ON THE SEDIMENTATION OF THE TABLE MOUNTAIN GROUP IN THE WESTERN CAPE PROVINCE.

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Thesis presented for the Degree of Doctor of Science at the University of Stellenbosch.

PROMOTORS:
PROFESSOR S. MASKE
PROFESSOR J. DE VILLIERS

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CHAPTER I

INTRODUCTION

This dissertation aims to illuminate some surprises which the deceptively uncomplicated Table Mountain sandstone had long been harbouring in its kloofs and cliffs. The major contributions of this first systematic investigation of an extensive portion of the Table Mountain sandstone are the discovery of diagnostic marine fossils by means of which the first direct dating of the deposit is effected and important sedimentological conclusions drawn; the stratigraphy of the Table Mountain sandstone is worked out in fair detail and a new subdivision and nomenclature suggested; the intraformational folds are analysed by means of electronic computation and their distribution traced, and a new theory presented to explain their formation; a massive onslaught is made on the major primary structure of the Table Mountain sandstone: cross-bedding, and the sequential paleocurrent history of the basin developed, this too, with the support of electronic computation; and the basin shape and successive stages in its development are shown in a step by step analysis in which use is made of trend surfaces. The basin is proved considerably deeper than accepted earlier, and the remarkably constant locus of maximum negative movement is indicated. A paleogeographical and sedimentological reconstruction synthesises all this information.

The discerning reader will no doubt point to many aspects of the Table Mountain sandstone neglected in this dissertation. For this the author offers no apology. The need for more detailed research on the Table Mountain sandstone is well documented in this report.

This study was started under the direction of the late Professor M.S. Taljaard in 1961 and the initial aim was to conduct a detailed petrographic-stratigraphic-sedimentologic analysis of the Table Mountain sandstone between Vanrhynsdorp and Port Elizabeth. Preliminary field work soon shrunk the study area to the tract between Vanrhynsdorp and Worcester (Fig. 1) but even in the smaller study area the unexpected large bulk of the sediments as well as their stratigraphic diversity shortly cast serious doubt on the practical implementation of such an expensive study program. The mountainous nature of the study area was an important factor in the early decision to concentrate on the shape of the basin, the dispersal pattern and the stratigraphic subdivision. These are important prerequi-
sites for further systematic detailed study of the Table Mountain Group.

The premature death of Professor Taljaard in 1966 brought the study to a halt and it was rounded off under the guidance of Professor S. Maske and Professor J. de Villiers.

The study area is a mountainous tract. Youthful to mature fluviatile dissection of the strike-faulted mild Cedarberg folds and the block-faulted syntaxis area to the south has carved an impressive mountain range complex since the earliest Cretaceous at least. The tectonic grain controls the topography closely and nearly all the individual ranges and intermontane valleys strike north to northwest. The major rivers which rise in these rugged kloofs are the south-flowing Breede, and the north-flowing Berg and Olifants Rivers. They are fed by winter rainfall which may locally exceed 200 inches per year but is normally 20-40 inches per year, supplemented by snowfall on the higher peaks.

The areas most neglected are the Sandveld between Redelinghuys and Vredendal, and portions of the Western Province. In the Sandveld an extensive sand cover combines with faulted near-horizontal strata to complicate correlation and other problems. In the Western Province time was insufficient to properly supplement the data obtained from peripheral and mountain pass studies. Fortunately many mountain passes traverse the entire study area and these afford good fresh exposures as well as easy access to numerous complete or near-complete stratigraphic sections.

If topography is at times an obstacle to the investigator, vegetation is seldom so. The very poor and sandy acid soil of the Table Mountain Group supports a sparse but typical plant cover. The Proteacea and Suurgrasses dominate and some of the more common indigenous trees are Olea africana (wild olive or olien), stunted Podocarpus falcatus and P. latifolius (yellow-wood), Rhus lancea (karree), Protea arborea (waboom), Ficus cordata (without), and the rare high altitude Widderingtonia cedarbergensis (Clanwilliam cedar).

Acknowledgements
First I want to express deep gratitude to my wife and family who gladly suffered innumerable inconveniences, spiritual as well as physical, some
foreseen but many unexpected, which this study enforced upon them over a period of more than seven years. Without their continued moral support and interest its completion would have been delayed many years.

My initial promotor, the late Professor M.S. Taljaard, was prevented by continued indisposition to materialise as actively as he wanted to the interest he had for this investigation. The data on the shape of the basin, the development of a computer program to deal with vectorial data, and especially the discovery of brachiopod shells were of much satisfaction to him.

I wish to thank my two subsequent promotors, Professor S. Maske and Professor J. de Villiers, for their interest in the project and their willingness to support my theses when this study was practically concluded and the data cast in near final form.

The behind-the-scenes computer programs which deal so effectively with vectorial data and trend surfaces were devised, written and tested by Mr. R.McD. Dodds during very many daylight and moonlight hours of hard work. Without his contribution and the facilities of the Computer Centre the sections on vectorial structures and basin analysis would have suffered immeasurably.

Many of my students, past and present, helped me over the years by collecting many hundreds of vectorial and other data on field trips, at times under my guidance but sometimes on their own. I cannot possibly name them all for fear of omitting one, but their incalculable contribution to this dissertation is a silent testimonial to their enthusiasm and unselfish scientific endeavour.

The Geological Survey and especially Mr. J.N. Theron were particularly helpful and readily supplied me with whatever information or assistance I requested.

During my 1965 nomadic field season my family and I enjoyed the hospitality of many. To all these new friends and to our parents I extend again my sincere thanks for all the kindnesses showered upon us during this time.

The Council for Scientific and Industrial Research generously supported this project. Without its financial aid this project would have been impossible.
CHAPTER II

REVIEW OF THE LITERATURE ON THE TABLE MOUNTAIN SANDSTONE

The literature falls into three groups: a) publications prior to the formation of the Geological Commission of the Cape of Good Hope; b) Annual Reports of this Commission (1896 to 1912) and contemporaneous publications; and c) modern publications, the more important of which appeared since 1925.

Very early papers

The Table Mountain Sandstone was named by A.G. Bain after the impressive Table Mountain at Cape Town (Draper, 1895). The term Cape Formation was coined by Schenck (1888) for the same sediments, but never used.

In 1852 A.G. Bain correctly estimated the age of the 10,000-12,000 feet thick unfossiliferous Table Mountain Sandstone to be Lower Silurian but three years later the Bokkeveld Beds were proved Devonian, thus suggesting that the Table Mountain Sandstone is somewhat younger than Lower Silurian. Bain found abundant large pebbles in the northerly trending ranges, and speculated on their provenance in view of the paucity of antecedent quartzose rocks and concluded: "... the parent of those numerous pebbles and conglomerates now lies buried in the depths of the Atlantic or Indian oceans." (Bain, 1856.) He correctly identified the sedimentary basal contact of the Table Mountain Sandstone and stressed the former connection between Table Mountain and the nearby interior ranges.

Annual Reports of the Geological Commission of the Cape of Good Hope, and other publications of this period.

The varied misconceptions of the Table Mountain Series in respect of its correlation and terminology have been discussed by Corstorphine (1896). The Table Mountain Series was then considered a non-fossiliferous shallow marine deposit, accumulated under conditions of balanced rates of subsidence and deposition; the Malmesbury slate and Cape granite highlands of that time now submerged beneath the Atlantic Ocean. Some conglomeratic sandstones between Klapmuts and Joostenberg were incorrectly correlated with the Table Mountain Series (Rogers, 1896).
Eastwards, at Gamka Poort "... a perfectly gradual passage..." occurs between the Table Mountain Series and the overlying Bokkeveld Beds (Schwarz, 1896).

Corstorphine and Rogers (1897) described red non-fossiliferous basal shales in the Peninsula, and near Stellenbosch, where another shale band occurs near the top of the Table Mountain Sandstone. Schwarz (1898) described the Upper Shale Band in the Franschoek Mountains, and noted "unusual variations in the composition of the sandstone" (p. 31, 1898) near it, but failed to identify tillite. Two years later, at Pakhuis Pass, the tillite horizon below the Upper Shale Band was identified by Rogers (1900, 1902, 1904) on: "(1) The peculiar character of much of the rock itself, viz., the distribution of large pebbles at intervals through a fine-grained sandy mudstone. (2) The flattened and faceted form of the pebbles, and their striated surfaces." (p. 239, 1902). He considered the tillite non-terrestrial and found no glacial floor, a peculiar circumstance in the light of later discoveries (J. N. Visser, 1962).

Prominent conglomerate beds at Lamberts Bay were reported (Rogers and Schwarz, 1900). Good exposures of the normally soil and rubble covered Upper Shale Band in the Cold Bokkeveld are "... composed of thinly laminated shales, slates, and mudstones, which, though very suitable for the preservation of fossils, have as yet yielded nothing in the way of organic remains." (p. 70, 1900). Corstorphine (1900), however, mentioned that "... one solitary piece of quartzite, from the Table Mountain Sandstone of the southern summit of the Zwartberg Pass, showing some worm-tubes, is the entire record." (p.xii, 1900). Schwarz (1903) apparently, discovered these organic remains.

Rogers and Schwarz (1900) demonstrated the northward thinning of the Table Mountain Series and noted that in the extreme north Dwyka glaciation was partly responsible for the reduced thickness. A prominent reddish thin-bedded shale horizon near the base of the Table Mountain Sandstone in the Piket Mountains and farther north was described by Rogers (1903). He also noted: "...The Table Mountain Sandstone is, on the whole, more conglomeratic in this area than in any other part of the Colony yet mapped." (p. 156, 1903).
Quartz and slate pebbles are common and jasper also occurs but granite is rare. The development of the conglomerate is insufficient to throw much light on its provenance. Schwarz (1905) found the basal reddish shale horizon dynamically metamorphosed, containing ottrelite towards Worcester. Again he referred to "... Very fine sections of the (upper) Shale Band..." (p. 268, 1905) in the Cold Bokkeveld and "... searched these beds most carefully for traces of graptolites and molluscan remains, but unsuccessfully. If we are ever to obtain fossils from the Table Mountain Sandstone it is in these beds that they will be found." (p. 268, 1905).

Rogers (1904) remarked on the predominantly quartzose nature of the Table Mountain Series, which has only two shaly horizons, of which the lower shows sun cracks and indicates extremely shallow water deposition. Important conglomerate horizons occur along the West Coast only. No organic remains were found except for some obscure trails or worm-castings, and the Table Mountain Sandstone contains no bands with marine fossils. Amongst others, this last aspect led Rogers (1904, p. 7) to think"... that the series was deposited by a river or several rivers running over a slowly subsiding area. On this supposition alone we get over the difficulty of the preponderance of coarse sediment, its great thickness, the occasional layers of pebbles, and the general absence of the fine-grained material which must have been produced by the denudation of the ancient land from which all the sand came. The fine mud must have been carried away and deposited beyond the area we have access to. As to the sources of the deposits it is unsafe to speculate, but there are good grounds for the belief that the northern part of the Cape Colony was the contributor."

The tillite and the Upper Shale Band were deposited in a lake. "From the absence of angular fragments freshly derived from the parent rock mass, we may suppose that the boulders were carried... a considerable distance; they were probably derived not from the immediate shores of the lake but from the country behind, by the combined agencies of glaciers, streams, and floating ice." (Rogers, 1904, p. 7).

Hatch and Corstorphine (1905) correlated the Waterberg and Matsap Formations with the Table Mountain Series, with which Schwarz (1905b) disagreed. Schwarz considered the Table Mountain Series too uniform in
character to approach the Waterberg Formation petrologically. Thinning of the Series northwards of Vanrhynsdorp indicated a nearby shore line. "...It was a sand laid down close in shore from off a vast continent, and it is inconceivable that similar conditions could have prevailed contemporaneously in the heart of the very continent whence it drew its supplies of material... and we can actually find jaspers, such as occur in Prieska, embedded in the Table Mountain Sandstone, in Clanwilliam." (p. 89, 1905b).

Schwarz (1906) thought the northward thinning of the Table Mountain Series resulted from progressive outlap of the upper beds, but Hatch and Corstorphine (1909) stated that its southerly increased thickness resulted from the lowering of the floor, later confirmed by Rust and Theron (1964).

Rogers and Du Toit (1909) reported tillite on top of Table Mountain at Cape Town. In a sedimentational reconstruction of the Table Mountain Series they pointed out that some 43,000 square miles were underlain by mostly quartz sand, and remarked on the absence of the concommittantly deposited finer-grained sediments. After considering and rejecting marine, lacustrine, and aeolian conditions they re-affirmed Rogers' fluvial theory.

Schwarz (1912) observed that "... along the coast at Vanrhynsdorp, there are coarse conglomerates at the base, showing that the actual shore was close by. The northern shore has not been satisfactorily traced." (p. 138, 1912).

Modern Publications since 1925

Contemporaneous folds, a most important and characteristic feature of the tillite, were first described in detail by Haughton, Krige and Krige (1925) for a few localities between Wellington and Gordon's Bay. Rennie (1925) reported similar folds on top of Table Mountain. Haughton (1929) showed that the tillite occurs between Pakhuis Pass and Gordon's Bay and eastwards to Swartberg Pass. The northernmost intraformational folds are near the Cedarberg Tafelberg. Haughton concluded that "... the folding was actually caused by the passage of an icesheet moving in general from west to east, the folds being formed just in front of the contact-line between the ice and the underlying unconsolidated water-logged sediments. As the ice moved forwards it planed or plucked off the tops of the anti-
clines already formed. On the retreat of the ice, much glacial detritus was brought down by sub-glacial streams and deposited as the fluvioglacial "Hard-Band" and possibly, when the supply of pebbles had ceased, as the succeeding Upper Shales". (p. 88, 1925).

Du Toit (1926) noted the occurrence of a tillite in Mitchell's Pass.

Haughton (1933) subdivided the south-western exposures of the Table Mountain Series into five zones: 1. Lower Shales, 2. Main Quartzite, 3. Glacial Band, 4. Upper Shales, 5. Upper Quartzite, and measured a maximum thickness of about 5,000 feet near Gordon's Bay. The 150-200 foot thick Lower Shales occur in the Cape Peninsula and in the mountains near Stellenbosch but were not observed at Sir Lowry's Pass. No conglomerate occurs at the base of the purple false-bedded, ripple-marked and sun-cracked flagstones where overlying granite; on Malmesbury rocks a thin white sandstone usually occurs at the base. Haughton discredited so-called "lamellibranchs" reported in the Lower Shales by Rogers (1909) and considered these impressions to be mud-pellets. The shales pass upwards into homogeneously arenaceous and occasionally pebbly Main Quartzite characterised by thick bedding with prominent cross-bedding. The Main Quartzite is between 1750 and 2500 feet thick. The Glacial Band occurs in the Hottentotsholland Mountains and on top of Table Mountain, and is easily identified by its unbedded structure and polygonal weathering, and by soled and striated pebbles set in an argillaceous gritty matrix. Excellent exposures of the glacial folds occur in Kogel Mountain near Gordon's Bay.

Between 1920 and 1937 a number of theses were accepted at the University of Cape Town dealing with the heavy minerals and other aspects of the Table Mountain Series in the Peninsula. (Duk, 1920; Malan, 1920; Rothkugel, 1920; Partridge, 1924; Cameron-Swan, 1935; Coetzee, 1937).

De Villiers (1944, 1956) reviewed the orogenic background, amongst others, of the deposition of the Cape System and noted that in the west the pre-Cape rocks had been severely disturbed so that a marked angular unconformity occurs at the base of the Table Mountain Series. "This pre-Cape diastrophism ... was followed by a long period of crustal stability. Erosion was active, and as a result there was produced in the west what must have been the most perfect peneplain in the geological history of Stellenbosch University http://scholar.sun.ac.za
South Africa. The surface must have been nearly level, as can be seen where
the post-Cape movements have not disturbed it much, e.g., in some parts of
the Western Provence." (p. 188, 1944). "The pre-Cape platform was an extreme-
ly flat peneplain and towards the end of the Silurian this plain was deflect-
ed southwards along a hinge-line stretching from Vanrhynsdorp to Zululand,
and inundated by the sea from the south. On this broad continental shelf,
where the water was shallow throughout, the Table Mountain Series was deposit-
ed ..... all the evidence support the contention that these sediments were
produced by deposition under epineritic conditions." (freely translated, 1956).

Rogers (in De Villiers, 1944) wrote "Probably the (upper) shale band
rests on the pre-Cape near Matsikamma or the Kobe Mountains." (p. 197, 1944).
He thought that the shale band persisted beneath the glacially eroded upper
sandstone and eventually interposed between the pre-Cape rocks and the Dwyka
tillite.

Conradie and Rabie (1944) deduced from a study of the size - shape
characteristics of a part of the Table Mountain Series at Botmanskop at
Stellenbosch that there were rapid variations in the transporting current
velocities at the very start of the Cape sedimentation.

Swart (1950) investigated the Table Mountain Series/Bokkeveld Series
contact at Wuppertal and presented a synthesis of current ideas of the sedi-
mentation of the Table Mountain Series. In the west a southward sloping
fluvially produced peneplain slowly subsided and marine erosion reduced it
to an almost plane surface. In the south-west the first deposits were
littoral sands and muds. Swart opposed Du Toit's (1937) delta theory and
considered the Table Mountain Sandstone a "... marine-produced, marine-trans-
water fluctuating currents left abundant cross-bedding in the sands. The
absence of fossils is due to unfavourable climatic and bottom conditions.
The easterly thinning of the tillite and the attitude of the glacial folds,
indicate a westerly centre of glaciation. The unsorted and unlaminated
tillite indicates the temporary absence of strong currents, but Swart did
not explain why; he merely discredited Du Toit's (1937) lacustrine theory
and stated that the tillite was also marine. Passing over the Upper Shale
Band, Swart remarked on the thin-bedded, feldspathic, pebbly, fossiliferous
upper sandstone below the Bokkeveld Series. Glacial agents played an im-
portant rôle during the deposition of this last facies of the upper sand-
stone, and the very sharp concordant contact between the Table Mountain
Series and the Bokkeveld Series in this area, as compared to the gradational
contacts east, could in some manner be due to glacial controls, or possibly
to climate change, or to a change in the rate of subsidence.

Du Toit (1954) gave preference to a partly aeolian derivation of the
Table Mountain Series: "... the region formed probably a vast coastal plain
or a series of broad deltas studded by wandering dunes that were periodical-
ly levelled and reworked by rivers or ponded streams." (p. 246, 1954).

Von Brunn (1959) reported the normal stratigraphic sequence, except
for intraformational folds, in the area north of Tulbagh.

Hardie (1959) considered the Table Mountain Series in Natal a shore-
line deposit representing a mixed terrestrial-stable platform facies in the
east which grades into the well-sorted shallow marine facies south of Zulu-
land.

Von Backström (1960) reported that south of Nieuwoudville the full
thickness of the Table Mountain Series is between 300 and 400 feet, and
that the northerly thinning is due to both depositional thinning-out and
erosion by Dwyka glaciation.

A diminutive isopach map of the Table Mountain sandstone (Woodward,
1961) shows an east-west basin edge from Calvinia passing north of Vanrhyns-
dorp, and a southeasterly trending trough between Citrusdal and Touws
River, more than 10,000 feet deep.

Visser (1962) noted heavy mineral rich beds in the Table Mountain
Series in a few localities along the West Coast. Von Backström (1962)
reported near Klawer white quartzites and purple shales between the Table
Mountain Series and the Malmesbury Formation, and tentatively but incorrect-
ly correlated these beds with the Klipheuwel Formation.

Near Villiersdorp the Table Mountain Series is 6250 feet thick
(De Villiers, 1962), increasing eastwards to 10,500 feet near Prince Albert
(Rossouw, 1962), and to somewhat less than 20,000 feet in the Groot Winter-
hoek Mountains (Marais and Dippenaar, 1962), where the upper shale horizon
is absent and the basal contact tectonically disturbed.
Visser (1962) subdivided the glacial zone into three members: a lower tillite, which occupies the troughs of the synclines in the underlying fold zone, and which is overlain by the "Hard Band" (an undeformed grit) and another, also undeformed, tillite. Glacial floors, heavy mineral distribution, and tillite fabric analyses indicate ice movement from north to south. The contemporaneous fold mechanism is obscure. Visser thought that the tillites were deposited by a number of individual glaciers. Differences in the upper and lower tillites resulted from appreciable differences in the mode of glaciation; the lower tillite was deposited by terrestrial glaciers and the upper when the area was inundated. He considered the "Hard Band" to be a local deposit derived from contemporaneous erosion of the lower tillite and its associated folds.

Theron (1962) reported Table Mountain Series tillite at Towerwater Poort but now considers it a misidentification (Personal communication).

Taljaard (1962) described a 12 feet by 5 feet bedding plane slab from the basal Table Mountain Series at Brandenburg, Clanwilliam District, with trails, tracks and burrows supposedly made by trilobites or eurypterids on it. The trails were made on wet tidal beach sand, and covered by a windblown layer before the quiet return of the tide. (This slab is in display at the Department of Geology at the University of Stellenbosch.)

The heavy minerals of the Table Mountain Series were investigated sporadically (see previous references to theses presented at the University of Cape Town; Hëllich, 1962; de Villiers and Wardaugh; 1962; Conradie and Rable, 1944; Swart, 1950.) Stable species, such as zircon and rutile are common; unusual species such as reddish zircon and garnet occur in the tillite while brookite occurs in the upper sandstone near Nieuwoudville.

Van der Merwe (1963) discussed the Table Mountain Sandstone in the northern Cape Peninsula. The 150-200 feet thick ripple-marked lower shale overlies the unconformity directly and its shale/siltstone ratio is about 1/4. Stream ripple marks indicate flow from N25E, which corresponds to cross-bedding data. The bedding thickness and grain size of the main sandstone increase upward. Ubiquitous cross-bedding resulted from streams which came from N32W to N22E. Small tillite exposures west of Maclears Beacon are briefly described and a fine-grained vein quartz conglomerate
noted below the tillite. The upper shale is not present.

De Villiers, Jansen and Mulder (1964) observed the lenticular habit of the basal shales in the area between Worcester and Franschhoek and remarked on the variable basal sediments: a coarse conglomerate at Slanghoek, shale near Franschhoek and sandstone at Stettynskloof, Brandwacht and Kaaimansgat. The monotonous pebbly quartz sandstone which forms the bulk of the Table Mountain Series contains some shale and rare arkose, and is ripple-marked and profusely cross-bedded. The current direction was to the south-southeast. Problematic tube structures occur in places. The tillite and intraformational folding are common, and topographic expression, karst-type weathering surface and resistant faceted erratics up to 10 inches in diameter typify the tillite. The upper shale has a regular thickness of about 150 feet but near Zachariaashoek, Franschhoek it is absent over short distances. In the Bier River delicate varves occur in the upper shale. The upper sandstone resembles the lower, but it tends to be reddish near Rawsonville, and contains more shale. The following thicknesses were either measured directly or calculated:

Lower shale plus sandstone plus tillite 3900 feet
Upper shale 150 feet
Upper sandstone 2200-500 feet

The total thickness of 6250 feet is 25% more than previous accepted estimates.

The sediments were deposited in shallow water, as indicated by cross-bedding and soft-sediment deformation by the glaciers which deposited the tillite. The upper shale indicates deeper water. The climate was cold.

Rust and Theron (1964) described the progressive northerward overlaps and outlaps of the Table Mountain Series near Vanrhynsdorp. A basal conglomerate of variable thickness and the main sandstone wedge out more or less simultaneously; northwards vari-coloured tillite overlies the pre-Cape rocks. No glacial pavements were found. North of Klein Kobe the Table Mountain Series is represented by the upper sandstone only.

Rust (in the press) announced the occurrence of brachiopods in the upper shale near Porterville, about 2200 - 2850 feet below the Bokkeveld Series. The provisional age is Lower Silurian.
CHAPTER III

THE FIELD GEOLOGY OF THE TABLE MOUNTAIN GROUP

Introduction

In the description which follows locality names present a problem because many of the geologically important sites described are geographically insignificant. The problem is aggravated in this dissertation by the absence of a geological map but the more important locality names appear in Figure 1. With one or two exceptions all the place names used occur on current 1:50,000 and 1:18,000 maps compiled by the Trigonometric Survey, Mowbray (Table 1), and most are indicated on the 1961 1:250,000 topo-cadastral maps.

Table 1. Index of 1:50,000 and 1:18,000 map names arranged in their relative geographic positions. Locality names referred to in the text occur on these maps.

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<th>Vanrhynsdorp</th>
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<th>Urionskraal</th>
<th>Lokenburg</th>
<th>Klawer</th>
<th>Nardouw</th>
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Most of the localities are given a double name: the first is specific, and the second refers to a nearby town, or the 1:50,000 map name.

The absence of a geological map is compensated for in part by some 60 cross-sections, some of which are constructed from topographic maps, but most are field sketches which merely show general field relationships without strict adherence to scale.
**The suggested new stratigraphic names**

The latest directive by the Geological Survey regarding stratigraphic nomenclature (1947) advocates the use of the twofold terms proposed by the Eighth International Geological Congress in 1900. This is supported by Verwoerd (1964) and is applied in part to the rocks under discussion by De Villiers, Jansen and Mulder (1964) and Du Toit and Haughton (1966) (Table 2).

<table>
<thead>
<tr>
<th>Geological Survey, 1947</th>
<th>Group</th>
<th>System</th>
<th>Series</th>
<th>Stage</th>
<th>Substage</th>
<th>Zone</th>
<th>Stage</th>
<th>Stratum</th>
<th>Text</th>
</tr>
</thead>
<tbody>
<tr>
<td>De Villiers et al., 1964</td>
<td>(?) Cape-Karoo Cycle?</td>
<td>Cape System</td>
<td>Table Mountain Series</td>
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<tr>
<td>Du Toit and Haughton, 1966</td>
<td>Cape System</td>
<td>Cape Mountain Series</td>
<td>Table Mountain Series</td>
<td>Upper sandstone</td>
<td>Upper sandstone band</td>
<td>Upper shale band</td>
<td>Glacial Zone</td>
<td>Lower sandstone</td>
<td>Lower shale</td>
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**Table 2.** Stratigraphic names in use and the suggested new nomenclature.
It is clear that the traditional South African stratigraphic nomenclature avoids application of formal names lower in rank than "series".

Without elaborating on the merits or demerits of this traditional nomenclatorial approach or its particular application, it is, firstly, not in accordance with international intent (Hedberg, 1961) and extensive usage (ACSN, 1961) and, secondly, shown in this dissertation and elsewhere (Johnson, 1966; Theron, 1967; Loock, 1967) to be unsuitable for detailed stratigraphic study.

The recognition of previously unknown marker beds, fossil zones and facies changes in the Table Mountain Group soon proved to the author that a nomenclatorial change was imperative. Application of the American Code of Stratigraphic Nomenclature (ACSN, 1961; Krumbein and Sloss, 1965) to the Table Mountain Group yields a nomenclatorial image which is formally rigid yet practically flexible, adaptable and sensitive. The ACSN separates or helps to separate objective features (lithostratigraphic and biostratigraphic) from subjective factors (chronostratigraphic). A possible objection against the ACSN is that its application leads to a multiplicity of names, but is this very ability of the Code not desirable when dealing with a complex sedimentary deposit? Many components need many names. The ACSN is an important step towards giving and using these many names in an ordered and consistent way.

New lithostratigraphic names for the components of the Table Mountain Group (Table 2 lists only the most important) have been given not merely for the novelty of change or experiment but with the conviction that the ACSN has appreciable merit and that its future usage, extention and refinement will lead to easier communication and more meaningful stratigraphic subdivision.

The six formations are grouped in the TABLE MOUNTAIN GROUP mainly on petrographical grounds, but their relationships to the basal unconformity as well as to the overlying Bokkeveld sequence are also distinctive. Fossils are not common in the Table Mountain Group but they are unlike those of the Bokkeveld sequence. The lithostratigraphic unit, Table Mountain Group, is identical to the succession usually referred to as the Table Mountain Series. Short motivations follow of the names chosen for the formations.
Piekenier Formation

This dominantly conglomerate and sandstone succession is well exposed in Piekenierskloof Pass near Citrusdal. Due to an important twofold member division the formation has two type areas namely Piekenierskloof Pass and the Piketberg Range.

Graafwater Formation

Reddish sandstone of this formation underlies an appreciable area near Graafwater but the type area is at Looprivier, 22 miles southeast of Graafwater, where the stratigraphic relationships are more clearly demonstrated.

Peninsula Formation

This white sandstone typifies the Table Mountain Group and it builds almost all the mountains of the Western Cape. The type area of the Peninsula Formation is between Porterville and Citrusdal but characteristic stratigraphic relationships are displayed in many places. The name alludes to the numerous exposures in the Cape Peninsula, notably in the face of Table Mountain.

Pakhuis Formation

First proof of a tillite in the Table Mountain sequence was furnished by the Pakhuis Pass road-cuttings and it is appropriate to assign this name to the Pakhuis Formation. The Pakhuis Pass exposures are still among the best and most easily accessible outcrops of the tillite.

Cedarberg Formation

In the Cedar Range the shaly Cedarberg Formation, which is normally readily identified and is a useful marker, is especially conspicuous due to the broad anticlinal structure of the mountainland. The type area of the Cedarberg Formation is near Porterville, however, where an important fossiliferous zone is developed.
Winterhoek Subgroup

Glacial control on the Pakhuis and Cedarberg Formations makes a subgrouping desirable, and where appropriate these formations, together with the Fold Zone of the Peninsula Formation, are referred to as the Winterhoek Subgroup in allusion to the typical development of glacial features in the Winterhoek Range north of Tulbagh.

Nardouw Formation

Although widely exposed, this sandstone formation seldom builds prominent ranges, with the exception of the Nardouw Escarpment between Clanwilliam and Vanrhynsdorp, and its northern extension into the Bokveldsberg Plateau. In the Cold and Warm Bokkeveld the Nardouw Formation forms the rugged Skurweberg Range. Scattered type areas have been assigned to the members of the Nardouw Formation.

The Lap-out Map.

The lap-out map (Fig. 2) indicates the distribution of the formations of the Table Mountain Group where they subcrop on the basal unconformity. This type of map has been jocularly called a worm's eye map (Levorsen, 1960).

A dip cross-section (Fig. 3) in the northern portion of the lap-out map shows that the upper formations progressively lap over the lower formations. All the formations thin towards the northeast and thicken basinwards. This stratigraphic relationship indicates transgression (Krumbein and Sloss, 1965). The subcrop strike of the transgressive formations parallels the original basin edge (Levorsen, 1960).

The central area of the lap-out map is a regular subcrop of the Piekenier Formation. The control on this subcrop is poor but the interpretation presented on the map is the most likely at present. The subcrop pattern of the Piekenier Formation indicates the initial size and placement of the Table Mountain basin.

Off-lap and lap-out relationships between the Peninsula, Graaffwater and Piekenier Formations occur southwest of Worcester. Only the broad pattern is shown. The Peninsula Formation subcrop in the Stettyn area results from a structural high on the basin floor, the Stettyn Rise,
which projected above the level of pre-Peninsula strata. On the northern slopes of the Stettyn Rise is some evidence that the Graaf-water Formation is off-lapped on the Piekenier Formation.

Very coarse basal conglomerate lenses consisting of locally derived angular and sub-angular quartzite pebbles and boulders (Plate 1E,G), occur between Slanghoek and Sybas Mountain. These lenses are isolated features of the unconformity in the south. The floor topography is visibly uneven at Sybas Mountain and in places resistant quartzite ridges project through the Piekenier and Graafwater Formations into the Peninsula Formation.

Regionally the profile of the plane of unconformity is impressively even. This is a reflection of the energetic pre-Cape erosional agent or agents, but it is a misconception to regard planation and sedimentation as separate and consecutive; sedimentation of any one formation was actively proceeding in the subsiding basin while simultaneously the basin periphery and environs were being planated and contributing debris. The unconformity represents a long time interval, and in the chronostratigraphic framework of the Table Mountain Group it is markedly diachronous.

The unconformity is normally mantled by a basal conglomerate, which, if recognised as a separate lithostratigraphical unit, is also diachronous. In certain areas the unconformity surface was swept clean of any pebbles, as at Sir Lowry's Pass, Somerset West, and Chapman's Peak Drive, Simonstown. This type of basal contact is a characteristic of much of the Graafwater and Peninsula Formation subcrops in the southern area of the lap-out map.

Large well-rounded exotic pebbles occur on the unconformity near Piekenierskloof Pass, (Plate 1 A, B) Redelinghuys, Elands Bay and Klein Tafelberg. The exotic pebbles indicate a significant change in the sedimentological character of the basal deposit. The Piketberg Ranges are devoid of such basal conglomerate beds.

No outcrops of the unconformity are known in the region bounded by Elands Bay and the Olifants River estuary on the West Coast, and inland between Renostershoek, Eendekuil, and the Bulshoek Barrage, Clanwilliam. This is due to the nearly horizontal attitude of the Table Moun-
tain strata, and to their great thickness. The unconformity may be exposed in the gorges east of Keerom, Citrusdal.

Below the Bulshoek Barrage and near Trawal, Klawer, the Graafwater Formation rests on the unconformity. An immature small-pebble conglomerate with sub-rounded vein quartz pebbles typifies the subcrop of the Graafwater Formation. The regional appearance of the unconformity surface is one of extreme regularity. Near Bellevue, Vredendal, a very prominent and coarse basal conglomerate (Plate 4F) may represent an in-filled channel or valley, but field evidence is inconclusive.

North of Klawer the Peninsula Formation subcrops on the unconformity and excellent outcrops occur on the mountain slopes between Giftberg and the Kobe. Normally only a few well-rounded pebbles are scattered on the unconformity (Rust and Theron, 1964, Plate I, Fig. A), as at Widouw Pass, Vanrhynsdorp, but as the pinch-out zone is approached, conglomerates become increasingly noticeable at the contact plane. Rust and Theron (1964) reported very coarse conglomerates at the base of the Peninsula Formation near Kobe, Lokenburg, but subsequent more detailed investigation has cast doubt on the correlation of these rudites with the Peninsula Formation. Structural relationships near Witkleigat in the Kobe Valley indicate that some of these rocks are of pre-Peninsula age.

Many conglomerate lenses occur in the lap-out zone of the Peninsula Formation. Near Krieskamma in the Kobe valley the pebbles and boulders are well-rounded and may be as large as 30 inches in diameter but most are between 4 inches and 6 inches in diameter (Plate 5E). In these exposures the basal conglomerate consists of about 15% jasper and jaspilite pebbles, 5% to 10% white spherulitic banded chert pebbles, about 35% to 40% boulders and pebbles of a dark grey coarse-grained quartzite, and the rest are vein quartz pebbles. The conglomerate lenses fill a low rolling topography near Matjiesgoedkloof on the western slopes of the Kobe Range (Plate 5A). The lenses are up to 25 feet thick and their long axes appear to trend 120 degrees. Near Witkleigat and Krieskamma portions of the basin floor were covered by patches of large in situ weathered blocks at the time of deposition of the Peninsula Formation conglomerate lenses so that sand and pebbles infiltrated into the spaces between the
underlying rubble (Plate 5E).

Still in the confines of the Kobe valley, the Pakhuis and Cedarberg Formations lap out on the unconformity.

Near Bergland, Kobe, the unconformity underlies the Pakhuis Formation and two fine striated and polished glacial pavements are cut into a prominent quartzite member of the Nama Group. The original topography of this floor exceeded 100 feet vertically, forming a single conspicuous hill which was buried only at the conclusion of the Pakhuis glaciation. Regionally, however, the contact is very even. The unconformity beneath the Cedarberg Formation is never clearly seen due to the similar weathering characteristics of the Cedarberg siltstone and shale, and the underlying Nama subgraywacks.

North of Rooikloof, near Klein Kobe, the Nardouw Formation lies on the unconformity (Rust and Theron, 1964, Plate VI, Fig. F). Once again a basal conglomerate becomes more prominent northwards. It is well exposed at Vanrhyns Pass, at the Nieuwoudt Falls, Nieuwoudville, and near Brandkom on the road to Loeriesfontein. The pebbles are mostly vein quartz, some with black tourmaline crystals, and are well-rounded to rounded, with diameters up to 19 inches but averaging less than 2 inches. The Bokveldsberg Plateau forms the most northerly outcrop of the Table Mountain Group, which is here represented in its entirety by the Nardouw Formation.

The unconformity beneath the impressive scarps of the Nardouw sandstone is remarkably even, and practically horizontal for many miles (Plate 21 D). The only noticeable irregularity in this monotonously flat plane is a very broad and shallow depression near Paardekraal, Nieuwoudville. Towards the east the Nardouw Formation is rapidly truncated by pre-Dwyka erosion and no further outcrops occur except for a few small inliers near Calvinia (J.N. Theron, personal communication, 1967).

Paleogeologic map of the unconformity

The paleogeologic map (Fig. 4) is a reconstruction based on outcrop data, extrapolated outcrop data and deduction. Two boreholes only are known to have penetrated the Table Mountain Group. In the northern and southern portions of the map the control is good but in the central area it is poor. The main intention of this map is to indicate the broad distribution of pre-Table Mountain rocks. Internal structural relation-
ships have not been indicated and the interested reader is referred to de Villiers (1944, 1956), Scholtz (1946), Brink (1950), Rabie (1948), Von Backström (1960), Jansen (1960), and De Villiers et al. (1965).

In the extreme north of the map only a small portion os shown of the presently extensive Namaqua Granite-gneiss Complex. This ultra-metamorphic complex probably played a subdued rôie in the provenance area of the Table Mountain basin.

The Malmesbury Group, which is a regionally metamorphosed and tectonically disturbed eugeosynclinal deposit of unknown thickness, is a conspicuous feature of the paleogeological map. Towards the south the Malmesbury Group is invaded by the syntectonic Cape Granite Complex (553 ± 8 m.y., Allsop and Kolbe, 1965) and here the combined outcrop distribution of sediments plus granite is indicated. Lumped with this unit are one or two possibly unrelated and relatively unimportant formations, of which the Franschhoek Formation (?) is the most important.

The Nama Group (Cambrian, sensu stricto, was probably exposed over an appreciable area north of the paleogeological map boundary. It is an unstable shelf deposit of immature sandstone and turbidite, shale, grit and limestone. The Nama Group is usually only moderately folded but it is steeply tilted at Vanrhyns Pass and the degree of folding increases westwards. The Nama Group was probably a major debris contributor to the Table Mountain basin.

The Klipheuwel Group (Late-Cambrian ?) is a succession of very coarse to very fine-grained reddish clastics. Its relationship to the Nama Group is not clear but the Klipheuwel Group is considerable younger than the Malmesbury Group.

The Klipheuwel beds are concordant to the Piekenier Formation at Redelinghuys, Elands Bay, Klein Tafelberg and Weltevrede near Aurora (Figs. 7,8,9,11). However, at Meerlandsvelei, near Piokenierskloof Pass, a low-angle unconformity proves the true relationship between the Klipheuwel Group and the Piekenier Formation (Fig. 6). When the Piekenier Formation was deposited a considerable area of the basin floor was probably underlain by the Klipheuwel Group.

The northwestern trend of a number of contacts on the paleogeological map reflects the regional grain of the area, and it is notable that the isopach maps of the Table Mountain Group formations partly mirror
this structural control.

The organisation and interrelationships of the elements on the paleogeological map reflect on the prolonged existence of a zone of crustal weakness which was sedimentologically exploited a number of times. The earliest sedimentary cycle is the Malmesbury eugeosynclinal cycle, which culminated in a magmatic phase. The Namaqua Craton probably grew by accretion of the Malmesbury eugeosynclinal sediments. In none of the subsequent sedimentary cycles was the crustal activity as intense as in the initial cycle, and magmatic phases did not develop again. The Nama and Klipheuwel cycles produced mostly unstable shelf conditions and the Table Mountain cycle was reduced to a stable shelf framework.

The Piekenier Formation

The type area of the Piekenier Formation includes Piekenierskloof Pass, and the Piketberg Range. Two members are recognised, namely, the conglomerate REST MEMBER, named after The Rest at the western approach to the pass, and the sandy DE HOEK MEMBER, named after De Hoek, south of Piketberg. This twofold division of the Piekenier Formation is an oversimplification of its stratigraphy and much tongueing and gradation occurs. It is at present convenient, however, to view the Piekenier Formation as consisting of these two members only. In the text illustrations the Piekenier Formation is indicated by the symbol $C_1$.

The Rest Conglomerate Member.

This member is very well exposed in the cuttings of Piekenierskloof Pass but its base and top are unfortunately obscured. It is a thick-bedded, profusely cross-bedded, open framework conglomerate with a sandy matrix, and a few thin beds of purple sandstone and shale near the base. The conglomerate beds are characteristically lenticular and extensive re-working took place before permanent deposition. The cross-bedding pattern (Fig. 84F) shows that strong currents flowed consistently from the northwest. The mean diameter of 50 largest pebbles is 2.3 inches. (This value will be referred to as the mean largest pebble diameter.) The pebbles are well-rounded and an estimated 98% or more are vein quartz; the rest consist of reddish and white quartzite, red jasper, gray banded and black chert, dark green hornstone, dark red
mafic and acid lava, argillite, diorite and hornblende gneiss. Near the base some obviously locally-derived intraformational shale fragments, as much as 18 inches long, occur in conglomerate.

In the slopes of Meerkoffs Kasteel, Eendekuil, (Fig. 5) the Rest Member becomes sandy and thin-bedded and contains only occasional pebbly sandstone beds as the Graafwater Formation is approached. The upper contact of the Rest Member here is the base of the first purple shale or siltstone bed which shows worm trails. So defined the Piekenier Formation is about 1800 feet thick.

At Meerlandsvlei, Eendekuil, coarse massively bedded conglomerate overlies the Klipheuwel grits with an angular unconformity (Fig. 6). The purple shale near the base at Piekenierskloof Pass is absent at Meerlandsvlei. At Skerpklip, Eendekuil, the conglomerate contains rounded vein quartz and quartzite pebbles up to 6 inches in diameter, set in a gritty matrix.

North of Jansekrail and Agterland, Eendekuil, and south of Piekenierskloof Pass the conglomeratic Rest Member grades into the sandy De Hoek Member.

Between Redelinghuys and Elands Bay the Rest Member is mostly a sandy small-pebble conglomerate. At Skuinskraal, Redelinghuys, the Rest Member lies conformably on Klipheuwel grit (Figs. 7, 8). The mean largest pebble diameter is 2.25 inches and the coarse basal conglomerate grades upwards into a pebbly orthosandstone. The 5% pebbles other than vein quartz are jasper, green and yellow quartzite and banded brown chert. The lower 300-400 feet of the Piekenier Formation are exposed at Skuinskraal.

At Muishoek Mountain, Elands Bay, the basal unconformity is exposed at the Verlorevlei lagoon. Once again reddish Klipheuwel grit appears concordantly beneath the Rest conglomerate, the lower foot or two of which is also reddish. The pebbles are predominantly vein quartz, and white, blueish-green, pinkish and gray quartzite, jasper and reddish quartz-porphyry. The pebbles are subrounded and the mean largest diameter is 2.1 inches. The conglomeratic Rest Member is about 350 feet thick and grades into a coarse pebbly thick-bedded orthosandstone, which is correlated with the De Hoek Member.
At Elands Bay the Piekenier stratigraphy is similar to that at Muishoek Mountain (Fig. 9). At Baboon Point the mean largest pebble diameter is 2.4 inches (Plate 1B). The De Hoek Member forms the cliff at Baboon Point while Klipheuwel grit is exposed below the high tide line near the jetty.

At Klein Tafelberg, Elands Bay, the unconformity is cut in quartzite and reddish grit of the Klipheuwel Group. The plane has small scale irregularities, filled in by lenticular coarse conglomerate of the Rest Member. Apart from the ubiquitous vein quartz pebbles, other lithological types are variously coloured quartzite, reddish sandstone, khaki argillite, hornstone and much black chert. The pebbles are well-rounded and the mean largest diameter is 2.5 inches. The Rest Member is only a few feet thick and most of the capping of Klein Tafelberg is built of thickly bedded coarse orthosandstone of the De Hoek Member.

At Lamberts Bay an estimated 2600 feet of coarse cross-bedded conglomerate and interbedded sandstone is correlated with the Rest Member although neither top nor bottom is exposed. Reddish orthosandstone becomes prominent lower in the sequence but there is no sign of an approach to the basal contact. The numerous sandy bays on the coast and the sand-covered inland make it impossible to determine whether the succession has been displaced by any of the common northwest trending faults in this area; the jointing is, however, moderate and brecciation is absent.

Farther south, at Steenbokfontein, more conglomerate is exposed, here, too, dipping northwards at about 10 degrees.

The conglomerate near Lamberts Bay ranges from an openwork lag gravel, with an infiltrated coarse sand matrix (Plate 1C) to a disrupted framework conglomerate (Plate 1D). Lag gravel indicates appreciable re-working and sorting prior to final deposition, whereas a disrupted texture results from very rapid deposition with no further re-working. Lag gravel frequently forms only a thin deposit, sometimes a single line of pebbles, but beds of disrupted framework conglomerate can be very thick. At The Point, Lamberts Bay, such conglomerate beds are up to 10 feet thick. The lithological pebble types are similar to those elsewhere in the Rest Member. The largest pebbles are 10 inches in diameter.
North of Lamberts Bay, opposite the camping site, thin-bedded fine-grained sandstone and purple shale of the Graafwater Formation are down-faulted against the Piekenier Formation.

At Kanon, near Donkin Bay, at least 850 feet and possibly as much as 1800 feet of very coarse conglomerate and massively bedded white orthosandstone contain rounded vein quartz pebbles up to 8 inches in diameter. This can probably be correlated with the Rest Member.

Southwest of Doorn Bay inaccessible off-shore islands are built of coarse conglomerate which surely belongs to the Rest Member. This conglomerate is succeeded by massively bedded medium- to fine-grained white orthosandstone correlated with the De Hoek Member. A fault separates the Piekenier Formation from the Graafwater Formation.

South of Bellevue, Vredendal, very massively bedded coarse-grained vein quartz conglomerate about 100 feet thick forms the basal sediment of the Table Mountain Group, which here overlies micaceous phyllite of the Malmesbury Group. At present all basal conglomerate in this area is classed as Graafwater Formation (next section).

The stratigraphy of the Piekenier Formation along the West Coast is not well known due to isolated and incomplete exposures. The very coarse oligomict orthoconglomerate indicates two features namely proximity to the basin edge, and the supracrustal nature of the provenance area. The cross-bedding diagrams (Fig. 85) show that very strong currents flowed from the northwest and the basin edge was probably in that direction.

The De Hoek Member

The main development of the De Hoek Member is in the Piketberg Range and south of it, although important exposures occur near Olievenboskraal.

At De Hoek, south of Piketberg, a thin-bedded medium-grained, occasionally gritty and thick-bedded, orthosandstone about 760 feet thick overlies Malmesbury limestone with marked unconformity. The relationship of this sandstone to the basal unconformity and the overlying Graafwater Formation here, and at Zebrakop (Plate 2E) and Grootkop, Eendekuil, and Swart-
rug, Goergap, shows it to be the stratigraphic equivalent of the conglomeratic Rest Member of nearby Piekenierskloof Pass.

At Oukraalsvlei, Piketberg, the contact between the De Hoek Member and Malmesbury sericite schist is sharply exposed. About 10 feet of coarse basal conglomerate with locally derived subangular vein quartz and mica schist pebbles grade into cross-bedded white orthosandstone with some thin purple ripple-marked shale beds. The upper boundary of the 825 feet thick Piekenier Formation is the uppermost white orthosandstone below the purple siltstone of the Graafwater Formation. At nearby Waboom are some indications of buried topography, probably an original ridge of dynamically deformed Malmesbury small-pebble conglomerate, but the relationships are somewhat obscured by post-depositional tectonic deformation. Normally the plane of unconformity is very even. At Grootkop, Eendekuil, the De Hoek Member has a thin basal conglomerate and the remainder consists of white orthosandstone (Fig. 10).

On the western slopes of Platberg, Aurora, no basal conglomerate overlies the Malmesbury phyllite, but about 120 feet of reddish thin-bedded fine-grained sandstone with very indistinct worm-like trails and faecal pellets grade upwards into pebbly orthosandstone which is well exposed on the eastern slopes of Platberg (Fig. 12). Scattered vein quartz pebbles up to 1 inch in diameter occur in the conspicuously festoon cross-bedded sandstone.

At Rooiberg, near Weltevrede, Aurora, thinly bedded coarse reddish sandstone (Plate 2C), in places quite massive at the base, and about 400 feet thick, lies directly on a very even plane of unconformity cut in mudstone and coarse petromict conglomerate of the Klipheuwel Group. The thin-bedded sandstone is overlain by about 400 feet of coarse white orthosandstone which grades upwards into a gritty vein quartz conglomerate (Plate 1F), about 200 feet thick. The mean largest pebble diameter is 1.25 inch.

The lower sandstone at Platberg and Weltevrede is correlated with the De Hoek Member, and the upper conglomeratic sandstone with the Rest Member. This vertical relationship is the opposite of that at Muishoek, Klein Tafelberg and elsewhere; this inversed arrangement is discussed later.
North of Piekenierskloof Pass, from Sewefontein, past Klein Alexandershoek to Lambertshoek, the Piekenier Formation is represented by the De Hoek Member, a white orthosandstone similar in all appearances, except stratigraphic position, to the Peninsula sandstone.

On Paardekop, Eendekuil, the De Hoek Member is at least 2800 feet thick and between Bobergvlei and Klipkraal it is probably more than 3000 feet thick (Fig. 13).

Near Klein Alexandershoek, Oliwenboskraal, a reddish sandstone zone with indistinct faecal pellets develops near the top of the De Hoek Member, and the white orthosandstone becomes noticeably thinner bedded higher, which may be the reason why a sharp contact between the De Hoek Member and the Graafwater Formation cannot be demonstrated anywhere in this area. The two formations probably grade into one another.

North of Lambertshoek, Oliwenboskraal, no further Piekenier Formation is exposed except for an isolated outcrop at Klipheuwel, Leipoldville. Originally classed as Peninsula, the gritty orthosandstone at Klipheuwel was reclassified because its mean cross-bedding flow direction of 103 degrees fits much better in the Piekenier flow pattern (Fig. 85) than in the Peninsula flow pattern (Fig. 90).

South of Piekenierskloof Pass, at Pietersklip, the Piekenier Formation consists of a thick coarse orthosandstone with a basal gritty vein quartz conglomerate and purple shale. The middle beds are thin-bedded ripple-marked fine-grained pinkish orthosandstone about 50 feet thick (Fig. 14; Plate 2G). The upper white orthosandstone shows an abrupt contact with the purple siltstone of the Graafwater Formation. The lower gritty orthosandstone and purple shale probably correlate with the nearby Rest Member, and the upper beds with the De Hoek Member.

The varying vertical relationships between the Rest and De Hoek Members indicate that these two members were deposited simultaneously in different areas according to local circumstances. A diagrammatical representation (Fig. 16) shows the relation between diachronous lithological member boundaries and synchronous planes (horizontal in the diagram). The facies map (Fig. 86) shows that the conglomeratic Rest Member probably wedges out to the southeast and the De Hoek Member becomes dominant; opposite relation-
ships apply to the northwest.

Towards the south, at Voëlvlei, Porterville (Plate 2A), and at Schoongezicht, Tulbagh, and at Gansekraal, Worcester (Fig. 15) the Piekenier Formation is represented by the sandy De Hoek Member. At Jasonsfontein and Slanghoek near Rawsonville and at Sybas Mountain, Breede River, lenticular coarse conglomerate with angular quartzite boulders up to 10 inches in diameter occurs in the Piekenier Formation. The formation has a very irregular thickness and is absent in many places as, for example, on the western slopes of Waaihoek Mountain and Olifantsberg. The irregular buried topography is well displayed on the northwestern slopes of Sybas Mountain opposite the footpath to the ski hut.

At Dwarsberg in Stettynskloof, and farther south and southwest the Piekenier Formation is not developed and the very sandy facies of the Graafwater Formation usually forms the basal formation of the Table Mountain Group (Fig. 2).

The Graafwater Formation

The Graafwater Formation, frequently poorly exposed, is characterised by its purple thin-bedded ripple-marked and mudcracked sandstone, siltstone and shale layers, and by its fossil content. In the sections the Graafwater Formation is indicated by the symbol $S_1$.

The type area is between Tierhoek and Klein Alexandershoek, Olievenboskraal, where the Graafwater Formation develops a maximum thickness of about 1400 feet, and can be subdivided into four lithostratigraphic members and one biostratigraphic zone. In the northernmost outcrops of the Graafwater a fifth lithostratigraphic member is recognised.

At Looprivier, Olievenboskraal, the Graafwater Formation is at least 1300 feet thick. The basal shaly member, which is about 600 feet thick, is poorly exposed but it is seen in part at Middelpos, near Looprivier, where purple shale is interbedded with fine-grained pinkish orthosandstone and shale-pellet conglomerate. In talus rubble can be seen festoon cross-bedding and primary flow lineation in the sandstone, and ripple-marks in the shale. The basal contact of the MIDDELPOS MEMBER with the Piekenier Formation can not yet be placed accurately due to a gradual decrease in
the shale/sandstone ratio from the Piekenier Formation to the Middelpos Member.

At the top of the Middelpos Member an abrupt change to the overlying sandstone member takes place. Near Looprivier this sandy member, a prominent cliff-maker, is a thin-bedded, festoon cross-bedded, fine- to medium-grained pinkish orthosandstone which is about 600 feet thick. Distinctive bedding characteristics, lithology and colour, stratigraphic position and relationships, and fossil content identifies the LOOP MEMBER.

In the Table Mountain Group the Loop Member is of special interest because it includes the lowermost fossil zone, namely arthropod (trilobite?) and worm-like tracks, trails and burrows (described in another section). At Looprivier such fossil remains occur in the uppermost 100 feet of the Loop Member. Further search may well lower the base of this biostratigraphic zone, referred to as the TRACK ZONE. Present indications are that the deposition of the Loop Member witnessed wide-spread organic activity.

The Loop Member is capped by a thick-bedded coarse white orthosandstone, about 100 feet thick at Tierhoek, and called the TIERHOEK MEMBER. In the Graafwater type area the Tierhoek Member is succeeded by an upper member similar in appearance to the Loop sandstone, called the FAROO MEMBER, and described presently.

The stratigraphic relationships in the Graafwater Formation are well displayed near Koperfonteinkloof in the Swartberg south of the pass up Klein Alexandershoek (Fig.17).

On the lower western slopes of Lambertshoek Mountain, Graafwater, fine-grained thin-bedded festoon cross-bedded pinkish orthosandstone of the basal Middelpos Member outcrops. The exposures on the slopes are insufficient to determine the complete sequence but the upper members can be recognised (Fig. 18).

At Brandenburg, Leipoldtville, purple and buff mudcracked (Plate 4D), ripple-marked (Plate 4A, E), and cross-bedded flaggy sandstone of the Loop Member develops a thickness of 300-400 feet along the eastern slopes of Grootkop. A further 200 feet of the Loop Member have been penetrated by a borehole. A remarkable track-marked bedding plane was discovered here (Taljaard, 1962), about 200 feet from the top of the Loop Member; other tracks trails and burrows occur higher in the member. The Loop Member is poorly exposed west of Brandenburg and the Track Zone has not
yet been located. The Tierhoek Member is a coarse-grained thick-bedded orthosandstone in this area. It caps Sandberg and Olifants Mountain and seems to thicken westwards.

The upper 700 feet of the Graafwater Formation are exposed at Farooskop and Laingshoogte near Graafwater. At Farooskop the Loop Member, of which about 400 feet outcrop, is thicker bedded and coarser grained than in the type area and the Track Zone has not yet been located. Enigmatic cylindrical bedding plane impressions, probably of organic derivation, occur about 45 feet from the top of the member (Plate 12G). The 100 feet thick Tierhoek Member is a coarse-grained white orthosandstone, with sparsely distributed vein quartz pebbles up to 2 inches in diameter, and with low angle planar cross-bedding laminae. The Tierhoek sandstone is succeeded by about 200 feet of thin-bedded fine-grained reddish sandstone very similar in appearance to the Loop sandstone. This is the Faroo Member. Worm trails and thin worm tubes occur from 35 to 65 feet from the base of the Faroo Member and there are also trails 12 mm wide, tubes 2-3 mm in diameter, and some arthropod-like burrows (Plate 14C,E) at about 40 to 70 feet from the top of the member.

Farooskop is capped by 30 feet of coarse-grained orthosandstone of the Peninsula Formation.

At Laingshoogte, Graafwater, thin-bedded reddish festoon cross-bedded sandstone of the Loop Member (Plate 3B) contains many thin purple shale and shale-pellet conglomerate beds. Ripple marks and primary flow lineation are common. A number of tracks, trails and burrows occur here and the Track Zone is therefore easily recognised.

At Heerenlogement, Graafwater, the upper 450-500 feet of the Graafwater Formation are exposed (Fig. 19) but a borehole at Grodtfontein, stopped in Malmesbury phyllite, indicates a total thickness of about 750 feet for the Graafwater Formation. According to the owner red mud only was drilled out, which establishes the absence of appreciable Piekenier Formation.

Near the shop at Heerenlogement festoon cross-bedded shale-pellet containing sandstone of the Loop Member grades into the 70 feet thick cross-bedded white Tierhoek orthosandstone. The Tierhoek sandstone be-
comes increasingly quartzitic near its top, where it contains vein quartz pebbles up to 3 inches in diameter. Large low angle cross-bedding laminae, and sympathetic ripple-marks are developed. Better exposures of the Loop Member farther south fail to demonstrate the Track Zone. The 235 feet thick Faroo Member is well exposed in Lekkransekloof and consists of the usual thin-bedded reddish siltstone and sandstone covered by salt efflorescence in places and riddled by thin vertical worm tubes in places. The contact of the Faroo Member with the Peninsula Formation at Lekkrans (Plate 2F) records an unusually abrupt and major change in sedimentary conditions. Locally the contact is slightly transgressive.

At Bamboes Bay near Doorn Bay the upper Graafwater Formation is exposed. A shaly Faroo Member, about 100 feet thick, and devoid of organic remains, interposes between creamy-white Peninsula orthosandstone and the Tierhoek sandstone at Poliesmangat. At Bruinpunt a wave erosion platform is cut in the upper Loop Member, which is here quite shaly in its uppermost 30 feet. The gently dipping reddish Loop sandstone abounds with shallow water primary structures such as small and large ripple-marks, mudcracks and small scale planar and festoon cross-bedding. Tracks have not yet been found but hemispherical trilobite-like burrows 6-8 inches in diameter occur in the Loop sandstone (Plate 14B).

The Bruinpunt shale is thinly bedded purple shale, interlaminated with pinkish sandstone and siltstone (Plate 3C), and it is riddled by small diameter worm tubes, which are characteristically confined to one bed at a time. A well preserved cast of a worm-like organism (Plate 18C, described in Chapter IV) was discovered in the Bruinpunt shale about 15 feet from its base.

About 20 miles east of Bamboes Bay important outcrops of the complete Graafwater Formation occur south of Bellevue, Vredendal. The Graafwater Formation at Bellevue is only 180 feet thick and it lies directly on Malmesbury phyllite. A prominent basal member from 50 to 80 feet thick (Fig. 20) consists of coarse ill-sorted orthoconglomerate with an open or disrupted framework and surrounded vein quartz, chert and quartzite pebbles 2-10 inches in diameter. The conglomerate is intimately interbedded with purple ripple-marked sandstone and siltstone,
those "trademarks" of the Graafwater Formation.

The question may be raised whether this member, referred to as the 
BELLEVUE MEMBER, should not be classed with the Piekenier Formation. 
Its character is, however, opposite to that of the supermature Piekenier 
ortho-conglomerate with its typical exotic pebbles and interbedded coarse 
orthosandstone.

The basal conglomerate is succeeded by a 25 foot sequence of purple 
siltstone with subordinate vein quartz grit, correlated with the Loop 
Member. Interbedded sandstone is characterised by 2-3 mm wide worm 
trails and faecal pellets. The Loop Member is followed by the 70 feet 
thick Tierhoek orthosandstone in which well-rounded vein quartz pebbles, 
up to 4 inches in diameter, are scattered and in places arranged vaguely 
along the bedding. The cross-bedding laminae are low angle planar sur-
faces near the base of the member, changing to smaller festoon units near 
the top. The Faroo Member is here almost identical to the Loop Member ex-
cept that it contains no organic remains. Included with the Faroo Member 
is a white soft-weathering massive gritty orthosandstone with peculiar 
small concretions.

Towards Klawer the thickness of the Graafwater Formation is reduced 
to a minimum and near Sandkloof only a few small faulted fragments remain 
(Fig. 23). At Bruinkrans, Klawer, the complete Graafwater Formation is 
only 70 feet thick (Fig. 22) but southwards it increases to 480 feet at 
Trawal (Fig. 24), to about 600 feet at Holfontein Pass (Fig. 25), and to 
about 650 feet at Kaapsekloof. Between Klawer and Kaapsekloof the 
Graafwater Formation consists mainly of purple thin-bedded siltstone and 
shale, with infrequent worm tubes and faecal pellets. The Bellevue Mem-
ber is a lenticular basal conglomerate, frequently coarse and chaotic, 
with an assortment of angular locally derived pebbles. The member 
thickens irregularly from about 2-4 feet near Klawer to about 50 feet at 
Holfontein, but farther south, near Breekwal, the Bellevue Member consists 
of 2 feet of vein quartz conglomerate followed by about 10 feet purple, 
fine-grained, petromict rudite with a disrupted framework (Plate 4C).

A lenticular sandstone in the lower third of the formation between 
Trawal and Kaapsekloof (Figs. 24, 25) probably corresponds to the Tier-
hoek Member. At Breekwal the Bellevue conglomerate is overlain by about
100 feet purple Loop siltstone and Tierhoek sandstone, which is here about 50 feet thick (Plate 4B). Organic remains are insufficiently developed to aid in more detailed and better founded stratigraphic correlation.

Southwards of the type area a continuous outcrop of the Graafwater Formation occurs on the western slopes of the Olifants River Mountains, Saron Mountain and the Elandskloof Range to Bainskloof Pass, where the formation is nearly wedged out. From Saron past Tulbagh the formation outcrops on the western slopes of the Witzenberg Range, Waaihoek and the Hex River Mountains, and it continues eastwards past Hex River Poort. The Graafwater Formation is mostly absent between Wellington, Rawsonville and Villiersdorp but it increases in thickness westwards to the Cape Peninsula outlier. The Piketberg Range is an important outlier with numerous fine exposures. Other notable outliers are Heuningberg at Porterville, and Riebeeck Kasteel.

At Pietersklip and at Dasklip Pass, Porterville, the 420 feet thick Graafwater Formation consists of a lower thin-bedded alternation of purple siliceous shale and siltstone with worm remains, and an upper sequence of soft-weathering thin-bedded pinkish sandstone which grades into the Peninsula Formation. The facies change makes direct correlation with the type area subdivisions difficult but the lower shale probably represents the Middelpos Member, and the upper sandstone the Loop Member.

Some of the more important relationships of the Graafwater members in the type area and environs are indicated diagrammatically in Figure 27.

Whereas the stratigraphy of the Graafwater Formation is vertically well differentiated in the north, lateral gradation characterises the formation in the south, with the result that the systematic subdivision of the Graafwater stratigraphy is not extended south of Porterville because such classification demands more information than presently available. The stratigraphy of a few selected localities in the Western Province is described empirically.

At Ons Rus, Wolseley, the Graafwater Formation is characterised by thin-bedded ripple-marked fine-grained sandstone with many interlaminated mudcracked purple shale beds (Plate 3F). Lenticular deposits
of similar lithology wedge out against pre-Graafwater topography at Waaihoek. At Gansekraal, Worcester, the Graafwater Formation thickens to about 200 feet (Fig. 15). The finer-grained siltstone and shale beds have developed ottrelite due to slight dynamic metamorphism. A basal thin-bedded siltstone member, which overlies the sandy De Hoek Member of the Piekenier Formation with a sharp contact (Plate 2B), is followed by a middle white orthosandstone member. An upper sequence of alternating fine-grained white orthosandstone and shale/siltstone, has yielded one trace fossil, a burrow, but no tracks.

At Dwarsberg, Stettynskloof, the Graafwater Formation is not developed although it is still present at Jasonsfontein, Rawsonville. At Kaaimansgat, Villiersdorp, the Graafwater Formation is absent, but near Wemmershoek Dam it is present in places (De Villiers et al., 1965). Northwards, along the Du Toitskloof and Hawekwas Mountains, the formation is probably absent but Groenberg, Wellington, is capped by a thin incomplete remnant of purple shale and sandstone which belongs to the Graafwater Formation.

Southwestward of a line connecting Wellington and Franschhoek the Graafwater Formation becomes more prominent and it lies directly on the unconformity.

The general character of the Graafwater Formation between Simonsberg and the Hottentots-Holland Range is well displayed in the rail- and road-cuttings at Sir Lowry's Pass, where the formation appears to be absent because thick-bedded white orthosandstone, typical of the Peninsula Formation, dominates the lower Table Mountain Group and directly overlies the unconformity. However, a number of thin purple siltstone and shale beds, some with organic remains, are interbedded with the orthosandstone beds in the lowermost 350 feet of the Table Mountain Group at Sir Lowry's Pass. The interbedded purple shale beds are also seen at Steenbras River Mouth. In Stellenbosch Mountain the Graafwater Formation is a succession of mainly white or cream orthosandstone with interlaminated purple siltstone beds, which show some worm remains. Trilobite-like tracks (Plates 18B, 19B) occur in the Graafwater Formation at Bosmanskop and Helderberg.
The Graafwater Formation in the Cape Peninsula is characterised by reddish to creamy-white thin-bedded sandstone, siltstone and subordinate shale; conglomerate is absent. The unconformity is regionally very even, but small scale irregularities, usually associated with aplite dikes in the granite, caused interesting lenticular lap-out relationships to develop during the progressive filling-in of the hollows.

The sandstones are profusely chevron cross-bedded and occasional large scale tangential troughs have developed. Two directions of stream flow almost at right angles to one another are recorded in the cross-bedding (Fig. 89R). Small normal ripple marks are common.

Slump deformation on Table Mountain Drive (Middelmost, 1967) is much larger than that on Chapman's Peak Drive, Hout Bay. In places normal loadcast structures grade into pinched-off balls to form disrupted bedding. Small as well as large scale mudcracks are developed in the shale beds, and in Chapman's Peak Drive one such bed can be traced for a few hundred feet. The original cracks are filled with sand and now superficially resemble clastic dikes, nodules or disrupted bedding due to irregular compaction deformation of the sand-filled cracks and the surrounding mud.

The Cape Peninsula exposures of the Graafwater Formation are reminiscent of the Loop Member and it may be more than fortuitous that an arthropod track (Fig. 78) occurs on Chapman's Peak Drive. The exact stratigraphic position of the Track Zone in the Western Province is not known but it is near the base of the Graafwater Formation.

The notable facies change in the Graafwater Formation between Hout Bay and Sir Lowry's Pass is in part due to an eastward increase in the coarseness ratio (sandstone/siltstone + shale ratio) from about 4:1 to about 32:1. Another conspicuous facies change dependant upon a change in coarseness ratio takes place north of the Graafwater type area where the ratio increases from about 1:1 or less at Looprivier, Oliewenboskraal, to more than 4:1 at Bellevue, Klawer.

The Peninsula Formation

In the area under discussion the original volume of the thick, homogeneous and monotonous Peninsula Formation was more than 10,000 cubic miles, of which practically all is supermature, cleanly sorted,
coarse quartz sand with occasional vein quartz pebbles the size of doves' eggs. Complete or near-complete sections of the Peninsula Formation are common and the type area is between Piekenierskloof Pass, Citrusdal, and Dasklip Pass, Porterville, for no special attribute other than near-maximum thickness (6700-4700 feet), and general accessibility. In the text diagrams the symbol $Q_1$ indicates the Peninsula Formation.

The Peninsula Formation consists of coarse-grained white orthosandstone with subordinate small-pebble openwork conglomerate beds; at Piekenierskloof Pass, for instance, such a rudite zone occurs near the middle of the formation. Whereas Piekenier conglomerate is characterised by jasper pebbles, Peninsula conglomerate from Du Toitskloof northwards frequently contains pebbles of a black and white oolitic chert. The bedding thickness of the Peninsula Formation is usually 2-6 feet and occasionally 10-12 feet, and tabular cross-bedding laminae are the almost solitary primary structure. Fossil remains are exceedingly scarce. A few isolated worm tube zones (Fig. 19; Plates 12C, F; 13D), worm trails (Plate 19A), and rare arthropod tracks (Plate 14D) are present.

Due to the lack of marker beds it is impossible to determine the stratigraphic position of isolated fragments of the Peninsula Formation. Only one indistinct and lenticular lithostratigraphic member has been recognised, namely the Slanghoek Member, which is an openwork small-pebble vein quartz conglomerate developed in the uppermost sediments of the formation in the area between Citrusdal and Cape Town. The intraformational folding below the Pakhuis Formation makes it difficult to determine the true thickness of the Slanghoek Member but it is mostly only a few tens of feet thick. The Slanghoek Member is well exposed in the Slanghoek Mountains south of Bainskloof Pass, as well as at Weltevrede, Porterville, at the Baths, Citrusdal, and on Table Mountain west of Maclear's Beacon. Present information is incomplete to utilise indications that natural subdivisions occur in the Peninsula Formation near Voëlvlei, Porterville, and at Maraisberg, Citrusdal, at Trekpoort, Clanwilliam and near Trawal, Klawer.

The lower contact of the Peninsula Formation is of various types. In the Peninsula type area the Graafwater Formation grades upwards into the Peninsula Formation by fairly abrupt increase of bedding thickness.
and decrease of the proportion of interlaminated siltstone beds. Additional features are a change in sediment colour from purple or pinkish to white, and an increased degree of induration and resistance to weathering. In the road-cuttings of Dasklip Pass and in the Cape Peninsula this gradation is clearly displayed. The contact with the Graafwater Formation can also be exceedingly abrupt, as at Lekkrans, Heerenlogement (Plate 2F). At Dwarsberg, Stettynskloof, the Peninsula Formation lies directly on granite and phyllite, and a thin basal conglomerate bed mantles the unconformity. Northeast of Klawer the Peninsula Formation lies on the unconformity for a few miles before the sandstone wedges out at Kobe Mountain. The plane unconformity is impressively even in this area, except near Matjiesgoedkloof, Urionskraal, where very slight original topography is buried by lenticular conglomerate beds (Plate 5A). On the slopes of Kobe Mountain some unusual Peninsula sediments occur, namely gritty khaki-coloured protosandstone near Oorlogsfontein, and mature, as well as immature, coarse, petromict conglomerate beds near Oorlogsfontein, at Matjiesgoedkloof, and at Kriekamma. Some of these conglomerate beds have a disrupted framework (Plate 5D), others are lag gravels (Plate 5B, C) and still others are in situ rubble deposits infiltrated by Peninsula cobblestones (Plate 5E). Rare imbrication showing long axes inclined northwards occurs in conglomerate at Kriekamma.

The upper contact of the Peninsula Formation is fascinating due to the overlying glacial Pakhuis Formation. It may be a normal, sharp concordant contact, or it may be gradational, or typically, it is a zone of complex intraformational glacial folds. Due to the genetic association these features with the Pakhuis Formation, they are described in the next section.
The Winterhoek Subgroup

The WINTERHOEK SUBGROUP typically consists of the Fold Zone, the Pakhuis Formation and the Cedarberg Formation. The member subdivision is diagrammed in Figure 62. In the diagrams the Pakhuis Formation is indicated by the symbol G, and the Cedarberg Formation by S2.

Important though subordinate down-dip changes lead to the interesting situation that two type areas are nominated for the Winterhoek Subgroup. Broadly speaking a line connecting Piketberg - Citrusdal - Wupper- tal separates the Winterhoek Subgroup with northern aspect from that of a southern aspect. The northern type are as between Vanrhynsdorp and Clanwilliam, where the upper member of the Pakhuis Formation is dominant. The southern type are as the Cold Bokkeveld Range, where the Fold Zone and the lower members of the Pakhuis Formation are important, and where the Kobe Member has undergone significant facies change to become the Steenbras Member. The lithology of the Cedarberg Formation exhibits subdued lateral changes and a systematic regional pattern is not apparent.

More than 80 profiles of the fascinating Winterhoek Subgroup were studied in some detail in this investigation; obviously most of these localities can be referred to only very briefly, in order to guide future investigators. Only field sketches are presented for some of the localities.

The FOLD ZONE is a zone of contemporaneous soft-sediment folding which is developed in the upper Peninsula Formation and which is causally related to the deposition of the Sneeukop Member of the Pakhuis Formation. The type area of the Fold Zone is in the Cold Bokkeveld between Wyekloof, Keerom, and Molenrivier, Gydo. The exposures at Bo-Rozendal, De Meul, may also be regarded as typical. The distribution, the stratigraphic depth of disturbance by contemporaneous folding, and the mean internal geometry of the Fold Zone is shown in Figure 99.

Macrofabric analysis reveals the essential character of any fold system and in the study of the Fold Zone conventional pole-plots were employed. The peculiar nature of the Fold Zone deformation causes the degree of exposure to control in part the ultimate appearance of the fabric patterns.
(Figs. 69, 83a, 101, 102) and from place to place certain parts of the fold complex may be over- or under-represented. (All fabric diagrams except those of Pakhuis and Witwater (Fig. 69) show the fold attitude prior to tectonic tilt.)

The typical Fold Zone fabric diagram pattern is a near-vertical girdle, as demonstrated at Die Drif, Grootrivier (Fig. 69; Plate 17A). A composite fabric diagram (Fig. 83a) of the area between Gydo and Wuppertal reveals an almost identical girdle. The diagrams illustrate a near-horizontal symmetrical flat-bottomed synclinal fold system in which the folds have steep, almost vertical or even overfolded flanks, "shaped like an egg with the top cut off" (Haughton, et al., 1925).

Field evidence proves the absence of flat anticlines which would produce a similar diagram pattern. Flat-bottomed synclines abound in the Fold Zone (Plates 8D, E; 9A, B, C; 15A; 16; 17A, C; Figs. 40, 41, 45, 46, 47, 48, 49, 53, 54, 55, 58, 61).

Not all diagrams show such a clear girdle as for the folds at Die Drif. If the folds are very shallow and the flanks are almost non-existent, as for example at Pakhuis, Clanwilliam, (Fig. 69), the girdle is incomplete, but the axial plane trend normal to the girdle can nevertheless be indicated (in such a case a pi-diagram is more graphic than a pole-plot). At nearby Witwater the folds are more clearly developed (Plate 8E; Figs. 40, 41) and the fabric diagram (Fig. 69) shows a more extended girdle and a very clear pi-diagram. When the exposures are such that the fold flanks dominate, two opposite maxima develop, as at Spitskop, Clanwilliam (Fig. 101A), Die Trap, Citrusdal (Fig. 101E), or Mitchell's Pass, Ceres (Fig. 102A).

Some diagrams are obscure (Vaalfontein, Cedar Mountains, Fig. 101B), or indistinct (Dwarsrivier, Cedar Mountains, Fig. 101C). They probably indicate superposition of two or more fold systems.

The intensity of deformation as well as the stratigraphic depth of deformation are maximal in the Worcester- Ceres- Cold Bokkeveld area, but both degree and depth of folding decrease westwards and northwards. When strongly deformed the Peninsula sandstone, or Slanghoek conglomerate, invariably exhibits abnormal thin-bedding (Plate 10A) which upon closer examination proves to be plastically deformed and drawn out cross-bedding.
laminae. Primary brecciation is never developed, not even in areas of most intense folding, but the Fold Zone has been affected by subsequent tectonic warping probably because it could not yield by bedding slip as did the rest of the succession. A prominent axial lineation is usually developed where the Slanghoek Member has experienced an extreme degree of folding. The contemporaneous deformation was by plastic lamellar flow in which original sandstone beds and cross-bedding laminae were attenuated to a fraction of their initial thickness. Enclosed pebbles caused linear "deformation shadows" in the matrix parallel to the direction of movement, which is also parallel to the axial trace of the fold as determined by fold fabric analysis. Both lineation and laminae are shown in Plate 10B. Neither the increased degree of folding nor the size of the folds affects the essential nature of the fabric diagrams. All sizes and types of folds occur, from small tightly folded structures (Plates 9B; 18A), to medium-sized, open or steep folds (Plates 9A, C, D; 16; 15A), to egg-shaped folds (Plate 8D; 17A), of which some are of enormous dimensions (Plate 17C).

If the axial trace directions are plotted and a gridded moving average axial trace superimposed on an isopach map of the Fold Zone, the resultant coherent pattern (Fig. 99) indicates that the folds were produced by a systematic force which operated over a large area. Stratigraphic evidence proves that this force was not tectonic and that it operated during a short geological time interval only.

Although the Fold Zone and the Sneekkop Member are closely associated, some aspects relating to their simultaneous or individual absence need be referred to. In Table Mountain, Cape Town, the Sneekkop Member is present but folds are but poorly developed, and the same applies to parts of the Elandskloof Range, Tulbagh, where folds are absent even though the complete formation is exposed. Only very dubious folds and no Sneekkop Member have been located in the Piketberg Range (Fig. 26). Although an almost continuous outcrop of the Winterhoek Subgroup stretches in practically a straight line from Piekenierskloof Pass to Klawer, the Fold Zone is essentially absent from the stretch and, especially towards the south, the Sneekkop Member is only poorly developed. At Kransvlei and Botuin, Clanwilliam (Fig. 39), at Brandberg Pass (Fig. 37), at Oskop (Fig. 36), and at Windhoek near Klawer, the Fold Zone is absent, and the Sneekkop
Member is quite thin.

The development of both the Fold Zone and Sneekkop Member is clearly subdued to the west and extreme north.

The type area of the SNEEUKOP MEMBER is between Molenrivier and Wye-kloof (also known as Kaffersdal, on the Meul River) in the Cold Bokkeveld (Figs. 53, 45, 46), but the member takes its name from Wellington Sneekkop (Plate 15A) where it was first described (Haughton, et al., 1925). The symbol G₁ denotes the Sneekkop tillite in the sections accompanying the text.

The Sneekkop Member is an unbedded pebbly quartzitic protosandstone or orthosandstone (Plates 6C, E; 8A), blueish-green when fresh, off-white when weathered. During the course of tillite fabric analysis (see description of method in a later section) abundant opalescent, rutilated, deep purple to light pink amethyst sand grains were observed in both the Sneekkop and Steenbras tillites. The apparent absence of the amethyst grains in the Kobe Member may be due to their greater dispersal in this immature argillaceous tillite. In thin-sections the colour intensity of the amethyst will be insufficient in order to recognise it from ordinary quartz (see, for instance, Visser, 1962 and Von Brunn, 1959). The amethyst has an important bearing on determining the provenance area of the tillite.

The Sneekkop Member contains abundant facetted and striated erratics (Plate 6D, F), mainly quartzite and other resistant sedimentary and metamorphic types), which are remarkable in having been well-rounded before being subjected to glacial abrasion. The erratics are frequently facetted on opposite sides and then look like flattened buns. Multiple faceting and striation on a single erratic are common.

The Sneekkop Member occupies synclinal troughs in the upper portion of the Fold Zone (Figs. 45 and following; Plates 8D, E; 9A, B; 15A; 16; 17C) and therefore has a most irregular thickness. Both tillite and folds have been truncated by a phase of intraformational erosion which preceded or accompanied the deposition of the Oskop Member and later sediments (Plates 8B; 7C, F: 9C, Figs. 36 and following). During the phase of intraformational erosion the uppermost portion of the Sneekkop tillite became water-sorted and was re-deposited locally (Plate 7C), so that the tillite may appear to grade into the Oskop Sandstone. Normally the contact of the Sneekkop Member with the Peninsula sandstone and Slanghoek
conglomerate is extremely intricate, but sharp (Plates 9B; 17C; Figs. 40, 45, 47, 49). Only at Daskop, Hangklip, is a continuous gradation developed between the Peninsula Formation (Slanghoek Member) and the Sneeukop Member. (It may be significant that the Fold Zone is absent at Daskop.)

In Franschhoek Pass the Sneeukop Member contains two abnormally interbedded sandstone beds (Fig. 61) but this is probably a local feature, perhaps due to intermittent water action towards the southern limit of the Sneeukop glacier sheet. At Sir Lowry's Pass the Fold Zone happens to be indistinct (Fig. 60), the zone is better developed towards the trigonometric beacon Bottentots Holland 17, only to disappear again near the Steenbras Filtration Works. In this area the Sneeukop Member does not occupy synclinal troughs as elsewhere but Haughton (1933) reported normal Fold Zone/Sneeukop Member relationships in Koësberg, 5 miles south of the Filtration Works.

North of its type area the Sneeukop Member can be traced almost continuously through the Cedarberg Range and from there intermittently to Windhoek, Klawer, where the Sneeukop Member is reduced to a single layer of typical erratics.

The northernmost occurrences of the Sneeukop tillite occupying Fold Zone troughs are at Spitskop, Clanwilliam, and west of Groot Kliphuis, Pakhuis Pass (Fig. 40). North of Clanwilliam the basal contact of the Sneeukop Member is concordant with the Peninsula Formation (Oskop, Fig. 36), or nearly so (Rietvlei, Fig. 38). In this area the upper boundary of the Sneeukop Member bears no evidence of the intraformational erosion which typifies the contact with the Oskop Member in the type area; the Oskop Member is, in addition, not everywhere developed (Figs. 37, 39, 43, 50, 51, 52).

The Oskop Member type area is between Wyekloof (Fig. 45) and Molenrivier (Fig. 53) in the Cold Bokkeveld, but this member takes its name from a locality 6½ miles southeast of Klawer, where its stratigraphic relationship to the two tillite members of the Pakhuis Formation was first identified (Fig. 36). The Oskop Member is not to be confused with the Hard Band of Haughton, et al. (1925), (Plates 8C; 15A; Fig. 62, 56). In the field sketches the Oskop Member is indicated by the symbol GQ.
The Oskop Member is a lenticular orthosandstone which unconformably overlies the Sneeukop Member and the Fold Zone. The Oskop sandstone is seldom thicker than 10 feet, and it is represented in places by a mere layer of pebbles (Fig. 55). Rarely the Oskop Member is concordant or pseudo-concordant with the underlying sediments (Figs. 44, 60, 36; Plate 7C). The lenticular habit of the Oskop Member is demonstrated clearly at Bergsig, Citrusdal (Agter-Bokkeveld), where the Oskop sandstone pinches out in a few yards, so that shale and tillite of the upper Pakhuis Formation rest directly on the Sneeukop Member (Plate 7A). South of Clanwilliam, at Kransvlei, the Oskop sandstone is about 10 feet thick, but at Bo­tuin, about 1 mile to the north (Fig. 39), and at Boskloof, 9 miles east of Kransvlei (Fig. 43; Plate 7B), the Oskop Member is not developed. The Oskop Member is an ill-sorted, openwork, small-pebble conglomerate at Molenrivier, Cold Bokkeveld (Plate 7F), where it contains irregular fragments of Sneeukop Tillite as well as many facetted erratics now modified by water abrasion. The Oskop sandstone is normally indistinctly or thickly bedded but cross-bedding is developed at Cedarberg Sneeuberg (Plate 7C), at Wyekloof, Keerom, and at Groenberg between Pakhuis and Wuppertal. The upper surface of the Oskop sandstone is frequently marked by asymmetrical stream ripples which from place to place show diverse flow directions; notable occurrences are at Wyekloof (Plate 8B), at Bergsig in the Agter-Bokkeveld, at Gevonden near Rawsonville and at Groot Kliphuis at Pakhuis Pass (Fig. 40).

Glacial grooves are cut into Oskop Member at Karookop, Pakhuis (Plates 15B; 17B; 11A), at Oskop and at Molenrivier (Plate 10C). The pavement at Pakhuis (Visser, 1962) has wide, deep and irregular grooves, some with enigmatic cross corrugations (Plate 11A), but thin striae on relatively flat surfaces also occur. The pavement resembles a plowed field and the appearance of the grooves indicates soft-sediment deformation. Minor structures and drag marks suggest ice movement from the northeast. The grooves at Oskop are wide and shallow, smoothly flat-bottomed with sharp ridges and fine striae, but only a few grooves are exposed. The Molenrivier pavement occurs at the very top of the Oskop Member and has fine striae on a flat surface (Plate 10C), but only a few square feet are exposed.

Following stratigraphically on the Oskop and Sneeukop Members are the
Kobe Member in the north and the Steenbras in the south. The Kobe and Steenbras Members are isochronous units in the Pakhuis Formation: they represent facies end-members of a single event. Both have separate characteristics and are easily identifiable in their respective type areas, but at present insufficient information is at hand to delimitate and analyse in detail the stratigraphy of the lateral transition which takes place between Pakhuis Pass and the Agter-Bokkeveld area. (In the text diagrams both the Kobe and Steenbras Members are indicated by the symbol G₂ and by black "erratics", but the Kobe Member is shown with a mudstone matrix, whereas the sandy Steenbras matrix is stippled.)

The type area of the KOBE MEMBER (pronounced: Koo'be) is in the Kobe Valley, 18 miles east of Vanrhynsdorp, where the succession is a thick, rudely stratified, varicoloured argillaceous paraconglomerate, characterised by abundant pebble-to boulder-sized, angular pentagonal chert and jaspilite erratics, as well as erratics of a great variety of other sedimentary non-resistant rock types. Almost all the erratics are striated, facetted and polished (Rust and Theron, 1964, Plate V, Fig. D). Rounded erratics such as typify the Sneeukop Member are very scarce in the Kobe Member. The basal contact of the Kobe Member is mostly obscured, but a striated glacial floor which is cut in Nama quartzite occurs at Bergland, Kobe Valley (Plate 10E). In the lee of the pavement the Kobe tillite rests in a valley excavated almost 100 feet deep in Nama shale and the contact is strewn with angular shale boulders (Plate 6A, B). At nearby Krieskamma (Fig. 31; Plate 72) and in places at Witkleigat (Fig. 32), the Kobe Member rests conformably on the Peninsula Formation, while on the western slopes of Kobe Mountain, at Matjiesgoedkloof, a thin varved mudstone with rafted pebbles forms the base of the Kobe Member (Fig. 33). To the south and to the east the Kobe Member normally lies concordantly on the Peninsula Formation, as at Oorlogsfontein (Fig. 34), at Kuilen, at Saulesfontein on the Matsikamma Mountain (Fig. 35), and at Giftberg, Vanrhynsdorp. South of Windhoek, Klawer, the Kobe tillite is underlain first by the Sneeukop tillite and then by the Oskop sandstone (Fig. 36).

The contact of the Kobe Member with the Soom Shale is gradational over a few feet, and invariably, where this elusive contact is exposed, as, for example, in the road from Matjiesgoedkloof to Blaauwpoort, Urions-
kraal, the Kobe tillite becomes laminated and interbedded upwards with thin-bedded mudstone in which rafted pebbles (Fig. 68) become gradually smaller and fewer so that eventually a thinly laminated micaceous shale, probably black when fresh but yellow when weathered, continues upwards as the Soom Shale. The placement of the upper contact of the Kobe Member is probably an academic question because this portion of the Kobe stratigraphy is hardly ever exposed, but it should be taken at the highest bed in the transition sequence which still contains rafted clasts, and which thereby indicates the presence of some ice, albeit floating, during Kobe sedimentation.

The Kobe Member is normally about 70-100 feet thick in the Kobe Valley (Figs. 29-32) but it wedges out northwards near Klein Kobe (Fig. 28) and it thickens to about 250 feet at Ondertuin at the southern extremity of the Kobe Valley.

The type area of the STEENBRAS MEMBER is between Wyekloof and Molenrivier in the Cold Bokkeveld (Figs. 45, 46, 53) but it takes its name from the Steenbras River near which it is exposed in Sir Lowry's Pass (Fig. 60).

The Steenbras Member is a dark blueish-green, unbedded or poorly bedded, quartz-rich pebbly subgraywacke or protosandstone which is called a tillite because of its faceted and striated erratics, its relationship to glacial floors at its base (referred to in the section of the Oskop Member) and because of its fabric. The Steenbras tillite is distinguished from the Sneeukop tillite, which it superficially resembles, by its darker colour; by its greater percentage clay matrix (which causes the Steenbras tillite to weather more easily); by the smaller proportion of erratics, which, though mostly quartzite and vein quartz, include a notable proportion of less-resistant types; by the smaller but much more regular thickness of the Steenbras tillite and lastly by its unconformable relationship to the Fold Zone. The upper contact of the Steenbras Member is seldom seen and the tillite is probably gradational into the Cedarberg Formation. At Sir Lowry's Pass (Fig. 60), Wellington Sneeukop (Fig. 56) and at Kweekkraal, Villiersdorp, the upper contact is, however, sharp.

North of Kweekkraal, Villiersdorp, the dark blueish-green Steenbras tillite persists for 80 miles continuously to Wyekloof, Keerom. However, thirteen miles north of Wyekloof, at Kunje and at Bergsig, a major facies
change replaces the Steenbras tillite by a sequence of thin-bedded shale, rafted shale and bedded argillaceous paraconglomerate, which will be referred to informally as the Kunje Sequence.

At Bergsig the Kunje Sequence follows on an uneven surface (Plate 7A) which is underlain by Sneeukop tillite and ripple-marked Oskop sandstone. The hollows are filled in by finely laminated shale which features abundant rafted sand grains. The basal shale grades upwards into tillite which, though bedded and distinctly argillaceous, nevertheless has some affinities to the Steenbras tillite. A second similar tillite occurs a few feet higher.

At nearby Die Drif (Fig. 50) the Kunje Sequence is not exposed but its presence is deduced, amongst others, from numerous weathered out erratics, here mostly Kobe-type chert, pebbles, so that the Kunje Sequence at Die Drif appears to have affinities to the Kobe Member.

The typical Kunje Sequence is closely repeated at Grootrivier about 10 miles east of Bergsig (Fig. 51). The tillite is a sandy yellow mudstone with strong Kobe affinities, but it is vaguely bedded in unites 2-4 inches thick and it contains many small rafted pebbles, a few up to 2 inches in diameter.

The basal shale of the Kunje transition sequence seems to extend northward into the area occupied by the Kobe Member because at Boskloof, Clanwilliam (Fig. 43; Plate 7B), an otherwise normal Kobe Member commences with a basal shale.

Visser (1962) reported the presence at Pakhuis of two mudstone tillite beds which are interbedded with a laminated rafted mudstone. In this dissertation this entire sequence is classed as Kobe Member. In addition, however, a thin sandy tillite bed is found to lie directly on the glacial pavement. It represents locally eroded sand re-deposited from the extreme sole of the glacier which made the grooves, and it should not be confused with the Sneeukop Member.

The composite dispersal map of the Pakhuis Formation (Fig. 95) combines all vectorial data from glacial striae and fabric analyses (Figs. 96, 97, 98) for all the tillite members. Not shown on the map but documented in the outcrop descriptions is the notable increase in petrographic maturity associated with the transition from the Kobe tillite to the Steenbras tillite, presumably in the direction of transport. As expected for tillites
so closely associated in time, though in part not so petrologically, no significant difference in ice flow direction for the various tillites can be detected. The flow pattern is practically identical to the fold axial trace map (Fig. 99) and the deduction is inescapable that the folds formed not transversely but parallel to the direction of glacier flow.

The type section of the CEDARBERG FORMATION is at Langvlei, Porterville (Fig. 63). The Cedarberg Formation consists of shale, siltstone and fine-grained sandstone beds, which are yellow when weathered but black, carbonaceous when fresh. The Langvlei type section also contains a biostratigraphic unit, the Brachiopod Zone. A notable exposure of the Cedarberg Formation occurs east of the Steenbras Filtration Works (Plate 11E shows part thereof) but here the Brachiopod Zone is not developed. In the field sketches the Cedarberg Formation is shown by the symbol $S_2$.

The SOOM SHALE MEMBER is the lower and lesser member of the Cedarberg Formation. It is named after The Soom in Matsikamma Mountain, Vanrhynsdorp (Plate 21C), which is the soft slope underlain in part by the Cedarberg Formation. The Soom shale is a thinly laminated micaceous shale, which is very seldom exposed. The lowermost beds of the Soom shale are commonly shale interbedded with rafted mudstone beds (Fig. 68). Downstream of the Bulshoek Barrage, Clanwilliam, about 65 feet of varved beds are exposed in the basal Cedarberg Formation, but the precise stratigraphic relationship is obscured. The Bulshoek varved sequence is probably a lentil in the Soom Member because the varves do not occur in the next nearest outcrops (Plate 7E, Figs. 37, 38, 62). The Soom shale can be traced from Tierberg, Lokenburg, to Sir Lowry's Pass (Figs. 29, 43, 60, 62). Good exposures of the Soom shale occur at Matjiesgoedkloof, Uriosn kraal, at Langvlei, Porterville (Fig. 63), at Gevonden, Rawsonville, at Kweekkraal, Villiersdorp, as well as next to the Toll House in Mitchell's Pass and in a road-side quarry at Bruinkrans, Het Kruis. Near Keurbos, Clanwilliam, the Soom Member is a fine-grained ash-white laminated shale about 25-30 feet thick.

A prominent orthosandstone bed, known as the Hard Band (Haughton, Krige and Krige, 1925), occurs in the Soom Member at Wellington Sneeukop Plates 15A; 8C; Fig. 56). It is here renamed the HARD BAND LENTIL (Fig. 62) for though it persists eastwards to Slanghoek Piles, it wedges
out on the northwestern slopes of Vaalkop (Lower Sneeukop). The Hard Band Lentil may be present at Tierkop, Bainskloof Pass (Fig. 54), but at Zuurfontein 230 in the Elandskloof Range and northwards, the Hard Band Lentil is definitely absent.

The DISA SILTSTONE MEMBER accounts for the bulk of the Cedarberg Formation. It is a thin-bedded argillaceous and carbonaceous siltstone and fine-grained protosandstone, and takes its name from the Disa Pool in the Cedar Mountains, near which rare fresh outcrops occur at Swartwatervalkloof. Apart from the type section at Langvlei, Porterville, easily accessible exposures of the Disa Member occur in a quarry at Modderfontein, Piekenierskloof Pass, in Franschhoek Pass and at the Steenbras Filtration Works, Gordon's Bay; the Disa Member is commonly well exposed in mountainous areas.

In the type section three zones are identified (Fig. 63). The basal zone is a thin-bedded fine-grained sandy siltstone, apparently devoid of macrofossils, and wanting primary bedding structures. The central zone is a more massively bedded, bioturbated fine-grained siltstone and mudstone, crammed with Brachiopod shells (see a later section). In the type section the Brachiopod Zone is confined to the central lithostratigraphic unit of the Disa Member. The upper zone is a fine-grained laminated protosandstone in which the bedding thickness increases upwards from about 6 inches at the base to about 12 inches at the top. The contact between the Disa Member and the Nardouw Formation is gradational due to appreciably improved sorting in the sediment and a further doubling of bedding thickness. This transition is clearly displayed at a number of localities, notably at Mitchell's Pass, where the Disa Member is abnormally coarse-grained, and at Franschhoek Pass, where the siltstone appears to be as much as 325 feet thick, and on the Doorn River, Klawer (Plate 11D), where the Disa Member consists mainly of micaceous subgraywacke.

The Brachiopod Zone is exposed in typical circumstances near Berghof, about 5 miles north of Langvlei. The fossiliferous zone occurs elsewhere on more or less the same stratigraphical horizon, and it is exposed in a bioturbated mudstone south of the historic Witzenberg Pass, Tulbagh, and in a fine-grained proto sandstone at Wellington Sneeukop. Reasonably
good exposures of the Disa Member are common north of Clanwilliam (Plate II C, D) but fossil remains are limited to indistinct worm-like tubes, trails and faecal pellets.

North of the type area the Disa siltstone is characterised by ripple-mark cross-bedding (Plate II B) and a conspicuous mica content. Sole structures are normally very indistinct but load-casting balls occur in places, notably at Modderfontein, Fiekenierskloof Pass.

East of the Filtration Works at the Steenbras Dam only the lower half of the Disa Member is a siltstone; the upper half is a thin-bedded (4-12 inches), fine-grained micaceous sandstone, dark gray when weathered, which contrasts sharply in contact with the cream-white Nardouw Formation sandstone (Plate II E).

The Nardouw Formation

The Nardouw Formation consists of orthosandstone nearly identical to that of the Peninsula Formation, but the Nardouw Formation can be subdivided into more members and it is more fossiliferous than the Peninsula Formation.

The Nardouw Formation consists of the basal **GOUDINI MEMBER**, first identified in the Slanghoek Mountains by Jansen in 1951, and the middle **SKURWEBERG MEMBER**, and the upper **RIETVLEI MEMBER**, initially recognised as a separate member by Swart (1950). The presently vague definitions of the member boundaries do not prevent the recognition of the members as separate entities of the Nardouw Formation.

The type area of the **GOUDINI MEMBER** is at Kliphoutskloof, Rawsonville, where this orthosandstone member is characterised by its reddish colour upon weathering and by its thinner bedding and finer grain than the rest of the formation. Neither the basal contact of the Goudini Member with the Cedarberg Formation, nor the upper boundary with the Skurweberg Member can at present be satisfactorily indicated and more detailed work is necessary before these limits can be fixed accurately. The typically grading contacts and interbedded purple siltstones are especially well exposed in Franschhoek Pass, and at the Steenbras Filtration Works (Plate II E), and to a lesser extent in Mitchell's Pass. In the type area the member is about 100-150 feet thick if topographic expression is any indi-
cation of its true thickness. In Franschhoek Pass, where the member is completely exposed, it is impossible, without additional information, to decide on its true thickness; it may be less than 100 feet.

Northwards of the type area the Goudini Member retains its character as far as Klawer, although in places it can be recognised only with difficulty. It may eventually prove advantageous to subdivide this member and institute two or more lateral members in its place.

At Molenvrij, Gydo, and at Horingkloof, Groot Winterhoek, the Goudini Member is about 250 feet thick and it contains scattered worm trails and ripple marks. At Langvlei, Porterville, the upper siltstone beds of the Cedarberg Formation grade fairly rapidly into yellowish or reddish protosandstone and cream-white orthosandstone beds of the Goudini Member, which are characterised by bedding thickness of 1-2 feet and by inconspicuous cross-bedding. Reddish micaceous flagstone beds are interbedded in the medium-grained orthosandstone which constitutes the bulk of the Goudini Member at Wyekloof, Keerom. Woolsack-like weathering characterises the white orthosandstone of the sandy Goudini Member near Die Trap and at Kunje, Citrusdal. Near Citrusdal the Goudini Member is almost indistinguishable from the Skurweberg Member but at Piekenierskloof Pass worm trails occur about 180 feet from the base of the Nardouw Formation. Similar trails characterise the top of the Goudini Member farther northwards, and the trails at Piekenierskloof Pass accordingly indicate the boundary between the lithologically similar Goudini and Skurweberg Members.

In the Vanrhynsdorp area many good exposures of the Goudini Member (Fig. 35) demonstrate a different field appearance from that at the type area but the stratigraphical correlation is not in doubt. In the Kobe Valley and environs the member is an ash-white orthosandstone, which is sharply contrasted with the buff sandstone above. The colour difference is mainly due to lichen, which may point to a significant trace element variation in the rocks of the two members. The Goudini Member bedding thickness is about 2-6 feet, which is more or less similar to that of the lower Skurweberg beds. For no apparent reason the Goudini sandstone weathers to woolsack-like, rounded exposures (Fig. 34) which form a subdued bench below the precipitous cliffs of the Skurweberg Member. The Goudini Member is characterised by two additional features namely by a conspicuous blueish-gray
siltstone bed at its top and, closely associated with the siltstone bed, by worm trails (described in Chapter IV). The siltstone and trails are well displayed in the road up Tierberg (Fig. 29), and at Boskraal and Witkleigat (Fig. 32) in the Kobe Valley. In the Kobe area the Goudini Member is about 70-80 feet thick but it reduces to 10 or 15 feet at Rooikloof, Lokenburg, where it directly overlies Nama graywacke. The absence of a basal conglomerate here is notable. The member wedges out abruptly north of Rooikloof and it is absent from Driefontein northwards.

The type area of the SKURWEBERG MEMBER is in the Gydo area, where it is responsible for the rugged Skurweberg Range. The Skurweberg is characterised by thick-beded coarse-grained white orthosandstone, which is profusely cross-bedded and which is fossiliferous in places. The grain size of the Skurweberg sandstone increases northwards. The lower and upper contacts of the Skurweberg Member are subtle and indistinct but the transition is notable in the field and in places it is quite abrupt. Due to its thick-bedding and good induration the Skurweberg Member is topographically the most prominent member of the Nardouw Formation – this is strikingly displayed in Matsikamma Mountain and in the Bokveldsberg Plateau near Vanrhynsdorp (Plate 21C, D).

The bedding thickness of the Skurweberg sandstone is normally 2-6 feet but it is frequently more than 10 feet, as for example at Widouw Pass, Vanrhynsdorp, and at Lelikloof, (Doringbos) (Plate 20A). To the south the bedding thickness attenuates gradually to about 3-4 feet at Wellington Sneewkop and at Daskop, Hangklip. The bedding surfaces are normally very even and flat, as for instance at Rozendal, De Meul, and they are only infrequently rippled, as at Modderfontein and Marcuskraal near Citrusdal.

Two types of cross-bedding are common. Ubiquitous planar cross-bedding (Plate 20B) is well exposed almost everywhere, but notably at Steenbras Dam, at Mitchell’s Pass, at Groot Kliphuis east of Porterville, and at Bodrif, Nardouw.

Very large planar cross-bedding laminae have been traced for 180 feet along strike at Wolwepunt north of Nieuwoudtville. Subordinate large scale lenticular cross-bedding (Plate 20D) appears to be more abundant north of Clanwilliam. Slumped laminae are very scarce in the Skurweberg Member.

In the Western Province the Skurweberg sandstone is a medium-
coarse-grained orthosandstone but it becomes noticeably coarser northwards. North of Clanwilliam small-pebble vein quartz conglomerate beds are common and in the Nardouw Range and at Lokenburg and Boskloof near Nieuwoudtville supermature rudite forms zones up to 100 feet thick (Fig. 65). The Skurweberg conglomerate is similar to the mature and supermature conglomerate of the Peninsula and Pieskenier Formations, and at Vanrhyns Pass and Wolwepunt, Nieuwoudtville, quartzite pebbles and boulders up to 10 inches in diameter occur in the basal conglomerate beds. Flat, biscuit-shaped shale and argillite pebbles 7 inches or more in diameter occur in the basal conglomerate in Vanrhyns Pass; the pebbles are probably of Nama derivation and have not been transported far.

From Driefontein, Nieuwoudtville, northwards the Skurweberg Member lies directly on the unconformity, which is a most impressively flat surface cut in Nama sediments and which is exposed in many places below the precipitous Bokveldsberg scarp (Plate 21D).

Fossil remains in the Skurweberg Member are limited to a variety of worm trails and tubes which are sparsely exposed from Witkleigat in the Kobe Valley in the north, to Daskop, Hangklip, in the south (Plate 12D).

The type area of the RIETVLEI MEMBER, which is the uppermost member of the Nardouw Formation, is between Wuppertal and De Meul, and it is named after a locality 12 miles south of Wuppertal, where its upper and lower contacts are clearly exposed and where most of its typical characteristics are displayed. The Rietvlei Member occurs throughout the investigated area except to the north of Oorlogskloof, Nieuwoudtville, where it has been removed by pre-Dwyka erosion. Between Boskloof and Lelikkloof, south of Nieuwoudtville, the Rietvlei Member is about 300-365 feet thick (Fig. 65) and it thickness increases to about double this value at Blaaskloof, De Doorns.

The Rietvlei Member is a white orthosandstone, characterised by a bedding thickness of about 1-2 feet (which is notably less than the bedding thickness of the Skurweberg Member) and the sandstone is feldspathic and finer-grained than the Skurweberg sandstone (Swart, 1950). Large scale festoon-type cross-bedding shows slumped laminae in places (Plate 20E). The sandstone is not well indurated and outcrops are generally poor. Some worm tubes occur in the Rietvlei Member (Plate 13A, BC; Figs. 65, 66).
At Rietvlei, Wuppertal, the sharp basal contact of the Rietvlei Member with the Skurweberg Member is marked by a low but prominent dip slope cliff. (A conspicuous terrace which is here underlain by Rietvlei sandstone may be an abandoned valley floor of the Moordenaarsgat River, which has now been incised along the Rietvlei/Bokkeveld contact.) Elsewhere the basal contact of the Rietvlei Member is usually topographically more subdued, but it nevertheless has some topographical expression, as for instance at Gydo Pass, Ceres, and near De Doorns, as well as south of Brandvlei, Worcester. The upper contact with the Bokkeveld shale, subgraywacke and mudstone is nearly always abrupt, but this contact is not frequently exposed; some notable exposures are at Hoogte near Rietvlei, at Witkrans north of Wuppertal, at Lelik Kloof, Doringbos (Plate 21A) and at Grootrivierhoogte, Grootrivier (Plate 21B). Rare primary flow lineation and ripple marks occur in the upper thin-bedded Rietvlei Member at Lelik Kloof, Doringbos.

A gradual passage from the Rietvlei Member to the Bokkeveld Group in the road-cuttings of Gydo Pass makes it difficult to decide on the precise location of the contact in the transition zone of interbedded white, fine-grained orthosandstone and carbonaceous Bokkeveld-type mudstone; probably the uppermost white orthosandstone bed would be the most practical and useful lithostratigraphical marker of the boundary.

A small-pebble conglomerate about halfway in the Rietvlei Member has been identified over a distance of 80 miles between the Kranagat River, Lokenburg, and the Groot River, Grootrivier. The conglomerate is characterised by angular to subangular gray stromatolitic chert fragments which are set in a sandy matrix. The pebbles are normally about 1/4 inch in diameter, but at Loerkop, Nieuwoudtville, the clasts are up to 12 inches in diameter. This conglomerate may eventually prove to be a useful time marker horizon.

Vertical worm tubes (described in the next chapter) have as yet only been located north of the type area but this may be due to the commonly poor outcrops of the Rietvlei Member. It is interesting that these tubes are in or near to gritty sandstone in some places, notably at Lokenburg and at Kerskop, Wuppertal (Plate 13A, C).
FOSSILS IN THE TABLE MOUNTAIN GROUP

Introduction

Fossils have always been and still are very scarce in the Table Mountain Group but for the first time sufficiently diagnostic types have been found to date the Table Mountain rocks independently, and to determine the depositional environment conclusively. A restricted, dominantly faunal assemblage occurs mainly in the Graafwater Formation, the Peninsula Formation, the Cedarberg Formation and in the Nardouw Formation. The Piekenier Formation has yielded only very dubious and indistinct faecal pellets in the Piketberg Range, whereas the Pakhuis Formation has as yet yielded no organic remains. More than 70 fossil sites have been located in the Table Mountain Group.

Four types of organic remains occur in the Table Mountain Group:

a) Worm-like tubes or burrows, and trails, faecal pellets and other markings;

b) Arthropod tracks and burrows;

c) Brachiopod shells;

d) Plants.

Tubes, trails and pellets

A considerable variety of markings are ascribed to worms or worm-like forms (Scolithos and ?Helminthoida) mainly on the grounds of the sinuous trails, and the vertical tubes ("pipe-rock") and other burrows in which the organisms presumably lived and fed (Seilacher, 1967).

The trace fossils are classified according to size. The smallest are 2-3 mm across and are rather common; the next larger forms are about 5-6 mm wide and the largest structures are about 12 mm across; a single burrow 35 mm wide is the very largest trace fossil of this type which occurs in the Table Mountain Group.

Three millimetre tubes occur only as vertical linear tubes which do not appear to be U-shaped. The tubes are invariably confined to a single bed at a time (there is no preference for a particular rock type) and
riddle the bed in their hundreds, destroying almost always all primary sedimentary structures (Plate 12A, C, D, F). The tubes are structureless and are frequently filled with material of different texture and colour than the host rock. The tubes are probably feeding as well as dwelling burrows which were dug only in freshly deposited and presumably nutritious sediment. The tube filling may be a faecal deposit.

Three millimetre "pipe-rock" is developed in the Faroo Member of the Graafwater Formation at Bruinpunt, Doorn Bay (Plate 12A), and at Heerenlogement as well as at Weltevrede, Aurora. The tubes are confined to the lower Graafwater Formation at Dasklip Pass, Porterville, and elsewhere tubes are developed only sporadically in the formation.

About 650 feet above the base of the Peninsula Formation at Heerenlogement, Klawer (Fig. 19), occurs a 20 feet thick zone of tube-riddled coarse-grained orthosandstone with slumped cross-bedding laminae. The tubes are up to 1200 mm long. A similar "pipe-rock" zone about 5 feet thick occurs 550 feet higher in a conglomeratic orthosandstone which contains pebbles up to 3 inches in diameter. It is improbable that two similar tube zones at Skimmelberg, Olievenboskraal, can be correlated with the Heerenlogement zones because the stratigraphic separation of about 1000 feet at Skimmelberg is almost twice that found at Heerenlogement, and in addition the stratigraphic position at Skimmelberg is uncertain because the base of the Peninsula Formation is not exposed here. At Boekenberg, Redelinghuys, a 12 inches thick tube zone near the summit of the hill overlies at least 1500 feet of apparently barren orthosandstone. The Boekenberg zone is also associated with conglomeratic sandstone which contains scattered well-rounded pebbles up to 4 inches long. Tubes are scarce south of this area and the only presently known exposure is in Bainskloof Pass about halfway or higher in the formation. This zone was not encountered in a traverse up nearby Sneeukop.

A tube zone about 120 feet thick in which individual 3 mm tubes are up to 1200 mm long, occurs at Kalmberg, Redelinghuys, in an isolated outcrop of thick-bedded bioturbated sandstone and conglomeratic, cross-bedded orthosandstone with pebble washes and erosion troughs. The stratigraphic correlation of the "pipe-rock" sequence is in doubt because the coarseness of the sandstone favours the correlation of the Kalmberg sequence with the Piekenier Formation but no tube zone is known anywhere in this
formation. The Graafwater Formation, though containing many beds with 3 mm tubes, is lithologically quite unlike the Kalmberg rocks, and the lithologically similar Skurweberg Member of the Nardouw Formation lacks common development of the thin tubes; correlation with these two formations is therefore untenable. By means of this process of elimination the Kalmberg sequence is provisionally correlated with the upper, coarse-grained Peninsula Formation (Slanghoek Member?).

Southeast of Vanrhynsdorp 3 mm tubes occur sporadically in the uppermost Peninsula Formation. Near Corlogsfontein, Urionskraal, tubes in a fine-grained pinkish orthosandstone are about 300-375 mm deep with a maximum depth of 600 mm in a zone which varies in thickness from 1-20 feet (Fig. 34). At Saulsfontein on Matsikamma Mountain, tubes riddle a white orthosandstone bed 6-10 feet thick (Fig. 35). The tube zone is sharply truncated by a thin cross-bedded unit devoid of organic action. At Matjiesgoedkloof, Urionskraal, the northernmost conglomeratic lenses of the Peninsula Formation contain interbedded tube-riddled orthosandstone beds. At Krieskamma, Lokenburg, the closely packed 3 mm tubes are up to 450 mm deep in a zone from 5 feet to 25 feet thick (Plate 12F). At nearby Koringland, however, no trace of any organic action is recorded (Plate 7D).

It appears that the tube zones are not reliable marker horizons in the Peninsula Formation.

Some indistinct 3 mm tubes occur in the Disa Member of the Cedarberg Formation at Windhoek, Klawer, and near the Bulshoek Barrage, Clanwilliam.

In the Nardouw Formation the very thin tubes are uncommon. The most important occurrence of 3 mm "pipe-rock" is at Daskop, Hangklip, where a profusely tubed zone a few feet thick (Plate 12D) is developed about 150 feet above the base of the Skurweberg Member. The tubes were originally normal to the bedding but are now slightly displaced tectonically in the manner of axial cleavage. The only other locality where 3 mm tubes occur in the Nardouw Formation is at Lelikkloof, Doringbos, where insignificant tubes, trails and pellets occur in the Rietvlei Member about 300 feet below its top. Associated with the "pipe-rock" are unidentified flat circular impressions 25 mm to 50 mm in diameter.

An organic structure resembling a double-headed match stick (Plate
12B) is frequently associated with 3 mm tubes in the Graafwater Formation. The "sticks" lie scattered on the bedding surface and seem to favour a muddy environment; their significance is obscure but the structure could have been formed by a worm leaving the surface exit of one burrow and entering a new burrow a few millimetres away. The structure is particularly common at Bruinpunt, Doorn Bay, as well as near Klawer and Trawal but it occurs only indistinctly at Grootkop, Eendekuil. Similar structures are developed in coarse orthosandstone in the Peninsula Formation at Table Mountain, Cape Town. Talus blocks along Table Mountain Drive are riddled with short stick-like impressions which range from 1 mm x 12 mm to 5 mm x 100 mm; no detail can be discerned in the impressions.

Three millimetre trails in the Graafwater Formation are usually sinuous, cross one another and have a semicircular cross-section (Plate 13G). The trails are widespread and are frequently associated with other evidence of organic activity, but not necessarily with 3 mm tubes. In the upper Peninsula Formation such trails occur as serpentine burrows on cross-bedding laminae at Schoongezicht, Tulbagh, at Franschhoek Pass and as far east as Swartberg, Prince Albert. These particular structures are feeding burrows along the cross-bedding laminae where nutrients were probably concentrated. In Widouw Pass, Vanrhynsdorp, an unusual V-shaped sinuous trail (Plate 13F) occurs in medium-grained cross-bedded orthosandstone about 200 feet above the base of the Skurweberg Member. The delicately grooved trace fossil is flanked by small but conspicuous rounded ridges.

Tubes larger than 2-3 mm in diameter are scarce. At Kerskop Pass, Wuppertal, a coarse-grained cross-bedded orthosandstone bed in the lower Rietvlei Member is perforated by numerous funnel-shaped and indistinctly ribbed tubes about 6 mm in diameter (Plate 13A,C). The funnels flare out abruptly at the top to about 20 or 25 mm in diameter (Fig. 66). The tubes cut through the cross-bedding and are from 450 to 600 mm deep. No U-shaped tubes or bottom cavities were observed, and the tubes are now filled with sandstone which weathers out easily. Similar tubes occur in the nearby Witteberg Group (Swart, 1950). At Lokenburg the tubes are up to 10 mm in diameter and 250 mm deep in coarse-grained pebbly orthosandstone of the lower Rietvlei Member (Fig. 65). A similar "pipe-rock" zone occurs about 200 feet below the Bokkeveld Group at Lelikkloof, Doringbos, and at Tengieterskloof, a few miles west of Lelikkloof, indistinct tubes occur
about 400 feet below the Bokkeveld Group. These individual tube zones probably belong to the same biostratigraphic zone in the Rietvlei Member between Wuppertal and Lokenburg but present stratigraphic control is inadequate to present a detailed correlation.

Large tubes occur in the lowermost Peninsula Formation at Kaapsekloof, Klawer, where indistinct 12 mm tubes are associated with equi-sized trails. At Matjiesgoedkloof, Urionskraal, 12 mm vertical tubes are developed in interbedded sandstone lenses in the Peninsula Formation, which is locally conglomeratic; the tubes at Matjiesgoedkloof occupy a stratigraphic position at the top of the formation. Nearby but not associated with the vertical tubes, are remarkably serpentine and spiral 12 mm diameter burrows in white orthosandstone (Plate 13D).

A probable U-shaped 12 mm tube about 125 mm deep occurs in the Faroo Member in Farooskop, Graafwater (Plate 14C). This tube-like structure is associated with many trilobite burrows (Plate 14C,E) and the possibility is not excluded that it may represent a cross-section through a trilobite burrow; however, the general shape of the burrows makes it unlikely that the U-shaped structure was excavated by a trilobite.

Six millimetre and 12 mm trails are large enough so that some internal details may sometimes be seen. Some of these trace fossils are true trails made on a surface of soft sediment (Plate 13D), others are feeding burrows, dug in soft sediment near or below the surface and were then filled with faecal deposit. Both trails and burrows are characterised by a distinct longitudinal median ridge and, in the adjacent grooves, by closely spaced track-like marks (Fig. 67) probably caused by cilia or similar locomotive appendages. The trails and burrows range from reasonably straight to extremely serpentine.

The first fossil zone which was recognised in the Table Mountain Group is characterised by 6-12 mm worm trails and occurs at the contact between the Goudini and Skurweberg Members of the northern Nardouw Formation. It was suggested in Chapter III that this biostratigraphic zone can be used to determine the contact between the otherwise similar Goudini and Skurweberg Members at Piekenierskloof Pass. Trails are also found at other stratigraphic levels in the Nardouw Formation as for example in about the upper third of the Skurweberg Member at Tengieterskloof,
Doringbos (Plate 13B), and near Witkleigat, Lokenburg. A variety of indistinct bedding markings as well as the familiar 6 mm trails occur at Molenrivier, Gydo, about 70 feet above the base of the Goudini Member. At Tierkop, Bainskloof, sinuous 6 mm trails in the basal Nardouw Formation include the normal indented and ridged type as well as a poorly preserved convex rounded trail with a vague median ridge.

The largest trail in the Table Mountain Group occurs in the uppermost Peninsula Formation at Swartberg Pass, Prince Albert. This trace fossil is a surface burrow 35 mm wide which shows the median ridge and adjacent "track marks" (Plate 19A; Fig. 67). The serpentine trail crosses itself and from the relationships of the faecal deposit which fills the burrow it is possible to determine the original sense of movement of the organism.

At Waterval, Oliwenboskraal, 12 mm trails and burrows (but no vertical tubes) occur in the Loop Member of the Graafwater Formation. Whorled and serpentine feeding burrows riddle some of the beds (Plate 13E). Surface feeding burrows are common between Graafwater and Tierhoek, Oliwenboskraal. Arthropod tracks are associated with the worm-like trails at all the localities.

A single impression which shows some detail of a "worm" was found in the Loop Member, Bruinpunt, Doorn Bay (Plate 18C). The cast occurs on the underside of a bedding plane and was originally impressed in purple mud and was then infilled with sand. Three incomplete impressions are preserved: the longest is 195 mm long and is straight, a second piece is 70 mm long and curves through about 80 degrees, and the last fragment is 40 mm long and is straight. The impressions are 7 mm wide and have a relief of about 1.5-2.0 mm. Prominent cross "corrugations" spaced at 2 mm intervals are indented with a longitudinal median groove (an original ridge). One end of the longest fragment is slightly flattened as if this could be a natural termination, but the preservation on this critical portion is so poor that any such interpretation (head? tail?) is unwarranted. The cast does not look like a crinoid stem but the impression bears some resemblance to Beaconites antarcticus (Vialov, 1962). The possibility that the Bruinpunt impression is a trail or burrow and not a cast of an organism is not excluded, though the fragmentary nature of the fossil speaks against such
an interpretation, but the resemblance of the impression to a contemporary deep-sea trail, 20 times larger (Vyalov, in Ager, 1963), is notable.

Short cylindrical impressions of uncertain generic significance occur at Farooskop, Graafwater, in the Loop Member of the Graafwater Formation. The casts are about 6 mm in diameter and 25-35 mm long and are scattered on the lower bedding surface of a fine-grained, cross-bedded sandstone (Plate 12G). Similar impressions at Die Drif, Grootrivier, are aligned parallel to the paleoflow of the Skurweberg Member.

A number of scattered crescent-shaped impressions of doubtful origin occur together with arthropod tracks on a bedding plane slab from Brandenburg, Redelinghuys. The markings are 15-35 mm across and are deepest impressed at the rounding so that the two arms flare out and shallow towards the ends (Figs. 79, 72). The central area is slightly humped in the larger specimens. This may be a rest-mark left by a swimming organism because no tracks lead to the impression, but it may also be an inorganic structure (Letter from Professor Seilacher, 1967).

On the same slab are two circular medusa-like rest-marks, 100 mm and 170 mm in diameter. No detail is visible except for the central raised mound and a very faint peripheral ridge (Fig. 73). A third circular structure consists of two concentric grooves (Fig. 75). The significance of these structures is obscure.

The very common subspherical or elongated pellets which occur profusely with signs of organic activity are considered faecal pellets (Plate 12E).

Arthropod tracks and burrows.

The first discovery of arthropod tracks in the Table Mountain Group at Brandenburg, Redelinghuys (Taljaard, 1962), has been succeeded by finds at Tierhoek and Waterval, Olievenboskraal, as well as at Graafwater and in the vicinity of Stellenbosch and Hout Bay, Simonstown. In these localities the tracks occur in the Graafwater Formation but tracks have also been found in the uppermost Peninsula Formation at Matjiesgoedkloof, Urionskraal. No exoskeleton has yet been found with the tracks. However, in the Brachiopod Zone a small trilobite tail (Silurian homalonotid? according to Dr. B. Rowell, British Museum) has been located, and an indistinct thorax impression was found in the Cedarberg Formation near
The arthropod tracks are commonly associated with hemispherical burrows 70-240 mm in diameter which are characteristically filled with coarser-grained material than the host rock. The close association of tracks and burrows supports the deduction that the burrows were dug by the same animals as made the tracks.

The most remarkable exposure of tracks in the Table Mountain Group remains the Brandenburg bedding plane on which no fewer than 12 or 13 different organic or possibly organic structures are preserved on an original area of some 20 square metres. In the following description the tracks will be treated as if each was made by a different individual (and presumably a different species) but this is an oversimplification of the problem because the same individual may leave different footprints due to a change in gait or due to differential preservation of the track, or juvenile and mature individuals of the same species may leave quite dissimilar tracks. A close study of the tracks nevertheless reveals many of the qualities of the animals.

The illustrations of the tracks (Figs. 70-78) are full-size carbon paper rubbings made directly from the original tracks. On some of the tracings the footprints and the stride sequence have been outlined for extra clarity.

Individual A left a track (Fig. 70) about 120 mm wide with distinct and sharply delineated, irregularly elliptical and pointed footprints, 8 mm x 15 mm or smaller in size. A very shallow, discontinuous, central tail drag mark with a median ridge is 12 mm wide and 120 mm long and is repeated regularly at 180 mm intervals. This interval corresponds to the cyclical footprint sequence which consists of 9 oppositely paired footprints. The footprints are thrust outwards at about 36 degrees to the direction of travel (down in Figure 70), and low thrust heaps occur behind the footprints.

Animal A walked with a strong, regular and purposeful gait on its stump-like feet, and, with the exception of its tail, its body was lifted on its feet clear of the surface. It walked with a slight see-saw motion so that its tail not only lifted and lowered regularly but it swayed gently from side to side. The animal must have been reasonably heavy because, notwithstanding the large surface area of its feet, the track is clearly
imprinted in the sand.

Individual B was probably related to A if the footprints are compared but here is an instance where the track of the same individual changes due to a change in gait. The first portion (Fig. 71) of a continuous track shows a double row of footprints 85 mm apart, separated by most prominent discontinuous central tail drag mark. The elliptical footprints are indifferently placed in a 7 or 8 footprint sequence, and are thrust outwards at about 50 degrees to the direction of travel. The tail was not merely dragged over the surface but was actually pressed into the sand; the drag marks are 100 mm long and are repeated every 140 mm. A second tracing of the same continuous track (Fig. 72) repeats the characteristically disordered footprint sequence but the median tail drag mark is not developed. The animal was probably highly articulated because not only could it execute an almost 90 degrees turn in about its body length (a 1:2 width/length ratio is common in trilobites, Moore, Lalicker and Fisher, 1952), but it could manipulate its tail by either pressing it down or holding it above the surface. The regular line of footprints indicate that all the legs of this determined walker were of the same length although the stride length was not constant.

Individual C left a track (Fig. 73) 60 mm wide, with irregularly triangular footprints resembling those of Animal A. The cyclical footprint sequence consists of 10 prints repeated every 150 mm but this is not duplicated by any tail drag mark. This animal is not noticeably different from the types described above.

Individual D left a track (Figs. 73, 74) in which the small, delicate footprints with their prominent thrust heaps record the very long stride of the animal as it executed graceful turns on the sand. The animal obviously had few, but long and spindly legs, with sharply pointed feet, like a crab, and relative to its flat-footed companions, it was fast and agile.

Diminutive Individual E was only about 30 mm wide and it had short stubby feet (Fig. 75) on which its body was carried clear of the surface. It is not possible to establish its footprint sequence which means that its stride length was very regular and that all its feet were alike.

Individual F is represented by a track (Fig. 76; Plate 14F) which
is characterised by two features: a continuous tail drag mark, and an asymmetrical distribution of the footprints flanking the drag mark. The irregular or elliptical footprints are indistinctly placed in a cycle of about 115 mm. More footprints appear on one side of the track, as though the animal walked askew, like a crab, and on the same side 4 or 5 regularly spaced crescentic impressions were probably made for added propulsion by some part of the animal's body other than feet. The tail drag mark is 12-15 mm wide and clearly shows the direction of travel. (The slab on which this track occurs is at the South African Museum, Cape Town.)

Individual G left a track (Plate 190) about 85 mm wide with no tail drag mark. The chisel-like feet were thrust outwards at about 45 degrees so that conspicuous thrust heaps formed behind the footprints. The feet were not dragged but lifted clear with each step and the stride length was at least 35 mm.

This concludes the description of the Brandenburg tracks. However, more tracks, some similar to but others different from the Brandenburg types, occur elsewhere in the Graafwater Formation. At Graafwater one track closely resembles type G; apparently the two animals differed only in size, the Graafwater individual being slightly smaller. A new kind of track was made by Individual H, which was presumably about 95 mm wide, and possessed of many small weak legs, which in their laborious efforts left two swaths about 25 mm wide of numerous indistinct footprints and drag marks. The reluctance if not inability of this sluggish animal to follow a winding course is reflected in its track, which is a bee-line for the full 3 metres exposed of it.

At Tierhoek, Olievenboskraal, the upper Loop Member contains a variety of tracks of types B, C, and G. One new track (Fig. 77) is characterised by a double tail drag mark. The track is 60 mm wide with regular oppositely spaced stubby footprints about 15 mm x 6 mm or smaller in size; the track is not sufficiently clear to determine the footprint sequence. Individual I lifted its feet clear and with an unvarying rhythm and it preferred a straight course. The depth of its footprints indicates that the animal was not a lightweight. The continuous central drag marks are 3 mm wide and 16 mm apart and were formed by a forked telson which was passively dragged on the sand.
The tracks near Stellenbosch are quite different from those described above and this dissimilarity is probably due to a significantly different faunal assemblage in the northern and southern areas of the Graafwater basin.

In Helderberg Mountain and in Botmaskop, Stellenbosch, a few short tracks in the Graafwater Formation are characterised by asymmetrical tracks (Plates 18B; 19B) with typical scooped-out footprints with toe scratches. Two types of foot can be identified: the larger, oar-shaped foot was about 10 mm wide and it had at least 4 pointed toes or projections which probably aided in digging. When in motion the movement of this foot was directed down and sideways towards the rear as if with a scoop-like action. The smaller foot was only 2 or 3 mm across and seldom shows toe scratches. The larger footprints are spaced about 15 mm apart, and the smaller are 10 mm apart; the tracks are 50-60 mm wide and show that the animal, Individual J, walked askew. Tail drag marks do not occur.

At Chapman's Peak Drive, Simonstown, a poorly preserved track (Fig. 78) is 90 mm wide and consists of a normal double row of lenticular footprints with a 9 step cycle repeated at 65-70 mm intervals. This animal, Individual K, probably did not differ much from type G.

The tracks and burrows in the Peninsula Formation at Matjiesgoedkloof, Urionskraal, are remarkable not merely because they are isolated features in the Peninsula Formation, but because these trace fossils occur at the very stratigraphic top of the formation; they are therefore separated from the Graafwater Formation by an appreciable time interval. Three tracks, 35 mm, 60 mm and 150 mm wide, and without tail drag marks, are associated with burrows (Plate 14D); unfortunately the degree of preservation is poor in the coarse-grained orthosandstone but the individuals appear to resemble types B, C and G.

Burrows are commonly associated with arthropod tracks in the Graafwater Formation (Plate 14). The burrows differ from those ascribed by Seilacher (1959, 1960, 1967) to trilobites in that the burrows in the Graafwater Formation are not elongated, scooped-out grooves, but are hemispherical hollows which show no digging marks or any internal structures. The burrows are from 60-240 mm in diameter and are normally truncated by erosion (Plate 14A, C, E). On the Brandenburg bedding plane the filling of the burrows projects above the bedding surface, thereby indicating
that the burrows were dug from a higher level than the surface on which the tracks occur. The burrows are always filled with material coarser than the host rock (Plate 14B, E): shale pellets at Bruinpunt, Doorn Bay, and a gritty sand with pebbles 5-6 mm in diameter at Gansekraal, Worcester. Cross sections through the burrows invariably demonstrate a cross cutting relationship with primary structures.

What made the tracks and the burrows?

The absence of exoskeleton remains in the Track Zone is inexplicable when the size and robustness of the organisms are considered. The preservation of delicate and ephemeral tracks makes it all the more difficult to understand why no trace has been found of even a fragment of what must have been the sturdy and substantial framework which these organisms possessed.

Recourse must now be had to an interpretation of the tracks in order to reconstruct the inherent qualities of the animals which made them. Although the tracks differ in detail, as for example in track width, in footprint pattern, and in the presence or absence of tail drag marks, the essential qualities indicate a faunal assemblage of uniform habit.

The undoubted robustness and strength of the animals are reflected in the wide tracks, and the clear and regular footprints testify to the ability of these organisms to move about with dignity. The thrust heaps behind the footprints indicate a forceful movement of the legs. The general absence of body drag marks proves that most of the animals were sufficiently sturdy to lift their bodies clear of the surface while walking about. The majority of the animals were walkers, but one was obviously a crawler and it appears as if another could run. The symmetry of the tracks indicates that almost all the animals moved straight ahead but one or two walked askew, like crabs.

The width of the tracks as well as the size of the footprints give some indication of the dimensions of the animals. Body width does not necessarily correspond to the track width but for arthropods the track width approximates the "size" of the animal. According to track width the Table Mountain arthropods had a maximum width range of 30-150 mm and the most common width varied between 60 mm and 120 mm.
ratio of Silurian and Ordovician arthropods commonly ranges from 1:1 to 2:1 (Moore et al., 1952) which makes the estimated body length of the Table Mountain arthropods 30-300 mm; most of the animals were between 60 mm and 240 mm long. Body length may perhaps also be reflected by the footprint sequence length or by the spacing of the recurrent tail drag marks, in which case it is notable that these observed distances fall within the estimated body size range. In addition this estimate of body size is in concert with the observed dimensions of the burrows which were presumably excavated by these organisms; this is significant because Seilacher (1959, 1960, 1962, 1967) states that the animal completely occupied the burrow while digging it.

A distinctive body feature is the tail or telson which was usually a flattened spatulate appendage capable of restricted vertical movement but which normally moved passively in rhythm to the gross body movements or was merely dragged along. Some of the organisms either had no telson, or otherwise one so short as not to leave a drag mark; one individual had a pronounced forked telson.

The oppositely paired feet were almost invariably moved simultaneously. The feet were simple stumpy-like cylindrical or elliptical appendages but pointed, and oar-shaped or chisel-like feet are also recognisable. Except for the tracks near Stellenbosch setae or toe-like projections are not apparent. Body marks are so rare that it is safe to conclude that feet only were used for locomotion. There is no indication that the animals could swim or move in any way other than walking. The predominantly straight tracks seem to indicate that the organisms were not highly articulated and were probably capable only of slow but determined movement. The problematic crescent-shaped marks which accompany the Brandenburg arthropod tracks may perhaps represent rest marks of swimming arthropod larvae.

On the nature of the burrows can only be speculated. Taljaard (1962) considered them to be egg depositories but no egg-like objects have yet been found in the burrows. Seilacher (1959, 1960, 1962, 1967) is of the opinion that such structures are probably feeding burrows or rest places. The generally coarse-grained filling of the Graafwater burrows suggests that the finer-grained sediment was probably winnowed out during
the fossorial activity of a feeding burrower. The organisms probably fed on microscopic algae, plankton and other organic debris in freshly de-
posited sediment. Like some modern arthropods the Table Mountain or-
ganisms were probably predators and scavengers. It is not possible to determine to what extent the arthropoda on the one hand and Scolithus and Helminthoida on the other supplied sustainance to one another.

The Table Mountain animals were most probably trilobites, as suggested by Taljaard (1962). After studying the tracks figured in this disserta-
tion Professor Seilacher revised his earlier private opinion that the tracks were made by eurypterids and now considers that the tracks were certainly made by trilobites (Letter, 1967). The now better known age of lower Ordovician (Chapter VII) for the Graafwater Formation places it in the period when trilobites generally flourished elsewhere. Eurypter-
ids are not considered to have made the tracks because they were apparently swimmers and their so-called walking legs seem much to diminutive for active use in dragging around, and much less in manipulating with ease, the large body with its pronounced unsupported telson. Eurypterids are not asso-
ciated with trilobites and brachiopods (Moore et al., 1952; Zittel, 1927).
Xiphosura could conceivably have made some of the tracks observed in the Graafwater Formation but the main reason for discarding this possibility is the minor importance of xiphosura during the Ordovivian.

The presumed presence of trilobites in the Graafwater basin and in the late Peninsula basin has profound bearing on the reconstruction of the depositional environment because the trilobites are undoubted marine animals.

Brachiopods

A systematic description by paleontologists of the British Museum (Drr. Cocks, Rowell and Brunton) of the forms from the type locality of the Brachiopod Zone (Fig. 62) is in an advanced state of progress, and accordingly no additional description of the forms will be attempted here. Some of the forms (still unnamed) are figured by Rust (in the press).

The forms are generally small (the largest shells measure about 25 mm across). The shells consist primarily of Orbiculoidea-type inarticu-
lates, and the articulates are "dominated by Eostropheodonta, and Isorhizie,
with rather uncommon associated Protatrypa, rhynchs and atrypoids" (Letter from Dr. L.R.M. Cocks). One small trilobite tail (Silurian homalonotid? according to Dr. B. Rowell) and a short segment of a crinoid stem are the only other associated forms found to date.

The whole fauna indicate a definite Lower Silurian (probably Lower Llandovery) age for the Brachiopod Zone. This means that the Track Zone in the Graafwater Formation is considerably older, possibly older than lowermost Ordovician. In addition, the brachiopod fauna indicate "shallow shelf marine conditions" (Letter from Dr. C.H. Brunton) during the deposition of the Cedarberg Formation. It is significant that the marine milieu deduced for the Graafwater Formation on the strength of the trilobite tracks, is independantly supported by brachiopod shells at a much later stage in the history of the Table Mountain basin.

Plants

(This information came too late for inclusion in Chapter II.)

Controversial algal fossils are reported from the Cedarberg Formation ("Upper Shale") near De Doorns by Plumstead (1967). She concludes that the ramifying sinuous strands on bedding planes of dark grey carbonaceous shale "are in all probability of vegetable origin and represent branched cylindrical algal thalli" (1967, p. 63). Similar structures were not found elsewhere in the Cedarberg Formation.

She bases the age of the Table Mountain Group on the age of the Bokkeveld Group, and regards the Cedarberg Formation as being of lowest Devonian or more probably of upper Silurian age.
PALEOCURRENT ANALYSIS

Introduction

Cross-bedding is the dominant vectorial structure in the Table Mountain Group. In the present study attention is focussed mostly on the vectorial significance of cross-bedding and no consideration is given to the distribution in space of the various types of cross-bedding which are developed. The orthosandstone beds are characterised by planar, tabular cross-bedding (Plates 2B,D; 20B, C) in sedimentation units which range in thickness from 1-10 feet, whereas festoon cross-bedding is typical of the fine-grained, thin-bedded units (Plate 3B). Lenticular cross-bedding (Plate 20D) is quite subordinate to the tabular type. Rare deformed cross-bedding occurs sporadically throughout the stratigraphic column (Plates 2D; 20E). The flow direction of more than 8000 cross-bedded units was measured.

The contemporaneous folds of the Fold Zone below the Pakhuis Formation are unusual vectorial structures. In order to determine the direction of iceflow which resulted in the intraformational deformation and in the deposition of the tillite of the Pakhuis Formation, almost 1500 measurements of fold flanks, axial plane traces and axial lineations in the Fold Zone were supplemented by measurements of the striae directions on a number of glacial pavements as well as by 11 fabric analyses of Pakhuis tillite.

Other vectorial primary structures, as for example ripple marks, primary flow lineation, soft-sediment down-slope slumping, sole structures, pebble imbrication and primary orientation of fossil debris, are of subsidiary importance in the overall paleocurrent reconstruction of the Table Mountain basin.

Due to the mountainous nature of the study area the rigorous spacing of sampling stations is impractical and it was decided therefore that at least two sampling stations per formation per quarter degree quadrangle should provide adequate sampling density. The paleocurrent maps which were compiled from the field data indicate that satisfactory control exists even though it was not always possible to maintain the required sampling density.

Most workers on cross-bedding data (Potter and Pettijohn, 1963) pre-
fer to reduce all readings to a vector mean flow direction and to com-
pute the standard deviation, but in this study in addition to such cal-
culations, the readings have been plotted on an equal-area projection so
that a complete graphical plot of the flow pattern results. The present
IBM 1620 computer program which was developed by Mr. R. McD. Dodds pro-
cesses 60 readings per 10 minutes during which time the raw information
is read in from punched cards, the effect of tectonic tilt on the vector-
rial data is removed, and the transformations and calculations completed,
and a graphical plot 10 inches in diameter is printed out, together with
the vector mean, vector strength and mean dip angle in the case of cross-
bedding diagrams.

The graphic equal-area diagrams are contoured lower hemisphere pole
plots in the case of the Fold Zone diagrams (such plots are normal proce-
dure for structural analysis of folds). The low dip angle of a cross-
bedding lamina results in its pole plotting near the centre of the diagram
which is unsatisfactory if the aim of the diagram is to depict the direction
of flow. However, a plot of the line of maximum dip, or nadir point, of a
cross-bedding lamina yields a point near the circumference of the diagram,
thereby endowing the plot with appreciably improved moment. The area of
maximum concentration of nadir points on a diagram indicates the extent and
direction of dominant stream flow for that particular collection of cross-
bedding measurements.

Early in the field study as many as 250 cross-bedding readings per
sampling were collected, but progressive development of the graphic plots
during collection of the data proved that 30 readings constitute a lower
limit, and 50-70 readings are an optimum number of readings from which
to fully identify the nature of the stream pattern at any particular
sampling point (Fig. 80). Cross-bedding is so prolifically developed
in the Table Mountain Group that 50-70 different measurable laminae are
easily found in a reasonably small volume of rock; only in rare in-
stances is it necessary to traverse a large volume in order to collect
sufficient readings. The gross lithologic homogeneity of the constituent
formations of the Table Mountain Group makes it unlikely that important
stream flow fluctuations could have taken place during the deposition
of any one formation, and as expected the cross-bedding plots, developed
progressively at different stratigraphic levels in the same formation, show no systematic change in the flow pattern (Fig. 80). In practice this means that as long as at least 30 measurements, or preferably 50-70 measurements, are made, the stratigraphical distribution of the cross-bedding sample is immaterial.

In order to determine the mean geometry of the Fold Zone at any one sampling station practical experience dictates that at least 50 measurements should be made on fold flanks, and that this information should be supplemented by as much data as possible on fold axial traces and axial lineation. In a few localities as many as 100 measurements were insufficient to clarify the structure whereas in others as few as 20 values gave dependable results.

Streamflow

Raw cross-bedding data were treated on three levels. On the first level contoured flow diagrams were computed for some 120 sample stations. The diagrams are based on from 30 to 200 readings each, and the streamflow patterns were investigated in an unsuccessful attempt to extract meaningful parameters relating to the hydraulic milieu of the paleocurrents. The lack of sufficiently sophisticated methods of three-dimensional statistical analysis (Fisher, 1953) hampered the proper exploitation of present data. The second level treatment entailed plotting the vector mean flow direction on base maps in order to determine the paleocurrent pattern for each formation. The paleocurrent maps were then subdivided into areas characterised by harmonious vectors, and the third level treatment consisted of the computation of composite flow diagrams for the sub-areas. The composite diagrams picture the paleocurrent flow pattern in areas where the regional stream flow direction was reasonably constant, and the large sample size (the 14 composite diagrams are on the average based on 565 readings each) overrides the effects of local fluctuations in stream flow direction.

(IMPORTANT: The composite diagrams, Figures 81-83, are reduced facsimiles of originally 10 inch diameter computer plots on which the circular outline and contour lines have been drawn. Magnetic north is indicated by 000 at the top of the diagrams; geographic north is about 23 degrees east of magnetic north. In the other diagrams and maps all the data
have been referred to geographic north.)

The flow diagrams of the Piekenier Formation (Fig. 84) exhibit strongly peaked unimodal distributions with an average vector strength of 76%. The composite diagram (Fig. 81a) is based on the cross-bedding data in the northwestern half of the paleocurrent map (Fig. 85). (The composite diagram of the remaining area is not shown because the diagrams are essentially similar in appearance.)

The currents which deposited the Piekenier gravel and sand flowed southeast in a gentle fan-shaped pattern. Locally as well as regionally the currents were consistently uni-directional as shown by the unimodal cross-bedding distribution and the high vector strength. The cross-beded units are thick (Plate 2D) and the sediment is coarse (Plate 1).

The combined evidence points to vigorous currents. In one or two localities, however, notably at De Hoek, Piketberg (Fig. 84G), inequivalent flow diagrams (vector strength 33%) indicate that relatively quiet-water conditions prevailed in places in contrast to the characteristic high energy deposition elsewhere in the Piekenier basin. At De Hoek the De Hoek Member is a thin-bedded sandstone and the Rest Member is not developed, as would be expected for quiet-water sedimentation.

The cross-bedding flow diagrams and paleocurrent map of the Graafwater Formation (Figs. 87, 81b, 88) differ from those of the Piekenier Formation in several important respects. The average vector strength of the Graafwater flow diagrams is 56% but sporadic high values do occur. Low vector strength values can be interpreted to mean that the currents which deposited the cross-bedding were mostly inconsistent in direction. A conspicuous reversal of the regional flow direction is documented at a number of sampling stations, especially to the northwest (see, for example, the flow diagrams of Doorn Bay and Graafwater, Fig. 87). This pattern of diametrically opposed flow direction is particularly well demonstrated by the composite diagram of the northwestern outcrops (Fig. 81b), and by means of a special method of calculating the moving averages, this feature has been incorporated in the paleocurrent map (Fig. 88).

The flow lines of the paleocurrent map more or less parallel the earlier fan-shaped pattern of the Piekenier Formation, but with the important added quality of reversed flow. It appears as if the Stettyn Rise affected the flow pattern in the south (Cape Peninsula, Fig. 87H) because
the trimodal flow pattern may be interpreted as indicating longshore and offshore currents.

Cross-bedding diagrams of the Peninsula Formation (Figs. 89, 82) have an average vector strength of 65%, which is an increase over that of the Graafwater Formation of 13.9%, but a reduction of 14.4% when compared to the mean value for the Piekenier Formation. With regard to vector strength the Peninsula currents were therefore "intermediate" between those of the Piekenier and Graafwater Formations. The contoured flow diagrams are conspicuously polymodal except near the paleoshore (Fig. 89) where unimodal distributions with high vector strengths are developed. The polymodal distributions are interpreted as resulting from the constantly shifting currents which must have endlessly re-worked the supermature sand which filled the basin during Peninsula time. The re-working currents were, however, controlled by paleoslope because the regional flow trend (Fig. 90) is remarkably constant.

The paleocurrent map of the Peninsula Formation illustrates a major change in the direction of stream flow in the basin. Where the flow was up to now largely axial with respect to the zone of maximum negative movement (Fig. 103), an obliquely transverse flow pattern developed during the Peninsual deposition.

The paleocurrent map, which is actually based not on the vector means but on the principal modes (direction of strongest flow at any sample station), reveals a peculiar southeasterly flow trend between Piketberg and Porterville (Fig. 90). Inspection of the composite diagrams indicates that the Porterville flow pattern (Fig. 82b) differs from the general pattern (Fig. 82a) not so much in the vector mean as in the distribution, which is markedly skewed to the southeast in the case of the Porterville flow diagram. The reason for this flow deflection is obscure. The regional flow pattern appears to be independent of local modes; that this is so is demonstrated not only by the individual plots (Fig. 89) but also by the composite flow diagrams: a line of symmetry bisecting the distribution of nadir points indicates the regional flow trend closely. This feature can be exploited when only a few cross-bedding readings are available or when the distribution is polymodal, because the bisectrix separates the regional trend from the locally dominant vector(s).

In view of the similarity in gross field appearance between the
Peninsula and Nardoëw Formations, it is not unexpected that the paleoflow pattern (Figs. 91, 92) and mean vector strength (60%) of the Nardoëw Formation closely parallel those of the Peninsula Formation. The flow diagrams (Fig. 93) of the Nardoëw Formation are markedly trimodal near the paleoshore (Melkkras, Fig. 93A; Nardoëw, Fig. 93C, and others not shown, for example Wolwepunt, Nieuwoudtville, and Landekloof, Doringbos) and early in this investigation it was thought that symmetrical trimodal flow distribution characterises the entire Nardoëw Formation, but away from the paleoshore the trimodal aspect changes to polymodal (Fig. 93D,G) while the composite diagrams are markedly unimodal and symmetrical, even for the areas nearest to the paleoshore (Fig. 83b). The trimodal and polymodal flow distributions are due to local small-scale deflections of the regional current pattern, but no satisfactory method has yet been devised to study the flow deflections as residuals of a regional trend surface.

No significant difference appears when comparing regional flow patterns based on the cross-bedding vector mean flow direction (Fig. 91) and the direction of maximum stream flow or principal mode (Fig. 92).

Trend surfaces (Krumbein, 1959) were calculated for the data presented on Figure 91. The linear surface accounted for 62.0% of the variation, the quadratic surface for 64.7%, and the cubic surface for 71.0%. The cubic trend map (Fig. 94) is a close reflection of the cross-bedding flow pattern, and shows paleostreams from the east entering the basin in the north, then turning abruptly south and continuing southwards over the major part of the basin before gradually swinging towards the southeast near Stellenbosch and Worcester. The wide, asymmetrical arc-like flow pattern seems to be strongly developed in the Nardoëw Formation.

The paleocurrent pattern of the Table Mountain Group has been compiled exclusively from cross-bedding data because ripple marks have only limited vectorial application and are in any case practically confined to the Graafwater Formation. Ripple marks in the Table Mountain Group are invariably short wave-length structures and both stream and oscillation types are developed (Plates 4A,E; 8B; 13G; 15C). In the Graafwater Formation ripple marks are commonly associated with mudcracks (Plate 4D), which indicates that the ripple marks are shallow water structures. The stream pattern deduced from ripple marks is usually similar to that derived from...
associated cross-bedding but in places it is clear that two different sets of conditions were responsible for ripple marks and cross-bedding respectively (Fig. 87). Stream ripple marks in the Oskop Member (Plates 13G; 15C) show no systematic relationship to the paleoslope as deduced from a vectorial study of the Pakhuis Formation (Fig. 95). The general impression gained from a comparison of cross-bedding and ripple mark data is that ripple marks are not reliable paleocurrent indicators in the Table Mountain Group.

Primary current lineation (parting lineation) occurs sporadically in thin-beded portions of the Table Mountain Group. At Lelikkloof, Doringbos, the thin-beded uppermost Rietvlei Member (Plate 21A) develops primary flow lineation parallel to the cross-bedding flow direction (Fig. 80a); this is a common relationship in sedimentary rocks (Potter and Mast, 1963).

Parallel orientation of the long axes of short fusilinid-like cylinders (similar to the structures on Plate 12G) to the cross-bedding stream flow direction (King, 1948, in Potter and Pettijohn, 1963) occurs in the Skurweberg Member at Die Drif, Grootrivier (Fig. 93). The degree of preservation is insufficient to determine whether the orientation is due to a life position or to after-death clastic orientation.

No in situ turbidity-type sole structures were found in the Table Mountain Group. A block displaying spatulate structures was picked up on the lower slopes of Stellenbosch Mountain and was interpreted as a steep-slope sandflow (Swart, 1950); presumably this specimen is derived from the Graafwater Formation. A problematic grooved (sole?) structure in the Goudini Member at Piekenskloof Pass parallels the regional paleocurrent direction (Fig. 91) but locally the attitude of the deformation is such that the groove may well be a tectonic feature.

Iceflow

Although the Pakhuis Formation occupies only a very small volume of the Table Mountain rocks, the tillite members have considerable paleogeographical implication, and it is important to determine whether the paleoflow direction of the glaciers conforms to the general dispersal
pattern of the Table Mountain basin as deduced from cross-bedding and facies data. Iceflow direction is recorded in the Pakhuis tillites by means of striae on glacial floors, and by depositional fabric as well as by the dispersal pattern of heavy minerals and erratic types (Visser, 1962). It was shown in Chapter III that the soft-sediment folds of the Fold Zone also indicate the trend of glacier movement.

Undoubted glacial pavements occur only near the northern basin-edge of the Pakhuis Formation. All these floors were cut by the Kobe glacier sheet. The striae at Bergland, Lokenburg (Plate 10E), indicate flow in a direction 250 degrees, and the grooves at Pakhuis (Plate 15B) show that the ice probably moved in a direction 227 degrees. At Groenberg, Pakhuis, scribing tools are still embedded in glacial grooves (Plate 15C) which record ice movement in a direction of 307 degrees; the stream ripple marks which are associated with this floor make it probable that the pavement was left by a grounded Kobe ice floe. Doubtful glacial floors are cut in the Oskop Member at Oskop, Klawer, and at Molenrivier, Gydo; the wide, flat-bottomed grooves at Oskop strike 272 degrees whereas the delicate striae at Molenrivier (Plate 10C) had a strike of 173 degrees prior to tectonic tilt.

Near Donkerkloof, Citrusdal, indistinct grooves (Plate 10D) strike about 160 degrees. Haughton et al. (1927) and Visser (1962) report pavement-like grooves (probably of Kobe = Steenbras age) at Wellington Sneeukop (strike 135 degrees) and at Kleinmond (strike 128 degrees), respectively. The southern grooves are admittedly vague but there is no serious reason why they should not be considered to be of glacial origin.

In order to determine the relationship of the microscopic tillite fabric orientation with respect to the floor striae and to the Fold Zone symmetry, fabric analysis were conducted on 15 selected oriented specimens of Pakhuis tillite in the area between Bergland, Lokenburg, and Wellington Sneeukop. The following experimental procedure was employed:

The original horizontal plane, or ab-surface, of the rock specimen was determined from field evidence, the tillite itself being normally unbedded, and the specimen was sawn through along this plane, after which the surface was smoothed with 400 grit. The magnetic north direction was then transferred to the ab-plane and a ¼ inch by ¼ inch parallel grid was drawn on the sawn surface. The specimen was mounted on plasticine and inspected under a stereomicroscope after wetting the smoothed
rock surface with thin oil. The orientation frequency of all apparently elongated sand-sized and larger fragments was determined by means of an ocular graticule engraved in 20 degree sectors; an average of 150 fragments per surface is sufficient to determine the sense and degree of anisotropy in a tillite specimen. (The specimens were too large to be mounted on a mechanical stage, and the grid drawn on the rock surface was used to orient the specimen against a reference line on the graticule while moving the sample along by hand on the microscope stage.) After completely traversing the ab-surface the vector mean was calculated and Tukey's test (Harrison, 1957) employed to determine the significant percentage of anisotropy; in only 4 cases out 28 was the significant percentage less than 99.5%. The specimen was then cut in an originally vertical plane parallel to the vector mean of elongated particles on the ab-surface, and the procedure described above was repeated in order to determine whether significant imbrication was present. By means of this method an essentially three-dimensional fabric analysis of a relatively large rock volume is determined quickly and without the necessity of making thin sections.

The fabric analysis data are presented as star diagrams (Fig. 96, 97, 98), two to a specimen. The mode of the long axis distribution (longest diameter on the diagrams) is interpreted to indicate the iceflow trend on the horizontal plane (ab-surface) diagrams, whereas the mode on the vertical cut (ac-surface) diagrams indicates the attitude of imbrication. (In both cases the original data were recalculated in terms of a standard mode diameter so that the degree of anisotropy of the data is reflected in ratios of mode diameter to other class interval diameters.) The microscopic information is supplemented by two contoured equal area diagrams showing the orientation of pebble long axes in the tillite (Visser, 1962).

All the ab-diagrams are markedly anisotropic (13 of the original 15 diagrams were anisotropic above the 99% percent significance level). The regional vector mean trend for all the tillites is almost due south (Fig. 95). The angle of imbrication (ac-diagrams) is more pronounced in the Sneukop tillite (Figs. 96H, 97B moving average, 98B, F) than in either the Kobe tillite (Fig. 96D) or the Steenbras tillite (Figs. 97E, 98D), but almost invariably the deduced direction of iceflow is towards the south. In two instances only is the imbrication opposite to the regional trend; at Oskop, Klawer, the deduced iceflow is from the west, where-
as it is from the south at Bergsig, Citrusdal (the latter result was ignored in the compilation of Figure 95 because the fabric of a duplicate specimen at Bergsig conformed to the regional trend). At Bergland, Lokenburg, the relationship between diagram mode and glacial striae is opposite to the expected (Fig. 96A); however, the sample was taken in the lee of the Bergland Hill (Plate 6A, B) and the observed fabric probably resulted from a transverse spill-over effect in the ice in the vicinity of the hill (unfortunately the hand specimen disintegrated in an attempt to determine the imbrication). At nearby Matjiesgoedkloof, away from topographic interference, the tillite fabric is normal (Fig. 96C, D).

As a test of symmetry one specimen of the Sneeukop tillite at Kunje, Citrusdal, was sectioned along the ac-plane (transverse to the ab-mode) and although it was necessary to smooth the data the expected symmetrical fabric emerged (Fig. 97C). This is not a test of tillite fabric (Harrison, 1957; Potter and Mast, 1963) but merely confirms that the other modes (Fig. 97A, B) are real.

The pebble long axis distributions (Figs. 96B, 98G, modified from Visser, 1962) are in complete agreement with microscopic fabric analyses.

The combined fabric analysis evidence favours the interpretation that the Pakhuis ice sheets moved generally from north to south (Fig. 95) and that there was an important influx from the east near the northern basin-edge.

The soft-sediment folds of the Fold Zone were first interpreted by Haughton, Krige and Krige (1927) as transverse structures which formed below an ice sheet while it moved from west to east over unconsolidated sediment. Martin (1961) found the same relationship and inferred the same mechanism for similar structures in South America. The contemporaneous character of the Table Mountain Fold Zone was never in dispute (Du Toit and Haughton, 1966). Visser (1962) however, decided that the intraformational folds formed not transverse to the glacier movement, but parallel to the iceflow, and he thereby postulated a radical change in the previously accepted (Du Toit, 1956) direction of iceflow in the Table Mountain basin. The present study supports Visser's deduction. The weight of indirect evidence points in this direction: firstly, the orientation of the glacial striae; secondly, the vectorial significance.
of the macro- as well as microfabric of the tillites; thirdly, the pro-
gressive southward decrease in the proportion of mechanically unstable
heavy minerals and erratic types in the Pakhuis tillites (Visser, 1962);
and fourthly, the unchanged regional flow pattern both before and after
the deposition of the tillites (Figs. 81-92). All these features more
or less parallel the axial plane trend of the Fold Zone (Fig. 99). Al-
though recognition of the axial plane of the Fold Zone deformation iden-
tifies the trend of ice movement, the direction sense cannot be determined
from a fold diagram and in this respect macro- or microfabric analysis of
the associated Sneeukop tillite promises to be of the most significance.
CHAPTER VI

THE SHAPE OF THE TABLE MOUNTAIN BASIN

Data collection

In the complete absence of geophysical or other subsurface information (except for one borehole) on the thickness of the various formations of the Table Mountain Group, the isopach maps are entirely based on surface information. In a number of cases thickness of the strata was measured directly by means of precision altimeter or by means of a Jacob's staff. Detailed plane table maps of various small areas (Drievlei, Lokenburg; the area south of Lamberts Bay; the coastal strip between Doorn Bay and Strandfontein; the trilobite track locality at Brandenburg, Redelinghuys; the Piekenierskloof Pass and Sir Lowry's Pass areas and the up-faulted fragment of the Graafwater Formation at the bridge over the Doorn River south of Klawer) were used to calculate some thickness values. Field controlled photogeological maps, notably of the Piketberg Range and of the Matsikamma Range, Vanrhynsdorp, were also used to determine the thickness of strata.

Most of the values on which the isopach maps are based, however, were obtained by means of geometric construction methods making use of the invaluable 1:50,000 and 1:18,000 topographic maps compiled by the Trigonometric Survey. Apart from the information collected in the field, most of the thickness values in the southern half of the area studied are based on published and unpublished maps and reports of the Geological Survey. The Cedarberg Formation proved to be an important stratigraphical and structural marker and its outcrop pattern was used to calculate strike and dip values in a few areas. However, the Cedarberg Formation itself accentuated the main shortcoming of extracting thickness data by means of map analysis, namely the inability to detect small differences in thickness.

Compilation of the isopach maps

For each formation the thickness values were plotted on a 1:250,000 base map, and contoured when sufficient information was assembled. Usually this entailed the calculation of additional values in order to
clear up ambiguous situations. It must be stressed that three factors combine to make the contoured maps somewhat biased. Firstly, the long narrow outcrop belt limits the area in which surface data can be obtained with the result that the trend of the contour lines along the short side, that is, from east to west, cannot be accurately determined; secondly, the density of the control stations is rather low due to the limitations of the surface methods used to determine the thickness values; and thirdly, the inherent error of especially the geometric measuring methods cannot be evaluated. Partly because of these limitations the isopach map of the Winterhoek Subgroup (excluding the Sneeukop Member and the Fold Zone) could be compiled only after resorting to the calculation of moving averages; the resultant map is not considered particularly accurate. The isopach map of the Table Mountain Group as a whole was compiled from the data on the constituent formations and in order to make maximum use of this information it was necessary to extrapolate the individual isopach surface in places.

The isopach maps have not been constructed on a palinspastic base map. Over a large area the Table Mountain Group is practically horizontal and the deformation, where present, is usually moderate. Removal of the effects of tectonic deformation from the isopach maps will result in a slight widening of the east-west aspect of the basin with proportional lowering of the floor slopes. For present purposes no essential features are obscured by the tectonic effects.

Mathematical models

For each isopach map linear, quadratic and cubic trend surfaces were computed (Table 3), using in all the cases except for the composite Table Mountain Group surface, the irregularly-spaced raw data on which the isopach maps were based originally (Krumbein, 1959). The computer program was written by Mr. R. McD. Dodds and the calculations were done at the Computing Centre of the Stellenbosch University.
Table 3. Percentage fit of trend surfaces to the thickness data.

<table>
<thead>
<tr>
<th>UNIT</th>
<th>LINEAR</th>
<th>QUADRATIC + L</th>
<th>CUBIC + Q. + L</th>
</tr>
</thead>
<tbody>
<tr>
<td>Table Mountain Group</td>
<td>6.6%</td>
<td>58.3%</td>
<td>67.5%</td>
</tr>
<tr>
<td>Nardouw Formation</td>
<td>27.3%</td>
<td>70.5%</td>
<td>83.9%</td>
</tr>
<tr>
<td>Cedarberg Formation</td>
<td>13.4%</td>
<td>81.4%</td>
<td>93.0%</td>
</tr>
<tr>
<td>Peninsula Formation</td>
<td>25.5%</td>
<td>76.6%</td>
<td>86.4%</td>
</tr>
<tr>
<td>Graafwater Formation</td>
<td>28.2%</td>
<td>50.6%</td>
<td>86.8%</td>
</tr>
<tr>
<td>Pickenier Formation</td>
<td>10.5%</td>
<td>55.2%</td>
<td>83.3%</td>
</tr>
</tbody>
</table>

Considering the reliability of the original data the trend surfaces should perhaps be regarded merely as mathematically smoothed isopach maps and a considerable proportion of the observed deviations can be ascribed to initial error. That some of the deviations are nevertheless significant is shown by their relationship to the basin as a whole.

The linear surfaces are not considered meaningful in view of the bilateral symmetry of the basin. The generally high percentage fit of a third order surface to the data is interpreted to mean that a cubic surfaces approximates the ideal basin shape. The 93.0% fit of the Cedarberg cubic surface may be partly the result of the relatively few (19) control stations used to calculate the surface. The other surfaces are based on between 32 and 50 control stations each.

The trend surfaces are very sensitive to the effect of values lying off the general north-south line on which most of the control stations are clustered. Because of this the trend surfaces are not particularly useful for the prediction of basin characteristics east and west of the outcrop area. However, the trend across the basin axis is well defined.

The final shape of the Table Mountain basin

The isopach map of the Table Mountain Group (Fig. 103) demonstrates the uncomplicated shape of the basin: a prominent trough striking south-east and flanked by marginal shelves. The original northeastern basin edge is preserved but the opposite boundary, though probably not very far removed, cannot be indicated. The maximum negative movement recorded in the basin is about 12,000 feet near Citrusdal. Present indications are
that the deepest part of the trough deviates slightly to the south towards Worcester. It is significant that all the constituent formations mirror this trough in the basin. This means that the dominant negative movement of the Table Mountain basin was singularly confined to the axis of the trough throughout the depositional history of the basin. The significance of the saddle-shaped basin floor is not clear; it seems probable from thickness values of the Table Mountain Group farther to the east that the trough persists in that direction but present indications are that the western hollow shown on the isopach map terminates to the west.

The two basin flanks are markedly dissimilar. The northeastern flank is particularly featureless and consists of a relatively steep slope rising rapidly from the trough which then merges into a low angle slope which persists right up to the basin edge. The southwestern flank has a much more gradual slope. The irregularity southwest of Worcester reflects local control on the rate of negative movement in the basin; this is the site of the Stettyn Rise. The fact that Stettyn Rise can be detected when considering the basin as a whole means that it was an active feature throughout the history of the basin.

The Stettyn Rise is also shown on the cubic trend surface deviation map (Fig. 105) in addition to impressive evidence of the "over-deepening" of the basin in the vicinity of the trough axis. The quadratic and cubic trend surfaces (the latter shown in Figure 104) fit the original data only reasonably well with the result that important anomalies or deviations are included in the deviation or residual maps. The dominant character of the basin trough is highlighted in the deviation map (Fig. 105).

The southeast disposition of the deepest portion of the basin agrees well with the regional grain of the paleogeologic map (Fig. 4). The coincidence of the Table Mountain basin trough with an appreciable subcrop of the Klipheuwel Group seems to be significant.

The sequential development of the basin

The information on which the isopach map of the Piekenier Formation (Fig. 106) is based, is rather meagre due to unfavourable exposures, especially along the West Coast. The first notable feature is the
presence of the trough and the second is the appreciable thickness (+3000 feet) which the formation attains. The zero isopach lines are shown as smooth contours but the actual depositional edge is markedly irregular. The western continuation of the Piekenier basin is unknown but interpretation of the paleocurrent and facies maps (Figs. 85, 86) would indicate that the western termination of the Piekenier basin was not far off the present West Coast. The eastern limit of the basin falls outside the area studied but apparently the formation is not developed very far to the east. This means that the Piekenier isopach map probably represents the major extent and shape of the initial Table Mountain basin. Of course, the Piekenier isopach map shows only that portion where deposition took place; it tells little of the areas where denudation operated actively during Piekenier time, except that those areas are outside the zero isopach lines.

The cubic trend surface (Fig. 107) accommodates 83.3% of the observed variation; although the longitudinal (east-west) disposition of the trend surface is unstable, the profile across the trough is strongly developed. The cubic deviation map (Fig. 108) stresses the negative anomaly between Lamberts Bay and Citrusdal where the trough is "over-deepened" because the actual thickness of the Piekenier Formation is more than indicated by the third order trend surface. The anomaly southwest of Ceres is probably a peripheral effect of no geological significance.

The persistent trough of the Table Mountain basin is mirrored by the isopach map of the Graafwater Formation (Fig. 109) and the regional appearance of the basin documents the transgressive character of the Table Mountain sedimentation pattern. To the north the zero isopach line more or less parallels the Piekenier basin edge, but the Graafwater shore shifted about 15 miles northeast, while to the south the basin was extensively enlarged. Over the Stettyn Rise the Graafwater Formation is either very thin or absent. To the south of the Stettyn Rise the Graafwater basin apparently deepened but lack of outcrops preclude tracing the development of the basin in this direction. Although the Graafwater trough is positioned over the previous Piekenier trough, the profile of the trough is slightly irregular and in addition has a pronounced saddle-shape, in contrast to the earlier uncomplicated shape of the basin. This is due to a change in the tectonic behaviour of the basin during
Graafwater time, which is not only indicated by a change in the basin shape, but also in part by the change in sediment type. The local development of the shaly Middelpos Member over the deepest part of the basin indicates that the rate of negative movement increased sharply at the onset of the Graafwater sedimentation, so that deep water conditions developed, but thereafter the rate moderated and sedimentological considerations indicate that the mean Graafwater rate of negative movement must have been only slightly more than the average value for the entire Table Mountain basin. The rate was probably more and not less because a relatively faster rate of burial is required in order to prevent the sediment reaching the supermaturity associated with, say, the Peninsula or Nardouw basins.

Due to the very poor control on the thickness of the Graafwater Formation in the western area of the isopach map the significance of the western hollow in the trough area is in doubt.

The cubic trend surface (Fig. 110) and the deviations from it (Fig. 111) demonstrate the prominence of the Stettyn Rise and the main trough as well as the irregularity of the deepest portion of the basin and the tendency of the southern part of the Graafwater basin to be "over-deepened".

The next stage in the development of the Table Mountain basin is shown by the isopach map of the Peninsula Formation (Fig. 112) which shows portion of a rather simple trough with its axis along the by now firmly established zone of maximum negative movement in the basin. (For practical reasons the Sneeukop Member is included in the Peninsula Formation.) Both trough axis and basin edge trend northwest but of the latter very little is exposed; locally the coast was extremely irregular but the continued northward transgression of the basin edge is shown by the appreciable northward shift of the zero isopach line. The first evidence of the development of marginal shelf is contained in the Peninsula isopach map. Extrapolation of the isopach contours to the southwest indicates that the basin edge was probably nearby but it is not known whether the Peninsula basin, like the Graafwater basin, developed a slope reversal to the south so that the nearness of the basin edge is apparent rather than real. The effect of the Stettyn Rise on the shape of the basin is marked.

The cubic trend surface (Fig. 113) predicts a southward increase in the thickness of the Peninsula Formation but the residual map (Fig. 114)
shows that in this area the formation tends to be thinner than the cubic model. The dominant feature of the residual map is the negative anomaly which in part reflects the prominence of the main trough.

Although the isopach map of the combined Cedarberg Formation plus the Pakhuis Formation (exclusive of the Sneeukop Member) is poor for reasons stated earlier, it nevertheless illustrates the invariant quality of the tectonic framework of the Table Mountain basin. The accurately placed zero isopach line indicates that the basin continued to expand during the glacial interlude. The basin edge shows more or less where the glaciers entered the basin; to the northeast of approximately the zero isopach line the ice sheet rested on pre-Table Mountain rocks. No regressive terminal moraines were preserved outside the basin perimeter; presumably such deposits were destroyed during the Nardouw phase of continued denudation, transgression and deposition. Extrapolation of the isopach data to the southwest strengthens the suspicion that the Table Mountain basin edge was nearby. However, the glacial paleocurrent map (Fig. 95) practically excludes the possibility that the southwestern positive area (if it existed) was glaciated and was contributing debris to the basin.

Trend surfaces were calculated for thickness data on the Cedarberg Formation, for which 19 reliable thickness values were used. The cubic trend surface (Fig. 116), though unstable in an east-west direction, closely repeats the main trough of the Table Mountain basin and predicts a basin edge to the nearby south of the area studied. In the absence of a Cedarberg isopach map a trend deviation map was not prepared.

The shape of the Nardouw basin (Fig. 117) is similar to that of the Peninsula Formation except that the trough is not as deep and the basin edge is displaced about 35 miles to the northeast. The offshore shelf is most prominently developed in the Nardouw basin. Extrapolation of the data to the southwest adds to the weight of rather vague evidence pointing to the existence of an elusive highland somewhere in the nearby southwest. The continued effect of the Stettyn Rise is clearly noticeable even at this late stage in the history of the Table Mountain basin.

The cubic trend surface of the Nardouw isopach data (Fig. 118) accounts for 83.9% of the observed variation. In general appearance it differs
from the previous patterns. This may be due to a lack of data towards the west or it may indicate a real difference in the basin framework. The choice between these alternatives is difficult if not impossible to make but it may be significant that the paleocurrent trend map (Fig. 94), which is based on completely different kind of information, corresponds well with the thickness trend map. In addition the isopach trend map also predicts highland to the west. Apart from the anomaly caused by the marked disparity between the isopach map and the western portion of the trend surface, the deviation map (Fig. 119) accentuates the "over-deepening" of the Nardouw trough as well as the irregularities of the Stettyn Rise.

In summary the major features of the development of the Table Mountain basin are the following:

(a) The basin is characterised by a zone of maximum downwarp trending southeast. The axis of this trough did not shift laterally during the active life of the basin. The amount of maximum downwarp in the Table Mountain basin is of the order of 12,000 feet and in terms of calculated third order trend surfaces all the stages in the development of the basin are over-deepened along the axis of the trough. This indicates that the Table Mountain trough marks a zone of continued and accentuated tectonic activity in the basin.

(b) The trough of the Table Mountain basin was flanked by two shelves which were especially prominent during Peninsula and Nardouw time. Of the two shelves the northeastern marginal shelf played an important role in the production of the supermature sediment of the Table Mountain basin. Because of tectonic control the marginal shelf was an area with an extremely slow rate of subsidence, and it was therefore mainly an area of transit in terms of the sediment delivered at the edge of the basin. Instead of being an important area of sedimentation, the marginal shelf was an area of active re-working, sorting and attrition. However, the hydraulic conditions even in the central basin were such that the finer grained particles which were washed out of the shelf area were transported through the trough zone also. The only sediment which eventually accumulated in the central basin was the supermature, coarse grained "spill-over" from the marginal shelf. Indirectly the tectonic control of the basin in-
fluenced the quality of the sediment which deposited on the marginal shelf as well as in the adjacent trough area. The presently exposed portion of the southwestern shelf seems to have been an intra-basinal shelf on which further "spill-over" from the trough area was deposited; this shelf therefore also obtained its sediment from the northeastern source and it was not a marginal shelf in the sense of adjoining the edge of the basin.

(c) The regular bilateral symmetry of the basin is disturbed by the Stettyn Rise only, an area which was initially a peninsula or a cluster of islands, and which persisted throughout the history of the basin as a region with a slower rate of subsidence.

(d) The transgressive nature of the basin (Fig. 2) is further illustrated by the isopach maps, and it is clear that the regular growth of the basin can be correlated with an equally regular rate of negative tectonism in the depositional area. The Piekenier isopach map outlines the initial size and shape of the Table Mountain basin, while the final disposition of the basin is shown by the Nardouw isopach map. Apart from the conspicuous progressive enlargement of the basin its essential nature at its close is practically identical to its initial character.
CHAPTER VII

RECONSTRUCTION OF THE CONDITIONS OF SEDIMENTATION AND THE PALEOGEOGRAPHY OF THE TABLE MOUNTAIN BASIN

The tectonic framework

An uncomplicated stable shelf model (Krumbein and Sloss, 1965) best fits the Table Mountain data. The basin shape is a simple trough with flanking off-shore shelves and the position of the axis of maximum negative movement never shifted significantly. This implies local and uniform tectonic activity.

The bulk of the sediments which filled the basin is extremely homogeneous. This may be due to a lithologically homogeneous provenance, and the rock samples of the source area which are preserved as pebbles in the various rudites indicate that quartzite, chert and vein quartz figured prominently in the denuded areas, but other more variable types, for example shale, lava, hornstone, limestone and the like were also present. Obviously the ratios between pebble rock types do not reflect the distribution of those rocks in the provenance area but is seems clear that a supracrustal rather than a plutonic provenance can be deduced. In the case of the Table Mountain sediments processes of transport and deposition eliminated all but the most resistant rocks and minerals, thereby contributing to the biased view now available of the provenance area. The elimination of practically all the fine-grained detritus must be attributed to conditions in the basin itself, namely persistent turbulence to keep in suspension material below a certain critical diameter, and sufficiently persistent currents to sweep away these suspended particles, at any rate, beyond the presently exposed area. On the whole this points to reasonably energetic conditions in a basin which was undergoing sustained and steady negative movement. The absence of limestone and evaporite indicates continued open circulation in the Table Mountain basin.

The sequential mathematical models of the various stages of the basin unfailingly reveal the overdeepening of the depositional area in the vicinity of the invariant trough axis of the basin. It almost appears as though tectonic forces acted mainly on the deepest portion of the trough,
dragging the crust down along the axis of the basin so that the adjacent parts lagged slightly behind. With the increased degree of downwarp the lateral negative effects compounded, thereby steadily enlarging the size of the basin.

The Stetyn Rise should not be seen as a positive element in the basin but rather as a slightly less negative unit.

The tectonic behaviour of the cratonic borderlands is indistinct. Positive as well as negative movements are envisioned; perhaps the picture is particularly complicated because we cannot see much and we do not know that to look for. Initially a mountainland to the west shed coarse detritus into the early Table Mountain basin, this elevated area will be referred to as the ATLANTIC MOUNTAINLAND. However, its role as a producer of debris was soon played out, whether naturally due to fluvial old age, or hastened due to negative tectonism, is speculation. The time available for complete denudation of a mountainland capable of delivering such coarse debris as it did seems insufficient and an Atlantis-like disappearance of the Atlantic Mountainland following the deposition of the Graafwater Formation may perhaps not be out of place. The mountainland flanking the northern basin shore apparently became an important feature only when the highland to the west was subdued. It is unlikely that this northern range, the BUSHMAN MOUNTAINLAND, was near the basin edge. Its lofty peaks supported the valley glaciers which at one stage waxed to coalesce as a piedmont sheet which invaded the basin itself. Positive tectonism may have aided the growth of the ice sheet but undoubtedly climatic control was decisive. Vague but persistent indications direct attention to the recurrent role of the Atlantic Mountainland during the close of the Table Mountain basin, and the history of later formations (Loock, 1967) confirms the presence, at a later stage, of elevated land to the west.

The life span of the Table Mountain basin terminated abruptly during the lower Devonian when an accelerated rate of downwarp changed the shallow water conditions of the Nardouw period into the deep water milieu of the Bokkeveld basin.

The marine environment

For the first time the marine environment of the Table Mountain basin,
hinted at by many, can be proved beyond doubt for the Disa Member by means of fossil evidence and can be similarly deduced with reasonable certainty for the upper Peninsula Formation and for a portion of the Loop Member. It is conspicuous that the lithology of both the Cedarberg and Graafwater Formations differs in several respects from the supermature aspect of the rest of the Table Mountain rocks and it is enticing to suppose that the fossiliferous formations represent the sedimentary record of isolated marine incursions into an otherwise non-marine environment. However, the absence of marine fossils in the bulk of the Table mountain rocks is no proof of the non-marine derivation of the sediments. The lack of recognisably marine fossils may be due to micro-environmental control as for instance bottom-sediment conditions, availability of nutrients, turbulence, water depth and temperature and the like.

Two lines of argument may be employed to further the case for a general marine environment. The unique lithology of the marine Cedarberg Formation is probably due to glacial interference rather than to other changes in the basin. An undoubted marine sequence nearest to the Cedarberg Formation is the lower Bokkeveld Group, which follows on the Nardouw Formation with complete structural concordance but with marked lithological change. The concordance indicates an uninterrupted process of sedimentation and the change in lithology is related to a significant change in the tectonic framework of the basin, which affected markedly the rate as well as the type of sedimentation. There is no need to postulate a marine transgression into what was previously a non-marine area at the commencement of the Bokkeveld sedimentation; the shallow sea of the Nardouw period merely became deeper due to accelerated negative movement in the basin.

If it is conceded that the Nardouw sandstone is a marine sediment then there is little reason to suppose that the lithologically similar Peninsula Formation did not also accumulate under marine conditions. Trilobite tracks in the Peninsula Formation would strengthen the case for a marine environment but such tracks are admittedly very scarce. However, it is significant that the Peninsula tracks found to date occur where sedimentological evidence dictates a beach environment. The extremely shallow water habitat of the Table Mountain trilobites is proved by the conspicuous shallow water primary structures in the Graafwater
Formation. The general absence of tracks in the Peninsula Formation (as well as in the Nardouw Formation) may therefore be attributed simply to excessive water depth rather than to a non-marine environment.

The marine environment of the Loop Member, and, by analogy, of the Graafwater Formation as a whole, must be modified in consideration of the sedimentological evidence. A marine tidal flat milieu accounts satisfactorily for all the observed features. Apparently then, the Loop trilobites were shallow water creatures equally at home in water and on the beach.

The Piekenier Formation is the only sequence which by its complete lack of fossil traces forces an environmental reconstruction based on sedimentological data only. A deltaic-epineritic milieu explains many of the observed features and the absence of fossil remains may be due to unfavourable conditions of preservation in the highly energetic zone of deposition.

On the whole, therefore, the case for a marine environment for the entire Table Mountain Group (with the partial exception of the Piekenier Formation) seems to be well founded. An argument against a marine environment and in favour of fresh water (lacustrine?) accumulation of especially the Peninsula and Nardouw Formations is that the groundwater of these formations is singularly free of dissolved salts, but the porosity of these formations is so large (intergranular porosity as well as joint, bedding and fault plane porosity) that it may safely be concluded that whatever connate sea water was originally present has since been flushed out by meteoric water.

The rate of sedimentation

The supermature aspect of the bulk of the Table Mountain sediments leads to the impression that a very slow rate of sedimentation prevailed in the Table Mountain basin and an involuntary question in the investigator's mind is: How many years of seemingly endless sorting, washing and reworking have gone into the single bed of quartz sandstone before me?

The reliable dating now available of the Brachiopod Zone makes possible an estimate of the rate of sedimentation in the last stage of the Table Mountain basin. If ages of 430 m.y (lowermost Silurian)
for the Brachiopod Zone, and 400 m.y (Devonian/Silurian, Holmes, 1965) for the onset of the Bokkeveld sedimentation are accepted, the interval of 30 m.y is represented by the Nardouw Formation and the upper Cedarberg Formation. The maximum thickness of this sequence is in the order of 3000 feet, which means that over a period of 30 m.y the basin filled at a mean rate of 1 foot of sediment per 10,000 years, or 1 mm per 33 years. The contribution of the Cedarberg Formation to this rate is negligible because on the one hand it accounts for about 5% only of the succession and conversely it appears that glacial outwash deposition can proceed at a very fast rate (as much as 1 foot per 20 years for Pleistocene varved shale, Holmes, 1965). The calculated rate is therefore referred to as the Nardouw rate, and it applies to supermature quartz sand deposition in the axial area of the Nardouw basin, where both the rate of downwarp and the rate of sedimentation were at a maximum. Obviously, towards the edge of the basin the rate was appreciably decreased.

The age of the Table Mountain Group

A combination of sediment thickness and rate of sedimentation leads to an estimate of the length of the interval during which accumulation took place. The knowledge of the Nardouw rate therefore permits an estimate to be made of the interval corresponding to the Table Mountain System and its subdivisions, as well as of the maximum age of the Table Mountain basin. The lower age limit of the Table Mountain Group is controlled by the middle Cambrian Cape Granite because sufficient time must be allowed for this intrusion to cool and be eroded quite extensively, especially in the southern portion of the basin. The upper boundary is fixed by the Nardouw/Bokkeveld contact, which for the purpose of this calculation is taken at the Silurian/Devonian junction.

It is reasonable to assume that the rates of sedimentation of the Nardouw and Peninsula Formations were similar. The rates of the other formations can be estimated with reference to the Nardouw rate, and in view of the reduced maturity it is probable that these rates were all faster to a greater or lesser extent. In estimating these rates the guiding factor was the extent to which the maturity of the formation in question differed from that of the Nardouw Formation. On this basis the rate for the accumulation of the Winterhoek subgroup was esti-
mated at 50% of the Nardouw rate (it is probable that the rate was even less but the small thickness of the subgroup minimises the effect of the choice of rate on the age of the Table Mountain Group). The rate of the Graafwater Formation was estimated at 75% of the Nardouw rate while the Piekenier Formation was perhaps deposited at an even faster rate, say 66% of the Nardouw rate. Obviously, these rates are merely estimates and there is little control over their values.

A conservative calculation of the age of the Table Mountain Group (Table 4) takes into account the adjusted estimates of the sedimentation rates, and applies these rates to a "mean maximum" thickness for each formation. This last value is the general thickness of each formation in the trough region in the basin and it therefore ignores the effects "over-deepening" on the rate of sedimentation.

Table 4. The estimated age of the Table Mountain Group.

<table>
<thead>
<tr>
<th>LITHOLOGIC UNIT</th>
<th>THICKNESS</th>
<th>RATE</th>
<th>INTERVAL</th>
<th>GEOCHRON</th>
<th>SYSTEM</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bokkeveld Group</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>DEVONIAN</td>
</tr>
<tr>
<td>Nardouw plus</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Disa</td>
<td>3000 feet</td>
<td>10,000 y</td>
<td>30 m.y.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brachiopod Zone</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Cedarberg</td>
<td>150 feet</td>
<td>5,000 y</td>
<td>5 m.y.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>plus Pakhuis</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Peninsula</td>
<td>5000 feet</td>
<td>10,000 y</td>
<td>50 m.y.</td>
<td></td>
<td>ORDOVICIAN</td>
</tr>
<tr>
<td>Graafwater</td>
<td>1100 feet</td>
<td>7,500 y</td>
<td>8½ m.y.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Piekenier</td>
<td>2500 feet</td>
<td>6,600 y</td>
<td>16½ m.y.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cape Granite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>GAMBRIAN</td>
</tr>
</tbody>
</table>

* RATE is the interval in years required for the permanent accumulation of 1 foot of sediment.
In this fashion the estimated age interval of the Table Mountain System is 105 m.y. The Table Mountain basin in the Western Cape therefore most probably came into being during the uppermost Cambrian, and a period of some 50 m.y is available for accomplishing the impressive erosion of the Cape Granite as well as for the deposition and partial erosion of the Klipheuwel Group prior to the development of the Table Mountain basin. The duration of the Winterhoek glaciation is in the order of 1 m.y., which is comparable to the Pleistocene Epoch, while the Nardouw and Peninsula Formations correspond almost to the entire timespan of the classical Silurian and Ordovician Periods.

The rate of transgression

The extreme age of the land surface encroached by the Table Mountain sediments is certainly a reflection of the efficiency of the agents of denudation, which in this case must have been impressive surf which ceaselessly pounded a coast already subdued by fluvial denudation, planing off the remaining topography to a reliefless surface which inexorably disappeared beneath successively overlapping layers of sand.

The rate of transgression at the northern edge of the basin is about 2 miles per million years for the Graafwater sedimentation but during the deposition of the supermature Peninsula Formation the rate of basin expansion decreased to only about ½ mile per million years. This very slow rate is obviously controlled by the tectonic activity in the basin and in the adjoining provenance area. The rate accelerated to 7 miles per million years during the Winterhoek interval but afterwards slowed to about 1½ miles during the Nardouw period. The momentary Winterhoek spurt was not so much controlled by tectonism as by the effect of the overload of the terrestrial glacier sheet which depressed the edge of the basin, thereby causing an abnormal encroachment following on the regression of the ice sheet. In the previous section it was assumed that the Peninsula and Nardouw Formations had similar rates of sedimentation but now it appears that the Nardouw rate of transgression was about four times as fast as for the Peninsula Formation. The prevalence of small-pebble rudite in the northern Nardouw Formation shows that the increased rate of basin expansion was due to a relatively increased rate of marginal down warp in the basin.
The rate of transgression of the Piekenier Formation was probably quite fast because a basin width of some 100 miles was reached in 16 million years; this is equal to a rate of 3 miles per million years for each basin edge. The relatively even aspect of the plane of unconformity is a testimonial to the energetic and extensive "site preparation" prior to the onset of Table Mountain sedimentation. The later expansion of the basin periphery no doubt took advantage of the subdued topography which resulted from the cumulative effects of the continued denudation.

The Piekenier Embayment and its delta

A reconstruction of the presumed Piekenier geography (Fig. 120) shows a medium-sized embayment open to the ocean to the southeast, and a bay-head delta of a major river which drained the Atlantic Mountainland to the west. At the coast the river delivered a large volume of very coarse detritus which was redistributed by wave action prior to permanent deposition. To the southwest was the low-lying peninsular Stettyn Rise, presumably a remnant of a drowned fluvial surface. Southwest of the Stettyn peninsula the Piekenier sea was actively eroding the seafloor and this area did not become a region of deposition until late in the Graafwater interval. The debris from the distant (emergent?) ranges of the Bushman Mountainland had little part in the sedimentation taking place in the Piekenier Embayment.

High energy processes were operating at the delta front and on the northern slopes and coast of the Stettyn Rise. Disrupted framework gravels were dumped by river floods and marine storms at the basin periphery. Some of the gravels were buried but most were sorted in the surf into sand beaches and lag gravels with well-rounded cobbles and boulders. The beach deposits developed large scale cross-bedding as endless combers rushed up the beach slope. The restricted shape of the embayment limited the approach of the waves and the cross-bedding formed in this fashion shows a strongly unimodal distribution; some of the sand beds were deposited as regressive beaches. Most of the sand, however, escaped the beaches and passed through the surf zone into the deeper water of the embayment itself where the conditions of deposition were much more subdued than at the edge of the basin. One or two low energy areas in the Piekenier embayment (protected bays? wave re-
fraction shadows?) are today recorded by thin-bedded sandstone and shale.

The facies distribution (Fig. 86) is controlled mainly by fluvial action but clear proof of fluvial deposition for the Piekenier Formation as a whole is lacking. The fan-shaped cross-bedding pattern resulted from an interplay of fluvial channel streaming and onshore wave action. Most of the Piekenier sediments were accumulated on the beach or under epineritic conditions.

The Graafwater Embayment and its tidal flat

The paleogeography of the Graafwater time is envisioned as a tidal flat (Fig. 121), flanked on the west by the now very much subdued Atlantic Mountainland, and on the north by the Bushman Mountainland, which was much more conspicuous at this stage than earlier. In the distant northwest a transgressive sea was slowly moving southeast; the significance of this postulation will be discussed in the section which deals with the Peninsula paleogeography. The shoreline of the Graafwater Embayment transgressed the earlier basin boundary (Fig. 2) and in the process the Stettyn Peninsula was reduced to a few islands, probably not so much by denudation as by drowning.

The previously energetic milieu was now transfigured into a relatively low energy environment in which deposition of comparatively fine-grained and immature detritus took place practically at sea level by the action of in- and out-going tides. At the northern shore many small rivers delivered immature gravel which was only partly resorted by the spent surf. The fluvial evolution of the Piekenier River was probably accelerated by negative tectonic movement in its catchment area because it was now in an old stage of development. It delivered very little sediment at the edge of the basin and its delta, formerly so prominent, was now almost nonexistent. At times portions of the tidal flat dried up and the mud crust cracked, only to be filled by blown or washed-in sand, or the small mud crust pellets were washed away to be deposited nearby as mudpellet conglomerate. In pools of standing water, wind waves formed small oscillation ripple marks. On the sandy stretches uncovered during the ebb, trilobites strutted about and dug their mysterious borrows. In the pools and in the shallow sea vast colonies of Scolithus and Helminthoida flourished from time to time, and in their fossorial life activities they practically
reconstituted the sediment they lived in.

In the central portion of the Graafwater basin the maximum rate of down-warp resulted in a well-defined vertical subdivision of the stratigraphic column, but towards the northern and southern basin periphery the rate of accumulation was very much slower with the result that lateral facies changes are more important than vertical differentiation. The slow rate of marginal sedimentation was not due to non-deposition but to continued re-erosion in these transit areas.

The major control on the Graafwater sedimentation was tectonic. The "red bed problem" of the formation has not been studied but the unique character of the Graafwater Formation may be due to a combination of tectonic control, shallow water/subaerial sedimentation and provenance pigmentation (Klipheuwel Group?).

The Peninsula sea passage

A major change in the paleogeographic relationships of the Table Mountain basin developed during the transition from the Graafwater interval to the Peninsula period. What was previously an embayment evolved into a sea strait as a result of a postulated connection with open sea to the northwest (Fig. 122). This theory is based on the paleoflow pattern of the Peninsula Formation (Fig. 90).

As a result of the development of the sea strait persistent and strong currents can be postulated flowing more or less from north to south in an area which formerly had no such flow pattern. The cross-bedding patterns of the Peninsula Formation seem to consist of two types. One type has a strongly unimodal distribution, and is restricted more or less to the vicinity of the northern basin edge. This pattern was produced by surf and is essentially identical to the Piekenier pattern. The second type has a polymodal distribution with a regionally constant vector mean, and is characteristic of the bulk of the Peninsula Formation. This type of cross-bedding pattern resulted from the action of the offshore current system in the Peninsula sea strait.

Adjoining the northern shore an impressive lowland developed and merged into an extensive beach. The rivers from the now prominent Bushman
Mountainland meandered through the piedmont plains and flood plains before reaching the sea. On land the sediment was brought to a reasonably high degree of maturity by fluvial action augmented by aeolian action. This was probably a stark and sterile landscape. The sand which eventually reached the beaches was there further reduced to the most resistant components only, quartz and zircon. A slow trickle of sand passed through the breaker zone and from there the marine currents swept southwards endless sheets of quartz sand. The offshore current system was a veritable conveyor belt. Occasional marine storms agitated the bottom sediment, enabling the current system to remove the sand in suspension and leave behind in places the small well-rounded pebbles, which are today seen scattered on some of the Peninsula bedding planes.

On land and epineritic sea floor alike the shifting sand formed an inhospitable habitat for organisms and for most of the time the sea floor was a biological desert. At times, however, sporadic colonies of Scolithus populated the sand, burrowing into it and from the safety of this home extended their antennae to fish for whatever microorganism passed by. On the beach a few hardy trilobites kept equally spartan Heminhoida worms company as they ate their way through the meagre sea sand fare.

Towards the end of the Peninsula period a precursor of the pending Pakhuis glaciation entered the basin. This was the conglomeratic Slanghoek Member, which derived its spate of small pebbles from the proglacial outwash which preceded the advancing ice front. The almost insensible transition from essentially normal climatic conditions to glacial circumstances indicates a regional change in climate and the Winterhoek glaciation does not seem to have been controlled by provenance tectonics.

**The Winterhoek glaciation**

The Winterhoek glacial epoch consisted of two transgressive glacial pulses and one intraglacial period of regression followed by the final deglaciation phase. During the first pulse the Sneeukop tillite was deposited, and, following on the regression of the Sneeukop ice sheet, the intraglacial Oskop sandstone was formed. The second and last glacial pulse resulted in the simultaneous deposition of the Kobe and Steenbras tillites. The deglaciation of the area is recorded in the deposition
of the glacial outwash sediments of the Cedarberg Formation.

The three different tillites are not indicative of either three different provenances or of three different glacier lobes. The Winterhoek ice sheet entered the basin at the northeastern coast and crossed the edge of the basin from the east and northeast but once in the basin itself the sheet flowed from north to south down the slight paleoslope of the basin floor. A reconstruction of the Sneeukop paleogeography (Fig. 123) shows the extent of initial glaciation.

In its passage from the glaciated Bushman Mountainland to the Table Mountain basin the Sneeukop ice sheet acquired its load of submature flu­vial and marine boulders, cobbles, pebbles and sand in the valleys and on the piedmont plain and on the lowlands and on the beaches, faceting in the process the previously rounded pebbles. The Sneeukop tillite therefore represents an almost unmodified sample of the original sediment as it was in transit to the basin, and the already low content of fine-grained material and of unstable mineral and rock particles is notable. The amethyst grains in the tillite indicate that the provenance was probably the pegmatitic area to the south of the present Orange River.

In the next section it will be indicated that the Sneeukop tillite is probably an englacial ground moraine which did not suffer any reworking by water currents but the peripheral regions of the sheet did experience aquatic interference. The Sneeukop sheet was either absent or very thin west of a line more or less through Cape Town - Piketberg - Klawer, and a southern limit of the sheet seems to have been at times between Franschhoek and Hangklip. The gradation between the Slanghoek and Sneeukop Members at Hangklip could be the result of deposition from a drifting lobe of the Sneeukop sheet; the melt-out debris was resorted by bottom currents and no sharp contact developed between the proglacial and glacial deposits. The interbedded sandstone beds at Franschhoek Pass probably mark minor stages of glacial regression at the edge of the sheet, during which stages watersorted sediment accumulated locally. Northwest of Clanwilliam it appears as though the Sneeukop detritus was to some extent redistributed by water, and especially towards Klawer the Sneeukop Member resembles a lag gravel distinguished by the very common unmodified faceted erratics.

In the wake of the Sneeukop glacial regression there followed surf-
working of the surficial debris uncovered by the waning ice sheet. The original surface emerging from the receding ice front was probably uneven and hammocky but it was planed off by wave action to a smooth surface. On this plane the Oskop sandstone was deposited in places. The conditions of sedimentation were probably similar to those of the preceding Peninsula environment because the fine-grained detritus was effectively removed and coarse-grained particles only accumulated. The Sneeukop ice sheet probably receded as far as the edge of the basin because the Oskop sandstone can be followed to within a few miles of the original coast.

The Kobe glacier sheet shortly re-occupied the Table Mountain basin. The direction of ice movement was identical to that of the Sneeukop sheet. To the north the glacial pavements of the Kobe sheet are preserved but southwards the ice probably drifted in very shallow water so that the sheet in places touched bottom and ploughed up the unconsolidated Oskop sand, or merely scratched the upper surface of the frequently ripple-marked sandstone. Melt-out debris accumulated in water beneath the ice sheet, the fragments dropping a few feet or less to the bottom through quiet, stagnant water, turbid with suspended clay. The argillaceous, mostly fine-grained detritus was quite sensitive to the aquatic milieu in which it was being deposited, and in places some of the rubble was rearranged by water currents into bedded pebbly mudstone and rafted shale. To the north, where the glacial pavements prove that the ice sheet was grounded, the glacier deposited debris as ground moraine. The varicoloured tillites of the Urionskraal area may represent minor phases of glaciation, or may be due to provenance control. The ground moraine deposits were not reworked by water.

It was stressed in Chapter III that the Kobe and Steenbras Members, though lithologically dissimilar, are contemporaneous deposits. They were actually both deposited from the Kobe ice sheet. When the Sneeukop piedmont sheet assembled at the edge of the Table Mountain basin it swept the lowland area clean of practically all the sediment on its way to the basin itself; this load was deposited essentially unaltered as the Sneeukop tillite. During the Oskop regressive stage the ice sheet outside the basin area was stagnant, but it resumed its southward journey when the Kobe glaciation set in. In the van of the Kobe sheet was
carried the last remnants of the original arenaceous debris which was picked up during the initial advance of the Winterhoek glaciation. This material was carried farthest into the basin by the Kobe sheet and was deposited as the Steenbras Member. The rear of the Kobe sheet had by this time cut through to bedrock (mostly Nama??) where it acquired its distinctive load, which was eventually deposited as the Kobe Member.

A flood of outwash debris followed in the wake of the final regression of the Winterhoek glaciation. For some time small icebergs melted in the returning sea and laminated pebbly mudstone beds were deposited in this interval. For a long time after the removal of the last traces of glaciation in the basin itself, the streams delivered to the basin fine-grained material only and the conditions were such that excessive sorting of the debris did not occur. That the conditions were marine is proved by the brachiopod fossils, but at least one fresh water lake recorded its existence in the Bulshoek varved shale lentil (Flint, 1947); this lake was perhaps a local ice-locked feature, which explains why the Bulshoek Lentil has such a limited distribution.

If the length of the Pakhuis glacial interlude is taken as 750,000 years (it was probably much less) the mean rate of transgression and regression of the glacier front in the Table Mountain basin was more than 1 mile per 100 years; this rate is therefore a minimum value.

A theory for the glacial folds

When the Sneeukop ice sheet entered the Table Mountain basin it was sufficiently thick to rest on the unconsolidated sand bottom of the Peninsula sea floor. The Peninsula sea is regarded as quite shallow (100 feet?) and for practical purposes the Sneeukop sheet placed the unconsolidated Peninsula sand under an overload of some 30 tons per square foot for every 1000 feet of ice (Flint, 1947). This overload deformed the sand into gigantic load cast structures so that the ice itself was pressed into the slumping sand. The "folds" which formed in this fashion are identical in all respects except size to the normal load cast structures commonly found in sediments which have been overloaded in the normal course of sedimentation (Pettijohn and Potter, 1964, their Plate 100a). If the ice sheet was stagnant the load cast theory predicts random basin-shaped "folds", but if the sheet was moving these structures would be drawn out parallel.
to the direction of movement. The original pods would be deformed into canoe-shaped bodies of glacier ice pressed into the unconsolidated (frozen?) sand. Due to the horizontal stress field the sand layers will also tend to yield by bedding slip parallel to the direction of ice flow thereby forming a bedding plane lineation while attenuating the beds by plastic deformation. The englacial debris will also conform to the stress field, and elongated fragments will imbricate and orient parallel to the direction of flow.

The essential feature of this theory is that the unconsolidated Peninsula sand deformed by plastic flow firstly due to the vertical overload of the ice sheet on the sand, and secondly due to the forward motion of the sheet. Such deformation can only take place when the ice is actually resting solidly on the sediment, and not when the sheet is merely skimming the surface.

It is required by this theory that the Sneeukop tillite is an englacial moraine and it is envisioned that, although the tillite was deposited below sea level, it was deposited in the load-casted pods directly from the sand-choked ice, and in the central basin area it was never subjected to any reworking whatsoever by water currents; this, of course, does not apply to the marginal areas of the Sneeukop sheet, as at Franschhoek Pass or at Windhoek, Klawer.

The Nardouw seascape

The gradual retraction of the ice sheets is marked by the progressive upward coarsening of the Cedarberg Formation as the Nardouw Formation is approached. The water depth in the Table Mountain basin gradually decreased and supermature sedimentation set in again. During the Nardouw period the shoreline of the Table Mountain basin shifted to its northernmost position and the landscape more or less repeated the Peninsula lowland with its extensive sand and gravel beaches flanking the sea strait to the south.

....00000....
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FIGURE 2

LAP-OUT MAP

TABLE MOUNTAIN GROUP

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 —— 100 Miles
0 —— 100 Kilometres

ATLANTIC

OCEAN

Stellenbosch University http://scholar.sun.ac.za
FIGURE 3. CROSS SECTION THROUGH THE NORTHEASTERN EDGE OF THE TABLE MOUNTAIN BASIN.
FIGURE 4

PALEOGEOLOGIC MAP

OF THE

TABLE MOUNTAIN GROUP BASAL UNCONFORMITY

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  VO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 100 Kilometres

0 100 Miles
OSKOP

RIETVLEI

GROOT KLIPHUIS

WITWATER

36

37

38

39

40

41

42
**BOSKLOOF**

**SNEEUKOP,**
*Cedar Mts.*

**WYEKLOOF**

**BERGSIG**

**DIE TRAP, Zuurvlei**

**DIE TRAP, Bo-Rozendal**
FIGURE 80

Lelikkloof, Doringbos.

Note the change in the size of the counting-out circle shown at single radii points in the various diagrams (Flinn, 1956).

a. Bietvlei Member
Uppermost 100 feet
20 cross-lamination
11 primary flow laminations

b. Bietvlei Member
Uppermost 100 feet
35 + 30 cross-lamination

c. Bietvlei Member
Uppermost 235 feet
40 + 30 cross-lamination

d. Lower Bietvlei Member
Upper Kromberg Member
Uppermost 500 feet of the Marlow Fi
70 + 67 cross-lamination
PIEKENIER FORMATION
CROSS-BEDDING DIAGRAMS
Total distribution and +5% maxima shown.

A. DOORN BAY. N=63
B. LAMBERTS BAY. N=56
C. KLEIN ALEXANDERSHOEK. N=69
D. VLAKKRAAL, CITRUSDAL. N=46
E. BARKATSFONTEIN, AURORA. N=55
F. PIEKENIERSKLOOF PASS. N=40
G. DE HOEK, PIKETBERG. N=54
H. DASKLIP PASS, PORTERVILLE.
PALEOCURRENT MAP

PIEKENIER CROSS-BEDDING VECTOR MEAN

The dots indicate where measurements were made.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  MO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0  100 Miles

0  100 Kilometres
FACIES MAP
PIEKENIER FORMATION

Percentage coarseness and maximum pebble diameter shown.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  Wo - Worcester  St - Stellenbosch  VI - Villiersdorp

0 100 Miles

0 100 Kilometres

ATLANTIC

OCEAN

Lamberts Bay

Cape Town

Con.

Ss.
GRAAFFWATER FORMATION
CROSS-BEDDING AND RIPPLE MARK DIAGRAMS
Total distribution and ±5% maxima shown.

FIGURE 87

A. BRUINPUNT, DOORN RAY.
N=60+27

B. BELLEVUE, KLAVER.
N=37

C. FAROOSKOP, GRAAFFWATER.
N=55

D. BRANDENBURG, PEDELINGHUYTS.
N=26

E. WATerval, OLIEVENBOSKRAAL.
N=42+12

F. BOVENSTELAND, Olievenb’l.
N=26

G. BOTMANSKOP, STELENBOSCH.
N=220

H. CHAPMAN’S PEAK, HOUT BAY.
N=113+9
PALEOCURRENT MAP
GRAAFWATER CROSS-BEDDING VECTOR MEAN

The dots indicate where measurements were made.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

100 Miles

100 Kilometres
PENINSULA FORMATION
CROSS-BEDDING DIAGRAMS

Total distribution and +5% maxima shown.

A. MATJIESGOEDKLOOF, URIONSKRAAL.
N=39

B. HOLBAK, DOORN BAY.
N=40

C. MARAISBERG, CITRUSDAL.
N=59

D. GROOTPLAAT, AURORA.
N=120

E. PIEKENIERSKLOOF, CITRUSDAL.
N=108

F. Molenrivier, Gydo.
N=89

G. FRANSCHHOEK PASS.
N=58

H. MUZENBERG, SIMONSTOWN.
N=132
PALEOCURRENT MAP

PENINSULA CROSS-BEDDING (MAXIMUM CURRENT DIRECTION)
GRIDDED MOVING AVERAGE

The dots indicate sampling stations.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  MO - Worcester  ST - Stellenbosch  VI - Villiersdorp

100 Miles

0 100 Kilometres
FIGURE 91

PALEOCURRENT MAP

NARDOUW CROSS-BEDDING VECTOR MEAN
GRIDDED MOVING AVERAGE

The data indicate where measurements were made.

VA - Vansynedorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 ——— 100 Miles
0 ——— 100 Kilometres

ATLANTIC

OCEAN

Lamberts Bay

Cape Town
FIGURE 92

PALEOCURRENT MAP

NARDOUW CROSS-BEDDING PRINCIPAL MODE
(MAXIMUM STREAM FLOW DIRECTION)

GRIDDED MOVING AVERAGE

The dots indicate where readings were taken.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Control  NO - Worcester  ST - Stellenbosch  VI - Villiersdorp

Atlantic

Ocean

-- 100 Kilometres

-- 100 Miles

Lamberts Bay

Cape Town
NARDOUW FORMATION
CROSS-BEDDING DIAGRAMS
Total distribution and +5% maxima shown.

A. MELKKRAAL, LOKENBURG.
N=178

B. MARCUSKRAAL, CITRUSDAL.
N=73

C. NARDOUW.
N=215

D. PAKHUIS PASS, CLANWILLIAM.
N=82

E. DIE DRIP, CITRUSDAL.
N=36+21

F. PERDEKOP, PIKETBERG.
N=48

G. THE BATHS, CITRUSDAL.
N=56

H. DASKOP, HANGKLIP.
N=32
NARDOUW FORMATION CROSS-BEDDING VECTOR MEAN

TREND MAP

CURRIC SURFACE  STRENGTH 71.0%

Contours are lines of equal flow direction.

VA - Vanswinkeloup  CL - Clanwilliam  CI - Citruedal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villieredorp

0 0 --------------- 100 Miles

0 0 100 Kilometres

270°

225°

180°

135°
PALEOCURRENT MAP

PAKHUIS TILLITE FORMATION

GLACIAL STRIAE

FABRIC ANALYSES

(Directional)

(Non-directional)

VA - Vanrhynsdorp
CL - Clarewilliam
CI - Citrusdal
PI - Piketberg
CE - Ceres
WO - Worcester
ST - Stellenbosch
VI - Viliersdorp

0
100 Miles

0
100 Kilometres

FIGURE 95

LIMIT OF TILLITE
PAKHUIS TILLITE FORMATION

FIGURE 96

FABRIC DIAGRAMS  \( ab = \text{Ice-flow} \)  \( ac = \text{Imbrication} \)

A. Glacial striae  N=150  \( ab \)

BERGLAND. KOBE TILLITE.

B. Pebble long axes  N=90  (Visser, 1992)

PAKHUIS PASS. KOBE TILLITE.

C. \( ab \) diagram  N=106

D. \( ac \) diagram  N=123

MATJIESGOEDKLOOF, VANRHYNSDORP. KOBE TILLITE MEMBER.

E. \( ab \) diagram  N=121

F. \( ac \) diagram  N=133

BOSKLOOF, CLAWILLIAM. SNEEUWOP TILLITE MEMBER.

G. \( ab \) diagram  N=183

H. \( ac \) diagram  N=171

DISA POOL, CEDAR MOUNTAINS. SNEEUWOP TILLITE MEMBER.
PAKHUIS TILLITE FORMATION

FIGURE 97

FABRIC DIAGRAMS  \( ab = \text{Ice-flow} \quad ac = \text{Imbrication} \)

A.  \( ab \) diagram  \( N=185 \)

KUNJE, ACTER-BOKKEVELD. SNEEUkop TILLITE MEMBER.

B.  \( ac \) diagram  \( N=230 \)

KUNJE, ACTER-BOKKEVELD. SNEEUkop TILLITE MEMBER.

C.  \( bc \) diagram  \( N=52 \)

KUNJE, ACTER-BOKKEVELD. SNEEUkop TILLITE MEMBER.

D.  \( ab \) diagram  \( N=350 \)

WYEKLOOF, COLD BOKKEVELD. STEENBRAS TILLITE MEMBER.

E.  \( ac \) diagram  \( N=105 \)
FIGURE 98

PAKHUIS TILLITE FORMATION

FABRIC DIAGRAMS  \( ab = \text{Ice-flow} \quad ac = \text{Imbrication} \)

A.  \( ab \) diagram  \( N=160 \)

B.  \( ac \) diagram  \( N=214 \)

C.  \( ab \) diagram  \( N=219 \)

D.  \( ac \) diagram  \( N=182 \)

E.  \( ab \) diagram  \( N=169 \)

F.  \( ac \) diagram  \( N=136 \)

WAAJHOEK, BREDE RIVER. SNEEUKOP TILLITE MEMBER.

WELLINGTON SNEEUKOP. STEENBRAS TILLITE MEMBER.

G.  \( 8\% \quad 4\% \quad 2\% \)
    \( N=111 \)

WELLINGTON SNEEUKOP. SNEEUKOP TILLITE MEMBER.

FRANSCHHOEK PASS. PERBLE LONG AXES IN SNEEUKOP TILLITE.

(Lower hemisphere. Modified from J.N.J. Visser, 1982.)
FIGURE 99

FOLD ZONE

FOLD ZONE THICKNESS AND AXIAL TRACES

GRIDDED MOVING AVERAGE FOR AXIAL TRACES

Dots show where measurements were made.

VA - Vansryndorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

LIMIT OF TILLITE

VA

Lamberts Bay

Cape Town
FIGURE 100

DUE TO AN OVERSIGHT THIS FIGURE NUMBER IS VACANT.
FOLD ZONE POLE DIAGRAMS

EQUAL-AREA PROJECTION. LOWER HEMISPHERE.
Total distribution and +5% maxima shown.

A. SPITSKOP, CLANWILLIAM. N=88
B. VAALFONTEIN, CEDAR RANGE. N=112
C. DWARSRIVIER, CEDAR RANGE. N=79
D. ALLANDALE, CITRUSDAL. N=46
E. DIE TRAP, SHURVELI. N=50
F. DIE TRAP, PO-ROZENDAL. N=109
G. WTEKLOOP, COLD BOKKEVELD. N=32
H. HORINGSKOOF, VISGAT. N=36
FOLD ZONE POLE DIAGRAMS

EQUAL-AREA PROJECTION. LOWER HEMISPHERE.

Total distribution and ±5% maxima shown.

A. MITCHELL'S PASS. N=51
B. TIERKOP, RAANSKLOOF. N=45
C. WELLINGTON SKEEKOP. N=68
D. DU TOITSKLOOF PASS. N=62
E. HEX RIVER POORT. N=29
F. SANDDRIF, DE DOORNS. N=48
G. FRANSCHHOEK PASS. N=72
H. VAAJMAANGAT, VILLIERSDORP. N=39

FIGURE 102
FIGURE 104

TABLE MOUNTAIN GROUP THICKNESS

TRENDS MAP

CUBIC SURFACE  
STRENGTH 67.5%

VA - Vanrhynsdorp  
CL - Clanwilliam  
CI - Citrusdal  
PI - Piketberg  
CE - Ceres  
WO - Worcester  
ST - Stellenbosch  
VI - Villiersdorp

0 0 100 Miles  
0 0 100 Kilometres

ATLANTIC  
OCEAN
TABLE MOUNTAIN GROUP THICKNESS

RESIDUAL MAP

Shaded: Sediment thicker than cubic trend model.

VA - Vanrhynsdorp  CL - Clanwilliam  CT - Citrusdal  PI - Piketberg
CR - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0  100 Miles  0  100 Kilometres

ATLANTIC  OCEAN

Lamberts Bay  PI

Cape Town
ISOPACH MAP

PIEKENIER FORMATION

BASIN EDGE AND TROUGH SHAPED

VA - Vanswylendorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  MO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0  0
100 Miles  100 Kilometres

ATLANTIC

OCEAN

Lamberts Bay

Cape Town
FIGURE 107

PIEKENIER FORMATION THICKNESS TEND MAP

CUBIC SURFACE  STRENGTH 83.3%

VA - Vanrhynedorp  CL - Clanwilliam  CT - Citrusdal  PI - Piketberg
CE - Ceres  NO - Worcester  ST - Stellenbosch  VI - Villieredorp

0  0 - 100 Miles  0 - 100 Kilometres
PIEKENIER FORMATION THICKNESS

RESIDUAL MAP

Shaded: Sediment thicker than cubic trend model.

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WD - Worcester  ST - Stellenbosch  VI - Villieredorp

0  100 Miles
0  100 Kilometres

ATLANTIC

OCEAN

Lamberts Bay  Cape Town

VA  CL  CI  PI  CE  WD  ST  VI
FIGURE 109

ISOPACH MAP

GRAAFWATER FORMATION

BASIN EDGE AND TROUGH SHAPED

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  MO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0  0  0  100 Miles

100 Kilometres

ATLANTIC

Lamberts Bay

OCEAN

Cape Town
GRAAFWATER FORMATION THICKNESS

RESIDUAL MAP

Shaded: Sediment thicker than cubic trend model.

VA - Vanrhynsdorp  CL - Clanwilliam  CT - Citruasdal  PI - Piketberg  
CE - Ceres  MO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 0 100 Miles  100 Kilometres
FIGURE 112

ISOPACH MAP

PENINSULA FORMATION

BASIN EDGE AND TROUGH SHADED

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 — 0 — 100 Miles
0 — 100 Kilometres

ATLANTIC

32 33 34

Lamberts Bay

3000'
4000'
5000'
6000'

32 33 34

Ocean

Cape Town

3000'
4000'
5000'
1000'
3000'

32 33 34

17 18 19 20

100 Kilometres
FIGURE 13

PENINSULA FORMATION THICKNESS

TREND MAP

CUBIC SURFACE STRENGTH 86.4%

VI - Vansuyndorp  CL - Clovis  CI - Citrusdal  PI - Piketberg
CE - Ceres  NO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 0 100 Miles

100 Kilometres

Atlantic

Ocean

Cape Town

32

33

34

32

33

34
FIGURE 114

PENINSULA FORMATION THICKNESS

RESIDUAL MAP

Shaded: Sediment thicker than cubic trend model.

VA - Vryheynsdorp  CL - Clawsilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WC - Worcester  ST - Stellenbosch  VI - Villiersdorp

0  100 Miles

100 Kilometres
ISOPACH MAP

CEDARBERG FORMATION AND PORTION OF THE PAKHUIS FORMATION

BASIN EDGE AND TROUGHS SHARDED

VA - Vanrhynsdorp  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WO - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 ———— 100 Miles

0 ———— 100 Kilometres

ATLANTIC

OCEAN
FIGURE 116

CEDARBERG FORMATION THICKNESS

TREND MAP

CUBIC SURFACE STRENGTH 93.0%

VA - Vourhynsdorp  CL - Clanwilliam  CI - Citruedal  PI - Piketberg
CE - Coers  WI - Worcester  ST - Stellenbosch  VI - Villiersdorp

0 100 Miles
0 100 Kilometres

ATLANTIC

OCEAN

Cape Town

Lamberts Bay
ISOPACH MAP

NARDOUW FORMATION

RATIN EDGE AND TROUGH SHADED

VA - Vryheidshoop  CL - Clanwilliam  CI - Citrusdal  PI - Piketberg
CE - Ceres  WD - Worcester  ST - Stellenbosch  VI - Williersdorp

0 --------- 100 Miles

0 --------- 100 Kilometres

ATLANTIC

OCEAN

Lambert's Bay

Cape Town

17  18  19  20

17  18  19  20

17  18  19  20

17  18  19  20

VA  CL  CI  PI  CE  WD  ST  VI
FIGURE 118

NARDUW FORMATION THICKNESS

TREND MAP

CUBIC SURFACE STRENGTH 83.9%

VA - Vanrhynsdorp CL - Clanwilliam CI - Citrusdal PI - Piketberg
CE - Ceres WD - Worcester ST - Stellenbosch VI - Villiersdorp

0 0 100 Miles 100 Kilometres

ATLANTIC OCEAN

Lamberts Bay

Cape Town
FIGURE 119

NARDOUW FORMATION THICKNESS

RESIDUAL MAP

Shaded: Sediment thicker than cubic trend model.

VA - Vanrhynsdorp
CL - Clanwilliam
CI - Citrusdal
PI - Piketberg
CE - Ceres
WO - Worcester
ST - Stellenbosch
VI - Villiersdorp

0 —— 100 Miles
0 —— 100 Kilometres
GRAAFWATER EMBAYMENT & TIDAL FLAT
FIGURE 123

WINTERHOEK GLACIATION: SNEEUWkop SHEET
PLATE 1

A. The contact between the Rest conglomerate (above) and the Klipheuwel grit (below) at Meerlandsvei, Eendekuil. Note the very well-rounded cobbles and boulders set in an open framework.

B. The Rest conglomerate at Elands Bay.

C. The Rest conglomerate at Lamberts Bay. This is a bedding plane view of a lag gravel. Note the poor rounding of the boulder on which the shaft of the hammer rests.

D. A disrupted framework orthoconglomerate bed of the Rest Member at Steenboksfontein, south of Lamberts Bay.

E. A very coarse-grained immature conglomerate consisting of angular, locally derived quartzite boulders. Piekenier Formation, Fouches-hoek in the Waaihoek Mountains, Wolseley. The scale is shown by the rucksack in the lower left corner.

F. A small-pebble vein quartz orthoconglomerate bedding plane surface in the Rest Member at Weltevrede, Aurora. The scale is in inches.

G. The immature Rest conglomerate on the northern slopes of the Stettyn Rise at Jasonsfontein, Rawsonville. Note the large angular boulders.
A. The De Hoek Member at Voëlvlei, Tulbagh. Note the gradual transition from the Piekenier Formation to the Graafwater Formation.

B. The contact between the De Hoek Member and the Graafwater Formation at Gansekraal, Worcester. Note the cross-bedding laminae in the white De Hoek orthosandstone, and the thin-bedding in the Graafwater siltstone.

C. The basal thin-bedded De Hoek Member at Weltevrede, Aurora. Note the prominent thin-bedding and the absence of cross-bedding. In the absence of sufficient stratigraphical control, this succession may easily be mistaken for the Graafwater Formation.

D. Rare overturned cross-bedding laminae of the planar type in the Piekenier Formation at Muishoek Mountain, Elands Bay. Note the characteristic thick-bedding of the Piekenier sandstone beds.

E. Zebrakop, Piketberg. In the lower slopes Malmesbury schist is exposed; the lower cliff is the Piekenier Formation, here represented by the De Hoek Member; the soft slope is underlain by the Graafwater Formation (see Plate 3E), and the upper cliffs are outcrops of the Peninsula Formation.

F. The very sharp contact between the Graafwater Formation (below) and the Peninsula Formation (above) at Lekkranskloof, Heerenlogement. Note the conspicuous difference in the quality of the two formations. A low-angle minor unconformity marks the contact plane here. About 35 feet exposed vertically.

G. The upper thin-bedded portion of the Piekenier Formation at Pietersklip, Eendekuil. This succession records local low-energy deposition in parts of the Piekenier basin (see also Figure C on this plate).
A. The cliff exposure of the Graafwater Formation at Grootkop, Eendekuil. The regular thin-bedding is notable in the 250 feet cliff. In contrast, see also Figure E on this plate.

B. The Loop Member at Laingshoogte, Graafwater. Note the thin-bedding and the festoon cross-bedding. The Track Zone occurs in this member.

C. The uppermost, shaly, Loop Member at Bruinpunt, Doorn Bay. Note the regular thin-bedding. Some of the beds are perforated by hundreds of 2-3 mm Scolithus tubes (see also Plate 12A). The organism illustrated in Plate 18C was found in this succession.

D. The conglomeratic Graafwater Formation at Bruinkrans, Klawer. Note the alternation of small-pebble conglomerate beds and thin-bedded purple siltstone and shale beds.

E. The argillaceous Graafwater Formation in Versveld Pass, Piketberg. In contrast to the cliff-making quality of the Graafwater Formation as a whole about 15 miles to the north of this locality (see Figure A, this plate), the rock in Versveld Pass slakes easily and the formation underlies a soft slope (see Plate 2E).

F. Thin-bedded sandstone and siltstone of the Graafwater Formation at Ons Rus, Wolseley. The cliff is about 80 feet high.
A. Oscillation ripple marks in the Loop flagstone, Brandenburg, Redelinghuys.

B. Contact between the Tierhoek Member (left) and the Loop Member (right) near Breekwal, Klawer. The steep dip is a local feature due to the nearby fault along the Olifants River. The bank of the canal in the foreground is about 3 feet high.

C. The lowermost Graafwater Formation near Breekwal, Klawer. On the extreme right is Malmesbury phyllite overlain by the Bellevue conglomerate. The left half of the area shown consists of the Loop Member. The bank of the canal is about 3 feet high.

D. Mud cracks in the Graafwater Formation, now shown in relief due to the sand which filled the original cracks. Ons Rus, Wolseley.

E. Irregular ripple marks resembling linguoid ripples; the Loop Member at Sandberg, Redelinghuys.

F. Massively bedded Bellevue Member near Bellevue, Klawer. The pebbles are mostly poorly rounded vein quartz fragments and are set in an arenaceous matrix. The rucksack in the lower left corner indicates the scale.
A. Lenticular deposits of the Peninsula Formation on the western slopes of Kobe Mountain at Matjiesgoedkloof, Urionskraal. The lower half of the prominent cliff in the left foreground, however, is Nama (?) quartzite (see also Figures B and C on this plate). The foreground lens wedges out on the right. In the distance is another lens.

B. Close-up of the unconformity of the lens shown in Figure A. The Peninsula sandstone is here remarkably thin-bedded (following on the basal conglomerate) and contains worm trails (Plate 13D) and trilobite tracks (Plate 14D). The height of the cliff is about 25-30 feet.

C. Close-up of the basal conglomerate shown in Figure B. This was originally a coarse openwork lag gravel.

D. Disrupted framework conglomerate of the Peninsula Formation at Matjiesgoedkloof, Urionskraal. Note the marked change in texture between the lower, almost unbedded conglomerate, and the upper thin bed of reworked lag conglomerate.

E. In situ quartzite boulders at Krieskamma, Lokenburg, in the basal Peninsula Formation. Note the well-rounded pebbles which have infiltrated into the original spaces between the large boulders. The pebbles are imbricated to the right (north) in the bed below the shaft of the hammer; this is probably surf imbrication on the beach.
A, B. Two views of the Kobe tillite in the lee of the Bergland Hill at Bergland, Lokenburg. In both views the lowermost rock is Nama (?) shale. The erratics are mostly shale and subgraywacke and are set in a fine-grained argillaceous matrix. Note the rude stratification of the tillite in Figure A. See also Rust and Theron, 1964, their Plates VI, Fig. E; V, Fig. D, for another view of the Kobe tillite as well as of facetted and striated erratics from the tillite.

C, E. Two views of the unbedded Sneeukop Member; Figure C is at Bergsig, Citrusdal, and Figure E is at Donkerkloof, Citrusdal (see also Plate 9B). Note the characteristic karst type of weathering, in part due to the unbedded nature of the tillite. The arenaceous matrix is crowded with pebbles, many of which are facetted.

D, F. Two facetted erratics in the arenaceous Sneeukop tillite at Molenrivier, Gydo. Many of the erratics are facetted on two opposite sides and are characterised by their water-worn outline, which antedates the glacial abrasion.
A. The irregular contact between the Sneeukop Member (below and left) and the bedded Kunje sequence at Bergsig, Citrusdal. The low dark cliff behind the rucksack is a black, thinly bedded shale with numerous rafted sand grains in it; the shale grades upwards into the bedded pebbly mudstone, and the mudstone itself grades into a yellow shale. The Oskop sandstone is not visible in this view, but a few yards to the left a thin ripple-marked sandstone bed interposes between the Sneeukop Member and the bedded mudstone illustrated here; this means that the basal shale merely fills small hollows.

B. The regular contact between the Sneeukop Member (foreground) and the Kobe Member (top) at Boskloof, Clanwilliam; see also Figure 43. The hammer lies across a basal shale of the Kobe Member. The Oskop Member is not developed here.

C. The normally unbedded Sneeukop Member (foreground) becomes bedded where it is overlain by the Oskop Member at Cedarberg Sneeukop, Wupperthal. Note the thick-bedding of the Oskop sandstone as well as rare cross-bedding. The rucksack in the middle foreground indicates the scale. See also Figure 44.

D. A remarkably sharp and concordant contact between the Peninsula Formation and the Kobe tillite at Krieskamma, Lokenburg. This exposure is near the Bergland glacial pavement.

E. The Pakhuis Formation on the western bank of the Olifants River below the Bulshoek Barrage. The Fold Zone is not developed in the Peninsula sandstone seen in the right background and the dark Sneeukop tillite follows concordantly. The Oskop Member is only indistinctly visible and in the right foreground the argillaceous Kobe tillite is exposed. The Bulshoek varved-shale is not developed in the Cedarberg Formation which is well exposed in the gullies on the left. The canal is about 15-18 feet wide.

F. The massively bedded conglomeratic Oskop Member at Molenrivier, Gydo, is about 10 feet thick and dips away from the observer. The Sneeukop/Oskop contact appears to be behind the person on the left; he points at a facetted pebble in the Sneeukop tillite (Plate 6F).
A. The uppermost Sneeukop Member at Daskop, Hangklip. The tillite has been reworked by water action. In this area the Slanghoek Member grades into the Sneeukop Member, and in addition the Fold Zone is not developed.

B. Stream ripple marks on the upper surface of the Oskop sandstone at Wyekloof, Keerom. The direction of streamflow was southeast. Note the hammer in the foreground. The low grassed ridge in the middle background is the Steenbras Member. The Cedarberg Formation underlies the soft slope in the background and in the extreme upper right the Nardouw sandstone can be seen.

C. The type section of the Hard Band Lentil at Wellington Sneeukop. See also Figure 56.

D. Egg-shaped glacial folds at Vaalfontein near Eselbank, Wuppertal (see also Figure 101). Note the regional dip of the Peninsula Formation on the skyline. The sandstone in the Fold Zone is abnormally thin-bedded. The rounded exposures in the foreground are Sneeukop tillite. The prominent cliff in the middle is about 20 feet high.

E. Gentle folds in the Fold Zone near Witwater, Wuppertal. The rucksack indicates the scale. Note the rare anticlinal ridge and the wide synclinal trough on the left where the pebbles of the Slanghoek conglomerate are clearly visible. Regional dip is towards the observer at 15-20 degrees. The diagram of this fold system is shown in Figure 69.
A. The Fold Zone at Donkerkloof, Citrusdal. Cedarberg Sneeuberg 61 in the right background is capped by Nardouw sandstone. Note the flat-bottomed syncline in the foreground; the axial plane strikes almost due north. Note the truncated flank on the right; a close-up of this surface is shown in Plate 10D.

B. A close-up of a fold at Donkerkloof, Citrusdal. Note the thin-bedded Peninsula sandstone in the near-vertical flank on the right. Note the contrast in the mode of weathering between the bedded sandstone and the apparently massive Sneeukop tillite in the centre of the fold. The photograph holder is about 1 foot square.

C. Looking south at the Fold Zone near Wyekloof, Keerom, on the water divide between the west-flowing Tee River and the east-flowing Meul River. The cliff is 250-300 feet high. In this remarkable exposure the Sneeukop tillite is absent although normal Fold Zone/Sneeukop relations occur nearby to the east. Plate 8B shows the Oskop sandstone near this locality.

D. The Fold Zone at Gevonden, Rawsonville. Note the truncated flanks and the clear indication of plastic deformation in the 75 feet high cliff in the foreground. On the skyline the Nardouw sandstone dips away from the observer at the normal regional dip of about 25 degrees.
A. Fold Zone at Wyekloof, Keerom. Note the abnormally thin-bedded Peninsula sandstone; the laminae in the foreground are deformed cross-bedding laminae. The regional bedding dip in this area is about 5 degrees to the right.

B. Axial lineation in the thin-bedded pebbly sandstone in the Fold Zone at Schoongezicht, Tulbagh. The lineation strikes about due north, parallel to the fold axial plane.

C. Presumed glacial striae on the upper surface of the Oskop sandstone at Molenrivier, Gydo.

D. An uneven glacial floor at Donkerkloof, Citrusdal, on the upper truncated fold flanks of the Fold Zone. The Oskop sandstone is not developed in this area. Part of the fold complex shown in Plate 9A, B can be seen in the extreme right background.

E. Polished and striated glacial floor cut in Nama (?) quartzite on top of the Bergland Hill at Bergland in the Kobe Valley. The striae are parallel to the edge of the compass.
A. Enigmatic cross-corrugations on the glacial pavement at Pakhuis Pass (see also Plate 17B). The ice apparently moved directly away from the observer.

B. Ripple cross-laminated, thin-bedded Cedarberg siltstone at Rietvlei, Clanwilliam.

C. The thin-bedded upper Disa Member at Eselbank, Wuppertal.

D. The upper and lower contacts of the Cedarberg Formation at the Doorn River where it emerges from the Nardouw Range on the left. Note the prominent development of the Goudini Member especially on the left. The conspicuous hill on the right is Brandberg and it is underlain by the locally conglomeratic Skurweberg Member. In the middle foreground the Peninsula sandstone outcrops in the banks of the Doorn River. The Fold Zone does not occur here (see also Figure 36).

E. The contact between the thin-bedded, fine-grained protosandstone of the Cedarberg Formation (left) and the thin-bedded white orthosandstone of the Goudini Member (right) in a road-cutting near the Steenbras Filtration Works, Gordon’s Bay. (The rucksack in the lower middle indicates the scale).
A. A bedding plane view of closely packed 3 mm. Scolithus tubes in reddish mudstone of the Loop Member at Bruinpunt, Doorn Bay.

B. "Match Stick" structures on a bedding plane of the Loop Member at Bruinpunt, Doorn Bay.

C. A bedding plane view of tubes in Peninsula orthosandstone in the beach zone at Krieskamma in the Kobe Valley; see this plate, Figure F for a vertical section of these tubes.

D. Vertical tubes in the Skurwebeg Member at Daskop, Hangklip. Note the profuse development of the tubes.

E. Faecal pellets in the Graafwater Formation at Grootkop, Eendekuil. The scale is in inches.

F. Vertical section of the Scolithus-type tubes shown in plan view in this plate, Figure C.

G. Enigmatic cylindrical impressions on the underside of a sandstone bed in the Graafwater Formation, Farroskop, Graafwater.
A. A tube-riddled bedding plane of Rietvlei orthosandstone in the roadside at Kerskop, Wuppertal. See also Figure 66.

B. A Nardouw bedding plane with 12 mm ridged worm trails; see also Figure 67. Tengieterskloof, Doringbos.

C. A close-up of the funneled tubes which occur at Kerskop Wuppertal, and at Kransrivier, Lokenburg. See also Figure 66.

D. Remarkably serpentine trails of 12 mm worms in the Peninsula orthosandstone at Matjiesgoedkloof, Urionskraal.

E. Tortuous Helminthoida (?) burrows within a Loop sandstone bed at Waterval, Olievenboskraal. The bedding dips to the right at about 30 degrees.

F. Delicate V-shaped trails in the Skurweberg Member at Nardouw Pass, Vanrhynsdorp.

G. Sinuous 3 mm trails on a cast of a ripple-marked sandstone bed in the Loop Member at Laingshoogte, Graafwater. The scale painted on the surface in the lower right hand corner is 5 inches long.
A. Trilobite burrows in a Loop sandstone bed at Waterval, Olievenboskraal (see also Plate 13E).

B. Burrows in the Loop Member at Bruinpunt, Doorn Bay. The filling is clearly coarser than the host rock.

C. A rare cross section of a few burrows (centre right), tubes (left), and one probable U-shaped tube (top) in Faroo sandstone at Farooskop, Graafwater. See also this plate, Figure E.

D. A 6-inch wide trilobite track in the Peninsula sandstone at Matjiesgoedkloof, Urionskraal. Note the way in which the animal skirted the pebble. A small burrow appears on the right.

E. Plan view of two large hemispherical burrows filled with debris obviously coarser grained than the host rock (the scale is in inches). Faroo Member, Farooskop, Graafwater.

F. Trilobite track with a continuous tail drag mark in the Loop Member at Brandenberg, Redelinghuys (see also Figure 76). This slab is now in the custody of the South African Museum, Cape Town.
A. Wellington Sneeukop on the left crowns the type locality of the Sneeukop Member. The Sneeukop tillite is indistinctly visible as a darkish, structureless outcrop in the middle distance immediately below the Cedarberg Formation. Some very large folds in the Fold Zone are seen in the cliff of the low spur in the background. The Hard Band Lentil is particularly conspicuous (see also Plate 8C).

B. The glacial pavement at Pakhuis Pass, looking west. Note the furrowed appearance of the parallel grooves (see also Plate 17B). The exposed area of the floor is about 150 x 150 yards. Bothasberg in the background is capped by the Nardouw Formation. Note the soft slope underlain by the Cedarberg Formation.

C. A remarkable glacial floor at Groenberg, between Pakhuis and Wuppertal. Note the scribing stones embedded at the heads of the grooves. Note in the background the asymmetric ripple marks about 4-6 inches below the striated bed of arenaceous tillite.
A stereopair (please view with a lens stereoscope) of a continuous exposure of some 2500 feet of the upper Table Mountain Group at Donkerkloof, Citrusdal. Note the conspicuous difference in the bedding thickness of the Peninsula and Nardouw Formations. The Fold Zone (see also Plates 9A, B; 10D) is developed to a depth of 50-150 feet and clearly shows the truncated upper contact as well as the manner in which the deformation dies out downwards. Note the lack of outcrops of the Cedarberg Formation.
STEREOPAIRS: Please use a lens stereoscope for best results.

A. The folds of the Fold Zone at Die Drif, Grootrivier; see Figure 69. The cliff is 75-80 feet high. The regional bedding dip is practically horizontal and the upper boundary of the Fold Zone is at the top of the photograph.

B. The glacial floor at Pakhuis, looking east in the presumed direction whence the ice came. Note the uneven appearance of the grooves.

C. Enormous egg-shaped glacial folds at Bo-Rozendal, De Meul. The figure in the foreground sits near the Pakhuis/Cedarberg contact, which is also seen in the distant background. The dark uneven exposures in the right middle ground are infolded Sneukop tillite in a very large canoe-shaped or egg-shaped fold.
A. A stereopair of small, tight folds in the Fold Zone at The Baths, Citrusdal. The regional dip is 60-70 degrees to the left.

B. A plaster cast of trilobite tracks in the Graafwater Formation, Helderberg, Somerset West. Note the very characteristic toe scratches in the scooped-out footprints. See also Plate 19B.

C. A cast of a worm (?) from the upper Loop Member at Bruinpunt, Doorn Bay.
A. An extremely large worm (?) burrow on the upper surface of an orthosandstone bed in the upper Peninsula Formation in Swartberg Pass, Prince Albert. Note the cross-cutting relationships of the faecal deposit in the burrow on the right. The central median ridge can be seen indistinctly.

B. Trilobite tracks of the skew type, shown in close-up in Plate 18B. Helderberg, Somerset West.

C. Trilobite tracks in Loop sandstone at Brandenburg, Redelinghuys. Note the clear footprints with the prominent thrust heaps, and the absence of body drag marks.
PLATE 20

A. The thick-bedding of the Nardouw orthosandstone (Skurweberg Member) at Lelikkloof, Doringbos. Note the figure in the circle. The thinner bedded sandstone in the upper third of the succession is the Rietvlei Member.

B. Planar cross-bedding in the Skurweberg Member at Weltevrede, Nardouw. Note the small-pebble lag gravel consisting of vein quartz pebbles only.

C. Planar cross-bedding laminae in the Peninsula Formation, Molenrivier, Gydo.

D. Lenticular cross-bedding in the Rietvlei Member near Doringbos.

E. A bedding plane view of slumped cross-bedding laminae in the Rietvlei Member at Lelikkloof, Doringbos.
A. The sharp contact between the thin-bedded orthosandstone of the Nardouw Formation (below) and the dark subgraywacke of the Bokkeveld Group. Lelikkloof, Doringbos.

B. The abrupt Nardouw/Bokkeveld contact at Grootrivierhoogte, Grootrivier.

C. The stratigraphic sequence exposed in Matsikamma Mountain, Vanrhynsdorp (see also Figure 35). About 1200 feet of the Table Mountain Group are exposed.

D. Looking south at the very even plane of unconformity in the Bokveldsberg Escarpment between Vanrhynsdorp and Nieuwoudtville. The indistinct white line on the sky-line is the cliffs of the Nardouw Formation; the rocks underlying the Nardouw Formation are Nama grit, sandstone, graywacke and shale.