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This book is the first major synthesis of our present knowledge of Papua New Guinea's land forms, their distribution, origin and development, the processes that acted on them - and are continuing to act. Such things as volcanic eruptions, earthquakes, landslides, and subaerial and subsurface erosion processes, changing river systems and their role in the formation of plains can be better understood by those who read this book: those working in Papua New Guinea and directly concerned with its development and environment such as geologists, agriculturalists, engineers and conservationists, and scholars generally concerned with problems of the geomorphology of the humid tropics.

Particularly valuable are the illustrations, which cover a wide range of land forms and illustrate vividly the often dramatic topography.
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Dr Ernst Löffler has a Ph.D. from Heidelberg University and is now a Senior Research Scientist with the Division of Land Use Research (formerly Land Research), CSIRO, which he joined in 1967. In that year he spent several months in Nigeria working on a research program concerned with thematic mapping of African environments. In addition he worked for several months in 1973 as a consultant on terrain evaluation in Iran and in 1975 was a visiting Senior Lecturer with the University of Papua New Guinea. With the Division of Land Research he took part in several land resource surveys and other research programs in Papua New Guinea during which he became familiar with most parts of the country. This book is based on his work there.

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Geomorphology of Papua New Guinea
Geomorphology of Papua New Guinea

Ernst Löffler

The Commonwealth Scientific and Industrial Research Organization, Australia in association with the Australian National University Press
Canberra 1977
To the memory of W. Behrmann and K. Sapper who under most difficult conditions pioneered geomorphological work in Papua New Guinea and to the new generation of Papua New Guinean geomorphologists who will continue this research in their newly independent nation.
Preface

Papua New Guinea has remained in the backwaters of scientific exploration and research for a considerable time and much of its fame in the world still appears to rest on this touch of unexploredness and its reputation of being the 'last unknown'. However, in the last two decades remarkable progress has been made in spite of the great difficulties imposed by the often forbidding terrain, dense rain forest and adverse climatic conditions. Not only have the 'blank areas' on the maps completely disappeared but the detail of knowledge has also greatly increased and particularly in the last decade work has gradually shifted from a purely exploratory and reconnaissance level to detailed scientific research.

Much of the credit for a quick breaking down of the frontiers of our knowledge on the landforms of this country is due to the availability of aerial photographs and other remote sensors. In addition the ease of access has been vastly improved not only through an expanding road network but also with the use of aircraft and helicopter transport. While researchers certainly have not set foot on every square metre or square kilometre of ground and are unlikely to do so for some time to come, we have now a reasonably complete knowledge of the surface features of this country. This is not to say that this also applies to our understanding of the landforms and the various interrelationships between them and with other environmental factors. In fact there is a great lack of quantitative information and detailed studies on these problems but the beginning has been made.

This book has resulted directly from the involvement of the Division of Land Use Research, CSIRO, in land resources mapping and inventory in Papua New Guinea which was undertaken at the request of and through the funding by the Papua New Guinea Administration and more recently the Papua New Guinea Government.

Throughout the drafting of the manuscript I have greatly benefited from discussions with and suggestions and criticism from colleagues both within and outside CSIRO.

I am particularly indebted to Dr J.N. Jennings who spent a great deal of effort and time on the manuscript and suggested many improvements and clarifications in matters of terminology, interpretation and style. Dr R. Blong, Mr H.A. Haantjens and Drs E. Bird, D.H. Blake, R.W. Galloway and C. Pain have also read drafts of the manuscript or parts of it and their advice and criticism is gratefully acknowledged.

I also wish to thank the staff of the Division of Land Use Research for editing and for transforming my poor sketches into
respective drawings and the Australian National University Press for bringing the book to publication.

My thanks are also due to the many Papua New Guineans who unwittingly helped with this book. They accompanied me into the field and helped me carry out all the strange things I did in spite of my difficulties in explaining to them the purpose of my work.

The following conventions regarding territorial names have been adopted. New Guinea refers to the entire island of New Guinea including the Indonesian part of Irian Jaya in the west and Papua New Guinea in the east. Papua has been used purely as a geographical term describing the southern and eastern parts of Papua New Guinea. If only the mainland of Papua New Guinea is referred to excluding the islands the term mainland is used. Highlands is also used in a purely geographical sense and not as a political and administrative unit.

It has not been possible, as originally intended, to include a geomorphological map of Papua New Guinea in this volume and some readers may find it difficult to locate certain landform types or topographic names. The distribution of most landform types has been shown on small text figures but since lettering had to be kept to a minimum not all topographic names mentioned are indicated. A geomorphological map of Papua New Guinea has been produced separately and is available from the Division of Land Use Research.

E.L.
Canberra, 1975
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Geomorphological research is at present very much engaged in quantitative approaches and it may seem anachronistic to publish a volume on the geomorphology of a country with few offerings on quantification, and a heavy emphasis on qualitative studies and descriptive regionalisation. This somewhat unbalanced treatment of the geomorphology of Papua New Guinea reflects, however, the present state of geomorphological research in this country where in spite of great progress in opening up this 'last unknown', accessibility is still a major factor in conducting geomorphological and geological research. About 75 per cent of the country is covered by primary rain forest, and here the research worker has been facing the problem of lack of visibility and vantage points from which to grasp the relationships of the different landform elements ever since he moved into these areas (cf. Behrman 1927). It is therefore not by accident that the few quantitative studies undertaken so far have been done in the easily accessible, mostly forest-free coastal areas (Chappell 1974a) and highlands, or rely heavily on measurements of aerial photographs (Simonett 1967; Williams 1972a). The important problems of slope development under primary rain forest have not been treated quantitatively although they are the most important land-forming processes in this country and there is certainly an urgent need for future work. Last but not least, the more qualitative approach reflects the author's own research work and data collecting on several land resources surveys, between 1968 and 1974, with the Division of Land Research, now Land Use Research, CSIRO.

As with any book dealing with a broad subject organisation has been a major problem. A strictly regional geomorphology such as Thornbury's (1965) *Regional Geomorphology of the United States*, dealing with each major region separately, did not appeal to the author in the case of Papua New Guinea, which is much smaller and in spite of great contrasts of landforms a more homogeneous spatial unit. On the other hand, a purely thematic treatment of the landforms and geomorphic processes without the important regionalisation did not appear a satisfactory solution either.

The author therefore compromised in combining the two approaches although this inevitably resulted in some repetition, especially in Chapters 2 (Regional Geomorphology) and 3 (Landform Types).

The aim of this book is to give a synthesis of our present knowledge of the landforms of Papua New Guinea, and their distribution and development, as part of a natural resources inventory. It is also hoped that this will provide a useful study of the geomorphology
of the perennially humid tropics, an area much neglected in geomorphic research, which tends to focus attention much more on the seasonally humid tropics (Tricart 1972; Thomas 1974).

Much of the book rests on the work of the Division of Land Use Research but other relevant work has of course been used. The author has tried to make it as comprehensive as possible regarding the modern literature on Papua New Guinea geomorphology, but older publications have usually not been quoted if they have been superseded by or incorporated in more modern work. This applies particularly to the numerous papers written before the first world war.

It is difficult to pinpoint the start of geomorphological research in Papua New Guinea. A certain amount of valuable geomorphological information was collected in most of the numerous exploratory sea and land expeditions which started with the sea voyages in the sixteenth century and ended with the final discovery of the New Guinea highlands in the 1930s. It would be beyond the scope of this book to go into any detail on the history of exploration and the achievements of the various explorers, and the reader is referred to Wichmann (1909, 1910, 1912) if he seeks detailed information, or for a briefer more popular account to Souter (1963).

It is probably fair to give the credit for the first genuine geomorphological research in what is now Papua New Guinea to K. Sapper and W. Behrmann, two German geographers who took part in several expeditions in the former German territories. Both published extensively on the geomorphology of Papua New Guinea and the humid tropics in general. Sapper worked primarily in the Bismarck Archipelago (New Ireland [formerly New Mecklenburg], New Britain [formerly New Pommern] and Bougainville) where he undertook several expeditions. As well as giving regional accounts (1909, 1910a, b) he published papers on erosion processes in the humid tropics (1914) and on New Guinea volcanoes (1917a, b, 1921). In addition Sapper wrote a book on the geomorphology of the perennially humid tropics (1935) which contains a great deal of information on the geomorphology of Papua New Guinea and has remained a standard reference.

Behrmann took part in the well-known Kaiserin Augusta Fluss (Sepik River) Expedition in 1912/13, the largest and most wide-ranging scientific expedition ever undertaken by the Germans in their New Guinea territories. The expedition lasted 20 months and though its main task was to map and survey the main river, extensive botanical, zoological and ethnological collections were made and field research undertaken. The map of the Sepik River basin at the scale of 1:250,000 was an admirable achievement of accuracy, particularly in view of the great difficulties of the terrain, and only very recently with the assistance of aerial photography have improved maps become available. During the expedition Behrmann undertook three larger traverses on land, advancing from the larger tributaries of the Sepik southwards into the central ranges. On his first land expedition he was in viewing distance of the grass-covered lower volcano-alluvial terraces of the Mt Hagen volcano, which would have afforded easy access to the heavily populated highlands,
but he had to turn back for lack of supplies. He published several papers and books on his observations, the most important of which are ‘Der Sepik und sein Stromgebiet’ (Behrmann 1917), ‘Das westliche Kaiser-Wilhelms-Land’ (Behrmann 1924), ‘Die Oberflächenformen im feuchteissn Kalmenklima’ (Behrmann 1927) and ‘Die Insel Neuguinea: Grundzüge ihrer Oberflächengestaltung nach den gegenwartigen Stande der Forschung’ (Behrmann 1928) which is a first synthesis of the geomorphology of the island of New Guinea. With the departure of the Germans from their New Guinea territories after World War I interest in scientific exploration decreased substantially and much of their work was not utilised or followed up and therefore never received the credit it deserved.

In Papua no comparable work in geomorphology was done before World War I, but geological exploration under the government geologist E.R. Stanley was very active.

Between the world wars attention was focused on the discovery of gold, which provided the incentive for several expeditions into the interior, eventually leading to the discovery by Europeans of the extensive, densely populated highlands.

In the coastal areas search for oil led to more detailed exploration of formerly little known areas but the emphasis of the work was more on geology and in general little attention was paid to geomorphic problems.

In the late 1930s S.W. Carey (1938) wrote another synthesis of the geomorphology of New Guinea, which like Behrmann’s (1928) paper has remained a classic study of New Guinea landforms. Carey describes the essential morphostructural elements of New Guinea and their interrelationships, and although much more information is now available, little change to his basic outline has been necessary.

Although World War II stopped all civil exploration it gave great impetus to a rapid compilation of all available geographical information and to the production of topographical maps making extensive use of aerial photographs. A series of regional reports primarily dealing with the Japanese-occupied north-east of Papua New Guinea was compiled (complete list in Manser and Freeman 1970).

Geomorphological research was stimulated in the early 1950s when scientific teams of the Division of Land Research, CSIRO, commenced a series of land resources surveys based on techniques gained from work in the Northern Territory of Australia. Systematic mapping and description of large areas previously unknown or little known was carried out using aerial photographs as a major source of information. The method used in delineating the different spatial units of the landscape became known as the land system concept (Christian and Stewart 1953, 1968), which defines a land system as ‘an area or group of areas throughout which there is a recurring pattern of topography, soils and vegetation’. As the mapping of the land systems is largely based on the recognition of landforms and other surface features on aerial photographs, geomorphology has played a major role in the mapping and definition of the different land systems and the land system maps
can to a large degree be regarded as geomorphological maps (Mabbutt and Stewart 1963; Blake and Paijmans 1973; Löffler 1974a, b). A total of sixteen land resources reports with accompanying maps have been published on Papua New Guinea between 1964 and 1975, and the latest three of these cover the major aspects of land resources, geomorphology, vegetation and land use for the whole of Papua New Guinea (Löffler 1974b; Paijmans 1975; Bleeker 1975).

Besides their contributions to the reports, CSIRO geomorphologists have published numerous papers on specific geomorphic topics such as weathering, denudation, morphoclimatic zonation, alluvial plains, volcanoes, Pleistocene glaciation, and karst, to name a few (Bik 1967, 1972; Blake and Bleeker 1970; Blake and Löffler 1971; Blake and Ollier 1971; Haantjens and Bleeker 1970; Jennings and Bik 1962; Löffler 1970, 1971, 1972a, 1974c; Mabbutt and Scott 1966; Reiner 1960; Reiner and Robbins 1964; Ruxton 1966a, 1967, 1968; Ruxton and McDougall 1967; Speight 1965a, b).

During the last five years geomorphological research in Papua New Guinea has gained momentum as a greater number of research workers have become involved in geomorphological studies. Most active were members of the Australian National University who undertook several detailed studies on karst, Quaternary history, coastal landforms and landslides (Williams 1972a; Williams and others 1973; Pain 1972b, 1973; Chappell 1974a). At present geomorphic research is still very much in progress and it has gained new stimulus with the establishment of the University of Papua New Guinea.
The geological mapping and exploration of Papua New Guinea have made remarkable progress in the last decade in spite of the great difficulties imposed by forbidding terrain, dense rain forest cover and adverse climatic conditions. Not only have the ‘blank areas’ completely disappeared from the geological maps but the detail of knowledge of large areas has greatly increased. Gravity, magnetic and seismic surveys have been undertaken in addition to standard geological surveying and mapping. Much of this work has yet to be published, but most information is available in the offices of the Bureau of Mineral Resources, Geology and Geophysics in Canberra and from the Geological Survey of Papua New Guinea in Port Moresby.

However, many of these data have been summarised in the 1:1,000,000 Geological Map of Papua New Guinea published in 1972 (Bureau of Mineral Resources, Aust. 1972) and in a paper by Bain (1973) which forms a valuable supplement to the map.

**STRUCTURAL CONTEXT**

New Guinea and its associated smaller islands are situated between and are part of two major crustal elements, the continental, relatively stable land mass of Australia to the south and the deep ocean basin of the Pacific to the north (Fig. 1). They include part of the highly mobile zone of the earth’s crust that surrounds the Pacific ocean, characterised by young folded and faulted mountains, con-

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**Figure 1**

Main crustal elements and major fault pattern

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Major faults

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Present Main plate boundary

---
temporary volcanic and seismic activity and curved chains of islands and oceanic rises, known as island arcs. According to plate tectonics concepts, the New Guinea area has lain in the zone of interaction between the northward-moving Australian continental plate and the westward-moving Pacific plate since about Cretaceous times and its structural development is therefore the direct result of the interaction of these two crustal elements. As the Australian plate is moving north and the Pacific plate is moving west the area is structurally complex and strike slip movements and possible

Figure 2
Generalised geology

**CAINozoic**

**Quaternary**
- Alluvium, raised coral
- Subaerial lavas and pyroclastics

**Pliocene**
- Marine and terrestrial fine-grained sediments
- Limestone

**Miocene**
- Limestone

**CAINoZoic**

**Quaternary**
- Greywacke, siltstone and conglomerate, volcanics and some limestone
- Marine volcanics (slightly metamorphosed in places)
- Basic and ultrabasic igneous rocks

**Triassic**
- Dacite volcanics

**Permian**
- Low grade metamorphosed clastic sediments intruded by granite and granodiorite

**Cretaceous and Jurassic (and some Tertiary)**
- Siltstone, shale, quartz, sandstone, conglomerate (unmetamorphosed)
- Greywacke, siltstone, conglomerate interbedded with marine volcanics (metamorphosed)
rotation of small lithospheric plates are the results (Davies and Smith 1971). Zones of seismic activity show that there are probably several small lithospheric plates as well as two major plates (Fig. 1). The present contact between the two major plates is situated along the north coast of New Guinea and along the northern margin of the Solomon Sea (Denham 1969). During most of the Tertiary, however, the contact was probably further south, along the major fault zone on the south side of the New Guinea mobile belt.

**STRUCTURAL REGIONS**

The complex geology of New Guinea is best described in terms of its major structural regions which, as will be shown later, provide the framework for most of its physiographic regions. The distribution of the major stratigraphic units is shown in Fig. 2. These principal structural regions are roughly from south to north (after Thompson and Fisher 1965; Bain 1973) (Fig. 3).

1. The Fly Platform
2. The Papuan Fold Belt
3. The Aure Trough
4. The Kubor Anticline
5. The New Guinea Mobile Belt
6. The East Papua Ultrabasic Belt
7. The South-east Papua Volcanic Province
8. The Cape Vogel Basin
9. The Sepik-Markham Depression
10. The Finisterre-New Britain Volcanic Arc
    a The Palaeogene volcanic arc
    b The Quaternary volcanic arc
11. The Toricelli-Bewani Ranges
12. The Bougainville-New Ireland Volcanic Arc
    a The Palaeogene volcanic arc
    b The Quaternary volcanic arc

*Figure 3*

Structural regions (after Thompson and Fisher 1965 and Bain 1973)
The Fly Platform

The Fly Platform (Bain 1973), also known as the Fly Digoel Shelf (Glaessner 1950), is an extensive stable shelf area comprising the vast plains and lowlands associated with the Fly and Strickland Rivers and the limestone plateaux to the north-east of the lowlands. Structurally it is part of the stable Australian shield which extends from Cape York underneath Torres Strait. Palaeozoic granitic basement forms rocky outcrops on several Torres Strait islands and also near Mabaduan on the New Guinea mainland. The granitic rocks are overlain by a largely horizontal sequence of Mesozoic and Cainozoic sediments, and since the Palaeozoic the shelf has been subject to only minor epeirogenetic movements with some gentle warpings (Fig. 4a). Only in the north-east where the shelf area borders the tectonically very active Papuan Fold Belt do some major faults and folds occur (Fig. 4b). Situated at the north-eastern edge of the shelf the Bosavi volcano, which has several adventive
cones extending south-west into the lowlands, and the Aird Hills, the erosional remnant of a volcano, are also evidence of increased instability at the outer edge of the shelf. The sediments overlying the granitic basement are largely dominated by limestone of Miocene to Pliocene age ranging in thickness from about 500 m in the south to about 2000 m in the north. This limestone forms spectacular karst scenery in the north-east but over most of the region it is covered by a sequence of terrestrial sediments, Pliocene to Pleistocene in age. These range in thickness from 150 to 300 m except in the Strickland Basin to the north-west of Bosavi where over 3000 m of Pliocene terrestrial sediments are present.

The Papuan Fold Belt

The Papuan Fold Belt (Bain 1973), also known as the Papuan Geosyncline (Thompson and Fisher, 1965), lies north and north-east of the Fly Platform and consists of a series of subparallel north-westerly to westerly trending folds and faults. Most of the fold belt is underlain at depth by the folded and upturned margin of the crystalline basement of the Australian continental plate (Smith 1965) except in the east where the fold belt merges into the deep sedimentary basin of the Aure Trough. The predominant
outcrop rock is limestone, which covers well over half the area, forming spectacular karst landforms and strike ridges.

The character of the folds varies from monoclines and overthrust anticlines in the south-west in the proximity of the Fly Platform to tightly folded anticlines with broad synclines on moving north-east towards the highlands (Bain 1973) (Fig. 5). Jenkins and Martin (1969) interpret this pattern as diapiric, while Bain (1973) suggests that probably a combination of diapirism and associated gravity sliding and north-south compression caused the foreshortening of the sedimentary sequence in the fold belt. Plutonic and metamorphic rocks are notably absent but Quaternary volcanic cones, basaltic to andesitic in composition, form impressive landmarks in the central and eastern parts of the belt.

The Aure Trough

This unit extends east and north-east of the Papuan Fold Belt. It partly overlaps both the fold belt and the New Guinea mobile belt to the north and its margins are therefore difficult to delineate exactly. Some authors include the Aure Trough in the Papuan Fold Belt (Thompson and Fisher 1965), while others (Bain 1973) regard it as a separate unit, primarily because the general trend is northerly rather than north-westerly and because of the great thickness of sediments present in the trough. In the Aure Trough subsidence and deposition during the Miocene and Pliocene were followed by intense orogeny manifested in tight compressional folding and high-angle thrust folding. Some 16,000 m of sediments of which 15,000 m are Miocene to Pliocene in age were deposited in this trough. Most of these sediments are relatively coarse-grained greywacke sandstones derived from volcanic rocks fringing the basin to the west and north.

The Kubor Anticline

The Kubor anticline is a large anticlinal structure in the centre of the Papua New Guinea mainland. Palaeozoic basement and metamorphic and igneous rocks are exposed between non-metamor-
phosed sediments of the Papuan Fold Belt to the south-west and the metamorphosed and deformed rocks of the New Guinea Mobile Belt to the north-east. The broad structure of the anticline, first recognised by Rickwood (1955), has been mapped and described in detail by Bain and others (1975). The Kubor anticline is the largest and most eastern exposure of crystalline basement in Papua New Guinea and has remained an important land mass from late Permian times onwards except for a brief marine transgression in the upper Triassic.

The New Guinea Mobile Belt
This unit lies north of the Papuan Fold Belt and Kubor anticline, extends along the entire length of mainland Papua New Guinea and also includes the chain of islands at its eastern tail. It abuts against the Palaeozoic Australian plate along the major north-westerly to westerly trending Bismarck and Lagaip fault zone. In the east the contact with the Australian plate is probably situated offshore in the Coral Sea (Fig. 1).

The mobile belt partly overlaps the Aure Trough, which is situated like a wedge between the western and eastern halves. The eastern part is also known as the Owen Stanley metamorphic belt (Thompson and Fisher 1965).

The northern boundary of the mobile belt is obscured by the thick sedimentary cover of the Sepik-Ramu depression (Dow and others 1972) in the west, but is marked in the east by the Owen Stanley fault zone which forms a distinct lineament.

The mobile belt consists mainly of low grade metamorphics and acid to basic and ultrabasic plutons, most of which have been intruded along major fault planes.

East Papua Ultrabasic Belt
The east Papua Ultrabasic (Ultramafic) Belt forms a distinct structural unit on the north-east side of the Owen Stanley Ranges, from which it is separated by a narrow fault zone. The ultrabasic belt dips at angles of 10°–30° to the north-east and consists of a layer of peridotite, 4–8 km thick, overlain by 4 km of gabbro, which in turn is overlain by 4–8 km of basalt (Davies 1971). This ultrabasic belt is considered to represent a plate of oceanic crust (gabbro and basalt) and oceanic mantle (ultrabasics) emplaced by south-westerly thrust over the sialic metamorphic core of eastern Papua (Thompson and Fisher 1965; Davies 1971; Davies and Smith 1971).

The South-east Papua Volcanic Province
This unit lies south of the Owen Stanley Ranges and extends from Port Moresby eastwards to Milne Bay at the tail end of the mainland. It consists predominantly of submarine basaltic lavas flanked to the south by highly deformed early Tertiary sedimentary rocks including sandstone, siltstone, shale, conglomerate, limestone and chert, which form low foothills along the coast. The lavas and sediments overlie the metamorphics of the New Guinea Mobile Belt and were deposited on oceanic crust away from the influence
of clastic sedimentation from a large land mass (Thompson and Fisher 1965).

The Cape Vogel Basin
The Cape Vogel basin extends from Morobe to Cape Vogel and overlies the Papuan Ultrabasic Belt to the south-west. It consists of up to 4000 m of Middle Miocene to Pliocene sedimentary rocks which are overlain by Quaternary volcanics and alluvial sediments. Two volcanoes, Mt Lamington and Mt Victory, are still active.

Sepik-Markham Depression
The Sepik-Markham depression is one of the major structural elements of New Guinea, extending through the entire island from Geelvink Bay in Irian Jaya to the Huon Gulf in Papua New Guinea where it continues as a submarine depression leading into the New Britain Trench (Borch 1972). The depression, also known as the central intermontane trough (Carey 1938), has been a zone of relative subsidence since the later Tertiary and its margins are locally marked by steep fault scarps. The trough is now filled with terrestrial clastic sediments forming extensive alluvial plains and fans.

The Finisterre-New Britain Volcanic Arc
This structural unit includes New Britain, the Finisterre and Saruwaged Ranges of the New Guinea mainland, and the chain of volcanic islands off the north coast. It has the distinctive features of an island arc; it is arcuate in shape and is fronted at its convex side by an oceanic trench, the New Britain Trench, which descends to 7880 m and continues westward into the Markham Graben. It also has a well-defined seismic zone dipping steeply northwards to a depth of over 500 km (Denham 1969) and a belt of active volcanoes on its concave side. Structurally there are two concentric arcs. Southern and central New Britain and the Saruwaged and Finisterre Ranges form an older, outer arc of Eocene basic to intermediate volcanic rocks and associated intrusives overlain by middle to upper Miocene limestone. Within this arc is a younger arc of Quaternary volcanoes comprising the chain of volcanoes along the north coast of New Britain and off the north coast of Papua New Guinea.

The Toricelli-Bewani Ranges
This unit lies north of the Sepik basin and is commonly regarded as the western extension of the New Britain-Finisterre arc (Thompson and Fisher 1965). Bain (1973), however, stresses that this is not certain and points out that there are a number of distinct features which indicate that it may be a separate unit. For instance there are no Palaeogene volcanics underlying the Neogene sediments and there is no Quaternary volcanism. Instead the Toricelli-Bewani Ranges consist of granitic and metamorphic basement which is exposed in the central parts of the unit, flanked by a succession of sediments, predominantly sandstone and siltstone, which become progressively younger towards the outer zones of the mountains.
The Bougainville-New Ireland Volcanic Arc

This arc, also known as the West Melanesian arc, extends from Manus Island in the north-west to the Solomon Islands in the south-east. It is not as distinct an island arc as its southern neighbour, and lacks the associated trench. However, there is a well-defined seismic zone dipping to the north along much of its length and a string of Quaternary volcanoes off the north coast of New Ireland and on Bougainville. Like the New Britain-Finisterre arc its foundations represent a Palaeogene arc composed of pre-Miocene metavolcanics intruded by a variety of plutonic rocks and overlain by Miocene limestone, clastic sediments and younger volcanics. The Quaternary volcanic arc includes the extinct volcanoes of Tabar, Lihir, Tanga and Feni Islands off New Ireland and the active volcano Mt Bagana on Bougainville.

GEOLOGICAL HISTORY

The fundamental geotectonic force for the development of the present land mass of Papua New Guinea as well as its associated islands is the Pacific-ward drift of the Australian continental plate from its late Palaeozoic position in the vicinity of the South Pole, with southern New Guinea forming the north-eastern front of the advancing continent. The depositional environments of the Australian continental plate and the Pacific plates were basically different especially along their margins. Terrestrial and shelf type sediments were deposited on the Australian continental plate, whereas geosynclinal sediments associated with marine basic volcanics and intruded by large bodies of plutonics were accumulated north of the continental plate in the zone of interaction between it and the oceanic plates.

During most of the Mesozoic, the Palaeozoic basement block in the south formed an area of terrestrial deposition with the deposits thickening oceanwards to between 10,000 and 20,000 m. Deposition continued during the early Tertiary and early part of the upper Tertiary with the formation of shelf limestone, except at the Kubor anticline, which remained a basement high throughout most of the Mesozoic and Tertiary. This sediment cover became folded and to a lesser degree faulted in the Neogene. According to Bain (1973) the underlying tectonic forces were compression in the western part, and in the eastern part southerly gravity sliding and diapirism caused by uplift of the northern part of the basement block combined with north-south compression due to the northerly movement of the continental plate. Folding, faulting and thrusting were followed by uplift which reached its maximum during the late Pliocene and early Pleistocene. Associated with this was the formation of several large strato-volcanoes.

In the mobile belt north of the continental plate (Dow and others 1972) a great thickness of marine basic volcanics and volcanically derived sediments was deposited in the developing geosyncline from late Triassic time onwards. Periods of volcanism alternated with periods of deposition of continentally derived sediments. In the Cretaceous and Eocene the eugeosyncline reached its full develop-
ment with the deposition of a great thickness of shale, turbidites, marine volcanics, and limestone (Dow and others 1972). A period of further volcanism in the lower Miocene was followed in the middle Miocene by intense tectonism accompanied by the emplacement of large plutons, including the Bismarck Granodiorite, the Maramui Diorite and the April Ultramafics (Dow and others 1972; Bain and others 1975). Faulting was very intense, resulting in a complex series of anastomosing faults with probably large vertical throws. Associated with this tectonism was island arc volcanism resulting in the deposition of a narrow discontinuous belt of volcanics extending in a south-easterly direction from the upper Sepik basin to central Papua.

In east Papua, east of the continental plate, the development was somewhat different as here Jurassic-Cretaceous oceanic crust overrode the Mesozoic sediments in the lower Eocene, causing large scale metamorphism. This was followed by extrusion of great volumes of basalt onto the deep ocean floor to the south-west of the metamorphosed zone, probably as a result of opening of the Coral Sea by rifting (Davies and Smith 1971). The emergence of the east Papua land mass began in the upper Oligocene and continued throughout the Miocene. It was accompanied by intense erosion leading to the gradual filling up of the two flanking basins – the Aure Trough to the west and the Cape Vogel basin to the east. The sediments in these basins became folded, faulted and uplifted in the Pliocene.

The New Britain-Finisterre and the Bougainville-New Ireland volcanic arcs north-east of the mobile belt developed during the lower Tertiary. Along these arcs periods of volcanism and periodic uplift were interrupted by more stable periods during which reef limestone was formed. According to Bain (1973), the Finisterre part of the New Britain-Finisterre arc was welded onto the mainland during the late Neogene as a result of a southerly movement of the arc. This same movement also led to the separation of the Gazelle Peninsula from southern New Ireland and its joining onto New Britain.

The Torricelli-Bewani Mountains were part of the extensive geosyncline to the north of the mobile belt but were uplifted considerably later than the mobile belt. The geosyncline received a large amount of sediments from the rising mobile belt during the Neogene and was faulted, folded and uplifted in the late Pliocene and early Pleistocene. Uplift has continued to the present.

The Sepik-Ramu basin was formerly thought to be a fault-bounded depression (Behrmann 1917; Krause 1965). Recent investigations of the south Sepik area, however, show that the general trend of faults is not westerly, parallel to the mountain front, but north-westerly cutting across the Sepik plain and probably joining up with faults in the Torricelli-Bewani Mountains. Dow and others (1972) therefore assume that the Sepik basin is caused by downwarping due to compressional forces between the continental and Pacific plates.

The present land mass of Papua New Guinea thus evolved as the result of a series of complex events which are all basically linked
with the relative movements of the two interacting plates. The present land mass started to develop in the lower Miocene with the emergence of the New Guinea mobile belt, which has remained land ever since and has been subject to intense erosion. However, it was not until the upper Pliocene, which is only a few million years ago, that the framework of the present landscape became visible with the emergence of the Papuan Fold Belt to the south and the Torricelli-Bewani Mountains and the island arcs to the north. The zone of contact between the two plates has moved Pacific-wards and is now situated along the north coast of New Guinea and the south coast of New Britain. Similarly the zone of most intense volcanic and tectonic activity has moved from the mobile belt northwards and is now centred along the north coast and on New Britain. Raised coral platforms ranging in age from Pleistocene to Recent give evidence of continuing uplift in these areas, with maximum rates in the order of 3 mm per yr, which is quite a spectacular rate in geological terms. Rates of a similar magnitude were probably operative in other areas of Papua New Guinea in the past and this instability is undoubtedly one of the most important factors of the landform development in this country.
The difficulties of exact delineation of geomorphological regions are well known to geographers (Linton 1951), and any attempt at regionalisation will inevitably lead to controversy over the exact location of boundaries even though there may be general agreement about the existence of certain regions. Speight (1974) has attempted to define geomorphological regions rigorously by trying to group exactly defined land facets into higher order groups; his approach would involve very detailed mapping and meticulous description of the entire area and is not practicable at this stage. Furthermore, the concept has not yet been applied in practice. The present author feels that regionalisation is essentially a qualitative approach and does not lend itself to a rigorous definition because of the extreme complexity of regions. In Papua New Guinea, regionalisation is somewhat simplified because of the very youthfulness of the landscape where in most cases geomorphological regions closely follow relatively well-defined structural regions (Figs. 3, 6). Also important are local relief and altitude which are shown in Figs. 7 and 8.

The main geomorphological regions are the Fly Platform, the southern plains and lowlands, the southern fold mountains, the highlands, the northern metamorphic ranges, the eastern metamorphic ranges, the ultrabasic ranges, the Cape Vogel basin, the intermontane trough, the northern coastal ranges, the southern Bismarck Island arc, the northern Bismarck Island Arc, the east Papua islands.
Figure 7
Local relief classes

Figure 8
Altitudinal zones
and finally the Trobriand-Woodlark Island group.

The southern fold mountains, the highlands and the northern and eastern metamorphic ranges together are often referred to as the central ranges or central cordillera because they form an uninterrupted chain of mountain ranges. However, because of their heterogeneity and size, they are not treated as one geomorphological region.

THE FLY PLATFORM

The Fly Platform is the largest tract of low-lying country in Papua New Guinea, occupying nearly a third of the mainland. It extends from the south coast northwards to the foot of the central ranges. The region is more or less identical with the structural unit of the Fly Platform but excludes the Darai Limestone plateau. It covers some 100,000 km² and has a maximum width of 400 km. The southern part between the coast and the lower Fly River, sometimes referred to as the Oriomo Plateau (Carey 1938; Blake 1971), constitutes for the most part a gently undulating plain with local relief generally less than 10 m. Low-lying, flat areas are poorly drained or swampy and are thus in marked contrast to the surrounding slightly higher areas (Plate 1). Only in the south-east, where limestone occurs at shallow depths, and near Morehead, does the relief rise to about 30 m and a low hilly landscape is developed. Slopes are about 1°-2° and are smooth with no sudden breaks of slopes except at the river banks. Two interesting features of this area are an inselberg at Mabaduan and some karst development on the limestone.

Plate 1
The southern part of the Fly Platform, a gently undulating plain with numerous poorly drained or swampy areas.
The Mabaduan Hill is a low granitic inselberg — Papua New Guinea's only inselberg — rising from the Palaeozoic basement complex which extends from Cape York underneath Torres Strait to the southern shoreline of Papua New Guinea. Though very close to the present shoreline there is no evidence in the form of terraces or marine gravel that the Mabaduan Hill was ever subject to marine transgression. There are also no terraces or gravel above the present-day shoreline at Dauan Island, which is situated some 10 km offshore in Queensland waters. The inselberg is therefore clearly a product of subaerial erosion.

The only genuine karst feature in the area is a small cave some 3 m high and 5-6 m deep on a gentle rise south of Kuru village near Oriomo Station. Numerous shallow saucer-shaped depressions filled with water occur along the south coast. The depressions represent former dolines which have been sealed off by 1-2 m of alluvium. The dolines must have developed when sea level was lower and the limestone was exposed to the atmosphere.

The Fly Platform north of the lower Fly is distinctly different from the area to the south, although formed on the same sediments and situated on the same structural element. The landforms are low, closely spaced dendritic ridges and narrow valleys with plateau remnants (Plate 2). Slopes vary between 10° and 30°. Characteristically the summit levels of the ridges are very even, forming a perfect plain and sloping consistently to the south; this is clearly an indication that the area constitutes a former continuous piedmont alluvial plain which is now being dissected (Blake 1971; Blake and Ollier 1970).

Plate 2
Very low closely spaced ridges of the northern Fly Platform with an even summit level
The north-eastern part of the platform is formed by extensive volcano-alluvial fans which stratigraphically overlie the predominantly fluvial sediments to the south and west. Topographically the transition from the relict alluvial plain to the volcano-alluvial fan is marked by a low but distinct step or break of slope in the south, but in the west the transition is more gradual and difficult to delineate. Several isolated volcanic cones rise above the volcano-alluvial fan, sharply breaking the monotonous horizon of the plain. According to Blake (1971) the uppermost deposits of the vast piedmont alluvial plain were laid down less than 27,000 years ago and incision took place even later. This assumption is, however, doubtful, not only because it would require a very high rate of erosion not found on any other plain land in Papua New Guinea, but also because the exact correlation of the rather uniform alluvial deposits over vast areas is very difficult, if not impossible, and the date obtained from one small area cannot be regarded as representative for the whole area. Observations in the Bosavi area show that the volcano-alluvial fans overlie the alluvial plain and are therefore clearly younger. They are also much less dissected than the piedmont alluvial plain. The formation of the volcano-alluvial fans must have taken place contemporaneously with the formation of Bosavi volcano and the other neighbouring volcanoes in the early and middle Pleistocene, which must be regarded as the minimum age for the piedmont alluvial plain.

The seaward margin of the region to the south consists of a series of beach ridges and swales and a narrow belt of tidal mangrove flats. To the east lie the extensive deltas of the Fly, Bamu and Turama Rivers with generally an outer tidal zone merging inland.
into a fresh-water environment.

THE SOUTHERN PLAINS AND LOWLANDS
In contrast to the relatively uniform Fly Platform this region is one of great heterogeneity. It consists of hilly lowlands and foothills and intervening depositional plains forming embayments, and extends as a narrow belt some 15–20 km wide between the coast and the steeply rising central ranges, from the Purari Delta in the west to Mullins Harbour in the east. The lowlands are dominated by low ridges formed on moderately to steeply dipping limestone, chert, sandstone, siltstone and mudstone (Plate 3). The rocks are nowhere massively resistant; spectacular strike or hogback ridges are therefore missing but there is generally a clear alignment of ridges and intervening broad strike lowlands (Mabbutt 1965). Concave footslopes are characteristic for much of the area. According to Mabbutt this area has a rather distinct geomorphic character because the low seasonal rainfall results in a savanna type morphogenetic regime similar to that of northern Australia but quite unlike the humid tropical regime dominating most of Papua New Guinea. This is also the reason for a degree of landform stability with landforms, surface deposits and weathering profiles inherited from an older landscape cycle not to be found elsewhere in Papua New Guinea, with the possible exception of the southern Fly Platform.

Some undulating terrain often associated with duricrust cappings indicates advanced planation.

The depositional plains form extensive embayments between the lowlands and are made up of partly littoral and partly fluvial environments as well as extensive swamp areas. The estuarine littoral plains are most extensive in the west where large rivers such as the Tauri, Lakekamu and Vanapa enter the sea, but become smaller eastward. Sandy beach ridges, beach plains, spits and beach barriers occur along most of the coastline. The estuarine plains merge inland into fresh-water swamps or fluvial plains.

THE SOUTHERN FOLD MOUNTAINS
This region corresponds to the structural region of the Southern Fold Belt and is characterised by strong structural and lithological control of the landforms. The region falls into three parts, the sedimentary Aure fold area in the east, the Kikori-Lake Kutubu karst area in the centre, and the Victor Emanuel fold area in the west.

The Aure Fold Area
The Aure fold area consists of rows of closely spaced homoclinal ridges the general trend of which is north-north-west, turning to a more westerly direction at the head of the Purari Embayment. The steep escarpments face south-west to south and accordingly the dip-slopes face north-east to north (Plate 4). Both the dipslopes and escarpments are dissected by short gullies joining larger streams which flow parallel to the strike. Structurally most of the landforms reflect south-westerly overturned folds. Rock types are variably bedded limestone, greywacke, siltstone and mudstone, mostly
Miocene in age. The youngest sediments involved in the folding are Pliocene, which again points to the extreme youth of the landscape.

**The Kikori-Lake Kutubu Karst Area**

The Aure fold area merges westward into the Kikori-Lake Kutubu karst area approximately at a line connecting the two volcanoes Mt Karimui and Mt Faven. This area is the largest more or less continuous area of karst in Papua New Guinea, covering approximately 15,000 km², and consists of a series of west-north-westerly to westerly trending plateaux and broad ridges, the surfaces of which are covered with spectacular cone and tower karst (Plate 5). Separating the limestone areas are narrow corridors formed on clastic sedimentary rocks. The underlying geological structure, which

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**Plate 4**

Rows of closely spaced homoclinal ridges with south-west facing steep escarpments in the Aure area. Rock types are inter-bedded limestone, silt-stone, sandstone and greywacke.

**Plate 5**

Radar imagery of Kikori-Lake Kutubu karst belt with westerly trending limestone plateaux and broad ridges which are covered with cone and tower karst. The geological structure consists of broad synclines and tightly folded anticlines the cores of which have been deeply eroded and now form narrow corridors. (Note north is to the bottom of imagery. This has been done to eliminate the effect of an apparent inverted relief caused by the radar shadow on the north facing slopes.)
is well expressed in the landforms, changes gradually from monoclinal folds and overthrust anticlines in the south-west to tightly folded anticlines and broad synclines in the north-east (Bain 1973, Fig. 5). The monoclinal folds form wide plateau areas in the south, like the Darai Karst Plateau, while in the north the more tightly folded anticlines and broad synclines form limestone ridge and vale country with the synclines forming the karst ridges and the narrow anticlines being deeply eroded narrow sedimentary corridors. The ridge and vale country is thus an example of a very young relief inversion.

The Victor Emanuel Fold Belt

This region lies to the west of the Bosavi and Sisa volcanoes. It is a more heterogeneous region than its eastern neighbour, consisting of rows of homoclinal ridges to the south and karst plateaux and strike ridges in the northern and central parts. Its structure is one of broad folding with a large amplitude anticline, the deeply eroded Müller anticline exposing Jurassic and Cretaceous sediments and at one small locality Palaeozoic basement. The dominant physiographic feature is the Hindenburg Wall, a spectacular cliff of limestone hundreds of metres high, forming a nearly impenetrable barrier to the country behind (Plate 6). Other impressive landforms are a series of westerly trending limestone ridges probably formed by gravity sliding on the underlying clastic sedimentary rocks (Jenkins 1974) (Plate 24), and extremely rugged tower karst and doline karst on limestone surfaces. Mt Capella, Mt Scorpio, and possibly Mt Wamtakin, the highest peaks of the area, have been modified by glacial erosion and show well-preserved

Plate 6

The Hindenburg wall, a nearly vertical limestone cliff close to 1000 m high. The limestone plateau forms the northern limb of an anticlinal structure the core of which has been stripped of its limestone cover exposing the underlying soft Cretaceous sediments. In these mudflows are prominent and cause undermining of the cliff.
The term highlands as used here does not coincide with the political districts.

Plate 7
ERTS scene from the central and western part of the highlands. To the south the westerly to west-north-westerly trending Kikori-Lake Kutubu karst belt forms a well-defined southern boundary to the highlands. The volcanoes Mt Giluwe, Mt Lalibu and Mt Hagen, the intramontane basins and the ridge and V valley landscape are clearly visible (see Fig. 9 for interpretation).

The Goroka-Kainantu Area
This area consists of shallowly dissected upland formed by rounded hills and ridges with mostly convex slopes developed on granodiorite, volcanics and clastic sedimentary rocks (Plate 8). Three intramontane basins, the Goroka basin (Plate 57), the Aijura basin and the Arona basin form larger areas of flat to gently undulating country. The Goroka basin is covered with alluvial fans which were deposited glacial features.

This region was first traversed by Karius and Champion in 1927 on one of the most difficult expeditions ever undertaken in Papua New Guinea and even today the area is one of the most inaccessible corners of the country.

THE HIGHLANDS
Although never clearly defined the term highlands has been used for decades as a regional term encompassing the upland areas of the central part of the central ranges which are characterised by a relatively low local relief, a succession of intramontane plains and broad upland valleys, and by several huge strato-volcanoes, the eruptive products of which are widespread over much of the area* (Plate 7, Fig. 9). Much of this area is densely populated and covered by anthropogenic grassland.

The four main subdivisions of the highlands are the Goroka-Kainantu area in the east, the Wahgi-Mt Hagen area in the centre, the upper Purari area in the south and the Mendi area in the west.
on an older lake basin. Their source area is the eastern Bismarck Ranges, which during a period of intense seismic activity seem to have shed coarse debris into the lake basin, cutting across and covering the fine well-bedded lacustrine sediments. The sudden
influx of coarse gravel was probably also responsible for the incision of the lake overflow and its eventual drainage. Two older basins, the Aijura and Arona basins, were formed during the early Pleistocene by upthrusting of a ridge, the Yonki ridge, blocking the northward drainage of the basins and causing their filling with sediments. The Ramu River eventually cut through the ridge and the unconsolidated deposits were dissected and now form rolling uplands.

The Wahgi-Mt Hagen Area

This region extends west of the Goroka-Kainantu area and comprises most of the catchment of the upper Wahgi. It consists of rugged mountain ranges flanking the broad basin of the Wahgi River, which is filled with extensive fan deposits (Plate 56). The Wahgi basin in its present form is thought to have been caused largely by damming of the formerly westward flowing Wahgi by the eruption of Mt Hagen volcano (Haantjens 1970) but the existence of a structural low between the Kubor and Bismarck Ranges prior to the volcanic events must be assumed to explain the present extent.

Plate 9

A striking feature of the Wahgi-Mt Hagen area is a westerly trending limestone hogback with its smooth northerly facing dipslope (foreground). The Chimbu River has cut through the ridge in a deep gorge. In the background is the northern flank of the Kubor Anticline, formed of a succession of consolidated clastic sedimentary rocks ranging in age from Jurassic to Cretaceous. The northerly dip of the sedimentary sequence is well reflected in the landforms.

Plate 10

Series of parallel north-westerly to north-north-westerly trending limestone hogbacks separated by corridors formed on less resistant rock characterise much of the Mendi area.
and configuration of the basin. A striking feature of the area is the west-north-westerly trending Elimbari Ridge, continuing westwards into the Porol Escarpment, a limestone hogback ridge with a south-westerly facing clifffed scarp slope and a north-easterly facing smooth but steep dipslope (Plate 9). South of the ridge extensive coarse clastic material originating from the cliff forms a very hummocky landscape.

West and south-west of the Wahgi two huge strato-volcanoes, Mt Hagen and Mt Giluwe, dominate the landscape, rising about 2000 m above the surrounding plains. Both volcanoes were covered by glaciers during part of the Pleistocene.

**The Mendi Area**

This is the westernmost highland area, extending westwards from the Mt Hagen and Mt Giluwe volcanoes. The area is characterised by a series of parallel north-westerly trending hogback ridges of limestone, separated by corridors of less resistant rock formations such as siltstone, mudstone and greywacke partly covered by airborne volcanic products and lahars (Plates 10, 92). Although structurally this area is part of the Southern Fold Belt, physiographically it is closer to the other highland areas. In particular a number of intramontane basins with alluviated floors are present within the strike ranges that give it its distinct highland character. Many of the basins undoubtedly owe their origin to damming of drainage systems by lava and agglomerate during the formation of the highland volcanoes. Some have been caused by backward tilting of the headwaters or faulting across valley floors, some appear to be largely structural in origin and others have been dammed by landslides. There is certainly no single cause for their formation (see Chapter 5) (Pain 1973; Williams and others 1972; Guilcher 1970; Bik 1967).

**The Upper Purari Area**

This is a rather poorly defined area forming the southern part of the highlands. It is dominated by high relief ridge and V valley landforms except at its southern fringe where it abuts on the southern fold mountains and where four volcanic centres, Mt Soaru, Mt Karimui, Crater Mountains and Mt Yelia, interrupt the uniformity of the ridges and V valleys. Marine clastic sedimentary rocks underlie most of the area except on the southern slopes of the Kubor Range and on Mt Michael where acid igneous rocks occur.

**THE NORTHERN METAMORPHIC RANGES**

While the southern fold mountains and the highlands are very heterogeneous regions and contain a great variety of landforms, the northern metamorphic ranges form a belt of rather uniform mountainous country. The ranges are bounded to the south by the southern fold mountains and the highlands and to the north by the Sepik depression. They extend from the Thurnwald Ranges in the west to the Bismarck Ranges in the east and include the highest, most rugged and remote areas of Papua New Guinea. Most of the ranges present textbook examples of the tropical ridge and V
Plate 11
Radar imagery of uniform ridge and V valley topography in the northern metamorphic ranges. Section shows Lagaip River area

valley landforms or Kerbtalrelief (Behrmann 1923, 1927) (Plate 11) with high, narrow-crested mountain ridges, and long steep straight slopes meeting to form V-shaped valleys that have very steep gradients and are filled with coarse gravel and boulders. This area has the highest relief in Papua New Guinea with falls of up to 4000 m over a distance of only 15-20 km. The highest mountain of Papua New Guinea, Mt Wilhelm (4509 m), which was extensively glaciated during the Pleistocene, is in this region. There are very few erosion surfaces of slightly lower relief present and structurally controlled landforms are also rare, though stream courses are frequently fault-controlled. This is probably due to the area having been subject to erosion for a longer period than any other area in Papua New Guinea, the Kubor Range excepted.

THE EASTERN METAMORPHIC RANGES

The eastern metamorphic ranges comprise the Owen Stanley Ranges and its flanking ranges, running the entire length of east Papua. Structurally and geomorphologically they are very similar to the northern metamorphic ranges and in fact can be regarded as their eastern extension. They are also part of the New Guinea Mobile Belt and are dominated by massive ridge and valley landforms developed on low grade metamorphics and basic volcanics (Plate 12). One important difference, however, is the presence of widespread relict surfaces, mainly along the main divide, but also on some secondary watersheds branching off it (Plate 13).

These relict surfaces vary in altitude according to the elevation of the main watershed and thus rise from about 200 m in the east near Samarai to about 2800-3400 m near Mt Albert Edward and Murray Pass, where they are most extensive. From here they fall again north-westwards to between 1500 and 2000 m around Lake Trist and Mt Missim. These surfaces are assumed to be the remnants of one major erosional land surface as there are nowhere two distinct surfaces developed and as these landscapes can be continuously traced along the main watershed of the Owen Stanleys. The highest
peaks of the Owen Stanley Ranges, Mt Albert Edward, Mt Scratchley and Mt Victoria, were glaciated during the Pleistocene and show well-preserved glacial landforms.

THE ULTRABASIC RANGES

The ultrabasic ranges are separated from the Owen Stanley Ranges by a spectacular fault trough which extends with some interruptions over 200 km. The fault trough is bounded to the south-west by a thrust plane along which the Australian continental plate is thought to have been thrust beneath the oceanic mantle and crust of the Pacific plate during the early Eocene (Davies and Smith 1971).

Plate 12

Eastern part of the Owen Stanley Ranges, the rugged backbone of eastern Papua New Guinea. Most of the area shown is dominated by massive ridge and V valley landforms developed on basic volcanics. In the foreground steeply sloping delta fans give evidence of very active depositional processes. Densely dissected relict fans with flat remnants in the foreground and centre left indicate very young uplift.

Plate 13

Relict surface on the summit of the Owen Stanley Ranges. The relict surface is characterised by a distinctly lower relief, gentler gradients and denser dissection pattern than the surrounding younger ridge and V valley landscape.
The buoyant sialic crust rose isostatically during the late Oligocene and Miocene and the fault plane became gradually exhumed and dissected by erosional processes.

The landforms of these ultrabasic ranges are similar to those of the Owen Stanleys but tend to be more massive in appearance. They are mostly massive mountain ridges with very steep relatively uniform side slopes and narrow crests (Plate 14). Except near Lake Trist there are no summit surfaces preserved; however, there are some plateau-like areas near Mt Lamington produced by burying of the original landscape by a thick cover of ash.

**THE CAPE VOGEL BASIN**

This area is identical with the structural Cape Vogel basin. It is situated to the north-west of the ultrabasic ranges and extends from Morobe in the north-west to the Cape Vogel Peninsula in the south-east. The region is dominated by two groups of active and extinct volcanoes rising steeply from low-lying plains and fans. To the west there is the Mt Lamington group consisting of the active Mt Lamington volcano (Plate 39), the deeply eroded Hydrographers Range, and several smaller well-preserved volcanoes forming part of the Managalase plateau. The Cape Nelson group of volcanoes consists of the potentially active Mt Victory and the deeply eroded Mt Trafalgar (Plate 37).

Flanking the volcanic complexes are extensive plains and fans. To the north-west of Mt Lamington extend the Kumusi-Mambere plains, a large tract of low-lying alluvial plains with central meander belts and wide flanking swampy back plains. In the proximity of Mt Lamington the alluvial plains merge with the volcano-alluvial fans of the volcano, gradually increasing in gradient. The floodplains change from meander plains to braided plains due to the greatly increasing gradient and sediment supply.

South-east of the Mt Lamington group is the Musa plain, a nearly flat meander plain formed by rapid sedimentation of the Musa River which drains a large part of the Owen Stanleys. The
Musa plain merges south-eastward into low-angle fans which originate from the abruptly rising Gorupu Mountains. The easternmost part of the Cape Vogel basin is formed by low hills and raised coral platforms.

THE INTERMONTANE TROUGH

The central intermontane trough, one of the main structural lineaments of New Guinea, extends from Geelvink Bay in Irian Jaya to the Huon Gulf in Papua New Guinea and continues eastwards as a submarine depression leading into the New Britain trench.

The Markham-Ramu Graben

The eastern part of the trough, the Markham-Ramu trough, is a narrow graben zone occupied by a series of low-angle coalescing alluvial fans (Plate 15). These fans are formed of coarse debris derived from the tectonically very active Finisterre and Saruwaged Ranges which rise steeply along a series of faults along the north-eastern margin of the trough. The larger fans have gradients between 1 and 3 per cent and are entrenched at their apexes. Along some of the apexes several terraces are developed. The number of these terraces is not consistent, varying from one to three (further discussion p. 100). The rivers have gradients similar to the fans and flow in highly unstable wide braided flood-plains with constantly shifting sand bars and channels (Plates 44, 45). In their lower courses the present flood-plain and the older fan surfaces merge and form a continuous plain.

The Sepik Depression

Westwards the narrow Markham-Ramu trough opens into the much more extensive Sepik depression, which is predominantly formed of vast swamps, meandering flood-plains, and a series of low-angle variably dissected fans along the northern flanks of the depression. Structurally the Sepik depression differs considerably

Plate 15
Markham-Ramu graben, which is part of a major structural lineament separating the central ranges to the south (foreground) from the northern coastal ranges. The graben is filled with alluvial fans which have been shed from the tectonically active Saruwaged Ranges.
from the Markham-Ramu graben as it is generally not fault-bounded but constitutes a large basin, the remnant of the Northern New Guinea Basin which during the Pliocene extended from the central ranges to the present north coast.

The area south of the meander belt of the Sepik River is formed by a series of nearly continuous swamps which extend right up to the foot of the abruptly rising central ranges. Some more or less isolated hills and mountains occur like islands within the swamp area (Plate 16). As already described by Behrmann (1917, 1924) the valleys draining into this swamp have the character of drowned valleys with the swampy alluvium extending into the mountain front. A great number of valleys also seem to have been cut beneath the present base level of erosion. This drowning can be the result of either the postglacial rise in sea level or subsidence or both.

North of the Sepik River swamps are markedly rare or absent. Instead variably dissected low-angle fans cover much of the area between river and mountain front, indicating that this has been an area of uplift rather than subsidence (Plates 54, 55).

THE NORTHERN COASTAL RANGES
The northern coastal ranges are the second largest mountain system of Papua New Guinea. Like the central ranges they traverse the entire island from east to west, not as a continuous chain but as several ranges in echelon which are separated by structural lows such as the lower Sepik-Ramu plain and the Gogol depression.

The ranges fall steeply to the coast and the fall continues offshore to a depth of 1000 m. A coastal plain is therefore largely lacking. Instead much of the coastline is formed by narrow coral platforms and terraces, beach ridges and swales, and steeply sloping delta fans.
The Saruwaged and Finisterre Ranges

In the east the Saruwaged and Finisterre Ranges form the most rugged and highest parts of the ranges with the formerly glaciated Mt Bangeta rising to 4121 m. Ridge and V valley landforms dominate the southern fall of the ranges while on the northern slopes sloping structural surfaces form important elements of the landscape (Plate 17). Some spectacular limestone plateaux are developed along the eastern part of the summit area. Although structurally controlled the irregular plateau surface clearly cuts across the bedding of the Miocene limestone and is therefore to be regarded as a relict surface, post-Miocene in age.

A flight of young coral terraces is developed along part of the north coast, indicating that this area is actively rising (Plate 765). Maximum rates of uplift are of the order of 3 m per 1000 yr (Chappell 1974a).

The Adelbert Ranges

The much lower Adelbert Ranges are offset to the north-west and are separated from the Finisterre Ranges by the Gogol depression. In spite of the lower altitudes the range has a vigorous relief and very steep slopes which, as recent events have shown, are highly unstable during earthquakes of relatively moderate force (Plate 77).

The Prince Alexander-Toricelli-Bewani Ranges

The ranges west of the Sepik Ramu plain form a continuous mountain chain rising to between 1000 and 1500 m in its central

Plate 17
Saruwaged Range with summit plateau formed on flat lying to gently dipping limestone and massive ridge and V valley landforms on the underlying volcanics (right). Parts of the summit plateau were modified by glacial action during the Pleistocene (cirques and U shaped valleys in the foreground and along summit). Sloping structural surfaces in the background left.
parts. The core areas are formed by relatively uniform rugged ridge and V valley landforms on metamorphic and igneous rocks with frequent fault-controlled river courses. The flanking lower sedimentary mountains and hills exhibit locally structural control, particularly to the south where impressive rows of homoclinal ridges with a westerly trend are developed. The Toricelli and Eastern Bewani Ranges have been subject to severe earthquakes during this century and numerous landslide scars still bear witness to these catastrophic events.

**THE SOUTHERN BISMARCK ISLAND ARC**

Numerous islands ranging in size from about 30,000 km² to a few square kilometres are situated off the north coast of Papua New Guinea and off its easternmost tip. Of these the Bismarck Archipelago comprises all the larger islands which form an oval ring about the Bismarck Sea. Structurally they consist of two arcs, the southern Bismarck Island Arc and the northern Bismarck Island Arc or New Ireland arc.

The southern Bismarck Island Arc includes New Britain and the chain of volcanic islands off the north coast of New Guinea. It is a typical island arc with crescentic shape, a deep oceanic trench, the New Britain trench descending to 7880 m, and a belt of active volcanoes along its concave north coast. Most of these volcanoes such as Manam, Bam, Karkar (Plate 40), Bagbag, Long Island, Pago (Plate 41), Langila, Ulawun and Bamus volcanoes are active or potentially active. The central and southern parts of New Britain are, however, non-volcanic and consist of rugged mountains and extensive areas of karst. The coastal areas are formed of raised coral terraces in the south, which indicates recent uplift, and coastal swamps separated by volcanic foothills in the north. Fringing reefs are abundant along most of the coastline.

**THE NORTHERN BISMARCK ISLAND ARC**

The northern Bismarck Island Arc or New Ireland Island Arc consists of Bougainville and Buka Islands, New Ireland, New Hanover, the St Matthias Group and the Admiralty Islands. Bougainville, the largest island of the Solomons, is formed of a central chain of mountains which include three active volcanoes, Balbi, Bagana and Loloru (Plate 42). Extensive volcano-alluvial footslopes and fans surround these and other volcanic cones.

Buka Island to the north-west of Bougainville consists of a narrow low north trending range at its western margin and an extensive uplifted coral reef rising about 100 m above sea level. Few surface karst features have developed, probably because the limestone has been exposed to the atmosphere for only a short time span. However, several caves have been observed near sea level (D.H. Blake pers. comm.).

New Ireland, north of New Britain, is a long narrow north-westerly trending island with two narrow necks. The south-eastern part of the island is mountainous while the central and north-
western parts are dominated by well-developed tropical karst landforms. A discontinuous belt of raised coral is developed along the northern coast and coral reefs fringe most of the coastline. The asymmetric form of the island is due to north-east tilting with the maximum rate of uplift along the south-west side.

The island of New Hanover to the north-west has a core of older volcanics surrounded by volcanic aprons and alluvial deposits. The original volcanic landforms have been all but destroyed by erosional processes.

Running parallel to New Ireland and New Hanover is a row of volcanic islands including Feni, Tanga, Lihir and Tabar Islands, and coral islets and reefs. Mussau, the main island of the St Matthias group, is a strongly dissected volcanic complex.

The Admiralty Islands to the west comprise Manus Island and some much smaller islands grouped south of Manus in a semicircle. Manus is mainly hilly but some parts are mountainous. Raised coral reefs form the east and west coasts and much of the island is surrounded by a complex barrier reef system.

The islands south of Manus are mostly of volcanic origin or represent raised coral reefs. Tuluman volcano south of Lou Island is an active submarine volcano.

THE EAST PAPUA ISLANDS

The islands north of the eastern tip of the mainland, the D'Entrecasteaux Group and the Louisiade Archipelago, are much smaller and structurally represent an extension of the mainland. They are formed by rugged mountains predominantly developed on schist and gneiss. Clusters of volcanic cones occur along the south coasts of Goodenough and Fergusson Islands. Fringing reefs are well developed along most of their coastline and raised coral platforms give evidence of young uplift.

THE TROBRIAND-WOOLDRARK ISLAND GROUP

The Trobriand Islands, which lie north of the D'Entrecasteaux Islands, are a group of raised coral atolls. The most important islands are Kiriwina, the main island of the group, Kaileuna, Vakuta and Kitava. The islands usually consist of an outer ring of low hills that represent the old reef and a low-lying mostly swampy centre which constitutes the former lagoon. This is best developed on Kitava Island which is the highest island of the group, rising to 142 m. Several terraces occur on this island indicating that uplift took place in several stages (Ollier and Holdsworth 1969). No such terraces are present on the other islands. Numerous caves and other karst features such as pinnacles and dolines are associated with these reef islands and are well documented (Ollier and Holdsworth 1968, 1969; Ollier 1975). Woodlark Island further to the east is also basically an uplifted coral platform but has a core of metamorphic rocks rising above the coral at its south-eastern extremity.
3 Landform Types

The landform types of Papua New Guinea have been mapped and briefly described by the author (Lößler 1974b) and it is probably of benefit if the following section is read in conjunction with the geomorphological map as it will allow the reader to place the different landforms more easily in their spatial context. Unfortunately it has not been possible, as originally intended, to include the geomorphological map in this volume. It has been published elsewhere and is available from the Division of Land Use Research, CSIRO, Canberra.

The landforms have been subdivided according to the geomorphological processes by which they are formed and no geochronological differentiation has been made except for the distinction between active and relict. This is because in geomorphologically young areas like Papua New Guinea landforms tend to be more closely related to their present formative processes than is the case in older landscapes such as Australia, where significant changes in processes have taken place through their long history.

DENUDATIONAL LANDFORMS

Denudational landforms are by far the most common and most heterogeneous group, comprising all the landforms that have been formed by denudational processes in the widest sense. These include fluvial erosion, mass movements, subcutaneous erosion, slope wash, solution, glacial erosion and marine erosion.

Landforms such as karst, glacial or marine erosional landforms are relatively easy to define because they have been formed by one major process or set of closely related processes. Landforms of fluvial erosion and mass movement are, however, more difficult to characterise, since they have been formed by a greater variety of distinctly different processes. These include, as the term implies, not only fluvial erosion and various kinds of mass movements but also slope wash, sub-cutaneous erosion and solution. It is therefore to be expected that this group will be more heterogeneous than the others. Two major subgroups have been distinguished below according to whether or not structural control is prominent and has clear surface expression.

Landforms of Fluvial Erosion and Mass Movement

*Ridges and V Valleys.* This landform type consists basically of ridges with steep slopes and narrow to knife-edged crests separated by V-shaped valleys with steep gradients. Although this is the most common erosional landform type in humid tropical areas there is
no well-established English term to describe it. Simonett (1968) has used selva landscapes in an article dealing with denudational processes in the humid tropics. Selva, a Spanish term, refers to the rain forest cover rather than to the landforms. The German term Kerbtalrelief, meaning V valley relief, was established by Behrmann (1917, 1927) during his work in New Guinea but it has never been taken over into the English terminology. It has the disadvantage also that it notes only the valley forms but not the ridges. The term ridge and valley would be good but has been in use for nearly a century in connection with the fold structures of the Appalachian Belt (Thornbury 1965). Ridge and ravine has occasionally been mentioned (Ruxton 1967; Loffler 1974b) but ravines are small drainage features not really comparable with the large V-shaped valleys characteristic of this landform type. The term ridge and V valley, awkward as it may sound, seems to be the best compromise term; it has no other connotation and gives a fair, abbreviated description of these landforms.

Figure 10
Generalised lithology

<table>
<thead>
<tr>
<th>Sedimentary rocks</th>
<th>Igneous rocks</th>
<th>Unconsolidated sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Non-calcareous</td>
<td>Volcanic</td>
<td>Alluvium</td>
</tr>
<tr>
<td>Limestone</td>
<td>Intrusive, acid-intermediate</td>
<td>Volcano-alluvial</td>
</tr>
<tr>
<td>Metamorphic rocks</td>
<td>Intrusive, ultrabasic</td>
<td></td>
</tr>
<tr>
<td>All grades</td>
<td></td>
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</tr>
</tbody>
</table>
The simple description of course fails to account for the great variety of these landforms. There are great differences in drainage density, dissection pattern, slope form, slope steepness, relief, crest pattern and slope symmetry. Although detailed quantitative investigations on the interrelationships of these attributes are still lacking, numerous qualitative observations largely based on air-photo studies supported by limited field observations have been collected by the author and other CSIRO workers and have shown that the dominant factors determining the above-mentioned attributes are rock type and tectonic structure in its response to weathering and erosional processes. The influence of rock type and structure tends to be clearest when weathering is not pronounced, i.e. when denudation rates outstrip or at least balance the rate of weathering. In the following section this statement will be substantiated with a number of typical examples.

Ridge and V valley landforms on igneous rocks (distribution of rock types shown on Fig. 10) can be broadly generalised as being of massive appearance, with a coarse dissection pattern and steep rather straight slopes; however, looking at the landforms in more detail, quite significant differences emerge not only between different types of igneous rock but also within what is generally regarded as one rock type.

Ridges and V valleys on ultrabasic rocks are probably the most easily recognisable ones, forming massive ridges with long, straight or slightly convex slopes (Plate 14, Fig. 11). The dissection pattern is coarse with relatively few rivers per unit area, which makes the contrast with neighbouring areas mostly formed on metamorphics very clear and easy to trace on aerial photographs. Vegetation usually accentuates the difference, ultrabasic areas commonly being completely forested with a dense low dark-toned forest while the non-ultramafic areas bear a more open higher and lighter-toned forest or secondary vegetation, or are cultivated.

Field observations show that gullies or small streams are not common in the upper or upper-middle slopes and the incision is generally shallow. Water often runs in shallow saucer-shaped depressions and definite V-shaped valleys form only at the lower slopes. Streams are incised in the bedrock and are conspicuously poor in rounded gravel. They contain some sand and very coarse boulders with rough corners, obviously derived from nearby bedrock outcrops.

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**Figure 11**
Ridges and V valleys on ultrabasic rocks

![Diagram of ultrabasic rocks](image-url)
The thickness of the soil mantle varies from half a metre to some 5 m, but there is generally a sharp though irregular boundary between the weathered material and the bedrock. A special feature of the ultrabasic areas is also the occurrence of subsurface erosion features in the form of tunnels, small caves and sink holes.

These characteristics suggest a similarity between ultrabasic rocks and limestone in their response to weathering and denudation processes, and this is probably due to the solubility of the magnesium in these olivine-rich rocks (see Chapter 4, section on subcutaneous erosion).

Ridges and V valleys on other plutonic igneous rocks such as granodiorite, diorite, gabbro and basic volcanics also tend to be massive in appearance. Ridge crests are sharp and slopes are straight and rather uniformly sloping with slopes of 35–38° (Plates 12, 17, Fig. 12). Though still coarse, the dissection pattern is clearly finer than on ultrabasic rocks. Rivers and streams are cut in bedrock but are loaded with coarse well-rounded gravel. Landslides of the debris avalanche type are common.

This type of ridge and V valley country can be found in the northern coastal ranges, in the south-western part of the Kubor Range and on Mt Michael.

On volcanics large areas of massive ridge and V valley country extend over East Papua (Plate 12) and along the southern fall of the Saruwaged and Finisterre Ranges (Plate 17).
When deeply weathered rocks are involved the appearance of the ridges and V valleys differs considerably from that already described. Ridges here have broad rounded crests and convex side slopes. Examples of this have been found in some highland areas, mostly in association with relict surfaces (Plate 8).

Ridges on metamorphic rocks are similar to those on most igneous rocks, being massive and straight in the overall slope profile. In detail, however, slopes on metamorphics are more irregular, being densely dissected by a close network of small streams and gullies (Fig. 13). Many of these tend to follow foliation planes. Weathering is very shallow and the soil profile consists predominantly of coarse colluvium overlying the bedrock. Rock outcrops are also common. Good examples of this landform type occur in the Owen Stanley Ranges and in the northern metamorphic ranges.

As is the case with igneous rocks, ridges and V valleys on deeply weathered metamorphics are different in appearance. They are much lower in relief, crests are broad and slopes are more gentle, and the dissection pattern is fine but highly irregular. Most of this type of ridge and V valley occurs in association with relict surfaces (Plate 13).

Ridge and V valley landforms on sedimentary rocks are the most widespread and are particularly prominent in the southern and central parts of the central ranges and in the northern coastal

![Figure 14](image)

Ridges and V valleys on soft, fine-grained sedimentary rocks

![Plate 18](image)

Densely dissected slopes on fine-grained sedimentary rocks west of Madang. The irregular vegetation pattern is due to the practice of shifting cultivation. Villages are on top of the narrow ridge crests.
ranges. They show great variability, which is not surprising consid­ering the great differences in composition, degree of induration, bedding, homogeneity within the layers, and degree of tectonic deformation.

Ridge and V valley landforms on soft, fine-grained sedimentary rocks such as marl, mudstone and siltstone are characterised by a very dense dissection pattern and highly irregular slopes with great variability in slope steepness because of frequent slumping and intense gullying (Fig. 14, Plate 18). Slopes can vary between 50° at slump headwalls to a few degrees at slump toes. On the aerial photographs the slope irregularity is largely masked by the forest canopy and is therefore not always apparent but on the ground the often chaotic relief is very obvious and makes those areas extremely difficult for any engineering activity.

Weathering is surprisingly mostly shallow and immature but some softening of the rock due to hydration to a depth of about 3 m seems to occur in most areas (Haantjens and others 1972; Löffler 1972b). Examples of this landform type on fine-grained sedimentary rocks are to be found in the southern lowlands and eastern Papua and in the northern coastal ranges.

With increasing degree of induration of the sedimentary rocks, dissection patterns tend to be coarser and slopes somewhat more uniform, though still irregular in detail compared with slopes on igneous and metamorphic rocks. Slumping is still an important process of mass movement but debris slides are equally significant particularly in areas of thinly laminated but relatively hard mudstones and shales. In some areas the structure of the rocks is more clearly visible in surface expression (see landforms of fluvial erosion and mass movement with prominent structural control). Slopes are usually steep and partly influenced by rock structures. Weathering and soil mantle are generally shallow though there is great local variability. Examples of this landform type are to be found in the southern fold mountains, the highlands (Fig. 15, Plates 9, 19) and in the northern coastal ranges.

Landforms on coarser-grained sedimentary rocks such as greywacke, sandstone and conglomerate show a similar sequence, with young poorly consolidated rocks forming a dense intricately dissected pattern of ridge and V valleys and more indurated and

Figure 15
Ridges and V valleys on consolidated fine-grained sedimentary rocks

Shale
older rocks forming coarser patterns with more uniform slopes. An example of the first type is shown on Plate 20 and for the more indurated coarse-grained rocks typical examples occur in the southern

Plate 19
Irregular ridge and V valley landforms on interbedded silt-stone, mudstone and greywacke

Plate 20
Very densely dissected ridge and V valley landscape formed on poorly consolidated coarse-grained fanglomerates in the vicinity of Rabaraba
and central parts of the central ranges (Fig. 16) and in the northern coastal ranges.

Ridgeline valley landforms on limestone are not common, as in most areas where limestone is the dominant rock type karst landforms are present or structurally controlled landforms such as hogbacks and plateaux are developed. In the few cases where simple ridges and valleys are formed on limestone, ridges are broad crested and have straight very steep slopes with frequent rock outcrops. Weathering is shallow and soil cover is very thin. Drainage density and slope dissection pattern are very coarse and slopes are relatively stable except at very steep angles and in areas of intense seismic activity. Examples of this landform type occur in the northern fall of the Saruwaged and Finisterre Ranges and the Torricelli and Bewani Mountains. In many of these cases, however, limestone is interbedded with clastic non-carbonate sedimentary rocks.

As mentioned before these characterisations of the ridge and V valley landforms on different rock types are generalities which apply if relatively homogeneous rock types covering larger areas are involved. The picture is much more complex if different rock types are closely interbedded, if intense fracturing by tectonism has taken place and if layers of superficial deposits such as ash cover the original bedrock. This is the case for instance in the Owen Stanley fault zone where basaltic lavas which normally develop a coarse dissection pattern with long steep slopes form an intricately dissected pattern with short irregular slopes. To the east of the fault zone some areas of ultrabasic rocks are covered by ash derived from the nearby Mt Lamington, and here in contrast to the normal very coarse ridge and V valley landscape on ultrabasics there is developed a very fine closely dissected pattern of ridges and V valleys.

*Ridge and V Valley Landforms Associated with Relict Surfaces.* These are widespread features in the Owen Stanley Ranges, mainly along the principal watershed and some off-branching secondary divides. In the highlands they also occur extensively but are more irregular in occurrence and difficult to delineate exactly. These landforms are characterised by a distinctly lower relief, less steep gradients of the rivers and a more mature and thicker weathering profile than the surrounding landscape. The transition from the older surfaces to the younger ridge and V valley landforms is often abrupt and marked by distinct breaks of slope in the rivers, often giving rise
to waterfalls and a clear change in the pattern of dissection from a fine-grained pattern on the older surfaces to a much coarser-grained pattern in the younger landscape below (Plate 13, Fig. 17). The older surfaces have clearly not been affected by the present cycle of erosion but the headward erosion of the present erosional system seems to be rapid.

In the Owen Stanley Ranges the relict surfaces rise from 200 m at the eastern end of Papua to over 3000 m on Mt Albert Edward. From here they fall again to between 1500 and 2000 m around Lake Trist and Mt Missim. Closely associated with these older landscapes are several intramontane basins such as the Neon Basin west of Mt Albert Edward (Plate 21) and the so-called Myola Lakes near Mt Suckling. Further west, much of the low relief areas around Kainantu and Goroka have the character of relict landscapes, but they are much more irregular and less well-defined than those of the Owen Stanley Ranges. There are some smaller local occurrences of summit surfaces in the western highlands (Perry and others 1965).

In the Owen Stanley Ranges, however, the summit surface can be traced along most of the watershed and it is very likely that it represents the relicts of one major phase of erosion. Observations by Smith (1970) from the Milne Bay area show that some of these

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**Figure 17**

Ridges and V valleys associated with a relict surface

*Granodiorite*

---

**Plate 21**

Neon basin, an intramontane basin west of Mt Albert Edward at approximately 3000 m. The basin occurs within the relict surface and may have formed before the main uplift relatively close to sea level.
surfaces are developed on Miocene and also possibly on Pliocene rocks, indicating that the surfaces here are post-Miocene or possibly late Pliocene in age. The surfaces developed on metamorphic rock are probably of the same age, as the surface continues without any obvious break from the Miocene sediments into the metamorphics.

It is possible that the relict surface continues northward into the Saruwaged Ranges where a structurally controlled relict surface is present, developed on Miocene limestone (Plate 17).

The continuation of this relict landscape into the highlands cannot be traced with any confidence because the relict surfaces there show great differences in altitude over short distances which may have been caused by differential uplift. They are also too isolated in occurrence to allow any interpretation as to their former extent and character.

The surfaces allow the following generalised interpretation. The post-Miocene emergence of the Papua New Guinea land mass was interrupted by a period of relative tectonic stability during which a hilly and in places mountainous landscape was formed. Most of this landscape developed probably relatively close to sea level under a tropical lowland climate, resulting in deep mature weathering profiles. Renewed uplift in the late Pliocene and during the Pleistocene raised these surfaces well above their former base level. In the Owen Stanley Ranges the magnitude of uplift increased from less than 200 m in the south-east to probably nearly 3000 m in the Mt Albert Edward area, indicating a westerly increase in the rate of uplift of these ranges. In the highlands uplift was of similar magnitude but its history cannot be traced because of a multitude of differential movements.

**Landforms of Fluvial Erosion and Mass Movement with Prominent Structural Control**

Studies in structural geomorphology are relatively few and as pointed out by Pitty (1971) the reasons may be that in contrast to dynamic or historical landform aspects, structural features are obvious and their interpretation unexciting or self-evident. Furthermore many geomorphologists may be lacking the specialised expertise and facilities to examine these landforms in macro or micro-scale (Pitty 1971). Also the interpretation of structural landforms does not easily lend itself to the formulation of global theories on landform development and these features are therefore often dismissed as anomalies, especially among researchers preoccupied with a climatic geomorphological interpretation of the landscape.

In Papua New Guinea hardly any geomorphological work has been done on these landforms by geomorphologists, with the notable exception of Bik (1967). The information on structural landforms is almost entirely the by-product of geological work. In particular relatively intensive geological mapping and exploration by the Australian Bureau of Mineral Resources and mining companies over the last few years in the southern fold belt have greatly advanced the understanding of structural landforms (Davies and Norvick 1974; Jenkins 1974; Findlay 1974).
There are few if any landforms in Papua New Guinea which are not controlled by rock or tectonic structure to some extent. The degree to which this structure is expressed topographically, however, varies greatly. This section deals only with those landforms where such structure has had a marked influence on shape.

The great majority of structurally controlled landforms in Papua New Guinea reflect bedrock structures such as planes of bedding and foliation formed in a variety of tectonic settings. There are also numerous examples of fault and joint-controlled landforms but they tend to be of limited areal extent.

Homoclinal Ridges, Hogback Ridges and Cuestas. These are by far the most common types of structurally controlled landforms, and develop where beds of varying resistance are present and have been brought into a more or less strongly inclined position by faulting or folding. Subsequent differential erosion, usually during phases of uplift, causes the less resistant layers to erode more rapidly than the resistant layers and the rock structure of these resistant layers is thus gradually given surface expression. The greater the difference in relative resistance to weathering and denudation the more pronounced will be the structural control. Homoclinal ridges and hogbacks are therefore best developed where limestone and hard porous sandstone form the resistant layers, alternating with softer shales, mudstones and siltstones.

The more gently inclined homoclineal ridges (10°–25°) are distinguished from the steeper hogbacks (over 25°) according to the angle of dip of the strata. At the lower end of the scale cuestas-which have dipslopes of less than 10° are further separated.

Homoclinal ridges and hogbacks are most common in the southern fold belt from which the examples discussed below have been drawn. They also occur in other areas but nowhere as extensively and conspicuously.

In the south-western part of the fold belt there are extensive rows of parallel, closely spaced north-north-westerly trending homoclinal ridges. The structural pattern is one of tight, mostly south-westerly overturned faulting (APC 1961). This is clearly expressed in the landforms and the drainage pattern of smaller streams and rivers although the main rivers do not follow the structural grain but cut across it, probably because of antecedent drainage. The ridges are mostly asymmetrical with steeper scarp slopes facing west-south-west and opposing gentler dipslopes (Fig. 18, Plate 4).
Plate 22
Southern flank of the Müller anticline with long smooth dip-slopes forming cuestas

Plate 23
Radar imagery of homoclinal ridges and cuestas in the middle Strickland area (Strickland River in centre) between the dissected relict alluvial plain of the Fly platform (right) and the southern flanks of the Müller Plateau (left). The major structures are the southern extremity of the broadly folded Müller anticline and several smaller and more tightly folded anticlines to the west of it. The broad folding has resulted in smooth gently sloping dipslopes while the much narrower folding has led to the development of closely spaced homoclinal ridges with steep dip slopes. The cuesta is formed of the very thick Darai limestone overlain by thin beds of mudstone and siltstone, while the homoclinal ridges are formed of interbedded mudstone, siltstone, sandstone and conglomerate. The two latter rock types are the ridge builders. Compare with Fig. 19 which represents an approximate N-S profile of part of the imagery. Note: north is to the bottom of imagery.

Locally the situation is reversed. The bedrocks are interbedded greywacke, mudstone, siltstone and minor limestone; the entire succession is up to 10,000 m thick. The greywacke consists largely of coarse volcanic detritus which is relatively resistant and is consequently the most common ridge builder. Both the outcrop slopes and dipslopes are intensely dissected by short steep gullies. Locally the intense dissection of the ridges leads to the formation of triangular facets called flat irons (Lobeck 1939), more commonly known from arid environments (Twidale 1971).

The general accordance of the ridges, the gentle southerly slope and their summit level have been interpreted as relicts of a former (late Pliocene) erosion surface (Ruxton 1967). However, none of the ridges has any surface remnants left and therefore the nature of the surface preceding the uplift cannot be inferred.

In the western part of the fold belt the structural pattern is simpler because faulting has been less intense and folding less tight. A sequence of homoclinal ridges and in places wide cuestas is developed
between the broad Müller anticline in the north and the Fly and Strickland alluvial plains in the south (Plates 22, 23). In the east a sequence of particularly well-developed homoclinal ridges has formed on upper Miocene and Pliocene non-carbonate sedimentary rocks overlying the Darai limestone which sheets the anticline. Prominent ridge builders, besides the Darai limestone itself, are Pliocene sandstone and conglomerate which are interbedded with siltstone and shale. The dips of the beds are relatively gentle close to the broad Müller Anticline, resulting in a relatively wide spacing of the ridges which are here transitional to cuestas. With increasing distance from the Müller anticline and approaching the much narrower Wai Asi anticline and finally the Cecilia anticline the dips become considerably steeper and the ridges more closely spaced (Fig. 19, Plates 22, 23). The dips are mostly southerly but are reversed on the northern flanks of the two smaller anticlines.

Similar sequences of homoclinal ridges and cuestas are developed further west where the centre of the Müller anticline has been stripped of its limestone cover and where the underlying Mesozoic sedimentary sequence has been exposed. Here the rock sequence ranges from Jurassic to Pliocene in age. The ridge-forming beds in

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**Figure 19**
Profile across Müller anticline (after Davies and Norvick 1974) (reprinted by permission Bureau of Mineral Resources, Australia)

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**Figure 20**

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Plate 24
Radar imagery of hogbacks north of the Müller anticline. The imagery shows a series of these ridges with steep smooth dipslopes facing north (top) and opposing near vertical escarpments. The ridges represent thrust sheets of the Darai limestone which became detached during the Pliocene orogeny, slid southwards, piled up and overthrusted at the Müller anticline. Compare with Fig. 20 which shows a N-S profile of part of the imagery.

the Mesozoic sequence are upper Jurassic to lower Cretaceous quartz sandstone and conglomeratic arkose. On either side of the exposed anticline, remnants of the overlying Darai limestone form spectacular plateaux with completely karstified surfaces and nearly vertical cliffs up to 1000 m high (Plate 6).

Upper Miocene and Pliocene sedimentary rocks also overlie the Darai limestone here but form only very small areas of homoclinal ridges at the southern margins of the anticline.

The most dramatic examples of hogbacks are to be found on the northern flank of the Müller anticline on either side of the upper Strickland River (Fig. 20, Plate 24). Here a series of four, locally five, limestone ridges with nearly vertical south-facing escarpments and very steep but smooth dipslopes are developed. Recent geological investigations have shown that these ridges are not a normal sequence of progressively younger beds but represent thrust sheets of the Darai limestone formation of similar age (Jenkins 1974). These thrust sheets became detached and broken up during the Pliocene orogeny and slid southwards on the upper Jurassic sediments, which thicken rapidly northwards and provide the southerly sloping glide plane. The thrust sheets were apparently piled up and overthrusted as their southerly movement was blocked by the Müller anticline, which implies that the latter predates the period of uplift to the north of it. This is supported by the lack of any thrust structure in the Müller anticline itself (Davies and Norvick 1974).

After the sheets came to rest, vertical movement and deformation set in, producing the broad dish-shaped fold structure that characterises the present-day landforms.

The nature of the land surface preceding the events described above is not known, but it is thought to have been a largely depositional surface possibly similar to the upper parts of the Fly Platform.
One reason for this supposition is that the drainage system appears to be largely antecedent. This is difficult to prove because of much local modification by river capture and possibly also by superimposition and river persistence.

A final example of particular interest because of its exciting shape is the Masi syncline west of Mt Karimui volcano (Plate 25). This nearly symmetrical syncline consists of three limestone plates thrust upon one another. The margins of the limestone sheets have been upturned by vertical forces, probably during uplift, and subsequent erosion has exposed the structure. During the eruption of Mt Karimui the sedimentary depressions between the ridges were partially filled with ash flows, but no significant changes took place to the ridges where karst landforms had developed.

**Sloping Surfaces.** Sloping surfaces are landforms transitional between homoclinal ridges and hogbacks on the one hand and plateaux on the other. Physiographically they could best be described as sloping plateaux as their summit surface forms an excellent plane of accordance that slopes consistently in one direction (Fig. 21). The slope of the plane of accordance is not simply an exposed rock structure, although it is in sympathy with the rock structure. The

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**Figure 21**
Sloping surface with structural control

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**Plate 25**
Radar imagery of Mt Karimui and Masi syncline. The syncline consists of three limestone plates thrust upon one another. Mt Karimui volcano to the east (left) erupted during the Pleistocene and filled the sedimentary corridors with ash flows. The volcano has a huge crater enlarged by erosion and breached to the south (note south is to the top).
erosional processes have exploited the naturally occurring weaknesses in the rock such as bedding planes and fault planes or planes of foliation as in the case of the Dayman Dome in east Papua (Plate 26). However, the preservation of these surfaces is not favoured, since the erosional processes are dominated by vigorous fluvial incision. The sloping surfaces are therefore like the plateaux relict surfaces.

Two factors could account for their development. Firstly they originated, like the plateaux, adjusted to a lower base level of erosion, but were subsequently tilted during the uplift from their originally more or less horizontal position. Secondly they could be fault planes which were formed by such rapid uplift that the erosion was not capable of keeping pace with it.

*Plateaux.* Plateaux are high-lying surfaces on flat-lying or gently sloping rock surrounded by distinctly different landform types (Plates 6, 17). They are formed on relatively resistant rock such as limestone and sandstone and are essentially the product of differential erosion and uplift. The plateau surfaces are irregular in detail and clearly represent not original bedrock surfaces but erosional surfaces.

Most plateaux on limestone have been intensively modified by karst erosion and here the summit levels of the karst cones appear to be in close sympathy with the rock structure. A few structural plateaux, however, were formed by fluvial processes and seem to have developed at a much lower altitude when the base level of erosion was lower.

In their present position these plateaux are being destroyed by vigorous headward erosion cutting back from the lower and younger erosional system into the plateau surface. This is well exemplified

Plate 26
Sloping structural surface of Dayman Dome formed on densely foliated schist
in the Saruwaged Range where the river network on the plateau is partly beheaded by the rapid headward erosion of the present erosional systems (Plate 17).

**Landforms of Karst Erosion**

The term karst refers to landforms 'with distinctive characteristics of relief and drainage arising primarily from a higher degree of rock solubility in natural waters than is found elsewhere' (Jennings 1971). This broad definition is preferred here as the emphasis is laid on rock solubility rather than on the presence of limestone and underground drainage (Sweeting 1973), which would exclude certain areas of tropical karst as well as distinctive karst features on non-limestone rocks.

There are numerous types of karst in Papua New Guinea ranging from minor solution features like karren, ripples and solution notches to larger surface and subsurface landforms such as dolines, dry valleys, cones, towers, pinnacles, caves and subsurface rivers. Furthermore, karst is not restricted to limestone but occurs also, though on a much more limited scale, on ultrabasic igneous rocks. Unlike the tropical limestone areas of Indonesia and the West Indies (Lehmann 1936, 1956), the karst landforms in Papua New Guinea have not attracted much attention among geomorphologists until very recently although they are as diverse as, if not more varied than, those of these classical study areas. Again the reason for this is the difficulty of access of which the reader can gain an appreciation from the accompanying photographs (Plates 28–33). However, after an initial reconnaissance study by Jennings and Bik (1962), who first noted that there may be an altitudinal zonation of karst so far not recognised elsewhere, Williams (1971, 1972a, b, 1973) has not only greatly increased our knowledge of Papua New Guinea's karst landforms but has also developed quantitative methods of describing and classifying karst landforms which may well prove to present a major step forward in our understanding of these. In the western part of New Guinea, Verstappen (1964) studied karst in the Star Mountains stressing the lithological and climatic characteristics for the development of these exciting landforms. The present author had the opportunity to study all karst areas of Papua New Guinea on aerial photographs during the compilation of a geomorphological map (Löffler 1974b), but his field knowledge is restricted to a few localities in the Orimo and Kikori areas, the highlands and the Saruwaged Range.

**Minor Karst Features.** Among the small-scale solution features karren and solution notches are probably the most widespread. Both rounded grooves (Rundkarren) and solution flutings (Rillenkaran) are developed on massive dense limestone. According to Jennings (1965) the rounded grooves developed under a former soil and forest cover which has been removed by man's agricultural activities and replaced by grassland. The solution flutings are attributed to subaerial solution. Jennings notes that there is a striking change in karren forms from a predominance of solution flutings to one of rounded grooves as one moves from the eastern to the
western parts of the highlands. A possible explanation of this would be that high population densities and resulting more intensive and frequent use and rotation of land were achieved earlier in the east than in the west. This would have led in the east to an earlier exposure of the limestone due to soil erosion and consequently a longer exposure of the limestone to rainfall.

Solution notches occur commonly at or near the base of low, isolated limestone pinnacles or larger boulders (Plate 27) and are due to the particularly high rate of solution at the zone of contact between soil and protruding rock (Jennings 1971). Many of these notches are now exposed and positioned well above the soil surface, indicating relatively rapid lowering of the ground, possibly because of accelerated soil erosion due to man’s impact (Löffler 1977).

**Major Karst Features.** Most of the landforms of tropical karst are normally described in terms of cone karst and tower karst. Cone karst landscapes are characterised by a sea of roughly cone-shaped or hemispherical hills surrounding enclosed depressions, while tower karst signifies landscapes dominated by more or less isolated tower-like hills with very steep to nearly vertical slopes, generally rising from an alluvial plain. The terms Kegelkarst, cockpit karst or Gunung Sewu karst (Lehmann 1936, 1955) are also used for the former type as well as Kugelkarst if the hills are more hemispherical in shape.

As stressed by Jennings and Bik (1962) and Williams (1971, 1972a) these terms are insufficient to describe adequately the variety and complexity of Papua New Guinea’s karst landforms. Jennings and Bik (1962) recognise two other morphological types which they term pyramid and doline karst and arête and pinnacle karst. Verstappen (1964a) describes a similar type to the pyramid and doline karst from Irian Jaya which he calls labyrinth karst.

Williams (1971, 1972a) stresses the great difficulties in rigorously defining these different types as there are many overlaps and transitions. His morphometric analysis led him to the conclusion that the differences in positive forms are of much less significance than previously thought. Since in all the karst forms mentioned above

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**Plate 27**
Solution notches at the base of an isolated limestone pinnacle in the Mendi area near Nipa
central sinks are surrounded by crudely polygonal topographic divides, Williams describes them all by the collective term polygonal karst.

Based on the studies quoted above and my own observations the following morphological classification of major karst landforms in Papua New Guinea seems appropriate:

1. Karst plains. Karst plains, also called karst border plains (Karstranđebene, Rathjens 1954) or karst margin plains (Jennings 1971), are flat to gently undulating plains on limestone and occur where the base level of karst erosion has been reached. Karst plains

**Plate 28**
Collapse dolines and dry valley on the Müller Plateau

**Plate 29**
Narrow karst corridor on plateau north-west of Kikori
develop largely by lateral solution. A veneer of alluvium of variable thickness generally covers these plains, which makes their exact delineation from aerial photographs difficult.

2. Doline karst. Dolines are closed depressions, circular to oval in plan and conical, cylindrical or bowl-shaped in cross-section. They vary in diameter from less than a metre to over 500 m and similarly in depth (Plate 28). Dolines can be covered with alluvium and can thus be largely sealed off from further karst processes.

3. Plateaux with karst corridors (Plate 29). These are flat-topped plateaux with narrow often discontinuous corridors cut in the limestone preferentially along joints and faults and other lines of weakness. The plateaux with karst corridors are interpreted by the author as former karst plains now being rekarstified by lowering of the base level.

Crevice karst (Jennings and Bik 1962) appears to be similar to plateaux with karst corridors but the limestone is much more cut up vertically and the channels are much smaller in size, more resembling ordinary grikes (J.N. Jennings pers. comm.). The two types could be related.

4. Polygonal karst (Williams 1971, 1972a). This is clearly the most widespread type of karst, characterised by the polygonal shape of the catchment of the depressions, but with great variability in the positive forms. The following styles have been observed:

a) Undulating surfaces incised by close-set depressions with gully-like channels leading to roughly central stream sinks.

b) Pyramid and doline karst consisting of bowl-shaped depressions and concave pyramidal hills (Plate 30).

c) Cone karst or Kegelkarst in the traditional sense consisting of conical or hemispherical hills with steep intervening depressions or cockpits (Plate 31).

d) Tower karst, steep conical and cylindrical hills with rounded tops enclosing deep depressions (Plate 32). The tower karst described by Jennings and Bik (1962) and Williams (1972a) is similar to the cone karst but the karst hills are steeper sided and the relief is higher.

Plate 30
Pyramid and doline karst in the Kikori-Lake Kutubu karst area
e) Arête and doline or arête and pinnacle karst (Jennings and Bik 1962; Williams 1972a) is a unique and spectacular type consisting of bare nearly vertical limestone ridges with razor-blade-like crests surrounding enclosed depressions (Plate 33).

Major Karst Areas. Three areas in Papua New Guinea are dominated by karst landforms: the Southern Fold Belt, the western and eastern parts of New Britain and the north-western part of New Ireland (Fig. 22). Besides these there are numerous smaller occurrences of karst in the highlands, the Saruwaged Range and the westernmost part of the northern coastal ranges.

Plate 31
Cone karst near Kikori

Plate 32
Tower karst south of Porema. Most of the towers are asymmetric because of the steep northerly dip (left) of the limestone.
The extent of the karst country in the southern fold mountains is truly impressive, covering some 20,000 km² in area and stretching over a distance of 450 km as a nearly continuous belt between 50 and 100 km wide from the Gulf of Papua in the south-east to the Irian Jaya border in the north-west. The area contains most of the karst types mentioned above. Near the coast at Kikori low hemispherical hills rise in groups or as individuals from a nearly flat karst plain which is overlain by a thin veneer of alluvium about 1–2 m thick. The plain extends laterally into the hills, which range in what is regarded as a degradational sequence from small, low hummocks at the margin of the plain to larger better preserved hills on moving away from the plain. Northwards the plain extends as a broad corridor upstream on either side of the Kikori River and as a narrow belt at the foot of the escarpment along which the Darai plateau rises from the alluvial plain of the Turama River. The Darai Plateau, structurally a gently north-eastward and south-eastward dipping monocline, exhibits two main karst forms, a plateau with karst corridors forming an extensive area to the south-
east and polygonal karst with a variety of styles but predominantly of the cone karst type occupying the remainder. The existence of the corridor karst next to the cone karst is problematic as there does not seem to be any structural or lithological break between the two karst types. This will be discussed below.

Polygonal karst also covers nearly the entire limestone country to the north-east of the Darai plateau. Here the limestone beds are broken up by the more intense faulting and folding and associated gravity-sliding into parallel trending slabs separated by the underlying clastic sediments. Topographically these limestone slabs form broad plateau-like ridges some of which represent synclines made prominent by relief inversion, while others are simple dipslopes due to faulting or southerly gravitational sliding. The structure of the limestone is generally reflected in the asymmetric shape of the karst cones with longer and gentler slopes in the direction of the dip and shorter and steeper slopes on the opposite side (Plate 32). There is also a tendency for the cones to become higher near the scarps, obviously as a function of the higher relief, but this is far from universal. The greatest concentration of these higher cones, termed tower karst by Jennings and Bik (1962) and Williams (1972a), is to be found along the northern margin of the karst belt, particularly in the Emia River area (Plate 32).

As one moves north-west the karst styles become more complex, as does the gross morphology, which is dominated by north-westerly trending tightly spaced strike ridges and intervening vales which are partly formed on clastic sedimentary rocks and partly filled with volcanic products from the Doma Peaks volcanic centre.

The dominant karst landforms can still be described as cone karst though larger but patchily distributed occurrences of pyramid and doline karst become more frequent. The spectacular arête and doline karst is restricted to three small localities at the northern extremity of the limestone country around Mt Kaijende (Plate 33).

The dominant karst styles in the western part of the fold belt are again cone karst interspersed with pyramid and doline karst which almost entirely cover most of the impressive plateaux, cuestas and hogback ridges which flank the denuded Müller anticline. There is a strong tendency for structural alignment of the karst forms. Where the anticline is still sheeted by the limestone an extensive plateau is formed which is pitted by both aligned and randomly distributed dolines, some of which reach enormous dimensions (Plate 28). The largest doline seen measured 600 m in diameter and 380 m in depth while the average range is between 50 and 100 m in diameter.

The second largest occurrence of karst landforms is to be found in New Britain. Here flat-lying beds of limestone cover extensive areas in the west and east of the island. The limestone varies according to Ryburn (pers. comm.) from a compact and massive limestone to a porous coral algal limestone or a well-bedded bioclastic limestone ranging in age from lower Miocene to lower Pliocene. The thickness varies between about 500 and 1000 m. In contrast to the limestones of the Southern Fold Belt these limestones have experienced very little tectonic deformation except for
uplift since the Pliocene.

Accordingly the karst landforms present a picture of great uniformity, especially in the west where the uplift has been very modest. Here typical tropical cone karst with transitions to pyramid and doline karst dominates the entire karst area. The summit level of the cones presents a gently undulating plain indicating very uniform rates of karst erosion. In the east the karst forms are more of the pyramid and doline type with narrow ridges surrounding enclosed basins. There is much more surface drainage than in the west because of ash cover originating from the volcanoes situated at the northern margin of the karst area. Curved lines of limestone ridges are also present, probably representing reef growth structures.

The third largest karst area is on New Ireland where most of the north-western half is covered by karst landforms. The karst styles change gradually from pyramid and doline karst in the south-east to cone karst in the north-west. Curved lineations of ridges reflecting reef growth patterns are again common. The landforms are developed on a gently north-eastwards dipping sequence of coral and algal biostromal limestone ranging in age from lower Miocene to Pliocene and possibly Pleistocene. The limestone is chemically pure, off-white to creamy in colour and varies from completely recrystallised at the base of the sequence to slightly recrystallised at the top (Höhnen 1970).

The present summit level of the karst cones is not in strict accordance with the bedding of the limestone. It is therefore assumed that the limestone was truncated by an erosion surface prior to the main uplift and tilting which was followed by the karst erosion. The present karst topography is therefore of very young age.

**Minor Areas of Karst.** Large areas of the Huon Peninsula and its western continuation are covered by limestone, which extends from the north coast to the summit areas of the Saruwaged and Finisterre Ranges. The largely detrial Miocene limestone is massive and well bedded and attains over 1000 m in thickness. Structurally the limestone sheet the Finisterre-Saruwaged geanticline. In the summit areas of the Saruwaged Range the limestone is subhorizontal, forming an irregular plateau surface ending in a dramatic cliff to the south, but to the north it dips at an angle of about 15°, forming a well-defined structural surface (Plate 17). In many areas, particularly to the west, the limestone is severely eroded and the underlying volcanics exposed. Most of the karst is concentrated on the summit plateau of the Saruwaged Range and its eastern continuation, the Cromwell Range. The summit surface is highly irregular with rounded hills rising to 300 m above the general level of the plateau. In the western part this has been accentuated by glacial erosion. The plateau is pitted with dolines which show up clearly in the grassland areas covering most of the higher western part, but occur also, though patchily, in the lower eastern part of the plateau. Most of these grassland areas in the east are situated in enclosed basins, the largest of which measures over 20 km² in area. The dolines occur preferentially along smaller drainage lines many of which have become inactive as a result. The main through-going stream, the
Mongi River, has, however, maintained its surface flow.

In the far north-west of Papua New Guinea there are several smaller occurrences of karst between the middle Sepik and the north coast. Most of the karst is again of the cone karst type with the exception of an area south of Mt Bougainville (near the international border) where poorly developed pyramid and doline karst is present.

On the Trobriand Islands karst development is largely restricted to caves and dolines. This will be discussed in more detail in Chapter 4 section on subcutaneous erosion.

Altitudinal Zonation of Karst. Jennings and Bik (1962) first pointed out that there is a possible altitudinal and thus morphoclimatic zonation of karst landforms in Papua New Guinea with certain types occurring preferentially at certain altitudes. They stress that their findings are based on limited observations only and that there is by no means a straightforward picture; it is therefore incorrect to quote them as having established that karst landforms in Papua New Guinea vary simply as a function of altitude (Corbel and Muxart 1970). Williams (1973) further elaborates on this theme and though he still basically supports Jennings’s and Bik’s findings he emphasises more strongly the significance of lithology and stage of karst erosion. The author, having access to more complete air-photo coverage than the previous authors, found that the variation in altitude of the different karst forms and styles is greater than previously thought (Fig. 23) and that therefore a morphoclimatic explanation for their distribution does not seem to be tenable.

The dominant type is quite clearly the cone karst that covers an estimated 80 per cent of the karst country and ranges in altitude from near sea level to about 3000 m. There is a great variety in the shape of the hills, ranging from low hemispherical to high conical or tower-like, but the differences can mostly be explained in terms of general relief and rock type. Hemispherical hills with smooth convex slopes occur preferentially in soft chalky limestone while the steeper conical and tower-like hills occur in hard, relatively dense

Figure 23
Altitudinal distribution of karst landforms

![Altitudinal distribution of karst landforms](image-url)
often recrystallised rocks. Increased tectonic relief tends to favour higher cones.

The pyramid and doline karst is also present in a wide altitudinal range from 100 m to 2900 m but since it occurs rather patchily within the cone karst areas a lithological explanation appears more likely than a climatic one, as also assumed by Williams (1973). The arête and doline karst is the only type which has a clearly limited altitudinal range but it must be pointed out that its occurrence is limited to three small localities in the vicinity of Mt Kaijende. It has not been observed in any other limestone area of similar altitude and is therefore more likely to be a result of local lithological or structural conditions than of the prevailing climatic conditions.

Doline karst occurs over virtually the whole altitudinal range in Papua New Guinea. The lowest (very small) occurrence has been observed by the author in the south-eastern part of the Fly Platform some 30 km north of Mabaduan. Here the nearly flat alluvial plain is densely pitted with roughly circular to oval water-filled depressions 40–60 m in diameter. The alluvium here is underlain by limestone at a depth of about 1 m. The dolines appear to be all inactive and the whole plain is under water during most of the wet season.

The largest concentration of dolines is to be found at higher altitudes. They are particularly prominent on the summit plateaux of the Müller Ranges and Saruwaged Range but are distinctly different in appearance. While on the Müller Plateau dolines are more like large bomb craters with diameters of the order of 100–300 m and more, with nearly vertical side walls (Plate 28), on the Saruwaged Plateau the dolines are much smaller, generally saucer-shaped, and resemble closely the dolines found in karst areas of the temperate zones. The Müller Plateau dolines appear to be formed largely by collapse due to a very active underground drainage system, also indicated by the frequent occurrence of dry valleys and large underground rivers rising from the edge of the plateau with large discharges of water. According to the Papua New Guinea Speleological Expedition (1974) a great number of these dolines are developed in calcareous siltstone which locally overlies purer limestone. The more active solution processes in the purer limestone lead to undermining, formation of cavities and eventual collapse of the siltstone forming the roof. The dolines on the Saruwaged Plateau on the other hand are predominantly solution dolines formed by surface solution of the limestone.

Solution dolines at similar altitudes have been reported by Verstappen (1964a) and by Shephard (unpubl. 1965) in the Star Mountains. The predominance of the doline karst at altitudes above 3000 m and the absence of any polygonal karst except for the arête and doline karst supports the notion that the tropical karst forms are replaced by temperate doline karst above this altitudinal limit.

However, a gradual transition with increasing altitude from a polygonal karst to a doline karst on the same rock type has nowhere been observed. For instance in the Saruwaged Range doline karst is developed on detrital limestone over an altitudinal range from 2000 m to 4000 m with no signs of polygonal karst.
Plateaux with Karst Corridors. The occurrence of plateaux with karst corridors or crevice karst in close proximity to cone karst is puzzling. Ground observations showed that the lithology was identical in both cases and that there was no alluvial veneer or volcanic ash cover on the plateau. It is also important to note that the cones of the cone karst area rise about 30–40 m above the level of the plateau (Fig. 24). Only in close proximity to deeply incised streams are the summits of the cones at the same level as the plateau. This rules out the possibility that the plateau represents the original surface from which the cones developed, since the plateau would have to be higher than or at the same level as the summit level of the cones. A probable explanation would be that two distinctly different processes have formed the two landform types. These are vertical karst corrosion in the cone karst area, and lateral solution and planation on the plateau at a time when the plateau area was close to the base level of erosion. The plateau is therefore interpreted as a former karst plain or karst margin plain which became inactive when the area was uplifted to the present height. Incision into the plateau followed but only the largest streams maintained their surface flow. The alluvial cover was removed and karstification started. This would also explain the occurrence of the cone karst areas on either side of deeply incised streams. Here the alluvial cover was removed much earlier than on other parts of the plateau. The plateaux with karst corridors can therefore be regarded as an early stage in the development of karst but not in the sense that the present cone karst area has developed from it. Karstification started at a much later date when the former karst plain that had developed parallel to the cone karst became inactive because of a change in base level most probably caused by uplift. The incipient development of tropical cone karst is well demonstrated in an area east of Middletown where the plateau is criss-crossed by a dense pattern of corridors obviously following structural lines of weakness. The more rapid solution processes operating along these lines will eventually lead to widening of the corridors and the formation of cones. The pattern shown here has a striking resemblance to the growth model of the evolution of karst established by Williams (1972a, Figs. 18, 19).

Karst Processes. Much of karst research in the past two decades has been concerned with the fundamental problems: why the karst landforms in the tropical areas of the world are so strikingly different from karst landforms in the temperate regions, and what role is

Figure 24
Karst corridors and cone karst on the Darai plateau
played by the rate of solution in respect of the formation of these peculiar landforms. It would be far beyond the scope of this book to go into any detail on this subject and the reader is referred to Gerstenhauer (1968/69), Jennings (1971) and Sweeting (1973), who deal with this at length, discussing most of the relevant literature.

In the Papua New Guinea context the following should be briefly set out:

The karst landforms in Papua New Guinea are predominantly of the tropical type. There are great variations within these forms which appear to be due to differences in lithology, structure and stage of development; within the tropical types there does not seem to be an altitudinal or morphoclimatic zonation. The rate of karstification appears to be quite rapid although this has not been quantified.

The amount of carbonate in rivers draining limestone areas does not seem to be exceptionally high, but this should be seen in relation to the high amount of precipitation and run-off (Jennings 1973). The hypothesis of Corbel (1959) that the rate of solution in these tropical areas is slower because of the lower saturation equilibria for carbon dioxide and carbonate solution in warm waters has met with considerable opposition among karst workers, who stress the role of biogenic carbon dioxide and organic acids, which of course are directly dependent on temperature and precipitation, as the main factors promoting rapid limestone solution (Jennings 1971; Jakucs 1973). This is very important in Papua New Guinea where nearly all the karst areas have a thin soil and dense vegetation cover. A dense root mat is nearly always developed which will have a high biogenic carbon dioxide content as well as a high content of organic soil acids. In addition the higher the temperature the higher is the rate of chemical reaction and the more favourable are the kinetic factors affecting mass transfer of solutes from rock surfaces (J.N. Jennings pers. comm.). Moreover, solution processes have to be seen in relation to the amount of water passing through or over the rock at any given time. The problem is, however, far from solved and further research will be necessary to determine more precisely the roles of lithology, structure, climate and geomorphic history of the landforms.

Landforms of Glacial Erosion

Glacial landforms represent a distinct assemblage that contrasts sharply with the surrounding non-glacial landforms. Although Papua New Guinea's mountains have now been free of ice for some 10,000 years the erosional and depositional landforms created by past glaciation(s) have been surprisingly well preserved. Post-glacial fluvial erosion and mass movement have hardly modified the glacial terrain, which indicates a relatively slow rate of subsequent erosion and mass movement, and a marked persistence of these relict landforms.

The high mountains of Papua New Guinea have attracted a disproportionately large number of geomorphologists and other natural scientists because they bear undoubted evidence of past climatic changes and there is increasing realisation that a proper
understanding of the history of these mountains may well be of vital importance for our understanding of the past climatic and geomorphological history of this region. Also the high mountains of Papua New Guinea occupy an important biogeographical position, a kind of link between the mountain environments of the Asian land mass and islands on the one hand and the Australian and New Zealand alpine environments on the other.

The Pleistocene glaciation of Papua New Guinea has been summarised by Löffler (1972a) and the following is based on that paper. Traces of Pleistocene glaciation are very well preserved on about 20 mountains that exceed the Pleistocene snowline, which was at about 3550 m (Fig. 25).

The type of glaciation varied with topography from ice cap glaciation on plateaux and broad volcanic shields to valley glaciation on deeply incised mountain terrain. Examples of the first type are Mt Giluwe, the Saruwaged Range (Plate 17), Mt Albert Edward (Plate 34) and Mt Scratchley. For the second type Mt Wilhelm (Plate 35), the Kubor Range and the Star Mountains are typical.

**Figure 25**
- Glaciated areas

**Plate 34**
Mt Albert Edward plateau with glacially smoothed and rounded rock faces, overdeepened, swampy basins and cirques (background)
The most extensive ice cover was on Mt Giluwe (4368 m) where the broad summit area was covered by a more or less continuous ice cap covering 188 km² and extending down to 3200–3500 m. On the north-western, northern and southern slopes valley glaciers up to 400 m thick descended a further 400 m from the ice cap, eroding deep glacial troughs and depositing extensive lateral moraines along its margins. The total amount of valley deepening is probably not due to glacial erosion alone since preceding fluvial incision seems to have been a major factor determining the depth of the valleys. Glacial erosion appears to have contributed mainly to the widening of the valleys.

On the south-western slopes where the ice accumulated on a largely undissected planeze surface the ice cap did not break up into outlet glaciers but ended as a broad shield at about 3200–3300 m, accumulating a broad belt of terminal moraine deposits about 1 km wide and 20–30 m thick. A similar situation existed on the eastern slopes but here short ice lobes extended from the ice cap.

A multitude of low but distinct recessional moraines characterises much of these undissected south-western and eastern slopes (Plate 36). Over 20 single moraine ridges in five groups can be distinguished. These five groups of recessional moraines, named stages 1 to 5 from old to young, represent oscillations during the retreat of the ice cap.

Löffler (1972a) gave C¹⁴ dates of about 7000 years B.P. and 3500 years B.P. for the two youngest stages (stages 4 and 5) but this needs revision in the light of new information*.

The minimum age of stage 2 is about 12,000 years B.P. while the ages of the following recessional stages are between 12,000 and 9000 years B.P. The date of 9000 years has been obtained from a summit bog on Mt Giluwe and gives a clear indication that at this time no ice was present on Giluwe. These dates are compatible with Hope's (1973) more complete sequence of dates from Mt Wilhelm.

Extensive glacial cirques are grouped around the main and eastern peaks of Mt Giluwe. They have very steep to precipitous back walls up to 700 m high and broad flat floors which are not over-

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* The material was collected by G. Hope and the author in 1973 and the dating was carried out by H. Polach, A.N.U.
deepened. This supports the notion that most of these cirques represent merely glacially modified amphitheatre valley heads and were not subject to a great deal of glacial excavation. Amphitheatre valley heads are typical of a great number of non-glaciated volcanoes in Papua New Guinea and other volcanic areas like Hawaii and the New Hebrides, and are due to structural control by the hard, outwardly dipping lava beds (Stearns 1966; Ollier 1969a).

The glaciation on Mt Wilhelm (4510 m) was typically alpine in character with extensive thick valley glaciers forming deep glacial troughs and stepped valley profiles, originating from over-deepened cirques with near-vertical headwalls (Plate 35).

Most of the cirques are situated west and south-west of the generally north-west trending summit ridge. This asymmetry in the distribution of cirques is attributed by Reiner (1960) to the preglacial asymmetry of the Mt Wilhelm massif, favouring extension of the ice in a westerly and south-westerly direction. The asymmetry is, however, not restricted to cirques and glaciers alone but the slope profiles also appear to be asymmetric with smoother and 'gentler' slopes facing east and irregular and steeper slopes facing west and south-west. An alternative explanation could be a climatic one. In tropical mountains east-facing slopes generally receive more insolation than west-facing slopes, because the rapidly increasing cloud cover during the day makes insolation during the afternoons less effective than during the usually cloud-free mornings. This difference has also been used to explain the distribution of certain plant communities on Mt Wilhelm which extend further up on east-facing slopes than on west-facing slopes (J. Smith pers. comm.). The difficulty of the climatic explanation is that this

Plate 36
Mt Giluwe, which supported a plateau type glaciation. Lateral moraines of main glaciation in foreground and numerous low recessional moraines and small glacial basins in the middle distance.
east-west asymmetry is generally not present on other glaciated mountains in Papua New Guinea except for Mt Albert Edward, where it is largely a result of the structure of the steeply easterly dipping schist (Löffler 1970a). One reason for the lack of this asymmetry on the other mountains could be that the rock structure present (horizontal to gently dipping lavas on Mts Giluwe and Hagen, well-bedded limestone on the Saruwaged Range) do not favour its development. In addition the depth of the preglacial incision will have been an important factor. The differential insolation on west and east exposed slopes can only operate significantly if valleys are deep, and this would clearly favour Mt Wilhelm, where valleys are considerably deeper than on the other mountains.

Most glacial valleys extending from the summit of the Mt Wilhelm area have lateral moraines forming distinct smooth ridges with sharp even crests that curve towards the valley floors to form terminal moraines at about 3200 m (Plate 35). In contrast with Mt Giluwe recessional moraines are rare. This could be due to the unfavourable topographic conditions. It is also probable that the thick valley glaciers were much less sensitive to minor climatic oscillations during the deglaciation than thin ice caps such as occurred on Mt Giluwe.

The glacial landforms on all the other peaks that had an ice cover during the Pleistocene were essentially similar to either the Mt Giluwe type ice cap or the Mt Wilhelm type valley glaciation.

The snowline was at about 3500–3550 m on Mt Giluwe and Mt Hagen and most of the other highland peaks except for Mt Wilhelm where it appears to have been slightly higher.

On Mt Albert Edward it was at 3600–3650 m and on the Saruwaged Range at 3650–3700 m. While the higher snowline on Mt Albert Edward is probably due to the more southerly position of the mountain the relatively high snowline on the Saruwaged Range is unusual. From evidence in other tropical areas one would expect the snowline to rise from the coast to the interior. Here the opposite is the case. The most likely explanation for this discrepancy is post-glacial uplift of the Saruwaged Ranges, and this is supported by the studies of Chappell (1974a), who established that the Huon coast has been rising at considerable rates (see section on raised coral terraces).

Nearly all glacial features in Papua New Guinea have been attributed to the last glaciation. There is now mounting evidence that this last glacial period was more or less contemporaneous with the final Würm or Wisconsin glaciation in the northern hemisphere. The glaciers in Papua New Guinea had receded by 12,000 B.P. and by 9000 B.P. all the mountains were ice-free.

There is also increasing evidence for glacial periods preceding the last one. Blake and Löffler (1971) reported palagonitic breccia on Mt Giluwe, which must have erupted under a substantial ice cover well before the last glaciation. Moraine material overlain by a volcanic lava flow also points to a pre-Würm glacial period (Löffler 1972a). The minimum age of the palagonitic breccia is 290,000 years B.P. and the lava flow overlying the moraine material has an age of 320,000 years B.P. Williams and others (1972), although not directly concerned with glaciation, have evidence for cooler condi-
tions in the highlands before 33,000 years B.P. The notion that the central mountains of New Guinea escaped earlier glaciations because they were simply not high enough at the time (Verstappen 1964b) can certainly not be upheld. There still remains the problem that the last glaciation was the most extensive, in contrast with the situation in Europe and North America. This is, however, typical of smaller mountain glaciers in the middle and low latitudes and could be because the last glaciation, although of shorter duration, had a more severe climate than its predecessors (Galloway and others 1973). Because of this the large continental ice sheets would not be able to develop their maximum size, but smaller glaciers could develop rapidly and wipe out the traces of preceding glaciations.

In Papua New Guinea there was no reappearance of ice after 9000 B.P., but in neighbouring Irian Jaya at least three minor ice advances have occurred since 2900 B.P. (Peterson and others 1973). Hope (1973), who investigated the vegetation history of Mt Wilhelm, has pollen analytical evidence that a general cooling took place after 5000 years B.P. and suggests that a small temporary snow cap may have been in existence on the summit of Mt Wilhelm above 4400 m.

**Landforms of Marine Erosion**

The scarcity of actively cliffed coasts and the relative insignificance of coastal erosion is typical of humid tropical coastlines (Tricart 1972). This is partly due to protection of the coast by extensive depositional landforms such as deltas, estuarine plains, tidal flats and beach complexes, which in turn reflect the greater supply of materials to the coastal zone caused by higher river discharges, more active weathering and slope processes associated with this climatic zone. Barrier reefs and fringing reefs also reduce the erosive capacity of the waves but not to the same degree. Also important is the absence of strong winds and the resulting relatively low wave energy in this climatic zone, so that coasts are not effectively eroded nor is material brought down by the rivers distributed widely. This generalisation holds true for large parts of the coastline of Papua New Guinea, but because much of its coastline has been rising at considerable rates steep coastlines are more common than in more stable areas of the humid tropics. However, cliffs are usually not well-developed and most of the steeper coastline can be described as vegetated bluffs with minor cliffing at the base where continuous sea spray prevents the establishment of higher plant communities.

Along the south coast bluffs and cliffs occur only in a few areas where the foothills reach the sea. They are prominent around smaller headlands and promontories such as the bluff west of Kerema, Redscar Cliff and Cape Suckling, and along much of the coast between Port Moresby and Paira Point (Plate 3). Further east between Mullins Harbour and Samarai a nearly continuous cliffed coastline is developed. The coastline here has the usual characteristics of a drowned coast with numerous narrow headlands and deep bays which, in contrast with the embayments further west, have not been filled in, because their catchments are small and relatively low.
Cliff formation in this area is favoured by the direct exposure of the coastline to southerly and south-easterly winds and the absence of a protecting barrier reef system.

The north coast of mainland Papua New Guinea is for the most part distinctly different from the south coast. It is a typical 'Pacific type coast' (Suess 1888) with a relatively straight seaboard and mountain ranges trending parallel to the coast and rising steeply from the sea. Evidence of active uplift is plentiful all along the coastline.

In spite of the predominance of steep seaboards, cliffed coasts are not very extensive. Large areas have narrow belts of beach ridges and steeply sloping coastal fans which overcome wave action, such as the coast between Wasu and Astrolabe Bay and Wewak-Aitape and Vanimo. In other areas such as between Goodenough Bay and East Cape at the eastern end of the mainland the coastline is protected from significant wave action by surrounding peninsulas and large offshore islands and no cliffs are being formed in spite of steep slopes.

Low cliffs with notches are common in areas where raised coral platforms are developed, such as between Morobe and Salamaua, along the north-east coast of the Huon Peninsula and around Madang.

The characteristics of a rising coast along the north coast are interrupted only along its north-east section between Morobe and Cape Vogel where subsidence or lack of uplift of the Cape Vogel basin has produced a coastline very much like that found along the south coast. There are extensive low-lying alluvial plains.

Plate 37
Cape Nelson with deeply dissected Mt Trafalgar volcano. The coastline has typical characteristics of ria coast with deep narrow inlets. Fringing reefs are developed around the headlands.
extending seaward into tidal flats and deltas. Most conspicuous is the drowned landscape around Cape Nelson where an attractive ria coast with deep narrow inlets is developed (Plate 37). It has been suggested that this drowning of the coast is caused by local subsidence due to the weight of the pile of volcanic rocks overlying

Figure 26
Wave cut notches on New Ireland's coast (after Christiansen 1963) (reprinted by permission Geografisk Tidsskrift)

Figure 27
Bathymetry of the Waria Canyon (after van der Borch 1972) (reprinted by permission Bureau of Mineral Resources, Australia)
clastic sedimentary rocks at depth (J.G. Best unpublished), but the post-glacial rise in sea level would also account for this feature.

Low coral cliffs, often with double notches, are developed around much of the coral fringed coastline of New Ireland and southern New Britain. The cliffs here are fronted at the seaward side by a reef flat, largely of dead coral material. The cliff base is densely pitted by solution holes and basins extending seaward into narrow channels. Between cliff base and near vertical cliff face a single or double notch is usually developed (Fig. 26) (Christiansen 1963). The widespread occurrence of the upper notches and the dead coral reef platforms give evidence of active emergence caused by tectonic uplift.

An interesting feature of the north coast is the existence of deep submarine canyons some of which, like the Waria and Markham canyons, have been cut well below Pleistocene eustatic sea level lows (Fig. 27). According to Borch (1972) these canyons were probably initiated during the Tertiary and have maintained their courses during the growth of the deltas because of their high bedloads. In addition repeated incision and filling in of parts of the canyons during the Pleistocene sea level fluctuations occurred.

**VOLCANIC LANDFORMS**

As in most areas of the circum-Pacific mobile belt active or recently active volcanoes are prominent features in the Papua New Guinea landscape, and are of particular importance to the population both as welcome suppliers of fertile land and as potent threats to safety. Although the history of eruptions in Papua New Guinea does not parallel in human loss or destruction that of some other better known areas in the world, in terms of volcanic violence the eruptions certainly measure up to most other volcanic areas and the eruption of Mt Lamington in 1951 must rank among the great volcanic events of this century.

Papua New Guinea contains a great variety of volcanic landforms ranging from the ever present strato-volcano to lava shield, ash cone, scoria cone and mound, mamelon, spine and caldera. Among the volcanic processes lava flows, ash flows, lahars and nuées ardentes are the more important ones.

Probably no other landform type in Papua New Guinea has had the same attraction for geologists, geomorphologists and laymen and consequently there is a great number of published papers and unpublished records available on this subject.

On the Papua New Guinea mainland volcanoes occur in two irregular clusters. The largest of these is centrally located in and south of the highland region. A smaller one occurs in the southeast in the Cape Vogel Basin. A third cluster is on the D'Entrecasteaux Islands. In addition there are two chains of volcanoes, so-called volcanic arcs, in the Bismarck Sea which together contain the largest number of active, dormant and extinct volcanoes in Papua New Guinea. The volcanic areas and the distribution of active and potentially active volcanoes are shown in Fig. 28a, b.
The Highland and Southern Fold Mountain Volcanoes.

The highland and southern fold mountain volcanoes extend over a large area between about 5°S and 7°S and 142°30'E and 145°E. They consist of 15 major centres including Mt Bosavi, Mt Sisa, Doma Peaks, Mt Ialibu, Mt Giluwe, Mt Hagen, Sugarloaf, Mt Murray, Mt Soaru, Mt Karimui, Crater Mts, the sister volcanoes Mt Duau, and Mt Yelia (Plates 7, 25, 38).

The great majority of these volcanoes are steep-sided conical strato-volcanoes. Mt Giluwe is an exception. It is a lava shield volcano with gently sloping convex sides, and has numerous eruptive sites on its lower flanks marked by scoria cones with craters and scoria mounds without craters, and lava domes and maars (Plate 38) (Blake and Löffler 1971).

Doma Peaks and Crater Mts are also complex volcanic centres with numerous cones and craters and associated lava flows. Crater Mts consist of an extensive and thick volcanic basement from probably three old centres surmounted by about twelve small

Figure 28
a. Volcanic areas; b. Distribution of active and potentially active volcanoes

Figure 28(a)
- Moderately dissected cones and domes
- Strongly dissected cones and domes
- Moderately dissected volcanic footslopes and volcano-alluvial fans
- Strongly dissected volcanic footslopes and volcano-alluvial fans
- Volcanic plateau

Figure 28(b)
- Major eruptive centre with recorded eruption
- Major eruptive centre, thermally active, no recorded eruption
+ Thermal areas
well-preserved cones, most of which have craters and well-preserved lava flows clearly discernible on aerial photographs (Bain and others 1975).

All the volcanoes except Mt Hagen are deeply incised on their southern and to a lesser degree their northern flanks, while their eastern and western flanks are generally less dissected, forming planeze surfaces (Plate 25). Mt Hagen, the northernmost volcano, is more deeply incised towards the north and west.

This asymmetry in dissection probably reflects the drainage and slope of the pre-volcanic landscape in most of the area. After the eruptions, the lowest base levels were to the south (except for Mt Hagen where it was to the north) and a southerly preference would therefore be expected. Only in the case of Mt Hagen are these trends reversed to the north. Higher rainfall on the southern slopes associated with the prevailing south-east trades may have been a factor contributing to the strong dissection of the south slopes on the southernmost volcanoes such as Mt Favenc, Mt Duau, Mt Murray and possibly Mt Bosavi (Ollier and Mackenzie 1974). The other volcanoes are at too high an elevation to be under the influence of the trades (Brookfield and Hart 1966).

The pre-volcanic drainage system was of course severely interrupted in places. Volcanic material from Mt Hagen volcano blocked the westerly and then northerly draining Wahgi and this led to filling up of the Wahgi basin and eventually to reversal of the drainage. The merging footslopes of Mt Soaru and Mt Karimui temporarily blocked the Tua River to form a lake which seems to have been quickly drained as the overflow cut through the barrier.

None of the highland volcanoes is active at present and probably most of them have been extinct for a considerable time. K-Ar dates from Mt Giluwe indicate that the last eruptions took place about 200,000 years B.P. Mt Giluwe and Mt Hagen and probably also Mt Kerewa were covered by ice, at least during the last glacial period (Löffler 1972), and there are no signs of any disturbance to the glacial landforms by volcanism, although some very thin ash layers can be found in the alpine peat soil. The origin of the layers

Plate 38
Mt Giluwe rising from the Kaugel intramontane basin (middle distacne). Grass covered summit area coincides approximately with the former extent of the ice cap.
is not known and a non-highland source is possible (R. Blong pers. comm.).

Solfataric activity has been recorded from Mt Yelia and Doma Peaks, and according to G.A.M. Taylor local inhabitants of the Doma Peak area claim that it erupted last century destroying several villages (quoted from Mackenzie 1970). It is also likely, judging from the state of preservation of the Crater Mts volcanoes and lava flows, that eruptions could have taken place recently. State of preservation can, however, be deceptive and is not necessarily a good indicator of age. For instance along the southern slopes of Mt Giluwe a group of well-preserved scoria cones and mounds are probably over 50,000 years old (R. Blong pers. comm.).

Cape Vogel Basin and D’Entrecasteaux Volcanoes

The Cape Vogel basin volcanoes are groups of closely associated and coalescing volcanoes, consisting of the Lainmngton group in the west, the Cape Nelson group in the east and in between the Managalase plateau and several small volcanoes along the Owen Stanley Ranges.

The Mt Lamington group consists, from north-west to south-east, of Mt Lamington, an active strato-volcano, the Hydrographers volcano, a deeply dissected and clearly extinct strato-volcano, the Managalase plateau, a complex of volcanic centres, and finally the Sesara volcano, like the Hydrographers an extinct volcano in an advanced state of dissection.

Mt Lamington was hardly recognised as a volcano when in 1951 it suddenly erupted most violently causing widespread destruction and heavy loss of life. The history of this eruption from the early stages of activity has been described in detail in a fascinating account by Taylor (1958), who with disregard for his personal safety observed most of the events at first hand.

The eruption included catastrophic nuées ardentes (glowing clouds) – explosions producing deadly hurricanes of incandescent volcanic ash lubricated by the gases given off by the ash and exploding lava fragments. Within a few minutes all living matter in an area of some 225 km² in the vicinity of the volcano was destroyed (Plate 39), and there was no escape for the 3000 people living there. Ash showers were recorded as far away as Port Moresby some 120 km on the other side of the Owen Stanley Ranges where they caused temporary closure of the airport.

Less violent explosions and nuées ardentes continued for two months after the initial climactic explosion and for six months powerful mudflows swept down the river channels, depositing sand, gravel and large boulders in swathes up to one mile wide and twenty miles long. Taylor (1958: 56) gives a vivid eyewitness description of these sudden mudflows:

They rose with dangerous rapidity, reaching full flood within a few minutes and, at times, advancing down the valley as a wall; some of them were hot viscous streams, moving at about 10 miles per hour and looking rather like a conveyor belt loaded with logs. Unlike the later roaring, less viscous torrents, their movement was silent apart from the rumbling impact of large boulders,
Mt Lamington shortly after the eruption in 1951. All living matter in the vicinity of the volcano was destroyed by catastrophic nuées ardentes.

The Hydrographers volcano, the north-western footslopes of which merge with the south-eastern slopes of Mt Lamington, is a much older volcano, being in an advanced stage of dissection. It was active 650,000–700,000 years ago and has since been eroded to a ridge-and-V valley landscape, the average rate of erosion being about 75 cm per 1000 yr (Ruxton and McDougall 1967). The original volcanic shape is recognisable only from the radial drainage and crest pattern and some triangular footslopes.

The Managalase plateau which joins the Hydrographers to the south-east is a much more complex volcanic structure formed on semi-buoyant in the dense stream. The flows rarely lasted longer than an hour or two and subsided rapidly, leaving on the margins of the stream and the adjacent valley floor stranded boulders and a deep layer of mud.
irregularly sloping and largely flat-lying lava flows, above which rise a whole variety of smaller volcanic landforms including mostly well-preserved lava cones and domes, cinder cones, ash cones, scoria cones and mounds (Fig. 29) (Ruxton 1966a).

The Sesara volcano is the oldest volcano of the group. It lies to the south of the Managalase plateau and has been dated at 5.4-5.75 million years (Ruxton and McDougall 1967). It is similar to the Hydrographers but in a more advanced stage of dissection.

The two volcanoes of the Cape Nelson group are Mt Victory and Mt Trafalgar (Plate 37). Mt Victory is, like Mt Lamington, an active strato-volcano with lava domes and flows in a rugged summit area which is flanked by gently sloping footslopes and volcanic alluvial fans. It has several young adventive cones on its flanks and its last eruption took place in the 1890s.

Mt Trafalgar, which is connected to Mt Victory by a saddle 800 m high, resembles the Hydrographers volcano in its state of dissection and it is assumed they are of similar age. However, its lower footslopes to the north-east extending below sea level are better preserved, forming extensive though densely dissected planezes. The valleys are cut below present sea level and have subsequently been drowned, producing a spectacular ria coast (Plate 37).

Besides these major volcanic areas there is Waiowa volcano, a small, isolated, recently formed volcano about 50 km south-east of Mt Victory at the foot of the Gorupu Mts. It erupted in 1943 in very much the same way as did Mt Lamington eight years later, with nuees ardentes causing much destruction in the vicinity and mudflows filling streams and rivers (Taylor 1958).

The volcanic centres of the D'Entrecasteaux Islands are concentrated at the southern end of Goodenough Island and the south-western and south-eastern extremities of Fergusson Island. They are all well-preserved large strato-volcanoes though no eruptions have been recorded. However, three volcanoes on the south-eastern extremity of Fergusson Island are thermally active and are regarded as potentially active.

The Southern Bismarck Sea Volcanic Arc

The greatest number of active, dormant and extinct volcanoes is situated in the Bismarck Archipelago where they are aligned along two volcanic arcs, the southern and northern Bismarck volcanic arcs. It is impossible to describe or even mention all of these volcanoes and therefore only the major ones are briefly dealt with.

The southern Bismarck volcanic arc stretches over 1000 km from the Schouten Islands in the west to Rabaul in the east. Some authors do not regard the Rabaul volcanoes as part of the arc proper because...
of the marked structural differences between the Gazelle Peninsula and the rest of the island. There are well over fifty volcanic centres, dominantly andesitic strato-volcanoes rising from the sea or surrounding plains. The best known are Manam, Karkar and Long Island volcanoes off the north coast of Papua New Guinea, Langila, Pago, Ulawun and Bamus volcanoes in western and central New Britain, and Vulcan and Tavurvur near Rabaul.

Manam volcano is a steep nearly symmetrical cone rising abruptly out of the sea to an altitude of about 1300 m. There is no well-defined crater but the summit is broken by two steep valley heads leading into deeply incised valleys through which during recent eruptions thick blocky lava flows moved, reaching the sea or terminating close to it. Manam has a long record of eruptions, the latest of which took place in 1973.

Karkar volcano is scenically one of the most attractive volcanoes of Papua New Guinea, combining the beautifully symmetrical shape of the cone with long smooth intensively cultivated footslopes leading down to a coral fringed coastline. The summit area is formed by two concentric calderas with nearly vertical walls encircling a central dome which has grown from the caldera floor during recent eruptions (Plate 40). The volcano was reported as active in 1895 (Sapper 1917b) and remained dormant until 1973 when new eruptions took place which have continued intermittently since.

Long Island is an irregularly shaped island probably representing a truncated cone (Fisher 1957). The central part of the island is formed by a spectacular caldera, the floor of which contains a lake. A small active cone has risen from the lake floor during recent eruptions. Fisher (1957) considers that formation of the caldera

Plate 40
Karkar volcano with two concentric calderas and a central dome or mamelon that has grown from the caldera floor during recent eruptions.
must have taken place in the last few hundred years during an eruption of extreme violence.

Umboi Island further east is the largest island of this chain and consists of several volcanic centres forming an irregular pattern of steep cones, the footslopes of which meet in the central parts of the island to form larger plateau-like areas.

Talawe and Tangi, two large strato-volcanoes rising to over 1500 m, form the western extremity of New Britain. Smaller eruptive centres occur in clusters of satellite cones with mostly well-developed craters on the saddle between the two coalescing volcanoes as well as on the east flank of Talawe. These latter cones, called the Langila Craters, have a recent history of strongly explosive eruptions, the earliest of which was recorded towards the end of the last century. In 1954 major explosive activity began again and has continued intermittently ever since (Johnson and others 1971).

Towards the east the footslopes of Tangi and Talawe coalesce with those of two much older and more dissected volcanic centres, Schrader and Andewe, which have probably not been active since the early Pleistocene.

The greatest concentration of volcanoes in New Britain is to be found along the north coast of central New Britain. They fall into three main groups, the volcanoes of the Willaumez Peninsula in the west, the Cape Hoskins Group of volcanoes in the centre and a north-eastern group around Bamus and Ulawun volcanoes. Most of the volcanoes are well documented (Fisher 1957; Lowder and Carmichael 1970; Blake and Bleeker 1970; Blake and McDougall 1973; Johnson and others 1972).

The Willaumez Peninsula is entirely volcanic except for the coral fringed coast. It consists of a north-south aligned chain of eleven strato-volcanoes, several rhyolitic extrusions and numerous small cinder cones. The north-south alignment of the volcanoes is in marked contrast to the prevailing east-west trend of the volcanic arc and may indicate a major fault zone at right angles to the coastline (Ryburn 1974). The most remarkable feature of the area is the Dakataua volcano, a roughly elliptical caldera 10 × 13 km in size at the northern tip of the peninsula. The caldera contains a large lake shaped like a horse shoe partly surrounding a central vent volcano, Mt Makalia, which probably last erupted at the beginning of this century (Fisher 1957). Like the other calderas of Papua New Guinea the Dakataua caldera has been formed by collapse along a nearly elliptical ring fracture.

The Cape Hoskins group of volcanoes consists of eleven strato-volcanoes in various stages of dissection ranging in age from about 900,000 years to recent (Blake and McDougall 1973). The group includes the active Pago volcano which consists of several tongue-like lava flows and a central cinder cone inside an older caldera (Plate 41). The youngest of the flows is nearly 7 km long and averages 100 m in thickness. It was emitted in 1914 and still has numerous fumaroles and hot springs associated with it (Blake and Bleeker 1970).

East of the Cape Hoskins volcanoes is a group which includes the two most impressive volcanoes of New Britain, Bamus and
Ulawun, also known as South Son and Father respectively. Both rise gently from the coast to a towering 2000 m, thus forming the dominant landmarks of the area. Other volcanic centres are the Sulu Range and Lake Hargy area to the south-east of Bamus and Ulawun, and Likuruanga volcano and Lolobau Island to the north-east.

Ulawun and Bamus are similar in size, height, shape and structure, Bamus being somewhat older judging from the degree of dissection. They are both beautiful examples of highly symmetrical, conical central vent strato-volcanoes. In detail the symmetry of both volcanoes is broken by escarpments and lineaments, the origins of which are uncertain (Johnson and others 1972). The slightly curved escarpment at the southern flank of Ulawun is probably the remnant of an old caldera, while the much straighter lineament running along the west flank of Bamus is more likely to be a simple fault.

Ulawun has erupted explosively several times this century, the earliest eruption recorded being in 1915. During the last decade activity has increased and ash falls, nuees ardentes and lava flows were emitted in 1970 (Johnson and others 1972).

Other notable volcanic features of the area are the Lake Hargy caldera and Lolobau Island. Lake Hargy is a caldera about 13 km across bounded by a wall between 100 and 400 m high. The caldera was formed by collapse of most of the central part of the Hargy volcano, which was probably of similar size to the Bamus and Ulawun volcanoes (Johnson and others 1972).
Lolobau Island also constitutes a truncated cone with a central caldera and several post-caldera cones rising from the caldera floor. One of these, Sili, is reported to have erupted in 1905 producing a lava flow about 2 km long.

The third concentration of volcanoes occupies the north-eastern extremity of the Gazelle Peninsula around the township of Rabaul. The main features are a large partly submerged caldera open to the sea on its south-eastern side and a series of post-caldera cones rising from the outer rim of the caldera or from the largely submerged caldera floor.

According to Heming (1974) the caldera was formed by collapse of one or two large volcanoes in two stages, the first occurring around 3500 B.P. and the second at about 1400 B.P. Each event must have been a catastrophe of major proportions and was accompanied by the eruption of pumice ash which spread over a wide area.

Two of the post-caldera volcanoes inside the caldera, Tavurvur and Vulcan, erupted in 1937, causing widespread damage and loss of life (Fisher 1957). Vulcan, which rose from the sea during the eruption, is almost entirely composed of pumice and ash and represents a perfect example of a scoria cone. In spite of its young age the cone has been relatively densely dissected by radial gullies. Ollier and Brown (1971) calculated that gullying and erosion had taken place at the extraordinary rate of 19 m per 1000 yr over the past 31 years. Most of this erosion probably took place in the first few years of the cone’s existence and decreased rapidly with the establishment of a vegetation cover.

Plate 42
Bagana volcano (left), the most active volcano of Papua New Guinea, and smaller inactive volcano to the right. Numerous recent lava flows are clearly discernible on the slopes of Bagana, which is emitting some steam.
The Northern Bismarck Sea Volcanic Arc

This includes numerous post-Miocene volcanoes on Bougainville Island, six volcanoes forming small islands off the north coast of New Ireland and about ten volcanoes in the Admiralty Islands.

The volcanoes on Bougainville occur in three main clusters (Blake 1968). In the north-east there are the extinct and deeply eroded Tore volcano, the complex dormant Balbi volcano and several extinct volcanoes, in the centre the extinct volcanoes of Numa Numa, Billy Mitchell and Rein, and the active Bagana volcano, and to the south-east the very extensive complexes of Takuan and Taroka volcanoes, the latter including the dormant Loloru volcano.

The only volcano on Bougainville active at present is Bagana volcano (Plate 42). It is the most active volcano in Papua New Guinea, having been in an almost continuous state of eruption since it was first sighted by Guppy in 1882 (Blake 1968). It is also an example of a larger volcano almost entirely formed of recent flows. The eruptions are partly explosive and associated with emissions of nuées ardentes, and partly consist of viscous slow-moving lavas (Blake 1968). These flows are composed almost entirely of angular blocks and rubble and exhibit a rock glacier-like flow pattern with a concave profile at the upper end and a convex profile at the front end where numerous transverse pressure ridges are present. These flow features are also present on many of the extinct volcanoes, particularly on Takuan volcano.

Balbi is an interesting complex strato-volcano consisting of a number of coalescing cones and a summit area that contains a variety of volcanic lava forms including ash cones, craters, domes, spines and solfatara fields (Fig. 30). Balbi last erupted between 1800 and 1850 in an explosive manner, ejecting nuées ardentes which covered a large area to the east of the crater. The extent of these deposits is still clearly visible on aerial photographs, as they carry a distinctly different forest type.

The volcanoes along New Ireland’s north coast are all densely dissected strato-volcanoes but solfataric activity has been reported...
from two of them, Lihir and Ambitle. The volcanoes in the Admiralty are situated to the south-east of Manus Island, the main island of the group. Lou Island consists of several coalescing cones, all extinct. South of the island a submarine volcano rose from the sea during an eruption in 1955. It formed six cones which partly coalesce, built up by the extrusion of lava.

DEPOSITIONAL LANDFORMS
Depositional landforms fall into two major groups, the fluvial depositional landforms and the littoral depositional landforms, according to the processes involved in their formation and the environment they build up. The fluvial depositional landforms include closely related and often transitional types of alluvial plains and fans while the littoral depositional landforms encompass three distinctly different types, the estuarine plains and deltas, the beach ridges and plains, and the coral reefs.

Most of these landforms are generally regarded as unexciting geomorphologically because of their great uniformity and vast expanse. In Papua New Guinea the study of these plain lands is certainly not encouraged by the often adverse environmental conditions, characterised by extreme humidity, high temperatures, and often constant flooding. In addition to this the field worker is constantly harassed by clouds of mosquitoes and other pests such as mites and the ever present leeches.

Access is in many areas even more difficult than in mountainous terrain and a general appreciation of the interrelationships of the different landform units is virtually impossible. Here again the availability of aerial photographs has been crucial to the improvement of our understanding of these landforms.

Active Fluvial Depositional Landforms
The usually high rate of denudation in the mountains is matched by equally active redistribution and eventual deposition of the material removed in the form of alluvial sediments in the mountain foreland, and on a smaller scale in intramontane and intermontane basins. The landforms built up by these alluvial sediments are generally referred to as alluvial plains and alluvial fans. The distinction is not a sharp one and essentially the alluvial fan is simply a special form of alluvial plain. Alluvial fans have been defined as depositional landforms whose surfaces approximate to a segment of a cone radiating downslope from the point where the stream leaves the mountain (Bull 1968). The fan shape is not always well-developed, especially where larger fans are concerned, because neighbouring fans hinder the spread of sediments. The term alluvial plain is much more general, indicating that the landform is a depositional plain formed of alluvial sediments. Alluvial plains usually slope more gently than fans, which leads some authors to distinguish the two by a critical angle, but this is artificial and would exclude very extensive fans that have very low angles. An important difference lies in the stream and flow characteristics of the rivers that build up these landforms. Fans are always built up by braiding streams while alluvial plains are formed predominantly by lateral accretion.
and overbank deposition by streams flowing in defined though migrating channels.

The dynamic centre of any alluvial plain is the flood-plain. The exact nature and behaviour of these flood-plains in Papua New Guinea has not been studied in any detail except in the case of the Angabunga River (Speight 1965a, b) and the Markham River and its tributaries (Holloway and others 1973). For this reason and

Figure 31
Development of the Angabunga River (after Speight 1965 (reprinted by permission Elsevier Scientific Publishing Co.).

- 1963 river course
- 1957 river course
- - - 1938 river course
- --- Prior channels
- --- 1938 coast line
- - - - Edge of high ground


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because the flow and channel characteristics of these rivers are representative of the whole spectrum of Papua New Guinea's rivers, they are described first and in more detail than others even though they are of relatively small extent. The following is largely based on the work of Speight (1965a, b) for the Angabunga River and Holloway and others (1973) and my own studies for the Markham River.

The Angabunga River. The catchment of the Angabunga River covers some 2500 km² and lies in the rugged Owen Stanley Ranges which rise to well over 3000 m near the watershed. The stream leaves the mountains at an altitude of about 50 m and flows across a swampy alluvial plain between 15 and 25 km wide, in a central meander belt that rises slightly above the marginal back swamps. The meanders are subject to rapid lateral migration. Aerial photographs taken in 1938, 1957, 1963 and 1972 allow closer observations of the changes in the river course (Fig. 31). In 1938 the river flowed largely in a southerly direction then turned sharply east and then south again to terminate in the Hall Sound Delta which it had built up. The total length of the river from the mountain front to the delta was 90 km.

In 1954 a crevasse developed about midway between the mountain front and Hall Sound Delta and initially the river overspilled and a great amount of its water flowed overland without a defined channel. This situation continued until about 1960 when the crevasse developed into a new channel (Plate 43) which took a direct course to the sea, shortening its length by some 30 km and doubling its gradient from 1:4000 (0.025 per cent) to about 1:2000 (0.05 per cent). In 1962 when Speight visited the area the old channel was permanently dry, indeed so dry that part of it was used by motor cars in preference to the nearby vehicular track. No change has occurred since except for minor variations in the meander pattern upstream from the point of crevassing.

Speight distinguishes two types of flood-plain with characteristic cross-channel sections (Fig. 32) and bankfull discharge. The upper

Plate 43
Angabunga River with old, now dry course (right) and new course which developed between 1954 and 1960. Photograph taken in 1973.
part of the river course upstream from the point of crevassing is characterised by a high bankfull discharge and highly unstable meanders with wide triangular cross-profiles and unstable point bars consisting of coarse gravel and sand. In contrast, the lower (now dry) course is bordered by distinct, continuous levees built up of finer deposits of the silt and clay fraction; it has a more rectangular cross-profile, and its bankfull discharge is only about half that of the lower course. The new lower course has not yet developed any distinctive flood-plain morphology but its cross-profile is similar to that of the old lower course.

Investigation of the meander sinuosity, defined as the ratio of talweg length to airline distance, shows that there is no relation between sinuosity and width-depth ratio nor is there any relationship to the predominant grain size of the sediments. In the 1938 course the sinuosity is virtually constant throughout.

The meander wavelengths defined by recurring directions of flow along the talweg were studied by Speight with the aid of spectral analysis. He showed that two or three wavelengths dominate the spectrum throughout but in the lower reaches the dominant wave-
lengths are somewhat suppressed and overlain by smaller oscillations, obviously because the higher levees impose restrictions on the development of meanders. Another important finding is that orderly downstream migration of meanders does not occur even though meander cutoffs are common. Speight's results are of significance as they provide a quantitative framework within which one can relate the other largely qualitative observations on alluvial plains in Papua New Guinea, and indeed as will be seen below the general pattern established on the Angabunga can be found in certain variations on most of Papua New Guinea's flood-plains.

The Markham River. The Markham River occupies the eastern part of the fault-bound Markham Ramu graben system which separates the central ranges to the south from the northern coastal ranges (Fig. 33). The Markham commands a catchment of some 12,000 km², part of which lies in the Saruwaged and Finisterre Ranges and part in the central ranges. The Markham River is quite unusual in the Papua New Guinea environment, being a braided stream all along its course. The main river leaves the mountains to the north at an altitude of about 450 m, flows in a more or less straight south-easterly direction diagonally across the graben and then runs along the foot of the central ranges in an easterly direction to empty into the sea south of Lae. Its course is mainly straight but some incipient meanders of very long wavelength are developed. The main channel is between 0.5 and 2 km wide and is characterised by an anastamosing network of interconnecting secondary channels separated by sand, silt and gravel bars (Plate 44). The plain course of the Markham is about 140 km long and its overall gradient is of the order of 1:330 (0.3 per cent), which is considerably higher than for any other plain stream of comparable discharge and catchment area (Table 1). Its gradient is not uniform and varies from between 1:870 (0.12 per cent) in the lower section to 1:170 (0.6 per cent) in the central and steepest sections. The channel

Figure 33
Markham Valley and Huon Peninsula, with areas of braided streams and fans

- Active fans and braided stream channels
- Fans, not active
- Alluvial plains
- Raised coral terraces
- Mountain ranges
profile is shallow and wide and highly irregular in detail (Fig. 32c), contrasting sharply with channel profiles from meandering or levee-bound channels (Fig. 32a, b). The channel sediments of the Markham are very variable and include gravel, sand, silt and clay. There is of course a general decrease in coarse-textured sediments downstream but this is greatly complicated by coarse sediment supply from some major tributaries, in particular the Leron River.

The main northern tributaries of the Markham have similarly braided stream courses but have higher gradients and consequently transport coarser bedload. There are considerable variations depending on the character of the sediments over which the streams flow.

Holloway and others (1973) distinguish three types of tributary stream: fan streams, midfan streams and piedmont streams.

Fan streams are the major tributaries which mostly emerge from the mountains in single channels and flow across the fans they have built up, mainly in several distributary channels, to join the Markham (Plates 44, 45). Some of the distributary channels do not reach the Markham as a significant proportion of the water is lost by seepage into the ground water body. Some of this ground water re-emerges downfan to feed a midfan stream. Some of these midfan streams go underground again and eventually reappear on the surface to form swamps.

The third type, termed piedmont stream, is a small stream with a very high gradient of some 5°–6° emerging from the foothills to

<table>
<thead>
<tr>
<th>River</th>
<th>Catchment area km²</th>
<th>Discharge cumec</th>
<th>Average gradient Lowland course</th>
<th>Mountain course</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lakekamu</td>
<td>5300</td>
<td>195</td>
<td>1:2000</td>
<td>1:20</td>
</tr>
<tr>
<td>Markham</td>
<td>12600</td>
<td>385</td>
<td>1:330</td>
<td>1:30</td>
</tr>
<tr>
<td>Kikori</td>
<td>16900</td>
<td>1200</td>
<td>1:1700</td>
<td>1:120</td>
</tr>
<tr>
<td>Ramu</td>
<td>18500</td>
<td>990</td>
<td>1:4000</td>
<td>1:50</td>
</tr>
<tr>
<td>Fly</td>
<td>69900</td>
<td>5670</td>
<td>1:40000</td>
<td>1:70</td>
</tr>
<tr>
<td>Sepik</td>
<td>77700</td>
<td>5000</td>
<td>1:20000</td>
<td>1:50</td>
</tr>
</tbody>
</table>

Plate 44
Braided channel of the Markham River (foreground) and active fan of the Maniang River (background right). Terraces of Umia River, the main tributary of the Markham, in the left background. Photograph taken in 1975.
form small but distinct fans (Plate 45). These piedmont streams usually go underground near the front end of their respective fans. The streams are basically not different from fan streams except for their size and provide excellent study objects for active fan development.

Most of the fan streams are entrenched into an older fan surface by up to 40 m (Plate 45). The entrenched fan surface gradually converges with the flood-plain of the fan stream to form a continuous plain (Plate 44). The Maniang River is not incised but emerges from the mountains in numerous braiding channels, the flow of which goes partly underground and partly joins in a single channel, and flowed into the Markham River until about 1958. None of the channels is incised into the fan more than about a metre. The changes since 1958 have been investigated by Laurenson (1972) (quoted in Holloway and others 1973). In 1959 the channel that collected most of the surface flow was abandoned and most of the flow moved overland in an easterly direction in a sheet-like flow pattern. The

Plate 45
Braided channel of the Leron River incised into the main surface. One erosional terrace in foreground and uplifted terraces in the background where river emerges from mountains. Numerous small active fans along mountain front.

Plate 46
Braided channel of the Maniang River which in 1975 had moved north-east towards the foot of the mountains burying large areas of native garden land and partly destroying a village. The two boys stand on a former garden plot which is now covered by about 0.5 m of sediment.
water later returned to the main channel but left it again in 1966 when it broke out to the west of the main channel. Subsequently it returned to the main channel but in late 1971 the river again changed its course to flow eastwards in an overland flood about 2 km wide. Only a negligible amount of water remained in the main channel. In 1974 when the author visited the area the flow had moved further north-east towards the foot of the mountains, destroying a native settlement and burying large areas of garden land (Plates 44, 46). Partly-buried tree trunks and house stilts give evidence of accumulation rates of up to 0.5 m during this event. Similar changes took place in the lower Leron River. Here the river had been shifting in an easterly direction for many years but in 1974 it suddenly shifted westwards, destroying some 2000 ha of grazing land. Raising of the stream bed here also was of the order of 0.5 m (J. Robinson pers. comm.).

The shifts of flow appear to occur largely in response to the high rate of sedimentation within the river system, which leads to continuous raising of the actual stream bed and eventual shifting of the stream to slightly lower-lying adjacent areas.

On the Erap River, another tributary of the Markham, Knight (1975) observed a basically similar pattern but pointed out that in addition to the general raising of the channel localised constriction of flow by newly forming bars is important since it causes a local rise of the water level with consequent overspill and crevassing of the banks.

These observations on the actively changing rivers and the associated fan formation are of significance as they allow some insight into the processes that have been going on in this area in the past, and which have led to complete filling of the graben area with these sediments.

The total amount of material accumulated through time is truly impressive. Gravity surveys revealed that in the Leron and Erap area the thickness of the sediment amounts to up to 1000 m in the southern part of the graben, while at the northern margin the maximum depths are in the range of 300 to 700 m (Pettifer 1974). It is also important to note that the processes are active under present climatic conditions and that there is thus no need to infer past climatic changes to explain the extensive fan formation.

The Fly and Strickland Rivers. The Fly and Strickland Rivers drain most of the south-western part of Papua New Guinea. Their catchments cover some 70,000 km² and are second only to the Sepik River catchment with which they share the western part of the main watershed. The rivers flow southerly from the central ranges incised into an extensive relict alluvial plain, and join at about half-way between mountain front and coast to form the lower Fly, which then turns in a south-easterly direction terminating in the Gulf of Papua. The active flood-plains of the rivers are restricted to a relatively narrow belt between 10 and 15 km wide on either side of the meandering channels. Their features include scroll complexes, oxbow lakes and swamps and back swamps and lakes (Blake and Ollier 1971). For most of their courses the rivers meander vigorously within this flood-plain belt. The general pattern observed on the Angabonga is
clearly visible. In upper reaches the meanders are highly unstable, consisting of numerous point bars and scrolls largely formed of coarse material. The central meander belt is flanked by swampy back plains. As one moves downstream the meander wavelength does not seem to change but the meander channels become more confined. Point bars become less frequent, being replaced by low levees consisting of very fine material. Meander migration caused by overland flow during floods becomes less frequent and is more and more replaced by the simple cutting off of meanders by intersection of meander limbs. The flood-plain eventually changes to a true levee plain or covered plain in the lower reaches where meandering ceases completely and the river flows confined between levees (Plate 47).

In the uppermost reaches of the plain courses of the Fly and particularly the Strickland there is a tendency for the flood-plains to develop braiding courses. This is consistent with the results of Leopold and others (1964), who point out that there is a close relationship between braiding and meandering and that braiding occurs at a steeper slope than meandering under a given discharge. Other factors involved are amount and calibre of bedload. A large amount and high calibre of bedload favour the development of braids (Tricart and Vogt 1967).

Because of their sizes and the fact that both rivers are incised...
into a relict alluvial plain an additional feature not present on the Angabunga occurs: the blocked or drowned valley swamp or lake. The occurrence of these landforms is a direct result of the dissection of the relict alluvial plain. During the period or periods of low sea level the main rivers incised below their present level. Incision, although nowhere very great because of the very gentle offshore slope of the Gulf of Papua, was greatest in the lower reaches, decreasing gradually upstream. The tributaries also adjusted to their new base level at their junction with the main rivers. With the post-glacial transgression the main rivers aggraded rather quickly because of the great amount of sediment supplied by their mountainous catchment. The smaller tributaries rising from within the relict plain did not receive enough sediment from their low-lying catchments to keep pace with the filling up of the main channels and therefore became drowned and possibly also blocked near their entrance into the main river (Blake and Ollier 1971). The two largest blocked valleys are the Aramia River valley and Lake Murray. Lake Murray is permanently under water while the Aramia valley is only seasonally inundated. In the dry season the water is confined to a levee-bound course but during the wet season extensive flooding occurs, extending into the numerous tributaries (Plate 48).

Numerous smaller blocked valley swamps and lakes are developed on the lower and middle course of the Fly River. Along the Strickland blocked valleys are conspicuously absent except for Lake Murray at its lower end. This is because the Strickland has a much steeper gradient than the Fly and consequently the lowering of the base level did not cause any significant change in the middle and upper
reaches. Similarly there are no blocked valleys along the upper reaches of the Fly.

The Sepik River. With a catchment of 78,000 km² and a length of 1100 km the Sepik River is the largest river system in Papua New Guinea. Unlike the Fly and Strickland which emerge from three main mountain streams and flow more or less directly to the sea without being joined by any major tributaries, the Sepik flows in an intermontane basin parallel to the mountain ranges and is joined by major tributaries commanding extensive catchments all along its course. The geological situation is also distinctly different; while the Fly and Strickland traverse a basically stable shelf underlain by continental crust the Sepik flows in a mobile basin underlain at depth by highly deformed rocks. The flood-plain of the Sepik and its tributaries reaches a width of up to 70 km. Most of this consists of back swamps (Plate 49) and the actual meander belt only amounts to between 5 and 10 km.

In spite of these gross differences between the Sepik and Fly Rivers the basic flow and meander characteristics of the two river systems are similar. The Sepik leaves the central ranges in the far west corner of Papua New Guinea with a strongly braiding channel flowing in a north-westerly and then a northerly direction. As reported by Behrmann (1917) the stream here is very fast-flowing and the banks consist of coarse gravel and sand. Approximately where the Sepik turns sharply to the east the braiding pattern changes into a meandering course characterised by series of point bars and scrolls formed of sand and gravel. The rate of change of meanders is extreme. Thanks to an excellent map of the Sepik River produced by the German Sepik Expedition the changes within an approximately 50-year period can be studied. In the upper and middle reaches the river has dramatically altered its course even though the meander belt as such has not shifted (Fig. 34). In some sections the change has been so great that it is difficult to trace the 1913 course. The
shift of meanders has been particularly great above and below the entrances of major tributaries such as the Leonhard Schulze River, the April River and the South River (Karawari River) (Plate 50).

Plate 50
Sepik River, with meandering channel, oxbows, scroll complexes and backswamps and lakes. Pioneering vegetation accentuates the scroll pattern with trees and shrubs covering the highest parts of the bars and swamp grasses the lower lying areas. The back swamp is covered by floating grass.
The rate of change in the meander pattern decreases downstream from the South River while the meander wavelength seems to increase. The last changes occur at the entrances of the Yuat and Keram Rivers but these are not as dramatic as upstream. From the Keram River mouth downstream the river has changed little since 1913.

Unlike the Fly and Angabunga Rivers the Sepik does not flow in a levee-bound channel except for short distances. Along straight stretches some low discontinuous levees do occur, but the dominant channel feature remains the point bar.

Along most of its plain course the Sepik therefore forms a true meander plain and as such has a more active and immature floodplain than the lower course of the Fly.

Major changes in the entire course of a channel like those on the lower Angabunga have not been observed on the Sepik proper, but on some of its southern tributaries there were significant changes in the not too distant past, as many traces of inactive meander patterns can be seen on the aerial photographs. The most recent

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Figure 35
Yuat River area, former and present river courses

- Present river course, active flow
- Former river courses, no active flow
- Mountain front

50 km
Schwarz (1972) has pointed out that there is growing evidence for lowering of sea level by nearly twice that amount of the changes has been that of the Yuat River which at one stage entered the Sepik some 25 km upstream from its present mouth. Its former courses are still clearly visible on the aerial photographs (Fig. 35). The changes happened before 1913 because Behrmann’s map shows the Yuat River mouth at its present location. Judging from the aerial photographs the change in course of the Yuat is largely due to the very active aggradation along the old Yuat; this built up the meander belt well above the surrounding swampland, which eventually led to overspill and crevassing similar to that of the Angabunga. Whether or not the place of crevassing coincides with a change in channel cross-profile and transition from a meander type of flood-plain to a levee-bound type cannot be determined from the aerial photographs. A narrowing of the channel seems to occur along the older stream course but this could be due to gradual filling up of the older channels and overgrowing by vegetation.

The whole of the Sepik plain is therefore a very young, vigorously changing alluvial depositional system. The formation of these vast alluvial plains has generally been linked with down-faulting of the plain along roughly east-west trending faults (Krause 1965) or downwarping resulting from compression acting in a north-westerly direction (Dow and others 1972). While there can be no doubt that the Sepik basin represents a structural basin created by some tectonic force, the present configuration of the plain, its very irregular outline and its often deep penetration upstream into tributary valleys along the southern mountain front (Plate 16) cannot be satisfactorily explained by subsidence alone. Furthermore it seems unlikely that the aggradation of the rivers, especially of the southern tributaries that have large and high catchments, would not have kept pace with the subsidence, the maximum rate of which would not have exceeded the rates of other tectonic movements like those along the Huon coast for which more precise data are available. The nearly complete drowning of the Sepik plain requires a relatively sudden and uniform rise in base level and this was most probably caused by the post-glacial sea level rise.

The bathymetric chart (Fig. 36) of the mouth of the Sepik River shows a very steep drop of about 1:10 from the delta to about 700 m below sea level. This means that any change in sea level must have had profound effects on the gradient of the Sepik. At a lower sea level the Sepik and its major tributaries would have had greatly increased gradients, probably with braiding flood-plains, and the whole plain would have been considerably lower than today. Assuming that in the past the Sepik River had a graded course as it has today, a lowering of the sea level by about 130 m, as is generally accepted for the last glacial period, would have led to a fourfold increase in gradient and to downcutting in the basin as high as the May River area* (Fig. 37). Above this the influence of the lower sea level would have decreased rapidly.

The post-glacial rise in sea level was a rapid event with mean annual rates of the order of 1-2 cm per year and peak rates of up to 5 cm per year (Fairbridge 1961; Galloway and Löffler 1972), and the aggradation of the rivers, even though increased by the reduction in gradient, would not have kept pace with this except

* Schwarz (1972) has pointed out that there is growing evidence for lowering of sea level by nearly twice that amount.
along their immediate flood-plain. This led to the gradual drowning of the entire plain. Sedimentation of inorganic material decreased with increasing distance from the flood-plains but here accumulation of organic material in the form of peat took place. This situation has prevailed until today when there is practically no inorganic deposition in the swamp areas (H.A. Haantjens pers. comm.). It is likely that this process of lowering of the basin and subsequent drowning and filling up took place repeatedly during the Pleistocene, but the stratigraphy of the basin is not known and this therefore cannot be substantiated.

Other Rivers. The lower Ramu, flowing in the same depression as the Sepik, is very similar to its neighbour although its course is less actively changing, with only five meander cutoffs since 1913.

The majority of the other alluvial plains along the south coast and north coast of Papua New Guinea are all essentially similar to the Angabunga type of plain. For most of their courses they represent meander plains with rapidly changing scroll patterns and meander cutoffs and the central meander belt rising slightly above the surrounding back plains. In the upper parts of the plain course where the stream leaves the mountain there is nearly always a tendency to develop braids while along the lower part of the course levees become more prominent, replacing point bars. But there is

Figure 36
Bathymetry off the north coast of Papua New Guinea (after unpubl. BMR record No. 1972/21) (re­printed by permission Bureau of Mineral Resources, Australia)

Figure 37
Present profile of the Sepik River and assumed profile during the Pleistocene low sea level stand
never a clear cut picture, and one has to be content with describing
the majority of alluvial plains in Papua New Guinea as combinations
of meander plains and covered plains or levee plains with the meander
plain probably forming the major parts of the stream course.

Relict Fluvial Depositional Landforms

Fluvial depositional landforms that are above flood level following
tectonic events, changes in base level or changes in the fluvial
regime, and are therefore no longer subject to active depositional
processes, are referred to as relict depositional landforms. They
include alluvial terraces, plains and fans. These landforms are in
varying stages of dissection and are thus transitional between de­
positional and denudational landforms, but their character as
depositional plains is still apparent in the form of undulating
plain land or intricately dissected plains or fans with even summit
levels and flat-topped plain or fan remnants.

Relict Alluvial Plains. Relict alluvial plains are extensive only in
the Fly-Strickland area where they form most of the Fly Platform.
A vast relict alluvial plain extends here from the coast to the central
ranges over a distance of nearly 400 km, covering some 10,000 km².

This relict alluvial plain is in various stages of dissection. South
of the lower Fly it is formed by a flat gently undulating plain with
a relief of usually less than 10 m, while north of the lower Fly it
becomes increasingly dissected by a dense drainage system and
consists of closely spaced ridges with small areas of plateau-like
remnants preserved in watershed areas (Plate 51). In the north-east
the alluvial plain merges with the volcano-alluvial fans of the Mt
Bosavi volcano. The transition from the dissected alluvial plain to
the largely undissected volcano-alluvial fan is, at its southern end,
locally marked by a low but distinct step while along its western
margin the transition is more gradual. Stratigraphically the volcano­
alluvial fan overlies the dissected alluvial plain.
The development of the alluvial plain is closely linked with the formation of the central ranges at its northern margin. Slope instability associated with the uplift of the central ranges provided large quantities of sediment that were deposited on the stable foreland. The main forms of deposition were probably bar deposition, lateral accretion and overland flow from meandering rivers as along the present flood-plain. Braiding was probably more widespread than it is today, because of the greater supply of sediments and consequently the higher load-discharge ratio. The exact courses of the major rivers are not known except that they appear to have left the central ranges at about the same points as today, because a deeply entrenched antecedent drainage system is developed. From here they traversed the extensive platform in braids and lower downstream in frequently shifting meanders, probably resembling very closely the present Sepik or Angabunga plains. The plain eventually became incised by the major rivers and their tributaries. The exact date of this event or events is not known. Blake (1971) claims that the incision took place less than 27,000 years ago and was due to the eustatic lowering of sea level during the last cold period. As discussed in Chapter 2, this date is unlikely to be representative for the plain as a whole and the upper beds of the plain appear to be considerably older. The incision into the plain was probably partly due to a lowering of sea level but as this happened repeatedly throughout the Pleistocene it is unlikely that the last eustatic lowering of sea level was solely responsible for the dissection of this vast area. In addition to the changes in sea level, changes in flow characteristics and particularly the load-discharge ratios of the rivers draining the mountains have to be considered. With the reduction of the seismic activity in the central ranges the sediment supply will have decreased and the resulting increase in competence of the rivers will also have favoured incision.

The lower course of the Fly is claimed to have been diverted by gentle warping along an east-west axis in a direction east of its former straight southerly course (Blake and Ollier 1970). Again the date of this event is unknown. The warping took place either concurrently with the incision or at some earlier stage.

Warping along the Oriomo uplift, however, does not entirely explain the direction of the lower Fly, since not only the lower Fly but also its northern neighbour the Aramia River and to a lesser degree the Wowoi River turn from a southerly to an easterly direction. Also it seems unlikely that a large river like the Fly would not have been able to maintain an antecedent course through the soft sediments. It is more probable that the easterly flow follows the predominant slope of the former depositional surface.

The striking difference in the dissection pattern between the southern and northern parts of the relict alluvial plain could be due to differences in the prevailing erosional regime, with surface wash and associated planation as the dominant forms of denudation in the savanna and monsoon forest, which have a strongly seasonal climate, and linear incision in the rain forest areas which have a permanently humid climate (for further discussion see Chapter 5).

Relict Alluvial Fans. Alluvial fans are common landforms in arid
and semi-arid environments where they have attracted much attention among geomorphologists and hydrologists in the last decade, especially in the USA (Beaty 1963; Bull 1968; Denny 1965). They are, however, not restricted to these dry zones but also occur in humid areas such as the European Alps (Fisher 1965), New Zealand (Carryer 1966) and arctic regions (Rapp 1957). In the humid tropical environment of Papua New Guinea fans are also common and occur in very much the same geomorphological setting as in other climatic zones, i.e. they are associated with tectonically active mountain ranges and their forelands. This can represent a graben system like the Markham-Ramu graben, an intramontane basin like the Wahgi Valley, or some other form of foreland serving as a sediment storage.

Series of variably dissected fans and associated terraces are situated in the Yodda-Kumusi Fault Trough. This fault trough extends from the Musa basin in the east to the Waria basin in the north-west over a distance of some 300 km. The fans are most extensive in the broad Musa basin but further west they form a narrow discontinuous strip of coalescing fans. Most of the fans slope north-east away from the Owen Stanleys at angles between 1° and 3°. The fans have been built up by two main processes, fluvial deposition and mudflow activity. The fluvial fans tend to have more regular smoothly sloping surfaces and consist of stratified deposits, largely gravel, sand and silt. The mudflow fans often have irregular surfaces due not only to their originally more irregular deposition but also to the frequent occurrence of slumping in these clay-rich deposits. The fans differ greatly in age and stage of dissection, but no detailed work on this aspect has been done except in the vicinity of Kokoda where Pain (1972a) investigated the inter-relationships of several fan surfaces with the aid of detailed soil studies. Pain differentiates between four surfaces. The upper and main surface consists of strongly weathered fanglomerates, the second surface has been cut into the upper surface and consists therefore of similarly weathered deposits at the base overlain by younger gravel and sand that is only slightly weathered. Both these surfaces are overlain by ash. The third and fourth surfaces are formed entirely of fresh unweathered boulders, gravel, sand and silt, and some reworked ash. The sequence obviously indicates a history of a major fan-building phase represented by the upper fan surface followed by phases of downcutting and alluvial capping, or filling in. The stratigraphic relationship is shown in Fig. 38. No exact dating is available but a tentative age of between 30,000 and 50,000 years B.P. is suggested by Pain, based on the thickness of the ash accumulation on the upper surface and comparison with

**Figure 38**
Stratigraphic relationship of fans in the vicinity of Kokoda (after Pain 1972a) (reprinted from Pacific Science by permission of The University Press of Hawaii)

- Weathered fan deposits
- Airfall volcanic ash
- Gravels and boulders with little or no weathering
- Sandy and silty alluvium
the rate of ash accumulation on the Hydrographers Range.

The fans in the Markham-Ramu graben occupy a similar position between two steeply rising mountain ranges but in contrast to the Yodda-Kumusi graben fans they slope south, and their main source of sediment supply is the Saruwaged and Finisterre Ranges to the north (Plates 44, 45). The fans cover the entire graben area, and their thickness amounts to an astonishing 1000 m in the eastern part (Pettifer 1974). The largest fan is the Leron fan with a radius of 20 km. The fan deposits are clearly fluvial in origin. They are stratified and consist of well-rounded gravel and boulders alternating with finer sediments. The coarse material is generally poorly sorted but there is a gradual downstream decrease in grain size. The deposits are unweathered in the Markham area but slightly weathered in the Ramu area, and the soil cover is very shallow and pedogenically very little developed, clearly indicating the very young age of these surfaces. Organic material collected from the lower part of the Erap fan gave a C14 age of 610 ± 150 years B.P.

The individual fans coalesce to form a largely continuous southerly to south-easterly sloping surface which is referred to as the main fan surface. This main fan surface is variably incised, the incision being greatest near the mountain front and decreasing gradually downstream until the main fan surface merges with the active channels. Some rivers have not incised at all but flow on the main surface, actively extending and building up their fans; this applies particularly to those of small and medium size (Plates 44, 45).

The amount and degree of incision increases on moving westward over the low watershed into the Ramu catchment. Here the fans are not only incised at their apexes by the fan streams but are also dissected by a system of subparallel gullies rising from the upper and middle parts of the fans. This is an indication of their greater age.

Two different types of terrace are developed, firstly those which occur below the main fan surface and secondly those which occur above the main surface.

The first type is best developed along the Umi River where three well-developed terraces are cut below the main fan surface and
run along the present river course for a considerable distance downstream into the Markham River (Plates 44, 52). These terraces seem to represent largely erosional phases of downcutting into the main fan surface. Similar terraces are to be found along some of the Ramu River tributaries, especially the Lanu and Surinam Rivers, but only one or two terraces are present. One such terrace is also developed along the eastern part of the Leron fan (Plate 45).

The second type of terrace occurs at the head of the Leron and Erap fans. In contrast with the Umi River terraces the Leron fan terraces are not only above the main surface but wedge out or are sharply truncated a very short distance from the mountain front (Plates 45, 53). They seem to represent either remnants of former fans truncated by the formation of the main fan surface or more likely are parts of the main surface that have been uplifted. The significant role of tectonic uplift is clearly demonstrated by the fact that the upper fan remnant is locally offset by faulting or slopes in an opposite sense to the present river gradient (Plate 53). Nothing is known about the age of these terraces or fan remnants except that they are all very young features and that there appears to be a westerly increase in age.

The active fan formation in the Markham Valley is clearly linked with the tectonic instability of the catchment area and the associated slope instability and resulting high sediment load of the rivers. The reasons for the fan entrenchment and terrace formation are more obscure, in particular since the situation varies from fan to fan. Causes of fan entrenchment and terrace formation are numerous and there has been much discussion of this (Blong 1975; Bull 1964, 1968; Beaty 1963). Important factors are changes in climate affecting rainfall intensity, vegetation cover and consequently stream discharge and sediment supply, changes in gradient or base level due to tectonic events or sea level changes, adjustments of fan gradient due to continuous downcutting in the mountains, channel diversions because of blocking of channels by mudflows and changes from mudflow activity to streamflow activity.

In the Markham Valley mudflows in the strict sense do not play

Plate 53
Terrace sequence at the apex of Leron fan seen from main surface (foreground). The terraces here only occur along the mountain front and are probably due to tectonic uplift. Note the difference in height of the upper terrace on either side of the small valley which probably marks a fault line. In the background steeply dipping Pliocene (?) fanglomerates form a densely dissected ridge and V valley landscape.
an important role, judging from the clearly fluviatile character of the sediments, and any causes of entrenchment related to mudflow activity can be discounted.

Climatic changes did of course take place, but were they significant enough to alter substantially the conditions of stream discharge and sediment supply? On present knowledge this is not the case; the repeated cooling and warming up during the Pleistocene in the lower and middle altitudes probably resulted merely in an up and down shift of the different rain forest belts rather than in any substantial alteration of the vegetation cover. Amount and intensity of rainfall could have varied but no evidence to support this is available.

An important argument against any climatic explanation is the fact that there is great variability from fan to fan, with entrenched and non-entrenched and terraced and non-terraced fans in close proximity to one another. This non-uniformity cannot be explained by climatic factors. Changes in base level did occur during the Pleistocene low sea level stand but because of its steep gradient little change would have occurred in the Markham River except at its lower part where some incision is to be expected. However, any trace of this has been obliterated by rapid aggradation during the subsequent sea level rise.

The concept that fan entrenchment is a self-regulating process of adjustment of discharge and load to the fan gradient and that incision would follow once a certain threshold of fan aggradation has been reached is attractive, since it would easily explain the great variety of situations in the Markham Valley. Unfortunately the concept is difficult to verify.

Finally, the tectonic events which are so obvious and well-documented have to be considered. Bull (1964) has shown that the rate of uplift of the mountain front relative to the rate of downcutting will control entrenchment near the mountain front. If uplift exceeds the rate of downcutting fan deposition will occur near the mountain front and no incision will take place; if on the other hand stream downcutting exceeds the rate of uplift entrenchment will occur at the mountain front and the loci of deposition will shift downfan. Both situations are present in the Markham, which would imply considerable variation in tectonic activity along the mountain front.

No definite answer can be given on the question of fan entrenchment and terracing in the Markham Valley but it is felt that the last two explanations are the most likely for this area.

The Sepik alluvial fans are situated at the northern margin of the extensive Sepik depression between the Sepik River flood-plain and the northern ranges. They occupy a distinctly different geotectonic position from either the Markham or Yodda Kumusi fans and their extent is not restricted by a narrow fault zone. The transition from fans to mountains is not determined by an abrupt fault scarp but is much more indefinite and irregular, and in fact it is locally difficult to delineate the boundary exactly. The spread of fan sediments here has been unhindered, except laterally by adjacent fans, and the fans are thus much more extensive and more gently sloping (maximum 1 per cent) than the fault trough fans, in spite of their
small and relatively low-lying catchment.

Fan sediments are predominantly of very fine grain, largely clay and silt, but locally sand and quartz gravel are interbedded. Except for the quartz the sediments are strongly and deeply weathered, indicating a very advanced age. Soils are similarly weathered and have maturely developed profiles.

According to Reiner and Robbins (1964) and Reiner and Mabbutt (1968) there are two generations of fans, an older dissected surface and a younger largely undissected surface (Plates 54, 55). The older surface is represented by accordant often flat-topped ridges near the mountain front. This surface has been truncated by the younger much more extensive surface which extends locally as far down as the Sepik River flood-plain (Fig. 39). The fan surface is most extensive in the east where it attains up to 40 km in width, but it becomes narrower westward forming an apron some 10 to 20 km wide along the foot of the northern ranges.

The formation of these fans has been attributed to phases of uplift and associated erosion in the northern ranges (Reiner and

Plate 54
Gently undulating fan surface in the Sepik area. Owing to man’s agricultural activity and burning practice rain forest has been replaced by grassland except along drainage lines.

Plate 55
Dissected older fan surface in the Sepik. The former fan surface is represented by accordant ridges. Slumping and associated mudflow activity are very prominent in these poorly consolidated sediments.

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Robbins 1964). These authors assume that the oldest surface was formed during the Pliocene or early Pleistocene following the emergence of the coastal ranges. The fans are thought to have been deposited on a marine bight which occupied much of the Sepik depression at the time. The second and main fan surface was formed during a second phase of uplift which led to the dissection and truncation of the older fan surface and redistribution of sediments. Reiner and Robbins consider that deposition resulted from a relative rise of sea level following a low stand. This second fan surface eventually became incised and the present configuration emerged. Incision is attributed to uptilting of the fans in the north, possibly associated with tectonic movements of the northern ranges and lowering of sea level. Little can be added to this generalised picture of development of the fans, but the exact role of relative sea level changes in fan formation is not clear. There have been a large number of sea level changes in the last 200,000 years alone (Chappell 1974a) and it is difficult to see how these could be linked with the two surfaces. Furthermore the position of the fans in relation to older sea levels is not known. It seems more likely that fan formation took place largely independently of sea level changes as a result of large sediment supply from the rising mountains to the north.

While the fans discussed so far are all situated relatively close to sea level and occupy major graben systems or other depressions stretching laterally over hundreds of kilometres, the fans in the

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**Figure 39**
Sepik fan surfaces (after Reiner and Robbins 1974) (reprinted from the Geographical Review, by permission of the American Geographical Society)

**Figure 40**
Intramontane basins in the highlands

- Valleys and basins filled with volcanic material, largely lahars and ash
- Valleys and basins blocked by volcanic events and filled with fluvial, lacustrine, mudflow and volcanic material
- Structural basins filled with alluvial and organic material
- Basins of unknown origin, possibly erosional or tectonic
- Basins blocked by tectonic events

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highlands are usually situated well above 1000 m and are of local occurrence, much smaller in extent and irregularly scattered. The highland fans always occur in association with intramontane basins which act as local sediment storages (Fig. 40). These intramontane basins are of diverse origin. The majority have been caused by disruption of drainage by volcanic activity, such as the Wahgi, Kaugel, Lai and Tari intramontane basins. Others like the Kainantu and Goroka basins seem to have been formed by tectonic events leading to temporary ponding of rivers and subsequent filling up of the basins with lacustrine sediments. Others again, like the Neon basin and Myola lakes in the Owen Stanley Ranges, are associated with relict surfaces and may have formed close to sea level before the main uplift. Continued backward tilting against the gradient of the rivers may also have contributed to preservation of these relatively low gradient basins as proposed by Jennings (1963) and Guilcher (1970) for the Ka and Lai valleys respectively.

In spite of the differences in development of the basins the fans that form much of the fill are of similar origin. There are two main types of fan: fluvial fans which occupy most of the larger basins and mudflow fans which are more common in small basins.

The Wahgi fans are the most extensively developed fans in the highlands. They have been shed from the Kubor and Bismarck Ranges into a centrally located basin largely formed by ponding of the westerly flowing Wahgi River by the Mt Hagen volcano. It is, however, unlikely that the basin was produced entirely by ponding, as it is situated in a structural low between the Kubor anticline and the Bismarck fault zone. The original westerly orientation of the flow was first postulated by Haantjens (1970), and field studies of the present author support this. The main lines of evidence for this are significant westward decrease in relative height of the fans above the present river, coupled with a gradual westward decrease in absolute height, indicating that the central river draining the fan area sloped westward, not eastward like the present Wahgi River. Similarly the degree of fan entrenchment and fan dissection as well as the depth of incision decrease westward. The same westerly slope has been noted for the bedrock underlying the fanglomerates (Haantjens 1970).

The shape of the present Wahgi Valley, which is widest in the upper parts near Mt Hagen, where it is swampy, and becomes narrower downstream, is also indicative of ponding at the Mt Hagen end of the valley. Whether the former westerly flowing Wahgi joined the Bayer River to drain northwards into the Sepik drainage system or the Nebilyer to feed the Purari catchment cannot be answered at this stage.

There are three surfaces, the main fan surface which covers most of the area and two younger terrace surfaces occurring close to the Wahgi River and along some larger tributaries (Plate 56). The main fan surface represents a gently undulating surface sloping on average about 1° towards the centre of the basin. It consists of strongly weathered sediments of fluvial origin, which near the mountain front can reach 50 m in thickness.

The two younger surfaces occur as discontinuous relatively
narrow river terraces on either side of the Wahgi River and some tributaries. Both terraces slope eastwards in the same direction as the present river. They consist of well-bedded fluvial gravel. The weathering status of the deposits is distinctly different, indicating a considerable time difference. The upper terrace consists of strongly weathered gravel but the degree of weathering is clearly less than that observed at the main fan surface. Much of the gravel is still intact though it can easily be broken with a hammer. The lower terrace in contrast, consists of only slightly weathered or unweathered gravel. The terraces seem to have formed during periods of downcutting and alluvial capping that followed the main fan-building period. The first erosional period may coincide with the drainage reversal caused by gradual capturing of the Wahgi by the Chimbu River.

Unfortunately no dating material has been found in any of these deposits and the likelihood of preservation of datable organic material is slight. Also the fans and upper terraces are certainly older than the C^{14} dating range. Ash stratigraphy as established by Pain and Blong (1976) could be more promising; however, the ash layers occurring on the fans are strongly weathered and undifferentiated. It is therefore difficult to speculate on the possible age of these landforms. The main fan surface must be of considerable age considering the advanced stage of weathering of its deposits. If it is accepted that Mt Hagen volcano caused the blocking of the formerly westerly directed flow of the Wahgi and the subsequent partial filling up of the basin then the age of the volcano will give a minimum date for the fans. However, only the youngest lava flows on Mt Hagen have been dated, at 200,000 years B.P. (Page and Johnson 1974), and this would only be a lower age limit.
for the fans. Both the volcano and the fans are considered to be of much greater age than this.

Extensive fans also occupy the northern part of the Goroka basin. The Goroka basin (Plate 57) is probably of tectonic origin. Uplift along a major faultline along its western end seems to have caused temporary blockage of the drainage of the Asaro, Bena Bena and Dunantina River systems and led to the formation of an extensive lake. These lake deposits are finely laminated and silty and sandy, and are well over 100 m thick in the southern part of the basin where they have been deeply incised by the Bena Bena River and form a ridge and V valley landscape characterised by steep slopes with numerous terrace-like benches. These benches are produced by differential erosion of the lake beds, which vary in grain size and degree of consolidation. They are clearly not river terraces as they are virtually horizontal and concordant with the lake beds, there is no river gravel associated with them, and finally they also occur in young creek and gully heads where fluvial terraces could not be present. This also excludes the possibility that they are old lake terraces.

In the northern part of the basin, which is drained by the upper Asaro and its tributaries, the lake sediments are overlain unconformably by coarse gravel forming an extensive series of fans. The thickness of the gravel increases northward from 3–5 m in the south to over 40 m in the north. There are three surfaces, the upper fan surface and two terraces (Plate 57). The upper fan surface is the most extensive, forming a flat to gently undulating south-westerly sloping plain. It is formed of strongly weathered fluvial gravel very similar in degree of weathering to that of the Wahgi fans. This upper fan surface has been entrenched in two major stages leading
to the formation of two terraces which are well-developed along the
main tributaries. The erosional nature of the terraces is evident in
several well-exposed sections, showing the very shallow capping
of the lake sediments with a veneer of gravel (Plate 58).

The reason for the quite dramatic change from the lake environ­
ment to the fan formation is not known but it is reasonable to assume
that the change was caused by a relatively sudden draining of the
lake and influx of coarse gravel from the catchment area. This
was probably triggered off by a phase of tectonic instability which
led to an oversupply of material into the rivers. The sudden draining
of the lake must have been due to a rapid lowering of its overflow.
The succession of erosional terraces similarly seems to be caused by
consequent changes in base level.

The Kaugel intramontane basins and fans at the foot of the
towering Mt Giluwe volcano (Plate 38) are the best investigated
in the highlands, thanks to the work of Pain (1973), who undertook
a detailed study of the basins and their sediments and geomorpholo­
gical history. The Kaugel intramontane basins are a series of south­
easterly aligned basins separated by narrow gorges cut through
lavas, sedimentary bedrock and Quaternary sediments. The basins
essentially consist of extensive central basin floors which are strik­
ingly flat and are entrenched by the Kaugel River and its tributaries.
Entrenchment of the Kaugel is 40-50 m. Marginal fans sloping
between 5° and 8° have been shed into the basin, mainly from Mt
Giluwe, and occupy a belt some 2 km wide between the flattish
basin floor and the volcanic footslope of Mt Giluwe.

According to Pain (1973) the Kaugel basins were formed as a
result of the general drainage disruption that followed the eruption

Plate 58
Goroka basin, middle terrace of
Zozoki River; finely laminated lake
sediments are overlain by coarse
fluvial gravel
of Mt Giluwe. Radiometric dates recently obtained from Mt Giluwe lavas indicate that the minimum date for this event would be in the vicinity of 220,000 years B.P. The basins were initially occupied by lakes that became filled up with terrestrial inorganic and organic material. Pain considers that the condition of deposition ranged from quiet lake environments to rapid infilling by subaqueous mudflows. This period of lake infilling ended with a major ash fall event, the deposition of the Tomba Tephra and Bune Tephra, with sources from Mt Hagen and southern Mt Giluwe respectively. Several C¹⁴ datings suggest that this event took place at about 30,000–31,000 years B.P.* Some smaller fans are also present on the eastern side of the basin, indicating that fan formation was not restricted to Mt Giluwe.

The widespread ash fall was followed by the deposition of the extensive fans that now form the margins of the basin. These deposits, termed Kaugel Diamictons by Pain, were interpreted as of fluvioglacial origin (Blake and Löffler 1971), but Pain considers a mudflow origin more likely. The deposits are characterised by a lack of bedding, a widespread distribution and the lack of any disconformities within the sequence even over relatively large distances. Pain therefore suggests that these deposits were laid down simultaneously during a major catastrophic event, probably following a severe earthquake.

After this event only minor changes to the landforms in the Kaugel valley took place. These were entrenchment of the Kaugel River and its tributaries into the Kaugel Diamictons and the underlying lake beds. During this incision a number of terraces were formed which could represent a pause in the process of incision or, as Pain points out, were more probably formed as slipoff terraces as the river swung from side to side within the trench.

Some minor fan-building also took place at the very margins of the basin. In contrast with the deposition of the Kaugel Diamictons, however, this was not a single event but seems to have been a series of minor events resulting from localised landslides and slumps on the surrounding slopes. C¹⁴ dating indicates that these events took place sporadically during the last 10,000 years.

Conclusions on Fan Development in Papua New Guinea

Active fan-building clearly takes place under the present humid tropical rain forest conditions. In Papua New Guinea widespread fan-building is at present restricted to the Markham-Ramu fault zone south of the actively rising Saruwaged and Finisterre Ranges and to the coastal areas between the Huon Peninsula and Astrolabe Bay to the north of these ranges (Fig. 33). In all other areas fan formation is either very localised and on a small scale as in the Gumants basin near Mt Hagen (R. Blong unpubl.) or has ceased, and the fan surfaces have been strongly weathered and dissected in varying degrees by streams and rivers which mostly lack braided flood-plains characteristic of active fan-building. Active fan development on a larger scale is thus restricted to the tectonically most active part of Papua New Guinea, an indication of the close interrelationship between fan formation and tectonic activity and asso-

* According to more recent investigations of these ash deposits by R. Blong (pers. comm) a date of at least 50,000 years appears more likely.
ciated slope instability.

Deposition on the alluvial fans seems to be largely due to the sudden spread of flow and loss of velocity at the points where the rivers leave the confined channel of the mountain valley (Plates 45, 46). The transition from mountain stream to fan stream is gradual and not marked by a sudden change of gradient. Loss of flow through the highly permeable fanglomerates, often referred to as the sieve effect, also seems to be of importance.

The relict fans usually consist of a series of coalescing fans forming a continuous main surface, and two or three terraces, apparently representing a major fan-building phase, and two or three periods of downcutting and alluvial capping. In the Sepik two fan building phases have been recognised.

The observations in the Markham-Ramu graben show that such a fan-building phase is not to be regarded as a single event. Even though topographically there may be a continuous surface connecting the coalescing fans, fan activity varies from fan to fan. Entrenchment and active fan formation can even take place simultaneously on the same fan. Stratigraphic correlation of the fan surface will therefore be difficult. Entrenchment is probably caused as a result of adjustment of discharge and load to the rising fan or because the rate of downcutting of the mountain front exceeds the rate of uplift.

The fan-building phases must be regarded as probably extensive periods of fan-building with the loci of activity constantly shifting across the fan and from one fan to another.

In the case of mudflow fans Pain (1973) claims that they can be formed simultaneously by a single earthquake event because of the complete lack of unconformities within the mudflow deposits.

The relative ages of the fans in Papua New Guinea is not known and since it is difficult even to establish the relative ages of fans that form a continuous fan apron, the age relationships between fans from different areas will pose a major problem. Detailed investigation of tephras as attempted by Pain (1973) and Pain and Blong (1976) could provide a stratigraphic basis for correlations if the identification of ash layers could be made with greater precision.

**Littoral Depositional Landforms**

The coastal landforms of the tropics are among their most characteristic. The idyllic 'tropical paradise' with its coral-fringed coast, its clear blue-green water and its white sandy beaches shaded by coconut palms is probably the best known example. Less attractive scenically but also very typical and scientifically interesting are deltas, estuarine plains and tidal flats covered with dense mangrove vegetation. These mangrove-covered plains generally form the seaward extension of the alluvial plains and are thus closely related to them even though their environment is distinctly different. Like the alluvial plains the deltas, estuarine plains and tidal flats owe their existence to the great abundance of suspended material brought down by the rivers. In Papua New Guinea the proximity of the mountain ranges, the high precipitation and subsequent intensive erosion, particularly of fine material, are highly favourable for the development of these landforms.
Deltas, Estuarine Plains and Tidal Flats. Deltas, estuarine plains and tidal flats occur extensively only along the south coast of Papua New Guinea, largely because here along a relatively stable coastline with a gentle offshore slope the conditions for uninterrupted deposition and seaward extension of the land are more favourable than along the actively rising north coast where the sea bed falls off very steeply (Fig. 41, Plate 59). The post-glacial rise in sea level also affected the two coasts differently. Along the south coast it led
to drowning of river inlets, reduction of stream velocities in the lower reaches and formation of embayments which provided ideal loci for deposition. Because of its steepness and active uplift most of the north coast was largely unaffected by this sea level rise except for the Sepik plain and the Cape Vogel basin.

The process of deposition in this littoral environment is relatively complicated as it is a zone of interaction of fluvial and tidal processes and of relatively strong wave-induced longshore currents. Sediments are therefore not as uniform as one might expect and range in size from fine sand to clay. Sand is normally present in higher concentrations in a narrow band in the outer areas of the deltas where wave scour and tidal currents are strongest, leading to a relative concentration of coarser material by removal of the fines. These sandy deposits usually form beach ridges. Once the sediments accumulate to about 1–1.5 m below mean sea level they are invaded by mangrove vegetation. This colonisation by mangrove is very important as a means of trapping further sediments, largely due to more rapid sedimentation caused by the increase in roughness of the surface and consequent reduction in velocity of the tidal current. Mangroves are also important in compacting these fine sediments and protecting them from further transport (Tricart 1972).

Mound-building crabs, e.g. *Thalassina anomala*, are also locally important in hastening the rate of accretion. These animals rework the tidal sediments and build mounds 2–5 m across (Plate 60) rising to 1–1.5 m above high water mark. They are most abundant near tidal channels where their distribution is locally nearly continuous, but they decrease rapidly in numbers inland from the channels.

**Plate 60**
Crab mounds rising above high water level. The mounds here are used by the people to grow food-stuffs.
(Ruxton 1967), and inland from mean sea level to high water mark. These mounds are of great importance to the local population as they are used to grow foodstuffs accessory to sago. Locally these crab mounds are enlarged to form islands up to 20 m² in area separated by narrow tidal channels (McAlpine 1969).

Tricart (1972), discussing the role of crabs in mangroves, suggests that they are in fact facilitating erosion by 'ploughing' through the deposits and keeping them friable and thus causing the removal of material. This does not apply to the crab mounds observed in Papua New Guinea, which are raised above high water level, and where room for further deposition is made by removal of material from intermound areas (Ruxton 1969). The gradual but irregular raising of the surface by the crabs will also increase its roughness and consequently reduce the velocity of tidal currents. Redistribution of material will certainly take place but the net effect of crab activity is accretion.

The Purari Delta (Plate 59) is undoubtedly the most typical mangrove delta in Papua New Guinea, showing great similarity in appearance, though not in extent, to the Niger Delta in Africa and the Amazon Delta in South America. There is a dense network of interconnected channels with roughly rectangular bends separating the delta area into a maze of islands. The channels are funnel-shaped at their mouth due to the seaward increase in the velocity and scour of tidal currents. The large mangrove-covered islands have slightly raised rims and central depressions due to the combined effect of the higher rate of deposition of suspended load near the channels and crab mound-building activity. Locally the picture is complicated by very low centrally located interchannel rises (Plate 59), which appear to be slightly above tidal influence and subject to freshwater inundation originating almost entirely from rainfall. These rises have a predominantly fresh-water vegetation including sago which provides the staple food for the local population, whose settlements, built on stilts along the tidal channels, are often located nearby (Plate 59). These central interchannel rises are interpreted as former bars or low islands from which the mangrove extended laterally. They are not now part of the dynamic mangrove system.

The raised rims of the mangrove areas are not to be confused with true levees. The overbank flooding in the mangrove areas is caused by the relatively slow and quiet advance and retreat of the tide and not by highly turbid flood currents as in fluvial environments (Mabbutt 1965). For this reason too the drainage system is one of interconnected channels which are not or are very rarely sealed off from one another. Submerged levees are also absent.

Near the entrances of larger rivers with great discharges of fresh water, mangrove is often replaced by Nipa palm vegetation which performs very much the same function as the mangroves in trapping suspended sediments. In many places Nipa also forms the inland transition belt from the mangrove to the fresh-water environment.

With few exceptions the tidal flats, deltas and estuarine plains are actively advancing seaward. Clear evidence of seaward extension is given by inland beach ridges such as at the mouth of the Fly River and near Malalaua where beach ridges can be found 20 and 8 km respectively inland from the present coastline.
**Beach Complexes.** Like deltas, estuarine plains and tidal flats, sandy beach complexes are extensively developed only along the south coast (Fig. 41). They include beach ridges and beach plains as well as spits and bars. Their deposits are largely derived from the sandy bedload which is carried down by the major rivers and deposited at their mouths. Here it is picked up by the generally strong longshore drift and moved along the coastline. Owing to the prevailing south-easterly winds the main direction of movement is north-westerly more or less parallel to the coastline. This north-westerly drift can change locally in response to orographic obstacles such as headlands and deltas. The sediment supply from cliffed coasts is negligible because these are rare, and where they do occur the wave action is nearly always dampened by fringing reefs. Coral sand and debris is, however, present on most beaches, especially the smaller ones.

Sand bars are described by Tricart (1972) as key depositional features and characteristic elements in the littoral landscape of the tropics and this applies fully to the south coast of Papua New Guinea. The formation of sand bars, sand barriers and sand spits...
is of vital importance in the process of seaward extension of the
land and is a kind of pacesetter for the progradation. A typical
situation is illustrated in Fig. 42 and Plate 61. On the side of river
entrances spits and offshore bars develop wherever there is some
protection from the main longshore current. The westerly longshore
drift is quite obvious, resulting in a westward growth of the spit.
Behind the spits and bars the longshore current is no longer active
and the delta-building even from very small rivers is greatly in­
creased. With the gradual extension of the small deltas the mangrove
also advances seaward with its pioneer species *Sonneratia* being
later gradually replaced by *Rhizophora* and *Bruguiera*. This process
has repeated itself many times and led to a step by step advance of
spits, bars and mangroves whereby the inland bars become part
of the mangrove. The existence of former beach ridges or barriers
within the mangrove is often clearly visible on aerial photographs
as a striated pattern running roughly parallel to the coast, often
in association with straight channels.

In areas of very active sand transport, series of closely spaced
beach ridges separated by narrow swales have developed without
any associated mangrove. The ridges are most prominent near
the shore where they reach up to 2 m in height, but gradually
decrease in height inland where they form a gently undulating sand
plain. The less pronounced form of the older inland ridges is pro­
bably due to continued modification by slope processes, particularly
slope wash, accelerated by man’s disturbance of this often intensively
used environment.

**Coastal Dunes.** True dune formation is rare in Papua New Guinea
and this applies to the inner tropics in general (Jennings 1964;
Tricart 1972) owing to a number of factors, the most important
of which are lack of strong winds, high soil moisture, and dense
vegetation. Bird (1968) is doubtful whether vegetation has much
effect because it would simply trap the sand and facilitate dune
growth and he therefore regards the failure of sand supply as the
crucial factor.

There is, however, one area on the south-east coast at Hood Bay
where coastal dunes are prominent and form an extensive sandy
foreland or strand plain (Plate 62, Fig. 43). Most of the dunes are

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*Plate 62*

Stabilised parabolic dunes (back­
ground) and old strand plain
(foreground) at Hood Bay
parallel coastal dunes representing a succession of foredunes associated with a prograding shore. In addition there is a series of westnorth-westerly (280°–290°) trending parabolic to linear dunes 20–35 m high and 0.5–1.5 km long which occur at the north-western end of the strand plain. Both the parallel and the parabolic dunes decrease in height on moving south-west away from the Kemp Welch River,
which is the obvious source area of the beach sand. All the dunes are well vegetated.

The parabolic dunes are covered by eucalypt savanna which has a dense grass layer while the parallel dunes are cultivated or also covered by grass. Some larger patches of degraded rain forest are also present, an indication that this would be the natural vegetation cover for most of the area.

The present foredune is covered by sand-binding creepers and grasses that have the dual function of binding the sand and trapping new sand blown from the beach. On the foredune active sand accumulation and blowouts indicate present activity.

The dune complex at Hood Bay raises two main questions: firstly why do well-developed coastal dunes occur only here and not along the other stretches of the coastline; and secondly under what conditions did the dunes form, in particular the parabolic dunes? The answer to the first question becomes obvious when one looks at the orographic position of the dune field along the coastline (Fig. 41). The dunes occur exactly where the coastline changes direction from a predominantly westerly to a north-westerly trend, and the Hood Bay area is clearly the part of the coast most exposed to the south-easterly trade winds that blow consistently and with great force during the dry season. There are no wind velocity measurements from this area but data from Port Moresby, which is less exposed than Hood Bay, show that wind velocities exceeding 11 knots occur on an average of between 20 and 23 days per month during the dry season. Also favourable for dune development here, although not a prime factor, is the proximity of the Kemp Welch River through which a great amount of sandy material seems to be discharged.

The second question is more difficult to answer as there are several possibilities, none of which is conclusive.

The dunes could have formed under different environmental conditions such as greater dryness or more wind. Man-induced changes to the environment could also have initiated dune formation, or it could have taken place under environmental conditions similar to those of today but the process of dune formation ceased as the beach prograded.

First the age of the dunes must be considered. Both the parabolic dunes and the parallel dunes are undoubtedly young features. There are no signs of weathering and there is virtually no soil development. Since the dunes are clearly adjusted to the present sea level their formation could not have taken place before sea level rose to approximately its present position at about 6000–5000 years B.P.

There are no direct data on the environmental conditions along the coast during the last 5000–6000 years, but evidence from other parts of Papua New Guinea (Hope 1973) and theoretical considerations (Nix and Kalma 1972) suggest that climatic and environmental conditions were similar to the present ones. There is certainly no evidence for considerably drier or windier conditions.

The second possibility of man-induced dune movement can be discounted because man’s agricultural and burning activities are at present more intensive than they ever were in the past, and although
most of the dune area has been cleared no sand movement takes place except along the present beach and foredune.

The third possibility is regarded as the most likely. Dune development would have started once the sea level became stabilised. The first foredunes to develop reached heights of some 30 m near the mouth of the Kemp Welch River, gradually decreasing towards the south-west. Blowouts developed as the foredunes grew in height and eventually enlarged to form parabolic dunes. The shoreline must have been stationary for a considerable time near these foredunes to allow this sand accumulation. Eventually the shoreline prograded south-easterly and the high dunes became separated from the beach and the active sand supply. The newly developing foredunes would also have provided some protection for the older dunes from strong onshore winds.

Prograding of the shoreline and foredune formation are still active at present, which supports the notion that the dune system could have formed under climatic and environmental conditions similar to the present ones. Whether parabolic dunes develop seems to be largely a question of the growth and size of the foredune, which in turn depends on the rate of progradation. Slower rates of progradation produce higher foredunes and increase the probability that parabolic dunes will develop.

Coral Reefs
Coral reefs are certainly the most fascinating features of the

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Figure 44
Distribution of reefs

- Fringing reefs
- Barrier reefs, patch reefs and atolls
- Bathymetry in metres

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humid tropical coastlines because here biological factors play their most elaborate and extensive role. Papua New Guinea is richly endowed with reefs and almost every type of reef and reef island is present (Fig. 44). In spite of their considerable extent and richness in coral growth very little is known about the geomorphology or biology of these reefs. Apart from Davis's (1922a, b) papers on the reefs of the Louisiade Archipelago, which are based on examination of the British Admiralty charts, there are only two actual field studies (Weber 1973; Stoddart 1972), one of which deals solely with damage to coral reef communities by earthquake (Stoddart 1972). In addition Chappell (1974a) briefly investigated modern coral reefs in association with his study of uplifted coral terraces.

Darwin's classical concept of reef development distinguishes between fringing reefs as the basic reef type and the barrier reef and atoll reef as its main derivatives. In spite of severe criticism as to the universality of Darwin's concept these subdivisions have been used ever since in the description of reefs, whether Darwin's model is accepted or not. In addition to these three types there are several subtypes, the most important of which are the round or ovoid platform and patch reefs which are to be found in water of moderate depth, generally on continental shelves (Fairbridge 1967).

**Barrier Reefs.** An extensive barrier reef system (Fig. 44) runs along the south-eastern coast of Papua New Guinea between 4 and 12 km offshore and continues eastward into the Coral Sea where it forms a discontinuous series of easterly trending reefs. At its eastern extremity it surrounds Tagula Island in a wide semicircle and continues north-westerly towards the D'Entrecasteaux Islands where it loops to the north around the Trobriand Islands.
This barrier reef system consists largely of elongated narrow ribbon reefs at the windward side and more irregularly shaped often atoll-like reefs in the lee side. Many of the reefs have sandy cays or uplifted coral islands associated with them (Plate 63).

The Tagula Barrier Reef surrounding an obviously submerged group of islands is a classical example of a barrier reef system and in fact Davis (1922a, b) has used it as an example in a forceful defence of Darwin's subsidence theory. Extensive barrier reef systems are largely lacking along the north coast of the mainland and around the larger islands but some discontinuous barrier reefs occur off the Morobe coast, off Madang, on the east coast of the Gazelle Peninsula, the north coast of Bougainville and around New Hanover and Manus Island.

Fringing Reefs. Fringing reefs occur along most of Papua New Guinea's coastline in spite of the generally mountainous hinterland and the considerable runoff and sedimentation associated with it. They are naturally most abundant around rocky headlands (Plate 37) and around islands, some of which are completely surrounded by them. Weber (1973) reports from the Port Moresby fringing reefs that rocky promontories are usually surrounded by well-developed narrow but vigorously growing reefs, while the bays in between are either devoid of reefs or have wide shallow reefs supporting little coral growth except at their seaward margin. A similar pattern of bays and headlands has been observed by the author in the Vanimo area where a nearly continuous fringing reef extends from Vanimo Peninsula to and beyond the Irian Jaya border (Plate 64).

In contrast with the Port Moresby reefs, reef growth here is virtually
absent from the entire reef flat. Along the north coast of the Huon Peninsula narrow fringing reefs occur in connection with narrow elongated shallow lagoons (Chappell 1974a). In the lagoons terrigenous sediments are negligible. A typical example of one of these reefs is shown in Fig. 45.

Best developed are the fringing reefs around the islands, especially the smaller ones. Nearly continuous fringing reefs border the shores of Tagula Island, the D'Entrecasteaux Islands, most of New Ireland and numerous smaller islands. Fringing reefs are, however, conspicuously absent from Misima Island and around much of Bougainville.

**Platform and Patch Reefs.** Platform and patch reefs are abundant between the Louisiade Archipelago and the Trobriand Islands where they occur inside the barrier reef system, sometimes forming part of it. All types are present, ranging from round to ovoid reef patches to atoll-like circular platform reefs surrounding shallow lagoons. Reef patches also occur sporadically along the north coast of the mainland and off the north coast of central New Britain.

**Atolls.** Most atolls and atoll-like reef structures occur like the patch reefs in the barrier reef area off the eastern end of Papua
New Guinea. The Conflict Group of islands is an oval atoll about 10 × 20 km in diameter with several sandy cays at its rim. The Egum Atoll situated half-way between the D'Entrecasteaux Islands and Woodlark Island is of similar size and represents a classical almost-atoll with a small volcanic island rising from the centre of the lagoon. Other atolls and almost-atolls are to be found north of Bougainville (Kilinailau Island, Tauu Island) and west of Manus Island, where the Ninigo Group includes seven atolls.

**Reef Development in Papua New Guinea.** Much too little is known about Papua New Guinea's reefs to enter into any detailed discussion on their development, but some general points can be raised. The distribution of modern reefs clearly shows a preference for reef development along the south coast of the mainland and off its eastern tail end, an area which in comparison with the northern parts of Papua New Guinea is relatively aseismic and tectonically stable. The presence of submarine ridges and platforms and a shallow shelf, even though it is relatively narrow, are also favourable. In contrast, conditions for reef growth along most of the north coast and around the larger islands seem to be less favourable and reef types are largely restricted to narrow fringing reefs, some of which are not actively growing. The reason for this appears to be the frequent seismic activity (Stoddart 1972; Brooks 1965), continued uplift (Chappell 1974a) and the generally very steep offshore slope which does not permit any extensive seaward growth of the reefs. In addition reef growth seems to be inhibited in areas of extensive volcanic ash fall. This is probably responsible for the poor development of reefs around Bougainville and parts of New Britain. Reef growth along much of the Solomon Islands, which are situated in the same geotectonic belt as northern Papua New Guinea, is similarly impoverished (Stoddart 1969).

The Papua New Guinea reefs are typical examples of 'mobile belt reefs' (Fairbridge 1967) with eustatic sea level changes superimposed on a complex history of tectonic movements, and there is certainly no simple or single answer to the question of their development and distribution.

**Uplifted Coral Reefs and Terraces**

Uplifted coral reefs and terraces occur in great numbers along Papua New Guinea's coastline, supplying ample evidence of relative movements of land and sea (Fig. 41). They range in form from simple raised reef platforms like the Trobriand Islands and Buka Island to spectacular staircase-like terrace sequences such as occur along the north coast of the Huon Peninsula, the south coast of New Britain and the north-west coast of New Ireland. The height, number and state of preservation of these terraces vary greatly from place to place and any attempt to correlate them on an altitudinal basis must fail.

On Misima Island up to seven narrow coral terraces occur along the southern coast, the highest of which is 460 m above sea level. The terraces decrease in number and diminish in height eastward and only a single terrace 5 m above sea level occurs along the eastern end of the south coast (Davies, unpublished BMR record).
Several uplifted coral islands occur south of Misima Island and form part of the barrier reef system, indicating that a simple subsidence model as envisaged by Davis (1922a, b) does not satisfactorily explain the occurrence of this barrier reef system.

The Trobriand Islands and Woodlark Island are situated on a well-defined submarine ridge extending from the Lusany Islands in the west to the Laughlan Islands in the east. Most of the Trobriand Islands form an uplifted barrier reef and atoll system. The former atoll structure is particularly well-preserved at Kitava Island, the easternmost island of the group. The island has been raised some 100-140 m above sea level and consists of a peripheral ridge, the former barrier rim, surrounding a closed basin, the former lagoon. At least five coral terraces are associated with this island (Ollier and Holdsworth 1970).

The northern part of Kiriwina Island is similar in structure but the peripheral ring is only about 50 m high and only partially encloses the lagoon. No distinct terraces are developed here or on any of the other smaller islands, which are all essentially similar to Kiriwina in structure and physiography.

In contrast with the Trobriand, Woodlark Island has a volcanic core around which an extensive coral platform has developed. Along its northern margin the platform rises abruptly from the sea with cliffs up to 100 m high. It slopes gradually southwards until it submerges and continues as a submarine platform, to terminate with a barrier reef rim forming an approximate semicircle around the southern part of the island. The volcanic hills protruding through the coral platform are situated at the southern end of the raised platform. The coral is underlain at a depth of about 50 m by recent submarine clay (Trail 1967), indicating a much greater supply of clastic sediments before the formation of the reef platform. This was probably during lower sea levels when much larger areas of rock were exposed.

The best known and most spectacular area of uplifted coral is to be found along the northern seaboard of the Huon Peninsula. Here there has developed an impressive flight of coral terraces rising to over 600 m and extending over a length of more than 80 km (Plate 65). Under the initiative of J. Chappell this terrace sequence has been subjected to detailed investigations in the last decade and a number of significant results have been produced, not only about the record of sea level changes and relative movements of land and sea (Chappell 1974a; Veeh and Chappell 1970; Bloom and others 1974) but also about climatic theories (Veeh and Chappell 1970; Chappell 1974c), tectonic problems (Chappell 1973b, 1974b) and computer simulation of geomorphic evolution of small valleys in the terrace sequence (Chappell 1974d). The following section is based on Chappell (1974a) and Bloom and others (1974) and is largely confined to the significance of the terraces as a record of the sea level changes in relation to the rising land.

There are over 20 single terraces, very well preserved in the lower third of the sequence. Two main terrace forms are developed. The first consists of a broad surface with a low frontal ridge and a wide shallow depression behind, obviously representing a barrier reef lagoon.
The rate of uplift has varied considerably along the coast as the terrace height decreases to the north-west and also there is considerable variation in the height ratio between terraces.

Close stratigraphic investigation of terrace sections and radiometric dating of the terrace reefs revealed sea level maxima at about 7500, 30,000, 40,000, 60,000, 85,000, 107,000, 125,000, 140,000, 185,000 and 220,000 years B.P. These dates are in relatively good agreement with similarly dated sequences from elsewhere, but appear to represent a more complete sequence.

Assuming a constant rate of uplift for each terrace section and comparing the dated terraces with palaeo-sea levels from other tectonically more stable areas, rates of uplift for each section are calculated and palaeo-sea levels estimated. Rates of uplift vary along the coast from 3 m to 0.5 m per 1000 yr. The sea level curve compares reasonably with Emiliani’s (1970) generalised palaeotemperature curve based on oxygen isotopic analysis of planktonic foraminifera from deep sea cores. There are some substantial offsets in the peaks of the two curves and more dating will be necessary to refine the record of the terrace sequence, which could well reveal a nearly complete history of Pleistocene sea level fluctuations. A major problem is the complete disagreement of Emiliani’s and Chappell’s curves with the Pleistocene climate curve established by Ericson and others (1965).

West of the Huon coast coral terraces are less prominent, although low discontinuous coral platforms occur along much of the coastline, in particular between Madang and Alexishafen, near Bogia, around Wewak and west of Vanimo. Some of the small offshore islands near Wewak and Aitape also consist of raised coral. Most of these terraces are only 3–7 m above present sea level; in some areas two terrace levels are developed (Panzer 1933). No dating has been done for any of these terraces.

However, Hossfeld (1964) has reported the find of a former tidal mangrove swamp near Aitape, some 13 km inland and 52 m above present sea level. Four C\textsuperscript{14} dates gave ages between 4400 ± 85 and
5070 ± 140 years B.P. This would imply an extreme youthfulness of this former strandline and a rate of uplift of about 10 m/1000 yr, which is considerably higher than the maximum rates determined by Chappell (1974a) on the Huon coast. Since there are no other geomorphological features associated with the Aitape site to lend support to this spectacularly high rate of uplift, the interpretation of the dates or of the site must be doubted. In fact the extensive development of beach ridges around Aitape indicates that the position of the shoreline has been relatively stable for some time.

Terrace sequences similar to those along the Huon coast have been observed along the south coast of New Britain and the east coast of Gazelle Peninsula, where up to six terraces occur rising up to 500 m at Jaquinot Bay. As along the Huon coast the terrace heights vary laterally.

The north coasts of New Ireland and New Hanover are almost entirely surrounded by raised coral reefs. Höhnen (1970) reports that broad terraces are developed at 300–400 m, 80–85 m and at close intervals from 65 m down to sea level. The sequence appears to resemble that of the Huon coast closely but nothing is known about their possible age relationships. Wave-cut notches occur extensively along the north coast. According to Christiansen (1963) a pair of notches is usually developed (Fig. 26). He assumes that the lower notch has been formed by present-day wave action, while the upper notch was formed during a eustatically controlled postglacial higher sea level. Höhnen (1970), on the other hand, observes that the elevation of the notches appears to increase south-eastward, although this trend is locally disturbed by recent faulting, and suggests that tectonic uplift rather than eustatic sea level changes have been the main cause for the development of the notch. There is also no good evidence that sea level in post-glacial times exceeded the present level (Bloom and others 1974).

As on other coastlines of Papua New Guinea, superimposition of eustatic sea level changes on tectonic movements is to be expected and without dating and more precise stratigraphic analysis of the notches the relative significance of these factors cannot be assessed. The raised coral platform of Buka Island and northern Bougainville represents largely a former barrier reef and atoll system. The reef is raised to a maximum of 100 m above present sea level and its coral reef features such as the outer barrier reef rim, the flat lagoon floor, patch reefs and lagoon islands are exceptionally well preserved. Incipient karst development has started with the formation of some solution dolines. Radiocarbon dating of a clam shell from the uplifted reef flat has shown that it is beyond dating range (Speight 1967). The state of preservation of the platform in comparison with the terraces on the Huon coast suggests that their age is not greater than upper Pleistocene.

From this brief summary it is clear that the uplifted coral terraces offer great possibilities for detailed investigation into the relative movements of sea and land. The work of Chappell and his associates has already provided an important addition to our understanding of the complexity of these processes, and further work in this direction is desirable.
This chapter deals with geomorphic processes which act on or beneath the surface and which are finally responsible for sculpturing the landscape. Investigation of most of these processes in a tropical area like Papua New Guinea is particularly difficult not only because of the often mentioned forest cover and lack of access but also because a great number of processes have been acting simultaneously or alternately on the same site for long periods. Short-term high-intensity processes such as landslides and mudflows, triggered off by earthquakes, volcanic activity and exceptional storms, are usually superimposed on long-term, generally low-intensity processes such as fluvial erosion and deposition, slope wash and creep. The interactions between these processes makes judgments about their intensities very difficult. This is clearly a field for quantitative process studies.

Although no long-term measurements of any specific processes have been made some estimates of the rates of total denudation and deposition are available. The estimates are based on a variety of indirect measurements or observations. Ruxton and McDougall (1967) calculated denudation rates on the Hydrographers volcano, the age of which was determined by potassium argon dating, by reconstructing its original shape and comparing the former and present cross-sectional areas. The denudation rates varied from 8 cm per 1000 yr in the area of lowest relief to 75 cm per 1000 yr in the areas of highest relief.

Simonett (1967) used quantitative analysis of aerial photographs to estimate the volume of earth lost through landsliding during the 1935 earthquake in the Toricelli and Bewani Mountains. This ranged from 40 cm in granite areas which were the most affected to 12 cm and 7 cm in weak and mixed sedimentary rocks respectively. Based on Brooks’s (1965) estimation of the frequency of occurrence of earthquakes in this area, Simonett expects the longer-term rates of denudation to be of the order of 10 cm every 70 to 100 years (1.00–1.43 m per 1000 yr). Newer data show that earthquakes occur with a considerably higher frequency (Denham 1969) and the denudation rates are therefore also likely to be substantially higher. Pain and Bowler (1973) calculated a loss of material equivalent to 11 cm surface lowering during the 1970 Madang earthquake.

Against these high rates data from other areas of Papua New Guinea are very modest indeed. Pain (1973) has calculated an average rate of 27 cm per 1000 yr for the Kaugel basin area. R. Blong (pers. comm.) has found a rate of only 1–2 cm per 1000 yr for the last 5000–6000 years in the upper Wahgi basin. My own estimates
from the Lake Trist area are similarly low, amounting to a mere 0.5 cm per 1000 yr in an area completely covered by rain forest. In the last two cases the erosion is probably solely due to slope wash.

This contrasts with an estimated average denudation rate of 50 cm per 1000 yr on Mt Giluwe, which is based on a calculation similar to that employed by Ruxton and McDougall (1967). However, the problem with this calculation is that it assumes uniform rates of erosion over very long time periods, which in the Papua New Guinea situation is unrealistic, particularly where volcanic landforms are concerned, which normally have much higher erosion rates in their early stages of development. This has been demonstrated by Ollier and Brown (1971), who showed that erosion rates on Vulcan, a young volcano at Rabaul, amounted to a staggering 19 m per 1000 yr, but that almost all the erosion took place in the first 5 years of existence of the volcano, so that extrapolation of the rate over a long period is not meaningful.

Data on rates of deposition are similarly inadequate but also point to the great variability in the rate of processes. The enormously high rate of deposition on some of the Markham Valley fans has already been mentioned. The accumulation of 1000 m of sediments in the graben also gives a good indication of the rate of deposition. From the upper Wahgi basin Blong (unpubl.) has established a number of sedimentation rates largely based on borehole records and C14 datings. The figures vary by nearly four orders of magnitude from 0.19 mm per yr to 1.3 m per yr, but the great majority of values are about or below 1 mm per yr.

As Blong points out the extreme rates also reflect the extremes of the measured time scale, which varies between 5000 years and 8 months; an indication of the difficulties in extrapolating short-term observations over long periods.

WEATHERING

Although weathering is seldom directly a land-forming process in the Papua New Guinea environment it is of great importance as a process of weakening and breaking down rocks and allowing slope processes to remove rock in the form of finer particles or in solution. Papua New Guinea is situated well inside the humid tropics and its climate is characterised by constant high temperatures, a high degree of humidity, and excessive rainfall. Approximately 90 per cent of the land area of Papua New Guinea receives rainfall in excess of 2000 mm and about 80 per cent has average daily and annual temperatures of over 20°C. The climatic conditions are thus highly favourable for intense chemical weathering. In spite of this thick weathering mantles with mature weathering profiles are not as common as one would expect (Haantjens and Bleeker 1970; Ruxton 1969; Löfler 1972b).

The processes of weathering are often discussed mainly in terms of precipitation and temperature and certain types of weathering are associated with specific climatic conditions.

While there is no denying that a correlation between climatic conditions and weathering exists, especially if one considers extreme cases, the detailed situation is often much more complex. This
applies very much to Papua New Guinea where in spite of a relatively large number of observations partly supported by analytical data the author finds it very difficult to make valid generalisations. Differences in rock type, slope steepness, drainage conditions, slope dynamics and age all seem to complicate the picture greatly. Even within one rock type weathering profiles often vary considerably. For instance in granodiorite they may vary from a typical red mature profile as on Kassam Pass to a similarly deep profile showing only granular disintegration, but no alteration into clay minerals, in the Gembola Valley to a nearly unweathered profile in the Bewani Mountains. Similar observations have been made on metamorphic rocks. The following statements on weathering are therefore to be regarded as rough generalisations.

Kinds of Weathering

The two main kinds of weathering are physical weathering, which is the fragmentation of rock by purely physical means, and chemical weathering, which involves chemical alteration of the minerals.

Plate 66
Summit area of Mt Wilhelm with steep rock faces and screes. Frost weathering active on castellated rock faces.
Purely physical weathering is rare in Papua New Guinea. Frost weathering is restricted to altitudes above about 4000 m where the night temperatures regularly fall below zero. However, the frost is not severe and only a few hours in duration; only very small rock fragments are thus loosened by this action. The extensive coarse rock screes which have accumulated below steep rock faces and cliffs on Mt Wilhelm were formed during the latter part of the Pleistocene when glaciers retreated and frost was much more severe than it is today (Plate 66).

Weathering by insolation, i.e. by rapid temperature changes, is also rare, not only because there is very little rock outcrop at the surface but also because direct exposure to sunshine is even rarer as dense forest and scrub cover all but the steepest rock outcrops. Only above the natural forest boundary at about 3900 m do rock faces become more common and here insolation may be.

**Plate 67**
Kandite weathering with mature weathering profile, Edie Creek road, Wau, altitude 2400 m. The metamorphics here have nearly entirely weathered into kaolin rich clay.
active, mainly in loosening thin platy rock fragments. Since frost weathering operates in the same areas it is difficult to distinguish the two processes.

The other kinds of physical weathering (Louis 1968; Ollier 1969b) do not appear to be involved under the humid tropical conditions of Papua New Guinea, except perhaps fracturing of rocks by fire which according to Haantjens and Bleeker (1970) has occasionally been observed; undoubtedly it is of very little significance.

Weathering processes in Papua New Guinea are predominantly chemical. Haantjens and Bleeker (1970) subdivide the chemical weathering processes into three types according to the types of clay minerals formed.

1. Kandite weathering, by far the most common weathering process in Papua New Guinea, is characterised by the formation of kaolin, halloysite or allophane. It occurs on most rock types.

2. Smectite weathering is characterised by the formation of montmorillonitic clay minerals and is largely restricted to basic and calcareous rocks in relatively dry areas.

3. Sesquox weathering is characterised by the formation of crystalline or amorphous sesquihydroxides and is restricted to ultrabasic rocks.

The degree of weathering is described as mature, immature, and skeletal. Mature weathering means that a very large amount of clay minerals and a very small amount of rock fragments are present, while the reverse is the case for skeletal weathering (Plates 67, 68). Immature weathering describes an intermediate position (Haantjens and Bleeker 1970; Ruxton 1967). Each of these terms may be further

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Plate 68
Skeletal weathering in finegrained sedimentary rocks at 1700 m altitude in highlands near Kundiawa
specified according to the depth of weathering. The degree of weathering is often but by no means always consistent with the depth of weathering. For instance very deep profiles of immature, even skeletal weathering can be found on granodiorite, while shallow but maturely weathered profiles have been observed on limestone and ultrabasics.

The great majority of weathering profiles studied in Papua New Guinea during CSIRO surveys can be characterised as relatively shallow immature profiles of the kandite weathering type (Fig. 46). Thick mature weathering profiles which are reported to be widespread in other tropical areas (Ollier 1969b; Wilhelmy 1958) are notably rare. In the mountain areas they mostly occur on broad crests and relict surfaces and in the lowlands on relict alluvial plains and fans. Recent alluvial plains are generally unweathered. Skeletal weathering is predominant on very steep mountain slopes which mostly show evidence of active mass movements. There is generally a strong correlation between depth and degree of weathering and slope dynamics. Rapid denudation processes seem to prevent the development of mature thick weathering profiles on most steep slopes.

There is some discrepancy in records on weathering depths in Papua New Guinea. While most authors claim that depth and degree of weathering are relatively shallow and immature, Bik (1967) reports relatively deep and mature weathering profiles in the western highlands. There is, however, no real disagreement as there is no denying that deep weathering profiles do occur. Bik’s observations apply to the western parts of the highlands where volcanic ash, lavas and tuffaceous sandstones, which weather easily, cover large areas of relatively gently sloping terrain. Deep, maturely weathered profiles are also common in much of the Kainantu-Goroka area, particularly on Kassam Pass where they are associated with a relict landscape.

Haantjens and Bleeker (1970) observed an altitudinal zonation of weathering with the proportion of mature weathering gradually decreasing with increasing altitude. Mature weathering was rare above 1800 m and virtually absent above 3000 m. According to the author’s observations this zonation is often upset and in many

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**Figure 46**
Degree of weathering in Papua New Guinea

- Unweathered
- Skeletal weathering
- Immature weathering
- Mature weathering

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instances the reverse situation has been noted with deep and more mature weathering profiles at higher altitudes. Examples of this occur along the highland highway, particularly at Kassam Pass and Dalau Pass, at Murray Pass near Mt Albert Edward, around Lake Trist and along the Edie Creek road near Wau (Plate 67). In most cases the occurrence of these mature weathering profiles is obviously related to relict surfaces. But it also appears that the survival of deep weathering profiles at high altitudes is favoured by the virtual absence of slope wash (see section on slope wash).

**Rate of Weathering**

The rate of weathering in the humid tropics is generally regarded as very rapid and some extraordinary rates have been reported in the literature, such as weathering to clay of freshly exposed granite boulders within a decade in Malaya and even within one year in Indochina (Scrivenor 1931; Blondel 1933). No such spectacular rates have been recorded from Papua New Guinea and the author feels that these figures should be treated with caution and not be regarded as representative of the rate of weathering in the humid tropics.

Weathering studies of dated ash layers derived from Mt Lamington in north-eastern Papua by Ruxton (1968b) revealed that at altitudes between 1000 and 1400 m volcanic glass changed almost completely into the clay mineral allophane between limits of 8000 and 27,000 years. On the assumption that alumina remained constant during the weathering process the average silica loss was 4–6 mg per cm² per year. Crystals such as hornblende and feldspar, on the other hand, showed little etching in ashes younger than 20,000 years.

Radiocarbon dating of alluvial deposits has given the following results. In the Wewak-Lower Sepik area peat associated with unweathered alluvium was dated at 3500 years B.P. while alpine peat overlying nearly unweathered rock at 3000 m gave a date of 5800 years B.P. (Haantjens and Bleeker 1970). On Mt Giluwe charcoal obtained from unweathered alluvium at 3100 m was dated at 3780 and 1800 years B.P. (Blake and Löffler 1971).

On the Fly Platform Blake (1971) reports maturely weathered sediments that are at least 27,000 years old. In the Kaugel Valley at an altitude of 2300 m a number of dates show that alluvium over 40,000 years is only slightly weathered (Löffler 1972a; Pain) 1973).

The observations listed do not allow conclusive statements as they vary considerably and are highly inadequate in number, but they show the following trend. At low altitudes skeletal weathering takes place in about 5000 years, immature weathering requires 5000 to 20,000 years, while mature weathering requires more than 20,000 years (Haantjens and Bleeker 1970).

At higher altitudes the rate of weathering is clearly slower though exactly how much is difficult to tell. The observations in the Kaugel Valley at 2300 m suggest that even within 40,000 years only immature weathering takes place.

Finally the question of volume changes resulting from transformation of primary into secondary minerals is to be considered. Weathering generally leads to expansion because high density
primary minerals are transformed into low density secondary minerals, but weathering at constant volume can also take place (Ollier 1969b; Haantjens and Bleeker 1970; Brewer 1964). Haantjens and Bleeker present some evidence that under certain conditions significant loss of volume can also occur. They examine two cases of kandite weathering and two cases of sesquox weathering and calculate the loss of volume on a constant aluminium basis. Their results show that the two cases of kandite weathering result in volume expansion of the order of 150–200 per cent while the two cases of sesquox weathering in ultrabasic rocks result in volume losses of 85–90 per cent. This would indicate that in ultrabasic rocks the removal of material in solution far outweighs the formation of clay minerals. Haantjens and Bleeker’s results are supported by the author’s investigations in ultrabasic areas near Lake Trist where genuine karst features such as dolines, small caves and pipes occur and can be explained only by solution of the magnesium which makes up about 50 per cent of the rock (Löffler in press).

**FLUVIAL EROSION**

Fluvial erosion is undoubtedly the most important process operating in the mountainous landscape of Papua New Guinea. Because of the great tectonic relief there is a huge amount of energy available for the rivers to erode deeply into the bedrock. The gradients of the major rivers and their discharges are shown in Table 1. The rivers contain large amounts of coarse gravel and boulders which travel considerable distances downstream (Plate 69). In the Strickland, for example, coarse gravel, mainly quartzite and limestone, forming extensive gravel banks and bars, is to be found up to 100 km downstream from where the stream leaves the mountains. Similar observations are reported by Speight (1965b) from the Angabunga River, Papua, and Löffler (1972b) from the Pual River, Vanimo area.

The river beds are narrow, mostly cut into bedrock with the river occupying nearly the entire width of the bed. Side slopes are very steep and straight and there are few lower slope concavities as most of the debris transported to the foot of the slopes is quickly

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Plate 69
Gravel in stream bed of tributary of Tauri River in the Aure area. Altitude 200 m.
removed by the rivers (Plate 70). Strongly oversteepened profiles are also rare as slope processes seem to keep pace with the rate of incision. Only in limestone areas are oversteepened slopes more common.

As already described in the chapter on landform types the dissection pattern varies greatly with rock type, rock structure and tectonic structure. In most acid igneous and metamorphic rocks the drainage pattern is coarse-textured and angular, reflecting the joint and fault pattern. In ultrabasic rocks and to a lesser degree in basic rocks the drainage pattern tends to be even coarser but clearly lacks the angularity. Instead most rivers flow in incised meanders. This is due to the more massive character of these rock formations, which have been much less faulted and fractured than other rocks.

In sedimentary rocks the drainage pattern is clearly finer and is often closely related to rock structures. Fault-controlled river courses are also common. The drainage pattern varies from very fine-textured in soft fine-grained sedimentary rocks such as marl and mudstone to fine and medium-textured in coarser and more consolidated as rocks such greywacke and sandstone.

Quantitative data on bedload, suspended load and the amount of dissolved solids carried by the rivers and streams are very few. There is virtually no information on bedload transport, although it is undoubtedly the major form of fluvial transport in the mountains of Papua New Guinea. There is a rough estimate by the author from the Bewani Mountains where a temporary dam was filled up with sediments including bedload and suspended load within 30 years or less (Löffler 1970b). The total volume of sediment was approximately $2.5 \times 10^6$ m$^3$ from a catchment of 78 km$^2$. Bedload transport in the braiding flood-plains of the Markham River and its northern tributaries appears to be particularly great. Buried tree trunks bear witness to these rapid processes. Henley (unpublished, quoted in Holloway and others 1973) reports that partially buried trees at some locations indicate that up to 2.7 m of gravel have been deposited within 2 years. J. Robinson (pers. comm.) reports a raising of the Leron River by at least 0.5 m in 3 months. These are extraordinary rates and should not be regarded as representative, yet they give

Plate 70
V shaped valley with very steep side slopes in the highlands
an indication of some maximum rates of bedload movement.

The calibre of the bedload is also truly impressive. Boulders of several cubic metres have been observed by the author in numerous larger rivers, and although in some cases they have not moved very far in many instances the source area has been found to be many kilometres upstream (Plate 71).

Data on suspended and dissolved load are similarly poor; long-term measurements especially are completely lacking. There are a few randomly collected data on dissolved solids which warrant mention in this context (Table 2). They do not allow any detailed analysis of the interrelationships of dissolved solids, discharge and catchment characteristic, as, for instance, Douglas (1968, 1969) has done for some rivers in tropical Australia and South-east Asia. However, the data are clearly consistent with his results. They

<table>
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<th>River/Area</th>
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<th>Mg</th>
<th>HCO₃</th>
<th>SiO₂</th>
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<td>14</td>
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show that the values for dissolved solids are nowhere exceptionally high and are similar to those from extratropical humid areas. Seen in relation to the usually very high run-off, the removal of material in solution is substantial but by no means as rapid as previously believed.

There is a significant difference (1.5 per cent) in the mean values of TDS and SiO₂ below and above 1500 m, obviously reflecting the decrease in rates of solution with increasing altitude.

 Suspended material is similarly low but much more variable than dissolved solids, reflecting events such as intense rain storms. In the case of the Gogol River an exceptionally heavy silt load (1700 ppm) followed landsliding after earthquake activity.

 The transport of material in these tropical rivers does not, according to these observations, appear to be fundamentally different from that in extratropical areas as often claimed by climatic geomorphologists (Büdel 1972; Bremer 1971, 1972). There is no evidence to suggest that rivers in the tropics erode their beds more by solution processes than by physical corrosion by the bedload. This applies not only to the streams and rivers in the mountain areas but also to those in the lowlands.

 The difference between humid tropical rivers and extratropical rivers appears to be more a matter of degree than of kind and is largely a result of the higher run-off associated with this climatic zone.

LANDSLIDES AND RELATED MASS MOVEMENTS

Mass movements were defined by Penck (1894) as displacements of rock and soil under the influence of gravity but without the aid of transport agents such as water, ice or wind. This basic definition has been generally accepted and will also be used here, realising of course that as always in nature the definition cannot be applied too rigorously.

Most classifications of mass movements have been derived from examples occurring in the temperate areas of Europe and North America (Sharpe 1938; Varnes 1958; Zaruba and Mencl 1969;
Louis 1968). In spite of this they provide a reasonable framework for the classification of the mass movements occurring in the humid tropical environment of Papua New Guinea.

The three basic types of mass movements are falls, slides and flows (Varnes 1958) all of which occur in Papua New Guinea. Falls are a relatively simple form of mass movement occurring when rock or weathered material and soil move through the air by free fall. Slips move en bloc or broken up into particles along one or several slip planes while flows move in the form of viscous fluids. The transition from highly deformed slips to flows is continuous and all gradations exist.

Falls

Pure falls are not very common in Papua New Guinea, mainly because the rapid rate of weathering restricts the formation of extensive areas of cliff and rock outcrops. In altitudes below 4000 m falls are largely restricted to areas of limestone which because of their particular properties tend to be more resistant to weathering and erosional processes, resulting in the formation or preservation of cliffs and rock outcrops. Such cliffs are commonly formed along river gorges (i.e. Purari, Erave, Hegigio, Strickland), along hogbacks (Elimbari Ridge) and on structurally controlled limestone plateaux such as the Hindenburg Plateau and the summit of the Saruwaged Range.

In other rock types falls only occur very locally along undercut river banks and along gorges. Rock falls have also been reported to occur during severe earthquakes, but again their contribution to the general rate of denudation is very small (Pain 1972b).

In areas above 4000 m where glacial erosion during the Pleistocene has led to the formation of oversteepened cirque headwalls and where present-day frost is active, rock fall is an important process (Plate 66). The areas involved are negligible in size; they are restricted to Mts Wilhelm and Giluwe and cover only a few square kilometres.

Slides

Slides are much more complex mass movements than falls, involving downslope movement of rock, rock debris, weathering mantle, and soil en bloc or in a multitude of fragments. The deformation results from a high shearing stress which exceeds the shearing resistance. In Papua New Guinea the two main types of slide commonly observed are rotational slides (also named slumps) and debris slides.

Rotational sliding or slumping is a slope failure en bloc along a mostly well-defined concave internal slip plane, often involving a more or less well-developed backward rotation of the slip block (Plates 72, 73). This backward rotation can lead to the formation of enclosed swampy depressions and occasionally short-lived lakes between the slip block and the slump headwall (Fig. 47).

Dynamically the backward rotation also restores the equilibrium of the moving block and the process of slumping therefore does not lead to displacement of material over great distances, as happens with other forms of mass movement. In fact a great number of
Plate 72
Very large slump on Mt Suckling in metamorphics. The slope failure involves en bloc movement of a whole slope segment as well as debris slides and debris avalanches.

Plate 73
Small slump in fine-grained sediments (shale and siltstone) near Kundiawa

Figure 47
Rotational slumps
slumps have very little erosive effect and tend only to modify the slope angle rather than causing actual removal of material.

Slumps vary enormously in size from miniature slumps with vertical displacements of the order of a metre or less to huge slumps with displacements of several tens of metres involving millions of cubic metres of rock (Plates 72, 73). These latter slides tend to be complex, involving not only slumping but also debris slides and flows.

Slumps most commonly occur in areas formed on relatively homogeneous sedimentary rocks such as mudstone, marl, sandstone and greywacke. They also occur on low grade metamorphics, though less frequently. On igneous rocks true slumps are rare because the more massive character of these rocks is not favourable for the development of deep-seated slip planes. On deeply weathered igneous rock small slumps do occur in the weathering mantle.

Studies of the distribution of slumps on aerial photographs have shown that slumping, especially of small size, is more common in grassland areas than in forest (Simonett and others 1970; and the author's own studies (Plates 55, 73)). This difference may, however, be more apparent than real as many slumps under forest are practically undetectable on aerial photographs and even on the ground the dense forest cover prevents wide views and appreciation of the frequency of occurrence of small slumps. Slumps in forest areas also tend to be revegetated very quickly and are therefore obliterated in a very short time (Plate 74). In the lower Purari area for instance slumps and associated mudflows are very common under primary rain forest and studies of the age structure of these forests show that there is an approximately 50 year cycle in forest

Plate 74
Old slump in low hilly area north of Puari delta. Slumps in rain forest areas are quickly revegetated.
replacement (K. White, pers. comm.). This means that the frequency occurrence of slumps is of the same order.

In a debris slide movement of material is not en bloc but in many broken-up particles. Debris slides are shallow in depth, involving only the soil and weathering mantle. Like debris avalanches into which they grade with increase in water content and velocity, debris slides can be wedge-shaped with the thick end near the ridge crest, where the actual slope failure takes place, and a long narrow trail projecting into the valley floor. Larger debris slides tend to lack this wedge shape, and the slope failure takes place along a wide subparallel trail which maintains its width from crest to valley floor (Plate 75).

Debris slides occur on a great variety of rocks, especially those where original bedding planes, planes of foliation, and intense fracturing by joints and faults provide pre-existing zones of weakness. They are thus most common on rocks such as thinly bedded shales or densely foliated metamorphics, where the structural planes provide ideal slip planes. Debris slides are also quite frequent on igneous rocks with shallow weathering; here the surface of the unweathered rock acts as a slip plane along which the debris moves downslope. On deeply weathered igneous rocks slides are replaced by debris flows.

Flows

The three main types of flow operating in Papua New Guinea are soil and debris avalanches, earth-mud and debris flows, and soil creep. The first two types are movements of relatively high velocity, avalanching being the most rapid form of flow. Soil creep on the other hand is a very slow form of mass movement, by definition imperceptible.

Together with the closely related debris slides, soil and debris avalanches are the most common form of rapid mass movement in Papua New Guinea. They involve very rapid downslope movement of soil and weathered rock debris along narrow avalanche trails which generally follow steep gullies (Plate 76). It is probable that

Plate 75
Debris slide in shales the bedding of which is approximately parallel to the slope angle. This debris slide occurred in May 1975. Several people lost their lives during this catastrophe. (Photograph by P. Pieters, Geological Survey of Papua New Guinea.)
all avalanches are of the wet type involving various proportions of water. Dry avalanches are unlikely to occur under the highly humid conditions prevailing in Papua New Guinea.

All avalanches studied in the field are shallow in depth, involving only a very superficial layer of soil and weathered rock as well as the vegetation cover. Pain (1972b), who studied the landslides associated with the November 1970 earthquake in the Madang area, measured an average depth of debris of 44 cm and a range of 20 to 60 cm for debris avalanches. Slopes are generally very steep; according to Pain the debris avalanches occurred on slopes between 40° and 50°. This value agrees well with the author's estimates from the Bewani Mountains and is probably representative for this kind of mass movement.

Debris avalanches are not restricted to any particular rock type; their occurrence appears to be linked with topographic and environmental conditions. These are primarily high relief, great slope steepness, shallow weathering and soil mantle, and probably also dense forest cover. The great majority of debris avalanches seems to be triggered by earthquakes (Plate 77). This has been clearly demonstrated during the two earthquakes for which we have good records, the 1935 earthquake affecting the Toricelli Mountains and the 1970 earthquake in the Adelbert Range (Stanley and others 1935; Simonett 1967; Pain 1972b; Pain and Bowler 1973). Other causes leading to much more local occurrences of debris avalanches are tree falls and intense rain storms. On slopes exceeding 50° a peculiar form of avalanche first reported from New Zealand (Schweinfurth 1966) seems to take place. The movement is here initiated merely by the weight of the forest cover once it has reached a critical height and weight. In Papua New Guinea this process has so far been reported only from the Bewani Mountains (Löffler 1972b) but aerial photograph studies suggest that it occurs in many other areas.

Earthflows, mudflows and debris flows are three kinds of flow involving flow of earth material with a high degree of fluidity during the movement. The predominant grain size and water content differ-

Plate 76
Debris avalanches in Toricelli Ranges with narrow trails mostly situated in gullies
entiates mudflow (fine material, relatively high water content), earthflow (fine material but relatively low water content), and debris flow (mixed material with large amounts of coarse debris and relatively high water content). In practice the distinction is difficult to make as transitional types are more the rule than the exception and the three types are therefore treated here as one main type of flow, referred to as mudflow, which appears to be the most common type.

Mudflows can occur singly as more or less isolated features (Plate 78) or in series of flows forming extensive mudflow fans. Mudflow fans are most conspicuous in the Musa basin where they form most of the basin fill (Plate 79). Haantjens and others (1967) describe several generations of mudflow fans in various stages of dissection and varying degrees of weathering. The younger mudflow fans are undisetced and have gently undulating surfaces with occasional flow patterns in the form of buckling and corrugations. The older fans are being dissected and subjected to secondary mass movements, mostly in the form of slumping (Plate 79). The depth of the flows varies between 10 and 70 m. They consist mainly of unsorted coarse rock debris in a clay to silt matrix.

Mudflows of similar composition have been reported by Pain (1973) from the Kaugel Valley in the highlands. These flows, termed Kaugel Diamictons by Pain, appear to be continuous as there are no weathering breaks or pedogenic discontinuities within the sequence, and are therefore attributed to a single major earthquake.

The likelihood that large series of mudflows are triggered off by major earthquake shocks is substantiated by their distribution...
pattern which shows that they occur mostly in the vicinity of areas susceptible to earth tremors, such as major faultlines or volcanoes.
Smaller and more isolated debris flows occur often at the toe end of slumps or in association with other mass movement features.
Small-scale mudflows causing hummocky, buckled slopes on

Plate 78
Debris flow in deeply weathered granodiorite in the Gembola Valley, upper Chimbu. Slope failure is entirely in weathered material. Tongue shaped flow in centre. This area is part of a relict surface situated at about 3000 m.

Plate 79
Extensive debris and mudflows in the Musa basin, forming several generations of fans
generally deeply weathered rocks or on soft sedimentary rocks are to be observed frequently in grass-covered areas in the lowlands and highlands (Plate 80). Under forest small mudflows have not been observed. However, as with small slumps, this difference may be due not to reduced mudflow activity under forest cover but rather to the fact that these small-scale features are not detectable under the dense forest cover (cf. Simonett and others 1970). Behrmann (1927) reported mudflows that operate underneath the root network in the rain forest areas of the south Sepik. He used the term 'subsilvines Bodenfliessen' (soil flow under the forest) for this process. No such process has been observed by the author or other CSIRO workers in Papua New Guinea and it is therefore difficult to assess how widespread it is and what significance it has as a denudational process.

Complex Mass Movements

More often than not mass movements are complex with several types of movement involved, or with transitional types merging into one another or one type of movement serving as the triggering mechanism for another. The greater the area and the volume of material involved in the mass movement the more complex the event tends to be. This is apparent on Plate 72 where a large slump on the upper slope merges into a debris slide further downslope. Some narrow debris avalanches are also associated with the event. On Plate 55 the slumps are all closely linked with mudflows that extend from the slump toes and have obviously been triggered off by the slumping. Similarly debris slides can merge into or be replaced by debris avalanches or debris flows as the water content and velocity of the moving debris increases. In many cases classification may therefore be difficult and artificial.

Soil Creep

In contrast with all the mass movements discussed above which are sudden events, soil creep is a very slow imperceptible motion of soil under the influence of gravity (Parizek and Woodruff 1957a). Although soil creep is thought to operate in most mountainous
environments of Papua New Guinea good evidence for its occurrence is largely lacking. Most of the evidence reported is indirect and inconclusive. Terraced slopes, curved tree trunks, and destroyed and displaced fences of native gardens are generally cited as evidence for soil creep. As shown by Parizek and Woodruff (1957b) curved tree trunks do not necessarily indicate soil creep, in particular if they are not uniformly and continuously curved from base to top. Phipps (1974) has argued even more strongly against the curved tree trunk-soil creep relationship and claimed that curvature and tilting of trunks happen as geotropic and phototropic responses to physical and physiological conditions not related to soil creep.

Observations by the author and other CSIRO workers in rain forest areas show that although there is generally a certain proportion of curved tree trunks present (20–40 per cent on steep slopes), they do not bend consistently downslope nor do they bend continuously from base to top as required if a continuous movement of the upper soil layer is assumed. The bent tree trunks probably reflect sudden dislocations of the trees in the early stages of growth by small-scale slumping or other slope failures rather than continuous movement of soil by creep. Small terracettes are more likely to indicate slow movement of soil, although it is by no means conclusive, as small slumps and mudflows form a similar microlief pattern (Plate 80). Without long-term measurements which would be very difficult to undertake it is impossible to evaluate the role of soil creep in this environment, but undoubtedly it is a relatively minor process of mass movement compared with the more rapid forms discussed above.

SLOPE WASH

Reports on the significance of surficial slope wash under rain forest in Papua New Guinea and other parts of the humid tropics are not consistent. Whereas Behrmann (1927), Sapper (1935) Freise (1936) and Tricart (1972) claim that slope wash is minimal or virtually non-existent because the dense forest canopy, undergrowth, and leaf cover reduce the amount of rain water reaching the ground and also greatly reduce the impact of falling rain drops, McLean (1919) and Vageler (1930) point out that the high silt load of many tropical rivers as well as the presence of young, shallow undeveloped soils on steep forested mountain slopes indicate that surficial slope wash must be of great importance. This latter claim has been supported by Rougerie (1960) and Roose (1967), who measured rates of slope wash of 1.5–3.5 mm per yr and 0.08–0.5 mm per yr respectively in West African rain forest areas.

Before discussing the process of slope wash one must first consider how much of the rainfall actually does reach the ground as throughfall. Freise (1936), who is often quoted and who was one of the first researchers to measure throughfall quantitatively, claimed that only about 33 per cent of the total precipitation reaches the ground as throughfall. Newer data from other rain forest areas indicate that the percentage can be considerably higher and that Freise's results are probably not representative. Measurements range from 41 per cent to 85 per cent but the majority of values are near the
upper end of the range and values between 70 per cent and 80 per cent are most common (Hopkins 1960; Kenworthy 1971; Nye 1961; Odum and others 1970, quoted in Turvey 1974). Turvey (1974) estimates a value of about 80 per cent for his rain forest test site east of Port Moresby.

In Papua New Guinea Ruxton (1967) studied slope wash processes in northern Papua and argued strongly in favour of a high rate of slope wash, while Bik (1967) considered that slope wash did not operate in his study area in the western highlands. The present author has collected further information on this process from other parts of Papua New Guinea, particularly with respect to the variation of slope wash in the different altitudinal zones. The observations generally support Ruxton's (1967) findings for the lower altitudes and Behrmann's (1927) and Bik's (1967) for the higher altitudes. The intensity of slope wash decreases significantly with altitude and changes from an effective surficial wash in the lowland rain forest areas to a subsurface wash of greatly reduced intensity operating below the dense root mat in the lower and upper montane zones.

Kinds of Slope Wash

Slope wash involves two physical processes; firstly the removal of material by the direct impact of rain drops on the soil surface and secondly removal of material by sheet-like flow of water. This second kind of wash is generally referred to as unconcentrated slope wash (Cook 1936; Ruxton 1967). The first kind of wash, that by mechanical impact, though much more efficient in terms of energy (Ekern 1950), is largely inhibited by the ubiquitous ground cover of the forest soils by leaves, twigs, roots, mosses and other organic material. In the lowland forest areas, however, this ground cover is rarely complete and rain splash erosion cannot therefore be ignored. The impact of the rain drops on the soil surface is also thought to be greatly reduced by the canopy leaves which intercept the rain. But it should be kept in mind that a fall of between 7 and 9 m is sufficient for rain drops to approach 95 per cent of their terminal velocity. Furthermore the interception of the rain water by leaves leads to the formation of larger drops, the impact of which even from very low heights is similar if not greater than that of small rain drops falling with maximum velocity.

Unconcentrated slope wash operates much more universally than splash erosion. It begins after intense rain as soon as the rate of infiltration is exceeded by the amount of precipitation. In the wet season this can happen in a matter of minutes, as most of the forest soils in Papua New Guinea are virtually continuously at field capacity (McAlpine and Short 1974). Unconcentrated wash removes not only particles thrown up by splash erosion but also other loose soil particles as well as organic matter such as leaves and twigs.

Freise (1936) pointed out that an important percentage of the precipitation is collected by leaves and branches and eventually finds its way down as stem flow on the main tree trunks. Again his estimates (36 per cent) do not agree with more modern results, which are much lower, ranging from 1 to 18 per cent (Nye 1961; Odum and others 1970). This water causes an increased removal of material
at the downslope side of tree trunks quite evident on older and larger tree trunks. However, it rarely leads to linear incision.

Altitudinal Variation in Slope Wash

The processes of slope wash operate most efficiently in the lowland forest areas that have high canopies, a relatively open undergrowth and usually an incomplete ground cover (Paijmans 1976). Evidence of splash erosion can be found in the form of miniature impact craters on the soil surface, small earth pillars scattered over bare soil patches (Ruxton 1967) and films of silt and sand on leaves close to bare soil patches.

Unconcentrated slope wash leads to undermining of root systems (Plate 81), stowing of material behind logs and other obstacles and so-called root steps (Plate 82). Root steps are the result of temporary protection from slope wash of soil material directly beneath larger roots. Scoured surfaces and accumulations of lag gravel are also common indications of transport of detached soil particles by relatively intensive wash (Ruxton 1967).

In the lowland rain forest areas the cover of the soil with leaf litter varies greatly from place to place. Some trees like *Pometia* and *Castanopsis* shed more leaves than others and consequently the vicinity of these trees is always extensively covered with leaf litter. However, although the leaf litter will inhibit splash erosion, it does not appear to reduce unconcentrated wash, which also operates under the leaf litter, as field observations clearly show. The indications for slope wash mentioned above, in particular the undermining of root systems and the occurrence of root steps, are not restricted to exposed soil patches but can also be observed under leaf litter.

The structure and composition of the rain forest change gradually above about 1000-1400 m. The forest becomes lower and less rich in species, and more important, the ground cover becomes much denser. Here the soil is covered not only by leaves and twigs but by a dense thick mat of surface rootlets often covered with moss and other organic matter (Plates 83, 84). This root mat absorbs like
Plate 82
Root steps are the result of temporary protection of soil surface by roots. Altitude 50 m, slope steepness 10°.

Plate 83
In higher altitudes the ground is covered by a dense and thick mat of roots, mosses and other organic matter. Some undermining is, however, apparent. Altitude 2100 m, slope steepness 25°.
will only slowly be released from this root mat which acts as a temporary water storage.

Closer examination of the root mat and the underlying soil material shows, however, that some removal of soil material from below the root mat must take place. In many instances the root mat is clearly undermined and there is a noticeable hollow space between soil surface and root mat, ranging from a few centimetres to half a metre (Plate 83). This makes walking off the foot tracks very strenuous and hazardous because there is always the danger of unexpectedly breaking through the root mat.

Soil studies also indicate that in spite of the thick organic cover removal of soil material must take place. On steep slopes soils are shallow and poorly developed and on very steep slopes stony lithosols occur, clearly an indication of continuous removal of material. The mechanics of this process and whether the material is transported mainly in solution or suspension are not known.

In the upper montane zone the vegetation becomes more stunted, the undergrowth is denser and the root mats are even thicker than at lower altitudes. In addition, heavy rain showers become rarer and long-lasting light showers and nearly continuous mist and cloud prevail instead, reducing the potential for slope wash. A further reduction in the significance of slope wash is to be expected.

In the high-altitude grasslands that extend above the forest zone slope wash might be expected to be much more active, but this is not the case because the slopes are covered by a layer of peat and peaty soil. This permanently water-logged peat layer serves as a protective cover against slope wash action. Observations during
long-lasting showers at these altitudes have shown that surface run-off is in the form of shallow sheet flow of clear water. The lack of significant slope wash is also shown by the very low content of suspended sediment in these high-altitude creeks and streams even after long periods of rain.

**Acceleration of Slope Wash by Man**

What has been said so far about slope wash applies entirely to the natural or, in the case of the high-altitude grassland, semi-natural environment. Any disturbance by man can dramatically increase the intensity of slope wash in any altitudinal zone (Löffler 1976).

In the lowland forest areas where both rain splash erosion and unconcentrated wash are operative even under primary rain forest the removal of the forest and ground cover will initially greatly increase slope wash, in particular rain splash erosion. However, the traditional practice of shifting cultivation ensures rapid re-vegetation of the cultivated plots, particularly as the removal of the forest is never complete. The garden plots revert quickly to bush and secondary forest and slope wash conditions are soon similar to if not more favourable than those prevailing under primary rain forest.

In marginal rain forest areas with a pronounced dry season the removal of the forest has been more permanent, since in addition to slashing of the garden plots regular burning was practised during the dry season. The rain forest was gradually replaced by grassland and savanna (Plate 3). The burning leads to nearly complete removal of the protective cover and makes the ground highly susceptible to both rain splash erosion and unconcentrated wash. Evidence of these processes can be seen in the form of earth pillars, lag gravel, undermined root systems and elevated grass tussocks. Slope wash for example has been responsible for almost complete removal of the top soil cover on most of the slopes in the Port Moresby area, irrespective of rock type.

In the highlands where the agricultural system has been much more extensive than in the lowlands, rotation of the garden plots...
has also resulted in almost complete removal of the forest and its replacement by grassland. In contrast to the lowland grasslands these are probably due not to marginal climatic conditions, but to the much higher population densities and more intensive use of the land. Here also fires have played an important role (Plate 85). The acceleration of slope wash is locally severe and evidence of this is apparent in old gardens and in unused grassland areas that experience frequent burning. Here, irregularly shaped grass-covered hummocks are enclosed by a network of deep rills often exposing the weathering zone. The big grass tussocks (mainly Miscanthus) firmly enclose the soil with their dense root systems and prevent its complete removal (Plate 86). In between the hummocks, however, up to half a metre of soil has been removed. Evidence of accelerated removal of soil, probably by slope wash, is also given by solution notches at or near the base of limestone boulders and pinnacles. These notches develop under a soil cover but are now 1–1.5 m above ground level (Plate 27).

A very similar pattern of grass-covered hummocks and intervening rills has been observed in high altitude grassland areas on slopes where frequent burning is evident. Burning of the tussocks not only temporarily removes the thick dense grass cover but also destroys much of the protective peat layer. Evidence of this was found on Mt Giluwe and the Saruwaged Ranges following widespread burning after exceptionally dry spells.

How serious the effect of slope wash has been in the long term is difficult to assess. There is little evidence in the stratigraphy of basin sediments of any exceptionally high rates of sedimentation since man moved into these areas, and many very steep slopes show
a surprising stability in spite of long and intensive use (Plate 87). In the Kuk swamp in the upper Wahgi basin, where extensive archaeological excavations are under way, Golson and Hughes (pers. comm.) found that at about 9000 years B.P. the rate of sedimentation increased and its character changed from predominantly organic peat accumulation to alluvial clay deposition, presumably because of man's impact. The increase is, however, modest; while before 9000 years B.P. the average rate of accumulation was of the order of 0.04–0.06 mm per yr, after this date the accumulation increased to 0.07–0.2 mm per yr. Blong (pers. comm.) estimated an average rate of erosion of 1–2 cm per 1000 yr for the Kuk swamp catchment area. Pain (1973) has calculated an average rate of 27 cm per 1000 yr from accumulations in the Kaugel basin. Although this is more than a magnitude higher it is still a slow rate in geomorphological terms and does not suggest any greatly accelerated rate of soil removal. However, none of these calculations takes into account the through-going drainage, which even in the swampy basins could be considerable.

PERIGLACIAL SOLIFLUCTION

Periglacial solifluction, also called gelifluction or congelifluction, is defined as movement of soil particles due to repeated freeze and thaw action. This can either be on an annual climatic cycle, as is high latitudes, or on a diurnal cycle, as in low latitudes. This frost and thaw action causes sorting of fine and coarse material, generally referred to as patterned ground, or simply a downslope

Plate 87
Chimbu Valley, a typical V shaped valley with straight side slopes which in spite of intensive use have maintained a surprising stability. Slope steepness is between 36° and 39°. Rock type is massive shale.
movement of unsorted debris.

In Papua New Guinea solifluction is obviously of very limited occurrence because a diurnal frost and thaw cycle is restricted to the very highest mountain peaks. Although frost occasionally occurs at altitudes of 2000 m and less, causing severe damage to native crops especially if combined with periods of drought (Brown and Powell 1974), regular diurnal frosts are clearly restricted to altitudes of about 4000 m and above.

According to the climatic data collected at the Mt Wilhelm Station of the Australian National University (McVean 1968; Hnatiuk and others, in press) the 0°C mean minimum isotherm is at 4250 m and the 0°C mean isotherm at about 4730 m. Regular night frosts are therefore not to be expected much below about 4000 m.

It is thus not surprising to find only a few and poorly developed solifluction phenomena in Papua New Guinea (Löffler 1975). Terraces with vertical, slightly overhanging faces occur on several high mountains from about 4000 m. They are widespread only above 4200 m, an altitude substantially exceeded only by Mt Wilhelm (4509 m) and Mt Giluwe (4368 m). These terraces are largely formed by the action of needle ice, the development of which seems to be favoured by the excessive water content typical of high-altitude peat soils.

Features of free gelification (unvegetated patterned ground and scree slopes) are restricted to Mt Wilhelm, where they occur from about 4350 m upwards, but even here their extent is very small.

Plate 88
Solifluction scree slopes on Mt Wilhelm at 4400 m
because of the predominance of pure rock faces. The most widespread forms are terraced scree slopes which extend to the east of the main summit. The terraces are about 20 cm high and 30–60 cm deep and they slope between 5° and 20°. The frontal part of the terrace is mostly formed by larger stones which seem to move more slowly than the finer debris and therefore cause some stowing (Plate 88).

In localities where the scree slopes are composed largely of fine material patterned ground is developed. These are small stone polygons and stone nets on flattish ground and stone stripes where the slope exceeds about 8° (Plate 89). The diameter of the polygons and distance between the stone stripes are of the order of 10–20 cm. The sorting of the patterns is only superficial; larger stones are usually not incorporated in the sorting.

In spite of the insignificance of solifluction as an erosional process the determination of its altitudinal limit is important for regional comparison, particularly with other tropical areas (Löffler 1975). These comparisons showed that the humid quasionceanic mountains of Papua New Guinea have a higher limit of periglacial activity and free gelification but a lower snowline than mountains in East Africa at comparable latitudes. This compression of the periglacial zone is thought to be due to the uniformity of the Papua New Guinea mountain climate with its lack of annual and particularly diurnal temperature extremes.

Plate 89
Patterned ground due to diurnal frost cycle on Mt Wilhelm at 4400 m. Sorting is only superficial.
SUBCUTANEOUS EROSION
Subcutaneous erosion has already been mentioned in connection with karst. In karst it is primarily due to the solubility of the rocks, and the removal of material is largely in solution on an ion by ion basis. This is not to deny that mechanical forces are also at work and larger cave systems especially owe much of their extent to mechanical erosion by underground river systems. Karst erosion is not the only form of subcutaneous erosion, although it is undoubtedly the most prominent and most widespread. A generally little known process of subcutaneous erosion is piping or tunnelling which is a process that forms karst-like conduits in insoluble clastic rocks. Here material is removed as suspended solid particles.

Subcutaneous Erosion in Limestone
Subcutaneous erosion is most common in limestone areas and is largely responsible for the very specific landforms that develop on this rock type. Subsurface drainage systems have been studied for decades in many parts of the world, particularly the ‘classical’ karst areas of Yugoslavia, but in Papua New Guinea they have remained a closed book until quite recently.

Williams (1972b) first reported the existence of a number of caves and remarked on the great caving potential of the karst areas of Papua New Guinea. Since then documentation of caves has become much more readily available thanks to the establishment of the *Ningini Caver*, a newsletter of the Papua New Guinea cave exploration group, in which new discoveries are reported and discussed. For instance its first issue contains a report on the Bibima Cave in the Chimbu area which at 494 m in depth is the deepest cave in the southern hemisphere (Wilde and Watson 1973). Another important document on caves and caving is the report of the Papua New Guinea Speleological Expedition NSRE 1974, which describes a number of caves, sinks and underground rivers from the south-western section of the Müller Range. Most of the caves visited appeared to be fault or joint-controlled; bedding had little influence on cave formation. They seem to have been formed predominantly by underground river systems and could be classified as vadose caves. Collapse dolines were also prominent.

Although many of the caves visited did not meet the high (or more precisely low) expectations of the cavers whose aim was to find the ‘deepest hole in the world’, two caves of over 300 m were discovered and there is undoubtedly the possibility of considerably greater depths.

Much more modest in size and depth are the caves on the Trobriand Islands, which are the best researched and documented, especially from the geomorphological point of view.

Ollier (1975) has grouped the caves into six descriptive classes: straight caves, collapsed tunnels, cenotes, phreatic caves, sea caves and miscellaneous caves. Most of these were initiated at or below the water table as water table caves or phreatic caves and subsequently modified by collapse. Straight caves and sink holes, however, are regarded by Ollier as having formed in the vadose zone, and sea caves, which are always small and poorly developed,
by marine erosion. The important general conclusion Ollier draws from his observations is that several processes of cave formation, namely water table solution, phreatic solution, vadose solution and marine erosion, can be operative in the same small area and that the different theories of cave formation should therefore be regarded not as incompatible but as alternatives each applying to certain conditions.

**Subcutaneous Erosion in Ultrabasic Rocks**

Ultrabasic rocks are similar to limestone in respect of solubility. Instead of calcium it is the magnesium which represents the soluble element, but since ultrabasic rocks contain only about 50 per cent of magnesium their solubility is not as great as that of limestone, where the calcium content often exceeds 90 per cent. Subcutaneous erosion on ultrabasics is therefore not as impressive and widespread as on limestone.

Nevertheless observations in the Papua New Guinea ultrabasic belt show that it does produce closed basins and subsurface pipes (Plate 90). Chemical analysis of rock and soil show that removal of magnesium in solution must be the prime cause of subsurface erosion (Löffler in press). Analysis of water running over ultrabasic rocks showed that there is no significant difference in the amount of magnesium removed from ultrabasic and non-ultrabasic areas (Table 2). The reason for this must be that the water collected had little contact with fresh rock but merely had moved over and through the leached soil and weathering mantle.

![Plate 90](image)

Closed basin with sink hole in ultrabasic rocks in the Lake Trist area. Altitude 2000 m.

**Subcutaneous Erosion in Non-soluble Rocks (Piping)**

Piping has been known to engineers and soil scientists for some
time as an important process by which water from a reservoir percolates through the foundation of a dam, causing the formation of tunnel-like passages and eventually leading to the collapse of the structure. A similar process can also occur in natural environments and is here also referred to as piping or tunnelling. Piping is not restricted to arid and semi-arid environments or the presence of particular clay minerals (Parker 1963), but occurs over a wide range of climatic and soil conditions (Löffler 1974c).

In Papua New Guinea piping has been studied by the author in the southern lowlands where it is relatively widespread. Pipes occur typically at or near the edges of plateau remnants, on terraces and on moderately steep slopes. The discharge ends of the pipes, which are up to 30 cm in diameter, are usually situated in gully headwalls or sidewalls (Plate 91). A shallow surface depression leads upslope from the pipe to one or several closed depressions which appear to be interconnected (Fig. 48).

The sediments in which piping occurred were all young, weakly consolidated deposits of mainly fine grain (60 per cent clay, 40 per
cent silt). The dominant clay minerals were halloysite and disordered kaolin.

The principal conditions for the development of piping in Papua New Guinea appeared to be a steep hydraulic gradient, the presence of weakly consolidated sediments that disperse readily, and a certain degree of permeability of the sediments. Permeability is probably aided by biotic activity. The significance of piping as a geomorphic process is in the initiation, backward cutting and undercutting of gullies and thus in the acceleration of general headward retreat.

The widespread occurrence of active piping in the tropical rain forest is a clear indication that contrary to the views expressed by Tricart (1974) mechanical erosion and linear incision can take place and are important processes in this environment.

DEPOSITIONAL PROCESSES

The processes of deposition have already been described in Chapter 3 and a short summary will therefore be sufficient. There is a much closer link between depositional landforms and depositional processes than between erosional landforms and processes. By far the most common form of deposition in Papua New Guinea is fluvial deposition in the form of bar and overbank deposition and lateral accretion. Bar deposition is characteristic of braiding streams with high gradients, while overbank deposition is typical of low-gradient streams in confined levee-bound channels. Meander streams with medium gradients are characterised by lateral accretion, bar deposition and overbank deposition. Deposition by braiding streams usually leads to the formation of fans, while deposition by meander and levee streams results in alluvial plain formation. The distinction is not a clear cut one and there are many transitional examples.

In the coastal areas the fluvial depositional processes interact with and are replaced by littoral processes, which are dominated by the tidal regime and longshore drift. Deposition in estuaries is by overbank flooding due to the slow, quiet advance and retreat.
of the tide. Biotic activity in the form of mangrove colonisation and crab mound-building facilitates the accretion of suspended material. At the seaward side of tidal environments strong longshore currents are common and are capable of transporting large quantities of relatively coarse sediment. This is deposited as sand bars, sand barriers and sand spits. These features greatly aid the seaward extension of the land as 'pioneer landforms'.

Coral reef growth is a special form of deposition. Strictly speaking it is not a land-forming process as it is active only under water, but slight changes in the relative position of land and sea will transform the reefs to landforms. Large areas of Papua New Guinea are in fact uplifted coral reefs.

**VOLCANISM AND SEISMIC ACTIVITY**

It is one of the characteristics of the circum-Pacific Mobile Belt that the very young volcanic processes and seismic activity have a great impact on the distribution of landforms and landform development in general.

Volcanic and seismic processes are geomorphological 'accidents' and are therefore difficult to fit into any concept that demands a relatively uniform and continuous course of the geomorphic processes.

The impact of volcanic processes has not been restricted to the formation of volcanoes; lava flows, ash flows, lahars, volcano-alluvial fans and above all air-borne ash falls have extended far beyond the actual volcanoes (Plate 92). In the Yuat Valley for instance lahar or nuée ardente deposits have been found 130 km from the Mt Hagen volcano where they originated (Dow and others 1972). Besides the creation of new landforms volcanism has severely disrupted existing drainage conditions, as has been observed on the Wahgi, Kaugel, Lai and Hegigio River systems. Ash falls have masked large areas causing not only disruption of plant life but also altering the properties of the soil and possibly changing the slope processes at least temporarily. In the Kaugel Valley for example Pain (1973) claims that the addition of thick layers of rapidly weathering ash

Plate 92
Lahars and airborne ash in the Nembi Valley near Porema. These deposits have extended far beyond their source area which was probably Mt Giluwe
Processes associated with volcanism are mudflows (lahars) that have a considerable capability to move large quantities of debris including enormous blocks of solid rock. This is largely due to their specific gravity which can reach values of up to 2. Large areas in the vicinity of Mt Hagen and Doma Peaks have been formed by these deposits (Plate 93).

Seismic activity in Papua New Guinea is very high, totalling 5–10 per cent of the world’s earthquake occurrences (Brooks 1965). Papua New Guinea is thus one of the most active seismic areas of the world and most of its land areas can expect between one and ten earthquakes of magnitude 6 or larger per square degree per century (Fig. 49) (Brooks 1965; Denham 1969). The islands of the Bismarck Archipelago and some parts of the northern ranges can expect more severe and more frequent earthquakes.

The effects of an earthquake on landforms and geomorphic processes in the Toricelli Mountains, where a severe earthquake increased the rate of slumping.

Plate 93
Extensive lahars from Mt Hagen volcano (background) have filled formerly V shaped valley and now form extensive areas of flat and gently sloping land. Limestone escarpment to the left is to the south-west of the Kubor Anticline and can be regarded as a southern counterpart to the Porol or Elimbari Escarpment to the north of the anticline.

Figure 49a
Distribution of earthquakes with magnitudes $\geq 5$, and $\geq 6$ between 1958–1966 (after Denham 1969) (reprinted from *Journal of Geophysical Research* by permission of the American Geophysical Union)

- Depth $\geq 70$ km
- Depth $< 70$ km
took place in 1935, have been described by Stanley and others (1935). 'Soil, subsoil with their covering tropical jungle had disappeared from 60 per cent of the slopes, baring the underlying bedrock. Below, valleys previously of sharp V-form were aggraded to heights of 50–60 feet above the old floor'. More recently Simonett (1967) has undertaken a quantitative study of the effects of this earthquake with the aid of aerial photographic analysis of photographs taken in 1938. He concluded that in the Toricelli Mountains, where the epicentre was situated, 2,526,700 m³ of debris were removed per km². The effect of the earthquake decreases with distance from the epicentre and becomes insignificant at a distance of about 80 km. The effect of the earthshock was much more severe in the granitic basement area than in the overlying sediments. In granitic areas the loss was equivalent to the removal of a layer of 40 cm of debris over the planimetric area affected by the earthquake while it amounted to 'only' 12 cm on weak sedimentary rock and 7.4 cm in mixed sedimentary rock (Simonett 1967).

The effects of a more recent earthquake in the Madang area have been investigated by Pain and Bowler (1973) and Pain (1972b). The earthquake was of magnitude 7.0 and its epicentre was situated some 32 km north of Madang. Dense landsliding of the debris avalanche type was the major form of slope failure. Pain (1972b) considers it likely that tree fall triggered off by the earthquake was a major contributor to the initiation of landsliding, as sliding was more frequent on formerly forested slopes than on garden land. The denudation of 11.5 cm calculated by Pain and Bowler (1973) agrees well with the figures given by Simonett (1967) for sedimentary rocks.

Earthquake-induced mass movements must be regarded as major factors in landform development in Papua New Guinea.

**BIOLOGICAL ACTIVITY AS A GEOMORPHIC AGENT**

Organisms generally play a minor role in land-forming processes
except in environments such as coral reefs which are almost entirely built up of organic material, or tidal flats where the seaward extension is encouraged by silt trapping by mangrove vegetation. Organisms are also important in weathering and solution processes (Ollier 1969b; Jennings 1972). These processes are not considered here since they have already been discussed in their relevant context. In the following section some microrelief features resulting from the activities of small animals are discussed.

**Earthworms**

The recognition that small animals like earthworms can in time transport large quantities of soil is not new, and has in fact already been described by Darwin (1881). In Papua New Guinea the activity of earthworms has been thought to account for the formation of a peculiar microrelief in the Sepik area (Haantjens 1965). This microrelief consists of stone and gravel stripes and pitted soils. The stone stripes occur only on slopes above about 5° and consist of poorly sorted gravel stripes up to 1 m wide separated by earth rises 30–40 cm high and 1–3 m wide. According to Haantjens (1965) the stripes are aligned at about 25° to the maximum slope but this could not be confirmed by the author, who found that the preferential orientation is downslope at maximum slope angle (Plate 94). On ridge crests the stripes are replaced by stone rings surrounding small wormcast hills (Plate 95) or vice versa, wormcasts surrounding a central stone concentration. The pattern has undoubtedly a great resemblance to patterned ground observed in periglacial environments, though it lacks the regularity.

Plate 94

Stone stripe pattern in the Sepik area near Roma
The second type of microrelief consists of pitted soils which never occur in association with the stripes but are restricted to flat and gently sloping ground with impeded drainage. The pits are very irregular and vary from roundish holes to narrow elongated trenches. They vary in depth from about 0.15 to 0.9 m. Although the pits can be very long they always form isolated depressions and are not interconnected (Plate 96).

Haantjens’s (1965) detailed soil investigation shows that processes of physical soil deformation cannot explain these features because of the absence of swelling clays and of any severe water stress. There is also no evidence for erosional processes such as deflation, piping and other forms of subsurface erosion. The abundance of large earthworms in the pitted soils and on the intergravel rises suggests that the activity of these animals is largely responsible for the microrelief. The pitted soils are caused by continuous removal of soil from depressions and its accumulation on the rises. The stone stripes are explained by initial gravel concentration due to weathering and erosion, followed by deformation by rheotropic flow and swelling and shrinking. Earthworms then act on this partially segregated gravel to produce the striking gravel stripes and earth mounds. Surface or rill wash are not considered by Haantjens as important factors in the formation of the stripes. This does not agree with the present author’s observations, which suggest that gravitational forces must play an important role in the differentiation of the stone stripes. This is indicated by the lack of stripe patterns on flat to gently sloping ridge crests. Instead roughly circular forms are present. These become more elongated and eventually form stripes with increasing slope angle. Further there is

Plate 95
Worm cast hills surrounded by gravel on ridge crest
some sorting of gravel with finer gravel preferentially on the outer edges of the stripes and coarser material in the centres. The author's suggestion therefore is that the stone-striped pattern is due to the combination of worm activity, leading to an initial separation of coarse gravel and fine material, with slope and rill wash causing a downslope orientation and rough sorting of the material.

Lee (1967) rejects the hypothesis that earthworm activity is responsible for the microrelief and instead suggests that the microrelief is primarily a result of burning of grass tussocks and associated deflation of the burnt patches. The microrelief simply provides a suitable habitat for the earthworms and their association with it is interpreted by Lee as an ecological adaptation for the survival of a previously forest-dwelling species in a new grassland environment. While there seems to be little doubt that the microrelief is of ecological significance to the worms, it is difficult to understand why they should not be actively involved in the creation of this suitable environment, especially since their activity is so obvious (Plate 95). Lee's (1967) explanation of the microrelief is unconvincing because of the lack of strong winds in this area, the lack of any directional pattern of the pitted soils, because there are no severe soil moisture deficiencies and the microrelief occurs under forest (Haantjens 1969). Also the pitted soil holes are much too deep and their rims too sharp and steep to be caused by wind action (Plate 96). Similarly the stripes are directed downslope and not aligned with any prevailing wind direction. It perhaps deserves mention, though not as a valid argument, that the microrelief is also attributed to worm activity by the local natives.
Wormcasts of much more sporadic occurrence are present in lowland rain forest areas, usually on gentle to moderately steep slopes with well-developed fine-textured soils. Here small pillars, 15-20 cm high and 5 cm wide, are irregularly scattered over the forest floor, protruding through the leaf litter (Plate 97). They do not form any distinct pattern but occur as isolated individuals. The fact that they raise soil above the leaf litter and make it thus more susceptible to rain splash activity seems to be of geomorphic significance.

**Termites**

The activity of termites is similar to that of earthworms. They also work through the soil shifting soil particles from lower soil horizons to the surface. While earthworms require a considerable amount of soil moisture for their ploughing activities and are therefore restricted to humid areas, termites occur predominantly in dry or seasonally dry areas. In Papua New Guinea, they are common only in the lower Fly Platform south of the Fly River and the Port Moresby area, which have a tropical savanna climate with a pronounced dry season.

Termite activity leads to the formation of termite mounds measuring up to 4 m in height and 2 m in width. The contribution of termites to slope and soil formation has not been studied in Papua New Guinea but investigations in northern Australia (Williams 1968) show that termites are capable of moving large amounts of soil. The annual rate of earth movement by termites was calculated to be 0.48 m$^3$ per ha per yr.

**Plate 97**

Small worm casts protruding through leaf litter make soil more susceptible to rain splash erosion. Site, lowland rain forest near Kerema, 150 m.
The significant contribution of termite activity lies in the transport of fine soil particles from depth to the surface, thus counteracting the removal of soil by surface wash.
The pre-Pleistocene history of the Papua New Guinea landforms is intimately linked with its geological history. The framework of the present landforms, in fact the entire shape of the present land mass, did not become apparent until the late Miocene and early Pliocene. During most of the Miocene much of what are now high mountain ranges were still covered by shallow sea or represented fluvial depositional plains. In the New Guinea Mobile Belt active island arc volcanism accompanied the tectonic and igneous activity which had taken place there since the middle Miocene. These volcanics supplied much of the clastic sediments that accumulated to the south of it, especially in the Aure Trough. In the east the Owen Stanley Ranges probably formed land areas from the middle Miocene on. In the Pliocene sedimentation from the rising land mass was still active around the margins of what are now the central ranges, and in particular in the Aure Trough sedimentation of clastic material continued well into the Pliocene. In the east in the Owen Stanley Ranges a landscape of hilly relief with rounded slopes and broad valleys developed relatively close to sea level. Remnants of this landscape are still to be found as summit surfaces all along the summit of the Owen Stanleys and possibly continuing into the Saruwaged summit plateau. Further west in the highlands remnants of former low relief landforms can also be found, but their irregular distribution makes any correlation impossible. Large areas of the highlands, however, have the character of relict landscapes with subdued relief, deep weathering profiles and rivers with relatively gentle gradients. The gradients steepen considerably where the rivers leave the highlands and plunge down into the lowlands.

The Pliocene landscape is therefore considered to have been hilly and probably mountainous in places, but it did not reach great altitudes.

During the Plio-Pleistocene this land surface was uplifted during a major phase of vertical earth movements. The rate of uplift was not uniform. In the Owen Stanley Ranges uplift of the Pliocene land surface ranged from about 200 m at its eastern end to over 3000 m in the Mt Albert Edward area. In the highlands much differential movement took place along major faultlines and as yet it has not proved possible to reconstruct the movements in any detail. The lack of any obvious stepped surfaces on the Owen Stanleys indicates that the uplift probably proceeded in a more or less continuous movement without major interruptions or pauses.

Accompanying or closely following the uplift there was extensive volcanism leading to the formation of the highland volcanoes. There is no information on the approximate date of the start of
volcanic activity. The oldest dates are 1.1 million years from Kara Plug south-west of Ialibu (Bain and others 1975) and 0.85 million years from Mt Iume north-west of Tari (Williams and others 1972), indicating that by mid-Pleistocene volcanoes must have been in existence for some time. The volcanoes probably formed at or close to their present altitudes since they erupted onto a deeply dissected mountainous landscape with a drainage system very similar to the present one. A number of drainage systems were disrupted by the volcanic activity leading to the formation of several intramontane basins. Some, like the Wahgi, Kaugel and Haibuga basins, were blocked by lava flows or lahars and subsequently filled in with lacustrine sediments, mudflows and fluvial deposits. Others, like the Lai, Nebilyer, Bayer and Nembi valleys, were simply filled up with volcanic material, mainly ash flows, lahars and nuées ardentes. Not all intramontane basins and filled in valleys are, however, the result of volcanic events (Fig. 39). The Goroka and Aijura basins were caused by tectonic movements resulting in temporary ponding of the drainage system, formation of extensive lakes and gradual filling in of the lake basins with lacustrine deposits.

Other basins are of structural origin, such as several largely swampy basins in the limestone belt. Lake Kutubu is also situated in a structural basin but its ponding is due to a lava flow at its southern end (Bayly and others 1970).

Finally, there are a number of basins the origin of which is uncertain. Some of these like the Neon basin west of Mt Albert Edward (Plate 21) and the Myola Lakes occur within a relict surface, and may have formed before the main uplift when the area was close to sea level. A similar origin has been suggested for the Kandep basin in the Mendi area, although no relict surface is apparent there (Jennings 1963).

Bik (1967) has pointed out that some of these intramontane basins in the Mendi area superficially resemble hill-bordered saucer-shaped valleys (Flachmuldentäler mit Rahmehöhen) as described by Louis (1968), and suggested that they may therefore be erosional in origin. He stresses, however, that he has not been able to prove their erosional nature, nor are the present-day processes compatible with the Flachmulden concept. Blong and Pain (in press) have strongly rejected this and regard the Flachmulden concept as irrelevant for the intramontane basins, since all the valleys they examined (particularly the Wahgi, Kaugel and Nembi valleys) were infilled by up to several hundred metres with volcanic deposits and lacustrine, alluvial and organic material.

As described above a great number of intramontane basins have undoubtedly developed as a result of infilling associated with the volcanic activity in the highlands. However, the origin of the basins is much more diverse than envisaged by Bik (1967) and Blong and Pain (in press) and no simple explanation is adequate; each basin may have its own geomorphic history which need not necessarily coincide with the history of other basins.

All the basins are undoubtedly of considerable age although little is known about their absolute ages or their relative age relationships. For the basins associated with volcanic activity the ages of the
volcanoes and lava flows that caused the ponding will give some idea of age. The best date is that from the Haibuga basin which was ponded by lava flows that erupted about 0.85 million years ago (Williams and others 1972). This gives a relatively precise date for the start of the basin formation. In the case of the Wahgi and Kaugel valleys which are associated with Mt Hagen and Mt Giluwe a number of lava flows indicate that by about 200,000 years B.P. the volcanoes had had their last major eruptions. These lava flows are not directly relevant to the pondings and the basins may have formed at a much earlier stage of development of the volcanoes. These dates are therefore not really meaningful. The advanced stage of weathering of many of the deposits filling the valleys also indicates considerable age.

Filling of the basins has in most cases ceased or is restricted as in the Ka Valley (Jennings 1963) and upper Wahgi Valley to minor over-bank deposition and organic accumulation of peat in swampy areas. The much more active deposition in the past, particularly the extensive fan formation, was probably a result of the greater tectonic instability during this time.

In the mountain forelands extensive alluvial deposits were accumulated by the rivers draining the mountains. Deposition was most active where the streams flowed onto a stable foreland such as the Fly Platform. There is no information on the exact nature of these processes but their distribution and morphological appearance indicate that they were similar to the present-day fluvial processes, with deposition varying from braiding fan deposition in the steeper reaches of the plains through bar deposition with intensive meander formation to over-bank deposition in the lowest reaches.

The landform development in the high mountains has been significantly influenced by glacial activity. There is now growing evidence that Papua New Guinea mountains experienced more than one glacial period (Blake and Löffler 1971; Williams and others 1972). On Mt Giluwe volcanic eruptions alternated with periods of ice cover, probably as long ago as 300,000 years B.P., and in the Haibuga basin pollen from samples that are stratigraphically well below the sample that was dated at 32,700 yrs B.P. indicate relatively cold conditions well before this date. The notion that the New Guinea mountains escaped earlier glaciations because they were not high enough (Verstappen 1964b) cannot be upheld, at least for this part of New Guinea. However, it is only the last glaciation which we know in some detail. Important contributions to knowledge of the environmental conditions during and after the last glaciation and up to the present have been provided by pollen analytical studies (Flenley 1967; Walker 1970; Powell 1970; Hope 1973).

During the last glaciation the snowline was at about 3550 m with minor variations due to small climatic differences, and all mountain peaks exceeding this altitude were ice-covered. Glacial erosion was considerable and its traces are still clearly visible today. The temperature depression in the mountain areas has been estimated at about 5°–6°C. In the lowlands reduction in temperature was probably less (3°–4°C) because of the likelihood of an increased
lapse rate due to cooler, drier air masses (Nix and Kalma 1972). This temperature depression resulted in a general depression of vegetation belts and climatic geomorphological zones with the amount of depression somewhat decreasing with decreasing altitude. There is little geomorphological evidence that this depression had any effect on the landform development in the periglacial altitudinal zone below the ice-covered area. From the limited observations available it seems that no significant periglacial activity took place here. In fact all the evidence for Pleistocene periglacial solifluction has been found inside the glaciated area and it must have occurred shortly after the ice had receded. This observation is not surprising if one looks at the present-day distribution of the periglacial zone. Periglacial solifluction is an important process only from about 4350 m upwards. This means that the periglacial altitudinal zone is compressed to a mere 200–300 m, which is seen as a result of the great uniformity of the climate. During the Pleistocene the ice extended well below the snowline, especially where larger ice caps were developed, and covered parts of the area that was climatically suitable for periglacial activity, i.e. the zone of diurnal frost and thaw action.

Extensive grasslands covered much of the area below the ice and here relatively little erosion would have taken place if one takes the processes in the present grasslands areas of the high mountains as a yardstick. Powell and Hope (1976) claim that the grassland communities extended down to 1900–2100 m in the period between 38,000 and 30,000 years B.P. and to 2000–2300 m between 30,000 and 12,000 years B.P. This enormous lowering of the grassland-forest boundary of nearly 2000 m from the present one is surprising and is difficult to understand. To explain this discrepancy would require quite considerable climatic changes in addition to the temperature depression. In particular it would demand considerably lower precipitation and perhaps a more ‘extreme’ diurnal temperature range, possibly similar to the present-day conditions on Mt Kenya where the vertical distance between treeline and snowline is in fact in the order of about 1500 m (Hastenrath 1973). There are, however, arguments against both assumptions. Firstly, the relatively low Pleistocene snowline – the lowest tropical snowline in the world – requires a high amount of precipitation, similar in magnitude to the present level. Any substantial reduction in precipitation would have resulted in a higher snowline. Secondly, in spite of the lowering of the sea level by 130 m or so, Papua New Guinea would have remained essentially a highly oceanic climatic province with minimal diurnal temperature extremes, being surrounded by deep sea except for the Torres Strait area. In Torres Strait, climatic conditions most certainly changed to drier conditions but this would not have affected the high mountains (Nix and Kalma 1972).

At about 12,000 years B.P. the ice receded rather rapidly and by about 9000 B.P. all the Papua New Guinea mountains were ice-free. This has been established from $^{14}C$ dates on Mt Wilhelm (Hope 1973) and on Mt Giluwe (Hope and Löffler unpubl.). Surprisingly neither glacial activity nor recession of the ice seem to
have resulted in any large-scale erosional or depositional activity as previously thought (Blake and Löffler 1971), since the stratigraphy of basins in the vicinity of glacially covered areas has not shown evidence of dramatic changes in the depositional regime during this time (Pain 1973). This could be due to the diurnal regime of melting typical of equatorial glaciers with only low discharges during daytime and thus a relatively low ability to transport material supplied by the glaciers. The formation of extensive well-developed lateral and terminal moraines by tropical glaciers is explained in the same way (Tricart and others 1962; Tricart 1971).

The period from about 8500 B.P. to the present is characterised by some minor climatic changes and, more important, by the onset of man's impact on the land and his widespread destruction of the forest, which is evidenced in the pollen record of some sites in the highlands and high mountains (Powell 1970; Hope 1973), and in the stratigraphy of the Kuk Swamp in the Wahgi basin where sedimentation changes from predominantly organic to inorganic at about 9000 years B.P. (J. Golson and P. Hughes pers. comm.). In the high mountains repeated evidence of burning activities is to be found at about 4000 years B.P., resulting in temporary slope instability (Blake and Löffler 1971).

Moving away from the high mountains to the coastal areas the record of the past geomorphic history becomes more vague. Observations by Mabbutt and Scott (1965) in the Port Moresby area suggest periodicity in soil development in what is now one of the driest areas in Papua New Guinea. These changes are interpreted as results of periodic climatic changes from humid to drier conditions. An alternative explanation could be that the slope instability reflects man's impact on the land which in this area has been particularly severe, leading to nearly complete removal of the rain forest and its replacement by grassland and savanna.

Finally the role of the lowering of sea level during the Pleistocene has to be considered in relation to present-day landforms. The worldwide repeated lowering of the sea level by some 130 m or more caused major changes in the distribution of land and sea only in Torres Strait where the Fly Platform was joined to the Australian continent. The closure of Torres Strait was not only of geographical and biological significance (Walker 1972) but also had profound effects on the climate of the area, as it caused widespread desiccation (Nix and Kalma 1972). This was not restricted to northern Australia and the exposed Arafura Plain but extended well into the Fly Platform, possibly to a line running along the Aramia River to the Fly-Strickland junction. This is suggested by the distribution of pisolithic iron concretions in the soil profiles of this area.

Along most of the other coastlines the drop in sea level did not result in any significant land gains because of the steep offshore slope; but the coastal streams and rivers had to adjust to the much lower base level of erosion. Smaller streams with steep gradients were little affected because it did not result in a significant increase in gradient, but larger rivers like the Sepik and most of the large rivers flowing into the Gulf of Papua were profoundly affected, as their gradient increased greatly.
This resulted in incision into older alluvium, as in the Fly Platform, or in the formation of deeply incised embayments in areas where an extensive shelf was missing. The consequences must have been most substantial in the Sepik since the gradients increased dramatically and the entire plain seems to have been lowered to the new base level, or close to it. Rivers with considerably steeper gradients such as the Markham experienced increases of gradients of a lower order, and probably no major changes in the river regime resulted. The subsequent rise in sea level led to drowning of most of the coastline and the rapid filling of the embayments and other low-lying alluvial plains with sediment and organic matter. Again the rise hardly affected coastlines with steep offshore slopes as well as rising coastlines. In the Markham Valley it could have led to an increase in the rate of aggradation.

The geomorphological history of Papua New Guinea is dominated by young tectonic processes which have been responsible for the entire framework of the present landforms. It is impossible to reconstruct the exact history of these processes because they do not operate concurrently and there is no stratigraphic basis for correlating tectonic events over a large area. Geomorphic processes have sculptured and modified this structural framework and the detail of the present landforms is the result of these processes. In contrast with many other parts of the world there is on the whole very little evidence for climatically controlled major breaks or changes in the character and kind of these processes, except at high altitudes where periods of glaciation are evident. However, the intensity and the rate of processes has clearly changed. In the past the rate of processes in the central ranges and their foreland must have been considerably higher than it is today. This not only applies to the obvious case of volcanic processes but also to those of fluvial erosion and mass movement. The reason for this must be sought in the greater tectonic instability of this area in the past. There is no evidence that drastic climatic changes could have been responsible for this increased rate of geomorphic processes. At present geomorphic processes are most active in the Saruwaged and Finisterre Ranges and their foreland.
6 Geomorphic Concepts and Landform Development in Papua New Guinea

In attempting to discuss the development of landforms in the framework of geomorphic concepts one must be aware that these concepts only serve as abstractions helping to organise and generalise the extreme complexity of landforms. Geomorphology is not an exact science because it offers little scope for reproducible experiment and lacks laws capable of expression in exact mathematical terms. Landforms are also not living organisms that can be reproduced, and their evolution cannot be traced through a short life span. Landforms are passive matter upon which a multitude of interrelated and independent forces act through time. These forces, though subject to the laws of physics and chemistry, cannot be described overall by abstract mathematical formulas, and neither can the forms they create. This is in spite of the fact that a great number of well-understood physical and chemical processes operate on the land surface and that certain relationships can be expressed in generalised mathematical equations. However, it is the interaction of a multitude of these processes and above all the historical aspect of geomorphology with often unpredictable changes of both exogenic and endogenic processes that present the major obstacles to exact scientific expression. Absolute proof is rarely possible and painful as it may be to many of us in most cases the geomorphologist has to content himself with probabilities.

As recently pointed out by Pitty (1971) and even more specifically by Jennings (1973) each theoretical concept and approach has to be seen and judged in its own right and historical perspective, and each concept has merits and shortcomings because it is ultimately an oversimplification. One has to beware of too rigid an application of theoretical concepts which will tend to limit the flexibility of the approach. As Tricart and Cailleux (1972) express it in their criticism of the Davisian concept 'Hypotheses should open up perspectives, not shut them' and the 'naturalist must patiently investigate the earth step by step in order to progress in the knowledge of its intimate nature' (Tricart and Cailleux 1972: 43).

It is becoming increasingly evident that in many cases apparently conflicting and incompatible concepts of landform development are in fact merely alternative or complementary explanations, and that one of the most frequent errors has been the claim of researchers of universal application of their concepts. The development of landforms in Papua New Guinea can be seen from several conceptual angles.
STRUCTURAL GEOMORPHOLOGY

The structural geomorphologist will find close harmony between landforms and structure in many areas, and as repeatedly pointed out in the preceding chapters the entire framework of the landforms is structurally dominated. A structural approach to the landforms of Papua New Guinea is therefore profitable and this has been well demonstrated by Carey's (1938) classical study. Structural influence is not restricted to the gross distribution of landforms but also accounts for considerable detail as seen in the landforms with prominent structural control. The southern fold mountains are dominated by structural landforms and in many other areas structural control is quite apparent. The influence of structure and the survival of structural landforms varies with rock type and the greater the resistance of the rocks to weathering and denudation the more pronounced structural control will be. The appreciation and understanding of structure is therefore very important but it is basically a passive factor and cannot entirely explain the complexity of the landforms.

Here the aspects of geomorphic processes come into play. The significance of geomorphic processes and their close relationship with climatic and biogeographical zones has been stressed by proponents of a climatic geomorphological interpretation of the landforms (Louis 1968; Bödel 1969, 1972; Tricart and Cailleux 1972). Dynamic geomorphologists also see process studies as the centre of landform development but tend to neglect the historical dimension.

CLIMATIC GEOMORPHOLOGY

In Papua New Guinea an explanation of the landforms in morphoclimatic terms has been attempted by Bik (1967). He distinguishes four different altitudinal zones with characteristic landforms and processes. The delineation of the altitudinal zones is largely based on vegetation characteristics, namely the lower montane forest zone, the montane forest zone, the natural grassland zone and finally the Pleistocene glacial zone. This classification is not consistent as the first three subdivisions are based on present-day processes and the upper zone on processes that have taken place in the past. It is also difficult to separate a particular natural grassland zone as very little of the high-altitude grasslands is natural, in fact natural grassland is restricted to altitudes above about 3900 m; below this altitude it occurs only in small scattered nuclei on topographically suitable terrain such as waterlogged glacial valley floors and basins (Paijmans and Löfler 1972). Furthermore the altitudinal extent of the grasslands has varied considerably during the Pleistocene and the present forest-grassland boundary is only a temporary one. If it had had geomorphological significance traces of its former greater extent should be evident at lower altitudes.

As far as this was possible from qualitative observations major differences in geomorphic processes with increasing altitude could not be established. There is a decrease in the rate of weathering and a change in the kind of weathering with altitude but this is often
offset by the occurrence of relict weathering profiles at higher altitudes. There is also a significant reduction in the erosive capacity of slope wash with altitude and the survival of deep weathering on steeper slopes at high altitudes may well be due to this factor. But it could not be established that the landforms as such differ to any noticeable degree. In fact all altitudinal zones are characterised by typical ridge and V valley landforms with V-shaped valleys, steep side slopes and sharp crests formed by fluvial erosion and associated mass movements leading a parallel slope retreat with maintenance of steep slopes. Differences do occur but are related to rock type and evolutionary history much more than to altitudinal change.

There are, however, two zones besides the relict glacial landforms that appear to have landforms which warrant a morphoclimatic interpretation. The first is the periglacial zone which extends above about 4000 m. It is very limited and is restricted to the two highest mountain peaks, Mt Wilhelm and Mt Giluwe. Landforms here are characterised by smooth though steeply sloping often terraced scree slopes. Much of this zone is dominated by bare rock faces, which were formed during the Pleistocene glaciation.

The second is a zone of (relict?) planation which extends in the seasonally dry areas along the south coast. Here the predominant process is not linear fluvial incision and parallel slope retreat, but scarp retreat by the process of pedimentation guided by erosion processes that are dominated by surface wash (sheet wash) on the pediment and rill wash on the hill slope. Present-day processes do not maintain the process of planation but cause dissection (Mabbutt and Scott 1965).

The process of planation by sheet wash is still in operation in the southern part of the Fly Platform. Here on a relict alluvial plain rising between 10 and 50 m above sea level incision is not very prominent; surface wash is the dominant process. The area is under a strongly seasonal climatic influence and its vegetation cover ranges from grassland and savanna to monsoon forest. The area is in striking contrast to the central and northern part of the Fly Platform which is under a wet tropical climatic influence and is covered by lowland rain forest and is intricately dissected. Here fluvial incision is the dominant process. This does not mean that surface wash does not operate, in fact it is quite substantial, but linear incision is dominant. Fluvial incision and headward retreat of valley and gully heads is aided by piping. These observations contrast with those of Tricart (1974) from the Amazon area where he claims that fluvial incision does not operate at present under rain forest conditions. He attributes the intricate dissection of the area to a former savanna environment.

It is very difficult if not impossible to delineate exactly the boundary between predominant sheet wash and linear incision, in particular as repeated horizontal shifts are to be expected during the Pleistocene. A tentative line could be drawn for the present about parallel to the lower Fly, but in the past it seems to have extended further north.

In summary it can be said that the mountainous landscape of Papua New Guinea does not fit easily into a morphoclimatic concept.
which is based more on latitudinal climatic changes over great distances than on altitudinal changes compressed into very narrow horizontal distances. The pronounced changes in the altitudinal climatic and biogeographical zones during the Pleistocene add further difficulties. The mountainous landforms of Papua New Guinea are best seen as belonging to one major morphoclimatic system, the humid tropical morphoclimatic zone, which is characterised by high rainfall and run-off, constant humidity and very uniform temperature regimes which vary from hot lowland tropical to cool highland tropical. Only in the very high altitudes where frost is of regular, i.e. diurnal, occurrence can a distinctly different morphoclimatic zone, the periglacial zone, be recognised. The glacial zone above this is not reached in Papua New Guinea but during the Pleistocene it formed a distinctive zone the traces of which have been well preserved.

In the lowlands a morphoclimatic differentiation can be recognised in the seasonally dry zones where planation replaces the predominant incision characteristic of the humid tropical morphoclimatic zone.

**CATASTROPHIC EVENTS AND UNIFORM PROCESSES**

The concepts of landform development in terms of abrupt or catastrophic changes on the one hand and uniform changes on the other have been the subject of heated discussions in the early days of geomorphology. Although the uniformists seemed to have won the discussion hands down there is now growing appreciation of a non-uniform if not catastrophic element in the rates of both endogenic and exogenic processes (Pitty 1971; Goosen 1974; Beaty 1974). Modern terminology has adopted terms like low-magnitude high frequency events and high-magnitude low frequency events for very much the same concepts (Wolman and Miller 1960).

The significance of catastrophic processes, particularly those related to volcanic and seismic activity but also others not associated with endogenic forces, has been stressed throughout the text and one cannot but be impressed by the rate of processes in some areas. Against this there are data indicating a surprising stability of landforms and survival of relict landscape features in spite of a high relief and consequent high erosion potential. It becomes obvious that one cannot apply a simple concept of uniformatism or catastrophism to the Papua New Guinea landforms but must accept that both catastrophic and uniform processes are of importance. It is one of the major problems of the geomorphological interpretation of the landforms of this country that the ratio between the two sets of processes is simply not known. However, many of the apparently contradicting observations and conflicting interpretations of Papua New Guinea landforms are simply the result of this interaction and superimposition of catastrophic and uniform processes, and can be explained by this. An added complication is that through time the rate of catastrophic and uniform processes has not remained constant, nor has the ratio between them.

In the Kaugel Valley for instance Pain (1973, 1975) considers that a very large part, in fact more than half, of the surficial basin filling originated from one major earthquake-induced landslide-
mudflow event that took place 20,000–30,000 years ago. Since then very little has happened within this basin. In many other intramontane basins geomorphic activity has similarly ceased or slowed down.

In contrast to this, activity in the Markham Valley and its northern catchment is at a high intensity. Much of the fan-building activity in the Markham graben has an element of catastrophism with maximum rates of aggradation of the order of several metres per year, and catastrophic shifts of channels.

In the catchment area landslides are a frequent occurrence and seismic activity is potentially high even though no major earthquake has been recorded since the area has been under European influence. The rate of uplift along the Huon north coast has a catastrophic character.

The dichotomy between catastrophic and uniform processes is, however, more apparent than real, and depends to a large degree on the definition of the two opposing terms and even more on the time scale considered. Processes that have recurrence intervals of several years, decades or even centuries and are certainly considered catastrophic in a man's lifetime are probably part of a much more uniform cycle of processes if one looks back over time periods of ten or hundred thousands of years. Clearly the longer the time period under consideration the more uniform any set of processes will appear.

THE HISTORICAL APPROACH

The historical approach is usually associated with the Davisian denudation chronology and it has become fashionable to blame Davis for causing stagnation in geomorphological research rather than advancing new concepts and encouraging new ideas. A great deal of the criticism lacks historical perspective (Jennings 1973) and it is simply not true that the concept caused stagnation in the advance of geomorphology. On the contrary, most of our present concepts have either developed as antitheses to the Davis concept, such as climatic geomorphology and dynamic geomorphology, or have greatly modified Davis's approach, such as in modern studies of historical geomorphology. Sediment studies aided by modern dating techniques and other laboratory facilities, quantitative information on the rates of geomorphic processes, a better knowledge of tectonic movements and the aid of remote sensing have greatly improved our understanding of the landscape and its interrelationships and historical dimension. In Papua New Guinea we are only at the very beginning of unravelling the complex history of the landform development with its complex interrelationships of structure, evolutionary history processes both exogenic and endogenic, and climatic change and associated change in geomorphic processes, and the problem of unpredictable geomorphic 'accidents' like earthquakes or volcanic activity. Close study of the stratigraphy of the sediments of intermontane and intramontane basins could provide important clues about the landform development in the respective catchment areas. However, the correlation from basin to basin will be a major problem.


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