
around some anomalous structures. Wells et al., (in press)
noted that the contact between the two formations was
sharp and suggested that they would not interfinger
laterally. However, a transitional unit has been recognized between the two formations and they have been seen to inter-finger in the Carmichael area and on the western end of the James Range (see Chapter 5).

The general stratigraphic succession of the Parke Siltstone in the measured type section on Dare Plain is given in Figure 2.7. It has been subdivided into four members described in more detail below, namely the basal Deering Siltstone Member, overlain successively by the Harajica Sandstone Member, the Dare Siltstone Member and the Amulda Member.

The age of the Parke Siltstone is known from fossils found in the Harajica Sandstone on Dare Plain and the Deering Hills. It is Late Devonian and probably no older than Late Frasnian. The Dare Siltstone Member and Amulda Member are most likely to be Famennian or slightly younger.

(1) **Deering Siltstone Member**

The Deering Siltstone Member, the lowest
member of the Parke Siltstone, has the most restricted areal distribution (Fig. 2.8), extending from Tyler Pass to Deering Hills in the north and from Areyonga to Mereenie Anticline in the south.

The member takes its name from the Deering Anticline at the western end of Missionary Plain (23°43'S, 131°40'E). It is very poorly exposed on the southern flank of this anticline where it is overlain conformably by the Harajica Sandstone Member.

The type section of the Deering Siltstone Member is in a tributary of Parke Creek on the northern flank of the Mereenie Anticline extending east-north-east from the eastern extremity of Mereenie Sandstone outcrop (24°02'S, 131°38'E; aerial photograph Lake Amadeus Run 1A/5012 10.0 cm E, 7.0 cm N to 12.5 cm E, 7.3 cm N of the southwest corner). The type section is shown in Figure 2.9 and consists largely of alternating laminated and massive light reddish brown siltstones. The basal 1 m is shown conformably overlying the Mereenie Sandstone in Figure 2.10. The siltstone and silty sandstone beds in this interval are all finely laminated and the calcareous medium grained sandstone is weakly cross-bedded and 10 cm thick. Occasional thin limestone or very calcareous siltstone beds are present, especially in the lowest 25 m. They become progressively rarer higher in the sequence. Sandstone beds, generally less than 5 cm thick, occur sporadically and are almost always fine or
very fine grained. The most important element is siltstone which forms over 90 percent of the sequence. In this member laminated siltstone is more abundant than massive siltstone and the latter is usually present in beds up to 60 cm in thickness. In the upper half of the member laminated green siltstones form a minor constituent interbedded with the light reddish brown siltstones, and towards the top occasional fine to medium grained, laminated or cross-laminated, thin sandstone beds are seen. A fragment of fish plate was found by Dr K.A.W. Crook in a thin massive light reddish brown siltstone 38 m above the base of the member. Sedimentary features in this member include ripple marks, hydroplastic deformation structures, small scale cross-bedding, mud cracks, ball and pillow structures and rain prints.

The thickness of the Deering Siltstone Member at the type locality is 93 m. Isopachs are given in Figure 2.11 and the rapid thickening from 0 m at Tyler Pass to 240 m north of the Carmichael structure is obvious. In the Tyler Pass area the basal siltstone is generally coarser grained with frequent, thin, fine sandstones interbedded with the red-brown laminated and massive siltstones. Greenish siltstones are again fairly rare in this area. On the southern flank of the Deering Anticline the Deering Siltstone Member again thickens westwards from 0 m in the east to 100 m in the west. Exposures are very poor in this area. Where visible, the formation is composed of red-brown laminated
to massive siltstone with quite frequent very fine to fine, thin sandstone horizons. On the southern flank of the Gardiner Range the member is once again poorly exposed but here it is also poorly defined because outcrop of the Harajica Sandstone Member is sporadic.

Where visible, the contact between the Deering Creek Member and the underlying Mereenie Sandstone is conformable. On the southern flank of the Deering Anticline this contact is largely obscured but it may well be disconformable. The upper boundary of the Deering Siltstone Member is gradational and is taken at the base of the predominantly sandstone outcrop of the Harajica Sandstone Member.

(2) **Harajica Sandstone Member**

The Harajica Sandstone Member overlies the Deering Siltstone Member and forms the bulk of the Parke Siltstone on the southern flanks of the MacDonnell Ranges. The distribution and isopachs of this member are given in Figures 2.12 and 2.13. The Harajica Sandstone Member can be recognized from Ellery Creek to the Deering Anticline in the north and extends southwards from the western end of the Gardiner Range to the Mereenie Anticline.

The member takes its name from the Harajica water bore 5 km east of the southern end of Stokes Pass (23°36'S, 132°09'E). The type section of the Harajica Sandstone Member is in the north flowing creek 2 km
south-west of the southern end of Stokes Pass (23°37'S, 132°05'E; aerial photograph Hermannsburg Run 10/5123 0.5 cm E, 13.3 cm N to 9.5 cm E, 10 cm N of the south-west corner). The type section is given in Figure 2.14 and consists of a uniform sequence of cross-laminated and flat bedded, fine to coarse grained sandstones with occasional mudclasts and rare subrounded pebbles of quartzose sedimentary rocks. Occasional thin ripple-drift laminated sands and thin sandy siltstone horizons break the monotony. The sandstone beds form poorly defined cycles commencing with the coarsest grained sediments as large-scale cross-stratified sets that frequently overlie a slightly eroded surface. In a complete cycle these give way to finer grained sediments in smaller trough sets which are overlain in turn by flat bedded and ripple-drift bedded sandstones and finally siltstone. Complete cycles are rare and frequently only the cross-stratified section is preserved beneath the succeeding cycle. Parallel and linguoid ripple-marks and mud cracks are occasionally preserved.

The Harajica Sandstone Member becomes progressively thinner, finer and more silty towards the south until it wedges out south of Mereenie Anticline. Outcrops on Dare Plain are thinly bedded and siltstone interbeds make up almost half the sequence (Figs. 2.15 & 2.16). The very fine to medium sandstones are generally laminated and less than 25 cm thick. Where cross-stratification occurs it is usually in tabular
sets although a few shallow troughs were seen. A few of the sandstone beds are lensoid although most, including very thin beds, are of uniform thickness over a wide lateral extent. The sandstones are separated by laminated light reddish brown siltstone. In this area ripple marks and mud cracks are common, and other sedimentary structures include rain prints, ball and pillow structures, slump folding, mudclasts and possible rill structures. Fish plates have been found 55 m above the base of the Harjica Sandstone Member on the northern flank of the Mereenie Anticline.

The thickness of the Harajica Sandstone Member is given in Figure 2.13 and varies from 0 m just east of Ellery Creek (section AM) to 500 m in the Harajica area. The unit thins to the south and is only 80 m at the Mereenie Anticline. The composition also varies, with the sandstones becoming more quartzose eastward and southwards from Harajica Bore.

The Harajica Sandstone Member is usually conformable on the Deering Siltstone Member and the contact is gradational. The only exception is in the Carmichael - Deering Anticline area. Here the Harajica Sandstone Member disconformably overlaps the Deering Siltstone Member and lies unconformably on the Mereenie Sandstone. The upper boundary of the Harajica Sandstone Member is also conformable and is taken as the last major sandstone horizon beneath the overlying siltstone.
(3) **Dare Siltstone Member**

The Dare Siltstone Member conformably overlies the Harajica Sandstone Member where the latter is present. The Dare Siltstone Member has a larger lateral extent than either of the previous members (Fig. 2.17) and extends from south of Alice Springs to the Cleland Hills in the north, and from Maryvale to the north-east corner of the Ayers Rock Sheet in the south. Where the Dare Siltstone Member overlaps the depositional edge of the lower members it rests directly on the Mereenie Sandstone either conformably or slightly disconformably. In the south it transgresses over the Mereenie Sandstone and lies unconformably or disconformably on the Carmichael or Stairway Sandstones.

The member takes its name from Dare Plain at the Mereenie Anticline (24°03'S, 131°40'E). The type section starts from the top of the Harajica Sandstone Member north of East Mereenie No.4 Well and follows Parke Creek and a south flowing tributary of it north to the base of the sandstone bluff (24°01'S, 131°38'E; aerial photograph Lake Amadeus Run 1A/5012 from 11.0 cm E, 9.5 cm N to 12.5 cm E, 13.7 cm N of the south-west corner). The type section (Fig. 2.18) consists of a cyclic succession of laminated and massive siltstones. Cycles are not well developed in the lower part of the member where laminated green siltstones dominate. They frequently show ripple-marks, and occasionally mud-cracks and halite pseudomorphs. The green siltstones are frequently interbedded with light
reddish brown siltstone laminae, and rarely with very fine to fine, thin, calcareous sandstone and thin, silty limestone horizons. From 70 m above the base exposures are good and the cyclic sequence is apparent (Fig. 2.19). The basal member of each cycle is a laminated greenish siltstone up to \( \frac{1}{2} \) m thick which contains abundant ripple marks and halite pseudomorphs. Black and green minerals coating some of these siltstones were determined by X-ray analysis to be manganese oxides and malachite respectively. Above the greenish siltstones there is usually a rapid transition to light reddish brown laminated siltstone, up to 1 m thick, which may contain ripple marks, mudcracks and occasional halite pseudomorphs. The latter are much less abundant than in the underlying green siltstone. The top of these laminated siltstones is transitional into the overlying massive and generally structureless light reddish brown siltstone. The massive siltstones are usually 2 to 5 m thick and the only features they possess are series of calcite filled vugs aligned approximately parallel to the bedding. Some horizons contain several such bands but most only have one or none. Rare instances of greenish mottling occur in the massive beds and very rarely discontinuous thin bands of green siltstone were seen in an otherwise massive bed. The contact between the massive siltstone and the greenish siltstone of the next cycle is nearly always flat and sharp. Thin or partially deposited cycles were noted but no irregular erosion has been seen between any cycles.
Average values for the 70 complete cycles measured in almost continuous exposure in the centre of the member are as follows. The average cycle is 5.25 m thick and consists of 10 percent laminated greenish siltstone, 30 percent laminated light reddish brown siltstone and 55 percent massive light reddish brown siltstone. The thickest single cycle is 10.4 m while the thinnest is 1.2 m.

Near the top of the member, exposures are poor but the siltstone cycles continue with the addition of rare very fine grained silty sandstones in the basal portion of the cycle. Occasional crystals and casts of gypsum were seen in the upper part of the member.

The thickness of the Dare Siltstone Member at the type locality is 505 m and the isopachs of this member are given in Figure 2.20. East and south of a line from Tyler Pass to Gosses Bluff and Mereenie Anticline the lower three members of the Parke Siltstone cannot be distinguished, and from the generally poor exposures the interval, where present, appears to have the characteristics of the Dare Siltstone Member, i.e., cyclic bedding and halite pseudomorphs. In fact the three basal members probably interdigitate near the depositional edge of the Harajica Sandstone Member.

The lower boundary of the Dare Siltstone Member is transitional in all areas studied while the upper boundary in the north is either sharp, and
possibly erosional, or gradational and is gradational in all exposures south of the MacDonnell Range.

(4) Amulda Member

The Amulda Member is a transitional unit between the Parke Siltstone and the overlying Hermannsburg Sandstone (Fig.2.21). It is included within the Parke Siltstone because it contains almost 50 percent siltstone, a feature very atypical of the Hermannsburg Sandstone. The Amulda Member has the same areal distribution as the Dare Siltstone Member although it slightly overlaps the depositional edge of the siltstones.

The Amulda Member takes its name from Amulda Gap 1.5 km north-west of Areyonga Native Settlement (24°02'S, 132°44'E). There the Parke Siltstone is very thin and poorly exposed and consists almost entirely of the Amulda Member. The type section of the Amulda Member (Fig.2.22) is on the lower portion of the eastern bluff of the second major valley east of Areyonga (24°04'S, 132°19'E; aerial photograph Hensbury Run 1/5011 9.5 cm E, 12.0 cm N. of the south-west corner). It consists of interbedded fine to coarse grained white to brown sandstones and sandy, micaceous light reddish brown siltstones. Siltstones are more abundant towards the base where the sandstone bands are fine to medium grained, thin bedded and infrequent. The siltstones are either laminated or massive, with the former predominating higher in the member. They frequently exhibit parallel and linguoid
ripple marks as well as mud-cracks and occasional rain prints. Halite pseudomorphs are absent. The sandstones are generally laminated or cross-laminated in cosets up to \( \frac{1}{2} \) m thick but usually in the range from 5 to 25 cm. All the cross-stratification is tabular except for one or two coarser grained thicker sets of low angle trough cross-stratification beneath flat bedded sands. Minor slump bedding was seen in some of the fine to medium sandstones and occasional mudclasts occur near the base of thicker units. Rare irregular tubular trace fossils were seen in sandstones of this member 8 km west of the type section.

Both the upper and lower boundaries of the Amulda Member are gradational. The thickness of the member at the type section is 135 m while over most of the area of outcrop it varies from 20 to 200 m becoming thinner to the south. In the MacDonnell Range east of Harajica Well the Amulda and Harajica Sandstone Members form almost all the Parke Siltstone. The Amulda Member varies in thickness from 0 m at Tyler Pass and 10 km west of Ellery Creek to 60 m at Glen Helen and 200 m near Mereenie Bluff. The Amulda Member probably forms the bulk of the Parke Siltstone beneath the Missionary and Brewer Plains, as it does in Palm Valley No.1 Well, West Waterhouse No.1 Well and at the western end of James Range.
HERMANNSBURG SANDSTONE

The Hermannsburg Sandstone was defined by Wells et al. (in press) as a red-brown sandstone with minor conglomeratic sandstone and conglomerate. The formation was named after Mount Hermannsburg in the Krichauff Range approximately 6 km south-south-west of Hermannsburg Mission (23°59'S, 132°44'E). The type section is near Stokes Pass in the Western MacDonnell Range (23°37'S, 132°05'E; aerial photograph Hermannsburg Run 10/5123 from 13.0 cm E, 10.7 cm N to 15.7 cm E, 9.2 cm N and continued from 20.3 cm E, 10.1 cm N to 20.8 cm E, 7.7 cm N of the south-west corner). The basal part of the formation is not very well exposed but exposures improve in the upper part. The type section (Fig.2.23) is composed of cross-stratified and minor flat bedded medium to very coarse sandstones. Fine sandstone interbeds are rare and usually flat bedded while sandy siltstone interbeds are very rare. The basal 210 m of the Hermannsburg Sandstone contain very rare to rare, rounded, 1 to 4 cm diameter pebbles of sedimentary rocks, dominantly quartzite. The sediments are predominantly litharenites and sublitharenites and the colour ranges from greenish to brown depending on the state of weathering of the lithic fragments. The basal unit is composed of trough and rare tabular cross-stratified beds with cosets varying from ¼ to 1½ m in thickness (most are ¼-1 m). The pebbles usually occur near the base of large cross-beds as do occasional concentrations of mudclasts. The latter indicate penecontemporaneous erosion of siltstone horizons which are not
preserved in the succession. Some massive and flat bedded sandstones occur interspersed with the cross-stratified deposits. Contacts between sedimentary units are generally poorly exposed but local scoured surfaces were noted beneath many of the trough cross-beds.

Above this basal unit there is an essentially similar succession for a further 275 m. The difference is that no pebbles have been recorded from this interval and the cross-stratified cosets are slightly thicker (average 1 m). The succeeding 460 m are characterized by thick cross-stratified cosets (up to 2 m) containing occasional to common rounded sedimentary pebbles throughout. The pebbles are 2 to 6 cm in diameter, subrounded to rounded, and consist largely of Mereenie Sandstone and Larapinta sediments with occasional representatives from the Pertaoorrta Group. The sandstones are mainly medium to very coarse grained litharenites. Occasional horizons show concentrations of mudclasts up to 10 cm long. Good outcrops in the ridge-forming sandstones in the middle of the upper unit show some cyclicity of bedding. The cycles are of all sizes from 1 to 15 m and the majority of them are incomplete, probably due to erosion prior to the deposition of the next cycle. A general cycle starts with coarse grained, trough cross-stratified lithic sandstones, in cosets up to $1\frac{1}{2}$ m thick resting on a locally scoured erosional base. The basal sandstone may contain occasional pebbles and mudflakes. Overlying this is a variable number of slightly smaller scale trough cross-
stratified cosets of finer grained, more quartz-rich sandstone. These beds usually form the top of the cycle but in a few areas massive or flat bedded, fine to medium grained, white sandstones overlie the cross-bedded units. The upper boundary of each cycle is usually an irregular erosional surface.

The uppermost 100 m of the Hermannsburg Sandstone contains progressively more pebbles and near the top thin conglomerate bands are quite numerous. The pebble size increases up to 8 cm and the contact with the Brewer Conglomerate is transitional. The top of the uppermost sandstone bed marks the boundary.

The Hermannsburg Sandstone overlies the Parke Siltstone conformably at the type section and in most other sections where both units are preserved. Local exceptions occur and an unconformable relationship between the two formations is seen at Ellery Creek. Over the Goyder Pass and Carmichael structures the boundary appears conformable but the Parke Siltstone shows noticeable thinning. On the Carmichael structure the Hermannsburg Sandstone overlaps the depositional edge of the Parke Siltstone and rests unconformably on Mereenie Sandstone and the Larapinta Group. East of Ellery Creek and in the north-eastern Amadeus Basin the Hermannsburg Sandstone transgresses the Parke Siltstone and disconformably or locally unconformably overlies the Mereenie Sandstone. Basal conglomeratic sandstones at Ooraminna
indicate that at least local erosion of the Mereenie Sandstone preceded the deposition of the Hermannsburg Sandstone. An interpreted paleogeologic map of the pre-Hermannsburg surface (Fig. 2.24) shows the distribution of the underlying units.

The contact between the Hermannsburg Sandstone and the overlying Brewer Conglomerate is conformable and transitional in many exposures but intensive local scour in adjacent areas produces slight disconformities. Unconformable relationships are present only east of Ellery Creek in the MacDonnell Ranges and in the northwest Hale River Sheet area.

The Hermannsburg Sandstone is the most widespread unit in the Pertnjara Group (Fig. 2.25) and is the dominant element in most outcrops. It is usually well exposed on the flanks of anticlines and where it overlies the Parke Siltstone it forms prominent bluffs. Complete sections of the Hermannsburg Sandstone are exposed in the northern central Amadeus Basin while over the remaining area the top of the formation has been removed by erosion. Isopachs of the preserved thickness of the Hermannsburg Sandstone are given in Figure 2.26. Figure 2.27 shows the isopachs derived by Laplacian interpolation of measured thicknesses using the computer programme of Crain (1969). The validity of minimum thicknesses determined by this method is in doubt because with a non-random scatter of data there is a tendency to
interpolate to the mean between any two widely spaced
data arrays or points. This accounts for the apparent
thinning of the Hermannsburg Sandstone in the Missionary
Plain Syncline. Interpreted isopachs of original thick-
ness of the Hermannsburg Sandstone are given in Figure 2.28.

(1) Ooraminna Sandstone Member

The Ooraminna Sandstone Member is a new name
proposed for the lower portion of the Hermannsburg Sand-
stone in the Waterhouse-Ooraminna area. This unit was
named the Orange Creek Member by Pearce (1968) in an
unpublished company report for Magellan Petroleum (N.T.)
Pty Ltd. However, the name Orange Creek is preoccupied
by a unit in the Bowen Basin, Queensland, and has been
changed in accordance with the Australian Code of Strati-
graphic Nomenclature. The term "Ooraminna Sandstone" was
used by Chewings (1891) to define the lowest part of the
Hermannsburg Sandstone west of Tempe Downs Station (see
p.6). In 1894 Chewings restricted the name "Ooraminna
Sandstone" to represent the Pertnjara sandstones occurring
at Ooraminna although he gave no formal definition of the
unit. Since 1894 the term has not been used in a formal
sense and thus it can be formally defined here in the sense
of Chewings (1894).

The Ooraminna Sandstone Member of the Hermanns-
burg Sandstone is named after Mount Ooraminna, 50 km south-
south-east of Alice Springs (24°05'S, 134°00'E). The type
section is located on the northern bank of Orange Creek at
the eastern end of Waterhouse Anticline (23°59'S, 133°38'E;
aerial photograph Henbury Run 1/5181 from 8.4 cm N, 14.9 cm E to 7.8 cm N, 15.6 cm E of the south-west corner). The type section is shown in Figure 2.29 and consists of a fairly uniform succession of flat bedded and cross-stratified medium grained quartzose sandstones. Occasional coarser grained trough cross-stratified sets containing scattered rounded pebbles of siliceous white sandstone and quartzite were noted. Fine grained strata are less frequent although occasional siltstones with associated ripple marks occur at 250 m. The beds vary from white to red brown and are characterized by moderate to good sorting and a high quartz content (65-90%). The beds at the type locality and on the outcrops around Ooraminna Anticline resemble reworked Mereenie Sandstone although lateral and vertical interdigation with the more lithic Hermannsburg Sandstone is known.

The contact of the Ooraminna Sandstone Member with the underlying Mereenie Sandstone is disconformable in most areas. The occurrence of pebbles of Mereenie Sandstone in the basal part of Ooraminna Sandstone Member indicates that its deposition was accompanied by contemporaneous erosion of the Mereenie Sandstone, even though there is no marked erosional surface between them.

The distribution of the Ooraminna Sandstone Member is restricted to the northern part of the Amadeus Basin east of Ellery Creek (Fig. 2.30). It also occurs as the major component of the Hermannsburg Sandstone in the
Camel Flat Syncline area. The distribution of this member coincides with the area where the Hermannsburg Sandstone directly overlies the Mereenie Sandstone. The thickness of the Ooraminna Sandstone Member in the type locality is 315 m and isopachs are given in Figure 2.31.

(2) Owen Springs Sandstone Member

The Owen Springs Sandstone Member is a lensoid body occurring in the middle of the Hermannsburg Sandstone section. It consists predominantly of white, cross-bedded, medium to coarse grained sandstone with occasional pebbles. It extends along the flanks of the James and Gardiner Ranges from the Hugh River in the east almost to Bowson Hutt in the west.

The unit is named after the Owen Springs Station (23°52'S, 133°28'E) and the type section is in an unnamed stream on the northern flank of the James Range adjacent to the south-west corner of the Owen Springs Station (24°08'S, 133°10'E, aerial photograph Henbury Run 3/5067 12.6 cm E, 6.7 cm N to 13.0 cm E, 7.7 cm N of the south-west corner).

The type section (Fig.2.32) shows the uniform nature of the member. The base is characterized by a change from reddish, non-pebbly sandstones cross-stratified on a ½ to 1 m scale to white or yellowish, coarse grained, quartzose sandstones showing large scale (1-3 m) trough cross-stratification and containing
scattered, subrounded pebbles of white sandstone and quartzite. Near the base of the member the pebbles are all less than 2 cm but the pebble size increases to a maximum of 10 cm approximately 40 m above the base of the member. Large scale cross-stratification is the most prominent bedding form in this member and the troughs are either solitary or are grouped into cosets with thicknesses of up to 5 m. The cosets are occasionally separated by flat bedded or apparently massive intervals of medium grained sandstone lacking pebbles. Mud flakes are generally scarce throughout this member.

The lower boundary of the member is transitional in all cases and is marked by a colour change from brown to white, an increase in the degree of sorting, higher quartz content, and the incoming of quartzite pebbles. The upper boundary is also transitional and the criteria are the opposite of the lower boundary criteria.

The member can be traced as a distinct unit on aerial photographs, on which it is light coloured with more distinct bedding than the overlying and underlying red-brown sandstone units. The base of the member weathers recessively in many areas and can be traced on aerial photographs as a weak linear depression. The areal distribution of the member is given in Figure 2.33.

The Owen Springs Member is fairly uniform in thickness with the maximum being 220 m at section CP. In
the type area the thickness is 210 m, while at its eastern and western extremities it thins to about 100 m and laterally interdigitates with the normal Hermannsburg Sandstone over a distance of about 2 km. This interdigitation was only seen on the eastern extremity but aerial photographs show a similar effect on the western end. Because of the uniform thickness and limited lateral extent of this member no isopachs have been drawn.

(3) Ljiltera Member

The term Ljiltera Member is proposed for the pebbly and minor conglomeratic sandstones of the upper Hermannsburg Sandstone. In most exposures in the central Amadeus Basin the Ljiltera Member forms approximately the upper half of the Hermannsburg Sandstone although south of the James and Gardiner Ranges most of the Ljiltera Member has been removed by erosion. The areal extent of this member is shown in Figure 2.34.

The member is named after the Ljiltera waterhole where Bagot Creek emerges from the Krichauff Range approximately 15 km west of Hermannsburg Mission (23°57'S, 132°38'E). It is well exposed in Bagot Creek south from Ljiltera waterhole although neither the top nor the base of the member is exposed. The type section is in the Glen Helen gorge 4 km south of Glen Helen Tourist Camp (23°43'S, 132°40'E; aerial photograph Hermannsburg Run 12/5193 from 20.7 cm N, 11.1 cm E to 19.4 cm N, 12.5 cm E of the southwest corner). A lithologic log of the type section is given
in Figure 2.35. The Ljiltera Member is essentially similar to other parts of the Hermannsburg Sandstone except for the presence of subrounded pebbles of earlier sedimentary rocks. The Ljiltera sediments are orange to reddish brown litharenites and rarer sublitharenites. Grain sizes vary from fine to very coarse sand, with most of the sediment falling in the central part of the range. The pebbles and associated mudclasts both range in size up to 7 cm. Bedding is dominantly trough cross-stratified in cosets varying from 1 to 5 m in thickness. Individual cross-stratified units rarely exceed 1 m in size. Pebbles and mudclasts are commonly concentrated near the base of such cosets and become less abundant higher in the coset. The coset frequently becomes finer grained upwards and may merge into small-scale tabular cross-stratified beds or flat beds. The sequence can be divided into numerous partially complete cycles which will be described in more detail later. Towards the top of the Ljiltera Member pebble and conglomerate horizons and conglomeratic sandstones become more abundant and the average pebble size increases with several pebbles in the 12 to 15 cm range.

The base of the Ljiltera Member is gradational with the lower Hermannsburg Sandstone and is defined as the base of the first pebbly sandstone which forms part of the overlying pebbly sandstone sequence. This definition excludes the pebbly sandstones which occur at the base of the Hermannsburg Sandstone on local structures such as Ooraminna and Waterhouse Anticlines.
and the Goyder Pass structure, because these pebbly sandstones are overlain by a relatively thick non-pebbly sandstone interval. On the southern flanks of the MacDonnell Range the upper contact between the Ljiltera Member and the overlying Brewer Conglomerate is usually transitional with progressively more pebbles and minor conglomerate bands high in the Ljiltera Member. The actual contact between the conglomeratic sandstones and the massive boulder conglomerate is usually sharp and either conformable or slightly disconformable due to local scour prior to deposition of the massive conglomerate. Around and east of Ellery Creek the contact is unconformable with the base of the conglomerate cutting down through the sedimentary sequence as far as the Jay Creek Limestone.

The thickness of the Ljiltera Member at the type locality is 480 m and the equivalent horizon at the type locality of the Hermannsburg Sandstone near Stokes Pass is 470 m. Isopachs of the depositional thickness of the Ljiltera Member are given in Figure 2.36. The rapid apparent thinning to the south, where the Ljiltera Member is only preserved in the centres of the synclines, is a result of erosion.

**BREWER CONGLOMERATE**

The Brewer Conglomerate is the youngest formation in the Pertnjara Group. It was named by Wells et al., (in press) from Brewer Plain south of Alice Springs where it
occurs as very low undulating ridges of poorly exposed conglomerate and lag gravel. The type locality of the Brewer Conglomerate is on the southern flanks of the western MacDonnell Ranges south of Stokes Pass. Figure 2.37 shows the general features of the lower portion of the Brewer Conglomerate at the type section. The upper part of the formation in this area forms rounded hills (Fig. 2.38) and is totally obscured by an unconsolidated mass of the weathered out phenoclasts.

In the type area the contact with the Hermannsburg Sandstone is a transition and the base of the Brewer Conglomerate is taken at the top of the last conglomeratic sandstone bed before the massive conglomerate starts.

Sections through the Brewer Conglomerate are generally very difficult to measure because of the massive conjunct nature of the conglomerate and the lack of visible bedding planes. For the most part, thicknesses can only be estimated from aerial photographs using the limited dip information available. In many outcrops of sandy conglomerate where the bedding can be recognized the angle of north dipping imbrication on discoidal pebbles is generally 3° to 10° less than the corresponding bedding plane inclination. In areas of massive conglomerate the orientation of discoidal pebbles can frequently be measured and the average of a number of such readings at a locality gives a rough estimate of the dip. With the use of dip information and a change in dominant pebble
types several angular unconformities have been recognised within the Brewer Conglomerate and these will be discussed later.

The generalized sequence exposed at the type locality is given in Figure 2.39. In most exposures the conglomerate has a continuous framework with subrounded to rounded pebbles, cobbles and boulders of all sizes up to -10 phi. However, boulders of this size are rare and those forming the bulk of the conglomerate fall in the -7 to -9 phi classes. In the lower 700 m of conglomerate the matrix consists dominantly of coarse grained lithic fragments tightly cemented with crystalline sparry calcite. Conglomeratic sandstone bands are quite common in the cliff forming unit (350-700 m above the base of the conglomerate) of the western MacDonnell Ranges which forms a prominent aerial photograph marker horizon in this vicinity. The conglomeratic sandstone horizons are all coarse grained with pebbles up to -7 phi. They are occasionally trough cross-stratified on a large scale, but most of the beds appear to be massive. This unit does not occur east of Arumbera Creek where the conglomerate is massive throughout the lower portion. The interval from 0 to 700 m shows a continual change in the abundance of rock types (Table 2.1). Throughout this interval metamorphic fragments are absent and vein quartz is very rare. In the first 450 m the abundance of greenish siltstone and fine sandstone, probably derived from the Cambrian and Ordovician limestones and siltstones, shows a marked
decrease from about 60 percent to less than 10 percent. Red sandstones and siltstones similar to Arumbera sediments retain a fairly constant proportion through this interval (20-40%) while clean white sandstones and quartzites similar to the Mereenie, Stairway and Pacoota Sandstones increase in abundance. Above 250 m the first occurrences of boulders and cobbles of Bitter Springs Formation and Heavitree Quartzite are recognized and these increase in abundance upwards although at 700 m they still form only 5 percent of the phenoclasts.

The interval from 700 to 1600 m has no outcrops although lag gravels are plentiful. However, these only show the proportion of hard sandstones, siltstones and quartzites since more labile fragments are presumably weathered out. At 900 m there are numerous fragments of clean white Cambrian-Ordovician sandstones which comprise up to 75 percent of the phenoclasts. These pebbles continue in abundance through the rest of the sequence. Arumbera and Mereenie-Pertnjara sandstones are also locally abundant and above 900 m occasional Arunta pebbles are seen. The latter only become abundant above 1600 m where they may comprise 25 to 50 percent of the phenoclasts.

The main feature within the Brewer Conglomerate is the variability of phenoclasts present and the sometimes abrupt changes of dominant lithologies (Table 2.1). This is probably due to local source area control rather than tectonic control as proposed by Prichard and Quinlan (1962).
Only one cycle of dominant phenoclast lithologies is present south of Stokes Pass with stratigraphically lower formations becoming prominent progressively higher in the Brewer Conglomerate ending up with an abundance of basement phenoclasts.

The areal distribution of the Brewer Conglomerate (Fig. 2.40) shows that it is confined almost entirely to the northern margin of the Amadeus Basin. The major deposits occur on the southern flanks of the MacDonnell Range. Isolated outcrops of conglomerate also occur 35 km west of Johnstone Hill on the Mount Rennie Sheet, in the Ross River Syncline east of Alice Springs, and near Larrier Bore and in the north-west Hale River Sheet area in the east. The exposures in the northwest Hale River Sheet area are unusual in that they show the Brewer Conglomerate lying unconformably on the Mereenie Sandstone and the Pertaoorrta Group. This is a similar situation to the Jay Creek area on the Missionary Plains where the Brewer Conglomerate unconformably overlies the Mereenie Sandstone, all the Larapinta Group and the Goyder Formation and Jay Creek Limestone of the Pertaoorrta Group. The pre-Brewer surface geology is given in Figure 2.41.

The Brewer Conglomerate is the thickest formation in the Pertnjara Group with a maximum thickness of approximately 3000 m south of Glen Helen as defined by seismic work. The thickness along the southern margin of the MacDonnell Ranges probably varies from 1200 to 3000 m.
with the unit thinning very rapidly to the south. Most sections measured through the conglomerate measure the depositional thickness exposed on the surface at the present time. Rapid stratigraphic thinning of the basal units southwards beneath the measured section account for the smaller true thicknesses indicated by seismic work. Isopachs of the preserved Brewer Conglomerate are given in Figure 2.42.

No fossils have been found in the Brewer Conglomerate except for those incorporated within the phenoclasts. The age of the formation is therefore uncertain and may range from latest Devonian to Carboniferous.

Undandita Member

This member of the Brewer Conglomerate is being erected to include the upper conglomeratic sandstones and to separate them from the massive undifferentiated conglomerate below. The Undandita Member has only been seen on the Missionary and Brewer Plains south of the Brewer Conglomerate outcrops (Fig.2.43). It may be present in the Mount Rennie Sheet area (not visited by the author) since conglomeratic sandstones have been described from the area west of Johnstone Hill. It probably also occurs but is poorly exposed south of Bingie Bore in the Hale River Sheet area.
The unit is named after the Undandita Native Camp situated 10 km north-west of Gosses Bluff in the western Missionary Plain (23°46′S, 132°14′E). The type section is on the Tyler Pass road southwards from 2 km south of the trig station to the end of outcrop in the creek (23°43′S, 132°21′E; aerial photograph Hermannsburg Run 11/5137 from 11.1 cm N, 6.2 cm E to 7.4 cm N, 7.5 cm E of the south-west corner). The type section (Fig. 2.44) consists of massive, and lesser trough cross-stratified, conglomeratic and pebbly sandstones interspersed with conglomerate horizons. Where visible the bedding appears fairly uniform with bed- sets in the order of 1 to 3 m thick. The sandstones frequently have pebbles scattered throughout, and cobbles and small boulders concentrated near the base. Conglomerate horizons are usually less than 30 cm thick but occasional bands are up to 5 m. The major component of the Undandita Member is coarse to very coarse litharenite with subangular to rounded grains. It is usually greenish white in colour due to the abundance of green lithic fragments. Pebbles and cobbles within this member are characterized by vein quartz and abundant metamorphic fragments including schists, gneisses, granite, amphibolite, etc. Metamorphic fragments form about 50 percent of the phenoclasts and the remainder are sedimentary rocks - the most of which are Cambrian and Ordovician sandstone and siltstone, and quartzite.
The Undandita Member has a variable thickness which usually cannot be measured because of lack of outcrop in the Missionary Plain. However, it probably attains a thickness of up to 1000 m and occurs as a lens shaped body between the Brewer Conglomerate in the north and the Ljiltera Sandstone Member to the south. The contact between the Undandita Member and these two units is assumed to be conformable, except possibly over local structures, and all these units are assumed to interdigitate. (The relationship between these units will be discussed in more detail in Chapter 5). The distinction between the three units is based entirely on the size and quantity of phenoclasts within the unit. The Brewer Conglomerate is a conjunct conglomerate with very rare sandstone lenses. The Undandita Member is a conglomeratic sandstone with numerous pebbles and boulders throughout and contains several thin conglomerate horizons. In contrast the Ljiltera Member only contains scattered pebbles and cobbles with very rare thin pebble bands.

FINKE GROUP

The Finke Group is a sequence of relatively flat lying conglomerate, sandstone and siltstone which outcrops in the Finke Sheet area and extends further north, east and west in subsurface. To the north and west of the Finke Sheet the Finke sediments interdigitate with the Pertnjara Group and the relationships will be discussed in Chapter 5.
The name Finke River Sandstone or series was first used by Chewings (1914) and was revised and formally defined by Wells, Stewart and Skwarko (1966). Chewings' (1914) original Finke River Sandstone included the Crown Point Formation and Mesozoic sediments but these have been excluded from the present definition. The Finke Group is, of course, named after the Finke River which flows diagonally through the Finke Sheet area and forms with the main drainage system of the Amadeus Basin.

The Group was divided into four formations by Wells et al. (1966) and a fifth unit, the Hakea "Formation" has been proposed in this study to describe the shale, siltstone and fine sandstone beds encountered in water bores 92 and 122 on the Finke Sheet. The basal Polly Conglomerate lies unconformably on older rocks including the Stairway Sandstone and Winnall Beds in the Black Hill Range, and further south it overlies Proterozoic sediments and large areas of Precambrian crystalline basement. The Polly Conglomerate is conformably overlain successively by the Langra Formation, the Horseshoe Bend Shale, the Idracowra Sandstone and the Hakea "Formation". The Finke Group is overlain unconformably by the Permian Crown Point Formation and various Mesozoic and Tertiary sediments.

The outcrop and subsurface distribution of the Finke Group (Fig. 2.1) does not include its eastward extension beneath sediments of the Great Artesian Basin.
sequence. The thickness of the Group over most of the area is uncertain due to poor outcrop and lack of seismic work. The thickness in outcrop is estimated at 900 m while subsurface in McDills No. 1 Well it is 1000 m. The Group is also known in subsurface from Mount Charlotte No. 1 Well, Witcherrie No. 1 Well and numerous water bores. The Group is absent, due to faulting and pre-Mesozoic erosion, from the Purnie No. 1 Well.

The age of the Finke Group is uncertain since no fossils have been found in it. It overlies Ordovician and older sediments and is unconformably overlain by the Crown Point Formation of latest Carboniferous to Early Permian age. With this evidence, and the fact that the Group is correlated with the Pertnjara Group (see Chapter 5), the Finke Group is assigned to the Late Devonian and Early Carboniferous.

POLLY CONGLOMERATE

The Polly Conglomerate was defined by Wells et al. (1966) as the basal formation in the Finke Group. The formation is named after Polly Corner (25°13'S, 134°11'E) and the type section is situated 4 km east of Polly Corner on the northern flank of the Black Hills Range (25°13'S, 134°13'E; aerial photograph Charlotte Waters Run 4/5197 from 14.8 cm E, 11.2 cm N to 14.8 cm E, 11.7 cm N of the south-west corner).
The type section of the Polly Conglomerate is shown in Figure 2.45. It unconformably overlies the Winnall Beds at the type locality and the basal 5 m are characterized by abundant subangular pebbles and cobbles of Winnall sediments up to 15 cm on the long axis. Sub-rounded to subangular pebbles of milky and rutilated vein quartz are also common, along with granite, pegmatite, schist, dolomite, sandstone and quartzite. The conglomerate is disjunct with abundant coarse sand to pebble gravel in a polymict matrix of quartz, feldspar and rock fragments. Where visible, bedding appears to be dominantly massive or tabular cross-stratified with bedsets varying from 10 to 75 cm. This basal conglomerate is overlain by a thin succession of fine to coarse grained poorly sorted, tabular cross-stratified sandstones with common pebbles. Overlying this unit most of the Polly Conglomerate consists of poorly sorted coarse sand to pebble conglomerate with subangular to subrounded pebbles of quartzite, quartz, granite, pegmatite, porphyry and some sandstone (Fig.2.46). Most pebbles are less than 3 cm long and in several areas imbrication is quite apparent. Bedding in this section is usually massive or flat bedded with rarer tabular cross-stratification. The uppermost horizons are poorly sorted pebbly silty sandstones which grade upwards into the Langra Formation.

In the Umbeara area the Polly Conglomerate overlies igneous and metamorphic basement rocks and is
a poorly sorted, coarse sand to boulder conglomerate containing phenoclasts of granite, gneiss, pegmatite, amphibolite, porphyry and vein quartz. Bedding in this area is dominantly massive or flat.

Outcrops of the Polly Conglomerate are confined to the Finke Sheet and the south-west corner of the Kulgera Sheet. To the south of these Sheet areas the Polly Conglomerate is not uniformly present over the whole area since water bores SG53/6-2 and 169 show a contact between the Langra Formation and basement granite. In the subsurface the Polly Conglomerate extends and thickens eastwards into the McDills and Dalhousie Sheet areas (Fig.2.47). The thickness of the Polly Conglomerate is therefore variable, being 40 m at the type locality and about 200 m north-west of Umbeara Homestead. In McDills No. 1 Well the maximum thickness of 393 m is recorded. Interpretive isopachs of the member are given in Figure 2.48. No definite age can be placed on the formation but it is probably Late Devonian.

**LANGRA FORMATION**

The Langra Formation conformably overlies the Polly Conglomerate and likewise outcrops are restricted to the south-western Finke Sheet area. This is in disagreement with the distribution proposed by Wells et al. (1966) but is in accordance with their field observations (see discussion p. 72). The Langra
Formation is believed to have a wide subsurface distribution underlying parts of the Kulgera and Rodinga Sheets and most of the Finke and McDill Sheets, as well as further east (Fig. 2.49).

The Langra Formation is named after Langra Well 16 km south-west of Horseshoe Bend Homestead (25°18'S, 134°08'E). The type section (Wells et al., 1966) runs from the northern flank of the Black Hill Range to Horseshoe Bend Homestead (25°13'S, 134°14'E; aerial photograph Charlotte Waters Run 4/5197 from 14.8 cm E, 11.7 cm N to 20.0 cm E, 17.0 cm N of the south-west corner). It consists of dominantly well to poorly sorted sandstone with interbeds of conglomerate and siltstone (Fig. 2.50). The sandstones are usually white to yellowish while the siltstones are red (Fig. 2.51). Conglomeratic sandstone horizons are quite common in the basal and central parts of the member and they may reach a thickness of 4 m although most bands are less than 1 m thick. The size of the phenoclasts varies with the largest reaching 12 cm in the thicker bands while pebbles up to 5 cm are common throughout. The range of rock types in the phenoclasts is essentially similar to that in the Polly Conglomerate but vein quartz becomes increasingly common towards the top of the Langra Formation (Table 2.2) where it commonly displays an unusual etched surface. The conglomerate horizons usually show massive or indiscernible bedding and they are separated by cross-stratified and flat bedded sandstones and laminated or massive thin silt-
stones. The sandstones are generally medium to coarse grained feldspathic sandstones with both rounded and angular quartz grains. Cross-stratification is a common feature with the bedsets varying from 0.25 to 2 m. Trough cross-stratification is more abundant than tabular forms and the troughs usually overlie a marked erosion surface. Sedimentary structures are abundant and include erosion channels, mudclast breccias, lensoid conglomerates, slump or contorted bedding, and current lineation. The sandstones usually form units 1 to 10 m thick between the conglomerate and overlying sandy siltstone, and they occasionally show an upward decrease in grain size. The sandy micaceous siltstones are usually red, up to 1 m thick, and may be laminated or massive. Linguoid ripple marks are occasionally seen in the laminated horizons. The 12 m thick siltstone towards the top of the formation is a marker horizon and has been named the Engoordina Siltstone Member. Above this member there are no conglomerates and the sandstones are generally fine grained with occasional siltstone inter­beds. The top of the Langra Formation is marked by an abrupt but conformable transition to the Horseshoe Bend Shale. The 50 cm below the contact are characterized by medium sandstone with occasional well rounded etched quartz pebbles, and a 10 to 20 cm greenish ferrous iron cemented band.
The contact with the underlying Polly Conglomerate is conformable and gradational wherever seen, as is the upper contact with the Horseshoe Bend Shale. Isopachs of the formation are given in Figure 2.52, although these are largely interpretive because of the small number of measured sections.

Engoordina Siltstone Member

The Engoordina Siltstone Member is proposed for the major siltstone unit within the Langra Formation. This siltstone was termed Unit 2 of the Langra Formation by Wells et al. (1966).

The Engoordina Siltstone Member is named after Mount Engoordina about 2 km north of Horseshoe Bend Homestead (25°11'S, 134°14'E). The type section of the member is within the type section of the Langra Formation nearly 1 km south of Horseshoe Bend Homestead (25°13'S, 134°14'E; aerial photograph Charlotte Waters Run 4/5197 at 18.6 cm E, 15.3 cm N of the south-west corner). Part of the type section is shown in Figure 2.53 and consists of 12 m of alternating laminated and massive, red, sandy, micaceous siltstone. The unit contains a few ripple marked horizons, but is otherwise devoid of sedimentary structures. It is distinguished from the underlying siltstones on the basis of thickness and the change in grain size from medium to coarse sandstone below to fine sandstone above. It is distinguished from the Horseshoe
Bend Shale on the basis of colour, thickness, scarcity of sedimentary features, and the feldspathic nature of the clean fine sandstone overlying it.

The unit is 12 m thick at the type locality and thickens to 27 m in Mount Charlotte No. 1 Well. It appears to thin to the south and is about 10 m thick on the southern flank of Black Hill Range, while its presence in water bores further south is uncertain.

HORSESHOE BEND SHALE

The Horseshoe Bend Shale was defined by Wells et al. (1966) to consist of red and green biotitic shales with occasional thin interbeds of fine sandstone. The formation was named after Horseshoe Bend on the Finke River and the type locality is on the eastern bank of the Finke River north from Horseshoe Bend Homestead (25°12'S, 134°15'E; aerial photograph Charlotte Waters Run 4/5197 from 20.0 cm E, 17.0 cm N to 18.9 cm E, 21.5 cm N of the south-west corner). The sequence at the type section is given in Figure 2.54 and shows a fairly uniform succession of laminated micaceous siltstones and very fine grained silty sandstones. The siltstones are dominantly red-brown and green in outcrop but tend to be grey in bores (Rochow, 1965). However, in Mount Charlotte No.1 and McDills No.1 Wells in the intervals 18 to 158 m and 1155 to 1240 m below ground level respectively the Horseshoe Bend Shale retains its
red-brown and green colouration with only minor grey calcareous siltstone bands. Red-brown siltstone usually predominates and green siltstone only occurs as irregular interbeds. In the lower part of the formation the bedding is rather obscure and thinly laminated but towards the top of the exposure at the type section bedding cyclicity becomes apparent (Fig. 2.55). The cycle is similar to that of the Dare Siltstone Member starting with a massive or laminated green siltstone overlain by ripple marked, interlaminated, red and green siltstone which is overlain in turn by a massive red-brown mudstone with occasional green flecks. This cycle is repeated numerous times with each cycle generally being 1 to 3 m thick. Sedimentary features are common in this formation and include ripple marks, mud cracks, pseudomorphs after halite, and thin lenses of limestone and gypsum.

The upper part of the Horseshoe Bend Shale is not exposed at the type locality but south of Black Hill Range towards Sister Hills it is exposed and is generally similar. However, gypsum bands up to 5 cm thick were seen in the greenish and red-brown siltstones. In this area the contact with the Idracowra Sandstone is transitional with the upper Horseshoe Bend Shale containing progressively more thin, fine sandstone interbeds. One unusual feature noted was the occurrence of a couple of thin bands consisting entirely of biotite.
The contact of the Horseshoe Bend Shale and the Langra Formation is conformable and gradational although there may have been a period of non-deposition between the two units. The upper boundary with the Idracowra Sandstone is likewise generally conformable or shows a slight local disconformity. The lithologic change is transitional in most areas which again suggests little erosion occurred between the two units. East and south of Lilla Creek Homestead the top of the formation is eroded and it is unconformably overlain by the Crown Point Formation and Mesozoic and Tertiary sediments.

The Horseshoe Bend Shale is probably the most widely distributed formation in the Finke Group (Fig.2.56). It is present over most of the Finke Sheet and extends west onto the Kulgera Sheet, north onto the Rodinga Sheet and eastward beneath a regional unconformity onto the McDills Sheet. It is not present at Witcherrie No.1 Well to the south-east and its north-east extent is uncertain.

The thickness of the Horseshoe Bend Shale at the type locality is 92 m although this is not the total thickness of the formation. South of Black Hill Range the thickness is 310 m. Thicknesses, where known, have been plotted on the interpretive isopach map (Fig.2.57).
The age of the formation is uncertain but it is probably latest Devonian or Early Carboniferous.

**IDRACOWRA SANDSTONE**

The Idracowra Sandstone was named by Wells *et al.* (1966) as the uppermost formation of the Finke Group. It conformably overlies the Horseshoe Bend Shale in the northern Finke Sheet area and it extends northwards onto the southern Rodinga Sheet area (Fig. 2.58). In the latter, the white sandstone conformably overlying the Horseshoe Bend Shale and similar to the Idracowra Sandstone in all characteristics has been named Santo Sandstone (Wells, Ranford, Stewart, Cook & Shaw, 1967). This sandstone is considered to be the same formation as the Idracowra Sandstone and thus the latter name, which has priority, is retained.

The Idracowra Sandstone is named after the Idracowra Homestead (25°00'S, 133°47'E) and the type section is situated on the side of a mesa 6.5 km south-south-east of the homestead (25°03'S, 133°49'E; aerial photograph Charlotte Waters Run 2/5070 20.7 cm E, 15.7 cm N of the south-west corner). The type section is given in Figure 2.59. The base of the formation is not exposed at the type locality and the top of the section is a silcrete covered erosional surface. The lowest 10 m consists of yellow to brown, poorly sorted, medium grained sandstone interbedded with thin red-
brown beds of sandy siltstone. The sandstones are either massive, tabular cross-stratified or flat bedded with the bedding usually poorly exposed. The overlying 20 m consists largely of trough cross-stratified medium sandstones with occasional coarse grained laminae and rare, small, rounded quartz pebbles. The cross-strata are either solitary or grouped in sets with individual cross-stratified sets varying in thickness from 15 cm to 1 m. The cross-stratified sets are frequently separated by flat bedded medium to fine sandstones which occasionally show current lineation. The uppermost 20 m at the type section is poorly exposed medium grained kaolinitic sandstone with poorly developed bedding. Mud clasts occur sporadically throughout, but no pebbles were seen in the upper portion.

The basal contact of the Idracowra Sandstone is exposed south of the type section and in most places appears to be conformable and gradational. The base is defined as the base of the first thick bedded, yellowish, medium sandstone above which siltstone only occurs as thin interbeds.

The thickness of the Idracowra Sandstone at the type section is 52 m and the type section lies approximately 10 m above the Horseshoe Bend Shale. At Mount Casuarina approximately 80 m of the formation is present while further east in the area around Mount Hakea the thickness is estimated to be about 280 m,
162 m of which was intersected in bore SG53/6-91. Interpretative thicknesses of the Idracowra Sandstone are given in Figure 2.60.

The Idracowra Sandstone forms the uppermost unit of the Finke Group north and west of Rumbalara Siding. Its top is an erosional surface over all this area and it may be capped with Tertiary silcrete or unconformably overlain by Permian and younger sediments. South-east of a line through Rumbalara Siding and Mount De Souza the top of the Idracowra Sandstone is preserved and is apparently conformably and gradationally overlain by a shaly siltstone and fine sandstone unit - the Hakea "Formation". However, in the south-eastern part of the Finke Sheet area the Idracowra Sandstone was either not deposited or has been removed by erosion.

The age of the Idracowra Sandstone is uncertain but it is probably Early Carboniferous.

**Hakea "Formation"**

The Hakea "Formation" is a new term proposed for the shales, siltstones, and fine sandstones occurring in water bores 70, 87, 92, 94, and 122, on the Finke Sheet (Fig. 2.61, after Wells et al., 1966). The "Formation" is named after Mount Hakea (25°12'S, 134°29'E) which is situated 6 km north-west of water bore SG 53-6/92 (25°13'S, 134°32'E).
From drill logs and cuttings the Hakea "Formation" consists of relatively thin bedded and fissile red-brown to greenish shales and siltstones interbedded with lesser quantities of slightly calcareous fine grained silty sandstones. The unit appears to be similar to the Horseshoe Bend Shale and was termed such by Wells et al. (1966). However, it overlies the Idracowra Sandstone with apparent conformity and the boundary between the two appears to be gradational. The upper boundary of the Hakea "Formation" is unconformable with the overlying Crown Point Formation. The thickness and lateral extent of the Hakea "Formation" is uncertain, but it appears to be about 50 m thick in water bore SG 53-6/122.

DISCUSSION ON FINKE GROUP NOMENCLATURE AND DISTRIBUTION

The Finke Group still has not been studied in sufficient detail for a precise definition of all the constituent units. This is largely because of the relatively poor and discontinuous sections through most of the area and the presence of Tertiary silcrete cappings on most of the mesas. The only good continuous exposures are alongside the Finke River north and south of the Black Hill Range. The type sections of the four oldest constituent formations all lie north of the Black Hill Range and there is no doubt about their identity north and west of Horseshoe Bend.
However, south of the Black Hill Range the author disagrees with interpretation proposed by Wells et al. (1966). An almost continuous section is exposed on the east bank of the Finke River from Polly Corner to Mount Musgrave. Only the Polly Conglomerate and Langra Formations are recognized in this area by Wells et al. (1966) but the section traversed by the author includes all four formations. Figure 2.62 shows the succession south of Black Hill Range which is almost identical to the succession north of the range. Since the type localities and successions have been defined by Wells et al. (1966) on the north flank of the range, the formation boundaries south of the range cannot be altered to include the Horseshoe Bend Shale and Idracowra Sandstone within the Langra Formation as shown on the Finke Sheet map.

A further consideration is that at Mount Musgrave the dip is 2°S at a maximum and this is not sufficient to enable the full section at Mount Musgrave, which stands at least 100 m above the plain, to fit beneath the nearest Horseshoe Bend Shale outcrops to the south unless faulting or a change in general fold character is invoked between the two outcrops. The inclusion of Idracowra Sandstone in the Sister Hills probably represents a cartographic error on the 1st Edition (1968) map because it is shown to overlie Langra sediments and to be stratigraphically equivalent to them.
The sediments exposed around and north of Mount Musgrave, as well as in the upper 53 m in Cloughs Bore, are almost certainly Horseshoe Bend Shale and Idracowra Sandstone as defined in their type areas. Thus, almost continuous exposures of flat-lying or very gently folded Horseshoe Bend Shale can be traced from 2.5 m south of Black Hill Range to 24 km east-north-east of Umbeara Homestead.

Likewise, unless faults are invoked between the Black Hill Range - Mount Musgrave section and the outcrops further east, it is impossible for the latter to be Langra Formation. The dip in the Mount Obstruction area is approximately 1° to 3° to the south-east with very gentle undulating folding modifying this in places. Projecting the solid geology eastwards, the sandstone at Mount Obstruction would overlie the Horseshoe Bend Shale and belong to the Idracowra Sandstone. The latter can then be traced northeast to Mount Hakea and Point Eremophila. The cross-section through water bores SG53-6/91, 92 and 122 (Wells et al., 1966, p.32) also shows incorrect formation identification and the present interpretation is given in Figure 2.61. The accompanying Finke Sheet map has also been amended to show the present interpretation of the formation distribution within the Finke Group.
CHAPTER III

SEDIMENTARY FEATURES

The study of sedimentary features has been divided into two parts. The analysis and description of individual features indicates the variability of the depositing medium while the manner in which the individual features are combined to form groups is indicative of the overall depositional environment. The descriptions of individual sedimentary features have been grouped according to the grain size of the deposit, i.e. siltstone, sandstone and conglomerate.

In general the terminology and grouping of the various structures follows the classification proposed by Conybeare and Crook (1968). Bedset thicknesses, where given in generalized size classes, follow the subdivision proposed by Conybeare and Crook, namely -

- small scale - less than 5 cm
- medium scale - 5 cm to 2 m
- large scale - 2 to 8 m
- very large scale - greater than 8 m

Cross-stratified units are named according to the classification proposed by Allen (1963b), except where slight modification is required for clarity and distinction. Likewise, ripple mark nomenclature and ripple indices follow the terms and definitions proposed by Allen (1963a, 1968) and Tanner (1967).
SEDIMENTARY FEATURES OF THE SILTSTONE UNITS

Sedimentary features are abundant throughout the siltstone units although they are rather restricted in variety. Ripple marks, mud cracks and halite pseudo-morphs are widely represented while erosional features are virtually absent except where the siltstones are interbedded with fine sandstone laminae. However, even in the latter case, erosional features are rare and are limited to very shallow, small, concave channels. Deformational features are likewise scarce except where the siltstones are overlain by sandstones.

BEDDING CHARACTERISTICS

Three types of bedding characterize the siltstone units - massive, laminated and rippled. Massive beds vary in thickness from 1 cm to 5 m. They predominate in the Dare Siltstone Member and upper parts of the Horseshoe Bend Shale. They may contain vugs or mottled patches (Fig.3.1), especially in the thicker units. Only rare signs of lamination were seen in the massive light reddish brown siltstones of the Dare Siltstone Member. X-ray radiography (see appendix 3) has helped to elucidate the internal structures of several of these beds. Most of the reddish brown siltstones appear to be massive and the only visible structures present in some samples are small, elongate, aligned calcite vugs (Fig.3.2).
However, a few samples contain faint or discontinuous parallel laminations (Fig. 3.3) which indicate that at least in these samples the sediments have not been churned or disrupted. Lamination on such a fine scale in siltstones is strongly suggestive of deposition from suspension. Most of the thin, apparently massive, greenish, micritic and dolomitic siltstones also show fine parallel laminae and were probably deposited from suspension (Fig. 3.4).

Laminated sediments and associated ripple-bedded sets (Fig. 3.5) occur throughout the Parke Siltstone and Horseshoe Bend Shale but are most abundant in the lower parts. Red, brown and greenish siltstone laminae are the only forms occurring with the massive beds in the upper parts of these formations (excluding the Amulda Member) whereas elsewhere fine sandstone and rare limestone bands are interbedded with the siltstones. In the limestone bands the laminae are usually slightly irregular and bedding may be continuous or discontinuous (Fig. 3.6). Small sandstone lenses and wavy laminations are quite common. Ripple marks and mud cracks locally disturb the laminations but churning of the sediment is never prominent and trace fossils are extremely rare.

RIPPLE MARKS

Ripple marks are probably the most abundant
sedimentary features in the siltstone units. They are predominantly symmetrical or slightly asymmetrical, with only rare modified forms. The ripple marks are all formed in siltstone or very fine silty sandstones and generally have a small wavelength and amplitude but considerable lateral extent. The small scale of the ripple marks is characteristic of this grain size (Allen, 1963a) and the shape is generally characteristic of wave formed ripples.

(1) **Symmetrical Ripple Marks**

Symmetrical and near symmetrical ripples with ripple symmetry indices between 1.0 and 1.5 are the most common ripple form in the Dare Siltstone Member (Fig.3.7). The ripple crests are parallel and generally straight or very slightly sinuous (straightness index between 5 and 20 for sinuous forms and infinite for straight forms). The bifurcation index was generally large (up to 30) but could rarely be measured because of the limited exposures.

In cross-section, the ripples may either show deposition preferentially oriented in one direction but planed off to form symmetrical ripple marks, or deposition may be parallel to both limbs of the ripple (Fig.3.8). The former represents modified current depositional forms while the latter is typical of suspended load deposition.
Six of Tanner's (1967) parameters have been calculated, where possible, for all of the ripple marks (Table 3.1). Insufficient field measurements were made of all the ripple criteria for a clear cut distinction of modes of ripple mark formation. However, all the symmetrical ripples fall in the wave dominated sections of the graphs and there is little doubt about their mode of origin (90-98% confidence, Tanner, 1967).

According to Tanner (1968), symmetrical wave ripples form when there are no currents and the water depths under the wave troughs and crests are almost equal. Water depth could not have been very shallow otherwise depth-biased, slightly asymmetrical ripple marks would have resulted.

(2) Asymmetrical Ripple Marks

Slightly asymmetrical ripple marks are also common in all the siltstone units. They are similar in most characteristics to the symmetrical ripple marks, i.e. they are relatively straight and continuous and generally have a low bifurcation index. However, their ripple symmetry index varies between 1.1 and 1.3 and their internal structure is dominated by prograded, foreset laminae (Fig.3.9). Tanner's (1967) indices indicate that there is a 90 to 98 percent confidence that these ripples were formed by wave action.

According to Tanner (1967), asymmetrical ripple marks of this type are due to current-biased
wave action and not to direct current action. The action of a weak current, often wind induced, is sufficient to induce slight asymmetry to the pattern of water motion and favour grain movement in one direction. A similar effect can be generated in very shallow water without currents where asymmetry is caused by a regular ripple pattern impinging on a slightly shelving subsurface. Such ripples can be termed depth-biased wave-formed ripple marks and they cannot be distinguished from current induced wave ripple marks. 

(3) **Interference Ripple Marks**

Interference ripple marks occur sporadically in the siltstone units although they are rare in the Dare Siltstone Member. They usually consist of one dominant ripple mark set with secondary ripple crests present in the troughs. Occasionally the secondary ripples produce nodal interference with the primary ripple crest with a resultant increase in crest height at that point (Fig.3.10). The secondary ripples are most common oriented approximately at right angles to primary ripple crests although the angular divergence between crests varies from $60^\circ$ to $90^\circ$.

Interference ripple marks are commonly produced in shallow water where two different wave sets unite. However, they can be formed by transverse drainage in exposed primary ripple troughs although these forms do not produce true
interference nodes. Gubler et al. (1966) noted that interference ripple marks are characteristic of an aquatic environment but that it may be fluvial, lacustrine, estuarine or marine.

(4) Modified Ripple Marks

Modified current ripple marks are rare and were only seen at a couple of localities in the basal Dare Siltstone Member. They consist of asymmetrical, flat topped ripples separated by flat bottomed troughs which occasionally show secondary transverse ripples (Fig.3.11). The latter were probably generated by flow in the ripple troughs which are partially filled with plane laminated muds. In cross-section the ripples are seen to be of Jopling and Walker's (1968) Type A ripple-drift cross-lamination with eroded sigmoidal cross-laminae (Fig.3.12). When plotted on graphs using Tanner's (1967) indices, the primary ripple marks fall in the wave generated field and the secondary modifications probably occurred under very shallow water conditions.

(5) Bidirectional Current Ripple Marks

Bidirectional current ripple marks were noted at one locality in the Horseshoe Bend Shale (section FG). They consist of homogeneous very fine sand ripples with gently rounded crests, shallow troughs and a wave length of 6 cm. The internal structure of the ripple
mark shows cross-laminae oriented in opposed directions at right angles to the crest axis (Fig.3.13). The lower north-west facing set is truncated by the overlying south-east facing set and both probably represent traction current deposits.

The mode of formation of these ripple marks is important for environmental reconstruction. Similar ripple marks of both small and large scale have been reported from tidal areas and especially tidal channels (Conybeare and Crook, 1968). The occurrence of upstream and downstream cross-stratification from the same bed has also been recorded from antidunes in the fluvial environment (Hand, Wessel & Hayes, 1969), although these generally only occur on a medium scale. A further method for the production of bidirectional current ripple marks may be postulated. They could form as a result of wind induced current flow in extensive, very shallow water environments such as playa and salt lakes. These lake level changes or seiches are well known and have been found to be quite marked in broad, shallow lakes such as Lake Eyre (Bonython, 1955). Bonython recorded up to $\frac{1}{2}$ m fluctuations in lake level about the mean, which were caused entirely by wind action. A decrease in lake level by $\frac{1}{2}$ m caused the shoreline to retreat almost 1 km and the period of this oscillation was usually 12 hrs or less.
Such oscillations in water level would have the same effect as tides and are probably capable of producing similar bidirectional current ripple marks.

HYDRODYNAMIC SOLE MARKINGS

Hydrodynamic sole markings are rare in the siltstone members and the only ones noted consist of shallow scour depressions usually filled with coarse siltstone or fine sandstone.

(1) Erosional Channels

Erosional channels are very rare and the only one seen consists of an even, shallow, concave erosional surface overlain conformably by lensoid, fine grained, laminated sandstones which grade laterally into coarse siltstone (Fig.3.14). The channel is 1.35 m wide and 15 cm deep. It must have been scoured and filled by an almost straight flowing current since it is very nearly symmetrical in cross-section.

(2) Toroids

Toroids or circular scour pits are infrequent and were only recognized in the Deering Siltstone Member and the upper parts of the Harajica Sandstone Member on Dare Plain.

The toroids generally have a smooth exterior which may show a slightly swirled shape (Fig.3.15). They vary in diameter from
8 to 75 cm, but are generally only 2 to 5 cm deep. They are usually scoured into siltstones and filled homogeneously with laminated, coarse siltstone or fine sandstone. They are assumed to form by the erosive action of stationary eddies or whirlpools in flowing water (Conybeare & Crook, 1968).

RHEOTROPIC STRUCTURES

Rheotropic structures are produced by the deformation of variably cohesive sediments under the action of external forces especially gravity. With the exception of convolute bedding (Sanders, 1960) the formation of rheotropic structures is independent of current flow and organic reworking. The terminology used to describe the nature of rheotropic deformation is that proposed by Elliott (1965).

(1) Contorted Bedding (Hydroplastic)

Structures formed by hydroplastic rheotropic deformation (Elliott, 1965) are not very frequent in the siltstone formations. However, in the Deering Siltstone Member, several examples of small scale deformation were noted (Figs 3.16 and 3.17). These structures are confined to siltstone beds 2 to 5 cm thick and consist of irregularly contorted, convolute or recumbent folds overlain and underlain by undisturbed laminae. They probably formed by gravity sliding, or deformation may have been induced by localized
Ball and pillow structures are rare and poorly defined (Fig. 3.18) in the upper interlaminated sandstone-siltstone portion of the Deering Siltstone Member. They probably result from lateral translation of hydroplastic sediment (Conybeare & Crook, 1968).

(2) Slump Structures (Hydroplastic)

Large scale hydroplastic, rheotropic deformation of a 1.5 m thick mudstone horizon was seen in the Langra Formation at section FH on the southern flank of Black Hill Range.

The mudstone horizon is massive but contains white and varicoloured laminae and is overlain in part by sandy siltstone and in part by a large scale (5+ m thick), laminated, trough cross-stratified set of probable aeolian origin. The deformation features visible include irregularly contorted folding (Fig. 3.84) probably caused by direct loading, and flow or intraformational recumbent folding caused by lateral loading (Figs 3.85 & 3.86). Figure 3.87 shows more details of the intraformational mudclast breccia shown in Figure 3.86 together with the planar erosion surface. The breccia probably formed in a syncline between overturned anticlines and the clasts are probably derived from the upper layer of the mudstone.

The features most probably formed as a result
of a sand dune encroaching onto the surface of a semi-consolidated lacustrine or playa-lake mudstone. Similar types of fold structures have been described by Brown (1969) from Coorong Lagoon, S.A., and are ascribed to a similar mode of formation with folds in lagoonal muds being buried by the encroaching dune and probably suffering further hydroplastic deformation.

(3) **Spongy Texture and Vugs (Hydroplastic)**

A few slightly calcareous sandy horizons in the basal Deering Siltstone Member on Dare Plain show the development of a spongy texture (Fig. 3.19). The sandstone contains numerous randomly scattered, bubble-like vugs with a maximum size of 4 mm. They may form by entrapment of air following rapid inundation of comparatively dry sand by high tides (Conybeare & Crook, 1968) or by rapid alluvial flooding.

Calcite lined vugs are found in most of the massive light reddish brown siltstones of the Dare Siltstone Member. They are rarely abundant although may be locally concentrated in pockets aligned parallel to the bedding (Fig.3.20). Their mode of formation is uncertain but may be similar to the above with calcite diagenetically lining the voids.

(4) **Mudcracks (Quasi-solid)**

Mudcracks are abundant features preserved in the laminated portions of the Parke Siltstone and
the Horseshoe Bend Shale. The cracks are frequently curved and the polygons produced are usually irregular with 3 to 5 slides although occasionally fairly regular hexagonal forms are developed (e.g. Fig.3.21). The mudcracks may form on planar beds (e.g. Fig.3.22) but they are equally frequent on all types of ripple marked beds (Figs 3.10 & 3.23). The cracks may be filled with the same or a different coloured sediment, and they are much more commonly associated with reddish brown siltstones than with greenish siltstones.

Whether these cracks are dessication cracks or synaeresis cracks is almost impossible to determine although using the definition of White (1961) they are probably dessication cracks, i.e. they do not occur between sandstone units, the shale is fissile and occasionally there is evidence of curling. However, they could have formed by the method outlined by Burst (1965) since some of the clays contain appreciable quantities of smectite (e.g. DH 54) and salinity values are known to change because of the solution and deposition of associated salt casts.

Gubler et al. (1966) noted that mud cracks are present in fluvial, playa, deltaic and lagoonal environments but that they are practically absent from tidal flats since with the latter the interval of exposure is too short for complete dessication.
DEPOSITIONAL STRUCTURES

Depositional structures are common in the siltstones and are a characteristic of them. The widespread presence of crystal casts is very significant in environmental reconstruction. Colour mottling is also an important but infrequent feature in the massive siltstones.

(1) Salt Casts and Gypsum

Salt casts are a characteristic feature of the Dare Siltstone Member and they also occur quite frequently in the Horseshoe Bend Shale. They are rare in the other members of the Parke Siltstone and have not been seen in any of the sandstone units.

The salt casts are most abundant in the laminated, green siltstones although they also occur in the laminated, brown siltstones. They are always preserved as casts on the lower surface of the overlying siltstone and they vary in size from 1 mm cubes to cubes with almost 2 cm sides (Fig.3.24). The largest salt cast seen extended 1 cm into the underlying sediment (Fig.3.25). O'Brien (1968) noted a similar consistent pattern with salt casts confined to the base of a bed. The salt casts have typical hopper shaped, crystal faces which distinguish them from other isometric crystals such as pyrite. The orientation of salt casts appears to be random which is to be expected if the cube has fallen through water, i.e. they could
have originated on the surface of the water and when their specific gravity exceeded their buoyancy, they would have sunk to the mud beneath. The formation of salt "biscuits" in saline lakes has been noted by Muller and Irion (1969). They noted that cubes with up to 1 cm sides form on the base of the biscuits and when the latter sink, the larger crystals would protrude into the underlying sediment. They also noted the formation of isolated salt-cubes on the lake margins. Salt crystals that have grown in situ would preferentially form parallel to the depositional inter-face. However, large salt casts probably resulted from continued growth of randomly oriented crystals after they had sunk to the bottom. Subsequent cohesion or subaerial drying of the sediment formed a mould about the salt crystal which would have dissolved at the next influx of lower salinity water thus allowing the mould to be filled by the incoming sediment. In most cases, salt mould horizons are overlain by coarse ripple-bedded siltstones and at section EN they are frequently filled with ripple-bedded, silty, very fine sandstones.

The salt casts are frequently associated with mud cracks and they are assumed to have formed in shallow very saline water or on mudflats periodically inundated with salt water. The most probable environment is a non-marine saline lake, subjected to periods of desiccation.

Gypsum occurs less frequently than salt casts
and except for a few crystals and casts in the upper Dare Siltstone Member, it is restricted to the upper part of the Horseshoe Bend Shale. In section FH several beds of crystalline gypsum varying from 1 to 5 cm thick occurred interbedded with predominantly green, biotitic, laminated siltstones and less frequent red-brown siltstones. The association is cyclic for about 15 m vertically with each cycle being 1 to 2 m thick, starting with a thin, red, laminated siltstone horizon and passing upwards into a thick laminated sequence of green siltstone which is in turn overlain by the selenite horizon. No salt casts were observed in this cyclic sequence but a few mudcracks were apparent. The gypsum was probably deposited in a warm climate from shallow water with restricted circulation - either marine or non-marine.

(2) Colour Mottling

Colour mottling is largely confined to massive siltstones in the Dare Siltstone Member and the Horseshoe Bend Shale. In the former, it occurs as irregular greenish mottles up to 2 cm in the uniformly light reddish brown massive siltstones. The mottles are infrequent and usually randomly dispersed, although occasionally several of them lie parallel to the bedding. In the Horseshoe Bend Shale, the mottles vary from random types to elongate lensoid pods up to 40 cm long (Fig.3.1). The mottles represent reduction spots in the otherwise
oxidized sediment and their mode of formation is uncertain. Possibly they are "pseudo-gleyed" sediments resulting from localized changes in permeability.

SEDIMENTARY FEATURES OF THE SANDSTONE UNITS

Sedimentary features within the sandstone units are more varied than in the siltstones. The most prominent features are erosional contacts and internal structures such as planar and cross-stratified bedding. Small-scale features are also common and include many varieties of current lineation, ripple marks, hydroplastic deformation structures, pebble and mudclast fabric, and sole markings.

CROSS-STRATIFICATION

Cross-stratification is the most prominent depositional structure present in the sandstone units of both the Pertnjara and Finke Groups. There are a great variety of types present but the most abundant are isolated sets and cosets of trough cross-strata. Cross-stratified sets comprise between 60 and 100 percent of the sandstone succession and individual bedsets range from a few centimetres to over 6 metres in thickness. The Harajica Sandstone Member is somewhat distinct in the southern exposures near Mereenie Anticline where
the bedding is dominantly small scale and laminated with only minor horizons and lenses of cross-stratified sands.

The individual cross-bed types are described below in the order given by Allen (1963b, Fig.3.26) without respect to their importance and relative abundance. The terminology has been modified slightly to distinguish distinctive structural variations recorded in this area.

(1) **Plane-Beta-Cross-Stratification**

Plane-beta-cross-stratification is a term used to distinguish beta-cross-stratification composed of plane cross-strata in sections perpendicular to the strike.

This type of stratification is rare within the Purtinja and Finke Groups and was only seen at a few localities within the Hermannsburg Sandstone. One 75 cm thick set is shown in Figure 3.27 where the basal and upper contacts are almost planar erosion surfaces. The sets are all medium scale and the lithologically homogeneous cross-strata are discordantly related to the lower boundary surface of the set. In plan, the cross-strata usually appear to be straight or very gently curved.

The dip of the cross-strata is variable and the high angle forms were probably built by solitary banks or bars of well sorted sand, migrating downstream. In such cases, the structure consists entirely of topset and foreset bedding with an angular contact between the foreset strata and the base of the set. The low angle plane-beta-cross-stratification illustrated in Figure 3.28
has a maximum dip angle of about 10° which is considerably less than the angle of rest of the grains. The origin of this bedding form is harder to explain and it may represent a semi-lateral accretion deposit on the banks or bars in a stream system. It could possibly represent a prograding beach deposit but no evidence of lacustrine deposits was present in a downcurrent direction. The presence of mud-flakes on some of the bedding planes and near the base of the bed indicates that the bedform was not a result of aeolian action.

(2) Curved-beta-cross-stratification

Curved-beta-cross-stratification is a term used to distinguish solitary, tabular, cross-stratified sets of any size where the cross-strata are concave upwards (Fig. 3.29). The size of these sets varies from small to medium scale and they are most abundant in the 5 to 25 cm range. Large scale beta-cross-stratification is absent in the sandstones studied but it may be the poorly exposed stratification present in the very coarse conglomeratic sandstones of the Brewer Conglomerate.

Sedimentary structures described as beta-cross-stratification may locally merge into and include less frequent alpha- and gamma-cross-stratified sets with non-erosional planar or erosional irregular basal contacts respectively.

Beta-cross-stratification is a fairly common bedform developed in all sandstone units of the Pertnjara and Finke Groups. The grain size of individual cross-strata
is usually uniform within a set although it may vary from fine sandstone to very coarse pebbly sandstone in different sets. Pebbles and mudclasts, where present, are usually aligned parallel to the cross-stratification and concentrated towards the base of foreset bedding. The maximum foreset bedding angle is generally in the range from $15^\circ$ to $30^\circ$ with the majority of beds in the $20^\circ$ to $25^\circ$ interval. In all cases, the angle of foreset bedding decreases towards the base of the set (Fig.3.30) and the cross-strata are tangential to the lower boundary. Topset beds are rarely preserved.

Small scale beta-cross-stratification is less frequent and usually occurs in association with flat bedded sandstones (Fig.3.31).

According to Allen (1963b) alpha-, beta- and gamma-cross-stratification probably form by a similar mechanism of downstream migration of solitary banks or transverse bars. Alpha forms would be deposited under relatively tranquil non-erosive conditions while the beta and gamma forms probably accumulated more swiftly under variable flow regimes which were strong enough to erode the underlying sediment.

Beta-cross-stratification could also originate from small deltas or bars migrating downstream. The experimental work of Jopling (1963, 1965) has indicated that the slope and basal contact of a delta front is dependent on sediment size, depth of water on downstream side of delta and the velocity of flow. At constant depth, low flow conditions favour the deposition of
steep planar foresets with an angular basal contact whereas higher velocities, which transport more sediment in suspension, produce tangential basal contacts and may develop into low-angle concave cross-strata. For constant flow, increasing depths on the downcurrent side of the depositional interface promote a change from concave to planar cross-stratification. Thus large scale concave cross-stratification requires higher flow conditions for its formation than does small scale cross-stratification.

The majority of both solitary sets and cosets of tabular cross-stratified sandstone in the Pertnjara and Finke Groups have low to moderate angle concave cross-strata and were probably deposited from moderate to high velocity currents within the lower flow regime. Some of the thicker sets have planar cross-strata and tangential contacts but this is a result of their size and probably does not represent a reduced flow condition.

(3) **Theta-cross-stratification**

Medium scale solitary trough cross-stratified units, with the boundary shape of eta-cross-stratification but filled in the same manner as pi-cross-stratification, have been found at several localities within the Hermannsburg Sandstone. They probably represent incompletely preserved, elongate theta-cross-stratified sets.

The theta-cross-stratified units generally have a regular, low angle, scoop shaped, basal erosion surface cut down into laminated, plane bedded sandstones (Fig.3.32).
There may be more than one theta-cross-stratified set at the same stratigraphic horizon, each separated by a short interval of non-eroded substrata. In cross-section the troughs appear to be concordantly filled with homogeneous sand strata which gently plunge down the axis thus giving a discordant basal relationship when seen in longitudinal section. The thickness of the solitary sets varies from 20 cm to 1 m and their width correspondingly varies from 1 to 7 m, although the larger varieties are more common. The theta-cross-stratified sets are usually overlain by plane-bedded or tabular cross-bedded sets.

Theta-cross-stratified sets represent infilled erosion channels cut into the underlying sediment. The scour and fill could occur as distinct acts separated in time (Shrock, 1948). However, Stokes' (1953) postulate of simultaneous erosion on the down-current face and deposition of cross-strata on the up-current face of an erosional hollow, formed by a moving eddying mass of water, is also a feasible mechanism for the formation of theta-cross-stratification.

(4) Kappa-cross-stratification

Kappa-cross-stratification is an infrequent structure in the sandstones of both groups but is most prominent in the basal part of the Hermannsburg Sandstone in the Dare Plain area. It is represented by irregular cosets of climbing, homogeneous, fine sand ripples with minor erosion between some of the ripples (Fig. 3.33). In cross-section, the sets appear to be
lensoid. They can be termed ripple cross-laminated sands and according to the classification of Jopling and Walker (1968) they are transitional between type A and type B cross-lamination. According to Walker's earlier classification (1963), they can be termed Type 2 cross-laminations. They typically occur in the upper parts of the cyclic sedimentation unit and the cosets vary in thickness from an average of 5 to a maximum of 40 cm.

According to Allen (1963a), such stratification develops by the subaqueous migration of trains of small-scale asymmetrical linguoid ripples where there is an abundant sediment supply. Jopling and Walker (1968) also propose migrating ripples but consider the sediment supply originates from the interplay of both traction movement of a grain carpet and suspension load deposition.

(5) Omikron-cross-stratification

Cosets of tabular or omikron-cross-stratification are fairly common in all the sandstone units although they are generally of a small to lower medium scale. Commonly, there is a progressive upwards decrease in thickness of the individual cross-stratified units within a coset. The basal set in a coset may reach a thickness of 1.5 to 2 m (Fig. 3.34) but most tabular cross beds are only 15 to 50 cm thick. Each set is underlain by an essentially planar or slightly curved erosion surface and the boundary between the cross-strata and the base of the set is typically tangential. The cross-strata are usually composed of homogeneous sand laminae although the inclusion
of small rounded pebbles and flat angular mudclasts is frequent, especially in the Missionary Plain area.

There are probably two main modes of formation of omikron-cross-stratification (Allen, 1963b). They could form by the building of solitary transverse banks or bars across one another in shallow but continuously rising water depths. However, Allen (1963b) considered that they probably formed by the migration of trains of straight crested asymmetrical medium to large scale ripples in moderately deep water, where the sediment supply was not sufficient to prevent erosion of the stoss slope of the preceding ripple.

(6) Pi-cross-stratification

Pi-cross-stratification is the most abundant structure in all the sandstone units. It usually forms at least 50 percent of the succession and in local areas it is developed to the exclusion of all other stratification types. Pi-cross-stratification occurs as cosets of variable thickness with individual sets ranging from large scale to small scale - the latter merging into mu-cross-stratification.

In most of the sequence individual cross-stratified sets are medium scale or smaller. The cosets may overlie an irregular unconformity surface (usual, e.g. Fig.3.35) or a plane bedded unit. The degree of erosion beneath the basal cross-strata is very variable and is related to the thickness of the overlying set. The maximum recorded is about 1 m (Fig.3.36). Pebbles
and mudclasts, where present, usually accumulate towards the base of the foreset beds near the erosional surface. The width of medium scale sets is variable and up to 10 m in very shallow sets (Fig.3.37). Axial length of the trough sets is usually hard to measure but values up to 25 m have been recorded in good exposures. More commonly measured lengths are from 4 to 12 m. In cross-section, the cross-strata may be concordant or tangential to the lower bounding surface (Fig.3.38) while in longitudinal section the cross-strata appear as low angle usually asymptotic laminae (Figs 3.39 & 3.40). The vertical sequence within a coset may show a uniform size of individual sets, or much more commonly there is an upward decrease in the size of individual sets (Figs 3.38 & 3.39). This is attributable to a decrease in current velocity and competency due to waning currents or lateral migration.

The gradation or continuum of pi-cross-stratification from small scale (mu-cross-stratification) to medium scale (up to 2 m) suggests that the mode of deposition is similar for both forms. Allen (1963b) reviewed the possible modes of formation of pi-cross-stratification and suggested that most units originate by the downcurrent migration of trains of asymmetrical ripple marks with pronounced curved crests. Similar results were found by Williams (1968). In 1968, Allen elaborated on this mechanism and showed that trough cross-stratification can be generated by all variations of ripple types from quasi-three-dimensional straight
ripples to lunate ripples. Troughs with symmetrical fill reflect troughs aligned with the current flow, whereas asymmetrically filled troughs are generated by ripples building out obliquely to the current flow.

The unusual laminated, convex bedding, erosionally overlying the trough cross-strata in Figure 3.40, is similar to that described by Reineck (1963). He considered that it formed on intertidal beach bars during periods of reduced flow across the bar. A similar situation can be envisaged on fluvial deposits subject to waning current strengths.

(7) Combined Omikron-Pi-cross stratification

In localized areas such as the upper portion of section CY, pi-cross-stratification varies from small to large scale and is marked by the general uniformity of bedding direction and sedimentary features and the angle of the cross-stratified sets. The lower bounding surface truncates the underlying sets and is usually a very broad, gently curved or planar, smooth surface (Fig. 3.41). The sets vary from very low angle broad trough sets to high and low angle tabular sets with slightly concave lower bounding surfaces. Set thickness varies from 1 to 3 m, with occasional units of generally steeper dipping strata up to 6 m thick. The width of the units is also variable and the largest one measured south of Black Hill Range was 30 m wide (Fig. 3.42). The cross-strata within a set may have concordant, discordant or tangential basal contacts —
the first one being the most common in section CY (Fig.3.43). The sands in these cross-beds are all well laminated and usually plane bedded (Fig.3.44). Only rare instances of oversteepened bedding was recorded from steeper dipping sets. The maximum dip angles for cross-strata were obtained from tabular cross-bedded sets (20° to 35°) while from the low angle planar bedded units, dips are usually less than 10°. Not all the cross-stratification in these sets is large scale. Several instances of small scale sets showing planar erosion surfaces and occasional dip reversals were recorded (Fig.3.45).

The sediment comprising the individual cross-strata has a uniform, fine to medium grain size, is moderately sorted and has a relatively high content of quartz. The units are characterized by an absence of all very coarse grained detritus especially pebbles and mudclasts.

The features described from these sequences of omikron-pi-cross-stratified sets are similar to features described for dome shaped dunes at White Sands National Monument, U.S.A. (McKee, 1966). The steeper tabular cross-stratification probably represents accumulation as slipface deposits on an advancing dune. The characteristic feature of dome shaped dunes is the abundance of low angle strata (3-10°) which may be continuous planes, or a combination of topset and very low angle foreset laminae. Dome shaped dunes are also characterized by trough shaped scours cut and filled in
the upper part of the dune. McKee (1966) concluded that dome shaped dunes develop in exposed localities under the action of one dominant wind direction. They form where dune height is inhibited by unobstructed strong winds eroding the windward face of the dune and moving much of the sand in suspension to the lee side where it accumulates in moderate to low angle laminae.

(8) Micro-deltas

Features termed micro-deltas have been distinguished in several exposures of the lower Hermannsburg Sandstone. They occur as isolated sets of inclined strata 20 to 40 cm thick and usually lying above a smooth gently concave erosional base. They are typically overlain by a massive or obscurely laminated sandstone.

The internal structure of the micro-delta is the same as classical large scale deltas with flat lying bottomset beds, concave upward foreset beds which are tangential to the base, and gently inclined topset beds (Figs 3.46 & 3.47). The topset beds slope at about 5° to 10° and can be traced continuously over the break in slope. The foreset beds are thicker than the corresponding topset bed and are usually inclined at 20° to 30° to the horizontal. They become progressively thinner away from the delta front and are continuous, with decreasing slope, into the bottomset beds. The most important feature of this bed-form is that as the delta progrades the foreset beds, due to their relative
thickness, loose their identity. The depositional sur-
face becomes almost planar from the top to the bottom
of the bed and with continued deposition this is replaced
by a convex upwards bedding plane with a sharply tan-
gential basal contact (Fig. 3.48). Large scale convex
bedding is confined to aeolian deposits (Conybeare &
Crook, 1968) while small scale examples have been
described from bar-finger sands of the Mississippi
Delta by Fisk (1961). No described medium scale
examples are known to the author.

This type of structure is assumed to represent
subaqueous deposition from a small scale transverse bar
migrating down current into a scour pool. The transverse
bar maintains its leading edge until the scour pool
begins to shallow appreciably when the effectively
raised base level alters the sedimentational regimen by
altering flow conditions from tranquil eddies to rapid
turbulent conditions.

(9) Oversteepened and Contorted Cross-Stratification

Oversteepened cross-stratification (Fig. 3.49)
is more frequent than contorted cross-stratification
(Fig. 3.50) but both have a sporadic occurrence in the
sandstone formations.

The oversteepening or contortion usually
occurs in a restricted interval within any one cross-
stratified set. They typically originate near the top
of the foreset laminae, attain a maximum dip or
amplitude in the middle of the foreset bed, and reduce
in intensity near the bottomset beds. The oversteepened and contorted intervals are usually 5 to 25 cm thick and the overlying normal sedimentary units thicken to fill the irregular hollows of the disturbed strata. The features must have been formed during deposition since they are overlain and underlain by undisturbed cross-strata with similar characteristics.

Larger scale contorted stratification is quite common in the Harajica Sandstone Member on Dare Plain. It may involve up to 1.5 m of laminated sandstone (Fig.3.51) and may be termed convolute bedding.

In all cases of contorted and convoluted stratification, the individual laminae are continuous and the internal features of the sediments are preserved. For this class of deformation to occur, the sediment is said to be in a hydroplastic rheotropic state (Elliott, 1965). The structures were probably formed by gravity slumping in conjunction with overloading.

PLANE BEDDING

Plane bedding is a common feature in all the sandstone units and is frequently interbedded with cross-stratified cosets. The plane bedded sandstones vary in thickness from a centimetre or less to two metres with most bedsets in the 10 to 50 cm range (Fig.3.52). Plane bedding may exhibit very low amplitude undulations although this is not common. Associated streaming lineation and parting lineation
is common.

Plane beds of two types have been recognized in
the sandstones. Thicker sets generally overlie a wide-
spread planar erosional surface with a sharp junction
(Fig. 3.53). These beds vary in grain size from fine to
very coarse with the majority of them being medium to
coarse. Occasional coarser grained, plane bed deposits
contained mudclasts aligned parallel to the bedding,
and small (less than 2 cm) rounded pebbles. Bedding
planes in these sets generally exhibit current lineation.
The other type of plane bed deposit is usually in thin-
er sets and either erosionally or gradationally over-
lies medium to small scale cross-stratified sets. It
may be associated gradationally with ripple marks and it
is frequently overlain by wavy, laminated beds and
laminated siltstones. Current lineation was occasionally
present within this interval of plane bedded sands.

The bedding types are characteristic bedforms
developed under two different flow regimes. The former
type is typical of the upper flow regime plane bed
deposits, while the latter represents deposition from
the lower stage of the lower flow regime (Simons &
Richardson, 1961).

MASSIVE BEDDING

Apparently massive bedded sandstones are quite
numerous in the Hermannsburg Sandstone (Fig. 3.54), especially from a distance. On close examination, many of the massive beds show traces, often very poorly discernable, of plane bedding or rarer cross-stratification, e.g. alignment of mud flakes in Figure 3.55. Massive bedding is probably developed by rapid continuous deposition of well sorted sediments over a short time interval. It probably results largely from plane bed deposition in the upper flow regime, although very low angle trough cross-strata may also be involved. The occurrence of massive bedding is accentuated by a lack of differential weathering of the individual laminae and thus it appears more common in the sequence than it actually is.

EROSIONAL SURFACES AND CHANNELS

Erosion surfaces between bedding sets are extremely frequent and form the most common type of local contact between sedimentation units. They may be either planar (Fig. 3.53) or irregularly erosional (Fig. 3.35) and they are discussed in terms of the overlying beds. Regular concave erosional surfaces beneath isolated trough cross-stratified units (Fig. 3.37) are not as common.

Infilled erosional channels are rarely preserved as isolated recognizable features. They are relatively more common in the Langra Formation where they occur as small steep sided erosional features extending from
20 to 75 cm into the underlying sediment (Fig. 3.56). They usually have an asymmetrical cross-section, may show undercutting of one bank, and are commonly infilled with poorly sorted pebble conglomerate. They are normally associated with a more extensive irregular erosion surface (Fig. 3.57) which may contain other pockets and lenses of pebble to cobble conglomerate. They probably formed as isolated channels in a river bed during periods of low flow.

The only large distinct erosional channel seen in the Pertnjara Group was at section CS. It consists of an asymmetrical channel cut 1 m down into the underlying fine silty sandstones and interbedded siltstones of the basal Hermannsburg Sandstone (Fig. 3.58). The greatest exposed width of the channel is 2 m. The channel is filled with trough cross-stratified medium sandstones, with rare mudclasts at the base, which become finer grained towards the top of the structure. The upper fine sandstones of the channel infilling can be seen extending laterally beyond the channel margin as a thin sandstone horizon (Fig. 3.59). The shallow side of the channel also shows the lateral extension of lower units of the channel sand into the adjacent flat-lying deposits. This feature undoubtedly represents an isolated, curved, alluvial channel flowing through a frequently flooded alluvial plain. The lack of distorted bedding, root casts and soil profiles indicates that few, if any, plants were established on the flood plain. Desiccation cracks and
Ripple marks are frequent in the interbedded fine sandstone-siltstone overbank deposits.

RIPPLE MARKS

Ripple marks in the sandstone units are relatively common in the finer grained sediments at the top of sedimentary cycles (see later). They are almost exclusively asymmetrical with only very rare symmetrical or interference forms. The range of forms is indicated in Table 3.2 although the proportions recorded in the table are not representative. The most common ripple types are elongate linguoid ripples typical of current deposition, while less abundant types are lunate or sinuous.

(1) Linguoid Asymmetrical Ripple Marks

Linguoid asymmetrical ripple marks are the most abundant form preserved in sandstone units, especially where not directly overlain by siltstone. They may be preserved as ripples (Fig.3.60) or ripple moulds (Fig.3.61) and they are generally elongate transverse to the current direction. The ripples vary from true linguoid shape to an irregularly cuspate, straighter ripple form. The wave length of the ripples is usually in the range from 5.0 to 9.5 cm. The ripple index commonly varies from 8 to 13 and the ripples are markedly asymmetrical (ripple symmetry index 3-6). The ripples are composed of foreset laminae which have been eroded on the stoss side. In Figure 3.62 original linguoid
asymmetrical ripples have been modified by the deposition of uniform laminae covering the whole ripple form. The latter probably accumulated from suspension under lower current velocities.

Table 3.2 gives the Tanner (1967) indices for linguoid ripple marks and in all cases they fall in the current ripple fields. Such ripple marks form by the action of traction currents in shallow, comparatively turbulent and swift flowing water (Allen, 1963a). They are known to occur in intertidal environments and are common in stream bed and overbank fluvial deposits (Allen, 1968).

(2) Lunate Asymmetrical Ripples

Lunate asymmetrical ripples are the least common ripple form found in the sandstone units. They have been recognized by their shape and the direction of cross lamination (Fig.3.63) and they are small scale analogues of the larger forms described by Allen (1963a). Their general dimensions are width 9 cm, wave length 12 cm, amplitude 1 cm and degree of crest closure approximately \(-150^\circ\).

Small scale lunate ripple marks are rare in the recorded literature but their mode of formation is probably similar to that of the large scale forms and barchan dunes as described in Allen (1963a). The latter form by traction movement when sediment supply is insufficient to maintain an adequate coverage of the depositional surface. However, in both recorded instances from the
Hermannsburg sandstone, this condition was only marginal and the lunate ripples are associated with straighter or linguoid asymmetrical ripples.

From the ripple indices, the lunate ripple marks fall into the current ripple field on all graphs which suggests that they were formed by tractive currents.

(3) **Sinuous Asymmetrical Ripple Marks**

Sinuous asymmetrical ripple marks show a wide variety of form in straightness, continuity and ripple symmetry. They vary from almost straight regularly asymmetrical forms (Fig.3.64) to irregularly sinuous forms with variable ripple symmetry (Fig.3.65).

The ripples may have sharp crests, reasonable ripple symmetry, open junctures (Allen, 1968) but lack bifurcations, as shown in Fig.3.66. These forms have many of the characteristics of wave formed asymmetrical ripples as defined by Tanner (1967) although the presence of ripple terminations rather than bifurcations suggests that formation by current flow cannot be ignored. They could have formed in a shallow, slow flowing, aqueous environment with ripples developed either by wind or water generated wave motion.

More regular ripple marks with higher ripple symmetry indices and rounded crests (Fig.3.64) are also relatively common especially in the sandstones immediately overlying the Parke Siltstone. Bifurcation, or zig-zag juncture (Allen, 1968) is quite common in these ripple marks and the split crests may be either
the same strength as the original or one branch may be stronger. The former is shown in the top left and bottom centre of Figure 3.64 while the latter is clearly demonstrated to the right. The bifurcation index of Tanner (1967) varies from less than unity in the upper left to 2.3 on the right of Figure 3.64 but larger indices are also apparent. The ripple indices for the figured set are given in Table 3.2 (sample DF 7) and these suggest that the ripples could have formed either by wave or wind action. However, the former is favoured because of the associated structures such as mud-cracks in the overlying siltstones.

The most common forms of sinuous asymmetrical ripple marks are rather irregular in lateral extent and may bifurcate or terminate (Figs 3.65 & 3.67). The ripple symmetry index of these forms is generally greater than 4 and the combined indices indicate depositional environments varying from current to wind or wave. The ripples shown in Figure 3.67 are partially filled with silt and the terminations shown in the centre left of the photo are probably post-ripple-formation erosional features associated with transverse drainage. They probably represent depth-biased, wave formed, asymmetrical ripples subjected to very shallow water or subaerial exposure.

An unusual set of sinuous asymmetrical ripple marks was seen in a shallow channel structure. A mould of the ripple marks is seen in Figure 3.68 and shows an interesting relationship between slightly asymmetrical
channel shape and ripple mark size. On the deeper side the ripple wave length is approximately 3 cm and the ripples are fairly uniform, whereas on the shallower side the wave length is only 1.5 cm to 2.0 cm and the ripples are far less regular. Ripples are virtually absent from the axis of the channel but this could be due to erosion. Another small channel containing ripple marks in section DS shows a slightly different relationship with fairly regular ripples along the axis which bifurcate towards the side of the channel (Fig. 3.69).

Allen (1968) notes that straight and sinuous small scale, asymmetrical ripple marks may form in tidal flats or in fluvial environments where the current flow is low and smooth.

(4) Interference Ripple Marks

These are not a frequent feature in the sandstone units and they usually only occur where the sandstones are overlain by a siltstone horizon. This, in itself, suggests decreasing current action and consequent suspension deposition in a standing body of water. Figure 3.70 shows quite a large area of regular interference ripples superimposed on a set of sinuous asymmetrical ripples. The nature of the nodal contacts suggests that the interference was caused by stationary waves in a bidirectional wave pattern system. Similar but more regular interference ripples (Fig. 3.71) were seen above linguoid ripples in section CJ. These interference ripples represent two intersecting wave patterns and
are overlain by a thin horizon of mud-cracked siltstone.

(5) **Wash-over Crescents**

These are a rare form of flat topped ripple only seen in section AK. They consist of crescentic depressions elongated, with decreasing amplitude, on one side (Fig. 3.72). Wash-over crescents have been described by Tanner (1960) on shallow water sand bars awash during falling tides. The flat topped nature of the ripple marks suggests planation in water depths of 2 or 3 cm. Wash-over crescents have not been described previously from fluvial sediments but they could probably form in an analogous manner during periods of falling water level.

**INTERNAL HYDRODYNAMIC STRUCTURES**

Internal hydrodynamic features are common in the plane bedded sands but relatively rare in cross-bedded units. Current lineation is the most common feature and may be present over a wide lateral extent on any one bed. Other features include current crescents, harrow marks and sand ridges but they occur infrequently.

(1) **Current Lineation and Crescents**

Streaming lineation (Conybeare & Crook, 1968) and parting lineation are common features of planar
bedded sandstones in the Hermannsburg, Langra and Idraccwra Sandstones, although suitable bedding plane exposures are infrequent in the latter two. Current crescents are always associated with streaming and/or parting lineation in the Hermannsburg Sandstone although they are never frequent in any one outcrop. Figures 3.73 and 3.74 each show one current crescent on a large flat bed exposure while Figure 3.75 shows typically developed parting lineation. Streaming and parting lineation commonly show a divergence of up to 10° in direction on any one sedimentation surface, and may show a 30° variation between various surfaces in any one planar bedded sedimentational unit. Streaming and parting lineation, by themselves, only give the orientation of current flow and not the sense. Current crescents, where present, indicate the sense of the current flow since they are elongated in a down-current direction. They usually consist of two shallow linear grooves and an intervening sand shadow extending a short distance down-current from the obstruction. Karcz (1967) ascribed the formation of similar structures found in ephemeral streams to secondary currents and vortices which originate when the flow is deformed by obstacles in the streampath.

Allen (1964a) has shown experimentally that current lineation in sands forms in the plane-bed stage of the upper flow regime while Sorby (1908) noted from field evidence that it was more characteristic of the plane bed phase of the lower flow regime.
In the present study, current lineation occurred more frequently in assumed lower flow regime plane-bedded units but it was also noted in the coarse and sometimes pebbly plane-bedded units of assumed upper flow regime deposition. The lower frequency of the latter form of current lineation can be attributed to the relative scarcity of upper flow regime plane-bedded deposits. The lack of experimental evidence for the formation of parting lineation noted by Conybeare & Crook (1968) is not surprising since it requires consolidation of the sand prior to splitting along a bedding plane.

(2) Harrow Marks

Possible harrow marks have been recorded in section CQ (Fig. 3.76). They consist of elongated rounded ridges separated by shallow troughs (10 cm wide and 1 cm deep) containing slightly coarser sediment. They are aligned approximately parallel to current lineation in the same bed and appear to merge laterally into flat bedded sandstones. Karcz (1967) has described similar recent structures and suggested that they formed by the action of regular systems of longitudinal helical flow patterns with alternating senses of rotation.

HYDRODYNAMIC SOLE MARKINGS

Except for large scale erosional features which may be either planar, concave or irregular, hydrodynamic sole markings are rare within the sandstone units. They
are all erosional features aligned parallel with the current flow and no tool marks were seen.

(1) **Flute Marks**

Small flute marks are infrequent and are usually scoured into muddy underlying sediments (Fig. 3.77). In Figure 3.78 very small, almost symmetrical flute moulds are preserved on the base of a sandstone overlying a thin mud cracked siltstone. The only example of a larger scale flute mould was seen in the Ooraminna Sandstone Member at Orange Creek (section BP). It is an elongate (17 cm), symmetrical structure which sharply tapers towards the deeper upcurrent end and becomes broader and shallower down-current. The origin of flute marks is attributed to the erosive action of either sub-horizontal or sub-vertical vortices impinging on relatively soft but cohesive sediment (Dzulynski & Walton, 1965).

(2) **Rill Marks**

Casts of enlarged rill marks or small erosional channels have been preserved at a few localities within the Harajica Sandstone Member on Dare Plain. They consist of roughly parallel elongate channels which show infrequent junctions (Fig. 3.79). The channels vary in width from 2 to 5 cm and when curved they show an asymmetrical cross-section up to 1.5 cm deep. They are filled with plane laminated sand. These channels probably formed by rivulets of water flowing down an
inclined plain – possibly a beach or bar face.

RHEOTROPIC STRUCTURES

Rheotropic structures are not preserved frequently in the sandstone units but do occur in restricted localities. The exception to this rule is the widespread occurrence of mud cracks in thin, interbedded siltstone horizons. The terms used in defining the rheotropic state of deformation are those proposed by Elliot (1965).

(1) Sand Volcanoes (Quasi-liquid)

Sand volcanoes were seen at one locality near the base of the Harajica Sandstone Member in Glen Helen Gorge (section AJ). In sections parallel to the bedding plane (Fig. 3.80) they appear as exfoliating, sub-circular mounds while in cross-section (Fig. 3.81) they consist of vertical pipes of disturbed material up to 10 cm in diameter, flanked by radially dipping laminae. They probably formed by the upwelling of water through quicksand.

A similar but small scale feature was seen in ripple bedded sands of the Harajica Sandstone Member on Dare Plain (Fig. 3.82). The central vent of sandstone is clearly exposed and breaches the dark silty horizons. The latter must have been in a quasi-solid rheotropic state prior to the "eruption" because they exhibit fracture deformation accompanied by slight
bending. The feature occurs in the trough between two ripple crests and the elevation difference may have been sufficient to build up hydrostatic pressure to the extent that rupture of the overlying sediments took place.

(2) **Slump Structures (Hydroplastic)**

Small scale slump structures were noted in the upper Harajica Sandstone Member in interlaminated fine sandstones and siltstones (Fig.3.83). The origin of this form of slump bedding is probably gravity induced movement on an oversteepened depositional surface. Similar structures have been noted from the upper parts of fluvial point bars after the recession of high water levels (Conybeare & Crook, 1968).

(3) **Load Casts (Hydroplastic)**

Post-depositional hydroplastic deformation features are relatively abundant where siltstone interbeds or large mudclasts are encountered. They usually consist of load casts preserved in the overlying sandstone and they are frequently oriented in a subparallel fashion, e.g. Figures 3.88 and 3.89. They probably formed as a result of loading on pre-existing linear features such as flute casts. This is especially evident in the mudclast mould in Figure 3.90. Of course, in mudclasts such linear features would probably have formed prior to their transportation and are therefore invalid current direction indicators. Load casts
with the general appearance of distorted ripple marks (Fig. 3.91) were also seen occasionally. They may have formed by loading on pre-existing ripple marks in the siltstone, or by imprinting ripples formed in the overlying sand (Conybeare & Crook, 1968).

(4) Mudcracks and Clastic Dykes (Quasi-solid)

Mudcracks are frequently seen, associated with thin laminated siltstones, in the Harajica Sandstone Member, the Hermannsburg Sandstone and the Langra Formation while they are infrequent (possibly due to exposure) in the Idracowra Sandstone.

The mudcracks vary considerably in size but predominantly consist of 3 to 5 sided polygons. They may be preserved as moulds in the underlying sandstone which are either flat or concave upwards (Figs 3.63 & 3.92). However they are most frequently preserved as casts on the underside of the overlying sandstone bed (Fig. 3.93). The mudcracks are frequently separated by clastic dykes of the overlying sediment and this is well shown by the large scale mudcracks in Figure 3.94.

The development of mudcracks during the deposition of the Harajica Sandstone Member and the Hermannsburg Sandstone must have been much more frequent and widespread than the ones preserved would indicate, since mudclasts are common throughout both units.

Mudcracks in the sandstone units are thought to represent desiccation cracks rather than synaeresis
Cracks largely because of their vertical association with presumed fluvial deposits. The distinction between the two types of cracks proposed by White (1961) does not appear to hold in this situation because although the siltstone is usually fissile, the mud-cracks occur between sandstone units, they may be parallel sided or wedge-shaped in the same unit, and there is frequently little evidence of curling.

(5) **Rain Prints (Quasi-solid)**

Rain prints are infrequently preserved sedimentary features which were seen at a few localities but were only well preserved in the basal Hermannsburg Sandstone in section DH. They are shown preserved in a sandstone bed overlain by siltstone (Fig. 3.95) and are slightly asymmetrical to the east. The raised external rim and crater are especially well preserved in the central impression while coalescing imprints are seen just above it. In most exposures, rain prints are poorly preserved (e.g. Fig. 3.96) and their recognition as such is based on the random distribution of imprints and the lack of internal sedimentary structures in the sandstone directly beneath the imprint.

**INCLUDED CLASTS**

Clasts included within the sandstone units are frequent especially along the northern margin of the basin. Mudclasts occur with variable frequency throughout the whole succession whereas pebbles have a more
restricted distribution. They are frequent in lenses throughout the Langra Formation but they are rare in the Idracowra Sandstone, lower Hermannsburg Sandstone (except for the base of the Ooraminna Sandstone Member), and Harajica Sandstone Member and only become abundant in the upper Ljiltera Member north of Missionary Plain.

(1) Mudclasts

Mudclasts are common features throughout the sandstones of the Pertnjara Group although they are less common in the Finke Group. They are most abundant in sandstones overlying widespread erosional surfaces, especially near the base of cosets of cross-strata. Very large mudclasts may be concentrated in localized pockets or may occur as breccias in erosion hollows (Fig.3.97). The largest clasts seen were about 30 x 20 x 5 cm, usually angular, and either flat (Fig.3.98) or slightly curved. Large clasts frequently show load cast features (see above). Large mudclasts may be deposited either as disjunct edgewise breccias in pockets or less frequently as isolated clasts aligned parallel to the bedding planes of cross-stratified sets. Smaller mudclasts are much more abundant than the larger varieties and they typically occur as isolated flakes rather than breccias. They are usually less than 10 cm on the long axis and frequently show slight signs of rounding. They are aligned parallel to the strata in which they occur and in cross-stratified sets they are most common towards the base of the foresets. Imbrication of mudclasts is rare in these sediments.
Mudclasts almost certainly represent transported silty laminae which had previously been broken into fragments by mudcracking. The angular nature of the larger clasts indicates very little transport while smaller mudclasts show abrasion rounding. Fagerstrom (1967) noted that mudclasts could be transported by flotation on algal mats in still water but this process would probably not be important in water sufficiently turbulent to produce the cross-bedding characteristic of these formations.

(2) Pebble Shapes

The majority of pebbles in the sandstone units are subrounded to rounded and subspherical (Fig. 3.99) although present solution features make roundness determinations on limestone clasts difficult. The long apical bisectricies of one suite of 40 discoidal pebbles from both the Langra Formation and the Ljiltera Member were determined by the method of Lenk-Chevitch (1959).

The results obtained were:

<table>
<thead>
<tr>
<th></th>
<th>Langra Fm</th>
<th>Ljiltera Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>boomerang-shaped bisectors (predominantly beach)</td>
<td>14</td>
<td>9</td>
</tr>
<tr>
<td>sigmoidal or straight bisectors (predominantly fluvial)</td>
<td>18</td>
<td>19</td>
</tr>
<tr>
<td>unclassifiable (too equant)</td>
<td>8</td>
<td>12</td>
</tr>
</tbody>
</table>

The predominance of fluvial forms is not overwhelming although most other sedimentary features point to a fluvial mode of origin. In most cases, however, the boomerang-shape is not very marked, as can be seen in the casts in Figure 3.100 and it may point to the fact that discoidal
pebbles, in a stream that is incapable of moving them, may be eroded by sand flowing around the ends of the pebble.

Broken rounds are not very common features in these sediments although in the Hermannsburg Sandstone broken rounds of laminated sandstone or shale were occasionally noted. The occurrence of a few broken rounds of etched vein quartz grains in the middle of the Langra Formation is hard to explain since percussion marks do not appear to be preserved on them (Fig. 3.101). They may have been split by frost action or diurnal expansion and contraction along planes of weakness (Sugden, 1964).

Many of the vein quartz pebbles in the Langra Formation are well rounded and have a peculiar etched surface pattern.

(3) Imbrication

Imbrication is a rare feature in the sandstone units of both groups. The pebbles are all rounded and isolated while the mudclasts are predominantly aligned parallel to the stratification. Large mudclasts occasionally show signs of imbrication in some apparently massive bedded sandstones, e.g. Figure 3.55.

An interesting feature in the basal fine sandstones of the Hermannsburg Sandstone at section CU is the up-current imbrication of slightly muddy sandstone clasts. These clasts form a continuous pavement of blocks rotated 5° to 20° in the up-current direction without any apparent lateral displacement (Fig. 3.112). The cohesion within the sandstone clasts would have been weak because of the low
silt-clay matrix and one must therefore postulate scour by low velocity currents. The break-up and rotation of the original sandy horizon cannot be attributed directly to gravity sliding since the bed is horizontal, although the effect of mass movement further up-current, causing localized horizontal hydrostatic stress in waterlogged sandstones immediately beneath the affected bed, cannot be overlooked. The lack of lamination for 3 cm beneath the imbricated bed might support this hypothesis.

Grain imbrication is more common and is especially evident in micaceous sandstones where large mica flakes are imbricated in an up-current direction.

(4) Pebble and Grain Trains

Pebble trains are relatively infrequent in the pebbly sandstones of the Pertnjara and Finke Groups but this is probably the result of infrequent pebbly bedding plane exposures. However, they were occasionally seen in cross-section (e.g. Fig. 3.103) where a large pebble or boulder has a downstream tail of smaller pebbles. The feature is equivalent to sand shadows and was formed by a similar eddying current system in the lee of a large clast. Where present they are a useful current direction indicator.

Similar features of a train of smaller grains behind a large one have been noted in several thin sections cut parallel to the current direction and normal to the bedding. These features have been termed grain trains in this study and they are small scale analogues of pebble
trains. They occur, or are recognizable, in poorly sorted sandstones and are aligned parallel to the bedding (Fig. 3.104). They may show a small degree of up-current imbrication (Fig. 3.105) and the largest grain is usually oriented in a stable position with respect to current flow. They probably represent deposition under slowly waning current flow, which becomes incompetent to transport the largest grain. Eddy current systems set up behind this grain promote the accumulation of further grains in the form of a train.

TECTONIC STRUCTURES

Tectonic deformational structures are a characteristic feature of the sandstone units around Gosses Bluff in the Western Missionary Plain. Less noticeable possible tectonic jointing was seen in several massive sandstone beds in the Hermannsburg Sandstone.

(1) Shatter Cones

Shatter cones and associated rhombohedral cleavage is extensively developed in the Harajica Sandstone Member around Gosses Bluff, where they also occur locally in the lower Hermannsburg Sandstone. These features are shown in Figures 3.106 and 3.107 and the scale of shatter cone development can be seen to be very variable. The very large, almost complete, cone in Figure 3.108 is unusual and most of the cones are smaller and flattened on one side (Fig. 3.109). Rhombohedral cleavage is usually
confined to finer, less indurated beds where shatter cones are not developed. These structures have been described by Crook and Cook (1966), Dietz (1967) and Glikson (1969) and analysed by Milton (pers. comm., 1969). They are considered to be formed by intense shock fracturing associated with an extra-terrestrially induced crypto-explosion (Glikson, 1969). Analysis of shatter-cone orientation indicates that the focus of shatter-coning is over the centre of Gosses Bluff approximately 1,525 m above the base of the Harajica Sandstone Member (D. Milton, pers. comm., 1969).

(2) **Polygonal jointing**

Polygonal jointing on exposed uniformly bedded sandstone exposures was noted at several localities. The polygons may be small to moderate size on a rectangular system with joint surfaces continuous in one direction and discontinuous in the other (Fig.3.110). Larger scale, more irregular, polygons with frequent curved joints (Fig.3.111) are also encountered.

These features may represent a form of exfoliation cracks caused by diurnal temperature fluctuations or they could be tectonic features formed in massive sandstone beds as tension joints during periods of folding. The relatively consistent and deep nature of the jointing and the lack of exfoliated slabs tends to favour the second hypothesis.
(3) Micro-fault scarps

Two intersecting sets of micro-fault scarps were noted on a plane-bedded sandstone unit in the middle Hermannsburg Sandstone at section CU (Fig.3.112). These features were studied in cross-section using X-ray radiography (Fig.3.113). Although no actual fracture could be detected the amplitude of displacement decreases with depth and the plane of displacement is concave upwards. The features probably represent slight rotational slumping which occurred while the sediments were still in a hydroplastic state.

DIAGENETIC FEATURES

With the exception of Tertiary weathering profiles which produced laterite, iron oxide horizons and general staining, and silcrete, there are few macroscopic diagenetic features in the sandstone units.

Calcareous concretions were seen at a few horizons and were most frequent in the basal sandstones in section CU. Poorly compacted, lustre mottled, fine sandstones occurred infrequently in the upper Harajica Sandstone Member on Dare Plain. This feature is probably early diagenetic (Conybeare & Crook, 1968).

Porosity is generally low and effective porosity is very low in the Harajica Sandstone Member and Hermannsburg Sandstone in the Missionary Plain area due to primary
matrix content and some breakdown and rarer compaction of lithic grains. Porosity in the sandstone units generally improves to the south and east and both the Langra and Idracowra Sandstones act as locally permeable aquifers with predominantly saline water, except where local recharge is frequent.

Hematite staining of quartz grains may be a diagenetic feature but will be discussed later.

SEDIMENTARY FEATURES OF THE BREWER AND POLLY CONGLOMERATES

Sedimentary features are rather scarce in the conglomerates because of the generally massive and homogeneous nature of the deposits. Sandstone beds are relatively rare but provide the only breaks in the conglomerate sequence.

BEDDING FEATURES

Massive or indistinctly bedded conglomerate is the dominant bedform especially in the lower part of the Brewer Conglomerate. No erosional features were recognized within the massive beds although several unconformities have been inferred. East of the type section of the Brewer Conglomerate only rare sandstone stringers are present in the lower part of the conglomerate. These horizons are usually thin (Figs 3.114 & 3.115), lensoid and plane bedded or very low angle trough cross-bedded. A combination of these
bedding features indicates rapid, almost continuous sedimentation from upper flow regime currents. This mode of deposition is typical of the lower or depositional portions of fanglomerate deposits (Lustig, 1965) where the initial dips may be relatively high (2 - 5°). Details of the mode of deposition will be dealt with later on a regional scale.

Coarse grained granule and pebbly sandstones occur near the base and top of the Polly Conglomerate in the Black Hill Range area and towards the top of the Brewer Conglomerate. A conglomeratic sandstone unit is also present in the Brewer Conglomerate 350 m above the base in the western MacDonnell Range. All these sandstones are characterized by poor sorting and a dominance of plane bedding with pebbles. Occasional low angle trough and tabular cross-stratified units are inter-bedded with the plane bedded units but are generally less than 0.5 m thick. In the Undandita Member, cross-stratified sets are more numerous and may reach 1.5 m in thickness although they are still conglomeratic. Bedding planes are virtually never exposed and hence obscure any hydrodynamic features that may be present.

**SEDIMENTARY BRECCIA**

A sedimentary breccia deposit of very localized extent was seen at one locality near Larrier bore (section GC). The breccia comprises the lowest 30 m of
the Brewer Conglomerate where the latter abuts against a steeply dipping scarp of Areyonga sediments. The breccia forms a talus deposit at the base of the scarp and is not a tectonic breccia for the following reasons:

(a) the size and angularity of the clasts progressively decreases up the contact and laterally away from it;
(b) the clasts are not monomict although they are predominantly so;
(c) where bedding is recognizable, it is tangential to the contact; while in the Areyonga Formation it is almost normal to the contact;
(d) cross-stratification azimuths in sandstones immediately overlying the breccia are directed away from the scarp.

The breccia deposit is characterized by angular slabs of the distinctive Areyonga chert breccia. These slabs vary in size considerably and the largest blocks were about 3 to 4 m (-11.5 to -12 phi) long. They occurred near the base and close to the scarp, and are assumed to represent fallen blocks. The breccia is a very poorly sorted lensoid deposit, merging laterally into conglomerate.

INTERNAL FABRIC

The conglomerate beds consist of a poorly sorted agglomeration of subrounded to well rounded, discoid, ovoid and subspherical pebbles, cobbles and boulders
predominantly of sedimentary rocks (Fig. 3.116). Broken rounds are relatively infrequent and the fractures usually occur parallel to the bedding in siltstone and fine sandstone clasts. No striated or faceted clasts were noted in the conglomerates.

The range in clast size in the conglomerates is very variable, usually from medium or coarse sand to boulders up to -10 phi in the MacDonnell Ranges, up to -8 phi north of Umbeara homestead, and generally up to -7 phi. Higher in the conglomerate, the overlying and laterally equivalent conglomeratic sandstones have most clasts restricted to the pebble size with only a few larger than -6 phi. The largest rounded clast measured was 1 m in diameter. The distribution of maximum clast sizes in the conglomerates and associated sandstones shows a general decrease in size away from the basin margins, although this effect may have been accentuated by erosion of equivalent sandstones towards the centre of the basin.

The internal fabric varies from conjunct with pebbles and boulders touching each other and the interstices filled with finer detritus (Fig. 3.116), to disjunct with the pebbles separated by a coarse sand and granule matrix (Fig. 3.117). All the conglomerates are of a closed-work type with poor sorting except for the removal of most very fine to fine grained detritus. In many cases, the coarse sand and granule sizes have an
open work fabric filled with sparry calcite cement, although this becomes less abundant higher in the sequence.

Pitted pebbles are very rare in megascopic exposures but were occasionally noted in thin section. Figure 3.118 shows a typical example of a sandstone pebble pitting the surface of a micritic limestone clast.

Imbrication of discoidal pebbles is a common feature in the conglomerates (Figs 3.115 & 3.119), although it is frequently hard to discern when bedding planes are absent and the exposures are at an angle to the strike. In most outcrops, imbrication is restricted to discoidal pebbles or cobbles although these may be widely separated by rounded randomly oriented clasts (Fig.3.119). Apart from imbrication, current directional features are largely absent or impossible to measure in the bulk of the conglomerate. However, where the contact above and below is gradational into cross-stratified units, general current directions for the interval can be determined and the imbrication was found to be in an upstream direction in all cases.
ASSOCIATION AND VERTICAL PROFILE ANALYSIS

In studying sedimentary structures a study of the relationship between structures is as important as a study of the structures themselves. To do this a series of vertical sections were measured from typical outcrops over most of the basin. An attempt was made to provide one or more sections from most of the members studied although it must be realised that the actual section selected was arbitrary and depended on the availability of good exposures. In most cases lateral associations within a few metres or tens of metres of the section were noted where they were prominent or unusual. With the exception of a section from the Langra Formation no broad lateral associations are dealt with in detail.

Walther (1893-94) advanced an empirical concept regarding the lateral and vertical association of sedimentary structures. Walther's Law of Facies states that, where there are no time breaks in the stratigraphic sequence, those sediments that were areally adjacent must succeed each other vertically as a consequence of the lateral migration of facies boundaries. As a corollary Vassozhevich (1959) added that in a sedimentary sequence whenever a mutual separation of beds occurs it implies that those beds are not normally juxtaposed in spatial or temporal sequence in a particular area. This basic concept of the relationship between facies, whether they differ radically or only slightly, is extremely
important to the realization of the way in which sedimentary processes control sedimentary sequences. As Visher (1965a) noted each fundamental process produces both a specific environmental facies distribution and a specific vertical profile. In most non-terrestrial sedimentary environments lateral variation cannot usually be studied because of the relatively wide distribution of the individual facies. Thus vertical profile analysis becomes one of the primary concepts available for the study of the inter-relationship of facies and the manner in which lateral migration of the facies takes place. Hence an overall analysis of the changing sedimentary environments can be deduced and used for building up sedimentation models to fit the depositional history.

In all cases studied from the Pertnjara and Finke Groups, there is a direct relationship between lateral variations of depositional forms and their successive occurrence in the vertical profile. The nature of the vertical succession, therefore, provides the clue to the sedimentary processes operating in the depositional environment. Thus cyclicity in the vertical succession represents successive phases of deposition from a laterally oscillating or sporadically occurring sedimentary process i.e. lateral migration of river channels, periodic turbidity currents etc.

The method of association and vertical profile analysis is summarized in Conybeare and Crook (1968)
together with useful tables giving the environmental occurrences of sedimentary structures. Visher (1965a) has described several models available for vertical profile analysis. These include the fluvial or valley fill model, the lacustrine model, the deltaic model, the regressive and transgressive marine models, and the bathyal-abyssal model.

**Association Analysis**

Association analysis can be subdivided into two fields. One studies the relationship of gross lithology to environment, while the other considers the relationship of individual sedimentary structure to the overall environment.

The following gross lithological associations are widespread - massive and laminar bedded siltstones, interbedded laminated fine sandstone and siltstones, interbedded sandstones and very thin siltstones, sandstones, interbedded sandstones and thin conglomerate horizons, and massive conglomerate. They are described with reference to Table 10 of Conybeare and Crook (1968).

The predominantly siltstone sections are characteristic of the Deering and Dare Siltstone Members and the Horseshoe Bend Shale. They are dominantly heterogeneous mudstones and siltstones, although thin horizons of very fine sandstone occur especially towards the upper and
lower boundaries of these units. Fossils are very rare and are restricted to a few spores and fish plates near the base of the Dare Siltstone Member. The sequences generally consist of massive silty mudstones overlying laminated brown and green siltstones. The laminated varicoloured beds probably represent non-marine lake and flood plain deposits while the massive beds could represent fossil soils. No signs of infaunal organisms, or plant roots and debris, were noted in any of the massive siltstone sections and they are definitely absent from the laminated portions. The massive mudstones rarely show any form of even indistinct bedding or textural variation although there are occasionally some small patches of colour mottling in horizons parallel to the bedding. These massive beds are very similar in description to massive mudstone beds reported from the "predominantly sandstone and mudstone section" of Conybear and Crook (1968), which they attribute to a non-marine origin - probably a lake deposit or windblown dust. The associated thin laminated and ripple bedded brown and green siltstones occasionally contain mudstone or sandy siltstone interlaminae and according to Conybeare and Crook could have been deposited in a marine environment close to a delta, or in a non-marine lake or river flood plain deposit.

The thinly interbedded and parallel laminated fine sandstone-siltstone sequence, which is characteristic of the Amulda Member and much of the Harajica Sandstone Member
on Dare Plain, could likewise be marine pro-delta or deepwater trough deposits, or non-marine lake or river floodplain deposits (Conybeare and Crook). These beds also lack fossils including infaunal trains but they exhibit occasional convolute and contorted beds. The convolutions die out above and below the beds, they are not truncated, and they could form either in a marine or non-marine environment. The occasional contorted beds associated with medium scale cross-stratification suggests a possible non-marine fluvial environment.

The colour of the mudstone and siltstone sequence may also be of environmental significance. The generally light reddish brown and greenish grey siltstones exposed at the surface may have suffered secondary colour modifications during the present weathering cycle. However, the constant cyclic rhythm of greenish laminated siltstone, brown laminated siltstone, brown massive mudstone is repeated over a vertical sequence of at least 400 m and the lateral sequence is uniform even in the largest exposures (≈250 m). Also the greenish siltstones usually have associated salt casts while these are rare within the brown siltstones. Thus from surface exposures it may be suggested that the colours represent depositional colours with only slight modification during weathering, since prominent weathering would have produced either a uniform or an irregular colour distribution. This contention is supported by finding similar colour variations in most water bores and oil wells including Tyler No. 1, West
Waterhouse No. 1, Mt Charlotte No. 1, Palm Valley Nos 1 and 2, McDills No. 1 and Witcherie No. 1.

According to Conybeare and Crook the greenish grey dolomitic and/or calcareous siltstones probably represent marine deposits while the reddish brown dolomitic and/or calcareous mudstones could represent marine deposits but reddish brown sediments are more typical of non-marine deposition.

The predominantly sandstone sediments of the Hermannsburg, Langra and Idracowra formations and the Harajica Sandstone and Ljiltera Members are characterized by medium scale cross-stratification. They may contain thin siltstone or conglomeratic horizons but rarely both. The Langra Formation is an exception and contains relatively thick (up to 1 m) siltstone and conglomeratic lenses.

Most of the cross-stratification is high angle (15–30°) and consists predominantly of cosets of trough cross-strata with less frequent tabular sets. It is frequently associated with flat bedded and small scale cross-stratified horizons in a generally cyclic pattern. This association of beds strongly suggests deposition in an alluvial environment.

The conglomeratic sandstones are usually flat bedded or show high angle cross-stratification and
conglomerate lenses frequently show a cut and fill relationship. Pebble imbrication is a feature of some of the conglomeratic beds. All these features suggest a fluvial depositional environment.

The massive nature of the Brewer Conglomerate generally prohibits the recognition of bedding features except where interbedded with flat bedded and high angle cross-stratified pebbly sandstones. The poorly defined bedding and poor sorting of the conglomerate suggests a probable non-marine alluvial fan environment of deposition.

The relationship of individual sedimentary structures can be accomplished best in a tabular form by noting the known environmental occurrences of each structure. Tables 3.3., 3.4 and 3.5, after Conybeare and Crook (1968), present the data for the siltstones, sandstones and conglomeratic units respectively and when used in conjunction with other information they provide fairly clear evidence of the depositional environment.

The siltstone units (Table 3.3) are clearly not piedmont, beach, aeolian or glacial deposits, and the presence of desiccation cracks, rain and/or hail prints, and salt crystal casts indicates that they were not deposited under normal marine conditions. This leaves three major environments in which they could occur - namely fluvial, lacustrine or tidal flat, channel
and delta. The presence of corrugated and convolute lamination and mud volcanoes favours fluviatile or lacustrine environments but they are certainly not diagnostic of them. The lack of erosional channels, the widespread uniform parallel lamination and the uniform silt and clay grain size associated with evaporites and chemically precipitated minerals are all suggestive of a widespread environment such as a large shallow lake. Negative evidence in the lack of biogenic activity cannot be diagnostic but it does not favour a marginal marine depositional environment. A lacustrine environment of deposition is favoured for the siltstones although the evidence is not conclusive.

Overall analysis of sedimentary structures present in the sandstone units is given in Table 3.4. Almost all the structures, with the exception of washover crescents, rill moulds and bimodal fabric, are characteristic features found in fluviatile environments of deposition. As noted in the discussion of individual features, some of the structures are more characteristic of beach, deltaic or aeolian deposition and probably represent localized facies variants with the beaches and deltas being developed on the shores of the large shallow lakes filled with siltstone.

Sedimentary structures from the conglomerate units point strongly to a piedmont or fluviatile, depositional environment. The former is favoured by the very coarse
boulder nature of the deposits, the general lack of bedding, the restricted areal distribution and rapid interdigitation with fluviatile sediments down slope, and the preservation of ancient fans characterized by specific rock types.

Thus the grouping of individual sedimentary structures merely provides confirmatory and generalized evidence regarding the overall environment of deposition.

**Vertical Profile Analysis**

Twenty five detailed sections covering an average of 15 m each were measured from most of the described sedimentary units. The Deering Siltstone Member and the Harajica Sandstone Member on Dare Plain were not measured but have been described from photographic sections as has the Polly Conglomerate at its type locality. No section was measured through the Idracowra Sandstone but in outcrop it appears to be very similar to the Hermannsburg Sandstone on the Camel Flat Syncline.

The arrangement of descriptions of the detailed sections is basically stratigraphic with descriptions of the Pertnjara Group preceding those of the Finke Group. Within both groups the sections are described in ascending stratigraphic order. A few exceptions in the Hermannsburg Sandstone are due to the grouping of sections with similar characteristics. Thus, the grouped sections AM1, BPl and CS come from the basal, mid and upper
Hermannsburg Sandstone respectively but they all have similar depositional characteristics. Likewise, the pebbly lower Ooraminna Sandstone Member (section BS) is described prior to the Ljiltera Member with which it has its closest affinity.

The purpose of detailed descriptions of the measured sections is to bring out the systematic variations and cyclicity exhibited by the sediments. Different cyclic units denote deposition under slightly different conditions and the object of this part of the study is to determine the basic depositional sequences present and, if possible, to fit models of deposition to these sequences.

The vertical profiles described below cover all the main sequence variations noted in the field with the exception of locally anomalous features. All the sequences described can be attributed to deposition in non-marine conditions including alluvial fans, fluvial systems and evaporitic lakes.

Section DHL (Figs 2.16, 3.120; 0-150 m)

No detailed section was measured through the Deering Siltstone or Harajica Sandstone Members on Dare Plain although they form quite distinctive units. The upper part of the Deering Siltstone Member is seen in the lower part of Fig. 3.120 where the base of the
Harajica Sandstone is the sandstone bed at 2 m. The contact is entirely gradational.

Both members are characterized by even, thin, parallel laminae of siltstone and sandstone, although the central part of the Harajica Sandstone Member contains thicker cross-bedded sandstones. Sandstone beds are infrequent and very thin in the Deering Siltstone Member which consists dominantly of laminated and massive mudstones and siltstones. The Harajica Sandstone Member is composed predominantly of flat bedded fine to medium sandstones with numerous thin siltstone laminae (Fig. 2.16). In the centre of the member small to medium scale tabular and trough cross-stratified sets are more common and are associated with slump bedding and occasional mud cracks.

The extensive thin laminae and alternations of sandstone and siltstone suggest subaqueous deposition of pro-delta deposits or possibly non-marine flood plain deposits. The former environment is preferred because of the general lack of desiccation cracks, and the presence of carbonate and fragmentary fish plates.

Section AB1 (Fig. 3.121, 137-159 m)

This section shows the contact between the Deering Siltstone Member (0-12 m) and the overlying Harajica Sandstone Member, at the type section of the latter.
The upper Deering Siltstone Member is characterized by the alternation of siltstones and sandstones. The basal 3.5 m in Figure 3.121 shows a typical sequence of the Deering Siltstone Member with laminated siltstones and minor massive mudstones and fine-sandstone lenses. The siltstones occasionally show contorted bedding and ripple marks (predominantly asymmetrical). Overlying this sequence and forming the uppermost units of the Deering Siltstone Member there are a series of fine cycles of alternating sandstone and siltstone. The sandstones overlie an erosional surface and are usually relatively thin (less than 50 cm). They are usually trough cross-stratified and may contain occasional mudclasts and rare small pebbles (up to 5 mm diameter). The sandstones are poorly sorted litharenites which vary in grainsize from medium to coarse sand. The trough cross-stratified sets are usually overlain by flat bedded sands which give way to interbedded irregularly laminated fine sandstones and siltstones with occasional ripple-drift bedded horizons. The top unit in cycle 5 contains several thin tabular cross-stratified sets interbedded with the siltstones.

The boundary with the Harajica Sandstone Member is gradational and its lowest two units consist of medium scale tabular cross-stratified fine to medium grained sandstones with occasional mudclasts and very rare pebbles up to 1.5 cm diameter. These units are overlain by small scale cross-beds, laminated horizons and rare ripple bedded sandstone. The overlying cycles are rather similar
except in thickness and consist of a coset of trough cross-stratified units overlying an erosional surface. The cross-strata are composed of fine to medium grained, sub-rounded sublitharenites which occasionally contain a few mudclasts aligned parallel to the stratification. The cross-strata are either concordant or asymptotic to the base of the set. The trough cross-stratified cosets vary in thickness from 50 cm to 2.5 m and are usually overlain by a variable thickness of flat bedded and occasionally ripple drift bedded fine sandstone. Occasionally mud cracked, laminated siltstone is preserved at the top of the cycle.

Section AB1 shows a gradational sequence in depositional environment from one of low energy, characterized by the deposition of thin bedded siltstone and fine sandstone, to one of higher energy which deposited the medium scale cross-stratified sets. In the latter, currents were capable of erosion and of transporting small pebbles but they were probably still confined to the lower flow regime. The thick sequence of siltstone and fine sandstone suggests deposition in a flood plain or lacustrine environment. The increasing sandstone content higher in the sequence could indicate closer proximity to the fluvial channel with the Harajica Sandstone Member representing channel sand deposits.
Sections AB2 and AJ1 (Fig. 3.122, 650-672 m; Fig. 3.123, 125-160 m)

Both these sections represent intervals within the Harajica Sandstone Member and they show somewhat similar characteristics. Fine grained and laminated units are less common in section AJ1 than in section AB2 and this gives section AJ1 the appearance of being a homogeneous sequence of cross-stratified sandstones. However, when studied in detail the erosional surfaces, thickness and grain size of the sets, and the abundance of mudclasts make the section divisible into 12 cycles.

A typical, almost complete cycle, is shown in cycle 3, section AB2. The base of the cycle is marked by a sharply defined erosional surface which is overlain by trough cross-stratified coarse grained sandstones. The lower sets frequently contain small pebbles (dominantly less than 1 cm diameter) of subrounded sedimentary rock fragments, together with more abundant mudclasts up to 10 cm long. The latter are aligned parallel to the bedding and do not show imbrication. Higher in the coset the trough sets become smaller, shallower and finer grained and they very rarely contain mudclasts or pebbles. This festoon cross-stratification gives way to flat bedded fine to medium grained sandstone, occasionally showing current lineation or horizons of ripple-drift bedding. Very rarely siltstone laminae are preserved at the top of the cycle. Another feature characteristic of several of the cross-beds
in this unit is the presence of contorted or slump bedded horizons in sets with steep sides or foresets.

Most cycles are not as complete as the one described above and they usually consist of only the lower cross-stratified portion. This is especially the case in section AJ1 where the highest unit exposed is usually small scale festoon cross-bedding. Also no pebbles were recorded from section AJ1 although mudclasts are quite abundant.

The cycles in the Harajica Sandstone Member conform well with the point bar profile described by Harms, McKenzie and McCubbin (1963) with the coarser pebbly sandstones at the base passing upwards into finer trough cross-stratified and flat bedded sets. The general paucity of bedding types, representing the upper part of the point bar profile, is somewhat anomalous since Allen (1964b) has described complete cycles from the lower Old Red Sandstone. The lack of upper units could be due to erosion prior to the deposition of the overlying beds. The incorporated mudclasts indicate that erosion was common between successive units but do not indicate how much erosion took place since mud polygons are commonly only loosely attached to the underlying sediment. The other alternative is that these sedimentary cycles do not fit the described point bar profiles completely and that the mode of deposition was not entirely from meandering streams. A braided or low sinuosity stream model may fit these features more exactly
Section DH2 (Fig. 3.124, 302-347 m)

Section DH2 shows a fairly typical sequence through the lower Dare Siltstone Member at its type locality. The cyclic nature of the bedding is readily apparent above cycle number 6 and this cyclicity continues throughout the rest of the member.

The basic cycle is rather simple and consists of three units. The base of the cycle is a flat surface which may be either erosional or gradational (Figs 3.125 & 3.126) and it is marked by the incoming of greenish laminated siltstones. These siltstones frequently show straight or sinuous wave ripple marks and towards the top they may contain numerous salt casts in certain horizons. The junction between the green laminated siltstones and the overlying light reddish brown laminated siltstones may be either gradational with the two types interlaminated or it may be a sharp junction. The light reddish brown siltstones also exhibit ripple marks but contain less frequent salt casts than the underlying beds. These laminated siltstones pass gradationally upwards into thick massive beds of silty mudstone and muddy siltstone (Fig. 2.20). These massive beds are characterized by conchoidal fracture and the occasional presence of calcite lined vugs in horizons parallel to the bedding. The size of individual cycles varies from about 3 to 8 m and they very rarely show any unusual features.
Towards the base of the Dare Siltstone Member, however, irregularities in the sequence are more common and occasional lensoid, fine sandstones and rare, silty, micritic limestones may be present. In cycle 5 an unusual feature is a very shallow erosional scour (Fig. 3.14) filled partially with very fine grained sandstone. In cycle number 4 the massive unit is missing entirely, while at the base of number 1 there are three alternations of laminated and massive siltstones in one metre.

The laminated brown and green siltstones must have formed in a subaqueous saline environment subjected to wave action rather than current action. The colour variation, as discussed before, probably reflects the redox potential of the depositional environment. The greenish siltstones indicate reducing conditions with most of the iron in the ferrous state, and as deposition continued the environment became more oxidizing and the brown siltstones have most of the iron in the oxidized ferric state. The nature of deposition of the massive units is harder to explain since no internal burrowing or churning has been recognised. They are calcareous and dolomitic, as are the laminated units, and they may represent deeper water deposits unaffected by surface currents, or they could be wind blown dust redeposited from the laminated units. If the latter explanation were correct erosional contacts and incomplete sequences would be expected in at least some horizons but this was not found to be the case. Alternatively they could be pedogenic deposits but since no disturbance of
primary bedding has been noted (Fig. 3.3) and there are no features or remains of floral or faunal organisms this alternative appears to be unlikely.

The most probable environment of deposition would be a widespread playa subjected to periodic inundation and desiccation.

Section CJ1 (Fig. 3.127, 203-218.5 m)

This section shows the middle portion of the Amulda Member at its type locality. The Amulda Member is characterized by an alternating sequence of sandstones and siltstones on various scales. Figure 3.127 shows the alternating cycles at about the largest scale seen (4.5-6 m). Most frequently the cycles are smaller in scale (10-50 cm) and many are hard to distinguish from irregularly inter-laminated sandstones and siltstones.

Figure 3.127 contains three complete cycles each starting with a medium, or medium to coarse grained, poorly sorted, massive to indistinctly flat bedded or cross-stratified sandstone containing mudclasts aligned parallel to the bedding. The basal portion of the sandstones lies on an erosional surface and is frequently massive. It is overlain by low angle trough cross-stratified sets and occasionally, by siltstones and mudstones. A few sandstone units lack the massive portion and are trough cross-stratified throughout. Occasionally the sandstones contain
calcite lined vugs usually in rough lines parallel to the base of the unit. The sandstones are gradationally overlain by interlaminated silty sandstones and sandy siltstones which rapidly give way to laminated, light reddish brown siltstones and massive mudstones in an alternating sequence. Laminated beds usually predominate in this section and there may be a few irregularly laminated and lensoid fine sandstone interbeds. Occasional asymmetrical ripple marks were also noted.

These features are comparable to the upper part of a point bar profile with the thick sandstone units representing migrating or very wide shallow channel deposits and the laminated siltstones with rare fine sandstone lenses representing overbank deposits. The massive siltstones could represent suspension deposits or accumulations of wind-blown dust.

Section DH3 (Fig. 3.128, 735-749 m)

Figure 3.128 shows the Hermannsburg Sandstone overlying the alternating laminated sandstones and light reddish brown siltstones of the Amulda Member north of Dare Plain. Two almost complete cycles are shown in the basal Hermannsburg Sandstone and these are overlain by a large succession of similar but less complete cycles forming the bulk of the Hermannsburg Sandstone. The first cycle is 5.75 m thick and consists of two large to medium scale, low angle, trough cross-stratified units
each overlying a moderately flat but sharply distinct erosional surface. The cross-strata within each set are either asymptotic or concordant with the erosional surface and frequently contain mudclasts up to 12 cm long near the base of the set. The mudclasts are usually aligned parallel to the stratification and are rarely present in sufficient numbers to show imbrication. These large sets are overlain by medium scale trough and tabular cross-stratified sets up to 75 cm thick. These sets have erosional bases, rarely contain mudclasts and grade upwards into festoon cross-bedding of small medium-scale, trough cross-stratified sets. The latter in turn grade upwards into laminated and ripple-drift bedded fine sandstones (Fig. 3.33). The uppermost unit of the cycle consists of laminated and mudcracked light reddish brown siltstone and sandy siltstone. The upper two units are infrequently preserved in most cycles within the Hermannsburg Sandstone and this makes the cyclicity less obvious in most sections.

The upper cycle in Figure 3.128 is almost a repetition of the lower cycle except for the absence of the siltstone horizon. The lateral continuity of these cycles for at least 450 m suggests that the environment must have been fairly widespread or that lateral migration of the units must have been relatively constant. The cycle described has many of the characteristics of the point bar profile described by Bernard and Major (1963) and the predominance of trough cross-stratification is
in accordance with the point bar deposits described by Harms et al. (1963). The main difference between the cycles is the absence of coarse gravelly sand from the base of the cycles shown in Figure 3.128. The lack of gravel could be attributed to non-availability due to the transport distance and the velocity of flow. All the structures present in the cycles represent lower flow regime conditions which were probably not strong enough to transport gravel over large distances. The abundance of mudclasts and erosional surfaces indicate that the stream velocities were high enough to be erosional but the mudclasts are subangular and indicate only relatively short transportation. The absence of overbank deposits suggests that either they were entirely eroded every time before the deposition of the succeeding horizon, or the river migrated rapidly in lateral directions, did not flood very often and did not build up a silty flood plain deposit.

Section EH (Fig. 3.129, 123-137 m)

Section EH, west of Titra Bore, is very similar to the basal Hermannsburg Sandstone in section DH3 and it is probably the equivalent horizon. It conformably overlies interbedded fine sandstones and siltstones of the Parke Siltstone (0-3 m) and consists of four cycles overlain by more uniform cross-stratified sandstones.

As in section DH3 the cycle consists of trough
cross-stratified, medium grained sublitharenites overlying an erosional surface with a relief of up to 20 cm. The basal sets are up to 1 m thick but overlying sets generally decrease in thickness and grade into cosets of small medium-scale festoon bedded trough sets. The top of the cycle usually consists of flat bedded and ripple bedded finer sandstones and occasional siltstone laminae. Occasional sets (e.g. number 2) are composed of small-medium-scale tabular cross-stratified sets with curved asymptotic cross-strata. The larger cross-beds near the base of each cycle may contain a few scattered mudclasts. The average thickness of the cycles is about 3 m.

As noted in section DH3 these cycles are very similar to the point bar profile except for the lack of pebbly basal beds and overbank deposits.

Section AJ2 (Fig. 3.130, 640-655 m)

The sequence in section AJ, from near the base of the Hermannsburg Sandstone at Glen Helen Gorge, is characterized by the abundance of thin, fine sandstones and interbedded siltstones.

Ten cycles of variable thickness are shown in Figure 3.130. The lowest cycle displays most of the characteristic features and it is 3.6 m thick. Overlying the basal erosional surface there is a 10 cm interval of pebbly, trough cross-stratified sandstones with
the pebbles up to 1 cm in diameter, dominantly of sedimentary rock fragments. A thin, flat bedded unit separates the lower and upper cosets of trough cross-stratified units. The sets in the upper coset are about 20 cm thick becoming thinner towards the top where they grade into the overlying flat bedded mudstones and siltstone. The other cycles contain all or part of this sequence with occasionally more interbedding of sandstone and siltstone higher in the sequence. Cycles 8 and 10 contain up to 50 cm deep erosional scours beneath larger trough cross-beds.

The proportion of laminated, fine sandstone with occasional interbedded siltstones is high (60%) in this section while the associated cross-stratified sets are usually less than 30 cm thick. These sediments correspond quite well with deposits characteristic of the upper part of the point bar profile and associated flood plain deposits.

Sections BN, CJ2 and EM (Fig. 3.131, 15-29 m; Fig. 3.132, 415-423 m; Fig. 3.133, 187-201 m)

Section BN is a typical portion of the lower Hermannsburg Sandstone sequence exposed on the northern flank of Waterhouse Anticline. Section CJ2 is an atypical sequence because of the interbedded fine deposits. It occurs 155 m above the base of the Hermannsburg Sandstone east of Areyonga and it is underlain and overlain by more typical cross-stratified sequences, similar to those in section BN. Section EM is similar to section BN and was measured north-west of Idracowra homestead.
A typical, almost complete cycle is shown by number 8 in section BN. It consists of trough cross-stratified sets overlying an erosional surface. The sets become progressively smaller upwards and consist of thin cosets separated by slight erosional breaks. The top of the cycle is represented by laminated, flat bedded or irregularly wavy bedded sandstone. There is generally an upwards decrease in grain size in each cycle and a few cycles have scattered mudclasts near the base. Several cycles consist of tabular cross-beds instead of trough cross-beds and these usually have concave upwards cross-strata, asymptotic to the base. Section CJ2 contains some thin siltstone inter-beds in the laminated upper portion of the cycle.

These cycles are similar to point bar deposits but on a finer grained and smaller scale. They represent repetitive sequences equivalent to the middle and upper portions of the point bar profile. They may represent deposits formed by migrating small streams but quite possibly they were formed in a wide, shallow, braided or low sinuosity stream environment.

Sections AM1, BP1 and CS (Fig. 3.134, 18-44 m; Fig. 3.135, 254-264 m; Fig. 3.136, 295-310 m)

These three sections from the Hermannsburg Sandstone are similar although their stratigraphic position is variable. Section AM1 is from near the base at Ellery Creek, section BP1 is near the top of the Ooraminna Sandstone Member at Orange Creek, while section CS represents the highest
exposure at Hugh River on the northern flank of James Range. They are all characterized by the abundance of trough cross-stratified sandstones occasionally interbedded with flat bedded sandstones.

Number 2 in section BP1 is a fairly typical cycle even though it is only 1.5 m thick. Medium to coarse grained, medium scale, trough cross-stratified sets overlie an undulating erosional surface. The basal sets usually contain frequent mudclasts aligned parallel with the depositional interface, and in some cycles they also contain small pebbles. The lower sets pass upwards into smaller scale trough and occasionally tabular sets with less frequent mudflakes and usually no pebbles. Overlying the trough cross-stratified sets there is a gradational change to flat bedded laminated sandstones with current lineation on some surfaces. Higher in the cycle ripple bedding and siltstone interbeds may also be present. Many of the cycles represented are incomplete and only contain the cross-stratified units while a few cycles in section AM contain a basal unit of flat bedded, pebbly sandstones. The latter represent deposition from upper flow regime currents where flat beds are a stable bedform. The overlying cross-stratified units represent succeeding periods of deposition when the current strength had waned to the lower flow regime.

These cyclic deposits are very similar to the ones described in sections AB2 and AJ1 and they were probably
formed in a similar manner from meandering or braided streams.

Section CY (Fig. 3.137, 350-365 m)

Section CY in the upper Hermannsburg Sandstone near Deep Well is quite distinct from the usual cyclic sequences. It can still be divided into cycles on the basis of cross-bed size but the individual cycles are much thicker (up to 15 m).

Only one complete cycle is shown in Figure 3.137 and it consists of large scale, gently inclined, curved planar cross-stratified sets overlying even, flat, erosional surfaces. The laminations within each cross-strata are very even, regular, continuous and parallel and the sets are characterized by a uniformly medium grainsize and the lack of pebbles and mudclasts. Individual sets vary from 1.5 to 4 m in thickness and are continuous for over 20 m. Towards the top of a cycle, medium and medium to large scale sets are more common and may be trough cross-stratified or planar cross-bedded in inclined sets with cross strata parallel to the lower boundary surface.

Cross-stratification of this type is typical of aeolian deposits especially those of dome shaped sand dunes (McKee, 1966).
Section GA (Fig. 3.138, 224-239 m)

Section GA is from the top of the Hermannsburg Sandstone on the northern flank of the Camel Flat Syncline. The section is relatively hard to subdivide and appears to consist of both fluvial and aeolian deposits.

The two basal cycles are probably fluvial deposits similar to those in section CS (Fig. 3.136) but with fewer mudclasts and pebbles. The first cycle starts with a concentration of siliceous pebbles at the base which pass up into trough and tabular cross-stratified units and then flat bedded sandstones. The cross-stratified sets are up to 1.5 m thick with the cross-strata either concordant or asymptotic to the base. Individual sets usually become thinner towards the top of the cycle. The second cycle is similar but lacks the basal pebbles. The third cycle is 5.5 m thick and consists of two large scale, curved, planar sets overlain by trough and planar wedge shaped sets similar to those described from section CY. The sediments in this cycle are all medium grained. The fourth cycle is similar to the first two and consists of medium and medium to coarse grained sandstones with rare mudflakes. They are trough cross-stratified on a medium scale and pass upwards into laminated sandstones.

This section probably represents a composite set of depositional conditions with cycles 1, 2 and 4 being dominantly of fluvial origin and cycle 3 aeolian origin.
Figure 3.139 shows the upper portion of the pebbly, coarse grained, quartzarenite unit of the lower Ooraminna Sandstone Member. It differs from most of the Hermannsburg Sandstone sections because it only contains rare, shallow, cross-stratified sets. It is characterized by massive and indistinctly flat bedded, pebbly sandstones alternating with non-pebbly massive and laminated sandstones.

The sequence, even though indistinctly bedded, can still be subdivided into cyclic repetitions of various bedforms. The most complete cycle measured is number 3 which starts with a massive medium to coarse grained sandstone overlying a flat erosional surface. The contact between this unit and the overlying pebbly sandstone is gradational. The latter is an indistinctly flat bedded, coarse grained quartzarenite with abundant -1 to -3 phi well rounded pebbles of silicified Mereenie Sandstone and vein quartz. The unit also contains quite common mudclasts. Overlying the pebbly sandstone is a massive or indistinctly flat bedded sandstone containing rare pebbles of quartz scattered throughout. This set grades up into small to medium scale, low angle, trough cross-stratified sets followed by laminated sands forming the top of the cycle. The other cycles are essentially similar with one containing a concentration of pebbles at the base and no cross-strata while the other contains no pebbles and a relatively thick...
cross-stratified coset.

The massive and flat bedded pebbly sandstones are characteristic bedforms produced in the lower part of the upper flow regime. Waning current velocities are represented by the cross-stratified and overlying laminated units. This may represent an actual waning of current velocity or it could be formed by the lateral migration of the focus of maximum current velocity. The features described do not closely fit any of the proposed environmental models but show the greatest similarity with the point bar profile. They may be attributed to a fluvial depositional environment because of the poor sorting with irregularly scattered pebbles occasionally concentrated in discontinuous lensoid conglomeratic horizons together with erosional surfaces, medium scale festoon cross-stratification, and the presence of rounded pebbles of the underlying silicified sandstone. They can be attributed to sheet flood deposition on the flanks of the gently arched Ooraminna Anticline.

Sections AJ3, BM and CM (Fig. 3.140, 794-813 m; Fig. 3.141, 285-300 m; Fig. 3.142, 500-526 m)

These sections all represent portions of the Ljiltera Member of the Hermannsburg Sandstone. They are characterized by pebbly cross-stratified sandstones and they show a similar cyclicity to the lower Hermannsburg Sandstone.
A typical cycle is shown in number 8, section AJ3. Medium scale, trough cross-stratified sandstones overlie a sharp erosional surface. They contain numerous mudclasts aligned parallel to the cross-strata and variable quantities of subrounded to rounded sedimentary and quartz pebbles typically less than 10 cm in diameter. The pebbles and mudclasts are usually concentrated towards the base of the sets but rarely show imbrication. The basal sets are commonly overlain by a series of trough cross-stratified sets which become finer grained, less pebbly and smaller in scale upwards. The uppermost trough sets frequently contain no pebbles or mudclasts. The trough sets are overlain by flat bedded, irregularly laminated or ripple bedded sandstones which very rarely contain interbedded siltstone laminae. Ripple marks where present are usually linguoid or sinuous asymmetrical forms with a wave length of between 5 and 15 cm. Current lineation was occasionally seen in the flat bedded sandstones. Slump bedding of individual cross-strata was occasionally seen in some of the larger trough sets.

These cycles all fit the general point bar profile and represent various sequential sequences of erosion and deposition in a fluvial environment lacking a well developed flood plain.
Sections AJ4, AM2 and BP2 (Fig. 3.143, 1261-1276 m; Fig. 3.144, 55-77 m; Fig. 3.145, 310-325 m)

Section AM2 represents the basal Ljiltera Member conglomerate at Ellery Creek while section BP2 represents the same interval on east Waterhouse Anticline. Section AJ4 represents the top of the Ljiltera Member and the basal Brewer Conglomerate at Glen Helen Gorge. The sequences are characterized by numerous erosional breaks and the lack of complete cycles.

A generalized composite cycle would consist of a basal massive or flat bedded, pebbly sandstone or conglomerate overlain successively by pebbly trough cross-stratified units and possibly flat bedded, fine sands. The conglomerates frequently fill large erosional scours (Fig. 3.146), may contain sandstone stringers and frequently exhibit clast imbrication. A few pebble trains were also noted. The conglomerate bands may have gradational or erosional boundaries with the flat bedded pebbly sandstones (Fig. 3.147). Mudclasts are infrequent in the conglomeratic sequences although they may be present in associated sandstones. The conglomerates are usually poorly sorted and contain subrounded sedimentary pebbles up to 25 cm in diameter while most of the clasts are less than 10 cm. Cross-stratified sets are infrequent in section AM2 but the other two sections contain trough sets which become thinner and less pebbly upwards. They
may be overlain by finer grained flat bedded sandstones.

These interbedded conglomerate and pebbly sandstone sequences correspond to the lower part of the point bar profile where channel processes predominate. Many of the conglomerates are lensoid in nature and the general poorly defined bedding and poor sorting suggest that the sequences were most probably deposited in an alluvial fan environment (Conybeare & Crook, 1968).

Section FG1 (Fig. 2.46, 15 m)

Figure 2.47 shows a typical section from the middle of the Polly Conglomerate at its type locality. It shows the indistinctly flat bedded, poorly sorted pebble conglomerate with its associated flat bedded, coarse grained sandstone stringers. The pebbles in the conglomerate occasionally show up-current imbrication to the south but apart from this there are very few distinctive features. The flat-bedded coarse grained nature of the sequence suggests that it was deposited under upper flow regime conditions in a fluvial environment.

Section FG2 (Fig. 3.148, 165-173 m)

Section FG2 is a typical section through the middle Langra Formation showing all lithologies from coarse lensoid conglomerate to sandy siltstone. The
section is well exposed in the low cliffs on the southern bank of the Finke River at Horseshoe Bend.

A typical cycle comprises the lowest four units in section FG2 (Fig. 3.148). The basal conglomerate overlies an irregular erosion surface containing small channels (Figs 3.56 & 3.57). The conglomerate is composed of well rounded pebbles and cobbles of sedimentary, metamorphic and granitic rocks (Fig. 3.149). It is lensoid in form and occupies hollows in the erosion surface. The conglomerate is overlain by a flat bedded pebbly sandstone, occasionally showing current lineation, which is overlain in turn by medium scale pi-cross-stratified sandstones containing rare pebbles and mudclasts. The uppermost unit in this cycle is a fine to medium flat bedded sandstone also containing some current lineation. The overlying cycles are similar in general form but lack the basal conglomeratic horizon. Cycle 4 contains a laminated red brown sandy siltstone which is the finest member present in the sequence.

The conglomerate and overlying flat bedded sandstone of Cycle 1 were probably deposited by currents in the upper and transitional flow regimes whereas the remaining beds are typical of deposition in the lower flow regime. These cycles are common in the lower and middle Langra Formation and they conform to the general point bar profile. The conglomerate would represent a lag gravel deposited in the stream channel and the
succeeding finer deposits represent lateral accretion on the inner banks of the laterally migrating channel.

**Section FG3 (Fig. 3.150, 305-311 m)**

Figure 3.150, from the upper part of the Horse­shoe Bend Shale at the type locality, shows the general cyclic nature of the bedding. The cycles are somewhat similar to those of the Dare Siltstone Member, but they are less regular and the base of the cycle is hard to define because of poorly exposed contacts.

The base of the cycle is taken at the base of the laminated or massive greenish siltstone horizons which form distinctive marker bands. The greenish horizons tend to be slightly irregular and lenticular and their base may be transitional or abrupt. Some of the horizons are very biotitic and have wavy or rippled lamination and a few contain salt casts. The green silt­stones are usually 5 to 50 cm thick and are commonly overlain by 10 to 150 cm of laminated and wavy laminated red brown siltstone. The upper part of each cycle is typically composed of massive red brown siltstone which may contain occasional laminated red brown horizons or lenticular pods of green siltstone (Fig. 3.1). These massive horizons vary from 5 cm to over 2 m in thickness and they form the most resistant horizons in the Horse­shoe Bend Shale (Fig. 2.56).
The siltstones in section FG3 were probably deposited in a large ephemeral lake or playa subjected to periodic desiccation. The alternating colour banding probably represents deposition under alternately oxidizing (subaerial) and reducing (subaqueous) conditions. The massive beds may represent wind blown deposits.

**FLUVIAL ENVIRONMENTS**

The Pertnjara and Finke Group sediments fulfil almost all the criteria proposed by Wolf and Conolly (1965) for the identification of terrestrial deposits. The two exceptions are that these sediments possess no plant fragments and contain numerous rounded grains. The first factor may be attributed to a rigorous and oxidizing climate while the second factor is readily explained by the reworking of older marine and terrestrial sediments composed of rounded and well sorted quartz e.g. Mereenie, Stairway and Pacoota Sandstones.

However, recognition of the nature of the environment is not always easy because of the wide range of possible fluvial environments. On the present land surface they vary from narrow upland valleys to wide flat swamps near sea level with all the numerous intermediate stages. With this great diversity of environments it is perhaps surprising to find that river systems can generally be classified into three major forms although these forms may merge into one another along the length of
the river. The three major forms are meandering, straight or low sinuosity, and braided rivers.

Stream systems cannot always be recognized with certainty in older deposits and thus one of the fundamental approaches in studying fluvial deposits is to use the concept of flow regime (Simons and Richardson, 1961). With increasing velocity the continuum of bed forms produced varies from a plane bed without sediment movement to small scale ripples, to dunes often with ripples superimposed on their backs, to dunes without ripples (Harms & Fahnestock, 1965). All these bedforms are characteristic of flow within the lower flow regime and they may be preserved in the stratigraphic record as parallel laminations originating from suspension deposition, ripple-drift bedding and cross-bedding. Transitional forms between dune bedding and plane bedding (usually represented by very low angle cross-bedding) occur when the flow regime is transitional between upper and lower flow states. Plane beds with sediment movement and antidunes are the main bedforms in the upper flow regime although antidunes are rarely preserved in the stratigraphic record.

The concept of flow regime is very useful since it results from the combined effects of flow velocity, turbulence, bedforms, channel geometry and slope, discharge and temperature. These features are rarely recognizable in ancient deposits and thus the characteristic flow conditions are described more usefully by an overall flow regime
parameter denoting the competence of the flow at the time when a particular structure was deposited.

Stream competence is defined by the size of the largest grain that it can move by traction as bed-load. The competence for large grain movement is proportional to the sixth-power of the flow velocity (Morisawa, 1968) and hence during periods of flood, the competence of the stream is greatly increased. The force required to initiate grain movement is greater than that required to keep them moving. The relationship of depth and slope is critical in initiating movement of small non-cohesive grains, while stream velocity is the most important factor initiating entrainment of larger particles.

Flume studies by Fahnestock and Haushild (1962) indicate that pebbles and cobbles would be frequent in a sand deposit only if the adjacent bedding within that deposit were planar. If this is true then the occurrence of pebbles at the base of trough cross-stratified sandstones of the Pertnjara and Finke Groups suggests that either the bedding upstream of the cross-stratified deposit was planar, or that flat bedded sands were deposited and eroded prior to the deposition of the cross-stratified sandstone and the pebbles represent a lag deposit left during erosion. The graph (Fig. 3.158) indicates that the velocity required for erosion of sand sized particles is less than that required for the erosion of either silt, clay or gravel. Silt and clay are hard to erode because
of their cohesive forces and close packing but once entrained they remain in motion even at extremely low velocities. Thus, such particles are readily flushed through a fluvial system provided ponding is not prevalent and velocities are sufficient to entrain the particles in the first place.

A downstream decrease in particle size is most apparent in aggrading streams where reworking and renewed movement of detritus is limited. A downstream decrease in size of gravel material has been noted on alluvial fans (e.g. Blissenbach, 1952; Lustig, 1965) while a decrease in sand size material has been noted by Leopold and Miller (1956) and Allen (1965b).

In a stable alluvial channel the mean weight percent of silt and clay (material finer than 0.074 mm) in the channel and banks is inversely related to the width-depth ratio (Schumm, 1960). In general aggrading channels have higher width-depth ratios than is indicated by the quantity of silt and clay, and channels containing little silt and clay are relatively wide and shallow. Streams containing a high proportion of silt and clay generally have narrow, deep channels and are most frequently braided. The mean percent silt-clay (M) can be calculated by the relation

\[ M = \frac{Sc \times W + Sb \times 2D}{W + 2D} \]

where Sc is the percentage of silt and clay in the channel, Sb is the percentage of silt and clay in the bank, W is
channel width, and D is channel depth. The only clearly defined channel noted in this study (Section CS) gave values of:

\[
\frac{W}{D} = 2.5
\]

\[
M = \frac{(19.25 \times 2.5 + 75 \times 2)}{4.5} = 44.03
\]

When plotted on a graph (e.g. Fig. 10, p.23, Schumm, 1960) the point falls below the regression line \( \frac{W}{D} = 255M^{-1.08} \) which suggests that the stream was possibly degrading its course slightly but was fairly close to equilibrium.

In relatively stable alluvial channels a high proportion of silt and clay in suspension promotes sinuosity since it becomes incorporated in bank materials, increases the cohesiveness of the banks, and thereby restricts rapid lateral migration or erosion (Brice, 1964). Thus, the South Loup River, Nebraska, which has 24 percent silt and clay, exhibits a meandering pattern while the North and Middle Loup Rivers which have essentially similar flow conditions but only 9 percent silt and clay exhibit a braided pattern. This supports the relationship of silt and clay to width-depth ratio proposed by Schumm (1960, 1968).

The hydraulic geometry of a channel depends on the slope, width, depth, flow velocity and bed roughness. Discharge, calibre and load of sediment are independent of channel geometry and are related to the geology, climate and relief of the drainage basin (Allen, 1965b). Direct
relationships between mean velocity, depth and slope of a fluvial channel are indeterminate because of the variability of the hydraulic roughness caused by changing bedforms (Maddock, 1969). The competence of the stream to carry debris is also variable and depends on the rate of discharge and depth of water amongst other factors. Thus there are different relationships for stream competence for low, medium and high discharge rates in a single channel (Maddock, 1969). The effect of climate on sediment yield has been studied by Langbein and Schumm (1958) and they noted that at 10°C sediment yield increases with precipitation to a maximum at 33 cm per year and then decreases again as rainfall increases to 100 cm per year because of the added effects of vegetation. Yield increases again at very high precipitation rates (Crook, 1967).

The size distributions and modes of deposition of detritus forming alluvial fans and talus deposits in Deep Spring Valley, California, have been described in detail by Lustig (1965). In general, fluctuation in size of clasts and poor sorting of the sediments are characteristic features of the environment. There is no consistent pattern to the variation in maximum or mean particle size over the surface of an alluvial fan (Lustig, 1965) although there is a general downfan decrease in size. This is especially noticeable in the granule size range where mechanical breakdown of the granules reduces their abundance from 15 to 20 percent at the fan head to only 2 to 5 percent in the centre of the basin. In general, the estimated tractive force required to
entrain and transport the large clasts can be estimated from the slope and depth of water (Lustig, 1965) and there is usually a reduction in tractive force away from the fan head. In semi-arid areas mudflows are a dominant mode of transport and because of their viscous nature they may carry relatively large blocks over the surface of alluvial fans. With increasing and more frequent rainfall mudflows become less common and normal aqueous fluvial transport predominates on alluvial fans which consequently tend to have a lower gradient.

Clast size distribution in the Brewer Conglomerate could not be studied in a strictly down-fan sense because suitable cross-sections of a single deposit are not available. Thus the distribution of clast sizes shown in Figure 3.159 is a combination of both vertical and lateral variation of clasts within the formation. In a vertical sense clast size does not appear to vary in any systematic form except in areas where the Brewer Conglomerate overlies an erosional surface and basal clasts are large. In a lateral sense there is a significant decrease in both maximum and mean clast size towards the south. Since some of the uppermost fans are almost horizontal this decrease in size probably reflects an actual decrease on the original fan surfaces associated with lower current intensities away from the apex of the fan. Conglomerate is virtually absent south of a line from Gosses Bluff to the Waterhouse Range and the pebbly sandstones of the Ljiltera and Undandita Members form most of the outcrops. Pebble size and
frequency continues to decrease southwards and in the James and Gardiner Ranges pebbles are rarely larger than 2.5 cm - the maximum size that can be transported in the upper part of the lower flow regime (Fahnestock and Haushild, 1962).

Size distributions of the clasts in the Polly Conglomerate have not been studied in sufficient detail to allow anything other than generalizations to be made. Thus the only known trend is for a decrease in size and an increase in roundness and sphericity of clasts down the paleoslope from Umbeara Homestead to the Black Hill Range.

Movement of debris in the pebble and cobble size range depends on the velocity and turbulence of flow and on the spacing of the clasts. Thus Leopold et al. (1966) have shown that a flow velocity of 0.31 m\(^3\) per second will not move 500 gram samples spaced one diameter apart while the same current will move 5000 gram samples if their spacing is greater than 5 diameters. Thus gravel accumulations and bars are a form of kinematic wave caused by particle interaction. Leopold et al. (1966) also noted that there was no relationship between transport distance and particle size for pebble and cobble clasts that can be entrained by the flow.

In very coarse material roundness changes are very rapid during turbulent flood surge conditions (Scott &
Gravlee, 1968). They noted that, with diorite clasts, changes from angular to subangular took place in a matter of 100 m and the clasts became rounded after travelling 2.5 Km. This very rapid rounding is partly due to the lack of fine detritus for cushioning effects. The rapid decrease in size downstream from -8 phi at 0.6 Km to -6.5 phi at 2.5 Km is mainly due to progressive sorting and only 10 percent of the size decrease is attributable to abrasion and breakage.

According to Ouma (1967) roundness evolves downstream at a medium rate in grades coarser than cobbles, faster in the cobble to granule range, and slowest in sub-granule grades. Roundness increases to a maximum downstream but with further transport it then declines as size decreases. In Fivemile Creek, Wyoming, Hadley (1960) has shown that gravel sized pebbles are rapidly rounded during early transport. Thus in the first 30 Km of transport roundness increases from 0.34 to 0.51 and sphericity increases from 0.52 to 0.65 on the Wadell scale. Beyond 30 Km increase in roundness and sphericity is attained relatively slowly. The following table after Hadley (1960) shows the variation of average roundness with distance, according to lithology:
Limestone pebbles become rounded in approximately 16 Km of fluvial transport (Sneed & Folk, 1958) while quartz and chert pebbles require larger transport distances. In this present study, limestone clasts are generally rounded whereas chert fragments are subangular to rounded. Quartz pebbles are all well rounded but they are rare and are probably reworked from earlier formations. Thus transport distance cannot be adequately defined although it was probably in excess of 16 Km during the deposition of most beds. At section AO sandstone and limestone clasts are subangular and up to -10 phi in size. They indicate that
the clasts were probably only transported a short distance from the source area. Fossil scree deposits are preserved at section GC where a breccia with angular clasts up to \(-11\) phi lies at the base of a presumed fault scarp. The deposit is unsorted and unstratified and probably represents non-transported debris.

In general, studies of pebble morphology are of very limited value in the Parndjara and Finke Groups because most of the clasts are sedimentary and they become rounded very rapidly. Even the limestone and chert fragments are moderately rounded which may indicate transport distances of 30 to 110 km. The use of the proportion of quartz pebbles in conglomerates as a climatic indicator (Dal Cin, 1968) could not be attempted in this study because the source area was not metamorphic and most of the quartz pebbles are probably reworked.

In the Colorado River, Bradley (1970) noted a distinct change in pebble size and composition with distance of transport. He noted that coarse granite diminished by 50 percent in size in about 250 km of transport whereas quartz and chert gravel decreased by 30 and 20 percent respectively. In this reach the Colorado River is essentially at grade and conditions are unfavorable for significant down-valley size sorting of the detritus (Mackin, 1963). Fresh granite would only be reduced in size by 10 percent in an equivalent distance of travel and Bradley (1970) explains the extra
40 percent reduction as due to more rapid abrasion of particles weathered during periods of temporary alluvial storage.

Changes in composition and size of pebbles in a downstream direction can be inferred from studies conducted across the Missionary Plain. On the northern margin pebbles are abundant in the Ljiltera Member and the range in composition is the same as that in the lower Brewer Conglomerate - namely a large variety of sandstone and siltstone clasts with less abundant limestone and chert clasts and rare quartz pebbles. The clast size varies from -2 to -8 phi with occasional sandstone cobbles being even larger. Roundness values vary from subrounded to well rounded and sphericity is moderate. At Gosses Bluff where equivalent horizons crop out, clast size rarely exceeds -6 phi and sandstone fragments other than quartzite are less abundant than to the north. There is little change in the average roundness or sphericity values. Still further south near the Gardiner Range pebble abundance has decreased considerably and they usually only occur as isolated clasts on bedding surfaces near the base of a coset. The size of the pebbles in these localities rarely exceeds -4.5 phi and the southward decrease in pebble size can probably be accounted for best by sorting as well as abrasion during transport. The mean roundness of the pebbles is generally within the rounded class and thus there is no direct relationship between roundness and distance. However, sphericity increases to the south
where most of the pebbles are ovoid to sub-spherical and discoid pebbles are less abundant. This may be an effect of sorting rather than abrasion since the pebbles were probably transported by flows in the upper part of the lower flow regime (Fahnestock & Haushild, 1962) where rolling is the main mode of movement of clasts of this size. Composition is also notably different in the south. The lithologies are less varied and quartzite, chert, limestone and quartz pebbles are the main representatives. A few siltstone clasts survive this distance of travel (50 Km) while non-cemented sandstone fragments are very rare. The presence of limestone clasts and the rarity of sandstone and siltstone clasts suggests that the latter were rapidly reduced by abrasion, and weathering (especially acid weathering) was minimal.

FLUVIAL MODELS

Fluvial models can be discussed from two points of view. The first is in terms of an environmental approach and the second deals with the structures produced in a single cycle of deposition.

The environmental approach to fluvial facies models was outlined very concisely by Allen (1965b). He outlined four major depositional facies found in fluvial environments undergoing aggradation and their characteristics are related to the river pattern and conditions under which they were deposited. The four facies are the
piedmont formed of alluvial fans, and deposits from a braided stream system, a low sinuosity stream system, and a strongly meandering stream system.

The alluvial fan system (Fig. 3.160A) is composed of a series of wedge shaped units of poorly sorted, coarse, clastic debris which thin radially away from the adjacent upland areas from which they are derived. The wedges usually intertongue with finer fluvial facies at the fan edge. Stream patterns on the fans may be braided or meandering depending on the size of the fan and on the conditions of climate, vegetation, debris load, gradient and discharge. The mean current direction is unimodal and approximately perpendicular to the upland area, and variance of the current direction is moderate (usually less than 360°, Selley, 1968). Gross internal features of the fan deposits are lenticular or sheet-like and the dominant sedimentary structures are trough and tabular cross-stratification in channel deposits and flat bedded sandstones on the fan surfaces (Bluck, 1965).

The braided stream floodplain environment (Fig. 3.160B) is dominated by deposits of bed-load detritus while fine grained and argillaceous overbank deposits are rare and usually very thin when present. The lack of overbank deposits may be attributed to erosion and reworking of such deposits as the relatively shallow river channels migrate rapidly across the whole flood plain, e.g. the Kosi River has migrated 115 Km to the
West in the last 220 years (Das, 1968). Mean current directions in a braided stream system generally show a low directional variance and in the Brahmaputra River variances of more than 1000 are uncommon (Coleman, 1969). The most important sedimentary structures in a braided stream system are trough cross-stratification of varying size, and flat bedded sandstones (Coleman, 1969; Doeglas, 1962).

Low sinuosity or straight stream floodplain environments (Fig. 3.160C) are also dominated by bedload detritus. The river is not confined to a meander belt by fine grained overbank deposits or clay plugs filling abandoned channels and it is therefore free to sweep backwards and forwards across its entire floodplain. Deposits formed by such a river consist of tabular to wedge shaped sheets of bedload detritus which may be overlain by discontinuous overbank deposits preserved beneath the erosion surface at the base of the next cycle of deposition. If almost complete cycles are preserved, it should be possible to distinguish braided and low sinuosity stream systems on a basis of cycle thickness in floodplains of approximately the same size. Low sinuosity streams would have a deeper channel than equivalent braided streams and would therefore produce thicker deposits in a single cycle. Mean directional features in low sinuosity streams have a small to moderate variance depending upon the degree of sinuosity of the talweg. Sedimentary structures dominant in low sinuosity stream deposits include large and small scale trough cross-stratification, tabular cross-stratification
and flat bedded sandstones (Harms & Fahnstock, 1965).

High sinuosity meandering stream floodplains (Fig. 3.160D) are characterized by channel sandstone deposits separated from each other by extensive vertical accretion floodplain deposits. The rivers are highly sinuous and are confined within a meander belt by the cohesive floodplain deposits and clay filled abandoned channels. The meander belt is frequently elevated above the surrounding floodplain and lateral movement of the meander belt occurs by avulsion. Levees and crevasse-splay deposits are also important features in the environment. Directional features have a high variance (commonly up to 6000), owing to the variation in channel direction within a meander belt e.g. channel directions range through 290° in the Mississippi River (Fisk, 1944; in Allen, 1967). Sedimentary structures within the deposits of a high sinuosity stream system form the classical fining upwards point bar sequence. This sequence is usually dominated by large and small scale trough cross-stratified sandstones and overbank deposits of parallel laminated siltstone interbedded with flat bedded and ripple-drift bedded fine sandstones.

These four main fluvial depositional facies are in fact just idealized stages in a continuum of sedimentary environments formed by rivers of variable pattern. Thus within a single flood plain a river may exhibit reaches of all or any of the three basic patterns due to local
conditions of gradient, grain size of detritus, erodibility of channel banks and effects of vegetation.

The point bar model was first proposed by Sundborg (1956) to describe the vertical association of sedimentary features produced by a single cycle of deposition on the inner bank of a meander loop. Further descriptions and amplifications of this model have been given by many authors including Bernard and Major (1963), Harms et al. (1963), Allen (1964b, 1965a,c, 1970), Visher (1965a,b), Moody-Stuart (1966) and Conolly (1969).

The general feature of all these models is based on the concept of lateral accretion during channel migration (Fig. 3.161, after Visher, 1965b) and the consequent idealized vertical sequence of sedimentary structures (Fig. 3.162, after Visher, 1965a). The basal erosion surface is cut into the underlying deposits and is overlain by the lowest poorly sorted coarse clastic sediments of the point bar profile. These may consist of lag gravel, pebbly flat bedded coarse sandstone, or trough and/or tabular cross-bedded sandstones of variable grain size. This zone is usually less than 1 m thick and transport is largely by bedload traction and rolling (Sundborg, 1956). The predominance of medium to large scale trough cross-stratification in the lower part of a point bar deposit of the Mississippi River has been described by Frazier and Osanik (1961). Individual sets vary in thickness from 10 to 150 cm with most being less than 1 m, and the
overall zone appears massive. Transport of detritus usually occurs in traction and suspension. Higher in the point bar, trough sets become smaller, tabular sets are more common, and frequently flat bedded sandstones form an integral part of the sequence. Flat bedded fine sandstones are probably deposited from a traction carpet moving over a plane bed. This may represent upper flow regime conditions since a plane bed with sediment movement occurs at lower velocities in finer sand than in coarser sand because resistance to flow caused by grain roughness is less (Simons & Richardson, 1966). The upper part of a point bar on the Mississippi River was found by Davies and Ehrlich (1966) to contain the following dominant structures: ripple-drift lamination, asymmetrical ripple lamination, minor areas of small scale trough cross-stratification and laminated beds with some convolute laminations in the finer sediments. Vertical accretion deposits overlying the point bar sequence and extending over the flood-plain consist of laminated siltstones interbedded with fine ripple-drift bedded sandstones.

Point bar deposits on the meandering Brazos River (Bernard & Major, 1963) consist of an ascending sequence from poorly bedded basal gravels to large scale cross-stratified fine sandstones. The upper portion of a point bar sequence may show contorted stratification, local scour surfaces and clay drapes. This sequence is then overlain by flood plain deposits of fine sand and red, grey and green mottled silt.
An account of overbank deposition in a floodplain environment has been given by McKee et al. (1967). They studied the deposits of a single flood in Bijou Creek, Colorado, and noted that the deposits varied in thickness from a few centimetres to a maximum of about 4 m. The deposits are usually characterized by flat bedded sandstones deposited during upper flow regime conditions. Other structures encountered include ripple-drift bedding, convolute laminations and small scale festoon and tabular cross-bedding. Sequences of structures are variable but typically start with flat bedded sands and progress upwards to cross-bedded and ripple-drift bedded sands. Only during the final waning stages of current action is there any appreciable deposition of silt and clay. This sequence is somewhat similar to some deposits attributed to channel deposition and thus their study indicates that the distinction between channel deposition and near-channel overbank deposition is not clear cut.

Deposition in braided and low sinuosity stream floodplain environments does not appear to differ radically from the basic fining upwards sequence proposed for deposits from meandering stream systems. The major differences between the deposits is in the proportion and thickness of each member of the fining upwards sequence, especially overbank deposits and this depends upon the rate and freedom of lateral migration of the channel, the depth and sinuosity of the channel, and the consequent size of the
scroll bars, point bars or braid bars. Thus, as noted by Allen (1968), there is a general hierarchical association of bedforms from small scale ripples and flat beds with parting lineation to large scale ripples and dunes, scroll bars, braid bars and point bars.

In low sinuosity portions of the Rio Grande, Harms and Fahnestock (1965) noted five main sedimentary structures which occur in the fining upwards cycle. Large to medium scale trough cross-stratification in sets from 6 to 70 cm thick is volumetrically the most important structure. It forms by dune migration in the upper part of the lower flow regime where water depths exceed 30 cm. Small scale trough cross-stratification in sets 3 cm thick forms as a veneer on the river bed and banks by ripple migration in the lower part of the lower flow regime. Tabular cross-stratification is less common than the above. It occurs in sets 6 to 70 cm thick formed by the migration of transverse or oblique bars in the lower flow regime. Horizontal stratification of sands is not a common feature in these sediments and it is produced by plane bed transport in the lower part of the upper flow regime. Parallel stratification of silt and clay is also a minor feature formed by the deposition of suspended material as flow waves. The latter is most common in areas of overbank flooding but also occurs as a thin coating over much of the channel. Mudcracks and rain prints occur in these sediments and the former produces platy clasts which are very commonly reworked in the succeeding sandstones.
Typical deposits of a low sinuosity stream system such as this, exhibit a fining upwards sequence of deposits similar to that of point bars. They consist of medium scale trough cross-stratified sandstones with set size decreasing upwards and frequently merging into small scale forms. Tabular cross-beds may occur in this sequence and they are most common as small sets near the top of the cycle. Horizontal stratification when present usually occurs in a variable position in the upper part of the sequence either above, below or in the middle of the small scale trough sets. Parallel laminations of silt and clay are infrequently preserved except as relatively thin sequences on the channel banks.

Similar sequences of sedimentary structures have been noted in sand predominated braided stream systems by Sakar and Basumallick (1968) and Coleman (1969). They noted that cross-beding, in cosets up to a couple of metres thick, forms the major portion of the preserved structure present in channel bars. There is rarely any coarse lag detritus and most cosets of cross-stratification overlie erosional surfaces and their lateral extent is very variable. Trough cross-beding is most abundant near the base of the cosets while tabular sets are often frequent towards the top. Trough thickness decreases upwards in each coset and many cosets contain ripple-drift bedded fine sands and silts. Smaller structures are preserved near the tops of shoals and bars and they consist of ripple-drift lamination, small scale trough
cross-bedding and flat bedding. Overbank levee and crevasse-splay deposits consist of poorly sorted silts and sands which display little internal structure except for occasional ripples. These massive portions are overlain by small scale trough cross-bedded, flat bedded and ripple-drift bedded sediments. Parallel laminated silts and clays are deposited in areas of slack water between larger bedforms in exposed parts of the channels and in overbank situations. Distorted and convolute laminations are common in most channel deposits of the Brahmaputra River. There are a large number of forms of distorted stratification and the most common are due to slumping and shear flowage.

Williams and Rust (1969) studied a braided portion of the gravelly Donjek River, Yukon, from a sedimentological aspect. The river forms a rather specialized case of a braided stream system since it contains abundant coarse, angular, glacial detritus, discharge is variable, it is largely frozen for part of the year, and is degrading. The river is thus characterized by abundant gravel detritus and this in turn affects the sedimentary structures present, although they are generally similar in form to those of sandy rivers.

The preceding discussion and description of vertical profile models has been restricted to deposits typical of flood plain environments. However, while most high sinuosity deposits occur in such environments, low
sinuosity and braided streams are also the main stream patterns associated with alluvial fan deposition.

The sediments in an alluvial fan vary from boulders and gravel, common near the apex, to finer grained sediments in a downstream direction. Discontinuities in the fan profile may be related to grain size diminution (Yatsu, 1955) or to tectonic activity, climatic change or a change in local base level (Bull, 1964).

Park (1967) briefly described Triassic border conglomerates from New Jersey where the clast sizes range from -2 to -9 phi and are predominantly in the range from -5 to -7 phi. The deposits are poorly stratified and individual beds are usually lenticular. The conglomerates are still recognizably cone shaped and they are considered to be fanglomerates deposited normal to the fault boundary of the basin of deposition.

Stratification in the fanglomerate deposits studied by Bluck (1965) was found to be poorly developed with large, low angle, tabular and trough cross-bed sets occurring in channel fill areas while flat bedding was predominant on the fan surfaces. Medium scale tabular cross-bedding is more common near the fan-head while trough cross-bedding becomes more common down the fan. Imbrication is common among discoidal clasts in fanglomerate sediments. Discoidal pebbles become relatively more abundant down the fan since their hydraulic geometry permits them to be
carried in suspension whereas more oblate clasts are rolled.

Thus in an alluvial fan system the typical vertical association of sedimentary structures would differ quite markedly from that of the point bar sequence even though fining upwards cycles may still be recognizable. Also, the type of stratification would vary depending upon the climatic conditions under which the fan accumulates. In semi-arid regions flows are infrequent and usually highly viscous because of high sediment load. Mudflows often result and stratification in such deposits is very poorly developed. In areas with a higher rainfall and more frequent discharge individual flows carry less sediment load and stratification is more apparent in the resultant deposits.

An alluvial fan model would contain three main depositional elements. An erosional channel surface would typically be covered with a variable thickness of poorly sorted coarse clastic detritus. This may consist of massive conglomerate or coarse sandy gravel and if stratification is preserved it would probably be flat bedding or antidunes formed under upper flow regime conditions. Overlying this coarse interval, which would be equivalent to the lag gravel deposits of the point bar profile, would be an interval of fine to coarse and pebbly flat bedded sandstone and medium scale tabular cross-sandstone representing channel fill deposits formed in the transitional flow regimes. Near the fan head this interval would be thin
whereas nearer the fan margins it would become thicker and contain more abundant trough cross-stratification. Overbank deposits are usually thin and represented by flat bedded fine to coarse sandstones. They may be ripple-drift bedded at the top due to waning current action.

From the foregoing study of detailed measured sections (p.p.141-166), and from other exposures noted during the course of this study, it is apparent that five basic cyclic sequences are involved in the formation of sandstone units in the Pernjara and Finke Groups. One of these sequences is of aeolian origin (section CY, Fig. 3.137) with extensive, large scale, low angle, curved planar cross-stratified sets. The sets are composed of fine to medium grained sand, lack mudclasts or pebbles and are quite distinct from the predominantly trough cross-stratified sets of the fluvial deposits. The remaining four types of cycles are of fluvial origin and are simply distinctive forms in a continuum of cyclic sequence variations. They are presented as generalized cycloths Figures 3.151-156 and they show almost all the features of the idealized fining upward fluviatile sequence. However, overbank deposits are rare in all the sandstone units studied with the exception of the lower Langra Formation, and the lower Hermannsburg Sandstone in the south-east.

The coarsest fluvial deposits are those of the Ljiltera Member and Brewer and Polly Conglomerates. The Brewer Conglomerate is predominantly massive with little
or no sign of stratification. Cyclicity of bedding structures is therefore absent even on a large scale. Within the Ljiltera Member and the Polly Conglomerate massive horizons of conglomerate alternate with flat bedded and very rarely cross-bedded, pebbly, coarse sandstones. The basic sequence is shown in Figure 3.151. The coarse and frequently massive conglomerate at the base of the cycle always overlies an irregular erosional surface. Asymmetrical channels up to 1.5 m deep have been recorded and are filled with poorly sorted conglomerate. Smaller channels and surface irregularities are common. The conglomerate consists of a poorly sorted admixture of sand, pebbles, cobbles and boulders up to 50 cm. Most of the coarse material is subrounded to rounded, ovoid, and only rarely shows upcurrent imbrication of the more discoidal clasts. The contact between the massive conglomerate and the overlying pebbly, coarse sandstone is usually flat and gradational. The pebbly sandstone is either massive, flat bedded or low angle cross-stratified in tabular sets. It is usually much thinner than the underlying massive conglomerate and often only occurs as thin, lensoid stringers. The thickness of these cycles varies from 50 cm to 12+ m with an average of about 2 to 3 m. The larger cycles may represent indistinguishable remnants of several smaller cycles where erosion surfaces cannot be recognized and sandstone stringers are absent. Conglomeratic cycles are volumetrically important within restricted stratigraphic intervals at the top of the Pertnjara Group and at the base of the Finke Group.
The cycle is almost identical to that proposed for deposition in alluvial fans in areas with moderate or seasonal rainfall. The sediments were largely deposited from currents in the upper flow regime at a moderate distance from the fan head. The massive nature of the Brewer Conglomerate indicates that sorting and removal of most of the fine detritus took place, and this would probably only occur where rainfall was moderately heavy and most of the transport on the fan took place under normal aqueous fluvial conditions rather than as mudflows. Extremely similar deposits have recently been described in detail from the Solund District, Norway (Nilsen, 1968a, b; 1969). Nilsen also considered that the conglomeratic sediments had been deposited as part of an extensive alluvial fan system and he noted that the sediments became finer and represented braided stream deposits towards the fan margins.

The second type of cycle is volumetrically the most important constituent in the sandstone members of both Groups. The three most typical forms of this type of cycle are shown in Figures 3.152-4. They vary in thickness from 1 to 7 m with most of them being in the 2 to 3 m range. The most common form of this cycle consists almost entirely of trough cross-stratified sets of fine to coarse sandstone overlying a flat to irregular basal erosion surface (Fig. 3.152). Individual sets near the base of each cycle frequently contain mudclasts aligned parallel to the bedding, and occasionally contain small, rounded, sedimentary pebbles. However, many sets are devoid of included clasts.
The trough cross-stratified sets usually decrease in thickness and in inclination towards the top of each coset. Likewise, the frequency of cross-strata being asymptotic to the base of the set, decreases up the coset. The diminution of grain size up the coset has been noted in many instances but it is not an invariable characteristic of these cycles. Overlying the trough cross-stratified coset the cycle may pass up into thin horizons of one or more of the following structures - flat bedded sandstone with current lineation, ripple-drift bedded sandstone, isolated ripple sets, or laminated, sandy, coarse siltstone. These upper units are usually very thin and are frequently discontinuous or absent. Thus sequences of trough cross-stratified cosets often overlie one another. Their cyclicity is still apparent from the variation in cross-bed thickness and the presence of mudclasts and pebbles. Figure 3.153 shows a common variant of the above cycle with the trough cross-stratified coset overlain by a medium scale, tabular, cross-stratified set and then the flat and ripple bedded sandstone and siltstone. A much less common variation (Fig. 3.154) starts with a large, low angle, tabular, cross-stratified sandstone overlying a fairly flat basal erosion surface. This is overlain by a coset of small to medium scale trough cross-stratified sandstone prior to the flat or ripple bedded sandstone and laminated siltstone.

This second type of cycle and its variants have extremely similar characteristics to the models for low
sinuosity and braided stream deposits and they differ from the high sinuosity stream point bar deposits in the lack of vertical accretion deposits. These deposits are also very similar to deposits described by Allen and Friend (1968) from Devonian Catskill facies in the Appalachian region. They considered that the Catskill sediments were deposited by a complex of river channels, possibly braided, that were unrestricted in their lateral wanderings. A similar mode of deposition can be suggested for most of the Pertnjara and Finke Group sandstones although the distinction between braided and low sinuosity stream deposits cannot be determined with certainty. The only indication is from the average thickness of the cycle compared with the extensive area of deposition. Thus, since the average cycle is only 2 to 3 m thick, deposition is most probably from a shallow braided stream system rather than a large number of shallow low sinuosity stream systems.

The third type of cycle is rather similar to the classical point bar profile although vertical accretion deposits are usually very thin. This type of profile (Fig. 3.155) is only common in the basal Hermannsburg Sandstone south of Gardner and James Ranges and in the lower half of the Langra Formation. The profile has its fullest development in the latter where the basal erosion surface is overlain by a flat bedded lensoid conglomerate of rounded sedimentary, metamorphic and granitic pebbles and cobbles set in a poorly sorted sand matrix. Small
asymmetrical channel features up to 50 cm deep are not uncommon, while broad erosional features with only one visible bank up to 80 cm high are more frequent. These broad features usually contain lenses of conglomerate and are probably formed by lateral migration of a channel rather than the presence of a very large channel. The conglomeratic horizons may be overlain by coarse, flat bedded, pebbly sandstones and these are overlain in turn by a coset of trough cross-stratified sandstones with trough thickness and inclination decreasing up the set, as does grain size and the number of included clasts. Small tabular sets may be present towards the top of the coset and this unit usually passes up into either flat bedded fine to medium sandstones or small scale "festoon" trough cross-beds. Both these units may be present in either order and in a complete cycle they are overlain by interbedded ripple-drift bedded sandstone and laminated siltstone. The most frequent variant of this cycle is the absence of the basal conglomerate and flat bedded sandstone in all occurrences in the basal Hermannsburg Sandstone. The thickness of these cycles is usually in the range from 2 to 7 m with the average being about 3 m. The basal conglomerate and flat bedded sandstone rarely reach a thickness of 1 m and likewise the thickness of the interbedded fine sandstone and siltstone at the top of the cycle is usually in the range from 10 to 70 cm.

This third type of cycle, especially as it is developed in the lower Langra Formation, possesses all
the characteristics of the point bar sequence. The sediments were probably deposited by a relatively high sinuosity stream flowing over a broad alluvial plain and the vertical accretion deposits often show desiccation features. The coarseness and abundance of lag gravel indicates that flow conditions must have frequently been in the upper flow regime and suggests that the source area was not too far distant. The cycles present in the basal Hermannsburg Sandstone, where it overlies the Amulda Member in south-eastern exposures, lack this basal lag gravel. This suggests that the deposits were laid down at a greater distance from the source area than those of the lower Langra Formation and that the depositional currents were dominantly restricted to the lower flow regime.

The fourth type of cycle is the least important volumetrically and is restricted in occurrence to sequences with abundant fine sand or silt. It constitutes most of the Amulda Member and parts of the basal Hermannsburg Sandstone in the Mt Duff area. The sequence is basically one of a relatively thin, massive or low angle, trough cross-stratified, basal sandstone unit overlying a flat or irregular erosion surface and overlain in turn by a thicker sequence of flat bedded, laminated or ripple bedded fine sandstone and/or siltstone (Figure 3.156). The fine sandstones and siltstones may be interbedded and variations in this type of cycle are largely confined to variations in the ratio of these
two constituents. These cycles are consistently thinner than the preceding forms and most are only $\frac{1}{2}$ to 1 m thick. Occasionally large sequences of interbedded sandstone and siltstone were encountered and rare sections reached a thickness of 5 m.

This fourth type of cycle is also basically similar to the classical point bar sequence although the streams were dominantly small and the currents were restricted to the lower flow regime. The abundance of vertical accretion deposits indicates that the area of deposition was frequently subjected to flooding with the consequent deposition of fine grained material. The stream systems were probably of a highly sinuous nature and they probably represent a distributary system developed between the low sinuosity or braided stream floodplain or alluvial fan sandstone deposits and the fine grained playa or lacustrine deposits of the Dare Siltstone Member.

PROXIMAL ANALYSIS OF FLUVIAL DEPOSITS

The concept of proximal analysis (R. Walker, 1967) could theoretically be applied to the study of cyclic sequences of fluvial deposits to determine proximity to the main channel area. In point bar deposits a proximity index could be determined from the ratios of cycles starting with lag gravels or flat bedded pebbly coarse sandstones, to cycles starting with cross-stratified sets, to cycles starting with vertical accretion deposits of
ripple-drift bedded sandstone and siltstone. However, it would be impossible to distinguish the latter deposits from normal overbank deposits formed in a thicker cycle because in a single flood several horizons of siltstone, sandstone and ripple-drift bedded sandstone may be deposited (McKee et al., 1967). In fluvial cycles a more reliable proximality index would involve thicknesses of the three main portions of the cycles as outlined above. The ratio of these thicknesses may be plotted on a ternary diagram and the distance from the conglomerate and flat bedded pole may be used as an index of proximality. Alternatively a numerical index may be derived to denote the proportion of conglomerate and flat bedded sandstone (A), and cross-stratified sandstone (B), in the whole cycle (T). Thus following R. Walker (1967) the proximality index (P) expressed as a percentage is:

\[ P = \left( \frac{A + \frac{1}{2}B}{T} \right) \times 100 \]

Proximality indices may be considered for each cycle separately or the interpretation may be simplified if average proximality indices are calculated as a moving average over a number of beds as proposed by R. Walker (1967).

An alternative approach to proximality analysis in cyclic fluvial sequences could be aimed at determining the relative distance from the source area by means of current competence. Thus progressing away from a source area, conglomerate should give way successively to increasing proportions of pebbly sandstone, sandstone and siltstone. At the same time stream power should decrease and stream sinuosity should increase unless there have
been substantial volume or load contributions from tributaries. One method of approaching this problem is to plot relative abundances of conglomerate-sandstone-siltstone, or conglomerate+pebbly sandstone-sandstone-siltstone, on ternary diagrams. With such a system the proximity to the conglomeratic pole would be a measure of proximity to source area. Plots of cyclothems of increasing distances from the source area in an aggrading system would be expected to follow a curved path from the conglomeratic pole to the siltstone pole via the sandstone pole. A sequence following this general course was noted when average cyclothem compositions for the Hermannsburg Sandstone were plotted on such a ternary diagram (Fig. 3.157). The second method of approach, in terms of stream power and sinuosity, was proposed by Allen (1970) in a model for coarse member cyclothem deposition. He considered the ratios of medium scale cross-stratified sandstones \((B_1)\) to flat bedded sandstones \((B_2)\) to small scale cross-stratified and ripple-drift bedded sandstones \((B_3)\) in terms of a ternary diagram. Allen concluded that proceeding from cyclothems dominated by \(B_3\) deposits to ones dominated by \(B_1\) deposits is equivalent to changing from a low-powered to a high-powered depositional regime. Also, moving from the \(B_3B_1\) axis to the \(B_2\) apex is equivalent to depositional systems changing from high sinuosity to low sinuosity. Most of the cyclothems in the present study are dominated by medium scale cross-stratification and the next most common form is flat bedded sandstone. Small scale cross-stratification and ripple-drift bedding is a
rare feature except in the basal Hermannsburg Sandstone in southern exposures. Thus according to the proposal of Allen (1970) the streams depositing most of the sandstone units were relatively high powered and of moderate to high sinuosity. This conclusion is further substantiated by the general lack of vertical accretion overbank deposits and by the low average thickness (2.38 m) of individual cyclothsms. All the above factors point to deposition in shallow, broad stream systems which migrated rapidly back and forth across the whole of the alluvial plain. The actual form that this stream system takes cannot be ascertained since the characteristic depositional forms in equivalent modern stream systems are not known in sufficient detail to allow differentiation, if in fact one stream system can be differentiated from another using sedimentary structures alone. Both braided and low sinuosity streams could produce the structures present in the Pertnjara Group since in both cases the talweg(s) are sinuous within the overall stream channel system. The presence of ripple-drift bedding in the basal Hermannsburg Sandstone in the south, and the presence of vertical accretion deposits in the cycles within the Amulda Member, both suggest that stream power decreased to the south and a meandering stream system was probably responsible for the deposition in this area. Following Walther's Law, the increasing stream power and decreasing sinuosity with increasing stratigraphic height from the Amulda Member to the Ljiltera Member, indicates that either the proximity to the source area increased or the stream power increased
substantially with time. Alternatively, a more static condition of continual increase in size of a coalescing series of alluvial fan systems to the north could cause the fans to gradually encroach on the finer grained lacustrine or playa sediments to the south. However, the increase in conglomerate and pebble content in the Ljiltera Member and Brewer Conglomerate to the north indicates that source area effects became increasingly obvious. The probable explanation for the changing vertical succession, therefore, lies in a combination of both the above hypotheses.

The sequence in the Finke Group is the reverse of that in the Pertnjara Group. Thus the basal Polly Conglomerate gives way with increasing stratigraphic height to the sandstones of the Langra Formation and the siltstones of the Horseshoe Bend Shale. The Polly Conglomerate becomes thinner northwards from Umbeara Homestead and probably wedges out north of Black Hill Range. Subsurface in McDills No. 1 Well, the thickness of the Polly Conglomerate increases which suggests that the unit extends a considerable distance eastwards. The overlying lower Langra Formation contains fining upwards cyclothems consisting of a basal conglomerate, cross-bedded sandstone, and massive siltstone. This indicates that with increasing stratigraphic height either the proximity to the source area decreased or the stream power decreased, or both, with time. This process continues with decreasing pebble and coarse sand content up the Langra Formation. Thin,
fine sandstone bands are present in the lower Horseshoe Bend Shale but are rare in the middle and upper parts. This sequence probably reflects the wearing down of a moderately elevated granitic and metamorphic terrain situated south-east of the present Amadeus Basin.

However, not all the sequences shown in the vertical profile analysis can be attributed to deposition in a fluvial or flood plain environment. Such sequences include the Deering and Dare Siltstone Members, the lower Harajica Sandstone Member on Dare Plain, and the Horseshoe Bend Shale. All these deposits are characterized by parallel bedded predominantly fine grained detritus deposited as widespread thin sheets. Such sediments may be compared with known sequences in lacustrine and deltaic environments.

LACUSTRINE MODEL

The vertical sequence of sedimentary structures to be found in a lacustrine environment is shown in Figure 3.163 (after Visher, 1965a). The extensive parallel laminated and ripple laminated very fine grained sediments of the Deering and Dare Siltstone Members and the Horseshoe Bend Shale were probably deposited in a lacustrine environment. The lack of reworked sediments and organic remains is also indicative of a lacustrine rather than a tidal flat environment. The Deering Siltstone Member is thin bedded throughout and contains current and slump structures, ripple
marks and parallel laminated bedding. There are very few structures indicative of subaerial exposure and the beds are all uniformly red-brown coloured with virtually no mottling. A fragment of fish plate was found from the middle of this member on Dare Plain. This member probably resulted from deposition in a permanent body of fresh water. The Dare Siltstone Member and the Horseshoe Bend Shale differ from the Deering Creek Member in that they are characterized by cyclic deposition and periodic subaerial exposure. Mudcracks, rain prints and halite pseudomorphs all point to subaerial exposure while parallel and ripple bedded micaceous siltstones indicate shallow water deposition. The massive red-brown portions of each cycle have virtually no internal features and they may have originated either as deeper water deposits or as loess deposits. The overall environment was probably very extensive, flat and playa-like and the lake was ephemeral with alternate inundation and desiccation.

DELTAIC MODEL

The vertical sequence and characteristics of deposits in a deltaic environment has been outlined by Visher (1965a, Fig. 3.164). The littoral, delta fringe and pro-delta deposits described by Visher have the same characteristic features as those found in the lower Harajica Sandstone Member on Dare Plain. Parallel bedding, wavy bedding, slump and current structures are all preserved as extensive sheets in the lower Harajica Sand-
stone Member and the lack of organically reworked sediments lends support to the hypothesis that sediments were deposited as a delta into a fresh water lake, rather than a marine environment. In the middle Harajica Sandstone Member on Dare Plain the deposits become coarser, more sandy, and trough cross-bedding is the dominant sedimentary structure. This interval is also characterized by abundant slump bedding. The presence of lustre mottling and calcite cement in these sandstones suggests that they were deposited in a fluvial rather than a lacustrine environment (Picard & High, 1968).

Towards the top of the Harajica Sandstone Member the sequence is reversed and silty horizons become more prominent again and individual thin beds have a wide lateral extent. Mud cracks and ripple marks are abundant and rain prints were also recorded from this interval. This indicates periodic subaqueous and subaerial conditions probably in a near shore environment as suggested by the rill marks in associated sandstones which may represent beach deposits. Thus the Harajica Sandstone Member on Dare Plain probably represents deposition from a transgressive and then regressive deltaic environment extending southwards into a large lake.
PALEOCURRENT ANALYSIS

Paleocurrent analysis of the Pertnjara and Finke Group sediments was undertaken to determine the directional trends and variability of the currents operating at the time of deposition. The main source of directional information came from the analysis of cross-bed azimuthal dip directions. A total of about 3600 such measurements was made, and most of these came from the Hermannsburg Sandstone. Measurements from the Finke Group were few because of the small number of sections studied. Other directional features studied include ripple marks, current lineation, grain and pebble trains, imbrication, and long axis orientation of included clasts. All directional measurements were corrected for magnetic diversion using the average value of 4.5° (Parkinson, 1965) for the whole basin.

Cross-bed measurements were taken from various stratigraphic horizons on all measured sections as well as from several spot localities within the basin. Cross-bed measurements on measured sections were usually taken within a 20 m interval about the stratigraphic height indicated on Table 3.6 and were not taken across obvious lithologic boundaries. Usually between 3 and 20 readings were taken at each horizon depending on the exposure, while some well exposed localities had up to 30 readings taken. Ripple marks, current lineation and other
directional features were also noted at these localities.

A computer programme was designed to correct the cross-bed readings for tectonic tilt and to bring them back to an original horizontal attitude for comparison. The programme is a modification of the programme given in Cook (1966), and it is presented in Appendix 4, along with modifications of the basic programme, to obtain average azimuthal directions over various stratigraphic intervals. However, the use of cross-stratification as an indicator of paleocurrent directions must be viewed with caution (Friend, 1965) since there may be a large scatter of values even within a single coset. Thus a large number of readings should be taken from any single locality to give a reliable pattern of current trends. Meckel (1967) noted that the variation in direction was much greater in trough cross-stratified units (standard diversion 60°) than in tabular cross-stratified units (standard diversion 24°) in the same deposit.

In meandering river systems there is generally a high variance associated with the measurement of preserved directional features (Allen, 1967). Generally fluvial systems exhibit variances in the order of 0 to 6000 (Potter and Pettijohn, 1963) and the variance of directional features in meandering stream systems is usually in the upper part of this range. The channel reaches may show a variation of between 75° and 290° in a meander belt while variation between sand waves in
a channel reach is $110^\circ$, and the variation of directional features within a sand wave is $170^\circ$ (Allen, 1967). Thus the variance and standard deviation of cross-bed readings from such an environment would be expected to be large.

In the braided Donjek River, Canada, Williams and Rust (1969) noted that there was generally little variation in the direction recorded by individual features from that of the vector mean of the same features. Current directions of the stream bed range over an arc of $160^\circ$ although their mean vector is unidirectional in a downstream direction. Thus both large and small scale current directional data give a good approximation of the orientation of the composite stream channel.

In the Brahmaputra River variability of current direction is related to the width of the channel (Coleman, 1969). Thus in reaches where the width is 1.5 to 3 Km, variation in channel direction ranges from $15^\circ$ to $52^\circ$ whereas with a width of 13 to 16 Km the range is from $101^\circ$ to $112^\circ$. Cross-bedding orientation in such a situation would have an even larger variability and a large number of readings would be required over a large area to provide an accurate interpretation of the paleo-current direction.

In general the variance encountered in the measurement of directional features from braided stream deposits is lower than that for meandering streams (Selley, 1968).
This was confirmed by Coleman (1969) who recorded variance values of less than 1000 from all measured stations in the Brahmaputra River.

With alluvial fan deposits Kelling (1969) noted that even the vector mean was only a crude representation of the currents responsible for the deposition of the fan. This is because of the radial nature and rapid migration of stream features in the fan.

The results from the cross-bed analysis are presented in Figure 3.165 which gives the mean azimuthal current directions for all beds at each location together with twice the standard deviation (95% confidence level) at each point. The length of the arrow showing the mean current trend is proportional to the vector magnitude at that point expressed as a percentage. Maps were also constructed to show the variation of current trends at each stratigraphic interval and for each 100 m interval. In almost all cases means for these intervals varied randomly through the interval, indicated by the 95 percent confidence limits of the overall mean for the locality. Thus changes in overall current directions throughout the period of deposition sampled are not significant and their interpretation is hindered by uncertain time correlation of stratigraphic intervals within the southward thinning sedimentary units. The only non-random variation of current directions with increasing stratigraphic height were recorded from the four localities on the north-west
flank of Gardiner Range. Each of these localities has a reasonably large standard deviation owing to the fact that currents on the lower part of the column trend north-westwards while those higher on the column trend towards the north-east. A simplified map showing the means of the group vectors and the interpreted paleocurrent system is given in Figure 3.166.

Paleoslope determinations in a fluvial environment can be estimated from the vector mean current direction patterns since the depositional currents are usually unidirectional in a downcurrent sense. Thus paleocontours and paleotopography can be estimated in a general fashion from the paleocurrent directions and reconstructed estimates of the paleocontours are indicated on Figure 3.165.

The range of variance values recorded at the group level in the present study is presented in Table 3.6 together with mean current trends and tests for significance. A total of 25.6 percent of the groups had a variance of less than 1000, while 55.8 percent had a variance in the range 1000 to 6000, 10.7 percent in the range 6000 to 12000, and 7.9 percent had a variance in excess of 12000. Groups which have variance values in excess of 12000 are predominantly composed of only 3 or 4 cross-bed readings and they generally have a random distribution using the Rayleigh test. Likewise several of the groups with variances in the range from 6000 to 12000 also have a random distribution. If these randomly distributed samples are ignored in the
analysis most of the samples fall in the range of variances (0 - 6000) normally attributed to fluvial deposition (Potter and Pettijohn, 1963; Kelling, 1969). An interesting feature is that while variance remains almost constant throughout the Pertnjara Group there is a noticeable increase in the variance values from the south-eastern outcrops from Mt Duff to the southern flank of Camel Flat Syncline. This general increase in variance down the paleoslope may correspond to an increase in sinuosity of the river systems in this area.

The thickness of individual cross-beds was noted by Kelling (1969) to increase in a downstream direction towards the centre of the basin. The increase in size is linked to the diminution of hydraulic energy down an alluvial fan towards the centre of a basin because of the decrease in slope. Studies of the average thickness of cross-beds from the Pertnjara Group indicate a similar variation in part of this study area and a reverse trend further down the paleoslope. Thus cross-beds from localities on the northern margin of the basin have a mean thickness of 61.87 cm and this increases to a mean thickness of 77.32 cm for all localities in the central and southern parts of the Missionary and Brewer Plains. Individual localities with the largest cross-bed thicknesses are generally found in the Undandita Member in the central part of these plains. When measurements are restricted to localities from the Hermannsburg Sandstone a similar increase in thickness down the paleoslope was
noted although it is not as pronounced. South-east from
the Brewer Plain cross-bed thickness decreases to a mean
of 48.30 cm in the Rodinga Sheet Area. While the increase
in cross-bed thickness may be attributed to a diminution
of hydraulic energy, the decrease in thickness further
down the paleoslope could be due to a loss of water owing
to soakage and a still further decrease in the hydraulic
energy.
4.1 PETROLOGY OF THE SANDSTONES

The sandstones throughout the study area are remarkably uniform in mineralogy and textural features. A total of 454 thin sections have been studied and the important mineralogical features are discussed below.

Modal analyses of all samples were carried out using a fourteen digit point counter usually at a magnification of 80X, although finer rocks were studied under higher powers. The errors incorporated in the point counting technique have been discussed by Chayes (1956) and Bayly (1960, 1965). The two main sources of error are controlled by the count length and the area sampled but also depend on the coarseness and variability of the grain-size of the rock. The average area of each section studied was about 300 mm$^2$ and the average median grain size ranges from 2 phi to 4 phi. The counting variance (Vc) equals $p(100-p)/n$ where $p$ is the percentage of the mineral counted and $n$ is the number of points counted (Chayes, 1956). For a count length of 400 points and a mineral forming 50 percent of the sample the counting variance equals 6.25. The coarseness index (IC) of the rock depends upon the length in which 50 grains are encountered. This can be generalized to 50 times the mean diameter of the grain size distribution expressed in millimetres. Thus the coarseness index of the sediments studied
varies from 100 to 250 and the equivalent sampling variance (Vs) for a mineral forming 50 percent of the sample varies from 0.3 to 1.8. The standard error \( \sqrt{Vs + Vc} \) lies in the range from 2.6 to 2.85 percent with the lower value representing the finer grained rocks and the larger one representing the coarser samples. Increasing the number of points counted produces a reduction of the standard error to 2.1 percent and 1.75 percent for 1000 and 2000 points respectively in the coarser samples, but it was decided that in this study the increased accuracy was not warranted with the time available and the number of samples to be studied. As it was, over 190,000 points were counted in this study most of which were for modal analyses. The expected standard errors (Bayly, 1960) for a count length of four hundred points on the Pertnjara and Finke sediments varies as follows:

<table>
<thead>
<tr>
<th>Percentage in sample</th>
<th>Standard error</th>
<th>Average standard error in framework grain percentages</th>
</tr>
</thead>
<tbody>
<tr>
<td>1%</td>
<td>± 0.5 – 0.6%</td>
<td>± 0.5 – 0.6%</td>
</tr>
<tr>
<td>3%</td>
<td>± 0.9 – 1.0%</td>
<td>± 0.9 – 1.0%</td>
</tr>
<tr>
<td>5%</td>
<td>± 1.1 – 1.2%</td>
<td>± 1.2 – 1.3%</td>
</tr>
<tr>
<td>10%</td>
<td>± 1.5 – 1.7%</td>
<td>± 1.6 – 1.8%</td>
</tr>
<tr>
<td>15%</td>
<td>± 1.8 – 2.0%</td>
<td>± 1.9 – 2.1%</td>
</tr>
<tr>
<td>25%</td>
<td>± 2.2 – 2.5%</td>
<td>± 2.4 – 2.6%</td>
</tr>
<tr>
<td>50%</td>
<td>± 2.6 – 2.8%</td>
<td>± 2.8 – 3.0%</td>
</tr>
<tr>
<td>75%</td>
<td>± 2.2 – 2.5%</td>
<td>± 2.4 – 2.6%</td>
</tr>
</tbody>
</table>
These standard errors are probably overestimates of the true errors, but using the equation derived by Bayly (1965) they only overestimate by 0.1% at the most. The above figures can therefore be used as maximum standard errors to be applied to the modal analyses given in Tables 4.1 and 4.3 (Appendix 5).

All rock and thin section numbers, except those labeled B.M.R. in Appendix 1, refer to specimens and thin sections held in the petrology collection of the Department of Geology, The Australian National University.

Quartz

Quartz is the most abundant mineral in all the thin sections studied and on an average forms 75 percent of the rock. The following varieties of quartz were recorded in this study:— (1) simple grains with straight extinction or extinction angles of less than two degrees; (2) simple grains with undulose but continuous extinction with extinction angles greater than two degrees; (3) semi-composite grains consisting of two or more interlocking grains with distinct undulose or straight extinction positions in very close optical orientation; (4) composite grains with undulose extinction and smooth or crenulated grain boundaries. This subdivision follows the empirical classification of Folk (1968) with the omission of two classes. The simple grains are here only divided into straight and undulose since, as Blatt (1967)
pointed out, true extinction angles cannot be measured on a flat stage. Composite metamorphic quartz was not recorded as a separate entity because of its rare occurrence. It was recorded, along with schistose metamorphic quartz, as a metamorphic rock fragment. Other features recorded for quartz grains include grain shape, number and type of inclusions, and for disaggregated samples, grain sphericity and surface textures.

Simple grains of quartz with straight extinction are the most abundant constituents in all the sandstones studied (Table 4.1). The average content varies between 40 percent and 60 percent of the framework grains and it is fairly uniform throughout the Pernjara and Finke Groups. Conglomeratic formations are exceptions to this general rule and they contain less quartz than the sandstones e.g. Brewer and Polly Conglomerate. Quartz grains with straight extinction are generally subrounded to well rounded, are free of, or contain a few or moderate number of, vacuole inclusions, and infrequently contain acicular crystalline inclusions. Rare rounded grains were noted with worn and rounded overgrowths. This type of quartz is the most common variety present in most sedimentary rocks and could have been derived originally from almost any relatively unstrained or unrecrystallized source rock. High proportions of straight extinction simple grains can point to a sedimentary source where the other quartz types have been selectively eliminated by prolonged polycyclic abrasion (Folk, 1968).
Quartz with undulose extinction is usually the second most abundant quartz type forming an average of 10 percent to 25 percent of the rock. The quantity of undulose extinction quartz is not directly related to the quantity of straight extinction quartz and it is not related to the intensity of folding undergone by the sediment. Thus samples from overturned and steeply dipping sections and from near large faults, do not contain appreciably more strained quartz. The strained nature of this quartz class is, therefore, a primary detrital feature not related to post-depositional effects. Undulose extinction quartz consists of simple, single grains which are generally subrounded to rounded and may contain variable numbers of vacuoles and acicular inclusions. Acicular inclusions typically occur in grains with only a few vacuoles and are generally absent from grains with abundant vacuoles. The latter generally have low extinction angles and probably represent reworked vein quartz. The general roundness of most of the grains in this class indicates that they are probably of polycyclic sedimentary origin, although they may have passed through fewer cycles than most of the more abundant straight extinction quartz.

Semicomposite and composite quartz grains form a minor part of the quartz assemblage - generally less than 8 percent and 2 percent of the framework grains respectively. Both vary in shape from subangular to rounded and contain a variable number of vacuoles but few acicular inclusions.
Individual grain shapes in semicomposite quartz grains are variable but they usually have planar grain boundaries. In composite quartz grains the individual grain shape is either equant with planar boundaries or stretched with planar or irregular boundaries. Regional variation of these two quartz types is random but at any one locality, with the Ljiltera Member or Brewer Conglomerate present, there is an increase in quantity of these quartz types with increasing stratigraphic height. This suggests that at least some of these grains are probably first cycle derivatives from the gneissic and granitic terrain exposed by erosion of the overlying sediments. In the Finke Group these quartz types are most abundant in the Polly Conglomerate and Langra Formation also indicating a first cycle metamorphic or plutonic origin for the grains.

Miscellaneous quartz types, distinctive enough to be recorded separately, are very rare and generally form less than one percent of the framework grains. In almost all cases these grains represent graphic intergrowths of quartz and potassium feldspar, with the latter usually forming a lattice pattern in the quartz. Such grains only occur with any consistency at stratigraphically high levels in the Hermannsburg Sandstone in the MacDonnell Ranges.

A systematic study of quartz grains from all the disaggregated examples yielded the following results.
Angular and subangular grains are rare except in the fine and very fine sand fractions where subangular grains form a minor part of the assemblage. Most angular and subangular grains have a low sphericity and are anhedral, although a few show prismatic crystal faces. Most of the grains are clear, show few inclusions, and have a dull or polished surface. Subrounded grains are the most abundant form in the fine sand class and are also common in the coarser grained fractions. Rounded and well rounded grains predominate in all the fractions coarser than one phi and they are also quite common in the one to two phi fractions. Sphericity is also related to grain size, with subrounded grains generally having a moderate sphericity and the coarser, well rounded grains having a high sphericity. The samples studied vary considerably in stratigraphic height and distance from probable source areas but all show a similar relationship between grain size and quartz roundness and sphericity. This indicates that roundness does not alter appreciably with distance of transport and is probably an imposed feature originating from the well rounded Larapinta Sandstones. Fine and very fine sand grains are largely protected from rounding processes by hydraulic pressures in an aqueous medium and they generally have lower fall velocities than equivalent rounded grains (Williams, 1966). Such small particles may owe their angularity in the present cycle to original angular and subangular grains in the source rocks (e.g. Larapinta and Pertaoorrta siltstones and fine sandstones), to chemical solution in the source area.
soils, or to breakage of quartz grains during weathering or transport (Moss, 1966).

Surface features of the disaggregated, subrounded and rounded grains show some systematic variation with geographic location. Most of the samples from the Pertnjara Group, west of Deep Well and north of Henbury, are characterized by polished or dull surfaces frequently coated with iron oxides. Frosted grain surfaces are rare even on the coarser, very well rounded grains. Sample BN6 is an exception and contains many frosted grains in the 0.5 to 1.5 phi fractions. This sample is from a white sandstone and several other white sandstones show similar increases in the number of frosted grains over their reddish brown counterparts. The differentiation of surface features between the two types of sandstones may reflect aeolian abrasion during transport and deposition, post-depositional chemical etching, or the effect of source area supply. Any of these alternatives could, and probably were, operative. Erosion of the clean, white, medium quartz of the Mereenie Sandstone may be responsible for many of the frosted grains in the Pertnjara Group and the white sandstone lenses in the latter may represent such reworked sands. Coarser grained quartz in the eastern and southern exposures of the Pertnjara Group and in the Finke Group is characterized by its well rounded form and the predominance of grains with frosted surfaces. In these areas frosted grains are common down to a grain size of 2.5 phi, eg. FG15, FJ11, FJ12, and GB30. The sandstones in these areas are
predominantly white and the origin of the frosted grains is probably similar to that described above. However, sedimentary structures, especially cross-stratification, in the eastern areas suggests that aeolian activity may have been responsible for frosting in the last cycle, as well as the derivation of frosted grains from the Mereenie Sandstone.

Thin section analysis of quartz roundness was conducted on a total of over 20,000 grains from all thin sections studied. The average visual roundness of 40 to 60 quartz grains in the 2 to 3 phi class, encountered during the point count modal analysis, was determined by comparison with the roundness scale of Powers (1953). Medium and mean roundness values were calculated for each sample (Appendix 6) using the rho scale of Folk (1955). Interpolated values were also calculated for the graphic and inclusive graphic standard deviation of roundness for each sample. These values are presented in columns four to seven in Table 4.2. Samples with no roundness values are either conglomerates or siltstones, where quartz grains are too scarce for reasonable determinations.

The variation in medium and mean grain roundness is only slight, with the vast majority of samples (83%) having a mean in the subrounded class (3.0 - 4.0\(^\circ\), Fig. 4.1). The average roundness of all samples is 3.34\(^\circ\). Less than 3 percent of the samples have a medium or mean roundness in the rounded class (4.0 - 5.0\(^\circ\), Fig. 4.2) and most of them come from clean, well sorted sandstones near the
base of the Pertnjara Group. These samples probably represent reworked lenses of Mereenie or Larapinta Sandstones. Samples with a medium or mean roundness in the subangular class (2.0 - 3.0) form the remaining 14 percent of the total. They are of two general types. Most of them represent very fine grained sediments from variable stratigraphic levels. Their sorting and grain size controls the grain shapes present in such deposits. Other samples are coarser grained, such as AJ187, BB969, BC976, DI52 and 55, and occur at stratigraphically higher levels. They represent samples from a more diverse source area where there is mixing of rounded polycyclic grains and more angular grains, derived directly from igneous and metamorphic terrains.

Roundness sorting, expressed in terms of roundness standard deviation, shows a random pattern when all samples are considered. Approximately 10 percent of the samples have very good roundness sorting and most of these are medium to fine grained sands, a few of which show features of aeolian deposition, eg. CY47, ET20, and GA43. Most of the sandstones have good (66%) or moderate (22%) roundness sorting while the remaining 2 percent show poor roundness sorting. Many of the latter are also bimodal and contain well rounded fine to medium grains together with subangular very fine to fine grains.
Sediments displaying anomalous rounding features are rare in the Pertnjara and Finke Groups. Grains with strongly curved re-entrant angles and peculiar protrudences are quite common in these sediments but in most cases, especially in the reddish brown sandstones, such features are clearly the result of, or have been accentuated by, quartz overgrowths (eg. Figs 4.3 & 4.4). Dismembered grains are very rare although fractured grains were seen in several thin sections. If the quartz overgrowths had not been clearly defined by iron oxide coatings these features, and similar features in the white sandstones, could have been attributed to solution of the quartz grains by weathering processes (Crook, 1969). However, even in the silty sandstones and siltstones lacking iron oxides such grains are very rare and this may indicate that, while such sediments were exposed subaerially, soil profiles were not developed and organic solvents were, at the most, minimal. This concept is supported by the lack of massive or disturbed laminae, representing soil profiles, and by the virtual absence of plant fossils.

Inclusions in quartz grains were also studied in some detail. The nature of the inclusions in the first 50 non-composite quartz grains encountered during the modal analysis of each sample were recorded in the following classes:- grains containing microlites or acicular mineral inclusions, clear grains with no inclusions, and grains containing few, common or
abundant vacuoles. The grains were studied under high power magnification and where possible acicular and microlitic inclusions were identified. Fine, hair-like needles of rutile are the most common type of crystalline inclusion and they are usually associated with sparse numbers of bubbles. Occasional quartz grains, especially in the MacDonnell Range samples, contain microlites of blue or brown tourmaline. Rare examples of grains containing microlites of zircon, epidote and apatite were noted. The quantity of these grains in any sample is variable but usually very low (about 2%). The range of percentage values of quartz with crystalline inclusions is given by the acicular index (BAI) in Table 4.2 (Appendix 6). There is little regional or stratigraphic variation in the frequency of quartz with acicular inclusions, except for the marginally higher average content in samples from the western MacDonnell Range.

The quartz free of inclusions and the quartz containing vacuoles were studied in an analogous manner to the roundness determination. The following class limits were used:

- clear 0.0 - 1.0
- few vacuoles 1.0 - 2.0
- common vacuoles 2.0 - 3.0
- abundant vacuoles 3.0 - 4.0

The median, mean, and graphic and inclusive graphic standard derivations were calculated for each sample from cumulative percentage frequency distributions.
All samples had a very uniform assemblage with means predominantly in the range 1.8 to 2.3 and a graphic standard deviation of 0.8 to 1.1. The uniformity of distribution points to the uniformity of detrital quartz types available in the source area and the lack of substantial contributions from any distinctive primary source rock. The assemblage is just what would be expected from a polycyclic sedimentary source lithology.

**Rock Fragments**

The presence of sedimentary rock fragments is a characteristic feature of the Pertnjara and Finke Group sediments. The abundance of lithic fragments ranges from 0 percent in the siltstones up to 96 percent in the Brewer Conglomerate, with most of the sandstones having between 10 and 30 percent. Sandstones from the MacDonnell Range area have a higher lithic content than the average while the white sandstone samples from south and east of Deep Well generally have a lower lithic content. Sandstones in the Finke Group are likewise low in lithic fragments and are characterized by their feldspar content. Some of the basal sandstones in the Pertnjara Group also contain few lithic fragments (eg. AK1, AK3, BP2) and are dominated by rounded quartz. These sandstones always directly overlie the Mereenie Sandstone and are considered to represent direct reworked sediments.
The rock fragment suite is predominated by siltstone fragments. Sandstone and chert fragments are usually present in lesser quantities while igneous and metamorphic grains are rare in most examples. Limestone fragments are also present in some samples especially from the MacDonnell Range area where they become increasingly abundant up the stratigraphic column.

Metamorphic Rock Fragments

The definition of metamorphic rock fragments used in this study differs from that of Folk (1968) in that the fine grained fragments have been classed as siltstone or shale unless definite evidence of oriented micas or other metamorphic features are present. The classification of all shale grains as metamorphic would seriously complicate source area determinations when a thick sedimentary sequence with numerous siltstones is known to underlie the formations being studied. The classification used in this study was supported by the relative scarcity of siltstone sedimentary fragments from the lower units of the Finke Group, which are known to have had a metamorphic and granitic source as well as a reduced supply of sedimentary detritus.

Metamorphic fragments generally form less than one percent of the framework grains and they are absent from many samples. The most common metamorphic fragments in these rocks are schistose and stretched metamorphic
quartz grains. These grains are generally subrounded and may represent first cycle or reworked metamorphic fragments. Slate fragments are either very rare or are insufficiently distinct to be recognized. Many of the composite quartz grains may have been derived from a metamorphic source but they are predominantly reworked and have been recorded separately.

In the Polly Conglomerate and the Undandita Member metamorphic and granitic fragments are more abundant (6-22%) with the latter usually predominating. Many of the fragments are subrounded and the sediments are characterized by their poor sorting and presence of angular detrital grains (Fig.4.5). Shale fragments are infrequent while schistose fragments are more abundant in the pebble and granule size range than in the sand size range. Many of the angular quartz grains and most of the feldspar grains are probably primary detritus from a metamorphic and granitic terrain.

Siltstone Rock Fragments

Soft rounded and occasionally contorted siltstone fragments form the most abundant lithic fragments present in most samples. The fragments are all fairly uniform in appearance consisting of occasional very small quartz and mica grains set in an indeterminable matrix of clay minerals. A few of the grains are slightly calcareous. In most samples compaction is minimal and the
original grain boundaries are clearly discernable. However, some samples show slight compaction, with the siltstone fragments being the only ones involved. In such cases the siltstone grains are compressed and contorted between neighbouring quartz grains but they can usually be recognised by partial grain boundaries or distinctive changes in matrix texture. The soft nature and degree of rounding of the siltstone rock fragments suggests that they are dominantly of sedimentary rather than metamorphic origin.

The distribution and abundance of siltstone fragments show some systematic pattern. With few exceptions siltstone concentrations in excess of 20 percent of the framework grains only occur along the northern margin of the Amadeus Basin. They are especially abundant in the Stokes Pass area in the western MacDonnell Range. The siltstone content of samples from north of the Gardiner and James Ranges is generally between 10 and 20 percent, but concentrations as low as 2 percent have been recorded from some of the white, well sorted, quartzose sandstones. South of these ranges, and especially in the area east of Deep Well, the quantity of siltstone fragments is predominantly less than 10 percent. The Finke Group is also generally low in siltstone fragments. There is a general decrease in the quantity of siltstone fragments from north to south and also from northwest to southeast across the Amadeus Basin.
The siltstone fragments may have been derived from any of the Pre-Cambrian to Ordovician silty formations underlying the Pernjara and Finke Groups. Some of them may also represent penecontemporaneously eroded siltstone beds of which very few remain in the sequence.

**Sandstone Rock Fragments**

Fragments of fine grained quartzose sandstones are also relatively common in most samples although they are usually less frequent than the siltstone fragments. Exceptions to this rule usually occur in the coarse grained samples especially in the Brewer and Polly Conglomerates. The variation in sandstone types is quite marked, although quartz and calcite cemented, very fine, quartz sandstones predominate. Occasional silty and micaceous sandstones were also noted. The sandstone fragments are generally subrounded to well rounded and they are harder than associated siltstone fragments. These samples are largely non-diagnostic of the formation from which they were derived and they could have come from any of the older, fine grained portions of the sedimentary sequence.

The distribution and abundance pattern of the sandstone fragments is generally similar to that of the siltstones but is less well defined. Samples from the Missionary and Brewer Plains areas generally have a sandstone rock fragment content of between 1 percent and 15 percent with an average of about 8 percent. South and
east of Deep Well the proportion of sandstone fragments
tails off to an average of about 3 percent, with the Polly
Conglomerate being an exception. The latter contains up
to 30 percent sandstone fragments most of which are in
the form of coarse grains or granules probably derived
from the underlying Stairway Sandstone.

Chert Rock Fragments

Reworked, detrital chert fragments are a per­
sistent minor feature of all the sediments studied. They
generally form less than 3 percent of the framework
grains except in one sample from the Brewer Conglomerate
which contains almost 30 percent detrital chert as
granules and pebbles. The quantity of chert fragments
shows a slight increase with stratigraphic height in
samples from the Hermannsburg Sandstone in the MacDonnell
Ranges (i.e. from 1 to 8 percent in section AJ). Over
the rest of the basin the occurrence of detrital chert
is sparse and random.

The chert fragments are usually subrounded and
consist of brownish or grey, finely crystalline quartz.
Rare chert grains grade into composite quartz with
some individual grains being larger than 20 microns.
Such grains were classified with composite quartz rather
than chert. The only inclusions, noted in a few of the
chert grains, were randomly oriented clay flakes. These
chert grains are similar in characteristics to chert
bands in the Cambrian and Pre-Cambrian limestones and were no doubt derived from these source rocks since chert is relatively rare in the Larapinta Sandstones.

Limestone Rock Fragments

Limestone rock fragments, occasionally showing signs of algal structures in the larger grains, have a sporadic but regular distribution in both the Pertnjara and Finke Groups. In the Pertnjara sediments limestone fragments occur preferentially in the Ljiltera Member and the Brewer Conglomerate and in the latter they comprise up to two thirds of the framework grains. In the Ljiltera Member the proportion of limestone rock fragments is less than 25 percent and their average occurrence is about 10 percent. Samples with a limestone rock fragment content of less than 5 percent are rare except in the MacDonnell Range area. This may be a depositional feature, or possibly samples with a low original limestone content had the limestone removed in solution by connate waters undersaturated in carbonate. In the Finke Group limestone rock fragments are largely confined to the Polly Conglomerate where concentrations of up to 20 percent may be reached.

The limestone fragments have features in common with both the Cambrian and Pre-Cambrian limestone formations but probably the bulk of the grains were derived from the Bitter Springs Formation.
Feldspar

Feldspar is a fairly consistent minor mineral in the Pertnjara Group sediments. Its abundance ranges from 0 to 11 percent with an average of only 1 to 2 percent (Table 4.3). Feldspar is more abundant in samples from the MacDonnell Ranges than elsewhere in the Pertnjara Group and here averages about 4 percent of the rock. It is also more abundant higher in the stratigraphic column frequently forming 6 to 8 percent of the sandstones of the Undandita Member (eg. BA970, BI972), and 2 to 3 percent in the Ljiltera Member (eg. DI50-55). The proportion of feldspar is also slightly higher in the Parke Siltstone (eg. AJ105, DB, DH) than in the overlying sandstones.

In the Polly Conglomerate and Langra Formation the proportion of feldspar is much higher than in the Pertnjara Group. The average content is about 13 percent while some samples from the Polly Conglomerate contain as much as 33 percent.

Potassium feldspars predominate in all the sediments studied and microcline is usually more abundant than orthoclase. Plagioclase is very rare in any of the samples studied and the only grains recorded were weathered, rounded, sodic plagioclase. The abundance and proportions of microcline and orthoclase were checked in some samples by staining with
sodium cobaltinitrite using the method of Bailey and Stevens (1960) with modified etching times and hydrofluoric acid concentration. Staining for plagioclase was attempted on a few samples but did not improve the recognition of this mineral.

Microcline is usually present as rounded to subrounded grains showing little sign of weathering (Figs 4.5 & 4.6). The subrounded grains are usually recognizable as broken rounded grains thus indicating a turbulent mode of transport. Orthoclase grains are usually rounded and frequently show signs of weathering with the development of white "sericite" concentrated along twin or cleavage planes. Most of the sediments are fairly impermeable and the degree of feldspar weathering is not considered to be due to the present weathering cycle. This concept is supported by the occurrence of very rare, unweathered, clear, feldspar overgrowths on partially weathered, rounded, detrital orthoclase (Fig. 4.7).

The rounded feldspar grains are usually of approximately the same size as the associated rounded quartz grains (Fig. 4.6). This indicates an anomalous rounding history since feldspar is less durable to abrasion than quartz. This is probably attributable to the variation in feldspar content and quartz roundness in the source area sediments. The quartz is probably derived largely from the Larapinta sandstones while the
feldspar probably represents smaller contributions from the Cambrian sediments, eg. the Arumbera Sandstone. The proportion of feldspar in the siltstones is probably due to abrasion of most feldspar grains to the silt size range.

The roundness of the feldspars and the abundance of microcline over orthoclase points to the probable reworked sedimentary source of these grains in the Pertnjara Group. The homogeneity of weathering of the feldspars also points to a single predominant source lithology. In the Polly Conglomerate and Langra Formation of the Finke Group, microcline and orthoclase are almost equal in abundance, are fresh and generally subrounded. They are frequently intergrown with quartz in granitic fragments and they represent primary derivation from a granitic and/or gneissic source area.

According to Folk (1968) well rounded, fresh feldspar when present in moderate quantities is an excellent indicator of an arid climate. Thus the feldspar present in the Pertnjara Group could represent feldspar derived from a predominantly sedimentary source in an arid or semi-arid climate. However, it should be pointed out that terrestrial vegetation was probably scarce in upland areas during the Devonian and consequently the organic acids responsible for much of the feldspar alteration in more temperate and humid climates would also be scarce. Thus the climatic intensity proposed by Folk (1968) may not be strictly correct when
applied to ancient terrestrial environments with rare vegetation and rapid surface run-off and this may account for the lack of deep weathering and soil profiles in such areas. Thus the climatic environment of deposition of the Upper Devonian sediments in the Amadeus Basin may have been more pluvial than is indicated by the weathering of the feldspar.

Further climatic implications can be based on the degree of feldspar weathering. In areas of low relief, high non-seasonal rainfall and subtropical temperatures containing abundant plant cover orthoclase is readily weathered (Todd, 1968). The presence of fresh, unweathered and rounded orthoclase and microcline in the Pertnjara and Finke sediments indicate that the combination of these conditions did not exist in the source or deposition areas.

Micas

Detrital micas, especially muscovite and biotite are relatively common in the siltstone members of the Pertnjara and Finke Groups. The relative abundance of muscovite and biotite is given in Table 4.11. Muscovite is dominant in most of the samples and forms between 0 and 75 percent of the accessory mineral assemblage. Of the framework grains the proportion of mica rarely exceeds 2 percent. Biotite is only prominent in the Polly conglomerate and Horseshoe Bend Shale of the Finke Group and in the Undandita Member of the Pertnjara Group.
The mica flakes usually show little sign of weathering although a few biotite flakes were leached. A few mica flakes exhibit post depositional weathering with the development of frayed ends expanding to fill voids while the remainder of the flake, embedded between framework grains or matrix, is virtually unweathered.

In the siltstones the mica is aligned parallel with the bedding while in the sandstones micaceous minerals are generally too infrequent and often too small for imbrication analysis using the method of Diessel (1966).

When the abundance of biotite is equal to or exceeds that of muscovite Folk (1968) suggested that either the sediment has received volcanic detritus (improbable in this case) or the erosion rate has exceeded the rate of weathering in the source area. The abundance of biotite in the Polly and Brewer Conglomerates and the Horseshoe Bend Shale, and the occurrence of thin (1 cm) beds of almost pure biotite in the basal Idracowra Sandstone south of Black Hill Range indicates that weathering must have been minimal during these periods. The association of these sediments with desiccation features lends support to the hypothesis of an arid or semi-arid climate during deposition of these beds.
Opaque and Accessory Minerals

Opaque and accessory minerals form a persistent minor constituent in the framework grains (each approximately one percent or less). Accessory mineral contents are slightly more variable than opaque mineral occurrences and are slightly higher than average along the northern margin of Missionary Plain. High proportions of accessory minerals in a few siltstones are due to the concentration of micas in these samples. The opaque mineral suite is dominated by ilmenite and leucoxene while the accessory minerals are largely of the more stable varieties such as zircon, tourmaline, rutile, muscovite. Individual variation of these minerals will be discussed in more detail later.

Authigenic Quartz

Quartz overgrowths are common in most of the moderately to well sorted fine to coarse sandstones (Figs 4.2 & 4.4). They are generally rare in very fine sandstones and in poorly sorted lithic sandstones. The overgrowths may form well developed crystal faces (Fig.4.8) or they may just fill the voids. Pressure solution features are absent from these rocks and the silica required for quartz overgrowths must have entered the system in solution. Even samples from 2500 m below the surface in Tyler No. 1 Well show no compaction of quartz grains and only minor distortion of lithic fragments.
Quartz overgrowths are generally easy to discern in the Pertnjara and Finke sediments because the quartz grains are generally coated with a thin layer of hematite which gives the reddish brown colour to the sediments (Fig.4.6). In white sandstones quartz overgrowths may be harder to recognize but there is frequently a thin layer of pale clay mineral or vacuoles between the detrital grain and the overgrowth. In almost all cases the quartz overgrowths formed by direct precipitation in originally porous sandstones. They were not seen to be replacing any other mineral and do not occur in sediments containing abundant carbonate cement thus suggesting that deposition of the quartz occurred after deposition of the carbonate. It was probably formed by connate waters migrating through porous zones in the sandstones. Non-silica-cemented porous sandstones probably owe their porosity to their lensoid nature and the lack of continuity of permeable zones.

The period when quartz overgrowths formed cannot be clearly defined except that it occurred after the deposition of the iron coatings on the grains. The general absence of fractured grains provides no time control with respect to tectonic events. The occurrence of shattered and healed quartz grains with overgrowths at Gosses Bluff provides some evidence about the continued action of authigenic quartz deposition. Quartz overgrowths had formed prior to the crypto-explosion, forming Gosses Bluff in the Lower Cretaceous (about 125 million years ago, D. Milton, pers. comm. 1969) and the deposition continued
after this event.

**Carbonate Cement**

The occurrence of carbonate cement, dominantly calcite, may be either primary or secondary. It is a primary feature in many of the siltstones in which it consists of a mosaic of very fine grained calcite and dolomite crystals. Such calcareous siltstones account for all the samples with over 25 percent calcite but not all siltstones are so calcareous.

Calcite may also occur as a finely or coarsely crystalline cement in various sandstones. When present it generally forms between 5 and 25 percent of the rock and, except on samples from the northern Missionary Plain, it rarely occurs in lesser amounts. This localization of occurrence strongly favours an authigenic depositional process probably due to the migration of carbonate rich connate waters through originally more porous beds. This concept is supported by the replacement of siltstone and chert fragments and the detrital matrix by crystalline calcite. However, in most cases there is little alteration or replacement of the detrital minerals and the calcite appears to occupy primary voids (Fig.4.9). The lack of alteration and good state of preservation in such samples suggest that cementation with calcite took place a relatively short time after deposition. In most samples it formed after the formation of ferric oxide grain
coatings. There is generally no association of authigenic quartz and calcite in samples with abundant calcite while in samples only partially cemented with calcite quartz overgrowths are quite frequent. This indicates the sequence of authigenic events - namely deposition of ferric oxide grain coatings, deposition of authigenic calcite, and finally deposition of authigenic silica.

In the Pertnjara Group carbonate cement is largely confined to samples in or just above the Parke Siltstone and samples in the Ljiltera Member and Brewer Conglomerate. In the Harajica Sandstone Member on Dare Plain several sandstone lenses are irregularly carbonate cemented and show lustre mottling which is almost certainly an early diagenetic feature. Carbonate cement becomes increasingly abundant higher in the Ljiltera Member and is virtually the only matrix component in most of the massive Brewer Conglomerate samples. It may be derived from in situ solution of carbonate pebbles (little evidence) or it may be brought into the area of deposition in solution. It was undoubtedly derived from the Cambrian and Pre-Cambrian limestones of the Amadeus Basin succession as were the associated carbonate clasts.

In the Finke Group carbonate is locally abundant in the Horseshoe Bend Shale as micritic calcite and dolomite. It also occurs as a major cement in subsurface samples of the Polly Conglomerate (FJ16-19). In the latter it was probably derived from the Bitter Springs
Formation as are many of the included clasts.

**Iron and Silica Matrices**

Iron oxides or silica form a significant proportion of the matrix of a few samples. These samples are all surface samples which have suffered the effects of deep Tertiary weathering with the consequent partial or complete replacement of the matrix and most rock fragments by laterite or silcrete.

**Detrital Matrix**

The term detrital matrix is used to include all material, including authigenic clays, finer than 35 microns (4.8 phi). This material includes fine detrital quartz grains, accessory and opaque minerals, especially micaceous hash, set in a matrix of various indeterminate clay minerals. The term excludes carbonate cement and secondary iron and silica concentrations. Clay minerals in the sandstone matrices are dominated by kaolinite in the white sandstones and by illite, chlorite and mixed layer clays in the brownish sandstones. Kaolinite and illite predominate in the authigenic clay cutans.

The proportion of detrital matrix in the samples is very variable but the sandstones have an average content of 15 to 20 percent while siltstones and silty sandstones contain an appreciably higher proportion. There
does not appear to be any overall significant pattern in the distribution of the detrital matrix content.

Voids

With the exception of sample BQ5, the term void is limited to recognizable primary holes in the rock thus excluding plucked areas. Voids, termed as such, were recognized by the grain packing features in non clay cemented samples, and by the presence of clay cutans partially filling and lining the voids in clay cemented samples. The latter is typical of the vast majority of sediments from both the Pertnjara and Finke Groups and thus the counts obtained are probably fairly representative. The void content, with few exceptions, is characteristically low (less than 2.5 percent) and this accounts for the lack of porosity and permeability in the majority of these sediments. More voids occur in moderate to well sorted white quartz sandstones in both formations and these are the only potentially useful aquifers.
Texture

The Pernjara and Finke sediments are characterized by poor sorting, due to bimodality, and immature to submature textures with the majority of samples showing source area controlled textural inversion.

The sorting is usually poor with a population of silt and clay grains usually forming between 5 and 20 percent of the sample (Fig. 4.10). A few very silty sandstones have silt contents up to 40 percent while, of course, the silty mode dominates in the siltstones (Fig. 4.11). A few samples are trimodal with the coarser traction deposited grains dominantly consisting of well rounded quartz (Fig. 4.12). Unimodal saltation populations are infrequent in the sandstones (eg. Fig. 4.13) and they are usually poorly sorted. They are largely confined to localized aeolian dune deposits and their poor sorting is controlled by the poorly sorted nature of the fluviatile deposits from which they were derived. Unimodal sediments are more common in the siltstone units where suspension deposition predominates. However, such unimodal deposits are only significant in the Dare Siltstone Member where they are associated with carbonates and evaporites. Most of the siltstones in the other units contain some coarser particles ranging up into the fine sand grade and this is especially the case in siltstones incorporated in the upper part of the fluviatile cyclic sequences.
The quantity of silt and clay in these sediments indicates that they should be classed as immature or sub-mature according to the quantity of detrital matrix and sorting values. Thus the energy expended at the site of deposition is insufficient to allow sorting processes to operate. However, the sediments are also characterized by their high roundness and sphericity values for the quartz grains (Figs 4.2 & 4.6). Even the fine and very fine sandstones contain a high proportion of rounded and subrounded grains. There is quite a wide variation in grain roundness within a single grain size fraction although angular and very angular grains are rare throughout the lower part of the sequence and only became significant in the Undandita Member. The presence of rounded grains in a poorly sorted sediment is indicative of textural inversion controlled by availability in the source area (Folk, 1968). In this case the well sorted marine and aeolian sandstones of the Larapinta Group and Mereenie Sandstone provide abundant, well rounded grains. Many of the Cambrian sediments also contain moderately rounded grains and thus the abundance of well rounded grains in the Pertnjara and Finke sediments is hardly surprising when the thickness of the earlier sedimentary sequence is considered. The Finke Group also had a metamorphic and granitic source and this is reflected in the higher proportion of primary detrital subangular grains in the Polly Conglomerate and Langra Formation.
The predominantly poorly sorted, immature sediments of the Pertnjara and Finke Groups are typical of sediment accumulation in loci, such as flood plains, alluvial fans, river channels, and lagoonal or shallow neritic environments where current action is weak or sediment deposition is rapid (Folk, 1968). Thus the sediments are not reworked and are not subject to post-depositional mechanical sorting or rounding.

Paine and Meyerhoff (1968) listed a set of criteria for recognizing fluvial sediments in thin section. Many of these criteria can also be applied to the Pertnjara and Finke Group sediments eg. immature textures, poor sorting of framework grains, bimodal character, moderately abundant flat grain alignments, and absence of unabraded marine fossils. The main differences between these criteria and the sediments studied are:— the high grain roundness and sphericity and the glauconite content of these sediments, both of which are controlled by source area provenance; the absence of plant material, and the scarcity of fresh water fossils which may be attributed to the lack of adaption of these organisms in the Devonian to an environment subject to periodic desiccation; and the presence of some authigenic calcite which probably reflects the content of derived limestone clasts and the abundance of limestone in the source area.
Iron Oxide Staining

Iron oxide staining of grains is an important feature in the Pertnjara sediments but is rare in the sediments of the Finke Group. It is present in, and accounts for the colour of, all the reddish brown sandstones of the Hermannsburg Sandstone while it is virtually absent from white sandstones such as the Oooraminna Sandstone Member and the Owen Springs Sandstone Member. It is most abundant on the northern margin of the Missionary Plain and becomes decreasingly abundant in similar stratigraphic horizons to the south and east. Iron oxide staining is also less abundant in the Harajica Sandstone Member and in the Brewer Conglomerate where greenish lithic sandstones are quite common.

Iron oxides, dominantly hematite, form a thin coating around almost all detrital grains in the reddish brown sandstones (Fig. 4.6). Its presence is most conspicuous on quartz and feldspar grains and in disaggregated samples these grains have an orange brown appearance. The presence of the iron oxide coating is a useful aid in recognizing detrital grain outlines, including those of siltstone fragments. Iron oxides were only seen to replace minerals and fill voids when the sediments had been exposed to Tertiary weathering and lateritization, although some hematite "halos" near biotite grains may represent early diagenetic alteration products.
The deposition of iron oxide coatings is one of the problems associated with the genesis of red bed sequences. The original mineral composition of the iron oxide coating cannot be determined since most forms would alter to hematite during diagenesis. Thus the iron could have been deposited as a colloid or as hydrated ferrous or ferric oxide although the presence of detrital ferric heavy minerals suggests that the environment must have been oxidizing and not too acid (Miller & Folk, 1955). The presence of ilmenite and magnetite in many of the sediments with iron oxide coatings indicates that these minerals were not necessarily the main source of iron oxides in the Pernjara Group. Likewise the scarcity of ferromagnesian minerals is probably due to source area control. If these environmental conclusions are correct they would also favour the deposition of ferric oxide grain coatings rather than ferrous or colloidal forms. On aging hematite may form from a variety of amorphous hydrated ferric oxides (T. Walker, 1967a).

The period of deposition of the iron oxide coating is also problematical. Iron coated grains may have been derived from pre-existing red beds eg. Arumbera Sandstone; it could form in the source area soils prior to erosion and deposition; it could form during or prior to the deposition of the beds in the depositional area; or it could form after deposition. In all probability each of these factors may have contributed to the iron oxide coating. Deposition of at least part of the iron oxide
coating prior to final deposition is supported by the presence of continuous iron oxide coating around grains which show obvious primary packing features (Fig. 4.14). If all iron oxide had been deposited after the deposition of the grains one would not expect to find a continuous uniform coating at the point of contact (Miller & Folk, 1955). Another feature is that small grains often have a thicker iron oxide coating than larger grains and this can be attributed to the smaller degree of abrasion which small grains undergo during transport in a fluviatile medium. Whether the predepositional iron oxide coating formed in the source area or the depositional area cannot be determined. It may have formed in the parent soil where leaching is minimal and oxidizing conditions prevail. Brown iron stained detritus formed during weathering is present in many modern streams (Van Houten, 1964). However, it could have formed in earlier fluvial deposits which had been subjected to chemical weathering (Baker, 1962) prior to reworking before final deposition. The distance that iron oxide coated grains from areas of red soil formation may be transported is not very great and they are usually masked in areas of deposition by non red detritus (T. Walker, 1967b). This favours the suggestion that iron oxide coatings may have formed in the depositional area and were locally reworked prior to final deposition. This hypothesis is also supported by the fact that some beds show thicker iron oxide coatings in non grain-contact areas which indicates that the deposition may have, and probably did, continue after the final deposition of the bed.
In all events the deposition of iron oxide grain coating must have been a very early diagenetic feature since all other cements and overgrowths do not interfere with the pattern of iron oxide distribution. T. Walker (1967a) noted that the process of hematite formation from hydrated ferric oxides took place in an arid climate in a time interval of about one to ten million years.

The relationship of iron content to red and green coloration in sediments was studied briefly by Friend (1966). He noted that the content of ferrous oxide remained essentially constant in both red and green sediments while ferric oxide occurred preferentially in the red sediments. The results of similar chemical analyses of closely associated red and green sediments from the Dare Siltstone Member show very similar results which are tabulated below:

<table>
<thead>
<tr>
<th>Colour</th>
<th>FeO Total iron as Fe₂O₃</th>
<th>Fe₂O₃</th>
<th>Fe₂O₃ x 100 Total iron</th>
</tr>
</thead>
<tbody>
<tr>
<td>DH88 red-brown</td>
<td>1.73</td>
<td>4.95</td>
<td>3.03</td>
</tr>
<tr>
<td>DH87 green</td>
<td>1.22</td>
<td>1.44</td>
<td>0.08</td>
</tr>
<tr>
<td>DH74 red-brown</td>
<td>1.40</td>
<td>3.45</td>
<td>1.89</td>
</tr>
<tr>
<td>DH73 green</td>
<td>1.58</td>
<td>2.15</td>
<td>0.39</td>
</tr>
</tbody>
</table>

The differentiation into red and green horizons in the Dare Siltstone Member is probably due to variations in the oxidation-reduction potential shortly after deposition. The environment at the time of deposition of the green siltstones was an evaporitic lake and such sediments are usually well cemented with calcite and dolomite. This
would cause such sediments to be impermeable and relatively non-responsive to changes in the oxidation-reduction potential of adjacent zones whereas less well cemented sediments could be oxidized under similar conditions. The process of oxidation had little effect on the clay mineral assemblage and this suggests that the red beds were originally richer in iron oxides than the green beds, and that redness developed in situ soon after deposition as a response to variations in the water table level (Friend, 1966).

The presence of red beds cannot be taken by itself to indicate climatic conditions at the time of deposition (T. Walker, 1967a). They can probably form under a number of climatic conditions and their formation depends largely on the availability of iron minerals and a relatively high Eh and pH in the depositional area. Such conditions are common in arid and semiarid climates where there is little vegetation and a low water table. The lack of evidence of soils or terrestrial vegetation during the period of deposition of the Pertnjara and Finke Groups may lend support to the hypothesis of a rigorous climate, since most Devonian vegetation would not have been adapted to withstand periods of desiccation. The presence of evaporites in the Parke Siltstone and Horseshoe Bend Shale also point to a rigorous climate subjected to desiccation. There is, however, no evidence about the degree of aridity. The climate may have varied from extremely arid with rare flash floods to a more moderate savanah type climate with marked
wet and dry seasons.

Semi-arid or arid climates have been postulated for similar red bed sequences in the Baja California and the late Paleozoic red beds of Colorado (T. Walker, 1967a). In both cases climatic implication is derived from associated evaporites rather than the presence of red beds.
Variability of Mineral Suites

The variation of mineral proportion with grain size was studied from four widely spaced localities from within the basin (AF, DI, CY & FG). Most of the minerals show an independent variation with grain size while some of them show a random distribution e.g. feldspar, metamorphic rock fragments, microlitic quartz, quartz overgrowths, calcite and voids (Appendix 11). Minerals such as straight extinction simple quartz, undulose extinction simple quartz, semicomposite quartz, opaque and accessory minerals, and detrital matrix, all decrease in abundance with increase in grain size. Composite quartz, chert, siltstone, fine sandstone and limestone all increase in abundance with increasing grain size. Thus the finer sandstones tend to be richer in quartz and matrix while the coarser sandstones tend to be lithic.

Appropriate correction factors can be determined from the slopes of the regression lines for each individual mineral content versus grain size (Appendix 11). These correction factors were then used to recalculate the modal composition to a predetermined mean grain size - namely 2.5 phi. A list of the corrected modal analyses is given in Table 4.3.

Tables 4.4 and 4.5 present the modal analysis data in the form of triangular diagram, end members for the original analysis and for the analysis corrected to 2.5 phi respectively. The parameters presented include the Q, L, M plot of Crook (1960), stability plots for
both total framework grains and for quartz types after Blatt (1967), and the relevant triangles proposed by Folk (1968) including an additional triangle for sandstone, siltstone, chert values which are the dominant forms present in this study. The tables also include the classification of the sediments according to Folk (1968). Subdivision of the main classification using the chert, limestone, sandstone-siltstone rock fragment daughter triangle of Folk (1968) indicates that 2.6 percent of the samples may be termed calcilithites, 1.1 percent are chert arenites and 96.3 percent are sandstone-shale arenites. Using the daughter triangle sandstone, siltstone, chert, 3.3 percent of the samples are chert arenites, 12.1 percent are sandstone arenites and 84.6 percent are siltstone arenites. Two general trends emerge from the rock fragment triangles. There is an increase in sandstone and chert fragments (and also limestone fragments on the northern margin) with respect to siltstone fragments with increasing stratigraphic height, and with increasing distance of transport. The first trend is provenance controlled while the second trend is controlled by the rapid abrasion of softer siltstone fragments.

Most of the sample localities at approximately equal distances away from the northern margin of the Amadeus Basin show rather similar distributions of points on the triangular diagrams. Thus the samples in the A localities are more similar to each other than to members of the other localities, and so on for B, C, D, E, F and G localities.
This allows generalisation to be made after considering a limited number of localities which show significant trends.

Results from Glen Helen Gorge (section AJ), the section studied in most detail, show analogous patterns for both uncorrected and corrected modal analysis data although the boundaries around the various groups are not necessarily coincident (Figs 4.15a, b). This is also the case with other localities and hence all results are discussed in relation to corrected values only.

The QLM plot (Fig.4.15b) shows some interesting variation in mineralogy with respect to stratigraphic height. The sandstones of the Harajica Sandstone Member can be divided into lower (AJ2-48) and upper (AJ53-102) groups. The lower group is more quartz rich than the upper group and it also has a fairly constant ratio of labile fragments to matrix with varying quartz content. The upper group is generally more labile rich and contains a fairly constant matrix content with varying quartz to labile ratios. The overlying Dare Siltstone Member plots near the matrix pole as would be expected. An interesting feature to note with the plot of Hermannsburg Sandstone samples is that it shows a similar general trend of increasing lithic content with increasing stratigraphic height as does the Harajica Sandstone Member. However, the initial Hermannsburg Sandstone samples are not as quartz rich as those of the Harajica Sandstone Member and the Hermannsburg Sandstone generally
contains a higher proportion of detrital matrix. Samples from the Brewer Conglomerate are characterized by their high content of lithic fragments.

Analogous results are indicated on QFR diagrams (Fig. 4.16, after Folk, 1968) although the distinction between the Harajica Sandstone Member and the Hermannsburg Sandstone is not as pronounced since the main difference is in the matrix content. The figure clearly shows the two cycles of increasing lithic content with stratigraphic height.

Results shown on the overall stability diagram (Fig. 4.17, after Blatt, 1967) are also similar to the same two cycles of increasing lithic content emerging. This trend towards an increased content of unstable framework grains in the form of lithic fragments is the opposite sequence to that normally expected from the degradation and weathering of a source area (Blatt, 1967; Gostin, 1968). It indicates that the erosion of the source area is exceeding the rate of weathering and the ratio of erosion to weathering increases with time during the deposition of both the Harajica Sandstone Member and Hermannsburg Sandstone and culminates with extremely rapid and extensive erosion of the source area during the deposition of the Brewer Conglomerate.

The plot of points on a sedimentary rock fragment diagram with the end members sandstone, siltstone, chert
(modified from Folk, 1968) also shows a double cycle in the Harajica Sandstone Member - Hermannsburg Sandstone sequence (Fig. 4.18). Both cycles start off with a predominance of siltstone fragments and become less silty upwards, with a marked increase in sandstone rock fragments higher in the Harajica Sandstone Member and an increase in both sandstone and chert fragments higher in the Hermannsburg Sandstone and in the Brewer Conglomerate.

The only departure from this double cyclic trend is shown by the stability plot of quartz types (after Blatt, 1967). In Figure 4.19 all the sandstone units have quartz stability ratios within a very restricted field and the ratios are not controlled by stratigraphic height. This indicates that all the quartz grains are derived from an essentially constant and uniform source - namely they are being reworked from mature, earlier Paleozoic sandstones. Quartz in these older sandstones is all probably of polycyclic origin and this accounts for the high stability of the quartz types present since in prolonged abrasion the less stable quartz types are preferentially eliminated (Folk, 1968). Samples from the Brewer Conglomerate (Fig. 4.19) have a higher content of unstable quartz types than the underlying units and this is due to the uncovering and erosion of a metamorphic and granitic terrain during the deposition of this unit. The introduction of the first cycle quartz reduces the overall quartz stability ratio although it is still influenced by stable polycyclic quartz grains.
A typical section taken down the paleoslope (see Chapter 3) from near the centre of the Missionary and Brewer Plains includes the following sample localities in order of increasing distance from the northern margin: AM, BO, BP, CY, ER, ET. Results obtained by plotting triangular diagrams for these localities show the same general trends as similar paleoslope controlled sets of localities to the east or west and, therefore, the trends shown by this set of localities can be taken as characteristic of the areal variation within the Pertnjara Group.

Ternary diagrams can be used to illustrate generalized compositional changes in a system as an alternative to numerous bar graphs for each individual mineral. The QLM diagram (Fig. 4.20) shows the general trend of increasing quartz content with distance of transport. The samples from localities AM, BO, BP are relatively rich in labile and matrix components although the AM samples are slightly better sorted and contain less matrix material. With increasing distance the labile components decrease steadily while the matrix content remains approximately constant. The latter is indicative of the generally similar, poor level of sorting characterizing the Pertnjara sediments while the decrease in labile components can be attributed to the rapid abrasion of sedimentary rock fragments during transportation.

The sandstone-siltstone-chert diagram (Fig. 4.21) gives a more detailed account of the relative rate of wear
of the sedimentary rock fragments. With the exception of one sample from both localities BP and ET which only contain recognizable chert fragments, there is a general trend from an original siltstone rich rock fragment assemblage to one dominated by fine sandstone and chert. This indicates that abrasion and weathering during transport affect the soft siltstone fragments more rapidly than the other types. The sandstone fragments are all very fine grained and well cemented with calcite or quartz. This accounts for their greater stability than the siltstone, and quartz cemented varieties are the only detrital sandstone types present in locations ER and ET. The last two locations also show a slight trend towards higher chert contents and probably, if further localities were present in a down current direction, chert fragments would begin to dominate over sandstone.

QFR diagrams (Fig.4.22) are poor indicators of mineralogical variation in the Pertnjara Group because of the general scarcity of feldspar. The diagram merely confirms the decreasing rock fragment content with increasing distance of transport.

The stability diagram for the framework grains (Fig.4.23) illustrates two important features. Firstly there is very little variation in the quantity of resistant minerals with respect to distance of transport and they remain consistently low throughout the Pertnjara Group. Labile sediments are composed predominantly of
rock fragments derived from the earlier sediments and quartz grains reworked from these earlier sediments. Secondly, the trend of decreasing rock fragment content with distance of transport is shown again.

The quartz stability diagram (Fig.4.24) shows an unusual feature. The quantity of unstable and resistant quartz types increases with distance of transport which is contrary to the theoretical and practical aspects of quartz stability (Folk, 1968; Blatt, 1967). However, the increases in the less stable quartz varieties can be attributed to mixing of sands from two or more source areas. Thus, the southern margin and north eastern margin probably contributed small quantities of primary quartz grains to the predominantly reworked sedimentary quartz types derived from the Amadeus Basin sequence.

Sediments of the Finke Group (F localities) show several trends that differ significantly from those of the Pertnjara Group. The QLM diagram (Fig.4.25) shows that the Finke sediments have a wide range of framework composition and matrix contents even within one formation. The Polly Conglomerate and its equivalent subsurface units generally have a relatively low matrix content but a wide variation in the quartz to labile ratio. Some of the basal sediments of the formation are very quartz rich and represent reworked earlier sandstones such as the Stairway Sandstone (e.g. FG2, FJ21). Above these basal sandy units the Polly Conglomerate contains abundant lithic fragments and the quartz
to labile ratio is generally less than 1.0. The Langra Formation shows a large variation in composition and this is due to its very varied lithologies i.e. interbedded conglomerate, sandstone and siltstone. However, most of the sandstone units in the Langra Formation are confined to the same field as the sandstones of the Idracowra Formation and this field corresponds to the QLM field occupied by most of the Pertnjara sediments. The Horseshoe Bend Shale is, of course, rich in matrix and it is consistently low in labile fragments. Interbedded fine sandstones are quartz rich and they produce a very elongate QLM field for the Horseshoe Bend Shale.

Stability plots of the framework grains (Fig. 4.26) show a fairly varied range in stability with a general increase in stability with increase in stratigraphic height through the group. The Polly Conglomerate has a range from almost pure quartz sandstone, representing reworked Stairway or Mereenie Sandstone, to very unstable lithic sandstones forming the bulk of the formation. The Langra Formation shows a somewhat similar distribution but all samples contain more than 50 percent of very resistant grains. The Horseshoe Bend Shale is characterized by a variable proportion of resistant and very resistant grains with the latter always being predominant. The Idracowra Sandstone contains small sub-equal quantities of resistant and unstable grains in a sandstone generally composed of at least 70 percent of very resistant framework grains.
Quartz stability diagrams (Fig. 4.27) show little variation from those of the two preceding quartz stability diagrams except that the stratigraphic order is reversed. In the Finke Group the Polly Conglomerate contains the greatest proportion of unstable quartz while the Langra Formation and Horseshoe Bend Shale contain progressively more stable quartz suites. The Idracowra Sandstone does not fit this trend and contains a higher proportion of resistant quartz. However, it still falls within the field of the Langra Formation and Polly Conglomerate although it is also in the field occupied by the Hermannsburg Sandstone in the Camel Flat Syncline area.

The QFR diagram (Fig. 4.28) shows an increase in the quantity of quartz with increasing stratigraphic height which is the opposite to the trend at Glen Helen. This sequence, with the most labile and feldspathic sediments in the Polly Conglomerate, indicates local source area control for the Finke Group sediments which is in agreement with the paleocurrent trends for the lower part of the Group (see Chapter 3). The Idracowra Sandstone falls in the same field but is slightly more feldspathic than the Hermannsburg Sandstone and this may indicate mixing of sediments from two source areas - one to the south and one to the north-west.

The sandstone-siltstone-chert diagram (Fig. 4.29) shows the very variable quantities of sedimentary rock fragments present in the Finke Group with most of the
sediments being dominated by siltstone and fine sandstone fragments. The variability of rock fragment proportions is greater than in the Hermannsburg Sandstone and once again the Idracowra sediments are most closely related to the Hermannsburg Sandstone.
4.2 PETROLOGY OF THE FINE GRAINED ROCKS

Fine grained sediments were studied from the siltstone and sandstone formations in the Pertnjara and Finke Groups in order to determine variations in the clay mineral suites over the area studied.

A few thin sections were cut of the Parke Siltstone and Horseshoe Bend Shale from their type localities. The sections typically contain very rare accessory minerals, rare grains of silt sized quartz, and moderately abundant fine mica flakes set in a matrix of very fine grained detrital and/or carbonate minerals (Fig.4.11). In general they show very little petrographic variability except in the quantity of carbonate.

A total of forty samples were studied by X-ray diffraction. Most of the samples were collected from surface outcrop and were as fresh as possible. Four samples (BC, BI, BR45 and BR53) came from core or cuttings from shallow stratigraphic drilling (B.M.R.) or seismic line shot points respectively. A further eight samples were obtained from exploratory oil well cores and cuttings - namely Tyler No. 1 (BJ), Palm Valley No. 1 (CM694-5), Mt Charlotte No. 1 (FK) and McDills No. 1 (FJ).
Analytical Procedures

The analytical procedures used in the clay mineral analyses are generally standard, although tests were conducted to determine the effect of iron coatings in the sample, and the best and most reproducible method of glycolation.

The use of clay mineral standards, extracted from the samples to be analysed, for quantitative X-ray diffraction analysis (Gibbs, 1967) was considered but precluded by the lack of suitable equipment.

Errors in clay mineral determination caused by different mounting techniques were studied by Gibbs (1965). Of the suitable techniques available suction on ceramic disks was utilized throughout this study, and with a scanning speed of $\frac{1}{2} \theta$ per minute Gibbs found this resulted in a rerun error of $\pm 6\%$ (95% confidence) for peak area above the background.

1) Iron Coatings: The effect of iron mineral coatings in argillaceous sediments was considered by Gibbs (1967) to produce considerable inaccuracies caused by incomplete mineral separation, the higher absorption coefficient of iron coated grains, and the higher background radiation. A series of six samples (AJ105, CM110, CS15, FG16, FJ9, FJ15B) were studied in duplicate to determine the effect of iron coatings on the Pertnjara and Finke sediments.
In one set the minus two micron fraction was separated and mounted with preferred orientation on ceramic disks. Iron was removed from the other set by the sodium bicarbonate buffered sodium dithionate - sodium citrate method of Mehra and Jackson (1960). About 4 gms of sample, 5 mls 1 M NaHCO₃ and 40 mls 0.3 M Na-citrate were placed in a centrifuge tube and warmed to 80°C in a water bath. One gram of solid Na₂S₂O₄ was added and the mixture digested for 15 minutes with stirring. Then 10 mls of saturated NaCl and 10 mls of acetone were added to promote floculation prior to separating and washing the clay with warm Na-citrate and water, and mounting on ceramic disks. Mehra and Jackson (1960) stated that the treatment has almost no destructive effect on iron silicate clay minerals, and Gibbs (1967) substantiated this.

Both sets of mounted clay minerals were X-rayed under identical conditions with CuKα radiation at 40 KV and 25 mA and an iron discriminator was used throughout. Although the background radiation was slightly higher in the untreated set of samples the clay mineral ratios remained essentially constant. This is partly attributable to the use of the discriminator, but may also indicate that the iron minerals in the samples studied are largely amorphous to very finely crystalline. Because of the similarity of results between the treated and untreated samples it was decided that the removal of iron from further samples was not warranted.
Glycolation: A study was conducted to determine the quickest and most reproducible method for glycolating samples mounted on ceramic disks. A series of tests was run on a representative sample, DH54, which contained a moderate amount (30%) of smectite.

A single sample was crushed, disaggregated and the minus two micron fraction separated for analysis. Four samples were mounted on ceramic disks by the method of Kinter and Diamond (1956) and placed in a dehydrant desiccator for 48 hrs to dry. X-ray diffractograms of these samples showed very little intersample variation of clay mineral ratios. A further sample was prepared in a 50% ethylene glycol aqueous solution prior to mounting on a ceramic disk and drying in a dehydrant desiccator in an ethylene glycol atmosphere for 48 hrs. On glycolation the presence of some water in the expanded layers does not alter the d-spacing (Mackenzie, 1948). An X-ray diffractogram of this sample showed that the smectite has a basal (001) spacing of 16.8 – 17.1 Å with a moderately broad peak.

The glycolated sample and one of the non-glycolated samples were studied under high power while drops of ethylene glycol were carefully placed on the clay to saturate the whole surface. There was no visible disturbance of the oriented clay material during this process. When allowed to dry for 12 hrs in an ethylene glycol atmosphere both samples produced X-ray diffracto-
grams similar to the original glycol saturated specimen but with a slightly sharper peak at 16.9 Å.

A further three samples were glycolated using Brunton's (1955) method with the modification proposed by Kunze (1955) and allowing the samples 1 hr, 3 hrs and 6 hrs respectively in the ethylene glycol vapour at 60°C. Even after allowing a further 24 hrs to come into equilibrium with cold ethylene glycol vapour the X-ray results from these samples were not as consistent as those obtained by glycol saturation. The basal (001) peak was generally lower and broader than the saturated peak and varied from 16.4 to 16.8 Å with increasing time of glycolation. While the loss of precision can possibly be attributed to the thickness of the clay layer and the lack of relative humidity controls on the diffractometer, it is apparent that vapour glycolation of ceramic disk mounted clay minerals is neither as quick nor as consistent as glycol saturation by direct application.

A further test was run on glycol saturated specimens to determine the time required to obtain ethylene glycol equilibrium and the rate of glycol loss by evaporation. A sample was glycol saturated and the wet sample was irradiated at 12 minute intervals. The first three diffractograms are reproduced in Figure 4.30 and show that, as soon as the surface of the sample appears dry after 24 minutes, the diffraction pattern is virtually identical to the previous glycol saturated
samples where 24 hrs had been allowed for the sample to come into equilibrium. A glycol saturated sample was left exposed to cold glycol vapour for 10 days and its diffraction pattern did not change. A second sample exposed to the atmosphere showed a slight but continuous decrease in the basal spacing after 3 days. A third sample was vacuum desiccated for periods of 15 minutes, 45 minutes and 3 hours without a change in the basal spacing but after a further 24 hours in a dehydrant desiccator the basal (001) spacing had decreased to 16.3 Å.

The conclusion reached from this study is that for oriented clay mineral samples mounted on ceramic disks glycolation can be achieved most successfully, quickly and reproducibly either by glycolating prior to mounting or by carefully saturating the clay surface with liquid glycol. With glycol saturated samples no adverse effects were noted when the diffractometer is not fitted with relative humidity control apparatus.

(3) **Procedure used for Clay Mineral Analysis**: Five to ten grams of each sample were crushed in an agate mortar, X-rayed, then disaggregated by irradiating in an ultrasonic bath for 15 minutes with about 250 mls of
0.1 N sodium oxalate. The minus two micron equivalent settling diameter fraction was separated and an oriented specimen from each sample was coated onto a freshly ground ceramic disk using the vacuum sedimentation method of Kinter and Diamond (1956). The mount was washed with distilled water, three times with dilute magnesium chloride solution to expand vermiculites to normal d-spacings, and with distilled water again. A further set of specimens was prepared by allowing the minus two micron fraction to settle on glass slides by evaporation. All samples were air dried in a dehydrant desiccator for at least 48 hours. The samples were analysed by X-ray diffraction analysis using nickel-filtered copper K\(\alpha\) radiation at 40 KV and 25 mA with a Philips PW1051/30 goniometer, PW1010 power source, and PW1053 discriminator.

A series of six diffractograms were run for each sample:

1. A crushed whole rock sample mounted in an aluminium holder without preferred orientation, run in the range 2 to 50\(^\circ\) 2\(\theta\) with 1\(^\circ\) slits at 1\(^\circ\) 2\(\theta\) per minute using a chart speed of 20 mm per minute.

2. The clay sized fraction treated with magnesium chloride and mounted with preferred orientation, run in the 2 to 26\(^\circ\) 2\(\theta\) range with \(\frac{1}{4}\)\(^\circ\) slits at \(\frac{1}{4}\)\(^\circ\) 2\(\theta\) per minute using a chart speed of 10 mm per minute.

3. As above after saturating with liquid ethylene glycol run in the 2 to 14\(^\circ\) 2\(\theta\) range with the same settings.

4. As above after heating to 450\(^\circ\)C for 30 minutes
followed by dehydrant cooling.

5. As above heated to 600°C for 1 hour followed by dehydrant cooling.

6. The clay samples mounted on glass slides, run in the 3.4 to 3.7 Å range using 1/4° and 0.1° slits at 1/8° 2θ per minute with a chart speed of 5 mm per minute, to distinguish kaolinite and chlorite peaks.

Semi Quantitative Clay Mineral Analysis

A semi quantitative clay mineral analysis was carried out on all forty samples using relative peak area intensity ratios for all species present and recalculating to 100 percent. A critical review of quantitative clay mineral analysis by van der Marel (1966) showed that there are many uncontrolled variables effecting X-ray intensities e.g. variable X-ray machine conditions, clay thickness and grain size, degree of preferred orientation, degree of crystallinity, masking of peaks, etc. Also the effect of grinding the sample may damage the crystal structures to an undeterminable extent. Therefore, areas measured under peaks on a diffractogram only give estimates of the main groups present unless the individual clay mineral species are separated by some method e.g. Gibbs (1967). Intersample comparisons are only meaningful if ratios are used to compare clay units rather than individual mineral percentages.
The measurement of the 10 Å illite peak area, the 7 Å kaolinite - chlorite peak area, and the 17 Å glycolated smectite peak area enabled the amount of these four main groups to be estimated. Pierce and Siegel (1969) noted that the most important factor affecting the clay mineral percentages is in the method of calculation from given diffraction results. Their second method was used throughout this study with the relative amount of mixed layer clays being calculated from the ratios of its area to twice the illite 10 Å peak area. The kaolinite-chlorite ratio was determined from the 3.58/3.54 Å ratio (Biscaye, 1964) and the 7 Å peak area was divided proportionally. The amount of chlorite was confirmed by measuring the 14 Å and 4.75 Å peaks. Inter-stratified clay minerals are abundant in almost all samples. The proportion of these minerals is given in Table 4.6; they may be classed as swelling or non-swelling inter-stratified clays.

Mineral Identification

The methods used for the identification of clay minerals were adopted from Warshaw and Roy (1961), Weaver (1958) and Biscaye (1965). The clay mineral "groups" identified include smectite, chlorite, illite, kaolinite and various mixed layer clays.

The term smectite is reserved here for that material which, upon glycol saturation, expands to 17 Å.
In the non expanded state the smectite has a variable basal spacing between 12 and 15 Å and generally shows a very broad and ill-defined peak on the diffractogram. Likewise the 16.8 - 17.0 Å glycolated peak is usually broad, but occasionally sharp when smectite is relatively abundant. The broad nature of the peaks indicates that the smectite is probably very fine grained and also poorly crystallized. An approximate crystallinity index was derived by Biscaye (1965) and is indicated on Table 4.6. The index is defined as the depth of the valley on the long spacing side of 17 Å divided by the peak height, and it may be positive or negative. The average crystallinity in the samples studied is 0.494 and the range is from -0.45 to 0.93. The frequency distribution of crystallinity index is given in Table 4.7. Chlorite minerals were identified by a combination of the 14, 7, 4.75 and 3.54 Å peaks. The distinction and proportion of chlorite and kaolinite was determined from the 3.54/3.58 Å doublet (Biscaye, 1964). In the samples studied the peaks vary from broad to sharp and they intensify and sharpen on heating to 450°C. After heating to 600°C the 001 peak was intensified and all other chlorite peaks destroyed. This suggests that the chlorite is moderate to well crystalline (Brindley, 1961). The generally high values of the corrected 002/003 and 002/001 peak height ratios (1.8-5 and 1.2-3 respectively) are usually associated with an iron rich chlorite (Brindley, 1961). The relatively low 003 peak also indicates that the chlorite is trioctahedral. No septachlorites were
recognized in this study.

Three samples were treated with 2 N hydrochloric acid at 80°C for eight hours to remove the chlorite. The result of these analyses confirmed the absence of kaolinite in sample DH98G and the presence of it in DH78 and CJ29.

The presence of vermiculite especially in samples BR45 and BR53 was suspected from the 14.4 Å peak but due to masking by mixed layer minerals its presence could not be confirmed on heating.

The term illite is used in the broad sense of Grim et al. (1937) to include all clay sized minerals of the mica group. The illite content of the fine-grained rocks was determined from the 10 Å peak assuming symmetry. Most of the illite peaks were sharp and this indicates that it is dioctahedral - probably largely muscovite. It must be noted (Bradley & Grim, 1961) that dioctahedral material gives more efficient diffraction and therefore emphasizes itself at the expense of probably lesser quantities of trioctahedral micas. Thus the relatively low 001/002 ratio, which indicates the presence of iron rich micas, suggests that trioctahedral biotite may also be present. This has been confirmed optically in the silt size range. The degree of crystallinity, or metamorphism, can be determined from the sharpness ratio of the peak height at 10 Å divided by the peak height at 10.5 Å (Weaver,
The frequency distribution of the illite crystallinity is given in Table 4.8 along with the metamorphic grade equivalents derived by Weaver. These sediments show a range in crystallinity from 1.20 to 6.75 and an average of 2.987 which is equivalent to a grade of incipient metamorphism. The crystallinity of the illite was not determined by the method of de Segonzac et al. (1968) because the amount of mixed layer minerals would cause considerable difficulty in determining the width of the illite peak.

The presence and quantity of kaolinite was determined from the 3.54/3.58 Å doublet and the 7 Å peak, and confirmed by the 2.384 Å peak. Kaolinite is usually quite abundant in the samples in which it occurs and it is typically well crystallized and has sharp diffraction peaks. On heating to 450°C the 7 Å peak was not greatly reduced (cf. Zen, 1959) but it was destroyed after heating at 600°C for one hour.

Mixed layer clay minerals are relatively common in all the siltstone samples but they are fairly rare in mud clasts within the Hermannsburg Sandstone. In non-glycolated specimens the basal spacings of the mixed layer clays vary from 10 to 14.2 Å although no 24 to 30 Å basal spacings were recognized. The variable basal spacings are due to random or ordered inter-stratification of clay mineral species. Three types of mixed layer clays have been recognized and are shown in Table 4.6. Some
samples show an irregular tail on the short spacing side of the illite peak from 8.5 to 9.5 Å. The origin of this mixed layering is uncertain but it may represent illite weathering to kaolinite. Expanding mixed layer clay minerals have a glycolated peak between 14 and 17 Å which collapses to 12-14 Å on heating. They may be termed expanded 2:1 minerals and may represent dioctahedral vermiculite, randomly inter-stratified chlorite-vermiculite, or complexed smectite. Non expanding mixed layer minerals have basal spacings between 11 and 13 Å and occasionally show a moderately developed 12 Å peak. These minerals probably represent inter-stratified illite - chlorite, although sepiolite could be represented by the 12 Å peaks.

Non-Clay Minerals

A comparative mineralogical composition ratio for the non-clay to clay minerals was carried out using the method of Deere and Bayliss (1969) with some modifications. From the diffractograms of the disoriented whole rock samples, the following peak area intensities were measured above the background level:-

1. Illite 001 at 10.0 Å
2. Quartz 101 at 3.35 Å
3. Feldspar 040 at 3.2 Å
4. Calcite 104 at 3.03 Å
5. Dolomite 104 at 2.88 Å
The illite intensity was modified proportionally to represent the whole clay mineral fraction by using the percentage of illite in the clay. The quartz peak was also modified to exclude the 003 illite peak. Each of the five peaks has a relative intensity of 100 for the mineral represented and an intensity ratio, or percentage can be calculated for each mineral (Table 4.6).

\[
\% Q = \frac{Q_{101} \times 100}{I_{001} + Q_{101} + F_{040} + C_{104} + D_{104}}
\]

Deere & Bayliss found that, in samples containing only quartz, calcite and dolomite, the percentages calculated by this method are in good agreement with chemical analysis. Although extending this method to include feldspar and clay minerals introduces numerous further sources of error it is considered that the ratios obtained are still meaningful for intersample comparison.

The calcite to dolomite area ratio can also be used to estimate the percentage of these minerals using the method of Bromberger and Hayes (1966) but the accuracy of measurement required for precise correlation with the graphs was not warranted in this study.
The following generalizations can be made from the data in Table 4.6. The whole rock analyses show a fairly consistent but variable quantity of quartz in all the samples studied. Most of the samples fall in the range from 35 to 77 percent and the mean content is 49 percent quartz. Likewise feldspar is quite widespread but only forms an average of 8.6 percent of the rock. The calcite and dolomite distributions are variable and are closely associated with siltstone horizons, especially in the Parke Siltstone on Dare Plain (DH) and the Horseshoe Bend Shale (FG22, 26 and FK). Both minerals are absent from the mudclasts and are generally less abundant in the silty interbeds in sandstone sequences. This may be a primary feature or may be due to post-depositional leaching in the more porous sandstones. The clay fraction forms between 5 and 82 percent of the samples with an average content of 25 percent. There does not appear to be any direct relationship between clay content and lithology.

The clay minerals themselves show some systematic variation but are generally fairly uniform. The contents of illite and chlorite show little variation with means of 55 and 13 percent respectively. However, chlorite is distinctively low or lacking from the mudclasts. Mixed layer clays of the three types noted are likewise fairly randomly distributed although there is a tendency for higher percentages to be recorded from siltstones inter-
bedded with sandstones. The Finke Group sediments have a slightly higher proportion of mixed layer minerals than the Pertnjara sediments. Smectite has a random distribution with the highest values being recorded from the upper and lower Parke Siltstone and the lower Horseshoe Bend Shale. Kaolinite is generally scarce or absent from the silty samples but is abundant in the mudclasts. One exception is the occurrence of 38 percent of kaolinite in a siltstone from the upper Langra Formation (FG18).

The illite and smectite crystallinity ratios are also variable and the latter are too random to be significant. The illite crystallinity shows a couple of trends. The highest values of the crystallinity ratios are usually associated with mudclasts. The ratio is also significantly higher in the Pertnjara Group siltstones (excluding the mudclasts) than it is in the Finke Group i.e. 3.04 cf. 1.72.

Conclusions from the study of the clay minerals indicate that only in the mudclasts is there a major change in mineralogy. The siltstones are characterized by the presence of calcite, dolomite, illite, chlorite, 14 to 16 Å mixed layer clays and variable quantities of smectite. In comparison the silty interbeds are generally low in calcite and dolomite but contain much the same range of clay minerals. The fairly constant clay mineral assemblage probably reflects fairly constant environmental and climatic factors during the period of
deposition. Leaching was probably not intense and the environment may have been reducing or mildly oxidizing.

The mudclasts form a distinctive group characterized by the dominance of quartz, kaolinite and illite. The illite has a higher than normal crystallinity ratio and this may be due to the preferential breakdown of less stable forms of mica during weathering. The restricted assemblage in the mudclasts may reflect a mature, strongly oxidized, acidic depositional environment, the effect of post-depositional acid leaching and oxidation, or a combination of both these factors.

The colour of the siltstones on Dare Plain is only weakly reflected in the mineralogy of the samples. In closely associated pairs of samples the greenish siltstones tend to be slightly more calcareous and contain an average of 7 percent more chlorite than the brownish ones, i.e. comparatively the greenish siltstones contain 35 percent more chlorite than the brown ones. Some of the brownish siltstones contain a considerably higher proportion of smectite than the greenish siltstones but in some brownish samples this mineral is absent. These changes probably represent periodic changes in the environment of deposition and possibly reflect two modes of transport or the mixing of two different assemblages.
Weaver (1958) noted that carbonates contain essentially similar clay mineral suites to associated shales and this concept was found to be true in the present study. The effects of post-depositional alteration of clay mineral suites in permeable rocks has been studied by Smoot (1960). He found that the mineralogy of surface samples of shales varied only slightly from their subsurface equivalents whereas clay minerals in permeable sandstones are usually extensively altered both in surface and subsurface samples and are unlikely to resemble the original clay minerals. Hence the abundance of kaolinite and well crystallized illite in mudclasts from the sandstone members must be viewed with caution. They may very well represent pluvial Tertiary weathering products and the concentration of well crystallized illite would be due to its greater resistance to weathering than the more disordered forms.

Illite is the most abundant detrital clay mineral in most clay suites and it is usually concentrated by the action of winnowing. It usually forms in a high-potassium environment, such as in conditions of temperate to semi-arid weathering (Folk, 1968). It may develop from other clay minerals by diagenesis, deep burial and slight metamorphism but most of it is probably derived from acid igneous and metamorphic rocks and older illitic sediments. It is common in, and probably responsible for the colouration of the green shales in the Parke Siltstone.
Chlorite is most abundant in areas where chemical weathering is not very intense. Chlorite is a primary constituent of many igneous and metamorphic rocks although it may also form by the alteration of iron silicates such as biotite, hornblende, etc. In the Pertnjara and Finke sediments the iron rich chlorite is probably detrital and would have been derived from the igneous, metamorphic and sedimentary basement rocks.

Smectite may form by the weathering of eruptive igneous or basic rocks in a magnesium rich alkaline environment subjected to very little leaching. It may also form as an alteration product of chlorite (Smoot, 1960). The abundance of smectite, or smectite-like clay minerals, could be due to two factors. It could be a primary weathering product of volcanic detritus such as the Mt Harris Basalt. However, volcanic rocks are infrequent in the proximal source area and it is considered to be unlikely that all the smectite could have formed in this manner. Smoot (1960) suggested that chlorite could break down to a chlorite-vermiculite mixture. The magnesium ions present in the vermiculite could then be replaced quite readily by calcium and sodium ions thus producing a degraded mineral which would not differ appreciably from smectite. Grim et al. (1960) proposed a similar method of formation of smectite from chlorite to explain the presence of the former in the Great Salt Lake, Utah. A similar method of smectite formation can be invoked for its presence in the
Pertnjara and Finke Groups where high percentages of smectite are generally associated with lower percentages of chlorite. This is especially noticeable in the mudclasts where smectite is often present at the expense of chlorite thus suggesting that it is a degradation or weathering product.

Kaolinite forms as an alteration product of aluminous silicates subjected to intense weathering in a potassium deficient acid environment. The general scarcity of kaolinite in the Pertnjara and Finke siltstone suggests that such environmental conditions were probably not present and that the small quantities of kaolinite were probably reworked from earlier sediments. The high kaolinite content of the mudclasts is probably a result of post-depositional alteration.

Mixed layer clay minerals usually form as degradation or weathering products of pure species although some are primary multilayer associations. The minerals which expand to 14 Å to 16 Å are probably either degraded chlorite or illite-smectite interlayered species. The non-expanding 10.5 Å to 14 Å minerals probably consist of illite in various stages of degradation. The 8.5 Å to 9.5 Å minerals may represent a clay species in the process of altering to kaolinite. All these mixed layer clay minerals may represent weathering or alteration products of purer species either in the source area, or possibly after deposition. Environments
of formation of the various minerals are not clearly known and some of them could be reworked detrital minerals from the earlier sediments.

The effect of depth of burial on the clay mineral suite is not particularly noticeable in these samples except that all samples from well holes contain little 14 Å to 16 Å mixed layer clays. This can be interpreted that either the original samples contained little smectite or that the depth of burial was greater than about 3500 m (Weaver, 1960).

The clay mineral association in these sediments is non-diagnostic for the determination of weathering or depositional environments. Such an assemblage could be found in most environments except a wet, strongly acid leached, continental deposit where kaolinite would be dominant. The dominance of kaolinite in the mudclast samples, however, is not considered to be reliable since it could have formed as a result of post-depositional weathering.

Red coloration and clay mineral assemblages in soils cannot adequately characterize exclusive climatic conditions (Power, 1969) except in extreme cases. Power found that clay assemblages vary through a single weathering profile and they are related to the duration and intensity of weathering, which is controlled in turn by climate and tectonism. The red coloration is due to the
presence of hydrated ferric oxides (hematite) in small quantities throughout the sediment. Their presence indicates a lack of preserved organic matter and may be a result of permanent or seasonal aridity, high temperatures combined with active microbiota, or the infertility of the soil (Power, 1969). The abundance of ferric oxides and the pronounced lack of organic matter in both the siltstones and sandstones of the Pertnjara Group is somewhat analogous to this situation. Likewise, they may have formed under a climate subjected to seasonal aridity since many of the beds are associated with desiccation cracks and halite pseudomorphs.

The assemblage of illite, smectite and chlorite with lesser quantities of mixed layer minerals, is known from the lacustrine Tipton Shale Member of the Eocene Green River Formation of Wyoming (Tank, 1969). It is considered to have been deposited in a fresh water lake with a strong reducing environment but the assemblage in the overlying hypersaline Wilkins Peak Member is also similar.

A succession of laminated, calcareous and dolomitic shales, somewhat similar to the Pertnjara and Finke siltstones, was described by West et al. (1968) from the Carboniferous Aghagrania Formation of County Leitrim, Ireland. They consist of unfossiliferous shales, including greenish grey unfossiliferous micaceous and silty mudstones with frequent desiccation cracks,
interspersed with dolomitic and calcareous shales with halite casts and gypsum beds. However, these sediments are also associated with algal colonies and marine shales and are assumed to have formed in a tidal flat environment.

Bissell and Chilinger (1962) record impure to relative pure dolomitic horizons on the Utah salt flats and playas. The dolomite is probably forming as an early diagenetic mineral from primary calcite or aragonite. Its production is favoured by a relatively high pH (> 9) and high alkalinity and is probably hastened by elevated temperatures. It may be precipitated directly in environments with high carbon dioxide pressures, high salinity and a pH of less than 8 (Chilinger and Bissell, 1963). Skinner (1963) also noted that calcite and dolomite can be precipitated simultaneously from shallow, saline water with a pH of 8.9 - 9.2 thus confirming the earlier observations of Strakhov (1953). These conditions are not conducive to the formation of kaolinite and the small quantities present in these lakes are probably detrital. The deposits are characterized by cycles of fresh, saline and hypersaline environments with periods of aridity, desiccation and salt crystallization alternating with lacustrine sedimentation.

Grim (1953) noted that sepiolite – attapulgite minerals are characteristic of saline lake deposits but
later studies have shown this concept to be incorrect. e.g. Grim et al. (1960), Eardley and Gvosdetsky (1960). Parry and Reeves (1968) also noted that sepiolite may occur in some, but not all, saline lake deposits.

The evidence from the study of the fine grained strata indicates that deposition of the Parke Siltstone and Horseshoe Bend Shale probably took place in a warm, shallow water, saline environment subjected to periodic desiccation and inundation with the associated crystallization and solution of salt crystals and gypsum. Analogous sedimentation in Recent playa lake environments and the lack of any definite marine associations strongly suggest deposition in a widespread basin(s) forming the nucleus of an internal drainage system in a seasonal climatic regime.

The clay mineral suites do not indicate any particular depositional environments for the sandstone units.
4.3 HEAVY MINERALS

A study of the heavy mineral suite of 30 representative samples from the Pertnjara and Finke Groups was undertaken to provide additional evidence of source area lithology and stratigraphic correlations.

The effect of post depositional removal of unstable heavy minerals by intrastratal solution has been discussed by many authors. Potter (1968) found that the distribution of garnet in a single formation was restricted to rocks with a high iron oxide or argillaceous content. This implies that impermeable rocks are more likely to contain a representative heavy mineral suite than are porous sandstones. Blatt et al. (1969) studied the relationship between heavy mineral suites in adjacent sandstones and shales and found that, while the main minerals were approximately the same, the shale samples had slightly more numerous and variable accessory minerals.

With the effect of intrastratal solution in mind, the samples selected for analysis were either shaly siltstones or fine to medium grained sandstones containing abundant matrix material.

It was reported by Henningsen (1967) that heavy mineral analyses using crushed indurated sedimentary rocks closely represent the actual heavy mineral distribution of
the sample. When analysing argillaceous sediments, crushing may not disaggregate the heavy minerals completely and thus reduce the proportion of the lightest heavy mineral species (lower specific gravity minerals).

**Heavy Mineral Analysis**

Ten to twenty grams of each of the 30 samples were crushed, and then irradiated in an ultrasonic bath for 15 minutes with 0.1N sodium oxalate to disaggregate the clay minerals. The minus 10 micron fraction was removed by washing with sodium oxalate and allowing the suspension to settle for 20 minutes in a 10 cm deep column. The material coarser than 10 microns was slowly boiled in 0.3N sodium oxalate with aluminium foil to remove iron oxides. It was dried with acetone and the heavy mineral fraction was separated using tetrabromoethane with a specific gravity of 2.90. The heavy mineral concentrates were mounted in arochlor (refractive index 1.66) on a graticule slide. Quantitative analysis was carried out by counting 400 grains on a number frequency basis using linear traverses across the whole slide. This is theoretically the best method for inter sample comparisons (Galehouse, 1969) but it has the disadvantage that weight frequency distributions cannot be calculated. However, in this study number frequency is more meaningful for representing a heavy mineral assemblage than number percent because of the non-uniform grain sizes encountered. It was considered that
the inaccuracy introduced by studying only one non-sized heavy mineral assemblage per sample (Young, 1966), would not mask any important trends which may be apparent from the sparse sample coverage.

X-ray Analysis

The compositions of several heavy minerals were checked by using the X-ray powder photograph technique.

Hand picked concentrates of both clear and pale pink garnets were obtained from sample AF83 and the powder patterns from both samples were almost identical. The unit cell dimension of the pink garnet was calculated using the method set out in Nuffield (1966), and found to be $11.525 \pm 0.001 \AA$. The refractive index is 1.792 and the specific gravity is 4.03. Using the triangular diagrams of Sriramadas (1957) three possible compositions would account for the measured result.

(1) Almandite 51%  (2) Almandite 64%  (3) Almandite 60%
Pyrope 29%   Pyrope 30%   Pyrope 36%
Spessartite 20% Grossularite 6% Andradite 4%

When plotted on Winchell's (1958) diagrams the compositions obtained are very similar but the specific gravity measured favours one of the first two compositions.

Similar hand picked concentrates of the garnet in samples FG28 and FH30 gave identical powder photographs and are assumed to have a similar composition.
Opaque minerals from sample AF83 were also studied. A concentrate of the platy opaque mineral proved to be hematite while the almost equidimensional rounded grains with aggregate metallic reflections were found to be ilmenite.

Grey to white translucent aggregates of extremely fine grained, crystalline material from sample AJ142 were determined to consist almost entirely of apatite with a very minor amount of quartz.

Results

The results of the heavy mineral analyses are presented in Table 4.9 and the range and average value of the most important constituents are given in Table 4.10. From the tables it can be seen that the assemblage is dominated by opaque minerals, tourmaline and zircon which together constitute an average of 83 percent of the heavy mineral suite. All these minerals are stable to weathering and abrasion and the suite is characterized by well rounded grains.

1. Opaque Minerals: The opaque minerals consist largely of well rounded to rounded, almost equidimensional grains of ilmenite with an aggregate reflection and metallic lustre in reflected light. Ilmenite grains are frequently seen with a whitish, translucent coating of leucoxene. Grains of leucoxene or fully coated ilmenite grains are
common (average 12.5% of heavy minerals). They are the dominant opaque minerals in the more porous sandstones e.g. in sample DI50 93 percent of the opaque minerals are leucoxene. In most samples the proportion of magnetite in the opaque minerals is small - as determined by magnetic separation. Likewise the proportion of hematite in most samples is low although it becomes progressively more abundant up the stratigraphic column. It usually occurs as a platy mineral and may be authigenic.

2. Zircon: In most samples zircon is an abundant heavy mineral which forms an average of 20.4 percent of the heavy mineral suite. The grains are typically elongate and well rounded to rounded in the sand sized fraction and decrease in roundness to subrounded in the coarse silt, and subangular in the medium silt sized fractions. Only a few samples contained subangular to angular euhedral grains e.g. EH1, FD11, GA43, AF83, AJ186 and AJ196. The last three samples are all from the Brewer Conglomerate and contain several prismatic grains with acicular needles of rutile. Prismatic zircon is especially abundant in AJ196 with lesser quantities of rounded zircon. Inclusions in the abraded zircons are varied and include needle-like crystals of rutile, rare negative crystals, occasional opaque spots, and relatively numerous bubbles or cavities. Most of the rounded zircon grains are colourless, with a few pinkish grains and rare yellowish grains. However, in the subangular and prismatic grains there is a higher proportion of pinkish grains, and
also a few more yellowish grains. Sample AF83 contains a subrounded zircon crystal with recemented fractures. Very few zircon overgrowths were seen and those present usually showed signs of wear.

3. Tourmaline: Tourmaline is, likewise, an abundant heavy mineral in most samples with an average content of 22.9 percent. Tourmaline is more abundant in the coarse silt to medium sand size range than in the finer fractions. In almost all samples the tourmaline is well rounded to rounded and grains vary in shape from spherical to ovate. Exceptions occur and most subrounded grains consist of part of a fractured, well rounded grain. Occasional to rare, subangular to prismatic grains of tourmaline occur in samples AF83, AJ186, AJ196, EH1 and FH30. A few of the tourmalines contain inclusions, especially bubbles or cavities and occasionally acicular crystals of rutile, e.g. in sample AJ32 occasional yellow-brown tourmalines contain large needles of rutile. There are numerous colour variations shown by the tourmaline and the main pleochroic pairs recognized are listed:

1. Pale Blue: colourless, dark blue, mauve, black. (NaFeLi indicolite)
2. Yellow-green: colourless, light yellow, light brown. (Schorl or Elbrite)
3. Brown: yellow, dark brown, black. (MgFe\rightarrow Fe Schorl)
4. Pink: dark blue, black. (Li rubellite)
5. Blue-green: yellow, blue
Occasional banded brown and blue-green tourmalines were noted e.g. sample CS12. In other samples (e.g. EM9) several tourmaline grains appear opaque except on thin edges where they are usually yellowish brown or greenish.

4. Rutile: Rutile is a fairly constant minor member of the heavy mineral suite. The grains are usually sub-rounded to well rounded, and euhedral rutile is very rare e.g. sample AJ142. The grains were rarely seen to contain inclusions, and varied in colour from orange to dark reddish brown with most of the grains being red.

5. Biotite: The quantity of biotite in the heavy mineral samples is variable and this may be due to natural abundance or irregular separation due to hydraulic shape. The flakes are variable in size and show pleochroism from colourless or yellowish to brown.

6. Garnet: Garnet occurs as colourless subrounded grains in many samples where it forms a low proportion of the heavy mineral fraction. It is only abundant in four samples (AF83, AJ196, FG28, and FH30) where it is present as subangular to subrounded equidimensional grains or angular anhedral grains. Both types show a variation in colour from colourless to very pale pink and have been determined by X-ray analysis to consist predominantly of almandite. In sample FH30 rare garnet crystals contain numerous acicular inclusions.
7. Apatite and Collophane: These minerals are almost always present as accessory minerals. In samples with less than two percent of the minerals they are dominantly extremely fine grained apatite (X-ray analysis) although a few samples contained rare crystalline apatite. Where these minerals are more abundant the major constituent is still extremely fine grained apatite but structureless or pelletal, yellow-brown, isotropic collophane grains are also important.

8. Epidote: Colourless to greenish yellow anhedral fragments of epidote occur sporadically through the samples but are only abundant in sample FH30.

9. Staurolite and Kyanite: Rare platy fragments of bluish kyanite and pale yellowish staurolite occur in only three samples e.g. AF83, AJ196, FH30.

10. Other Minerals: These form from 0 to 3.7 percent of the heavy mineral suite. The most common of these minerals are greenish beryl, and greenish brown hornblende. Pyroxenes occur in a few samples. Very rare heavy minerals include spinel, andalusite, clinozoisite, and possibly monazite and topaz.

A fairly detailed heavy mineral study was conducted by the French Petroleum Co. (Aust.) Pty Ltd, (Leslie, 1965) in the Alice Springs, Rodinga, Hale River and Finke Sheet areas. The results of this study
are presented in columnar form in Figs 4.31-38.

In addition to these heavy mineral studies accessory mineral counts were made on all the thin sections studied. These counts have been corrected to a constant area for intersample correlation (Appendix 7) and they are presented in Table 4.11.

For most of the minerals present in thin section the descriptions given above for the heavy mineral concentrate are adequate to explain all the features seen. Opaque minerals were not separated into two groups, although in the few sections studied with reflected light a decrease in the white opaque count and an increase in the leucoxene coated ilmenite grains was apparent thus indicating that the leucoxene on many grains is just a coating. The most plentiful mica is muscovite although in a few samples biotite dominates. Only three varieties of tourmaline were recognized in thin section with the distinctive colours being blue, yellow-brown and green to green-brown. Rounded detrital grains of glauconite are relatively common in some samples, especially those from the northern margin of the Amadeus Basin, while they become less abundant to the south and east, with local exceptions. The grains are usually bright to dark green in colour and show an aggregate polarization. Occasional grains in several samples are weathered to a yellowish green or greenish brown but they are still recognizable as detrital glauconite grains.
The grains grouped under the term "others" are predominantly epidote which occurs in many samples in a variable but low concentration. Samples FE, FF and FI are unusual in that epidote is relatively abundant as schistose epidote-quartz fragments. Beryl is a relatively common minor constituent, while hornblende, actinolite, pyroxene, chlorite and possibly kyanite and staurolite have a very restricted occurrence usually in stratigraphically higher samples.
All the samples selected for heavy mineral separation were in the fine sand to coarse silt range, showed little sign of weathering, and were moderately well cemented without obvious replacement of detrital minerals. This minimizes the chances of post-depositional alteration of the heavy mineral suite and also allows direct comparison of samples since they are all of essentially uniform grain size.

The most abundant heavy minerals include rutile, zircon, tourmaline and opaque minerals such as ilmenite and magnetite.

Rutile is the least abundant of these minerals and forms an average of 1 to 4 percent of the heavy mineral assemblage. The grains are all moderately to well rounded and very rarely show recognizable crystal faces. They are probably all of polycyclic origin being derived from any of the earlier Amadeus Basin sediments. They would have originally come from an acid igneous, especially granitic, or metamorphic terrain.

Zircon is another stable heavy mineral frequently forming up to one third of the suite. In most samples it occurs as clear subangular to well rounded grains with the latter predominating. Rounded pinkish and yellowish zircons also have a sporadic occurrence but never form the dominant type. Roundness of zircon is a deceptive feature and cannot be taken as a direct indicator of distance of
transportation (Saxena, 1966). However, the dominance of rounded zircons is indicative of a polycyclic sedimentary source. In the Ljiltera Member, Brewer Conglomerate and Finke Group euhedral zircon crystals occur together with the more abundant rounded forms. A few of the euhedral crystals are clear and free of inclusions and could possibly have formed in sedimentary rocks such as the Bitter Springs Formation. Most of them contain bubble and negative crystal inclusions and are probably derived directly from an igneous or metamorphic terrain.

Tourmaline is a common heavy mineral but like zircon the grains are all dominantly rounded or well rounded and are frequently circular in section. Angular grains are rare except for broken rounded grains, although they are slightly more common at stratigraphically higher levels. Numerous pleochroic varieties have been recognized although their distribution does not appear to be systematic or significant. They have probably all been derived from a polycyclic sedimentary source rather than a granitic or pegmatitic source. The variety of tourmaline present indicates derivation from all main source rocks with probably the largest proportion coming from pegmatite injected metamorphic terrains (Krynine, 1946). The presence of minor quantities of varicoloured tourmaline grains is also indicative of a granitic pegmatite source (Staatz et al., 1955).
The opaque heavy mineral suite is dominated by ilmenite and leucoxene with lesser quantities of magnetite and secondary hematite. The occurrence of magnetite and ilmenite is indicative of an oxidizing environment both in the source and depositional areas (Miller and Folk, 1955) since both these minerals are unstable in reducing conditions. The grains are usually subrounded to rounded and are probably of a polycyclic origin from an original igneous or metamorphic terrain.

Sedimentary apatite, glauconite and collophane usually occur as subrounded to rounded grains although collophane is occasionally in a curved, tabular, platy form probably representing exfoliated oolite fragments. There is no evidence for any of these minerals to have formed in situ and they are all considered to be detrital derivatives of pre-existing marine sedimentary rocks in the source area. The occurrence of these minerals in the older rocks has been discussed earlier (Chapter 1) although the most important sources would probably be the Pacoota and Stairway Sandstones.

Accessory heavy minerals fall into two classes. Fluorite, monazite, beryl, topaz and some epidote and garnet may occur as subrounded grains scattered throughout most samples of the Pertnjara and Finke Groups. They represent minerals derived previously from a granitic or metamorphic source but they may have been through one or more sedimentary cycles before being incorporated in
these sediments. Accessory minerals such as actinolite, kyanite, clinozoisite, some epidote, hornblende, staurolite and garnet typically occur as infrequent angular to subangular grains in the Ljiltera Member, lower Brewer Conglomerate and lower Horseshoe Bend Shale. They become more abundant, especially almandite garnet, towards the top of the Horseshoe Bend Shale and in the Undandita Member. They are all characteristic of a metamorphic terrain of schists and gneisses but some could also come from igneous sources especially granites and pegmatites.

Leslie (1965) concluded that heavy mineral assemblages in the Pertnjara and Finke Groups are generally unreliable for purposes of correlation. Heavy minerals which are common in most localities are of little significance in correlation. This category includes a major proportion of the heavy minerals e.g. opaque minerals, zircon, rutile, tourmaline, micas and ferromagnesian minerals (predominantly epidote). Rare heavy minerals likewise cannot be relied upon for correlation. This leaves only four mineral varieties - namely apatite, collophane, glauconite and garnet. Leslie (1965) noted that anatase is a significant heavy mineral characteristic of the Mereenie Sandstone. It only occurs in the basal Pertnjara Sandstones and can be used as a check on the base of the Pertnjara Group.
Sedimentary apatite, collophane and glauconite are important accessory minerals on the margins of the Amadeus Basin and they are most abundant in the MacDonnell Range area. Away from the basin margins these minerals decrease in abundance and are replaced by the stable zircon-tourmaline assemblage. All three are quite readily abraded during transport and Wermund (1964) noted that fluviodetrital glauconite is only common within 40 km of the source area. This decrease in abundance away from the margins is paralleled by the decrease in soft lithic fragments and indicates that the process of weathering and/or abrasion was quite severe. The distance and occurrence of these three minerals away from the source area gives an indication of their stability when relative abundances are considered. Sedimentary apatite is more stable than glauconite and is a relatively constant minor constituent of the heavy mineral units in the centre of the basin (0.5-2%). Collophane is the least stable of the three and is almost restricted to the margins.

Almandite garnet is a relatively significant heavy mineral. Its occurrence as euhedral and angular anhedral grains suggests that it is a first cycle derivative of a metamorphic terrain. Its association with staurolite, kyanite and epidote supports this view as does its relatively restricted occurrence in the Brewer Conglomerate of the Pertnjara Group. It is only abundant near the top of the Brewer Conglomerate, in the Undandita Member, where garnet forms over half of
the heavy mineral suite in sample AF83. In the Finke Group garnet is present in the Polly Conglomerate, rare or absent in the Langra Formation, and becomes increasingly abundant in the Horseshoe Bend Shale and basal Idracowra Sandstone. Sample FH30, from a very micaceous, very fine sandstone at the base of the Idracowra Sandstone, contains abundant garnet along with epidote, kyanite, staurolite and other metamorphic minerals. The pinkish garnet from this sample is identical to the almandite garnet of sample AF83 and the similarity of the assemblages suggests a possible correlation between these units. However, this cannot be validated because of the lack of stratigraphically equivalent samples from intervening areas. The porous sandstones from the mid and upper Idracowra Sandstone do not contain any unstable heavy minerals and their assemblage consists predominantly of zircon and tourmaline. This may be a primary feature or it may be due to post-depositional weathering of the less stable minerals.

The interval from 795 m to 1170 m in Witcherrie No. 1 Well, which was identified as Mereenie Sandstone by Magnier (1964) was found to have a heavy mineral assemblage including abundant zircon, pink-blue and green-brown tourmaline, moderately common rutile, apatite, and garnet, and infrequent amphibole (Leslie, 1965). This assemblage is not typical of the Mereenie Sandstone and Leslie related it to the Horseshoe Bend
Shale although it is probably more characteristic of the Polly Conglomerate.

The heavy mineral suite was subdivided into fractions with varying stability to weathering and abrasion. The stable fraction includes zircon, tourmaline and rutile, while the unstable minerals encountered include glauconite, collophane and sedimentary apatite. All other constituent minerals were included in the moderately stable group. Results from plotting triangular diagrams using these end members were rather disappointing because of the lack of unstable minerals in many samples.

Samples selected from various flow lines down the paleoslope (see Chapter 3) were plotted using the stable to moderately stable to unstable ratios expressed as percentages. Examples of such flow lines include the following sections AM, BO, BP, CY, ER, and ET; or AS, AR, BV, BS, GA, and EV; or CB, DC, DI, and DF. In all cases only one pattern emerged - namely the sections on the northern margin contain more unstable minerals than any sections further south and the latter all have points along the stable to moderately stable base line of a triangular plot (Fig.4.39). No pattern emerged from this stable to moderately stable ratio with either distance or stratigraphic height. Sediments of the Finke Group cannot be linked with confidence to those of the Pertnjara Group on paleoflow lines. They form
a distinct group on the moderately stable side of the Pertnjara sediments (Fig. 4.40) due to the abundance of opaque minerals and mica. This change in heavy mineral suite may reflect the change in provenance area or it could also be due to the effect of winnowing and mixing of sediments from northern and southern provenance areas.

Sediments from the northern margin do show some systematic variation with stratigraphic height. This is illustrated most clearly in the Glen Helen section (AJ, Fig. 4.41) which has the largest number of samples studied. The Harajica Member contains the most abundant unstable heavy minerals, including glauconite, collophane and apatite, which form up to 42 percent of the entire suite. The occurrence of stable and moderately stable minerals is variable but the latter usually predominate. The lower Hermannsburg Sandstone contains a generally similar suite but with less frequent unstable heavy minerals (less than 25%). Pebbly sandstones of the Ljiltera Member, and the Brewer Conglomerate are characterized by high quantities of moderately stable heavy minerals, especially opaque minerals and mica, and they form a distinct group near the moderately stable pole.

Other combinations of heavy minerals were plotted on ternary diagrams but with little success in differentiating distance or stratigraphic height when
single localities or successions of localities down the paleoslope were plotted. For example the combination of stable minerals (rutile, zircon, tourmaline), opaque minerals, and other minerals produced random plots on most paleoslope locality successions (eg. Figs 4.42 and 4.43). Finke Group sediments produced a wide and random scatter between opaque and other minerals with most samples having less than 25 percent stable heavy minerals (Fig. 4.44). This contrasts with the Pertnjara sediments which often contain up to 50 percent, and occasionally 70 percent, stable heavy minerals. Localities on the northern margin generally produce random plots which fail to differentiate the Harajica Member from the lower Hermannsburg Sandstone (Fig. 4.45). However, the pebbly sandstones of the Ljiltera Member, and the Brewer Conglomerate are characterized by higher proportions of opaque minerals. The section at Tyler Pass (AF, Fig. 4.46) is an exception because the Harajica Member can be separated from the Hermannsburg Sandstone. Also the basal sandstones of the Harajica plot as a separate group with a high proportion of stable heavy minerals. Such sandstones appear to be similar to the Mereenie Sandstone and are probably reworked directly from it.

The zircon to tourmaline ratio was calculated for each sample but was shown to be randomly variable throughout the basin. Most of the samples have ratios in the range between 0 and 1 indicating a general
excess of tourmaline over zircon. A few samples have excess zircon with ratios of up to 3, and the 12 samples with a ratio of 99.99 in Table 4.11 indicate that while zircon is present there is no tourmaline and thus the ratio is infinite.

All the sections were drawn up with the zircon to tourmaline ratio plotted against relative stratigraphic height (Fig.4.47). The following generalizations can be made. The ratio in the Hermannsburg Sandstone is generally lower and more consistent than in the sandstones of the Parke Siltstone. The ratios in the Brewer conglomerate are very variable and are characterized by extremes as is the lower Finke Group in McDills No.1 and at Horseshoe Bend. Several of the sections (eg. CJ, CS, GA) show an increase in zircon to tourmaline ratio in the upper part of the Hermannsburg especially at the onset of pebbly sandstones.

The increase in the zircon to tourmaline ratio at the start of pebbly sandstones in the James Range area and further east may be of use in stratigraphic correlation and lends support to the suggestion that the Ljiltera Sandstone Member and Brewer Conglomerate are lateral equivalents. In many cases, especially towards the top of the stratigraphic column, the increase in detrital zircon is largely due to the addition of more angular euhedral zircon crystals to the assemblage.
4.4 GRAIN SIZE ANALYSES

In the past, several methods have been proposed for determining equivalent sieve grain size distributions from consolidated rocks. It has been realized that disaggregation of consolidated sediments very rarely yields the individual grains in the same form in which they were deposited. This is especially true in the fine grained range and where labile fragments are common, since diagenetic alteration is likely to be more pronounced in such cases. Hence, great care must be taken in the selection of samples to be analysed by disaggregation and sieving. The obvious answer is to determine a correction factor whereby thin section size analyses can be converted to the equivalent sieve size distribution. The most important conversions are probably those proposed by Packham (1955) and Friedman (1958). A summary of conversion methods and their usefulness is given by Friedman (1958). A review of grain size parameter is given by Folk (1966) although he does not go into the methods of obtaining the size analysis in any great detail.

Packham (1955) devised a method for reconstructing sieve size distributions from thin section data using the Rosiwal method of planimetric analysis. Packham's method uses a volume frequency rather than a number frequency, and cumulative percentage curves are drawn up. The reconstruction of the equivalent sieve
-308-

size diameter is calculated by determining percentages under this curve and multiplying them by the appropriate correction factors given in Packham's Table 1. The main features in favour of Packham's method are the use of volume frequency and a short axis definition of thin section grain size measurements. Both these measures are analogous to those used in sieve size analyses.

Packham's method can be criticized on several counts. Firstly the method is derived mathematically on the basis of spherical grains and close packing - features virtually never attained in nature. The derivation of the size distribution by means of the Rosiwal method of planimetric analysis incorporates a serious limitation on the number of classes that can be recorded unless repeated traverses are made along each line. The correction factors derived depend upon numerous calculations each incorporating errors of reading percentages off a graph or obtaining interpolated percentages. To minimize these errors, especially in the tails of the distribution, both log-arithmetic and log-probability graphs have to be drawn and measured. The R : 1.05R correction factor has been shown to be incorrect by Vistelius (1958) and he also considers Packham's method is less accurate than it appears. The examples given by Packham are not generally applicable to consolidated sediments because of the method of preparing the thin section. The packing of loose grains in a plastic medium is hardly likely to yield a similar
packing configuration to natural sediments due to the viscosity of the medium and convection currents if it is heated. It is thus unlikely to yield a truly random sample of thin section areas and the excellent corrections recorded by Packham are not reproducible in natural sediment (Friedman, 1958, and this study). Packham noted that the reconstruction is inaccurate around sharp changes in gradient on the cumulative size distribution curve and that they are best when the sediment shows poor sorting over the whole size range. Most natural sediments, however, consist of mixtures of two or three log-normally distributed populations which characteristically produce sharp gradient changes on cumulative percentage curves thus making Packham's method less applicable. A further important limitation of Packham's method is that no measure of the coarse tail of the distribution can be obtained except by extrapolation i.e. to reconstruct the one percentile one has to accurately know, or extrapolate to find, the 0.3 percentile. Usually extrapolation is necessary, unless a very large number of grains are counted.

Friedman (1958) derived regression equations and graphs for converting thin section grain size measurements to their equivalent sieve size measurement. His study was very comprehensive as far as analytical techniques and possible sources of operator error are concerned. However, the sandstones he studied were dominantly well sorted orthoquartzites with more than
90 percent detrital quartz. Exceptions are the Desse, Cardium and Gallup sandstones which are subgreywackes but still contain over 70 percent detrital quartz. Friedman compared the median and quartile measures obtained from log-probability graphs of thin section long axis measurements (500 grains per sample) and sieve analyses of the disaggregated samples. He determined regression equations for each individual parameter and for the combined parameters and found that the latter fitted the data equally as well as the individual equations. The overall regression equation (in phi units) that he obtained for the three parameters is:

\[ Q_{\text{calculated}} = 0.3815 + 0.9027 \times Q_{\text{thin section}} \]

The main disadvantages of Friedman's method are that it was derived for quartz rich, well sorted sediments and has not been shown to be correct for labile and poorly sorted sediments. The direct application of Friedman's regression equation to poorly sorted sediments would have to be checked and the equation possibly modified to fit the conditions.

A criticism of Friedman's regression equation is that it does not allow for the mixing of two log-normal populations in the fine grained range. Thus at about four phi the measured long axis is apparently equal to the sieve intermediate axis and finer than four phi the measured long axis is shorter than the sieve intermediate axis. This could mean that the concept of random thin sectioning of grains no longer applies in this region.
However, the spread of data points above 3.5 phi is greater than below it and there is no control between 4 phi and 5 phi, therefore the extrapolation of the regression line beyond 3.5 to 4 phi must be suspect.

An additional error incorporated in Friedman's method is apparent when one attempts to extend it to the study of labile sediments. It is due to the use of the long axis grain measurement in the thin section size analysis. The relationship between the long axis measurement and the equivalent sieve size (determined by the intermediate and short axes of the grain) is entirely dependent upon the axial ratio in quartz. Since the long axis does not play a part in sieve analysis it should not be used to correlate thin section analysis with sieve size analysis. Smith (1966) has shown that the long and short axes of quartz are highly correlated, thus indicating why Friedman's method can be applied to quartz-rich sands. The long to short axis ratio in labile components is likely to be significantly different and thus to cause a much higher scatter about any regression line.

Other proposed methods for correlating thin section and sieve size analyses have been proposed by Kittleman (1964) and Sahu (1966, 1968). Kittleman proposed using Rosin's distribution to determine size distributions of log-normal grain populations in a sediment. Sahu (1966) developed a theoretical correction procedure using weight frequency moment measures. In 1968 Sahu
went on to develop correction factors based on the probability theory using the Edgeworth series. He considers that the effect of random thin sectioning is most serious towards the finer sizes of the distribution and is negligible in the coarsest sizes since "the observed sizes in the thin section is greater than or equal to the true size of the grain". This last statement cannot hold and is contradicted later in the paper thus leaving doubts about the measure used in the correction factor.

Grain size analysis of the Pernjara and Finke Group sediments is a difficult problem because they are usually too indurated for sieve size analysis. They generally contain a high proportion of matrix, are poorly sorted, commonly consisting of two or three log-normal populations, and are dominantly litharenites and sublitharenites. Thus thin section grain size analysis by either Friedman's (1958) or Packham's (1955) methods is unlikely to yield consistent and accurate results. The poorly sorted nature of the sediments suggests that Packham's method might be the most appropriate but it was found to be less accurate than the method developed during this study. The latter is analogous to Friedman's method but uses correction factors based on the short axis measurement. The study of the relationship between thin section analysis and sieve analysis was conducted because the previously suggested conversion methods cannot be applied directly to the sediments. Also Rosenfeld, Jackson and Ferm (1953) stated that corrections for grain size
conversion from thin section to sieve parameters is probably of little use unless specifically determined for the rocks to which they apply.

Theoretical Considerations

Methods which utilize a number frequency analysis are not suitable for comparison with the weight frequency distributions obtained from sieve size analysis. Thin section point counting gives a volume percent not a number percent distribution (Stauffer, 1966). A thin section is assumed to give an unbiased sample of a volume of rock and if a random array of points is examined on this surface every point gives an equally weighted random sample of the area and hence volume of the constituent in the rock. Volume percentages are equal to weight percentages when the effect of density variation is very small. Thus a point count method is admirably suited for comparing thin section and sieve size distributions.

The selection of which grain axis to measure has also been the subject of some debate. Packham (1955) used the short axis since this gives a direct comparison with the axis controlling sieve sizes. Friedman (1958) used a long axis definition. The definition of a short axis measurement should follow Griffiths (1966) who defined it as the short axis of the minimum circumscribed rectangle. This definition is the two dimensional analogue of the smallest circumscribed cylinder, about a grain,
which controls the grain movement in sieve size analysis. It is realized that this equivalence does not strictly hold for tabular or platy grains but a long axis definition is of still less value.

Connor and Ferm (1966) produced further evidence in favour of using the short axis measurement. They found that although the short axis is not significantly better than the long axis (at the 95 percent confidence level) as a measure of grain volume it accounts for 36 percent of the variation in grain volume compared with only 24 percent for long axis measurements.

The number of points to be counted has been estimated by various workers and ranges from 50 to 500 (Friedman, 1958). Friedman counted 500 per thin section but noted that in many samples fewer grains gave similar results. Rosenfeld, Jackson and Ferm (1953) noted that 100 to 200 counts are necessary for each analysis to satisfy 95 percent confidence limits of the mean. Lumsden and Pelletier (1969) found that a count of 100 grains established a histogram which did not vary in shape when more grains were counted. However, they still counted 200 to 300 grains per analysis. A similar result was found in this study with constant histograms being produced after 50 to 100 grains depending on the degree of sorting of the sediment. It was decided to limit the number of grains counted in this study to 100 to 200 grains with larger counts for poorly sorted and coarser...
grained sediments. This was because the time required for more accurate analyses of the 450 samples would be excessive since the results obtained would still be unsuitable for comparison with statistical grain size parameters derived from unconsolidated sediments (Chappell, 1967). However, it was hoped that they may be useful for internal comparisons.

**Determination of Sieve Size Parameters from Thin Section Data**

As noted previously many attempts have been made to convert thin section size analyses to the equivalent sieve size distributions. Yet another method was attempted in this study in which it was hoped that a combination of the best features of Packham's (1955) and Friedman's (1958) methods (as discussed earlier) would yield promising results. The method followed in this study, therefore, uses the analytical procedure proposed by Friedman and the short axis measurement of grain size.

The relationship of the conversion parameters calculated by this method to those derived by Friedman (1958) are discussed, and the results are compared to those obtained by using Packham's (1955) method modified after Vistelius (1958).

The main limitation in carrying out such a study on rocks of the Pertnjara and Finke Groups is that few
samples are sufficiently friable and unweathered to enable accurate disaggregation for sieve analysis to be carried out. The eleven samples discussed in the section on sieve analysis fulfilled these conditions and also provide a fairly representative collection of the rock types encountered in these two Groups.

Sieve Size Analysis

Eleven samples were selected for sieve analysis on the basis of minimal weathering (determined from thin section) and maximum friability of the sample. Approximately 20 to 30 gm of each sample were immersed in distilled water and subjected to vacuum treatment until bubbles stopped appearing. They were then crushed by hand, with no grinding, to minimize grain damage especially to lithic fragments. Ultrasonic irradiation helped in the disaggregation and the processes were continued until less than 1 to 2 percent of aggregates were visible when the dispersed sample was studied under a binocular microscope.

Each sample was then washed through a 44 micron sieve using distilled water and the minus 44 micron fraction collected, evaporated to dryness and weighed. The coarser fraction was dried with acetone and sieved using a set of standard \( \frac{1}{4} \) phi interval sieves. Each fraction was corrected for aggregates where present although the percentage in any one sieve was usually
very low. Cumulative percentage graphs were plotted on both log-arithmetic and log-probability paper (Figs 4.48a-k). Most samples plot out as two nearly normal distributions of medium and very fine grained populations with the graphs showing the change from one population to the other at about 3 to 3.5 phi. Several samples also show a separate coarser grained population.

Analysis of the minus 4.25 phi material was not attempted because diagenesis rapidly affects this range and the chances of obtaining unaltered clay and silt minerals, with the same size distribution as they had when deposited, is very remote. Although the samples had minimal apparent diagenesis most of them showed some alteration and redeposition in the clay size range. Thus on log-arithmetic graphs the grain size of the minus 4.25 fractions were only estimated by extrapolating the curve to 100 percent at 14 phi.

Thin Section Size Analysis

A grain size analysis was determined for all thin sections studied. The short axis size of every second to fourth grain (consistent within any one sample) was measured using a micrometer eyepiece during a modal point count analysis. The spacing interval was determined before the analysis was started and depended on the degree of sorting of the sediment. Thus for poorly sorted sands every second grain encountered was measured
while well sorted sands only required every fourth grain to be measured. A total of between 100 and 200 grains were measured in every sample giving an overall total of more than 65,000 grains measured during this study. The grains were grouped into 1/86 mm classes (equivalent to one micrometer division at 80x magnification) from the finest measured class (3/86 mm) to the coarsest grain present in the analysis. The coarsest grain present in the sample, if not already measured, was also noted to determine the one percentile as accurately as possible. All grains finer than 3/86 mm (4.8 phi) were included in the detrital matrix and their size was not measured. Grain sizes below 4.8 phi can only be determined by extrapolation of the grain size distribution to 100 percent at 0 mm.

The average size of the grains measured in each sample was calculated in terms of 1/86 class divisions using the computer programme given in Appendix 8a. This size was then used to determine the effect of size variation in the modal analysis of the sample and after correction factors were determined all modal analyses were corrected to 2.5 phi.

The percentiles and graphic grain size parameters were determined by straight line interpolation and extrapolation using the computer programme given in Appendix 8. In addition to the frequency distribution in the grain size classes this programme also required the percentage
of matrix constituents expressed as a percentage of matrix plus framework grains. Provision is made in the programme to recalculate the percentage of matrix if voids and void filling cements have been determined. Both the latter are excluded because they would affect the volume frequency distribution in the thin section analysis yet do not affect the weight frequency distribution in sieve analyses.

The results of the unmodified thin section analyses are given in Table 4.12. It must be realized that these results are subject to various errors both in measurement and assumption. The percentile values may have standard errors of up to 5 to 10 percent of their value and the percentage error becomes higher in coarser grained samples where the phi value approaches zero. The main sources of error would be due to count length in both the number of grains measured and the number of points counted in the modal analysis. Both these errors could be reduced by longer counts over larger thin section areas (Bayly, 1960) but the time required for this would have been excessive. Errors are also introduced by using straight line interpolation between class divisions. However, Figures 4.48a-k show that the grain size distributions determined in this analysis have data points sufficiently close for this source of error to be minor in comparison with the counting errors. Other lesser errors include error in measurement of the "b" axis length, and operator error. Standard errors of the
Even though the errors in determining grain size distributions from thin sections are high the results should show some internal consistency since all values will be subjected to similar errors. Thus both the percentile measurements and grain size parameters will show up regional patterns, if present, although the interpretation of these patterns is open to question and will be discussed later.

Regression Equations

Thin section quartile measurements (in phi units), obtained from the unmodified grain size analysis programme, were plotted against their equivalent quartile values obtained from the sieve analysis (Fig.4.49). All the points, except one quartile value, with a thin section size finer than the finest class interval measured, fall with a reasonable degree of scatter about a straight line. A regression line for the quartile measure was obtained using the computer programme given in Appendix 9 and it showed an 87 percent correlation between thin section grain size and sieve grain size. The regression equation is:

\[ Y(\text{sieve}) = -0.3067 + 0.9481 \times X(\text{thin section}) \]

In this study it was decided that, since it was hoped to be able to find an overall regression line to
enable any percentile value to be converted to its equivalent sieve size value, then at least all the percentile measurements used in the graphic interpretation should be used to produce a regression equation. This is one of the discrepancies of Friedman's method where he produced a regression equation to correct the central fifty percent of the grain size distribution and he then used this equation to convert the whole spectrum of a grain size analysis to sieve size equivalents using a specially constructed graph paper.

It was apparent from a plot of all the percentile measurements used in this study (namely the 1, 5, 16, 25, 50, 75, 84, and 95 percentiles, Fig.4.50) that a single regression equation could not fit the data, when assuming a straight line extrapolation to 100 percent at 14 phi for the unanalysed fine detritus. There is a break in slope at about 4 phi with grains finer than 4 phi showing a very wide degree of scatter but with a general linear trend. Such points should probably be discarded because their original extrapolated values are based on assumption rather than fact. However, it was decided to proceed with a full analysis to determine the best method of representing the conversion factors. An overall regression equation was determined for the data:

\[ Y_{(\text{sieve})} = -2.8500 + 1.9104 \, X_{(\text{thin section})} \]

This equation accounts for 88 percent of the correlation between the sieve and thin section results. However, by inspection, the fit is not very good over the percentile
range from 1 to 75 and it is too general to provide an overall best fit. It was, therefore, decided to find two equations to fit the data. The first equation was determined for all the percentile values coarser than 4.8 phi (the smallest value actually measured). The regression equation for these points is:

\[ Y_{\text{sieve}} = -0.5742 + 1.0260 \times \text{ (thin section)} \]

and this equation accounts for 92 percent of the correlation between thin section and sieve sizes thus indicating a better fit for this data than was obtained with the overall regression equation. Figure 4.51 shows the actual variation in slope between the estimated regression lines for each sample and the derived regression equation for combining these samples. The variation in slope is large for some samples (e.g. FG 15) but most of the remaining samples show reasonable agreement of slope even though intercepts may differ. This indicates that the correct curves for most samples should parallel the sieve curves even if they depart from them slightly in value. The discrepancy of slope in sample FG 15 can be partly attributed to the non-log-normality of the coarsest fractions in the sample.

The regression equation for percentile values with grain sizes finer than 4.8 phi was also derived although, as it has been noted, very little reliability can be placed on it. The values it incorporates consist largely of 84 and 95 percentiles and thus parameters using these measures must also be suspect i.e. all the
The regression equation for these values is:

\[ Y_{(sieve)} = -11.1944 + 3.3860 \times (\text{thin section}) \]

and this equation only indicates a 44 percent correlation between thin section and sieve grain sizes in this region.

The equation derived for correcting quartile measures can be compared with the equation derived for all the percentile measures with thin section values greater than 4.8 phi to see whether these equations are comparable forms for describing the correction factors i.e. whether they differ significantly from one another. The percentile equation may be considered as fixed since it is calculated from a larger data set and the quartile equation may be compared with it.

\[ Y = -0.5742 + 1.0260 \times (\text{percentiles}) \]
\[ Y = -0.3067 + 0.9481 \times (\text{quartiles}) \]

Measures for the quartile equation are:

\[ \sum_{i=1}^{N} x_i^2 = 272.8677 \]
\[ S_x^2 = 0.3435 \quad \text{(variance of X)} \]
\[ S_y^2 = 0.3533 \quad \text{(variance of Y)} \]
\[ N = 30 \quad \text{(number of data points)} \]
\[ b = 0.9481 \quad \text{(slope of quartile equation)} \]
\[ S^2 = \frac{(N-1) (S_y^2 - b^2 S_x^2)}{(N-2)} \quad \text{(overall variance)} \]
\[ = 0.1119 \]
These can be tested in the following manner for a normal area distribution with $N(0,1)$ i.e.

$$|Z_1| = \frac{(a-\alpha).S_x.N.(N-1)}{\sqrt{\sum_{i=1}^{N} X_i^2}}$$  
(tests intercept)

$$= 0.8366$$

$$|Z_2| = \frac{(b-\beta).S_x.(N-1)}{S}$$  
(tests slope)

$$= 0.7347$$

$$P(1.96 < |Z| < \infty) = 0.05$$

Thus, since the tests for both intercept and slope are not significant at the 95 percent confidence level, it can be assumed that there is no significant difference between the regression equations and that both equations are of equal merit. Thus Friedman's (1958) assumption of an overall regression factor is in fact validated in the limited number of samples studied.

The regression equation for all percentile points can be tested against the 1:1 line to test whether any correction is warranted. The same test is used to the one above with the equations:

$$Y = X$$

$$Y = -0.5742 + 1.0260 X$$

The relevant statistics are then:

$$|Z_1| = 5.538$$  
(tests intercept)

$$|Z_2| = 0.684$$  
(tests slope)

$$P(1.96 < |Z| < \infty) = 0.05$$
Thus at the ninety-five percent confidence level the slope of the regression equation does not differ significantly from the slope of the 1:1 line while the intercept is significantly different. This implies that a constant correction factor of approximately half a phi unit can be applied throughout this study independent of the grain size under consideration at least in the range from 0 to 4 phi.

To determine the relationship between Friedman's (1958) regression equation and the equations determined in this study an intermediate step has to be calculated. This involves determining the relationship between the long and the short axes of the grains in these sediments and then using this relationship to test the two forms of correction to equivalent sieve size parameters.

Relationship between Length and Breadth of Grains

Smith (1966) found that there was a high degree of correlation ($r^2 = 0.984$, or 98%) between the measured "b" or short axis length of quartz grains in thin section and the "a" or long axis measurement of the same grain. Smith studied 13 samples, measured a total of 520 grains and derived a regression equation for the average values for each sample.

i.e.

$$X = 0.514 + 0.972Z$$

where $X$ equals the "b" axis length and $Z$ equals the "a"
The validity and general application of this regression equation was tested in the present study of measuring the length and breadth of 336 grains, both quartz and rock fragments, on point count traverses of 6 thin sections. The results were not averaged for each sample but were treated as a single population. Each grain was plotted on a graph and a reasonable linear correlation is apparent in Figure 4.52. A regression equation was determined for all these points using the same notation as above:

\[ X = 0.660 + 0.9365Z \]

This equation shows that the "a" and "b" axis lengths are 79 percent correlated in this study. The reduction in correlation is probably due to the measurement of the non-quartz grains which have differing elongation ratios.

With a sample size of 336 a normal distribution can be assumed in determining whether the present equation and the one derived by Smith are significantly different. The tests involve the same standard statistics as those used above with the number of degrees of freedom being \( N-2 = 334 \). The resultant values for \( Z \) are:

\[ |Z_1| = 2.497 \quad \text{(tests intercept)} \]
\[ |Z_2| = 1.558 \quad \text{(tests slope)} \]
\[ P(1.96 < |Z| < \infty) = 0.05 \]
\[ P(2.576 < |Z| < \infty) = 0.01 \]

Thus in the present study the slope of the derived
The value of the intercept depends on the degree of roundness of the grains and illustrates that, for the samples studied, the quartz in the Gatesburg Formation is more rounded than the grains in the Pertnjara Group. However, one important conclusion to be reached from the almost constant slope of the regression equations is that the ratio of grain length to grain breadth is proportional during the process of rounding and that the long axis is worn down faster than the short axis. This effect would be expected from either corrosion or the effect of mode of transport on abrasion of grains. Thus for the long axis to be reduced fastest implies that the main mode of transport for the sand sized grains is by saltation rather than rolling since most abrasion from rolling or sliding would be expected on the short and intermediate grain axes. The conclusions reached from this brief study are not necessarily true in general and a far more intensive study of grain measurement and roundness would be necessary to validate them.

Comparison with Friedman's Equation

The inter-relation of Friedman's (1958) overall
regression equation, for correcting grain size for errors encountered in thin section grain size analysis, and the percentile regression equation calculated during the present study can be determined statistically. Friedman's equation expressed in phi units is:

\[ Y = 0.381492 + 0.9027Z \]

where \( Y \) is the equivalent sieve grainsize for the measured long axis grain length, \( Z \). This equation has been determined with a sufficiently high degree of accuracy to enable us to consider that the intercept and slope values given in the regression equation are, to all intents and purposes, reliable estimates of the population coefficients for this regression. Likewise, the relationship between the lengths and breadth of the grains in the present study is known from a relatively large population whose distribution can be considered as normal about the regression equation

\[ X = 0.660 + 0.9365Z \]

In determining the relationship between Friedman's equation and those derived in the present study the above equations can be considered as fixed and the relationship can be tested much more simply by just considering the variance of the derived regression equations.

In the first case the regression equation for all the percentile measures was tested against Friedman's equation using the following statistics where \( Y \) equals sieve grain size, \( X \) and \( Z \) represent the short and long thin section axes respectively, and the variance notation
is as above for the percentile regression equation.

(1) \[ Y = 0.3815 + 0.9027Z \] (Friedman)

(2) \[ Y = -0.5742 + 1.0260X \] (percentile regression)

(3) \[ X = 0.6601 + 0.9365Z \] (length-breadth regression)

(2) + (3) = (4) \[ Y = 0.10296 + 0.96085Z \]

\[
S^2_c = s^2 \frac{\sum X_i^2}{N(N-1)s^2_x}
\]

\[
S^2_d = \frac{s^2}{(N-1)s^2_x}
\]

When \( \alpha, a + a_1 \) are the intercepts and \( \beta, b + b_1 \) are the slopes of equations (1), (3) + (4) respectively:

\[
|Z_1| = \frac{\alpha - a_1}{\sqrt{S^2_c + a^2 S^2_d}}
\]

\( = 2.049 \)

\[
|Z_2| = \frac{\beta - b_1}{b \cdot S_d}
\]

\( = 1.633 \)

\[ P(1.96 < |Z| < \infty) = 0.05 \]

\[ P(2.054 < |Z| < \infty) = 0.04 \]

These results indicate that while the slope of the percentile regression equation does not differ significantly from that of Friedman's, the significance of the intercept differences between the two equations is marginal. Thus at the 95 percent confidence level the intercepts are significantly different while at the
96 percent confidence level they are not.

A similar test between the quartile regression equation derived in this study and the quartile regression equation of Friedman yielded the following results:

(1) \[ Y = 0.3815 + 0.9027Z \]  (Friedman)

(2) \[ Y = -0.3067 + 0.9481X \]  (quartile regression)

(3) \[ X = 0.6601 + 0.9365Z \]  (length-breadth regression)

(2)+(3) = (4) \[ Y = 0.319 + 0.8878Z \]

\[ |Z_1| = 0.2974 \]  (tests intercept)

\[ |Z_2| = 0.2189 \]  (tests slope)

\[ P(1.96 < |Z| < \infty) = 0.05 \]

Hence, there is certainly no significant difference at the 95 percent level between the quartile regression equations derived by Friedman (1958) and during the present study. It can be assumed that while both methods have their advantages and disadvantages there is no significant difference between the results they will produce at least for rounded, moderately sorted sediments. The present method, when further validated, may prove to be more useful where dealing with lithic and/or non-rounded sediments since both these factors limit the degree of correlation between length and breadth of grains and thus induce errors in Friedman's basic assumption that the long and short axes always have the same relationship to each other.
Grain Size Correlation

The correlation factors for all percentile points, derived in the preceding section, were used to correct all thin section grain size distributions measured during the current study. For all thin section percentile values coarser than 4.5 phi the correction \( Y = -0.5742 + 1.0260 \times \) was used and for all measurements finer than 4.5 phi the correction \( Y = -11.1944 + 3.3860 \times \) was used. The results of the corrected grain size analyses are given in Table 4.13.

The first object is to see whether these grain size parameters provide better estimates of the sieve size distribution than the uncorrected data, and to determine whether this method is better than Packham's method.

Tests for Goodness of Fit

Numerous statistical tests were carried out to determine whether the measurements corrected by the method developed in this study gave a significantly better approximation to the equivalent sieve size distribution than the uncorrected thin section measurements or the measurements corrected by Packham's method.

None of the goodness of fit statistics showed any significantly better method to use for either the
quartile measures alone or all the percentile measures (Table 4.14). The corrected thin section values and the values obtained by Packham's method are only significantly better than the uncorrected values when a T-test is applied, and not for any of the other tests. However, it can be noted that with other tests they are always significant at a higher level than the corresponding uncorrected values. None of the phi squared values for individual percentiles were significantly different from a normal distribution.

In comparing the actual sample parameters in terms of mean deviation, variance and standard deviation it will be noted that for the quartile measures both Packham's method and the corrected thin section values are distinctly less than the corresponding uncorrected thin section value. This indicates that the latter method incorporates larger errors than either of the other methods and is, therefore, a less reliable measure of the sieve size distribution. The relative merit of Packham's method and the method developed in this study cannot be determined. When all the percentile values are considered a similar pattern emerges with the method developed in this study being superior to both Packham's method and the uncorrected thin section data. However, while Packham's method gives consistently higher variation from the sieve size data than the present method this is only in relative terms and cannot be upheld statistically.
Discussion

From this study one can conclude that the method developed for correcting thin section grain size distributions to sieve size distributions is analogous to and is not significantly better than Friedman's method or Parkham's method. However, it may be argued that the present method is more likely to yield reasonable correction factors in labile and angular sediments than Friedman's method since it does not incorporate a length-breadth assumption. With respect to Packham's method, the method developed in this study has the advantage of ease of calculation coupled with a more accurate correction for the upper and lower quarters of the grain size distribution.

Limitations in applying this method of grain size analysis is that sediments with percentile measurements greater than 0 phi have not been studied. The extrapolation of the regression equation into this region may yield inaccurate results because the number of grains studied in any one thin section will be less likely to be representative of the population and the effect of random thin sectioning will produce results on the low side of the true value. Likewise in the very fine material, thin section grain size analysis is virtually impossible and a consistent method of extrapolation should be employed.

The necessity for attempting to correct thin
section grain size distributions may also be questioned at this stage. Would it not be better to accept the fact that grain size corrections will probably never be sufficiently accurate to enable environmentally significant grain size parameters to be calculated? Even if this goal were reached, Chappell (1967) has noted that the effect of lithification decreases the resolving power of grain size parameters and that the general application of such parameters to consolidated rocks is problematical and uncertain. The value of the grain size parameters is further reduced in consolidated rocks by the significantly less accurate grain size analysis that can be obtained due to diagenesis of fine material, breakdown of lithic and sublabile fragments and the necessity for thin section grain size analyses. Thus the prime object of attempting to correct the grain size distribution is lost.

Hence, the statistical parameters derived from grain size analyses in this study are probably meaningless for detailed environmental discrimination when compared with modern values from known environments. However, they may be validly used for internal comparisons of samples from the Pertnjara and Finke Groups which have been analysed by the methods outlined.

It may well prove to be far more practicable not to correct thin section grain size analyses. It is known that uncorrected grain size analyses give meaningful patterns for internal comparison within a region. If a
single method of obtaining thin section grain size distributions was widely used and accepted, parameters based on such a distribution would probably also have environmental significance at least to a general degree.

Klovan (1966) outlined a potentially very useful method for handling grain size data. He utilized the full grain size analysis results without calculating any parameters and fed it through a Q-mode factor analysis programme of an analogous type to Appendix 10 to obtain orthogonally rotated factor axes and their loadings. From their factors he was able to define varying depositional energy levels without a prior knowledge of where the sediment came from.

This concept is very good but its application to ancient environments needs detailed studies which could not be incorporated into this work. There are two obvious drawbacks. One is the grain size analysis itself - if the rock can be disaggregated will the fine detritus still represent the original material or has it been diagenetically altered; if the rock is indurated can similar results be obtained from thin section grain size analysis? The second drawback is in the range of grain sizes that can be studied - very rarely can one obtain an accurate analysis of fine grained material from consolidated rocks or even semiconsolidated rocks. Thus to be generally applicable samples from known ancient environments need to be studied by a uniform
thin section method using closer phi intervals in the sand and coarse silt range to determine whether depositional or environmental analysis is practical utilizing factor analysis.
Grain Size Parameters

A summary of the relevant graphical grain size parameters from Tables 4.12 and 4.13 is given in Table 4.15. Values and ranges are given for both corrected and uncorrected thin section grain size distributions and they indicate the overall variability of these parameters. In general ranges for the corrected size distributions show a greater variability than the equivalent uncorrected parameters. This is especially obvious with the median and mean grain sizes which in the corrected samples range from very coarse sand to very fine silt. However, the majority of samples analysed have median and mean grain sizes within one phi unit of the average.

Graphic and inclusive graphic standard deviations both have a similar range of values with the variability again greater in the corrected samples than in the uncorrected ones. All values are relatively high and they reflect the bimodal nature of the sediment with the standard deviation increasing as the proportion of the fine mode increases. On the verbal scale of Folk (1968) 3.7 percent of the corrected samples fall in the moderately well sorted class, 3.5 percent in the moderately sorted class, 47.8 percent in the poorly sorted class, 41.2 percent in the very poorly sorted class and 3.8 percent in the extremely poorly sorted class.
The skewness values of the sediments are characteristically positive with only a few near symmetrically skewed distributions (0.9%) representing well sorted reworked older sands or aeolian deposits. A further 3.5 percent of the corrected samples are fine skewed while strongly fine skewed samples represent the bulk of the sediments studied (95.6%). Kurtosis values are also predominantly strongly positive due to the presence of abundant silt and clay in the sediments. Only 3.7 percent of the samples are mesokurtic while 10.8 percent are leptokurtic, 31.3 percent are very leptokurtic and 54.2 percent are extremely leptokurtic. The combination of fine skewed and very leptokurtic parameters together with high standard deviation values points strongly to a fluvial mode of origin with very rapid sedimentation and little or no reworking at the site of deposition. Friedman (1967) noted that these characteristics of fluvial sediments are generated by a three fold mode of deposition. The coarse grains are deposited by rolling or sliding, the bulk of the sand sized sediment is deposited by saltation and the fine fractions are deposited from suspension by becoming trapped between coarse grains. Alternatively fine detritus may be deposited along with saltating grains when the current wanes and the suspended load becomes oversaturated. The latter cause is quite common on sandy fluvial systems where stream water rapidly percolates into the underlying sediment, with a consequent reduction in volume and velocity of the stream.
As noted previously the significance of the calculated grain size parameters is uncertain. However, the general uniformity of relationships between the parameters points to a fairly uniform depositional environment over quite a long period of time.

The points of interest are that there is a general decrease in the standard deviation with decrease in mean grain size. This indicates that all the sediments are probably bimodal with a sand sized mode and a silt sized mode and they only became well sorted in almost pure siltstones. Exceptions to this general rule include sandstones reworked from the Mereenie Sandstone (eg. AJ3) and some of the aeolian sandstones (eg. D013). The slope of the regression line on a graph of mean grain size versus standard deviation is relatively low thus indicating that the source area provides numerous indistinct modal contributions which become mixed but unsorted in the depositional environment (Folk & Ward, 1957). Plots of mean size versus skewness show a general decrease in skewness with increase in size, although the relationship is not very clear cut. In most samples there is variable mixing of sand and silt modes thus producing a variable positive skewness for most mean sizes studied. The relationship between mean size and kurtosis is likewise variable with a few mesokurtic samples in a leptokurtic to extremely leptokurtic population. High values of kurtosis indicate that the sorting of the central portion of the distribution
remains relatively good while the tails of the distribution are poorly sorted and probably represent accumulation of detritus from different modal populations ie. many of the sediments are bimodal and several of the coarser ones tend to be trimodal. These features probably represent both the detritus available in the source area and the variable competence of the depositional current. The good sorting in the sand size range probably reflects detritus reworked from earlier moderate to well sorted marine sands.

Plots of standard deviation versus skewness or kurtosis show a rather random scatter in the positively skewed and high kurtosis regions which is of very little diagnostic value. It merely indicates the mixing of sand and silt modes with poor sorting. Groups of skewness versus kurtosis show almost all points in the positively skewed leptokurtic field thus indicating the predominance of one size mode over another - in this case a dominant sand mode and a subdominant silt mode. The few sediments with near normal skewness also have near normal kurtosis values and they represent clean well sorted sands - either reworked Mereenie Sandstone (eg. AJ3, FJ20) or aeolian sands (eg. D013, FG21A).

All the samples studied have very similar grain size parameters and are characterized by poor sorting of the various modal contributions. These characteristics are typical of fluvial deposits (Koldijk, 1968) although they are not diagnostic of fluvial deposition. The
distinction between fluvial bedload and suspended load deposits and dune deposits of reworked fluvial sand could not be determined. However, the characteristics of the sediments studied also fit fluvial dune deposits using the criteria of Hails and Hoyt (1969).

C/M plots (after Passega, 1957, 1964) were produced for all groups of samples for both the uncorrected thin section size measurements and the corrected measurements. Both sets of data yielded essentially similar distributions (eg. Fig.4.53), which have most of the elements of tractive current deposits, although it must be realized that few sections were studied from siltstones and very coarse sandstones. Thus all the graphs are dominated by sediments typifying graded suspension deposits of a tractive current system and only a few samples represent the coarser and finer extremes.

Corrected size distribution C/M plots of sediments from Glen Helen, together with a few other coarse samples from the Ljiltera Member and the Brewer Conglomerate, are given in Figure 4.53a. The sample from the Dare Siltstone Member (AJ105) plots as a solitary point in the area of suspension deposits, indicating relatively good sorting confined to the silt and clay modes. The majority of samples from the Harajica Sandstone Member and the Hermannsburg Sandstone plot out in a group aligned parallel to the C = M line in the area assigned to deposition from graded suspension. Coarser samples,
largely from the Ljiltera Member and Brewer Conglomerate, complete the tractive current deposit pattern with a predominant sand mode but containing coarser grains introduced into the system by tractive rolling. Although this pattern was produced by selecting samples from a wide stratigraphic interval it is considered to be fairly representative of patterns for subunits since throughout the column the sedimentation characteristics are fairly constant and consist of fining upwards cycles (see Chapter 3).

The sample localities AA to AU, BA to BU, CB, to CY, EH to EV, and GA to GF produce essentially similar C/M patterns to that shown by the Glen Helen samples (eg. Fig. 4.54). Very few siltstones were sampled from these localities and hence the absence of suspension deposits from the patterns is not a genetic feature but one caused by sampling bias. The scarcity of coarse grained samples is also partly a feature of sampling bias although such rocks are rare at the C, E and G sample localities. The patterns for each of these sets of sample localities are, therefore, dominated by groups parallel to the C = M line representing deposits from graded suspension.

Samples from the D sample localities show additional features to those described above. The C/M pattern (Fig. 4.55) shows the development of two groups. The group within the area characterized by deposition
from suspension, includes all the moderately sorted, non-sandy siltstones from the Deering and Dare Siltstone Members. The samples represent a fairly random stratigraphic collection and they indicate that the depositional processes must have remained fairly constant. Most of the sandstone samples once again plot out as a group parallel to the C = M line. A notable feature is the greater spread of samples above the C = M line indicating that some of the samples have quite large tails of silt and clay sized detritus. Most of these poorly sorted samples are closely associated with siltstone horizons in outcrop and they reflect a degree of mixing of the two size modes thus implying variable and incompetent depositional currents.

Samples from the Finke Group (Fig.4.56) once again show the three main sections typical of traction current deposits but, as in samples from D localities, some of the sediments are very poorly sorted and plot well above the C = M line. Once again this indicates the mixing of two size modes and the beds in question are associated either with siltstone or conglomeratic horizons.

The cumulative percentage curves may also be taken as indicative of depositional processes. The curves presented in Figures 4.48a-k are all basically similar to the grain size distribution of fluvial sediments presented by Visher (1969). Most of the
samples consist of two distinct populations, one related to deposition from saltation and the other to deposition from suspension.

The saltation population is dominant in almost all sandstone samples and in itself is moderately well sorted. The amount of suspension population is usually restricted to less than 20 percent and this indicates relatively high depositional current velocities. The few samples which show a good mixing of the suspension and saltation populations are indicative of highly variable current strengths during deposition. In coarser samples occasional larger grains form a coarse tail to the grain size distribution. This fraction represents a surface creep population moved by sliding or rolling of individual grains. This coarse population may be truncated at the maximum clast size indicating that size is controlled by current strength, or it may be gradational indicating control by provenance supply.

Doeglas (1968) proposed a very simple way of providing grain size information using three or five digit numbers. He classified samples with quartile phi values expressed as integers, and for finer environmental differentiation he also used the 1 and 99
percentile values.

Table 4.16 classifies all the samples studied, according to the quartile measures, and also lists the main environments in which such sediments occur (after Doeglas, 1968). A conspicuous feature is that most of the quartile classifications represented by the Pertnjara and Finke Group sediments are known to contain fluviatile constituents. Of the 33 environmentally classified quartile measures 22 are dominated by fluviatile constituents and 9 of these are characteristic of fluvial sediments only. Four of the remaining classes are dominated by beach or dune sediments and the remainder are characteristic of lagoonal, tidal flat and estuarine sediments. The latter represent fine grained sandstones and siltstones and include many of the samples from the Parke Siltstone and Horseshoe Bend Shale. Samples in the classes 222, 223, 233, 333 which are characteristic of beach and dune sediments can be further differentiated when the 1 and 99 percentiles are considered. The 99 percentile (obtained by extrapolation) usually falls in the 8 to 10+ integer phi range while the one percentile varies from 1 to 2. These additional criteria strongly favour a fluvial or dune origin for the sediments rather than a beach. The overall pattern is, therefore, one of a predominantly fluvial environment of deposition, with possibly some aeolian reworking, for the sandstone deposition and a quiet shallow water lagoonal, tidal flat, or estuarine
depositional environment for the finer grained constituents.

While the grain size analysis has not yielded a very variable or differentiable set of parameters it has at least provided further evidence for the overall uniformity of depositional processes. The samples from all the sandstone units indicate deposition by tractive currents containing suspended load and rolled bedload detritus. The sandstones are all positively skewed and leptokurtic thus indicating a degree of incompetent sorting by the depositional current with the subsequent mixing of a subdominant silt mode with a dominant sand mode. The relatively scarcer conglomerates and conglomeratic sandstones also have the addition of a pebble and boulder mode to the sand and silt modes. This indicates that even the stronger currents were ineffective for sorting the sediment. The only well sorted sediments represent reworked Mereenie Sandstone (eg. AJ3, FJ19) or aeolian deposits (eg. DI60, DO13), although not all the latter have normal skewness and kurtosis parameters. Siltstone samples reflect deposition in a relatively constant quiet water environment not subject to reworking or current action strong enough to move sand sized detritus.

The overall picture is, therefore, one of deposition from a tractive current aqueous media which is relatively inefficient at sorting the material supplied to it from the source area. These are
characteristic features of deposition in a fluvial regime. The siltstones were probably deposited in a relatively quiet water lacustrine environment.
4.5 CLASSIFICATORY ANALYSIS OF PETROGRAPHIC DATA

Classificatory analysis of the petrographic data given earlier in this chapter was undertaken using two techniques - factor analysis and cluster analysis. Computation was performed using either an IBM 360 computer or a CDC 3600 computer. Apart from the programmes given in Appendix 10 the following programmes were also used in various combinations:— MULTCLAS (Lance and Williams, 1967a,b); GOWER (Gower, 1966, 1967); GOWECOR and GROUPER (Lance, Milne and Williams, 1968).

Factor analysis is a technique designed to simplify relationships between large amounts of data. Each variable or sample may be defined by the coordinates of a point in multidimensional space. The line from the graph origin to this point gives a vector representation of the point and the angle between vectors to different points gives an indication of their similarity i.e. 0° identical, 90° totally uncorrelated, 180° perfect inverse relationship, and points in between show varying degrees of correlation. Orthogonal factor axes are selected to provide the best representation of a cluster of vectors and the projection of each vector onto these axes gives the vector loading (Harbaugh & Merriam, 1968) on the respective factor. The loading is arbitrarily defined as positive on one side of the graph origin. The sum of the squared vector loadings for each variable (the communality) indicates the extent to which a factor
defines all the vectors. When the factors are interpreted using a diagnostic programme such as GOWECOR the contribution of each vector to each factor can be obtained. This information is used in turn to define the temporal or spatial relationship between variables or samples and thus provide a simplified means of interpreting the original data.

In cluster analysis all the samples are compared and a similarity coefficient is computed for each pair of samples. The samples are then combined into an undefined number of groups with the samples in each group being more similar to each other than to those in other groups. The similarity between the groups is also determined and a hierarchical dendrogram of decreasing inter-group similarity is produced. These distinct groups and their relationships can then be used as the basis for geological interpretation.

Factor analysis may be considered from two angles. R-mode factor analysis is used to determine the relationship of the variables from a large number of samples. Q-mode factor analysis is used to determine the inter-sample relationship when each sample has a large number of variables - in this case 58 variables.

Q-mode factor analysis is restricted by available computer memory and a maximum of 129 samples can be analysed on the CDC 3600 computer using the programme
MAXGOWER, and 92 samples on the programme GOWER. The Q-mode analysis programme in Appendix 10 can handle over 120 samples each with 60 variables using an IBM 360 computer. R-mode factor analysis using the programme in Appendix 10 can handle large numbers of samples but is restricted to about 120 variables which can be compared simultaneously.

R-Mode Factor Analysis

The spatial relationship of petrographic parameters is an important feature in deducing the petrology of an area. R-mode factor analysis is an aid in this problem and factor loadings on the first third of the vectors usually account for a high proportion of the variability between samples. Thus they provide most of the important relationships between the variables.

R-mode factor analysis of all samples from the Pertnjara and Finke Groups yielded some interesting results, and confirms the general trends shown by the modal analyses. A total of 24 rotated factors were requested (Table 4.18) and these account for 89.1 percent of the variability of the measured parameters defining each sample. Factor loadings greater than ± 0.3 were considered to be significant in this study.

The first factor accounts for 17.4 percent of the variability and shows the relationship between quartz and rock fragments. Straight and undulose extinction quartz
and the associated quartz ternary diagram end members are positively correlated with the mean, median and one percentile grain sizes (in phi units), and with distance. These are all negatively correlated with stratigraphic height and siltstone, sandstone and limestone rock fragments and the associated labile ternary diagram end members. This means that with increasing distance from the northern margin of the Amadeus Basin the grain size of the sediments decreases and the quantity of quartz increases at the expense of rock fragments. It also indicates that with increasing stratigraphic height, labile constituents increase at the expense of quartz. The second factor shows a positive correlation between detrital matrix content, muscovite and decreasing grain size and a negative correlation between these parameters and the quantity of total quartz, sandstone rock fragments and overgrowths. Correlation between distance and height is negligible. This relationship could be expected in any sedimentary suite and merely indicates that the grain size depends to some extent on the amount of matrix present. The positive correlation between opaque minerals, zircon and various types of tourmaline in Factor 3 and their independence of distance and stratigraphic height indicates that they were probably derived from a similar and constant source throughout the period of deposition. Factor 4 shows the positive correlation of quartz overgrowths, glauconite, collophane and unstable accessory minerals and a negative correlation between these minerals and distance from the northern margin and the quantity of
moderately stable heavy minerals. This indicates that the collophane and glauconite content decreases with increasing distance of transport. Factor 5 indicates an increase in feldspar content with distance and this is due to the abundance of feldspar in the Finke Group sediments. There is a slight decrease in feldspar content with increase in stratigraphic height. Factor 6 simply indicates that as the proportion of undulose extinction and semicomposite quartz increases the proportion of straight extinction quartz decreases. Factor 7 indicates the relationship of mean, median and one percentile grain size to the grain size standard deviation, matrix and limestone rock fragments. As the grain size increases the standard deviation of the samples become higher due to the increasing content of matrix. Factor 8 shows that roundness of quartz grains and the number of inclusions in them varies inversely with distance. The decrease in roundness and roundness standard deviation with increasing distance from the northern margin can be at least partly explained in terms of decrease in grain size with distance (Factor 1) since larger quartz grains generally show a higher degree of roundness and sphericity than the smaller grains. The change in the number of inclusions can also be explained in terms of grain size since small grains are generally clearer than large ones. However, it could also be partly due to the preferential weathering and abrasion of the quartz grains with abundant inclusions. Factor 9 indicates that straight extinction quartz decreases with increasing stratigraphic height while the proportion of composite
quartz and less common accessory minerals increases. This relation indicates the uncovering of a metamorphic source during deposition of the higher beds. This is confirmed in Factor 10 by the strong increase in garnet and metamorphic rock fragments with increasing stratigraphic height. Factor 12 indicates a general increase in the quantity of muscovite and biotite, whereas Factor 13 shows a general decrease in rutile, microlitic quartz and quartz overgrowths, with increase in distance. Factor 14 shows a positive correlation between the quantity of limestone rock fragments, carbonate cement and stratigraphic height and a negative correlation between these factors and detrital matrix and distance. This suggests a genetic relationship between carbonate cement and limestone fragments although the relationship could also be due to source area control. Factor 20 also shows the relationship of limestone and chert rock fragments increasing with stratigraphic height thus suggesting the relationship of limestone and chert which is known to exist in the source area. Factor 19 shows an increase in opaque minerals with distance of transport and a concurrent decrease in the quantity of stable heavy minerals especially yellow-brown tourmaline. Factor 24 indicates the positive relationship between kurtosis and the quantity of detrital matrix. Thus as the quantity of silt and clay modes in the sample increases the sample also becomes more leptokurtic.

R-mode factor analyses were also carried out to test the variation of averaged parameters from each
locality in the lower Hermannsburg Sandstone (0-250 m) and in the total Hermannsburg Sandstone. Both analyses included the sandstone members of the Finke Group. In general both analyses are very similar (Tables 4.19 & 4.20), account for 95 percent of the parameter variability and need not be discussed separately. They also show many similarities to the overall R-mode factor analysis of all samples. They will be discussed in terms of the lower Hermannsburg Sandstone.

The first factor once again distinguishes quartz rich and labile rich sediments and shows that the quartz content increases with decreasing grain size and with increasing distance from the northern margin. The second factor shows a positive correlation between muscovite, biotite, matrix, resistant minerals, and the mean, median and one percentile grain size and skewness of the sediments and the distance away from the northern margin. At the same time there is a decrease in the content of stable accessory minerals, very resistant quartz and chert fragments. Thus further from the northern margin the sediments tend to be finer grained and have a higher proportion of micaceous detrital matrix at the expense of quartz and chert fragments. Factor 3 indicates that the quantity of unstable and composite quartz and chert varies inversely with the quantity of very resistant quartz. Factor 4 shows that the unstable accessory minerals glauconite and collophane are negatively correlated with moderately stable accessory
minerals and the distance from the northern margin thus indicating their preferential occurrence in samples from near the source area. Factor 5 shows that the quantity of very resistant quartz varies inversely with the quantity of resistant quartz types and is also negatively correlated with skewness. This suggests that samples with abundant straight extinction quartz are characterized by better sorting and they probably represent reworked, older, clean, well sorted sandstones, such as those from the Larapinta Group and Mereenie Sandstone. Factor 6 shows the positive correlation of opaque minerals, zircon and tourmaline and their decrease in abundance with increasing stratigraphic height and distance from the northern margin. Factor 7 shows an increase in garnet and metamorphic rock fragments with stratigraphic height and distance from the northern margin. Factor 8 shows a direct relationship between detrital matrix content and the grain size standard deviation and the inverse relationship between these parameters and the one percentile and quantity of quartz in the sediment. The positive correlation between rutile, microlitic quartz and carbonate cement content is shown by Factor 9, while Factor 10 shows a positive correlation between matrix content, skewness, kurtosis and distance from the northern margin. Thus the sediments become more fine skewed and leptokurtic with the increase in quantity of silt and clay sized detritus associated with increased distance of transport. Factor 12 just shows the relation between stable and moderately stable accessory mineral suites, and indicates that their proportions are not
related to height or distance. Factor 13 shows the influence of the Finke Group sediments with the positive correlation of garnet, moderately stable accessory minerals, feldspar, carbonate cement and distance from the northern margin. Factor 14 shows that the zircon to tourmaline ratio and skewness of the sediment both increase with distance from the northern margin. Factor 17 shows the negative correlation between quartz mean roundness and the number of inclusions in the quartz grains. This suggests that there are at least two populations of quartz in these sediments and that during the process of rounding, grains with abundant inclusions are preferentially weathered and abraded. Factor 19 indicates that mean roundness of quartz decreases with decreasing grain size and this occurs more at higher stratigraphic levels. The possible genetic relationship between carbonate cement and limestone rock fragments is again shown in Factor 23.

An R-mode factor analysis of all samples from the MacDonnell Ranges on the northern margin of the Amadeus Basin also yielded valuable and interesting results (Table 4.21). Twenty-four factors accounted for 93.7 percent of the variability of the parameters analysed. The first factor accounts for 23.7 percent of the variability and again shows the relationship between quartz-rich and labile-rich sands. The content of straight and undulose extinction quartz (and the associated ternary diagram parameters) is positively correlated with stable accessory minerals, median, mean
and one percentile grain size and negatively correlated with moderately stable labile fragments (especially sandstone and siltstone rock fragments), unstable quartz, standard deviation of the mean grain size, and stratigraphic height. Distance can be neglected in this analysis since all sample localities are essentially equidistant from the presumed source area. This concept is supported by the lack of correlation of the other factors with distance except for Factor 21 where distance has a high loading but does not correlate with anything else. Thus the first factor shows the relationship of increasing quartz content with decreasing grain size and the samples with high proportions of detrital rock fragments are generally coarser grained and become more abundant higher in the stratigraphic column. Factor 2 indicates the positive correlation of biotite, garnet, uncommon accessory minerals, metamorphic rock fragments, and undulose extinction quartz with stratigraphic height and the negative correlation of the parameters with more resistant quartz. Both these factors point to a metamorphic source which became more prominent higher in the stratigraphic column, while the stable quartz which is abundant near the base of the column represents polycyclic quartz derived from the underlying Larapinta sediments. Factor 3 indicates that kurtosis correlates positively with the proportion of matrix and carbonate cement and negatively with the quantity of quartz. This merely means that the kurtosis is dependant on the proportion of silt and clay modes present in the sediment. Factor 4 shows that the sediments become coarser with
increasing stratigraphic height and the coarser sediments have a higher standard deviation and contain higher proportions of undulose extinction quartz, chert and limestone rock fragments and carbonate cement. Once again there is a direct relationship between chert and limestone fragments and carbonate cement suggesting that they were all derived from the same limestone source area. The increase in feldspar content and carbonate cement higher in the stratigraphic column shown in Factor 5 is in accordance with the similar increase in limestone and metamorphic fragments up the column. The fresh nature of much of the feldspar is also indicative of a primary metamorphic origin. Factor 6 simply shows that the stable heavy minerals zircon and tourmaline are positively correlated and that they correlate negatively with moderately stable heavy minerals. The positive correlation, in Factor 7, of unstable and resistant quartz with a number of inclusions suggests a genetic relationship between these parameters. They also correlate positively with rutile content and negatively with resistant types of quartz. Factor 8 shows a positive correlation of matrix and carbonate cement with semicomposite quartz while Factor 9 indicates an increase in muscovite, rutile, glauconite and detrital matrix with stratigraphic height. Factor 10 shows that quartz overgrowths, collophane and unstable accessory minerals, decrease with increasing stratigraphic height and with matrix content. Factor 13 shows an increase in chert and unstable quartz types up the stratigraphic column while Factor 15 shows that quartz overgrowths and mean roundness increase with increasing
grain size and associated decrease in matrix content. This supports the observation that quartz grains are more rounded in coarse sandstones than fine ones and that the former usually contain a higher proportion of voids filled with secondary quartz.

Q-Mode Factor Analysis

Q-mode factor analysis of the total Hermannsburg Sandstone and the Finke Group was undertaken using the sequence of programmes MULTCLAS, GOWER and GOWECOR (Lance, Milne & Williams, 1968). It has yielded some useful information on the patterns of mineral distribution in the Amadeus Basin during the Devonian - Carboniferous period of sedimentation. Plots of positive and negative unrotated factor loadings (Harbaugh & Merriam, 1968) for the first seven factors (Table 4.22) are given in Figures 4.57-63 and lists of the 20 most important distinguishing characteristics for each factor are given in Table 4.23.

Factor 1 (Fig. 4.57) distinguishes the localities into two basic groups on the proportion of labile to non-labile fragments. Negative loadings are largely restricted to the northern margin of the basin and represent samples with higher quantities of labile rock fragments, especially sandstone and siltstone rock fragments, unstable quartz types and unstable accessory minerals. Positive factor loadings to the south and south-east indicate that these samples are finer grained, and contain more straight
extinction quartz, detrital matrix and mica. Factor 2 (Fig. 4.58) distinguishes samples with high quartz and chert contents (negative factor loadings) from those containing non-quartz constituents. The latter may be micas, detrital or carbonate matrix, feldspar or rock fragments and these samples are also characterized by a higher proportion of moderately stable accessory minerals and a higher grain size standard deviation. Factor 3 (Fig. 4.54) provides a subdivision based on accessory minerals and quartz types. Negative factor loadings indicate samples with higher proportions of very resistant quartz, tourmaline, zircon and opaque accessory minerals, and siltstone rock fragments whereas positive loadings indicate samples characterized by undulose extinction, semi-composite and composite quartz types and feldspar.

Factor 4 (Fig. 4.60) has no significant positive loadings and the positive area represents low proportions of rutile, tourmaline, zircon, collophane, glauconite, apatite, carbonate, feldspar, microlitic and composite quartz and quartz with acicular inclusions. Factor 5 (Fig. 4.61) indicates that areas with a positive factor loading are characterized by higher values of grain size skewness and kurtosis together with higher proportions of unstable and resistant quartz types, rutile and detrital matrix. In contrast negative factor loadings indicate a higher proportion of straight extinction quartz associated with an increased roundness of the grains, and a higher proportion of feldspar. Factor 6 (Fig. 4.62) distinguishes positively loaded localities characterized by a higher grain size standard
deviation, mean number of bubbles and sandstone rock fragment content from negatively loaded localities containing a higher proportion of undulose extinction quartz and unstable accessory minerals and a finer one percentile grain size. Factor 7 (Fig. 4.63) has a rather random distribution of positive and negative factor loadings largely dependant upon accessory mineral contents. Areas of positive loading indicate a higher proportion of unstable accessory minerals, carbonate matrix, and quartz with a larger number of inclusions. Negative loadings correspond to samples containing a high proportion of stable and moderately stable accessory minerals such as opaque minerals, zircon, garnet and green tourmaline.

Overall mineralogical patterns can be deduced from a consideration of the combination of the first seven factors. The localities on the northern margin of the basin are characterized by labile rock fragments including sandstone, siltstone, limestone and metamorphic fragments. They contain the highest proportion of unstable accessory minerals (collophane and glauconite), unstable quartz types and carbonate matrix. They are also coarser grained and have higher values of skewness and kurtosis. Moving down the paleoslope the samples become more quartz rich and are dominated by a mineral assemblage of straight extinction quartz associated with the stable accessory minerals tourmaline and zircon. The quartz generally contains more inclusions in the form of bubbles than further north and rock fragments are less important.
although sandstone and chert fragments are both consistent minor constituents. The localities in this group occupy a large area in the central and western Amadeus Basin and the boundary of the group is gradational and transitional in all directions. Further south-east down the paleoslope the mineral assemblage is essentially similar except for slightly increased proportion of undulose extinction, semicomposite and composite quartz types and feldspar. The increased proportion of these minerals can be accounted for by mixing of sediments derived from two or more source areas. This group of localities is most prominent from the Camel Flat Syncline to the north-west Finke Sheet Area. Most of the samples from the Finke Group form a distinctive group in the south-east which is characterized by feldspar. These samples also have high values of skewness and kurtosis. They reflect derivation from a local granitic and metamorphic terrain to the south.

Q-mode factor analysis of the lower Hermannsburg Sandstone and all members of the Finke Group yielded slightly different results from that of the total Hermannsburg Sandstone (Tables 4.24 & 4.25).

Factor 1 (Fig. 4.64) shows an intricate pattern of areal interdigitation of positive and negative factor loadings due to variation of matrix and labile components. Positive factor loadings are correlated with high proportions of matrix, quartz and micaceous minerals and an associated decrease in grain size and increase in grain
size standard deviation. Negative factor loadings are associated with higher proportions of labile rock fragments especially sandstone and siltstone fragments. Unstable quartz types and quartz overgrowths are also more abundant in areas with a negative factor loading. Factor 1 clearly shows the decreasing abundance of rock fragments southwards and south-eastwards from the northern margin of the basin. Factor 2 (Fig. 4.65) shows the area dominated by quartz and associated stable accessory minerals as a negative factor loading. Quartz poor sediments are only abundant near the northern and south-eastern margins of the basin but the relative importance of labile and feldspathic constituents in these sediments cannot be ascertained from Factor 2. The negatively loaded northern area depicted by Factor 3 (Fig. 4.60) is characterized by very resistant quartz types, siltstone rock fragments, tourmaline, zircon and opaque accessory minerals, labile fragments and a coarser grain size. Further south down the paleoslope less stable quartz types become more abundant e.g. composite, semicomposite and undulose extinction quartz types. Rare accessory minerals also become more frequent southwards as do metamorphic and chert rock fragments. The increase of unstable quartz types and associated unstable minerals is unusual. It could indicate the mixing of the sediments from two source areas, it could indicate southward stratigraphic thinning of the Hermannsburg Sandstone and thus the inclusion of stratigraphically higher horizons further south, or it probably represents a combination of both these causes.
The positive factor loadings of Factor 4 (Fig. 4.67) delineate areas with a moderately stable mineral assemblage including very resistant quartz, moderately stable accessory minerals, sandstone rock fragments and voids. The presence of more common voids in this area indicates moderate sorting of the sediments and the lack of matrix cement. Further north and east the mineral assemblage is more unstable with unstable accessory minerals such as glauconite and collophane being important on the northern margin of the basin.

Less stable quartz types and feldspar are important in the east in areas of negative factor loading. Likewise rutile, carbonate matrix, and yellow and blue tourmaline are also more common in negatively loaded areas and the grain size is generally coarser. Factor 5 (Fig. 4.68) shows a rather random distribution of positive and negative factor loadings with the former being more common in the north and the latter in the south. Positive factor loadings are correlated with higher proportions of feldspar, quartz roundness, straight extinction quartz and unstable accessory minerals. Negative factor loadings denote localities with finer grained sediments, and higher values of skewness and kurtosis. They also contain higher proportions of detrital matrix, resistant quartz types, stable heavy minerals and limestone and chert rock fragments. The distribution of minerals shown by Factor 5 does not appear to be directly related to paleocurrent trends. Negative factor loadings in Factor 6 (Fig. 4.69) denote finer grained areas containing higher proportions of undulose extinction quartz and unstable accessory minerals. Such localities
dominate in the centre and south of the basin. Positive factor loadings are correlated with accessory mineral suites of zircon, tourmaline, opaque minerals and uncommon accessory minerals. These localities also contain a higher proportion of composite quartz and quartz with inclusions especially acicular needles. The grain size standard deviation is also higher in positively loaded areas.

Negative factor loadings on the seventh factor (Fig. 4.70) are not very clearly related and show that these localities generally contain quartz with more inclusions and a higher standard deviation of inclusions. They also contain a higher proportion of quartz and carbonate cements, microlitic quartz and biotite. Localities with positive factor loadings contain a higher proportion of garnet, green and blue tourmaline, zircon, opaque minerals, apatite, feldspar and metamorphic and chert rock fragments. They also have a higher zircon to tourmaline ratio, and a higher grain size skewness and kurtosis.

When only the lower Hermannsburg Sandstone and Finke Group are studied the overall mineralogical variation across the basin does not differ significantly from the pattern derived for the total Hermannsburg Sandstone. The northern margin is again characterized by a labile mineral assemblage including various rock fragments, semicomposite and composite quartz, unstable accessory minerals and quartz and carbonate cements. The grain size is generally coarser and the skewness and kurtosis values higher than in the remainder of the basin. Down the paleoslope the mineral
content varies with unstable rock fragments and accessory minerals rapidly becoming less important. The central part of the basin is characterized by quartz sandstones, often moderately rounded, with associated stable accessory mineral suites. Cements are less common and many of the sandstones are more porous and have a better degree of sorting. Further south and south-east the sediments become finer grained, contain more detrital matrix and micaceous minerals, and sorting again becomes poorer. Concurrently with the increase in matrix there is also an increase in resistant and unstable quartz types, feldspar and metamorphic rock fragments. This assemblage reflects mixing of sediments from both northern and southern source areas. The preferential occurrence of metamorphic fragments in southern areas is the opposite to that determined for the total Hermannsburg Sandstone. This can be accounted for by the restriction of metamorphic fragments in the Hermannsburg Sandstone to the uppermost samples from localities on the northern margin of the basin. The change from significant siltstone rock fragment proportions in the north to chert in the south reflects the relative durability of these fragments to mechanical and chemical processes during transportation.

Cluster Analysis

Cluster analysis of the same petrographic data that was used for factor analysis provides additional evidence of the petrology of the basin. The spatial
relationship of cluster groups affords a subdivision of the broader grouping obtained from individual factors although the overall petrological conclusions reached are very similar.

Cluster analysis was performed using the programmes MULTICLASS and GROUPER (Lance, Milne & Williams, 1968) on a CDC 3600 computer.

The full analysis, using 58 variables, of the lower Hermannsburg Sandstone (0-250 m) and the members of the Finke Group shows a broad areal subdivision into three main units (Fig. 4.71). Only 3 localities in the south-east do not fit this general pattern and are more closely related to the northern clusters. The Horseshoe Bend Shale (FG2, FH, FK2) also forms a small group in the south-east which is characterized by a high matrix content and abundant biotite, muscovite, carbonate and feldspar (Table 4.26). It is also characterized by finer grain size and has higher grain size standard deviation, skewness and kurtosis.

The three major subdivisions are numbered 134, 135 and 136 from south-east to north-west (Fig. 4.71) and they have similarity coefficients of between 2.4 and 3.2. The defining parameters of these groups are given in Table 4.26 and the position and order in which they unite is given in Fig. 4.72. Thus groups 134 and 135 are the most similar and group 119 is quite distinct from any of
Group 136 is characterized by unstable accessory minerals, especially glauconite, and siltstone and sandstone rock fragments. In contrast group 137 (= 134 and 135) is dominated by straight extinction quartz and stable accessory minerals. Group 135 is characterized by stable accessory minerals especially tourmaline, straight extinction quartz, siltstone rock fragments and a finer grain size than group 134. Group 134 contains more feldspar, moderately stable heavy minerals, semicomposite and composite quartz grains than group 135.

Two important factors can be deduced from this cluster distribution pattern. Groups 136 and 135 show the effect of distance of transport on the mineralogical content of sediments. Both were derived from the northern margin and in all cases paleocurrents flowed from group 136 to group 135. The rapid decrease in abundance of soft unstable accessory minerals such as glauconite and collophane and the general decrease in the sedimentary rock fragments with increasing distance is attributable to the rapid abrasion, and possibly weathering, of these less resistant components during transportation.

The distinction between groups 134 and 135 cannot be attributed solely to the effect of transportation but rather it reflects the mixing of sediments from source areas of different lithologies. Paleocurrents indicate
that the Finke Group sediments were derived from the south and south-west while localities in the eastern Rodinga Sheet Area received detritus from the north and north-east. In both these localities the underlying sedimentary succession is thin either due to non-deposition in the south or erosion in the north-east and the higher feldspar and unstable quartz contents indicate derivation of at least part of the sediment from the metamorphic basement exposed in these areas. Mixing of this locally derived detritus with the sediment being transported through group 135 accounts for the similarity of these two groups.

When the 20 most significant variables, from the lower Hermannsburg Sandstone, determined from the similarity coefficient matrix, were used in a C.S.I.R.O. cluster analysis a somewhat similar pattern emerged (Fig. 4.73). The relationship between the groups is given in Figure 4.74 and the following groups were considered significant: 124, 126, 130, 133, and 136. Group 132 consists of 4 samples on the south-western portion of groups 124 and 130 and it is not significant.

Group 136 is related to group 128 on the basis of carbonate content. Group 136 is rich in labile siltstone and sandstone rock fragments and is slightly finer grained than group 128 which is characterized by its feldspar and garnet content, higher grain size standard deviation and greater distance from the northern margin (Table 4.27). Thus although the two groups are related they have
characteristic minerals which indicate their separate source areas.

Groups 137 (= 136 and 122) and 135 (= 124, 130, and 132) have similar matrix contents and are distinguished on the labile to non-labile fractions. Thus group 136 contains abundant siltstone rock fragments and feldspar while group 135 is characterized by straight extinction quartz and stable accessory minerals. The subdivision of group 135 into groups 124 and 130 is based on the higher proportion of siltstone rock fragments and stable accessory minerals in group 124, and the higher proportion of undulose and unstable quartz types in group 130. Group 124 is also finer grained than group 130.

The overall interpretation is thus the same as that derived from all variables — namely a decrease in labile fragments from north to south-east in the lower Hermannsburg Sandstone and a mixing in the east with sediments containing more unstable quartz types and feldspar.

The analysis of the whole Hermannsburg Sandstone using 58 variables is shown in Figure 4.75. The significant groups 129, 135, 136, 138, 139, and 142 are plotted in Figure 4.76 and they show somewhat similar but more complex distribution of lithologies than those obtained from the lower Hermannsburg Sandstone.
Groups 135 and 138 (=140) are the most similar groups represented. Group 138 is characterized by stable accessory minerals and undulose extinction quartz and the sediments are finer grained than those in group 135 (Table 4.28). Group 135 is characterized by its feldspar content and the proportion of unstable and moderately stable accessory minerals. The sediments in it are situated further from the northern margin and have a higher grain size standard deviation than those in group 138.

Groups 136 and 129 (=141) are distinguished on the basis of stable to unstable detrital fragments. Group 136 contains unstable accessory minerals such as glauconite and collophane, resistant quartz types and labile fragments especially limestone. Group 129 is characterized by stable heavy minerals and straight extinction quartz.

Group 141 contains abundant rock fragments and very resistant quartz types and this distinguishes it from group 140 which is characterized by unstable quartz types, feldspar, moderately stable accessory minerals. The groups are also distinguished on their distance from the northern margin of the basin.

A few samples from the northern margin form a distinct group (group 139) which is distinguished from the other localities in the basin on the basis of its higher proportion of labile rock fragments, unstable
quartz types and unstable accessory minerals and its consequently lower proportion of straight extinction quartz and stable accessory minerals.

Group 142 contains localities dominated by siltstone and shale and is therefore characterized by a high matrix and mica content.

Cluster analysis using the 20 most significant variables from the total Hermannsburg Sandstone and the Finke Group once again yielded a similar pattern of cluster group areal distribution and controlling factors (Figs 4.77 & 4.78 and Table 4.29). Thus group 141 has a high rock fragment and unstable accessory mineral content whilst 134, 138 and 139 have a higher quartz content and are distinguished by the presence of siltstone rock fragments in 139, undulose extinction quartz in 134 and feldspar and garnet in 138. Group 140 again represents siltstones and is characterized by a high matrix and mica content.

The conclusions reached from cluster analysis groups of the total Hermannsburg Sandstone are much the same as those for the lower Hermannsburg Sandstone. The quantity of labile rock fragments and unstable accessory minerals is greatest on the northern margin of the basin and the quantity of these soft and easily abraded fragments rapidly decreases down the paleoslope. Thus away from the northern margin the sediments become more quartz rich and have a rather restricted stable accessory
mineral suite. In the east mixing of sediments is apparent from the increase in feldspar and unstable quartz contents. These minerals are first cycle metamorphic derivatives from source areas in the north-east and south.
CHAPTER 5

STRUCTURE

The structural features and trends present within the sedimentary sequence of the Amadeus Basin are essentially simple which is in strong contrast to the complex folding and deformation which is characteristic of the basin margin especially to the north and north-east. Structures within the basin consist of a series of parallel or en echelon folds with relatively sharp, often asymmetrical anticlines separated by broader, often flat bottomed synclines. These fold belts form a broad regional arc convex to the south (Fig. 5.1). Individual fold axes trend north-westwards in the west, east-west in the centre, and north-eastwards in the east of the basin.

The Amadeus Basin consists of a downwarped remnant between regional anticlinoria which separate the Officer, Amadeus and Ngalia Basins. The margins of the Amadeus Basin are characterized by complex folding and faulting with nappe structures facing the basin from both the north and the south. An outline of these structural types has been given by Forman, Milligan and McCarthy (1967) and the structures are being mapped and studied in more detail at present. The southern margin of the Amadeus Basin was intensely folded during the Petermann Range Orogeny (Forman & Hancock, 1964) and was not affected greatly by the Alice
Springs Orogeny which produced most of the major structures on and near the northern margin (Forman et al., 1967). The intensity of structural deformation during the Alice Springs Orogeny decreased from north to south across the Amadeus Basin and in the south folding is largely limited to broad low amplitude undulations. This is also the major fold pattern in the west and south-east of the basin.

Folds formed during the Alice Springs Orogeny are numerous, regular and long (up to 240 Km) with folding styles ranging from gentle to closed, symmetrical, asymmetrical or overturned (Forman, 1968). The anticlines are generally tighter than the box-like synclines and frequently show faulted and contorted strata in their cores. Many of the anticlines are slightly asymmetrical due to thrusting or faulting on one or both margins e.g. Waterhouse Range. These folds are similar to the classical Jura type folds in many cases especially in the Rodinga Sheet area. Seismic, aeromagnetic and gravity surveys have delineated the basin margins and also indicate that the basement was not mobilized or involved in the folding and faulting seen in the sedimentary sequence (Krieg & Froelich, 1967). This supports the theory of Wells et al. (in press) that folding took place on one or more decollement surfaces owing to the lateral pressure and gravity sliding caused by overfolding and thrusting on the northern margin of the basin. The main plane of decollement is probably within the evaporites of the Bitter Springs Formation since this formation is frequently present in
the cores of the anticlines while the Heavitree Quartzite is always absent. In the north-eastern part of the basin there may be a second plane of decollement within the Chandler Limestone. Local diapirism may also be associated with these zones of decollement and many of the diapirs were in existence and grew during most of the period of lower and middle Paleozoic sedimentation as noted by local thinning over the structures.

Folding of the Amadeus Basin sediments was not confined to a single episode but rather a series of movements, starting at the Ordovician-Silurian boundary, and occurred with increasing frequency and intensity until they culminated in the intense deformation of the Alice Springs Orogeny. The age of the orogeny has been defined by mineral age determinations from the Arunta Metamorphics to be about 367 to 420 million years ago (Forman et al., 1967). The movements recorded which lead up to the Alice Springs Orogeny are the Rodingan and Pertnjara Movements (Wells et al., in press) and the Henbury Movement (Cook, in press). Several other unnamed movements occurred during the deposition of the Pertnjara Group and they are represented by unconformities and disconformities within the group especially near the northern margin.

The Rodingan Movement

The Rodingan Movement occurred prior to the deposition of the Mereenie Sandstone and it only had
indirect effects on the Pertnjara Group. It consisted of a simple epiogenic uplift of a large area beyond and including the north-east portion of the Amadeus Basin. It produced a low angle unconformity beneath the Mereenie Sandstone in this area but it did not result in the production of immature sediments. Hence, the depth of erosion through the pre-existing sediments was probably small. Renewed growth of localized diapiric structures may have occurred at this time since thinning of the Mereenie Sandstone over them has been noted.

A similar tectonic event is known in the Georgina Basin and there is a possible correlative in the Ngalia Basin. Thus the movement was fairly widespread north of the Amadeus Basin and it may be correlated with the Benambran Orogeny of eastern Australia since they both occur at approximately the same time (Forman, 1968).

The Pertnjara Movement

The Pertnjara Movement is a mid to upper Devonian event which produced a disconformity between the Mereenie Sandstone and the Pertnjara Group in the north-eastern part of the Amadeus Basin. This probably originated from regional uplift similar to that of the Rodingan Movement. However, localized folding and thrusting occurred in many places along the northern margin including the Ooraminna and Waterhouse Anticlines, the Hugh River Arch, the Goyder Pass - Carmichael structure and probably the
Johnston Hill diapir and other diapirs in the western portion of the basin. These localized uplifts were all of low amplitude and the only record of their occurrence is in the coarse pebbly sandstones deposited on their flanks and the thinning of the lower members of the Pertnjara Group over their crests. The south-western parts of the Amadeus Basin were unaffected by the uplift and folding in the north and in these areas the Parke Siltstone lies conformably on the Mereenie Sandstone. Uplift may have also occurred in the south-eastern part of the basin at this time and it resulted in the northward deposition of the Polly Conglomerate and Langra Formation.

The Pertnjara Movement can be correlated with the unconformity beneath the Mt Eclipse Sandstone in the Ngalia Basin (Wells et al., 1968) and a disconformity in the Georgina Basin. The presence of poorly sorted sediments in the Amadeus and Ngalia Basins suggests that they were relatively close to the uplifted area while the well sorted sands in the Georgina Basin indicate a greater distance from the source of detritus (Forman, 1968). Thus the Pertnjara Movement is also of regional significance and it may be correlated on age with the Tabberabberan Orogeny of eastern Australia.

The Henbury Movement

The Henbury Movement was defined by Cook (in press) as the uplift which occurred in the north-eastern
part of the Amadeus Basin after the deposition of the Parke Siltstone and prior to the deposition of the Hermannsburg Sandstone. It was proposed to account for the absence of the Parke Siltstone over the Illamurta Structure (see p. 396).

Unnamed Movements

Several other unnamed movements probably occurred during the deposition of the Pertnjara Group to produce localized unconformities on the northern margin and south as far as the Gardiner and James Ranges. Unconformities are rare within the Hermannsburg Sandstone but they have been recorded at the top of the Ljiltera Member and within the Brewer Conglomerate.

The Alice Springs Orogeny

The preceding movements culminated in the Alice Springs Orogeny which produced the major folding and faulting on the northern margin and within the sedimentary sequence of the Amadeus Basin. The orogeny probably occurred at the height of the crustal warping either late in the Devonian or early in the Carboniferous. The unnamed movements preceding the orogeny probably represent pulses of tectonism at the start of the orogeny rather than disconnected movements. The Brewer Conglomerate is probably a synorogenic deposit (Forman et al., 1967) rather than a pre-orogenic deposit (Forman, 1968).
latter concept is hard to visualize since the upper portions of the Brewer Conglomerate, especially the Undandita Member, are sub-horizontal and have obviously not been involved in a major period of folding and faulting. The unconformities in the Brewer Conglomerate adequately explain the attitudes of the formation as depicted in Figure 5.2. The only alternative method to explain the attitudes of the Brewer Conglomerate would be to propose a series of variable angle thrust sheets concave downwards away from the zone of folding. However, the presence of thrust sheets in such massive sediments without a suitable media for flowage is unlikely and they have not been delineated in any of the seismic sections. Thus the earlier concept of the Brewer Conglomerate being a synorogenic deposit appears to be more plausible than the suggestion that it is pre-orogenic.

Contemporaneous folding and faulting occurred in the Ngalia Basin, and to a lesser extent, in the Georgina Basin. This orogenic episode may be related to the Kanimblan Orogeny in eastern Australia.
THE RELATIONSHIP OF SEDIMENTATION TO INDIVIDUAL STRUCTURES

Western Amadeus Basin

The area west of Mt Solitary was not visited by the author and the results presented here have been compiled from B.M.R. Reports and the seismic survey report by Krieg and Froelich (1967) for Magellan Petroleum (N.T.) Pty Ltd.

The Mt Rennie and Mt Liebig Sheet Areas are transected by a series of north-west trending anticlines which form the Gardiner, Glen-Edith, Watson, Cleland and Gypsum Ranges. These anticlines are sharp, asymmetrical, and have thrust faults on their south-west margins. They apparently formed as a result of thrusting from the north with the consequent uplift and folding of the anticlinal structures. Uplift may have started during the deposition of the Larapinta Group since it thins over the crest of the Cleland Anticline. The anticlines are separated by broad, shallow, box-like synclines. Faulting across the Gardiner and Watson Anticlines has resulted in dextral displacement of the anticlinal axes. Thinning of the Pertnjara Group over these structures cannot be determined due to erosion and poor exposures.

Diapiric structures have been recorded at Johnstone Hill and possibly at the western end of the surface expression of the Gypsum Anticline. The Johnstone Hill diapir has been rotated through ninety degrees and it is well
exposed in cross-section. The core of the structure is composed of gypsum from the Bitter Springs Formation and this pierces the whole stratigraphic section from the Cleland Sandstone to the Hermannsburg Sandstone. Stratigraphic thinning of the units of the Larapinta Group over this diapir indicate that the structure was being uplifted during the period of their deposition. The Mereenie Sandstone also thins over the structure but once again no evidence is available for the Pertnjara Group.

The relationship of sedimentary facies to structure has not been studied in any detail in the western part of the basin. The siltstone in the Cleland Hills area is very thin (40 m), and it contains calcareous and dolomitic horizons as well as numerous thin sandstone interbeds. This suggests that the siltstone may have been thinning towards a local topographic high. However, the subsurface expression of the siltstone is not known in this area and the deposit at Cleland Hills may represent a locality near the depositional edge of the Parke Siltstone. At Johnstone Hill diapir there is an unexposed interval between the Mereenie and Hermannsburg Sandstones which may represent a thin (25 m) siltstone interval. Forty-five kilometres west of Johnstone Hill diapir the Hermannsburg Sandstone rests directly on the Mereenie Sandstone with apparent conformity. It is overlain by pebbly sandstone and conglomerates probably equivalent to the Ljiltera and Brewer Conglomerates.
The Carmichael - Deering Creek Structures

These two related structures probably extend north-eastwards as an arcuate ridge to the Goyder Pass diapir and in both areas growth of the structure probably ceased after the deposition of the Harajica Sandstone Member.

The Carmichael-Deering Creek structures are part of a faulted anticlinal trend which started to grow in the early Ordovician since the Larapinta Group thins in this area. Stratigraphic thinning is most prominent at Carmichael and the units become thicker westwards towards Deering Creek. The southern limb of the structure is overturned and faulted while the northern limb dips at a moderate angle into the Idirriki Range Syncline. The Pertnjara Group on the northern flank of the Carmichael structure shows a complex system of stratigraphic thinning combined with lateral facies changes.

On the southern flank of the structure tectonic influence is more apparent with periods of uplift giving rise to local unconformities and facies changes. Erosion of the Mereenie Sandstone took place prior to Pertnjara deposition. Uplift in the Carmichael area produced a pebbly, coarse quartzose sandstone (the Harajica Sandstone Member) which gave way to the west and south, and probably to the east, to quieter depositional environments where the Deering Siltstone Member was being deposited. As uplift and erosion continued the locus of coarse
sedimentation moved westwards in the Deering Creek area, and the grain size and abundance of clasts diminished from east to west. No erosion breaks or locally derived deposits occur within the upper Parke Siltstone or Hermannsburg Sandstone in the Deering Creek area and there is no evidence for continued growth after the deposition of the Harajica Sandstone Member.

**Tyler Pass - Carmichael Area**

The Tyler Pass-Carmichael area is a very important area showing lateral facies variations within all the units of the Pertnjara Group. Part of the area is shown in Figure 5.3 and it has not been directly affected by structural growth during deposition. In the south-west the Parke Siltstone abuts against the uplifted Carmichael-Deering Creek structure and has low dips to the south-east swinging to north-east adjacent to the structure. Dips within the Pertnjara Group steadily increase to the north-east until around Tyler Pass the dips are almost vertical and they continue as such into the adjacent Goyder Pass structure.

The Deering Siltstone Member forms a wedge shaped sheet which thickens south-westwards from its depositional edge between Tyler Pass and Stokes Pass (approximately at section AD). The siltstone is dominantly light reddish brown but it contains numerous thin sandstone lenses throughout the section, especially near the base and top
of the unit. On the flank of the Carmichael structure thin sandstone beds are prominent and the unit rapidly wedges out against the structure. Thinning is attributed to non-deposition rather than erosion. The Harajica Sandstone Member is relatively thin and coarse near Tyler Pass and Goyder Pass and it thickens to a maximum near Stokes Pass before becoming thinner and finer towards the Carmichael structure. In the north-east the Harajica Sandstone Member disconformably overlies the Mereenie Sandstone, is more quartzose and contains frequent pebbles of sandstone. At Stokes Pass it is more lithic and towards the Carmichael structure where it rapidly wedges out it contains siltstone intervals at the base and top of the unit. Where the Harajica Sandstone Member overlies or underlies siltstone units the contacts are always gradational. The Dare Siltstone Member is rather atypical in this area and it merges into the thicker and more widespread Amulda Member. These units consist of thinly interbedded micaceous light red brown siltstone and fine micaceous sandstone. They are absent over the Goyder Pass diapir and wedge out on either side of it. From Tyler Pass to the Carmichael structure these units become thicker and siltstone becomes more prominent except adjacent to the Carmichael structure where they both rapidly wedge out.

The Hermannsburg Sandstone conformably overlies the Parke Siltstone over this whole area and the contact is usually gradational. West of Tyler Pass the lower Hermannsburg Sandstone forms a recessive unit of lithic
sandstone which becomes more silty towards the Carmichael structure. The Ljiltera Member also becomes finer and less pebbly towards the south-west where it merges into non pebbly Hermannsburg Sandstone. This unit is the first unit which overlaps onto and over the Carmichael Structure and continues south of this structure. The Ljiltera Member has a gradational upper contact with the Brewer Conglomerate. In the Tyler Pass area the conglomerate is massive throughout while at Stokes Pass there is a pebbly and conglomeratic sandstone unit 350 to 700 m above the base of the conglomerate. Pebby sandstone units become more frequent in the lower Brewer Conglomerate towards the Carmichael structure and some of these units are even shaley. The overlying conglomerate is massive but it gives way up the section to the pebbly lithic sandstones of the Undandita Member.

Thus there is a general and repetitive trend for each sedimentary unit to become finer grained from the Tyler Pass area towards the Carmichael structure. Paleo-current patterns indicate that this sedimentary trend follows the general paleocurrent directions determined for the Harajica Sandstone Member and the Hermannsburg Sandstone. The lateral fining of facies away from the source area indicates decreasing current competence and probably decreasing stream gradients. Although this sequence in the Western MacDonnell Range probably has a distinctive geological setting the conclusions reached may be of broader application to the rest of the MacDonnell Range homocline.
Goyder Pass Diapir

The Goyder Pass diapir is a rotated diapiric structure on the eastern end of the thrusted Carmichael-Deering Creek Anticline. It is exposed in cross-section and clearly shows the thrust faulted south-eastern boundary of the structure (Fig. 5.4). The Larapinta and Mereenie units all show stratigraphic thinning over the crest of the structure indicating that the structure was in existence and growing during the period of their deposition. The Deering Siltstone Member wedges out west of the structure and does not reappear on the eastern side of it. The contact between this member and the overlying Harajica Sandstone Member is not erosional indicating that the area represents the eastern depositional edge of the Deering Siltstone Member. The basal Harajica Sandstone Member contains numerous pebbles of siliceous sandstone, and is coarser and more quartz rich over the Goyder Pass diapir than on either side of the structure. This indicates that local reworking of the Mereenie Sandstone provided quartz rich detritus on the top and flanks of this structure. The Harajica Sandstone Member also becomes thinner over the crest of the structure although the minimum thickness of this unit occurs west of the axis of thinning exhibited by the lower formations. This could indicate that rotation of the Goyder Pass diapir had already commenced. Neither the Dare Siltstone Member or the Amulda Member are present on the crest of the diapir although they occur on both flanks. Once again this is due to non-
deposition rather than erosion. The Hermannsburg Sandstone shows evidence of only minor thinning over the structure although the sediments tend to be slightly coarser. The structure has had no effect on the Brewer Conglomerate and thus structural growth must have terminated during or at the end of the deposition of the Harajica Sandstone Member.

The MacDonnell Range Homocline

The MacDonnell Range Homocline extends for approximately 270 Km in an east-west direction and it forms the northern margin to the preserved Amadeus Basin. The area west of Tyler Pass has already been described, as has the Goyder Pass diapir. East of Goyder Pass the structure dips vertically as far as section AK from whence the dip gradually decreases to the east. At Ellery Creek the dip is about 55°S and over the Hugh River Arch the dip is about 40°S.

The Deering Siltstone Member is absent east of the Goyder Pass diapir and the Harajica Sandstone Member maintains a fairly constant thickness from Goyder Pass to section AK. Throughout this area it is a moderately lithic medium grained sandstone, except for the basal 10 or 20 m which are quartzose and may contain a few siliceous sandstone pebbles. The Harajica Sandstone Member appears to overlie the Mereenie Sandstone conformably but local erosion of the latter produces the quartz rich facies in the overlying sandstone. This quartz rich facies becomes
thicker eastwards until in the Ellery Creek area it occupies the total thickness of the Harajica Sandstone Member. A silty interval overlying the Harajica Sandstone Member extends almost continuously from the eastern flank of Goyder Pass diapir to 5 Km east of section AK. It probably represents the Amulda Member since it is composed of brown dolomitic siltstone and interbedded thin fine sandstone lenses. This silty unit is conformably overlain by the Hermannsburg Sandstone which merges upwards into the Ljiltera Member. Between section AK and Ellery Creek the lower Hermannsburg Sandstone becomes coarser grained and pebbly and merges laterally into the Ljiltera Member. Thus at Ellery Creek a conglomeratic facies of the Ljiltera Member directly overlies the quartzose sandstones of the Harajica Sandstone Member.

West of section AK the contact between the Ljiltera Member and the Brewer Conglomerate is usually conformable and gradational. However, east of section AK the contact is unconformable and cuts down through the underlying units (Fig. 5.5). Thus east of Ellery Creek and over the Hugh River Arch the Brewer Conglomerate progressively cuts down through the Hermannsburg Sandstone, Harajica Sandstone Member, Mereenie Sandstone, Larapinta Group, Goyder Formation and Jay Creek Limestone. These units reappear east of the Hugh River.

Within the Brewer Conglomerate colour variations are quite apparent on aerial photographs (e.g. Fig. 5.4)
and these can be related to changes of dominant clast lithologies. Thus areas with dark colours usually contain abundant red-brown sandstone clasts such as Arumbera or Hermannsburg Sandstone whereas light coloured areas are dominated by white sandstones of the Larapinta Group. These areas of different colour have variable shapes but many are lensoid. They are frequently differentiated by a change in dip and overlie local unconformities. They are assumed to represent deposits from individual large alluvial fans and the clasts present within each deposit reflect the lithologies present in the drainage basin for that fan. The dip of each successive deposit is less than the preceding one and some of the highest deposits, such as the one indicated on Figure 5.4 have dips of 10 to 15° and possibly indicate the original surface expression of the fan which has been modified very little by post-depositional tilting or folding. Thus the Brewer Conglomerate is probably a synorogenic deposit with each successive deposit being less affected by the tilting and folding (Fig. 5.2). Tectonic movement during deposition would account for the presence of unconformities within the sequence.

**Waterhouse Range**

The Waterhouse Anticline probably existed as a low and intermittently uplifted structure throughout the period of Pertnjara deposition. The structure is arcuate and convex to the south and the southern margin of the structure
is thrust faulted with a maximum displacement of 2400 m. The existence of the fault was deduced by Krieg and Froelich (1967) from seismic reflection data. Uplift of the Waterhouse Anticline probably commenced during the deposition of the Mereenie Sandstone and Pertnjara Group. The most noticeable features produced by this structural growth are the unconformities within the stratigraphic sequence, although another feature is the stratigraphic thinning on the northern margin of the structure, especially within the Mereenie and Pertnjara sediments. The unconformity beneath the Mereenie Sandstone is an accentuated example of the regional unconformity in the north-eastern Amadeus Basin. On the Waterhouse Anticline the unconformity is most pronounced on the eastern end of the structure. Another local unconformity underlies the Ooraminna Sandstone Member and is probably responsible for the abundance of quartz in the latter. The Ooraminna Sandstone Member is thickest at its type section in Orange Creek and it becomes thinner to the west. This suggests that uplift was again most pronounced in the eastern end of the structure. Further uplift occurred at this same locality during the period of Pertnjara deposition and it produced a local intraformational unconformity at the top of the Ooraminna Sandstone Member. The uplifted area was probably dome shaped and the unconformity surface cut down through the Ooraminna Sandstone Member and the upper Mereenie Sandstone and is exposed on the nose of the present structure. On the flanks of the present structure the unconformity merges into a disconformity
and it can only be recognized by the increase in pebble variety. Overlying the unconformity there is a series of lensoid conglomerates, predominantly consisting of rounded sandstone clasts, and relatively quartz rich, coarse sandstones. The pebbly sandstones become more lithic up the section where they closely resemble the Ljiltera Member.

The Waterhouse Anticline also provided a barrier to the southward deposition of the Brewer Conglomerate and only pebbly sandstones are known south of the structure.

Ooraminna Anticline

The Ooraminna Anticline formed a local structural elevation at the start of Pertnjara deposition. It is a slightly asymmetrical structure which plunges gently to the south-west where it is continuous, via a saddle, with the Orange Anticline. There is a paraconformity between the Mereenie Sandstone and the Ooraminna Sandstone Member over the anticline and this surface is represented by a pebbly sandstone facies 5 to 20 m thick. The asymmetry of the anticline is represented by a coarser facies of the Ooraminna Sandstone Member on the southern flank of the structure than to the west and north. The Ooraminna Sandstone Member is a clean quartz rich facies of the Hermannsburg Sandstone and it probably owes its existence to the erosion and reworking of Mereenie Sandstone during the period of Pertnjara deposition. Uplift of the Ooraminna Anticline probably did not continue during the deposition
of the Pertnjara Group since the sublithic Hermannsburg Sandstone conformably overlies the Ooraminna Sandstone Member and it was deposited without any erosional breaks.

**Ross River Syncline**

In the Ross River Syncline the distinction between the Mereenie Sandstone and the Ooraminna Sandstone Member is hard to pick because there is no angular discordance or pebble band separating the units. Higher in the Pertnjara sequence the sandstones become more lithic and there may be local disconformities separating the Ljiltera Member from the underlying Hermannsburg Sandstone and from the overlying Brewer Conglomerate.

**The Gardiner - Tyler Uplift**

A structural high trending north-eastwards from the Gardiner Range via Gosses Bluff and the Tyler structure to Glen Helen was delineated by seismic work (Krieg and Froelich, 1967). It may have existed as a low feature influencing sedimentation since the Cambrian because many changes of facies appear to coincide with this trend (Wells et al., in press).

**Gosses Bluff**

Gosses Bluff is situated on the north-east trending Gardiner-Tyler-Glen Helen structure and it has been
variously described as a diapir, a volcanic or crypto-volcanic explosion crater, and an astrobleme. The latter explanation now appears to be the most plausible and it explains all the intense shock deformation features present in and around the Bluff.

There is no evidence for structural growth of the Bluff during deposition of the Pertnjara Group although the Gardiner-Tyler-Glen Helen structure may have existed as a slight ridge prior to deposition thus accounting for the almost complete lack of the Deering Siltstone Member at this locality. The stratigraphic sequence at Gosses Bluff starts with 40 m of interbedded fine to medium sandstones and thin light reddish brown siltstones with the sandstones predominating. The following 170 m contain several siltstone interbeds scattered throughout the interval and this is overlain by 526 m of massive and cross-bedded fine to medium sandstone with virtually no recognizable breaks in the sequence. This whole 736 m sequence probably represents the Harajica Sandstone Member although the lowest interval could possibly represent the top of the Deering Siltstone Member. In this area there is no evidence of erosion between the Mereenie Sandstone and the Pertnjara Group and the beds always appear conformable. The lower sandstone sequence is abruptly overlain by interbedded red-brown and greenish siltstones of the Dare Siltstone Member which attains a thickness of about 180 m in this area. It differs from the type section of the Dare Siltstone Member since it contains
thinner and poorly developed cycles, and it contains frequent thin interbeds of fine sandstone which occasionally show ripple-drift bedding. Towards the top of this member thin sandstone beds become more numerous and the contact between the Dare Siltstone Member and the Amulda Member (46 m) is gradational and arbitrary. Likewise, the contact between the Amulda Member and the Hermannsburg Sandstone is gradational and the lower Hermannsburg Sandstone is composed largely of cross-bedded sandstone with rare silty interbeds. To the north the lower Hermannsburg Sandstone merges upwards into the pebbly coarse sandstone of the Ljiltera Member and finally into the conglomeratic sandstones of the Undandita Member. The massive Brewer Conglomerate is not present around Gosses Bluff owing to a southward lateral facies change from massive conglomerate to pebbly and conglomeratic sandstone.

Gardiner Range

The central portion of the Gardiner Range may have had a slight surface expression from the Ordovician to the Devonian because units of the Larapinta Group and the Mereenie Sandstone thin slightly over the structure and this is especially noticeable on the northern flank near Areyonga. The Parke Siltstone, although poorly exposed, also appears to be slightly thinner over the Gardiner Range (Fig. 2.6). Confirmation of the existence and growth of the Gardiner Range during the period of Hermannsburg Sandstone deposition was obtained from the
paleocurrent patterns. During the whole of this period paleocurrent directions west of Hermannsburg Mission all have a northerly or westerly component which is directly opposed to the general south-easterly trend. Thus the central Gardiner Range was an effective barrier during the deposition of the Hermannsburg Sandstone and it would have acted as a subsidiary source of detritus. However, no coarse material was derived locally from this source which leads to the conclusion that it probably had a low relief throughout the period of deposition.

Uplift of the Gardiner Range was probably due to a combination of folding and thrusting. The thrust fault on the northern margin of the Gardiner Range is estimated to have a displacement of between 4300 m (Wells et al., in press) and 6700 m (Krieg and Froelich, 1967). The fault is a south dipping reverse thrust fault and the inclination of the fault plane decreases southwards.

The Illamurta Structure

The Illamurta Structure has been described in detail by Cook (in press) and he considered both thrusting and structural growth hypotheses to account for the exposed features. He noted that there is progressive stratigraphic thinning over the crest of all the units in the Pertaoorrta and Larapinta Groups, and in the Parke Siltstone. This suggests that structural growth during deposition was important since thrusting would tend to preferentially
remove the less competent siltstone horizons rather than the interstratified sandstones. Below the Mereenie Sandstone the Rodingan Movement initiated the erosion of the upper Larapinta Group from the crest of the structure although thinning within the Mereenie Sandstone is not apparent which suggests stable conditions during this time interval. The Pertnjara Movement did not affect this area and on the flanks of the structure the Parke Siltstone conformably overlies the Mereenie Sandstone. The absence of Parke Siltstone over the crest of the structure has been attributed to erosion following further uplift - the "Henbury Movement" of Cook (in press). However, it could also be attributed to non-deposition over the growing structure because the contact of the Parke Siltstone and Hermannsburg Sandstone is gradational to the north, east and west although the relationship between the two units was not studied adjacent to the structure itself. No stratigraphic thinning was recognized in the Hermannsburg Sandstone although cross-bed orientations from the lower Hermannsburg Sandstone indicate that local paleo-currents were radial away from the structure.

The Seymour Range Structure

The Seymour Range structure is poorly exposed on the western end of Seymour Range. Details of the structure are not known but it causes a marked change in strike at the end of the range which is accompanied by stratigraphic thinning of the Deering and Dare Siltstone Members in a
westerly direction. The Harajica Sandstone Member overlies the Mereenie Sandstone with apparent conformity and it is more quartz rich in such areas. Although uncertain, stratigraphic thinning and wedging out of the siltstones suggests that this structure was also growing during the deposition of the Pertnjara Group. Cook (in press) suggests that this structure is related to the Illamerta and Goyder Pass structures along the trend of a basement ridge.

Larrier Bore

The relationship of the breccia and conglomerate adjacent to the fossil fault scarp at Larrier Bore has already been discussed (p.129-130).

Black Hill Range

The Black Hill Range is a prominent east-north-east trending anticline which is faulted on its southern margin. The structure was formed after the deposition of the Finke Group since it has no influence on these sediments and both the Polly Conglomerate and the Langra Formation were deposited northwards, as sheets, uniformly across the present structure. The base of the Polly Conglomerate is unconformable on the Winnall Beds and locally derived included clasts are present in the basal 5 m of the formation.
RELATIONSHIP BETWEEN THE SEDIMENTARY UNITS

There are two main hypotheses which need to be considered in the relationship between the sedimentary units. They are:-(1) simple vertical relationships between the sedimentary units with possibly some lateral facies variation to allow correlation of the Pertnjara and Finke Groups; and (2) lateral variation through facies changes to account for the distribution of all units.

With the first hypothesis, the sequence in the Pertnjara Group can be accounted for by considering widespread uniform depositional environments. Thus the Deering Siltstone Member would be a wedge shaped unit thickening from zero on the northern margin to 150 m in the centre of the Amadeus Basin. It is overlain by the widespread wedge shaped blanket of the Harajica Sandstone Member which thins in a southward direction. The Dare Siltstone Member forms another southward thickening wedge and it is overlain by the fairly uniform blankets of the Amulda Member, Hermannsburg Sandstone and Ljiltera Member, and by the possibly southward thinning wedge shaped Brewer Conglomerate. The Finke Group also consists of a fairly widespread sequence of blanket-like deposits possibly thinning northwards and thickening eastwards.

When such widespread environments are envisaged the relationship between the Pertnjara and Finke Groups is rather restricted. One possibility is that the Polly
Conglomerate and Langra Formation were deposited in the south-east and if present in the centre of the basin then they must have been eroded prior to the deposition of the contiguous Parke Siltstone and Horseshoe Bend Shale. The Hermannsburg Sandstone would then be equivalent to the Idracowra Sandstone and Hakea "Formation". Other alternative inter-relationships between the groups all involve facies changes and are best considered in terms of the second hypothesis.

The second hypothesis considers lateral variation of the sedimentary units through facies changes. The facies variations noted in specific localities together with the gradational nature of almost all the sedimentary units both lend strong support to this hypothesis. This hypothesis is also in accordance with present day terrestrial environments which are characterized by lateral facies variations rather than widespread uniform environments. There is no direct evidence for this hypothesis based on age determinations since the formations are virtually unfossiliferous (see later).

The most probable relationships between the sedimentary units of the Pertnjara and Finke Groups are shown in columnar form (Fig. 5.6) and diagrammatically (Fig. 5.7). Most of the units are probably time transgressive and denote the existence of a particular depositional facies in space and time. Thus the Polly Conglomerate, Langra Formation and Parke Siltstone are probably all time
equivalent units which are coarsest near the source areas in the north and south-east and progressively fine towards the central part of the inland drainage basin. The Hermannsburg Sandstone is probably equivalent, at least in part, to the Horseshoe Bend Shale. These latter units indicate the change in depositional basin shape with the deepest portion of the basin moving south-eastwards away from the locus of maximum Parke Siltstone deposition. The Brewer Conglomerate is probably equivalent to the Ljiltera Member, upper Horseshoe Bend Shale, Idracowra Sandstone and Hakea "Formation". The spatial relationship between these units of the Pertnjara and Finke Groups at various intervals of time is given in Figure 5.8. Of course, this interpretation is not the only possible combination of the units in space and time but it is thought to be the most plausible one based on the evidence known to date.

VOLUME CONSIDERATIONS

The total volume of Pertnjara and Finke Group sediments preserved at present is in the order of 50,000 km$^3$. The original volume can only be estimated from present maximum thicknesses. The Missionary and Brewer Plains area contains the greatest thickness of Pertnjara sediments (approximately 3500 m). Over the remainder of the basin the Pertnjara sediments would probably have formed a wedge-like sheet of sediment, with preserved thicknesses in the order of 1500 m in the north thinning to about 500 m in southern exposures. These sediments would have largely
originated from the northern margin of the basin and, therefore, the southward thickening of the Polly Conglomerate and Langra Formation, derived from a south-eastern source area, has been omitted from the volume considerations. The volume of sediment derived from the northern source area to form the Pertnjara Group may be calculated and is approximately 132,000 Km$^3$. Original thickness of the Pertnjara Group was probably thicker than present exposures show and therefore this figure is a minimum value for the volume. The source area between the Amadeus and Ngalia Basins is approximately 700 Km long and 80 Km wide, and 2.35 Km would have to be eroded off this whole area to provide the detritus forming the Pertnjara Group. This is quite feasible since on the northern margin of the Amadeus Basin approximately 6 Km of pre-existing sediments were eroded to expose the Arunta metamorphics which occur as abundant clasts towards the top of the Brewer Conglomerate. Thus, to supply the detritus for the Pertnjara Group either the northern margin of the basin was an exceedingly high mountain range throughout the period of deposition or uplift of the range occurred either continuously or sporadically throughout the deposition of the Pertnjara Group. If the former hypothesis were correct coarse debris would be expected towards the base of the group and the detritus should become finer as the mountain chain was worn down and eroded back from the area of deposition. The alternate hypothesis of continuous or sporadic uplift would account for the increasing coarseness of detritus up the stratigraphic column and also for the
intraformational disconformities and unconformities. The growth of localized structures has already been discussed and these also point to a continued growth of structural relief during deposition. Climatic implications cannot be ignored since increase in precipitation in non-vegetated areas could cause increase in sediment yield and coarseness of detritus.

**PALEONTOLOGY**

Considering the number of workers who have examined the Amadeus Basin it is surprising to find that very few fossils have been discovered in the Pertnjara Group. Rare fish plates and spores have been found in the lower part of the Parke Siltstone on the Mereenie Anticline and in the hills south of Deering Creek, while the only other fossil is a plant fragment in the Lower Hermannsburg Sandstone (450 m above the Mereenie Sandstone), north-east of Tempe Downs homestead (section DR). It was identified as *Sigillaria* sp. by Leslie (1960) but Gilbert-Tomlinson (1965) considers that it is too poorly preserved for reliable identification.

The only fossil found during this study was a single fragment of a fish plate in the middle Deering Siltstone Member on Mereenie Anticline, although a few isolated, winding, non-branching, unsegmented trace fossils were seen in the Lower Hermannsburg Sandstone.
At the northern end of Lawrence Gorge (section BN) in the Waterhouse Range, a flattened ovoid trace fossil (Fig. 5.9) was preserved on a sandstone bedding plane 18 m above the Mereenie Sandstone. It is very similar in appearance to the trace fossil genus Urohelminthoida, with a tail-like appendage at the turn of what could have been part of a broad meander. Isolated occurrences of trace fossils with the general form of Planolites were seen in the sandstones of section EV and section EP (17 m and about 60 m above the Mereenie Sandstone respectively). They both consist of slightly flattened, ovoid burrows of about 5 mm diameter, which irregularly penetrate the sandstone at an oblique angle to the bedding. They were not seen to branch and the burrow infilling is slightly darker than the surrounding sediment.

A series of spine-like raised impressions was collected by Frome - Broken Hill from the Hermannsburg Sandstone at Steel Gap (section EW). Taylor (1959) considered that the origin of these impressions is organic but problematical.

Biogenic sole markings are very rare. Moulds of epifaunal trails were only seen in the basal Hermannsburg Sandstone at section DH. They were formed on the upper surface of a siltstone horizon which also contained poorly developed mud cracks (Fig. 5.10). Trails of two sizes are visible – thin randomly winding trails similar to the trace fossil genus Gordia, and straighter 1 cm wide symmetrical
grooves flanked by slightly raised rounded ridges. The former were probably formed by worms while the latter could have formed as a soft bodied organism ploughed through the semi-consolidated mud. Irregular load casting has modified part of the surface.

Placoderms

Fragmentary plates of the dermal armour of the antiarch Bothriolepis Eichwald were discovered by R.M. Hopkins of Magellan Petroleum Corporation, in the uppermost fine sandstones of the Harajica Member (180 m above the Mereenie Sandstone) on the northern flank of the Mereenie Anticline. They were identified and described by Gilbert-Tomlinson (1965). The two identifiable fragmentary plates are anterior dorsolateral plates of the trunk armour of different individuals. They are characterized by a rugose reticulate ornamentation and they have the dorsolateral ridge developed as a keel. They resemble the Bothriolepis illustrated by Hills (1959, pl.8, fig.B) from the Dulcie Range, Northern Territory (Georgina Basin), and they also have similarities with the Victorian B. gypslandiensis Hills.

A further four poorly preserved fish plates were collected by L.G.G. Pearce from the pebbly Harajica Sandstone (239 and 244 m above the Mereenie Sandstone) in the hills south of Deering Creek. All the plates are angulate and are probably parts of the trunk armour of
an antiarch (Gilbert-Tomlinson, in Pearce, 1968). Two of them have a keel at the angulation and are similar to the Dare Plain Bothriolepis while another has a sculpture of distinct tubercles similar to a Bothriolepis in the Dulcie Range (Hills, 1959, pl.8, fig.C). Thus, although identification is uncertain, these plates probably belong to the genus Bothriolepis.

The unidentifiable arthrodiran fragments found by Cook (1966) in the middle of the Harajica Sandstone (unit Pz(2)) have now been determined to originate from an overturned outlier of the upper Mereenie Sandstone (D. Milton, pers. comm., 1967). Petrology of the quartzose sandstone in which the fish plate was found has confirmed this diagnosis and the rock is petrologically distinct from the underlying Harajica Sandstone at the same locality.

A single fragment of fish plate was found by Dr K.A.W. Crook and the author in the middle Deering Siltstone Member, 5.55 Km east of the former Dare Plain locality. It occurred in the light reddish brown, massive but thin bedded siltstones 35 m above the Mereenie Sandstone, and 100 m stratigraphically below the former locality. Part of the bone material was still preserved and consisted of clear hydroxylapatite (basic calcium phosphate, identified by X-ray powder diffraction analysis). The fragment is 30 mm X 9 mm, is not angulated, and does not show any areas of plate overlap. The ornamentation (Fig. 5.11) is quite distinctive with random, almost circular, tubercles at one end merging
gradually into a subparallel array of elongate tubercles at the other end. The elongation of the tubercles is almost parallel to the length of the fragment and towards the upper margin (in the orientation of the photograph) the tubercles decrease quite markedly in size. The circular tuberculate ornamentation is somewhat similar to that of one of the Dulcie Range Bothriolepis described by Hills (1959, pl.8, Fig.C). The evidence provided by this fossil is insufficient for identification beyond saying that it is a fragment of the dermal armour of a fish - possibly a placoderm.

Age of the Placoderms

Evidence for the time range of Bothriolepis in Australia comes from eastern N.S.W. and Victoria where Bothriolepis first occurs just above late Frasnian or younger marine faunas (Gilbert-Tomlinson, 1965). This is probably the time at which Bothriolepis arrived at the then coast of Australia, and hence it may be inferred that the central Australian specimens will be no older than late Frasnian or early Framennian.

Correlations based on Placoderms

On the basis of the general and morphological similarity of the Bothriolepis and Phyllolepis found in the Amadeus and Georgina Basins to those in N.S.W. and Victoria, it has been suggested (Gilbert-Tomlinson, 1965) that the sedimentary cycle in which these terrestrial beds were
deposited occurred in a relatively short period of time near the close of the Devonian Period. The assemblages in these areas are dissimilar to those of northern Queensland and Western Australia. This suggests that the Devonian fish inhabited chains of lakes connected by a common river system with the drainage from the central Australia area towards the south-east.

**Palynology**

The only successful palynological study carried out in the Amadeus Basin to date is the report by Hodgson (1965) on spores found in Exoil Mereenie Water Bore No. 2 (23°59'25"S, 131°33'06"E). All cuttings from the well, except for the interval between 207 m and 213 m, are too oxidized to contain palynomorphs. The fossiliferous zone is in the Upper Parke Siltstone immediately above the Harajica Sandstone Member on the northern flank of the Mereenie Anticline (approximately 185 m above the top of the Mereenie Sandstone). Hodgson identified the following species from the few, well preserved, carbonized spores that he obtained:

- **Leiotrilites liebigensis** Hodgson (moderate)
- **Geminospora lemurata** Balme (relatively abundant)
- **Radiaspora darensis** Hodgson (rare)
- **Lophozonotriletes** sp. (rare)
- **Auroraspora** cf. **A. micromanifestus** (Hacquebard) Richardson (rare)
- **Ancyrospora parke** Hodgson (rare)
- **Ancyrospora amadei** Hodgson (rare)
- **Ancyrospora** cf. **A. simplex** Guennel (abundant)
From the abundance of Ancyrospora and Geminospora limurata Hodgson concludes that the sample is of late Middle or early Upper Devonian age - most probably Frasnian.

A series of twenty samples (Table 5.1) were analysed by the author for palynomorphs using the method outlined in appendix 12. However, all samples were too oxidised to yield any positive results. I acknowledge the assistance and advice of Drs E.M. Kempe and D. Burger, and the use of the facilities of the Bureau of Mineral Resources, Canberra, to carry out this study.

Paleoecology

The ecological habitat of Bothriolepis is uncertain, especially with regard to salinity tolerance. The genus has a very widespread geographical distribution and this suggests that it must have been capable of marine dispersal. Possibly they could adapt to a wide range of salinities as do some modern fishes of the two surviving classes, or perhaps the late Devonian placoderms were anadromous, spending most of their lives at sea and at maturing entered fresh water to spawn and die.

It is interesting to note that no placoderms have been located in the Upper Parke Siltstone to date. This could be due to increasing salinity and frequent desiccation, since this part of the column contains abundant salt casts and mud cracks which is in contrast to the Lower Parke
Siltstones and Harajica Sandstone Member where they are infrequent.

The almost complete absence of higher plants, especially the lycopod *Leptophloeum australe*, from central Australia is a distinct contrast to eastern and western Australia where it is widespread in both non-marine and near shore marine sediments. The distribution of *Leptophloeum* suggests that these plants probably required low-lying areas near the coast where there was a constant supply of water. Their geographic spread inland was probably primarily controlled by the availability of water with the correct salinity. Other climatic factors such as low humidity, hot drying winds and wide fluctuations in temperature probably also controlled the spread of these early plants.

The presence of spores in the basal Upper Parke Siltstone, however, indicates that at least some plant life was probably present during this period since even wind borne pollen and spores generally travel less than 160 Km (Andrews, 1961, p.446). The source area for hydrophilous transportation of the spores is considered to be local and hence does not alter the fact that plants must have been present. The spore *Triletes*, and hence probably *Leiotriletes*, is generally attributed to the lycopods (Andrews, 1961, p.252), thus possibly some lycopods spread to this area even though *Leptophloeum* has not been found.
The presence of *Bothriolepis* and spores in the lower third of the Parke Siltstone suggests that during this period of deposition the climate was equable and there was a continuous aqueous environment within the region. This, together with the evidence supplied by sedimentary structures, indicates that the Parke Siltstone probably accumulated as a lacustrine deposit with the Harajica Sandstone in this area, extending in the form of a delta. In the Upper Parke Siltstone period of deposition conditions must have become increasingly arid or seasonal since the sedimentary structures such as mud cracks and pseudomorphs of halite became increasingly common.

**PALEOLATITUDE**

The paleolatitude of the Amadeus Basin in the Upper Devonian to Lower Carboniferous would have been between 15°S and 25°S (Irving, 1964) with latitudinal lines being subparallel to their present position or possibly slightly oblique (west-north-west to east-south-east). A set of twelve samples were collected from the red-brown Hermannsburg Sandstone at the Waterhouse and James Ranges by the author and their paleomagnetic properties were measured by Dr F.H. Chamalaun, Department of Geophysics, A.N.U. The results were not encouraging and the magnetism reflected post-depositional and post-folding (probably Tertiary) weathering profiles (Chamalaun, pers. comm., 1967).
ENVIRONMENT OF DEPOSITION

Details of the most probable environments of deposition for the sediments of the Pertnjara and Finke Groups have been discussed in Chapter 3 and only a summary is presented here.

At the start of deposition of the Parke Siltstone the area to the north of the Amadeus Basin probably had a subdued relief except perhaps for low epirogenically uplifted areas to the north-east. Most of the southern margin of the basin probably also had a subdued relief except in the south-east where local uplift and deep erosion provided the sedimentary, metamorphic and granitic detritus for the Polly Conglomerate. This uplift in the east may have also formed the barrier to eastward flowing drainage from the Amadeus Basin, thus initiating the period of internal drainage with its associated saline environment. During the deposition of the Parke Siltstone uplift occurred to the north of the Amadeus Basin and this initiated the deposition of the southward thinning wedge of the Harajica Sandstone Member. In its southern extremities this member changed from fluvial to deltaic in character and it merged laterally with the lacustrine facies of the Deering and Dare Siltstone Members. During the period of deposition of the Dare Siltstone Member the climate must have become more arid and evaporation from the lake must have exceeded both precipitation and inflow. The presence of halite pseudomorphs indicates at least seasonal aridity and that the lake(s)
probably had no outlets (Borchert and Muir, 1964).

Further uplift to the north of the Amadeus Basin initiated another southward transgression - the fluvial facies of the Hermannsburg Sandstone. Current directions indicate that drainage was predominantly to the south-east where the Hermannsburg Sandstone becomes more silty and probably merges into the lacustrine environment of the Horseshoe Bend Shale. The latter also contains halite pseudomorphs and beds of gypsum and probably represents another centre of internal drainage.

The deposition of the Brewer Conglomerate was initiated by strong uplift and folding on the present northern margin of the Amadeus Basin. It probably accumulated as a series of alluvial fans spreading southwards into the pebbly sandstones of the Ljiltera Member and further south-eastwards into the upper Horseshoe Bend Shale, Idracowra Sandstone and Hakea "Formation". During this period of deposition the Amadeus Basin was probably connected to eastern Australia to form part of a very large drainage network.

The lack of preserved fauna and flora during this period of Upper Devonian-?Lower Carboniferous sedimentation can be attributed to the following conditions: - a saline lacustrine environment subjected to periodic desiccation during the deposition of the Parke Siltstone; rapidly migrating non-cohesive sandy river channels, probably ephemeral or at least dry for part of the year, during
the deposition of the sandstone units; the probable lack of adaptation of most Upper Devonian flora to inhabit sandy areas subjected to periods of desiccation; and the probable lack of connection with eastern Australia, during the period of deposition of the Dare Siltstone Member and the Horseshoe Bend Shale.

COMPARISON WITH OTHER AREAS OF DEVONIAN DEPOSITION IN AUSTRALIA

The Pertnjara and Finke Groups are very similar in lithology, inferred depositional environment, and stratigraphic position to the Mt Eclipse Sandstone of the Ngalia Basin (Wells et al., 1968) and the upper parts of the Dulcie Sandstone of the Georgina Basin. The Mt Eclipse Sandstone was described by Cook and Scott (1967) as a coarse continental deposit consisting predominantly of pebbly litharenites. It is characterized by frequent lateral variations in lithologic facies, contains Lower Carboniferous plant fossils near the top of the formation, and was probably deposited in fluvial and piedmont environments. It was derived from a south-western source area and probably represents a synorogenic deposit equivalent to the Pertnjara Group, especially the Brewer Conglomerate, although it was probably deposited at a greater distance from the source area. If this correlation is correct it provides evidence that the deposition of the Pertnjara Group probably also extended up into the Lower Carboniferous.
The Dulcie Sandstone is a cleaner, more quartz rich sandstone than the Mt Eclipse Sandstone or Pertnjara Group (Smith, 1967) and it probably represents deposition at a greater distance from the Arunta Block source area. The upper part of the Dulcie Sandstone has been dated as Upper Devonian (Hills, 1958).

Further eastwards the Pertnjara and Finke Groups can probably be correlated with the Buckabie Formation in the Adavale Basin. This unfossiliferous clastic sequence of red sandstones, conglomerate, and variegated mudstones and siltstones was probably deposited on an extensive flood plain incorporating saline lake basins (J.J. Tanner, 1968). The Buckabie Formation is very similar to the fluviatile and partly paralic sequence of the Mt Wyatt Beds in the Drummond Basin which drained northwards to the coast. The Buckabie Formation is also similar to the Upper Devonian - Lower Carboniferous Mulga Downs, Hervey and Cocoparra Groups of western, central and south-western New South Wales (Conolly, 1965a,b, 1966). These groups all consist of poorly sorted clastic material varying from conglomerate to sandstone and siltstone. In many areas fluvial cycles can be recognized and the environment appears to be one of widespread terrestrial deposition. In south-western New South Wales the base of the groups is usually erosional and is overlain by coarse conglomeratic deposits, probably piedmont fans e.g. the Barrat Conglomerate, of the Cocoparra Group (Conolly, 1966), and the Rock Holes Member of the Snake River Sandstone (Warris, 1967). Such conglomerates
are derived from local source areas to the south-west but are rather reminiscent of the Polly Conglomerate deposited in a similar structural situation 700 Km to the north-west in the south-eastern Amadeus Basin. The Hervey Group contains fining-upwards sequences and was deposited probably by meandering streams on a broad floodplain (Conolly, 1965a,b). General drainage was to the south-east and east where the terrestrial and paralic Catombal Groups give way to the marine Lambie Group. Continuous marine sedimentation in the Devonian and Lower Carboniferous was confined to the geosynclinal sediments of the Tamworth Trough (Crook, 1964, 1968).

The Grampians Basin probably existed as a separate area of terrestrial deposition, with sediment movement from north to south during the deposition of the Upper Devonian to Carboniferous Grampians Group (Spencer-Jones, 1965).

Thus the Upper Devonian of eastern Australia is characterized by the deposition of widespread terrestrial and marginal paralic deposits (Brown et al., 1968). In New South Wales terrestrial sediments were probably deposited as extensive sheets by a river system, or systems, which discharged into shallow marine environment to the east. Fluvial current directions are shown as general flow lines on the paleogeographic reconstruction of Australia (Fig. 5.12, modified after Brown et al., 1968 and Crook, 1968).
In western Australia Upper Devonian to Carboniferous sediments are known from the Carnarvon, Canning and Bonaparte Gulf Basins. These areas are all dominated by shallow water marine sediments including sandstones, shales and carbonate reef complexes. The landward margins of these basins contain wedges of deltaic and coarse fluvial sediments. The latter become more prominent higher in the sequence but are not as widespread as the equivalent facies in eastern Australia.

Thus the Upper Devonian–Carboniferous sediments of the Amadeus Basin can be linked with those of eastern Australia along extensions of the paleocurrent systems recorded in the Amadeus Basin. On paleontological evidence these sediments are probably more closely related to those of New South Wales than to those of Queensland.
REFERENCES


———, 1965: The sampling error in modal analysis. 
_Am. Miner., 50_, pp. 196-211.


BISCAYE, P.E., 1964: Distinction between kaolinite and chlorite in recent sediments by X-ray diffraction. 
_Am. Miner., 49_, pp. 1281-1289.

———, 1965: Mineralogy and sedimentation of recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans. 


BLISENBACK, E., 1952: Relation of surface angle distribution to particle size distribution on alluvial fans. 

_Sedimentology, 4_, pp. 225-245.


———, in press: The Illamurta Structure and its position on an 
important central Australian geologic trend.  

———, and SCOTT, I.F., 1967: Reconnaissance geology and petrography, 
Ngalia Basin, Northern Territory.  

CRAIN, I.K., in press: Computer interpolation and contouring of 
non-equispaced two-dimensional data. Geoexplor.

CROOK, K.A.W., 1960: Classification of Arenites.  

———, 1964: Depositional Environments and Provenance of Devonian 
and Carboniferous Sediments in Tamworth Trough, N.S.W.  

———, 1967: Tectonics, climate and sedimentation.  
Preprints, 7th International Sedimentological Congress.

———, 1968: Upper Devonian sedimentological provinces in 
eastern Australia and their controlling factors.  
(In Oswald, D.H. (Ed.) International Symposium on the 
Calgary, Alberta.)

———, 1969: Weathering and roundness of quartz sand grains.  
Sedimentology, 11, pp. 171-182.

———, and COOK, P.J., 1966: Gosses Bluff Diapir, Crypto-Volcanic 

DAL CIN, R., 1968: Climatic significance of roundness and percentage 
DAS, K.N., 1968: Westward shift in the course of the Kosi. 

DAVIES, D.K., and EHRLICH, R., 1966: An objective evaluation of the 
*sedimentary structural complexity of a fluvial bar.* 
*Sedimentology*, 7, pp. 179-190.

DEERE, R.E., and BAYLISS, P., 1969: Mineralogy of the Lower Jurassic 
pp. 133-154.

DIESSEL, C.F.K., 1966: The determination of the direction of transport 
of fluviatile arenites by orientation analysis of the detrital 

DIEHTZ, R.S., 1967: Shatter cone orientation at Gosses Bluff Astrobleme. 

DOEGLAS, D.J., 1962: The structure of sedimentary deposits of 

———, 1968: Grain-size indicies, classification and environment. 

DZULYNSKI, S., and WALTON, E.K., 1965: Sedimentary features of 
flysch and greywackes. In *Developments in Sedimentology*, 7, 
Amsterdam, Elsevier.

EARDLEY, A.J. and GVOSDETSKY, V., 1960: Analysis of Pleistocene 
core from Great Salt Lake, Utah. 

EAST, J.J., 1889: Geological structure and physical features of 

ELLIOTT, R.E., 1965: A classification of subaqueous sedimentary 
structures based on rheological and kinematic parameters. 
*Sedimentology*, 5, pp. 193-209.

FAGERSTROM, J.A., 1967: Development, flotation, and transportation 
of mud crusts - neglected factors in sedimentology. 
*J. sedim. Petrol.*, 37, pp. 73-79.

FAHNENSTOCK, R.K. and HAUSHILD, W.L., 1962: Flume studies of the 
transport of pebbles and cobbles on a sand bed. 


*Sedimentology*, 6, pp. 73-93.


———, and WARD, W.C., 1957: Brazos River Bar: A study in the significance of grain size parameters.  

FORMAN, D.J., 1966: Regional geology of the south-western margin, of the Amadeus Basin, central Australia.  


———, and HANCOCK, P.M., 1964: Regional geology of the southern margin, Amadeus Basin Rawlinson Range to Mulga Park Station.  


*J. Geol.*, 66, pp. 394-416.

———, 1967: Dynamic processes and statistical parameters compared for size frequency distribution of beach and river sands.  
*J. sedim. Petrol.*, 37, pp. 327-354.


HENNINGSEN, D., 1967: Crushing of sedimentary rock samples and its effect on shape and number of heavy minerals. Sedimentology, 8, pp. 253-255.

HILLS, E.S., 1959: Record of Bothriolepis and Phyllolepis (Upper Devonian) from the Northern Territory of Australia. J. Roy Soc. N.S.W., 92, pp. 174-175.


*Sedimentology, 2*, pp. 115-121.


———, and WALKER, R.G., 1968: Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts. 
*J. sedim. Petrol.*, 38, pp. 971-984.

KARCZ, I., 1967: Harrow marks, current aligned sedimentary structures. 
*J. Geol.*, 75, pp. 113-121.


*Soil Sci.*, 81, pp. 111-120.

KITTLEMAN, L.R., 1964: Application of Rosin's distribution in size-frequency analysis of clastic rocks. 

KLOVAN, J.E., 1966: The use of factor analysis in determining depositional environments from grain size distributions. 

*Sedimentology, 10*, pp. 57-69.

*Report for Magellan Petroleum (N.T.) Pty Ltd (unpubl.).*

KRYNINE, P.D., 1946: The tourmaline group in sediments. 
*J. Geol.*, 54, pp. 65-87.

KUNZE, G.W., 1955: Anomalies in the ethylene glycol solvation technique used in X-ray diffraction. 
*Clays Clay Miner.*, 3, pp. 88-93.


LENK-CHEVITCH, P., 1959: Beach and stream pebbles.  
J. Geol., 67, pp. 103-108.


LESLIE, R.B., 1960: The geology of the Southern part of the Amadeus Basin; Northern Territory. 
Rep. for Frome-Broken Hill Co. Pty Ltd, 4300-G-28 (unpubl.).

LESLIE, W., 1965: The Upper Palaeozoic deposits of the Amadeus Basin. 


Z. Geomorph, 9, pp. 82-114.
In G. H. Dury (Ed.) Essays in Geomorphology Heinemann,
———, 1967: Denudation Chronology in Central Australia:
Structure, Climate, and Landform Inheritance in the
Alice Springs Area. In J. N. Jennings and J. A. Mabbutt (Eds)
Landform Studies from Australia and New Guinea.

McKEE, E.D., 1966: Structures of dunes at White Sands National
Monument, New Mexico (and a comparison with structures of
dunes from other selected areas): Sedimentology, 7, pp. 3-69.
———, CROSBY, E.J., BERRYHILL, H.L., 1967: Flood deposits,
Bijou Creek, Colorado, June 1965.

MACKENZIE, R.C., 1948: Complexes of clays.

MACKIN, J.H., 1963: Rational and empirical methods of investigation

MADDOCK, T., 1969: The behavior of straight open channels with
movable beds.

MADIGAN, C.T., 1932a: The geology of the Western MacDonnell Ranges,
Central Australia.
———, 1932b: The geology of the Eastern MacDonnell Ranges.


MAREL, H.W. van der, 1966: Quantitative analysis of clay minerals
and their admixtures.

MECKEL, L.D., 1967: Tabular and trough cross-bedding: comparison
of dip azimuth variability.
J. sedim. Petrol., 37, pp. 80-86.
*Clays Clay Miner.*, 7, pp. 317-327.


O'BRIEN, D.C., 1968: Cubic casts as indicators of "top and bottom" in the Shields Formation (Precambrian Belt Supergroup). 

PACKHAM, G.H., 1955: Volume-, weight-, and number-frequency analysis of sediments from thin section data. 
*J. Geol.*, 63, pp. 50-58.


PARKINSON, W.D., 1965: Errors and limitations of the magnetic compass. 


PASSEGA, R., 1957: Texture as a characteristic of clastic deposition. 


PICARD, M.D. and HIGH, L.R., 1968: Sedimentary cycles in the Green River Formation (Eocene), Uinta Basin, Utah. 


POWER, P.E., 1969: Clay mineralogy and paleoclimatic significance of some red regoliths and associated rocks in Western Colorado. 


ROCHOW, K.A., 1965: The geology and occurrence of groundwater, Finke 1:250,000 Sheet area (S.G. 53/6), Northern Territory. 

*J. Geol.*, 61, pp. 114-132.


*Geol. Mag.*, 97, pp. 409-421.


———, 1968: Speculations concerning paleohydrologic controls of terrestrial sedimentation.


SELLEY, R.C., 1968: A classification of Paleocurrent models.
*J. Geol.*, 76, pp. 99-110.


———, 1966: Resistance to flow in alluvial channels.

SKINNER, H.C.W., 1963: Precipitation of calcian dolomites and magnesian calcites in the southeast of South Australia.
SLATYER, R.O., 1962: Climate of the Alice Springs area. 


SMITH, R.E., 1966: Grain size measurement in thin section and in 

SMOOT, T.W., 1960: Clay mineralogy of Pre-Pennsylvanian Sandstones 
and Shales of the Illinois Basin Part 1. - Relation of 
permeability to clay mineral suites. 

River, Texas. A study in particle morphogenesis. 

SORBY, H.C., 1908: On the application of quantitative methods to 
the study of the structure and history of rocks. 

SPENCER-JONES, D., 1965: The geology and structure of the 
Grampians area, western Victoria. 

SRIRAMADAS, A., 1957: Diagrams for the correlation of unit cell 
edges and refractive indices with the chemical composition 

STAATZ, M.H., MURATA, K.J. and GLASS, J.J., 1955: Variation of 
composition and physical properties of tourmaline with its 

STAUFFER, P.H., 1966: Thin-section size analysis: a further note. 
Sedimentology, 7, pp. 261-263.

STOKES, W.L., 1953: Primary sedimentary trend indicators as 
applied to ore-finding in the Carrizo Mountains, Arizona 

STRAKHOV, N.M., 1953: Diagenesis of sediments and its significance 
for sedimentary ore formation. 


Walther, J., 1893-4: Eintleitung in die Geologie als historische Wissenschaft; Beobachtungen über die Bildung der Gesteine und ihrer organischen Einschlüsse, bd 1: Jena, G. Fischer.


* * * *