THE LATE CENOZOIC HISTORY AND PALAEOENVIRONMENTS OF THE COASTAL MARGIN OF THE SOUTH-WESTERN CAPE PROVINCE, SOUTH AFRICA

Dissertation

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ANTHONY JAMES TEMPEST TANKARD

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PREFACE

The research for this dissertation was carried out at the South African Museum from 1970 to 1975. Most of the results of this research have either been published or are in press (see references). For various reasons several of these papers have been co-authored: Tankard & Schweitzer (1974, and in press), Tankard & Krinsley (1974), Kilburn & Tankard (1975).

The two papers with Mr F.R. Schweitzer of the South African Museum present analyses of cave sediments which were excavated by him as part of his archaeological programme. But these papers were entirely geological in approach and the analyses and interpretation were entirely my own work. Mr Schweitzer provided the archaeological framework. The paper co-authored by Professor D.H. Krinsley of New York University was also entirely my own work, but I learnt a great deal from discussions with Professor Krinsley. The paper with Mr R.N. Kilburn, on the other hand, was the result of several years cooperation. I benefited greatly from his knowledge, both of molluscan taxonomy and of the relevant literature. The resulting publication was largely written by Mr Kilburn although several of the species descriptions were my work. This joint paper is not integrated into the text of the dissertation, but some notes have been extracted and included in Appendix 3, together with other original palaeontological notes.

ACKNOWLEDGEMENTS

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ABSTRACT

This thesis examines the Late Cenozoic history and palaeoenvironments of the coastal margin between Elands Bay on the west coast and Die Kelders on the south coast. This study is introduced with a detailed discussion of eustatic sea level oscillation. The history of the existing ice sheets, sea floor spreading, isotopic composition changes of the oceans, and isostatic responses of the crust to varying loads are reviewed with regard to their bearing on sea level changes.

A detailed account of the Neogene stratigraphy of the south-western Cape Province is presented. The Middle to early Late Miocene Saldanha Formation is characterised by shallow marine phosphatic sandstone and phosphorite. It is thought to have been deposited in a warm transgressive sea.

The Pliocene Varswater Formation was deposited during a secondary transgression induced by seaward tilting of the coastal margin during a time of worldwide regression. The Varswater Formation is characterised by pelletal phosphorites. It includes marine, estuarine, and fluvial facies. The estuarine sands and peats contain a rich fossil mammal fauna. Depositional environments of the Pelletal Phosphorite Member are examined by means of conventional grain size analysis to show that deposition took place on a shallow sublittoral platform dominated on the outer edge by a breaker-bar. Accretion of the breaker-bar to form a barrier-island allowed the development of an estuarine complex on the leeward side. Post-depositional diagenetic changes were examined by means of scanning electron microscopy.

A detailed account of the petrology and geochemistry of the phosphorite and pelletal phosphorite is presented. The apatite mineral is a carbonate fluorapatite. It is concluded that the phosphorite is related to upwelling of phosphorus-rich waters.

Instability of the west coast in the Pleistocene reduced the Namaqualand 45-50 m transgression complex to 10 m a.s.l.in the Saldanha area. There were three sea level peaks in the last interglacial: 6,3 m a.s.l., 2–3,5 m a.s.l, 0 m a.s.l. In the area from the Cape Peninsula to Die Kelders elevated shorelines up to 30 m a.s.l. were described, but it was found impossible to correlate these even over short distances. Thermophilic molluscs found in the last interglacial estuarine–lagoonal facies are suggestive of a warmer hydroclimate than today.

Palaeoclimatic inferences were made for the period during the Würm lowering of sea level by examining Die Kelders cave sediments. It was found that the period of glacial advance in high latitudes was accompanied by a cold, wet climate in the southwestern Cape when the coastal plain was unoccupied by man.

An attempt is made to reconstruct the history of transgression, palaeoclimate, and tectonism for the South African coastal margin. It is shown that the tectonic history could best be explained by a combination of seaward tilting of the western and southeastern margins on a broader tilting of the entire subcontinent. The Neogene was characterised by a single eustatic transgression in the Miocene, with secondary transgressions in the Pliocene being the result of coastal warping. The Miocene strata are characterised by a cosmopolitan mollusc fauna suggesting deposition prior to development of the major Antarctic ice sheet. Ocean temperatures became colder through the Pliocene and Pleistocene. The last interglacial thermophilic molluscs are attributed to a southward shift of the South Atlantic anticyclone, and solar heating of the sheltered embayments.

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CHAPTER 1

CENOZOIC SEA LEVEL CHANGES : AN INTRODUCTION

I. INTRODUCTION

Palaeogeographic studies show that all the major coastlines of the world have been subjected to alternating periods of submergence and emergence. Old high-level shorelines and submerged terraces show that sea level has oscillated considerably in the past. In 1842 Maclaren introduced his concept of glacial control to explain an oscillating sea level during the Quaternary Period. According to "classical" theory a series of stepped marine terraces was produced by a sea level which fluctuates in response to waxing and waning of continental ice sheets, and each successive high sea level was lower than the previous (Trowbridge 1954). To cite an example, transgression complexes on the Namaqualand coast have been recognized at 75-90m, 45-50m, 29-34m, 17-21m, 7-8m, 5m and 2m, and these range in age from basal Pleistocene to Recent (Carrington & Kensley 1969).

The assumption that the order of decreasing altitude of the elevated shorelines corresponds with a decreasing age was confirmed by Flemming (1968). He showed that an oscillating sea level will not necessarily obliterate previous erosion features, but that there is a general absence of random positions of high-level shorelines. A descending chronological order implies that the shorelines are the result of either a series of sea level oscillations of decreasing magnitude, or the result of a series of oscillations of approximately equal magnitude superimposed on an overall regression (Flemming 1968). The fact that the Pleistocene glaciations were of nearly equal magnitude (Flint 1971) would support the second option.

There have been several attempts in recent years to synthesize data on eustatic sea level changes (e.g. Fairbridge 1961; Guilcher 1969; Hey 1971). As Hey points out, any attempt to interpret the field-evidence becomes an exercise in correlation. South African literature on the subject of high-level shorelines is generally of doubtful quality as there are no means of dating these shorelines and reliance is placed on long-distance correlation with the "firmly established" Moroccan and other sequences. The horizontality of a shoreline over considerable distances does not disprove the possibility of tectonism (Cotton 1963; Bloom 1967), while the southern African coastlands, in particular, have a poor record of Cenozoic stability. Not only can we not separate the climatic from the tectonic influence on sea level, but it is also common practice to ascribe high-level shorelines to named Late Cenozoic stages on the mistaken assumption that the northern hemisphere climatic history is well known and the stages adequately defined.

The recent publication of new evidence bearing on eustatic sea level studies makes a discussion opportune. The purpose of this chapter will be to discuss some of the more important factors and data dealing with eustatic sea level changes. Current research on the existing ice sheets, deep sea cores and oxygen isotope studies, sea floor spreading, and the bearing these have on eustatic sea level studies will be discussed. Particular attention will be paid to glacio-eustatic sea level oscillations of the last 125 ka (125 000 years) because only for that period are there adequate and reliable data, and the effect of the ocean floor spreading would be expected to be minimal. In this thesis the term "eustatic" will apply to changes of sea level relative to a fixed datum, say the centre of the earth, and which are synchronous over the whole globe.

II. GLACIO-EUSTATIC CONTROL OF SEA LEVEL

If high-level shorelines, higher than 30m above present mean sea level (a.s.l.) for instance, were of glacio-eustatic origin it would imply that the existing continental ice sheets had melted and again built up several times in the Quaternary. The history of these ice sheets shows that this has not happened.

Although there is little evidence for the existence of large continental ice sheets prior to the Middle Miocene, some evidence suggests the presence of calving glaciers in the Eocene (Denton *et al* 1971). The first major cooling is closely associated with the Eocene-Oligocene boundary (Devereux 1967; Kennett *et al* 1974), and although Oligocene temperatures on Antarctica were close to freezing there was no development of an extensive ice sheet (Shackleton & Kennett 1974). But Margolis and Kennett (1970) document evidence of some glaciation during the Oligocene. Evidence of glaciation prior to 22 m.y. B.P. is provided by radiometrically dated basalts overlying a Tertiary tillite (Craddock et al 1964; Rutford et al 1968). It seems highly likely that the glaciers were only mountain or valley glaciers. A rich and varied fossil flora suggests not only that extensive ice sheets had not developed in the Lower Tertiary, but also that until at least the Early Miocene the climate was generally temperate (Denton et al 1971).

Analysis of cores from the floor of the Ross Sea (Deep Sea Drilling Project leg 28) suggests extensive glaciation on eastern Antarctica at least by the Early Miocene (Hayes et al 1973). Shackleton and Kennett (1974) (Deep Sea Drilling Project leg 29) have found that the Antarctic continental ice sheet developed to present thickness between the early Middle Miocene and early Late Miocene, and that by the Late Miocene – Early Pliocene the Antarctic ice sheet was much more extensive than at present. This development was accompanied by a major regression. The maximum of ice accumulation 4–5 m.y. ago was followed by an abrupt melting and ice retreat to the present position (Hayes et al 1973). Subsequent fluctuations in the Antarctic ice cover have been minor. Studies of deep-sea cores show that Antarctic glaciation has been continuous for the last 3 to 5 m.y. (Koster 1966; Goodell et al 1968).

On radiolarian evidence Fillon (1973) has shown that for the last 3 m.y., surface water temperatures south of 65^{o}S were never warmer by more than about 3^{o}C than they are today. Mercer (1968a) has found that at some time during the Pleistocene Antarctic temperatures could have been 7 to 10^{o}C higher than today, but he notes that the present -10^{o}C January isotherm lies close to the coast. A low amplitude temperature fluctuation could not lead to significant ice melting.

In the northern hemisphere full glacial conditions have prevailed in Alaska since the Late Miocene (Miller 1953; Bandy et al 1969; Denton & Armstrong 1969). In a study of several cores from the Arctic Ocean Clark (1971, 1974) showed that since at least the Early Pliocene the

Arctic Ocean has been continuously ice covered and that the present thickness of the ice is a minimum. He found that the Arctic Ocean was ice-free until the Eocene, but that the ice sheet must have formed sometime between the Eocene and the Pliocene. Shackleton and Kennett (1974) suggest a Late Pliocene development of northern hemisphere glaciation.

Melting of the existing ice sheets (east and west Antarctica, Greenland, and others) would cause sea level to rise 65m (Flint 1971). But there is overwhelming evidence that the land-based eastern Antarctic ice sheet was stable throughout the Pleistocene and could have contributed little to sea level movement. The Antarctic ice sheet possibly underwent periodic surges into the Southern Ocean, but there is little evidence to support this (Denton et al 1971). Data from deep-sea cores argue against surges, but do not eliminate the possibility of smaller surges.

Mercer (1968b, 1973) has found that the marine ice sheet in western Antarctica is more vulnerable to climatic change and its melting would have caused a eustatic rise in sea level of about 5m. Isotopic studies of the Greenland ice (Dansgaard et al 1969) suggest that only 15 per cent of that ice is residual from the last glacial and preceding interglacial age. This led Emiliani (1969) to suggest that high Late Pleistocene sea levels resulted from melting of this Greenland ice sheet and that melting would cause a eustatic rise in sea level of about 10m. Taking a conservative view, that both the Greenland and western Antarctic ice sheets contributed to Pleistocene eustatic sea level movement while the large eastern Antarctic ice sheet remained stable, the maximum glacio-eustatic rise of sea level could not have exceeded 15m at any stage in the Pleistocene.

These conclusions are broadly supported by oxygen isotope studies of deep-sea cores. Accumulation of the vast isotopically negative Pleistocene ice sheets resulted in a lowering of sea level. At the same time the oceans became isotopically positive and more saline. The original aim of oxygen isotope studies was to measure palaeo-temperature (e.g. Emiliani 1955, 1966). Shackleton and Opdyke (1973) have published a record of ocean isotopic composition changes for the last 800 ka. They note that such a record is of greater stratigraphic value than a record of temperature change.

Based on a rough equivalent of 0,1‰ to 10m sea level change, Shackleton and Opdyke (op. cit.) have drawn a glacio-eustatic sea level curve for the past 130 ka. The remarkable agreement between positions of sea level shown on this curve, and those measured on Barbados (Broecker et al 1968) and New Guinea (Veeh & Chappell 1970) proves the correctness of the method. It thus becomes possible to use the record of oxygen isotopic composition in core V 28-238 (Shackleton & Opdyke 1973, figure 9.) as a sea level curve for the past 800 ka. This record shows that sea level during isotopic substage 5e (+ 6m) was the highest in the last 800 ka, and that only twice in this interval did sea level exceed present sea level, at about 400 ka and again at about 320 ka.

In conclusion, it must be emphasised that a sea level curve derived from oxygen isotope measurements gives a reasonably clear record of glacio-eustatic sea level changes without the complicating effects of tectono-eustasism or isostatic adjustments of the globe to changing loads. The history of the existing ice sheets shows that the glacio-eustatic component could not have caused a sea level rise much above present sea level in the Pleistocene. The major Antarctic ice sheet appears to have existed in its present form since the Pliocene. It is unlikely that the history of these ice sheets could account for the major Late Cretaceous and Tertiary (Eocene and Miocene) transgressions.

III. SEA FLOOR SPREADING AND TECTONO-EUSTATIC CHANGES

The new theory of global tectonics has transformed numerous branches of geological research, and the field of eustatic sea level change has not emerged unscathed. By this concept crustal plates grow along the mid-ocean ridges and sink or override each other where they again come into contact. Changes in the rate of accretion at mid-ocean ridges will lead to changes in the depth of the oceans, since elevation or subsidence of sea-floor is a function of ridge activity. The idea that ridge activity may affect sea level has been postulated by Hallam (1963, 1971), Russell

(1968), Menard (1969), Frerichs and Shive (1971), Flemming and Roberts (1973) and Vine (1973). Flemming and Roberts have attempted to correlate eustatic changes with contemporaneous global spreading discontinuities, and can explain the major Late Cretaceous, Eocene and Miocene transgressions in this way. Hallam (1971) and Hays and Pitman (1973) believe that the Late Cretaceous transgression was due to oceanic-floor uplift consequent upon accelerated sea-floor spreading. The Miocene eustatic rise due to spreading rate changes could be of the order of hundreds of metres (Flemming & Roberts 1973). If there had been no ridge activity in the Cenozoic, and allowing for isostatic adjustment, sea level would today be 350m lower than in fact it is (Vine 1973).

According to Jacoby (1972) the effect of plate movement is to create an environment of changing density in which continental blocks of constant density will float up or down. For instance, in an environment of greater density caused by cooling of the upper mantle the continental blocks would float upwards, causing a regression. Conversely, decreasing density would result in transgression.

Rona (1973) has recorded average rates of sediment accumulation from wells on the continental shelves and slopes. He found maxima of sediment accumulation, corresponding to major transgressions, in the Middle Eocene and Miocene. Each maximum is separated by a minimum which corresponds to regressive phases. Rona equates the transgressions with a volume increase or the mid-ocean ridges resulting from fast spreading and orogenic quiescence of continents. Regression corresponds to decrease of mid-ocean ridge volume with slower spreading and orogenic activity of the continents.

Increased ridge activity in the Eocene and Miocene would adequately account for marine transgressive complexes of those ages around the southern African coast. Tertiary marine sediments are preserved in depressions in the Precambrian basement south of Luderitz. Haughton (1963) suggested that most of the fossil fauna was Miocene, and that Eocene deposits were preserved at higher elevations. In situ Middle Eocene marine sediments also occur in Mozambique (du Toit 1954). Along the South African coast the only evidence for an Eocene transgression consists of reworked deposits at Uloa (Frankel 1968) and Birbury (Bourdon & Magnier 1969). Otherwise the major transgressive complex deposits in South Africa are of Miocene age (King 1953; Frankel 1968; Ruddock 1968; Tankard in press a).

These same authors also describe Pliocene transgressive complexes. The west coast Pliocene deposits can be attributed entirely to tilting and there is no need to invoke a eustatic sea level rise. Furthermore, Frankel (1968) disputes that there are Pliocene marine sediments along the Zululand coastal plain.

Hallam (1963) suggests that due to continued subsidence of the ocean floors, sea level was generally regressive through the Pliocene and Pleistocene following the Miocene peak. A change of elevation of the ocean floors implies a subcrustal transfer of mantle material. Depression of the ocean floors would transfer mantle material towards the continental blocks and would cause them to float upwards. But there would be a transitional zone of flexure where the rising continental block was coupled to the downwarping ocean floor. Mention has already been made of the maximum of ice accumulation 4 to 5 million years ago in the Antarctic which was followed by abrupt melting and ice retreat to the present position. The time lag between this melting and attainment of hydroisostatic equilibrium could have been marked by a Late Pliocene transgression, albeit minor compared with the Miocene transgression.

IV. PLEISTOCENE HIGH-LEVEL SHORELINES

Assuming a uniform regression from the Miocene eustatic high, +300m according to Flemming and Roberts (1973), the basal Pleistocene sea surface would have been 50-60m above present. This general regression, the result of downwarping of the ocean floors, would probably be accompanied by uplift of the continental blocks and continued flexuring of the margins. The increased Late Pliocene melt water suggests a basal Pleistocene sea level even higher than 50-60m before hydroisostatic equilibrium was achieved. But this is all speculation since little is known about the course of events following a Late Pliocene melting of Antarctic ice and since our knowledge of sea floor spreading is far from complete. Bloom (1971) warns that it "would be foolhardy to infer anything about glacial-eustatic control of sea level during one of the early Pleistocene glaciations, for instance, in the face of evidence that the ocean basins are widening at rates of up to 16cm per year ..."

To obtain sets of discrete high-level shorelines such as have been described for the South African coast (Carrington & Kensley 1969; Davies 1970-1973, and others) one could speculate that it would be possible to superimpose a glacio-eustatic curve derived from oxygen isotope composition changes in the Pleistocene on a sea level regressing due to subsidence of the ocean floors. It is doubtful, however, whether this would produce meaningful results. Shackleton and Opdyke (1973) record a sea level at stage 9 (approximately 560ka) near present sea level. To raise this shoreline to +30m either by uniform rate of uplift of the coastal area, or tectono-eustatic effects, would ensure that the 120ka shoreline (+ 6m) would today be recorded at +15m. Davies (1970) suggests that the Natal 60m shoreline is of Cromerian age. The "Cromerian Complex" probably extends from 350ka back into the Matuyama Epoch (Shackleton & Opdyke 1973). Furthermore, Davies (1970, 1973) claims that the 30m and 60m beaches contain Acheulian artefacts, but the lower limit of Acheulian time is about 700 ka (Klein 1974). The oxygen isotopic composition record shows a peak at about that time, stage 19, close to the present sea level. At a uniform rate of uplift to elevate such a shoreline to +60m, the 120ka shoreline would be found today at about +16m. What is believed to be the 120ka shoreline has been identified in Zululand at +8m (Hobday in press), and in the Saldanha area at +6,3m (Tankard in press b). This type of reasoning suggests that the 30m, 60m and higher shorelines, could be no younger than the Early Pleistocene on a "stable" coast.

V. LAST INTERGLACIAL SHORELINES.

In the Late Quaternary a more realistic appraisal of sea level fluctuation is possible because: (1) the effect of sea floor subsidence due to spreading changes would be minimal, (2) elevations of the shorelines can be explained by changes in the ice sheets, and (3) adequate data is available. In this discussion, the last interglacial will be taken as equivalent to isotope stage 5, a time range 128 to 73ka (Suggate 1974).

Emiliani (1961), Shackleton (1969), Shackleton and Opdyke (1973), and Emiliani and Shackleton (1974), have analysed deep-sea cores spanning the last 500 to 800 ka. In all of these studies substage 5e of the last interglacial registers a higher palaeotemperature than the maximum post-glacial (stage 1) or any earlier stage. Substage 5e corresponds to a sea level maximum at 120 ka. Shackleton and Opdyke have derived, from oxygen isotope measurements, a glacio-eustatic sea level curve for the past 130ka. The curve shows four maxima which have been compared with estimated sea levels on Barbados (Broecker et al 1968), and New Guinea (Veeh and Chappell 1970). Only the substage 5e sea level (Barbados III) at 120ka is higher than the present sea level. There are also two other maxima in the last interglacial, both 10 to 20m lower than present sea level: BII at 100 ka and BI and 80 ka.

Uranium series dating of corals from shorelines between 1,5 and 9m above present datum from various parts of the Indian and Pacific Oceans yield dates of the order of 120ka (Veeh, 1966). Undoubtedly the most detailed and most reliable results in recent years have come from Barbados and New Guinea. Both of these islands have been uplifted at a uniform rate. On Barbados Broecker et al (1968) have recognized three last interglacial sea levels at elevations +6m (BIII), -13m (BII), and -13m (BI), and which have been dated at 122ka, 103ka and 82ka, respectively. A similar sea level curve based on a series of transgressive-regressive cycles identified from a series of coral reefs on New Guinea independently confirms each sea level stand on Barbados (Veeh & Chappell 1970). The evidence from Barbados and New Guinea for two -13m peaks at 80ka and 100ka explains why these levels are seldom recognised from continental coastlands.

On Mallorca Butzer and Cuerda (1962) recorded Tyrrhenian II shorelines at 12,5m and 7,2m with a 120ka age, and a Tyrrhenian III shoreline at 2,2m with an age of 80ka. Molluscs from the western Mediterranean and Moroccan coasts are suggestive of sea level stillstands in those areas at 120ka and 80ka.

The shoreline between 2 and 7m above present sea level which is widely recognised in places remote from plate boundaries, and which formed at 120–130ka, is becoming commonly used as a Late Pleistocene sea level datum (Chappell 1974). If this shoreline is situated higher than +7m it would infer tectonic uplift (op. cit.). Hobday (in press) has recognised three last interglacial shorelines from the St Lucia area of Zululand; 8m, 3,4–5,3m and 4,5m. Three last interglacial shorelines have also been recognised in the Saldanha area of the southwestern Cape: 6,3m, 2–3,5m and 0m (Tankard in press b). In both of these cases the highest beaches would agree with the so called Late Pleistocene datum and should therefore be unaffected by displacements such as hydroisostatic adjustments. It therefore becomes difficult to reconcile the two low levels with the BII and BI levels, although they agree well with the Mallorca series described by Butzer and Cuerda (1962). Perhaps all three of these last interglacial shorelines from South Africa should be equated with BIII.

VI. SEA LEVELS OF THE LAST GLACIAL AGE

A. Interstadial Sea Level

In North America Milliman and Emery (1968) cited 15 radiocarbon dates on carbonate samples, to predict an interstadial sea level as high as the present at 35ka. Numerous other authors have followed them (see Thom 1973, for a full discussion). However, a considerable amount of evidence has been published which shows that such a high interstadial sea level is unlikely. This subject has been treated in great detail by Thom (1973) and only a few comments will be given here.

A glacio-eustatic sea level curve derived from oxygen isotope measurements on deep sea core V28-238 shows sea level to have been considerably lower than the present level during the interstadial (Shackleton & Opdyke 1973, figure 7.). This curve agrees with the New Guinea data (Veeh & Chappell 1970) although it was derived from independent lines of reasoning, and must therefore be basically correct. The NG III data shows a corrected shoreline at –20m at 35ka. According to Broecker and van Donk (1970) sea level could not have been less than 18m from present sea level between 35 and 45 ka.

Glacio-climatic evidence proves the impossibility of sea level being close to present sea level during the interstadial (Mörner 1971). The last glaciation did not possess a warm interval of comparable intensity or duration to that which exists at present or prior to 60ka, and oxygen isotope measurements on Greenland ice cores failed to show values equivalent to an interglacial or to present conditions (Thom 1973). Continental ice fronts during the interstadial lay in southwestern Sweden and at the line of the Great Lakes in North America (Fairbridge 1971).

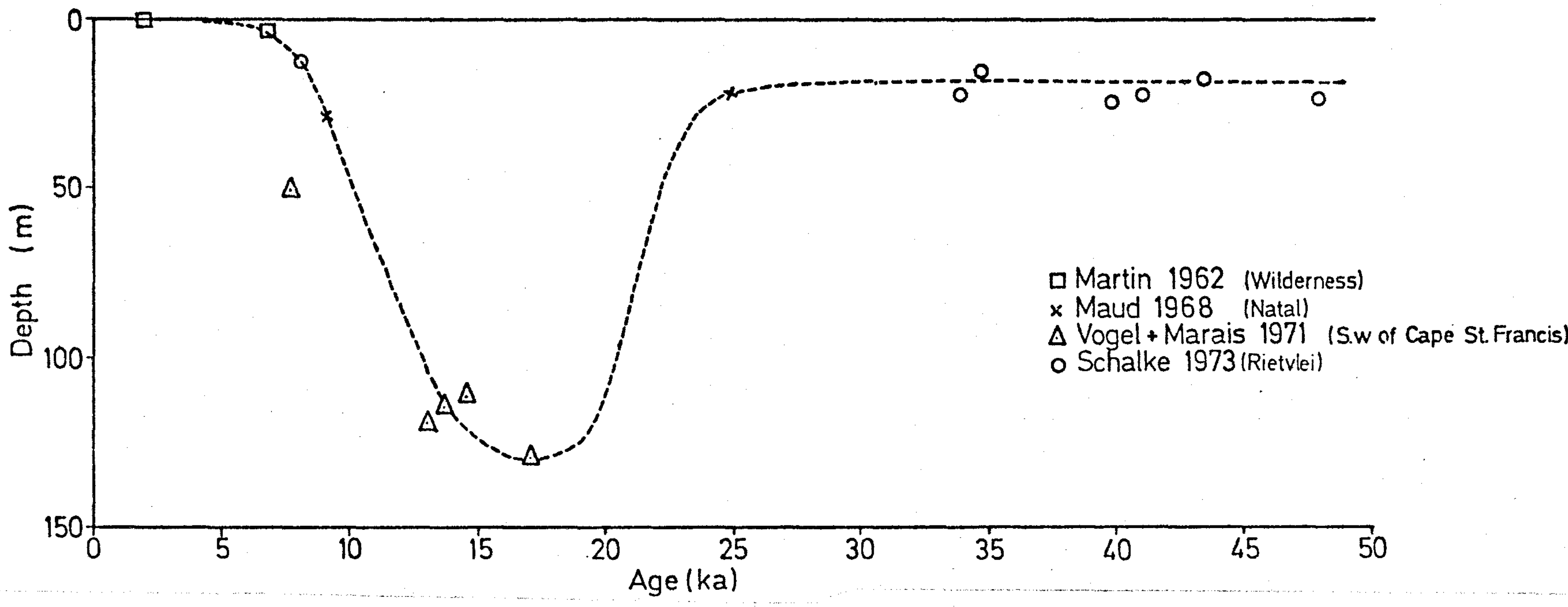
Fairbridge has recorded discussions at the eighth INQUA congress. It was stated that a high interstadial sea level to +3 to 5m relies on dated carbonate material. Studies carried out by the Columbia University radiocarbon laboratory show that all shell dates older than 20ka are subject to miniscule contamination which gives meaningless "dates" and which should be reported as "greater than .....". Fairbridge concludes that any "postulated glacio-eustatic level for this epoch anywhere near present sea level is absolutely out of the question".

Figure 1.1 is a South African time-depth plot of sea level over the last 47ka based on 20 radiocarbon dates derived from wood, peat or carbonate samples from below sea level. The chance of contamination by atmospheric CO_2 is thus minimised. These samples are all identified with shallow nearshore environments, but are unfortunately representative of widely scattered localities from west of the Cape Peninsula to Zululand. A line is drawn through the points to draw the eye, and should be regarded only as a crude approximation to sea level history over the past 47ka.

The curve suggests that between 47ka and 25ka there was a sea level maximum at approximately -20m. A lengthy stillstand at this level would be expected to leave topographic evidence. Rocky Bank, south of the Cape Peninsula, was most likely formed by a sea level at this elevation (Flemming in press), as was a prominent submerged cliff at -20m along parts of the west coast of the Cape Province and South West Africa (Wright 1964; Murray et al 1970).

Figure 1.1

Relative sea level curve for last 47ka for South Africa.



B. Glacial Maximum Sea Level

Evidence for low sea levels between 15 and 20ka, during the glacial maximum, have recently been discussed by Chappell (1974). Dated shallow marine deposits vary in depth from -65m to -150m. The most consistent results are from the Texas continental shelf. There a sea level lowering of 130m at 15ka compares with a lowering of 90m at 17ka for the eastern continental shelf and 130-170m for Australia (op. cit). Using Walcott's (1972) corrections for elastic warping, Chappell calculates a minimum sea level of -135m. As he points out this agrees very favourably with Flint's (1971) estimate of -130m calculated from ice volumes. Oxygen isotope results suggest a lowering to -120m (Shackleton & Opdyke 1973).

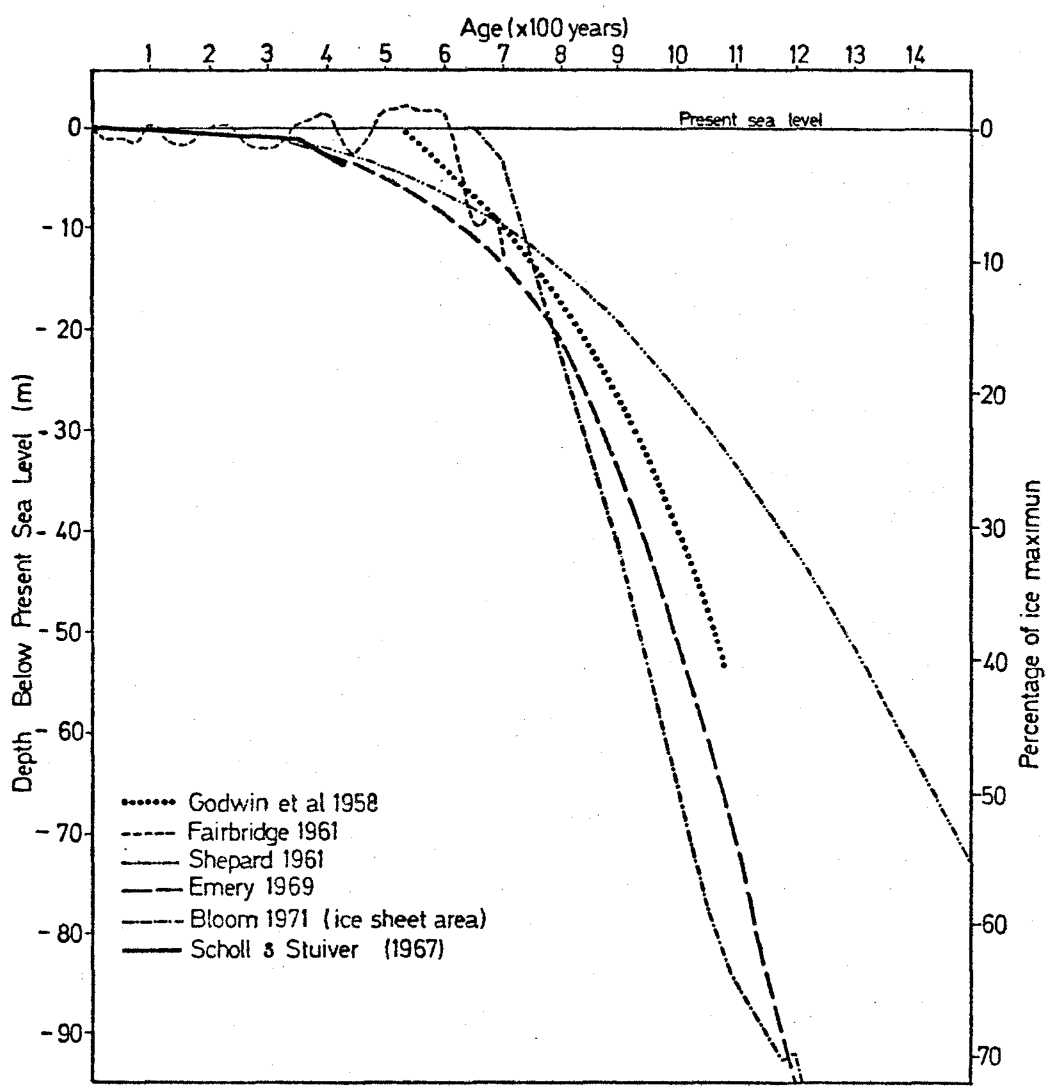
The South African time-depth curve (Figure 1.1) suggests a rapid fall of sea level with advance of the final Würm glaciation and that a minimum sea level of -130m was reached at 17-18ka.

VII. POST GLACIAL SEA LEVEL

In this section I propose to discuss the changes in sea level during the Holocene, and mainly the past 6000 years. Widely divergent opinions about the course of sea level rise persist because little account is taken of the isostatic responses of the earth crust to changing ice and water loads. The hydroisostatic effects will vary with location according to continental shelf geometry and structure of the underlying mantle. A prerequisite in any Holocene shoreline study is the accurate identification of the 125ka datum (+5 $\pm$ 3m) in that area (Thom & Chappell 1975).

Following the maximum glacial advance, the rate of retreat of the ice was nearly constant, and the present extent was reached at about 6,5ka (Bloom 1971, figure 4). Bloom's curve is reproduced in Figure 1.2 along with several estimates of sea level rise. Generally there appears to have been a rapid rise of sea level, but with the rate decreasing with diminishing time.

Figure 1.2 Holocene sea level rise, showing different concepts.



Ignoring the isostatic responses of the crust to changing loads, there is a threefold division of opinion on the course of sea level rise (discussed in detail by Jelgersma 1971). These are (Figure 1.2):

(i) The oscillating sea level concept (Fairbridge 1961). Fairbridge finds evidence, mainly from Australia, for postglacial sea levels at 3-5m a.s.l. (5ka), 1,5-2m a.s.l. (3,7ka) and 0,6-1,0m a.s.l. (2,3ka).

(ii) The steady sea level concept (Godwin *et al* 1958) suggests sea level rose rapidly until reaching the present position at 5,5ka. Sea level has, supposedly, remained constant since.

(iii) Finally, there is the continuously rising sea level concept (Shepard 1961, 1963). According to this concept the postglacial rise of sea level has been asymptotic and for the last 5ka sea level has been rising continuously. Shepard's data indicates a rapid rise from 17 to 6ka when sea level stood at -6m. This was followed by a slow rise to -1,8m between 3,5 and 2ka.

A very detailed work is that of Scholl and Stuiver (1967) on the stable Everglades coast of southern Florida. They see a rapid rise of sea level to -1m at 3ka, since when sea level has risen slowly.

Scholl and Stuiver could find no evidence in the Florida region in support of high Holocene sea levels as postulated by Fairbridge (1961). They pointed out that a slight rise in sea level in the Everglades region would have left a very substantial record. Furthermore, they question the significance of radiocarbon dates from the Everglades that were used by Fairbridge. Shepard (1960, 1963, 1964), Russell (1963), Shepard and Curray (1965), Hailes (1965) and Thom *et al* (1969) are rather sceptical of the significance of Australian radiocarbon dates used by Fairbridge (1961), Gill (1961) and Ward (1965). Thom *et al* (1969) have found no morphological or stratigraphical evidence for sea levels higher than the present in eastern Australia, which supposedly has a stable coastline, between 2985 and 9000 B.P.

Recent studies (e.g. Bloom 1967, 1971; Walcott 1972; Chappell 1974) demonstrate that the continental margins are not necessarily stationary, but that changing ice and water loads induce displacement of the earth's surface which may take the form of rapid elastic adjustments and slow viscous mantle flow (Chappell 1974). The course of sea level rise relative to the continental margins varies with location, and is affected by shelf geometry and physical structure of the underlying mantle (Thom & Chappell 1975). Bloom (1971) argues that, because the shelf off the Florida Everglades is very shallow, downwarping due to hydroisostatic compensation would be minimal, and that the curve drawn by Scholl and Stuiver (1967) would be a realistic glacio-eustatic curve. Walcott's (1972, figure 1) map of elastic warping strengthens this argument since the Florida Everglades lie on his 100 per cent contour, i.e. no vertical displacement due to elastic deformation of the earth. Florida is also remote from plate boundaries. According to Walcott (1972) no substantial change of sea level is necessary to explain sea level of the last 6000 years.

Deglaciation and the consequent sea level rise following the -130m low at about 17ka involved an average depression of the ocean floors of about 8m, and an average upward movement of the continents of about 16m (area of the continents about half that of the oceans) in the last 7ka (Chappell 1974). Furthermore, there would be a zone of flexuring where ocean basins and continents meet. These effects differ with locality. If there is a vertical movement of the continent and flexuring of their margins relative to the ocean floors, then the most desirable place to measure glacio-eustatic change would be on mid-oceanic islands which follow the movement of the ocean floors (the "dip-stick" effect).

Data from eastern Australia and Christchurch, New Zealand, both stable areas, allowed Chappell (1974) to draw a continental coast sea level curve. This curve shows present sea level to have been attained by 6000 B.P. and to have remained static relative to the continent since then. Recalculation of the sea level data from the continental curve to give movement of sea level relative to ocean basins shows that sea level has been rising continuously throughout that period and that the course of sea level rise relative to the ocean basins agrees with the results from midoceanic islands and the Florida Everglades.

The South African sea level curve (Figure 1.1) agrees with the Queensland-New South Wales results (Thom & Chappell 1975), although many more radiocarbon dates are required to substantiate its correctness. After the eustatic low at 17ka sea level initially rose very rapidly with deglaciation (167cm per 100 years). At 9000 B.P. sea level, relative to the continental coast, stood at -25m, and was within a metre of present sea level between 5000 and 6000 B.P. Detailed work in Zululand (Hobday in press) and the Saldanha area (Tankard in press b) has not revealed any Holocene shorelines higher than present sea level. Furthermore, identification of the Late Quaternary datum in these areas at 8m and 6m respectively shows comparative stability during this period.

VIII. CONCLUSIONS

Pleistocene sea level research is probably on the verge of a major revolution and studies such as that of Shackleton and Opdyke (1973) on isotopic compositional changes in the oceans should soon give us a clear account of glacio-eustasism through the Neogene and Pleistocene. Already this record, and the history of the existing ice sheets, show that glacio-eustatic sea level could not have been much higher than present datum at any time in the Pleistocene. It will require a basic change in philosophy if the record of ocean isotopic compositions is to be accepted, and a reaction against these ideas similar to that which greeted the radiocarbon dating technique can be expected. In addition it is clear that more information on the effect of sea floor spreading and tectono-eustasism in general is required.

Recent quantitative studies (e.g. Walcott 1972; Chappell 1974) show that the earth's crust responds isostatically to changing ice and water loads, and that the best place to observe the movement of sea level relative to the ocean basins is on the midoceanic islands which are remote from plate boundaries, and move with the ocean floor in its response to changing water loads.

Studies on Holocene shorelines along the continental margins take on a new dimension because of these recent trends. No longer is there need for absolute change in sea level to explain shorelines of the last 6000 years.

By measuring differences in the Holocene record with locality on the continental margins, and comparing these with data from the midoceanic islands, for instance, we should have a means of examining the structure of the underlying mantle. Holocene shoreline studies will, in future, demand a more sophisticated understanding of sedimentary, marine and crustal processes, than has been the case in the past in South Africa and elsewhere.

IX. CENOZOIC CHRONOLOGY

The Cenozoic boundaries suggested by Berggren (1971) and Berggren & van Couvering (1974) are accepted as a framework for this thesis. The Oligocene-Miocene boundary is taken at 22,5 m.y. B.P., the Miocene-Pliocene boundary at 6 m.y. B.P., and the Pliocene-Pleistocene boundary at 1,8 m.y. B.P. The Miocene may be subdivided into Lower (Aquitanian and Burdigalian), Middle (Langhian), and Upper (Tortonian).

Ideally, geological time is subdivided into zones which are based on phylogenetic successions in rapidly evolving forms which have, in turn, been related to the radiometric time scale (Berggren 1971). However, in the last few million years of the Neogene, time was too short for a satisfactory palaeontologic zonation, and a considerable amount of discussion is still taking place on the Pliocene-Pleistocene boundary and the subdivision of the Pleistocene. At the 18th International Geological Congress in 1948 it was recommended that the "Pliocene/Pleistocene boundary should be based upon changes in marine faunas, since this is the classic method of grouping fossiliferous strata" and that the "Lower Pleistocene should include as its lower member in the type-area the Calabrian formation (marine) together with its terrestrial (continental) equivalent the Villafranchian" (Berggren 1971).

Zagwijn (1974) discusses the uncertainties about the Calabrian stratotype and mammal faunas in relation to absolute datings. He believes that the base of the Pleistocene is probably at about 2,5 m.y., and not at 1,8 m.y. as previously thought. This would still be in agreement with the recommendation of the 18th International Geological Congress that the base of the Calabrian equates with the base of the Pleistocene. For convenience, in this thesis the Pliocene-Pleistocene boundary at 1,8 m.y. is adopted.

The guide fossils of the Calabrian marine stratotype are Hyalinea baltica and Arctica islandica, neither of which are known from southern Africa.

Whereas the 18th International Geological Congress has recommended the base of the Calabrian (1,8m.y.) and its terrestrial equivalent, the Villafranchian, as the base of the Pleistocene, it is found in fact that the base of the Villafranchian is older by at least 1,5 m.y. Deposits in Eurasia and Africa assigned to the Villafranchian on the occurrence in them of Equus, Elephas, and Bos (Leptobos) are probably younger than the type Villafranchian where these species are absent (Berggren 1971). Bos is not known in southern Africa and Elephas and Equus appear in Africa in the Pliocene.

In marine cores the Plio-Pleistocene boundary can be fairly accurately recognised by the planktonic foraminifera and coccolith content. The boundary is marked by the extinction of discoasters and the first evolutionary appearance of Globoratalia truncatulinoides. Onshore these fossils are not generally preserved because most of the coastal marine rocks reflect high energy shallow marine and beach environments, environments which are not favourable life habitats and which are characterised by poor fossil preservation.

The boundary between the Pliocene and Pleistocene is not as marked as other Tertiary or older boundaries. The boundary between the Pleistocene and Holocene (post-glacial) is no different from the boundaries between the preceding glacials and interglacials and it is therefore more logical to consider the Holocene as the most recent stage of the Pleistocene (West 1968). But the boundary between the Holocene interglacial and Würm glacial is accepted as 10 000 years B.P. measured in radiocarbon years by the INQUA "Subcommission on the Study of the Holocene" (1969).

The subdivision of the Pleistocene (Table 1.1) is based on West (1968), Van der Hammen et al (1971), and Shackleton and Opdyke (1973). Shackleton and Opdyke equate isotope stage 22 with the base of the glacial Pleistocene.

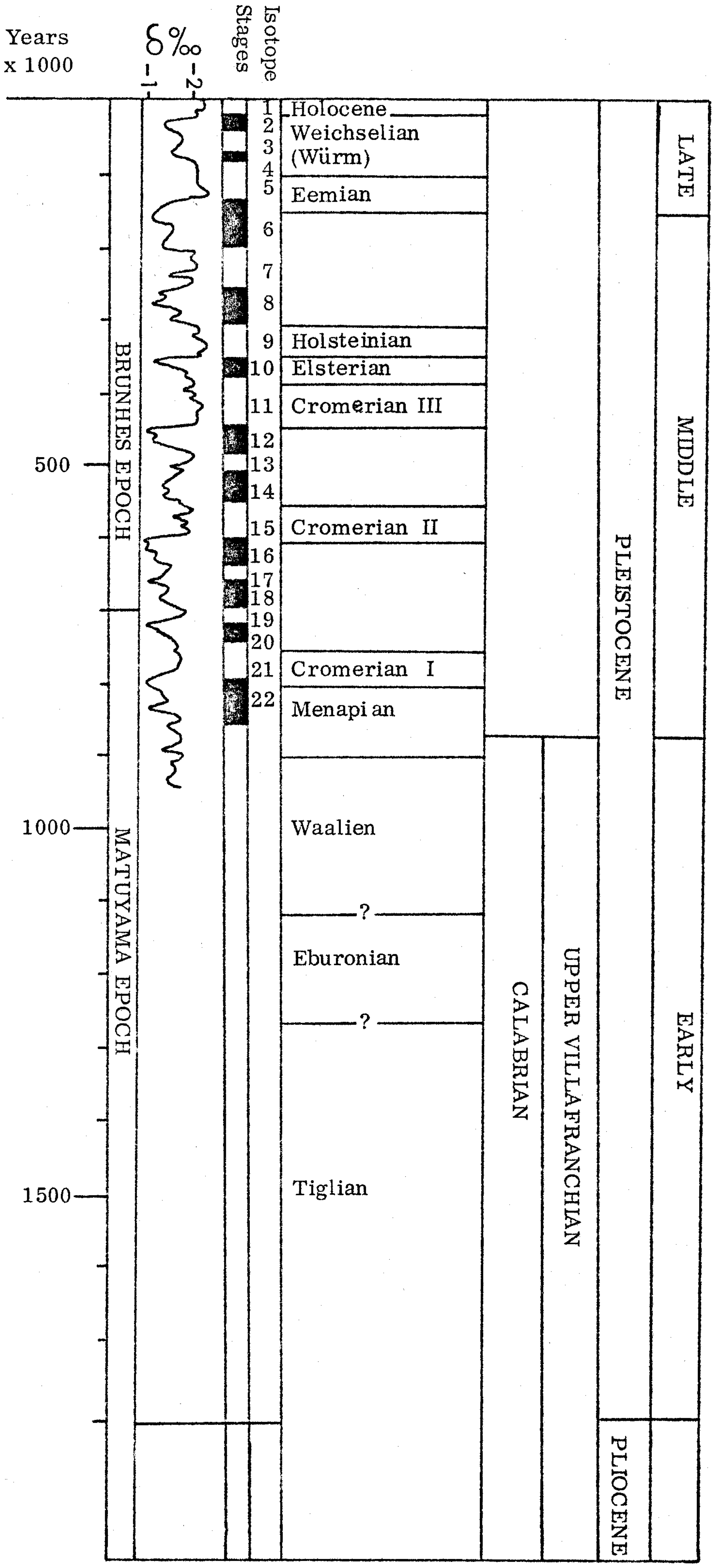


TABLE 1.1 Pleistocene Chronology

Cromerian I is characterised by a relict Tertiary vegetation in Europe (e.g. Eucommia). They restrict the Elsterian glacial to isotope stage 10, but concede the possibility that both the "Cromerian complex" and the "Elsterian complex" may extend back from 350ka into the Matuyama Epoch. The first magnetic reversal (Brunhes/Matuyama) is found in the lower part of the "Cromerian complex" (van der Hammen *et al* 1971).

The oxygen isotope record from deep-sea cores (Table 1.1; from Shackleton & Opdyke 1973) illustrates numerous global climatic changes over the last 800 ka. But the stratigraphic record in formerly glaciated continental regions is suggestive of more than four glaciations, each of which is separated by interglacials, in the last 700 ka. Kukla (1973) writes that the oscillations inferred from deep-sea cores did occur, and the apparent discrepancy between this and the continental record reflects the incomplete record in the latter, and the tendency of stratigraphers to accept the simplest model. Kukla has examined sedimentary deposits in Brno, Moravia, Bohemia, northern Austria and Slovakia, and has found evidence for eight sedimentary cycles, each of which starts in an interglacial. This would be in broad agreement with deep-sea core data.

In this study terms such as "last glacial", "last interglacial", "Eem", "Würm" ("Weichselian"), will be used as time-climate units based on European glacial events.

X. THE AIMS OF THIS THESIS

The theme of this thesis is a study of the Late Cenozoic (Neogene and Quaternary) history and palaeoenvironments of the coastal margin of the Cape Province between Elands Bay on the west coast and Die Kelders on the south coast. Because of its palaeoenvironmental bias the study will of necessity concentrate on those parameters which best describe a particular point in time. Sediments, their chemistry, and their fossil assemblages are intimately related and are the end products of processes which operated in particular environments, so that the nature of the rock and the fossil assemblage may be used together as indicators of processes and environments (Allan 1948). Allan concludes that any particular point in the geological record should be viewed as part of a landscape with physical and chemical attributes.

The Middle Miocene was a time of worldwide transgression, characterised in regions adjacent to areas of modern upwelling of cold subsurface water, by precipitation of authigenic apatite. Although the Pliocene was generally regressive, downwarping induced a secondary transgression on the west coast. In the Langebaanweg-Saldanha area the depositional record is suggestive of rapidly changing facies in a transitional zone.

Pleistocene history, on the other hand, was marked by a more rapidly oscillating sea level which left elevated shorelines close to the modern coast. Mollusc fossils from these old shorelines provide the best means for studying the environments. In the last interglacial, in particular, extant temperature sensitive molluscs and their known present-day geographic ranges provide a basis for interpreting Late Pleistocene palaeotemperature changes. But with the onset of northern hemisphere glaciation during the Würm sea level fell below present datum. Examination of cave sediments provides a wealth of data to supplement and confirm the climatic model derived from the last interglacial molluscs.

Finally, an attempt is made to reconstruct, in broad terms, the marine history of the whole of South Africa by comparing geological sequences from the south-western Cape Province with other documented sequences. But the marine history is incomplete without a discussion of the tectonic implications. Such a discussion is presented, and the picture which emerges is very much more complex than was presented in the past.

The subject matter for this thesis will be presented in order of diminishing age, starting with the Miocene.

CHAPTER 2

PHYSICAL FEATURES AND PRE-TERTIARY GEOLOGY

The southwestern Cape has a Mediterranean-type climate with hot, dry summers and mild winters. The temperature range at Cape Columbine is $19^{\circ} - 11^{\circ}C$. Average annual rainfall is 262mm with a winter maximum. The coastal waters have a temperature range of $13^{\circ} - 15^{\circ}C$ on the west coast and $15^{\circ} - 21^{\circ}C$ on the south coast (Shannon 1966). The west coast is characterised by active upwelling of cold subsurface water in the summer months. An evergreen Cape Macchia vegetation predominates and grasses are rare so that the surface is never adequately protected against wind erosion. The vegetation is denser east of the Cape Peninsula. Soils are generally skeletal.

A low-lying area, drained by the Berg River and its tributaries lies between the Hottentots Holland mountains and the Atlantic Ocean, reaches a width of 110 km at latitude $33^{\circ}S$. The higher eastern part of this lowland is known as the Swartland, and the lower western part as the Sandveld. It has been suggested (Talbot 1947; Mabbutt 1956) that the Swartland was planed by a shallow sea during the Pleistocene. Other than its plain-like surface, there is no evidence to suggest a marine origin. The Swartland and Sandveld are characterised by thick accumulations of Late Pleistocene and Recent dune sands. Typical of an aeolian landscape, the Sout River has many ponds and marshes.

A feature of interest is the salt content of the soils and rivers. Farmers east of the Sout River till a Malmesbury soil. Deep ploughing brings salt to the surface which results in a poor crop yield. A study of the Berg River (Harrison & Elsworth 1953) showed that chloride concentration increases in two steps. The first sharp increase occurs where the Berg River receives tributaries draining the 90m level. The second sharp increase takes place in the lower reaches. Du Toit (1928) wrote that NaCl is not an authigenic mineral in the Malmesbury rocks and that the chloride had a marine origin.

The pre-Tertiary geology of the west coast (Figure 2.1) comprises rocks

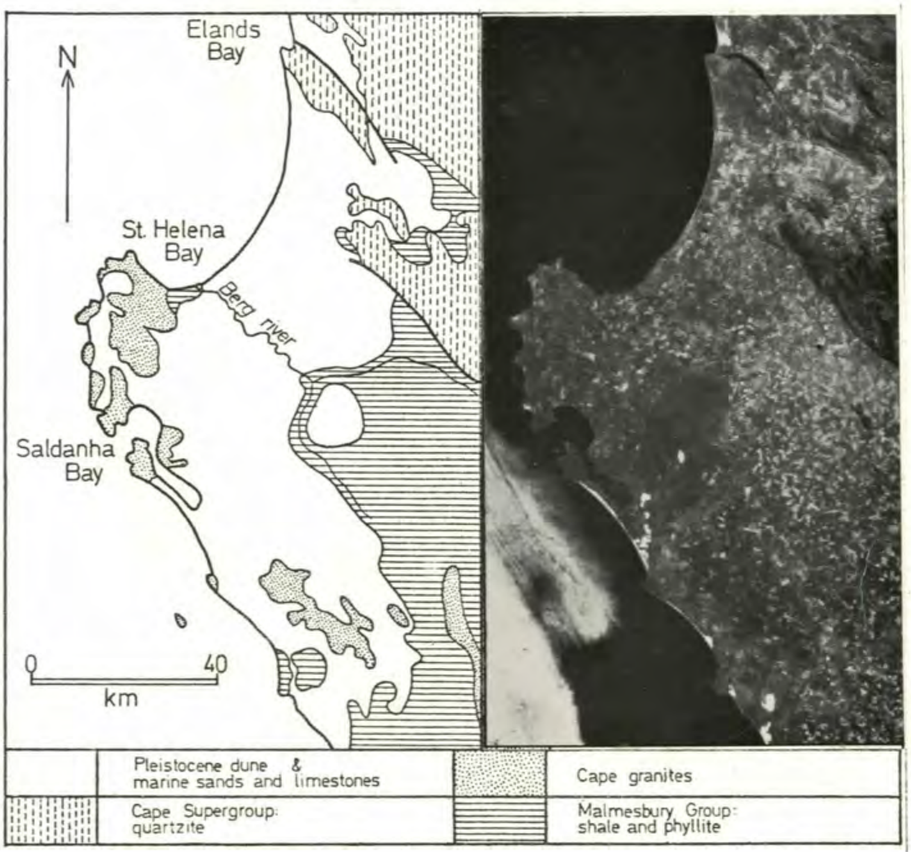


Figure 2.1 Comparison of geological map of part of the west coast with ERTS - 1 photograph shows the effect of lithology on topography

of latest Precambrian and Early Palaeozoic age. The oldest rock assemblage, the geosynclinal Malmesbury Group, consists of fine-grained greywackes and slates with associated phyllites, quartzites, and felspathic grits (Truswell 1970). On the farm Drommelvlei north of Hopefield (32°46'S; 18°25'E) a local consortium drilling for oil penetrated 3300 m of subhorizontally bedded shales and siltstones without reaching the base. In the earliest Cambrian granites were emplaced into the core of the Malmesbury anticline in two phases of similar age. The earlier phase of intrusion is marked by a medium-to coarse-grained biotite granite, and is followed by a finer grained quartz-porphyry. There are diorites with gabbro xenoliths at Ysterfontein.

The Cape Supergroup generally follows unconformably upon the Malmesbury Group and the Cape Granites. In the Elands Bay area the Cape Supergroup mainly comprises mature conglomeratic sandstones overlying red thinly bedded siltstones, shales and sandstones which are exposed along the southern shore of Verlorevlei.

Figure 2.1 shows the influence of rock type on the regional geomorphology, where the geology is compared with an ERTS-1 satellite photograph. The western part of the area is marked by the north-west trending Darling and Vredenburg plutons which form a hilly terrain. In the north and east the resistant Cape Supergroup sandstones form mountain ranges. From Elands Bay the conglomeratic sandstones form a south-east trending range which forms a natural groyne at Cape Deseada. The northern boundary of this sandstone block is a fault-scarp. Less resistant Malmesbury rocks form a negative relief. The Berg River follows the strike of the Malmesbury geosyncline. Whereas the granite and sandstone feature rocky shores, the St Helena Bay coastline on the Malmesbury Group is entirely sandy. That Malmesbury rock underlies this entire low coastal plain is shown by the frequent inclusions of that rock in the Pleistocene marine deposits between Slippers Bay and Cape Deseada.

From a point only 28 km west of Cape Columbine and trending south-southwest is the Cape submarine canyon (Simpson & Forder 1968). This canyon has probably affected sedimentation along the St Helena Bay coastline by acting as a sediment drain.

The Cape Peninsula massif consists of Cape Supergroup lying upon a granite platform. The lowest part of the Cape Supergroup, the Graafwater Formation, consists of about 70m of fine grained sandstones and subordinate mudstone layers (ratio 4:1). The Graafwater Formation is succeeded by the Peninsula Formation, nearly 1000m of orthoquartzites. Separating the Cape Peninsula from the Hottentots Holland Mountains are the Cape Flats, which are underlain by Pleistocene sediments resting on a granite-Malmesbury platform.

From Gordons Bay, on the north-eastern corner of False Bay, folded Cape Supergroup rocks form the coastline. Coastal morphology is a function of lithology and structure. Seaward dipping or sheared sandstones are conducive to the formation of a coastal platform rising gently from sea level. Where the sandstone dips steeply inland, undercutting and collapsing by high-energy waves forms a low cliff.

Surface runoff is more extensive on the south coast than on the west coast, and river systems are controlled by the structure. The largest river is the Bot River which enters Botriviervlei east of Kleinmond. The river has followed a syncline to form its course on Bokkeveld Group rocks.

Walker Bay stretches from Hermanus to Die Kelders. It has formed by the sea eroding back an inlier of Cape Granite.

CHAPTER 3

THE PRINCIPAL FEATURES AND THE DISTRIBUTION OF THE COMPONENTS OF THE WEST COAST NEOGENE DEPOSITS

I. INTRODUCTION

For the last decade the South African Museum has been involved in research directed at Late Cenozoic mammalian fossil sites of the south-western Cape. The most important of these sites , situated within the Chemfos Limited, New Varswater Quarry at Langebaanweg, is one of the richest occurrences of its type in Africa. It is the only Pliocene site being investigated in southern Africa. Since the site was first discovered in 1958 (Singer & Hooijer 1958) several accounts of the fossil fauna have been published (Singer 1961; Boné & Singer 1965; Hendey 1969; 1970a; 1974). But despite the importance of this site both palaeontologically and geologically, very little has been published on the geology of the various phosphate deposits of the area. Du Toit (1917) described in detail the aluminium phosphate deposits in the environs of Saldanha Bay, while Haughton (1932) and Frankel (1943) have briefly mentioned the bedded calcium phosphate deposits of Langebaanweg. Tankard (1974 a and b) discusses the stratigraphy and the petrology of the phosphorite in detail. A recent report of the Cenozoic sediments is that of Visser and Schoch (1973). Other accounts deal only superficially with the geology of the area (Talbot 1947; Mabbutt 1957, 1959; Boné & Singer 1965; Singer & Wymer 1968; Hendey 1970a, 1970b,1974).

The Neogene geology comprises strata of Middle Miocene and Pliocene age. Rocks of Miocene age include the phosphatic sandstones and microsphorite of the Saldanha Formation (Tankard in press a), and marine and aeolian limestones of the Bredasdorp Formation. The Pliocene Varswater Formation is characterised by a cyclothem dominated by pelletal phosphoritic sands, with beach gravels, kaolinitic clays, peat and carbonaceous sand, and fine quartzose sand. These bedded Neogene sediments occur principally around the perimeter of an erosional basin between the Berg River and Saldanha, while a small inlier occurs at Ysterplaat, Cape Town.

The aim of this chapter is to describe the geographic distribution of the Saldanha and Varswater Formations.

II. SALDANHA FORMATION

The Saldanha Formation is a Miocene marine transgressive complex taking its name from the town Saldanha. The stratotype is situated about the "Bomgat" on the Hoedjiespunt peninsula at Saldanha Bay ($33^{o}01,7$'S; $17^{o}57,4$'E) (Figure 3.1, Site 1). The phosphatic sandstone exposed in the floor of the New Varswater Quarry at Langebaanweg (Figure 3.1, Site 2) would perhaps have been preferable, but being a mine it is unlikely to remain exposed indefinitely. However, it will be used as a reference stratotype so as to describe the formation more completely.

Another occurrence of phosphatic sandstone of Miocene age occurs at Ysterplaat, Cape Town (Figure 3.1, Site 3). North of Hondeklipbaai ($30^{o}19$'S; $17^{o}16$'E) phosphatic sandstone is preserved as bedrock-hollow infills and extends up to 36m above sea level where it is truncated by a wave-cut erosion surface (Tankard 1974a).

The Hoedjiespunt microsphorite and the Langebaanweg phosphatic sandstone have been referred to informally as the "basal bed" (Tankard 1974a, 1974b, 1974c). Hendey (1974) included the Langeberg (Langebaanweg) phosphatic sandstone in his "bed 1", the remainder of "bed 1" consisting of Varswater Formation sediments. Du Toit (1954) equated the Hoedjiespunt microsphorite with the eastern Cape Alexandria Formation.

All occurrences of the Saldanha Formation appear to belong to a Middle Miocene transgressive complex. The most important single feature of these rocks is the occurrence of autochthonous phosphorite. They are lithologically distinct and sufficiently well developed to merit designation as a formal stratigraphic unit. This recommendation was accepted by the South African Committee for Stratigraphy: Tertiary-Quaternary Working Group in February 1975 after a field inspection.

The only other onshore deposits of suspected Miocene age consist of limestones exposed in a quarry 4km northwest of Saldanha. They consist

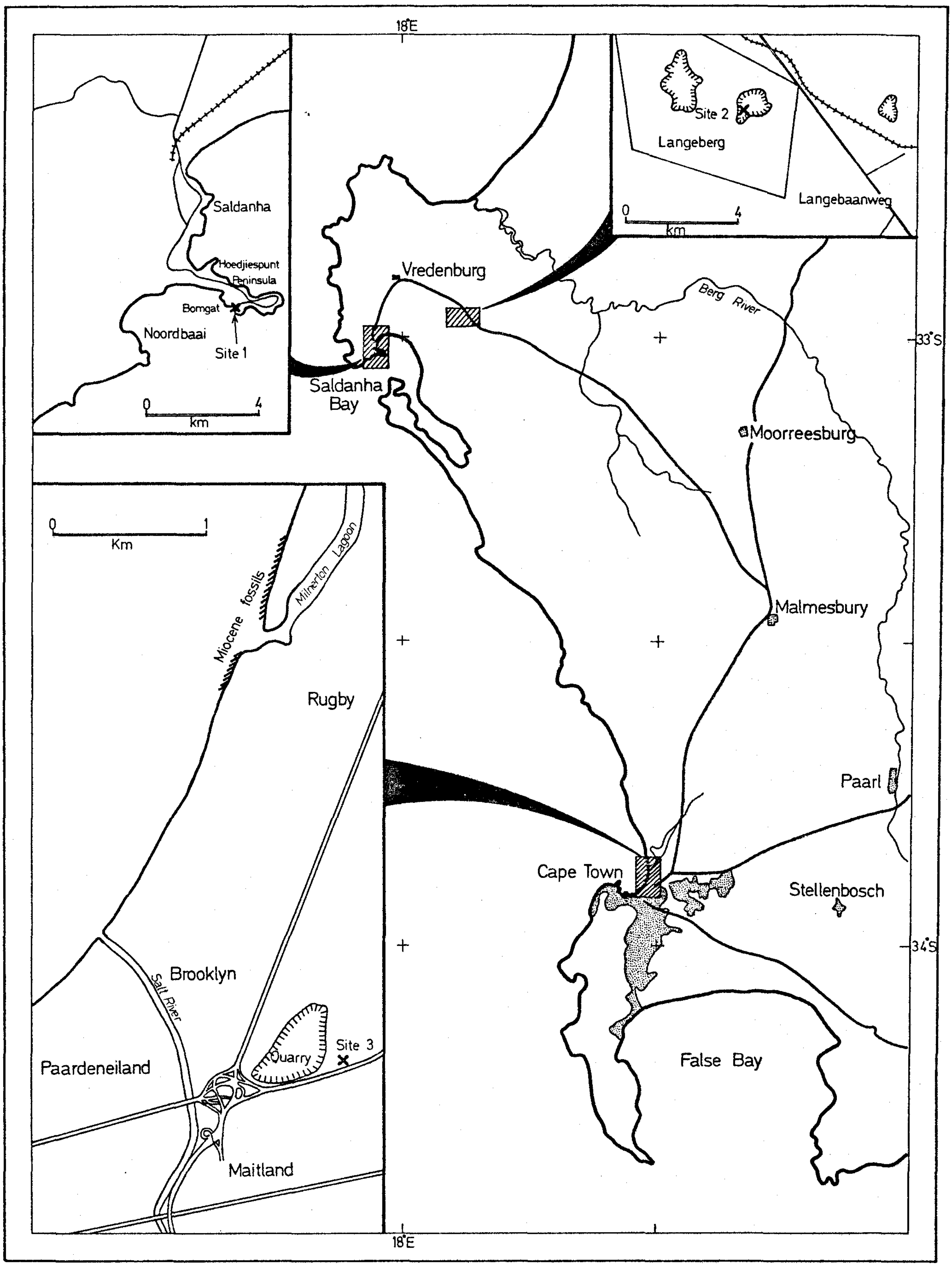


Figure 3.1 Locality map for Saldanha Formation.

of a massive marine limestone, more than 3m thick, overlain by high-angle cross-bedded aeolianite. The interface is at 56,4m a.s.l. At Die Kelders Bredasdorp limestone overlies steeply dipping Peninsula sandstone.

III. VARSWATER FORMATION

Although Hendey (1970a) used the name Varswater bed for the sedimentary succession at Langebaanweg, he later accepted (1974) that these sediments should be termed the Varswater Formation. That it merits designation as a formation is obvious in view of its observable lithologic separation from adjacent sediments both above and below, as well as the fact that it is recorded over a wide area, viz. Paternoster, Saldanha Bay area, and Langebaanweg area, and it consists of a distinctive succession of lithologic types. The Varswater Formation takes its name from the New Varswater Quarry ("E" Quarry of Hendey 1970a, b), the Chemfos Limited mine at Langebaanweg, wherein much of the sedimentary succession is visible. A complete description of the stratigraphy, sedimentology, and petrography of the succession will be given. The field work for this study began in 1970. Much use has been made of borehole logs, over 120 logs on Langeberg and a further 95 borehole logs on Witteklip-Sandheuwel-Langlaagte, besides numerous borehole logs from other uneconomic prospects. Borehole samples from 54 boreholes on Langeberg and Witteklip have provided ample material for analysis.

The name Varswater Formation was accepted by the Tertiary-Quaternary Working Group of S.A.C.S. in February 1975.

Hendey (1974) applied the name Varswater Formation to the phosphatic sandstone, silty sands, and pelletal phosphoritic-quartzose sands exposed in the New Varswater mine on the farm Langeberg at Langebaanweg. His "bed 1" consists in part of a phosphatic sandstone which is now included in the Saldanha Formation. The Saldanha Formation phosphatic sandstone at Langebaanweg was truncated at 30m a.s.l. by the transgression which gave rise to the Varswater Formation.

The stratigraphy of the Varswater Formation is summarised in Table 3.1,

TABLE 3.1 - STRATIGRAPHY OF THE VARSWATER FORMATION AND ASSOCIATED SEDIMENTS

Age	Stratigraphic Unit		Maximum thickness	Lithology
Quaternary	Pleistocene aeolian sands		41 m	Calcareous - quartzose sands, medium to fine grained, moderately sorted.
Quaternary	River channel sediments		2 m	Greenish-white clayey sands with channels of clayey sand and phosphatic sandstone
Pliocene	Varswater Formation	Pelletal Phosphorite Member	25-28 m	Upper and lower boundaries defined as 2% P_2O_5 cutoff. Moderately sorted, medium to fine phosphatic - quartzose sands. Phosphate present as sub-spherical pelletal phosphorite. Lenses and concretions of phosphatic sandstone.
Pliocene	Varswater Formation	Quartzose Sand Member	7 m	i. Estuarine facies: yellowish brown sandy silt, poorly sorted. 30-40% mud. Little phosphate. Mammal fossils abundant. Grades laterally into peat. ii. Fluvial facies: coarse sand intercalated with fine sand. Moderately sorted quartzose sands.
Pliocene	Varswater Formation	Gravel Member	0,5-2 m	Consolidated quartzose sands, frequently phosphate mineralised. Well-rounded beach cobbles and gravelly-sands. Shell casts of molluscs of warm water affinity.
Pliocene	Varswater Formation	Kaolinitic clay (freshwater)	6 m	Grey-black carbonaceous clay, pyrite, grass and _Podocarpus_ pollens.
Pliocene	Varswater Formation	Quartzose sand (unknown origin)	?	Fine quartzose sand

UNCONFORMITY - TILTING

Age	Stratigraphic Unit	Maximum thickness	Lithology
Middle Miocene	Saldanha Formation	0,5-1,5 m	Generally poorly sorted, fine grained phosphatic - quartzose sandstone. Hoedjiespunt: massive lens of bedded microsphorite, less than 1% quartz.
Middle Miocene	Pre-Saldanha Formation		

TABLE 3.2 – Correlation of terms used in this dissertation with those of Hendey (1974)

<table>
<tr><th colspan="4">Hendey (1974)</th><th colspan="3">This Dissertation</th></tr>
<tr><th>Age</th><th colspan="2">Stratigraphic Unit</th><th>Faunal Unit</th><th>Age</th><th colspan="2">Stratigraphic Sequence</th></tr>
<tr><td rowspan="2">QUATERNARY</td><td colspan="2" rowspan="2">Surface Bed</td><td rowspan="2">–</td><td rowspan="2">QUATERNARY</td><td colspan="2">Pleistocene aeolian sands
41m</td></tr>
<tr><td colspan="2">River channel sediments
2m</td></tr>
<tr><td rowspan="8">PLIOCENE</td><td rowspan="6">VARSWATER FORMATION</td><td>Bed 3b
- - - - - - -
Bed 3a
7,5m</td><td>–

Estuarine faunal unit 2</td><td rowspan="6">PLIOCENE</td><td rowspan="5">VARSWATER FORMATION</td><td>Pelletal Phosphorite Member
25–28m</td></tr>
<tr><td>Bed 2
2m</td><td>Estuarine faunal unit 1</td><td>Quartzose Sand Member
(Estuarine facies, fluviatile facies)
7m</td></tr>
<tr><td rowspan="4">Bed 1
1m</td><td rowspan="4">Marine faunal units 1 & 2</td><td>Gravel Member
0,5m – 2m</td></tr>
<tr><td>Kaolinitic clay
(freshwater)
6m</td></tr>
<tr><td>Fine quartzose sand,
unknown origin</td></tr>
<tr><td colspan="2">- - - - - - unconformity - - -</td></tr>
<tr><td colspan="3" rowspan="2">Clay</td><td rowspan="2">MIOCENE</td><td colspan="2">Saldanha Formation
(basal bed)
0,5–1,5m</td></tr>
<tr><td colspan="2">Mottled silty clay</td></tr>
</table>

while Table 3.2 shows the correlation with Hendey's terminology.

The type area of the Varswater Formation is the New Varswater Quarry on Langeberg (Figure 3.2). At its type locality the Varswater Formation is only about 10m thick, but elsewhere the total thickness is 39-43m. It consists largely of unconsolidated fine grained quartzose sands with the Pelletal Phosphorite Member rich in exploitable pelletal phosphorite. The upper boundary of the Varswater Formation has been taken (Tankard 1974a) as the 2 per cent P_2O_5 cutoff. The base of the formation is defined by the quartzose sands, kaolinitic clay, gravels and cobbles, or peats and carbonaceous sands. In the absence of all these lower units the base of the formation is taken as the base of the Pelletal Phosphorite Member which is the 2 per cent P_2O_5 cutoff, or the contact between higher concentrations of pelletal phosphorite and the Saldanha Formation.

The gross lithology of the various units of the Varswater Formation are summarised below, starting with the youngest.

A. Pelletal Phosphorite Member

Maximum thickness: Langeberg 25m; Sandheuwel 28m.
Moderately sorted, fine phosphatic-quartzose sands. Quartz rounded to well-rounded, polished. Phosphate present as pelletal phosphorite which is largely subspherical in shape, medium to fine sand size. Occasional iron oxide staining of quartz and pelletal phosphorite. Lenses and concretions of phosphatic sandstone common throughout demonstrating post-depositional phosphate mineralisation. Heavy minerals always less than 1 per cent. Ilmenite constitutes about 97 per cent of the heavy minerals, garnet 3 per cent, zircon 1 per cent. Shark teeth, mollusc shells and shell casts occur throughout; limited terrestrial fossils at base. Microfauna: phosphatised foraminifera fragments, mainly *Elphidium*; echinoid spines; minute fish teeth, coprolites. Maximum elevation 54m on Langeberg, Witteklip-Sandheuwel and Paternoster; 50m on Groot Springfontein; 47-50m on Duyker Eiland.

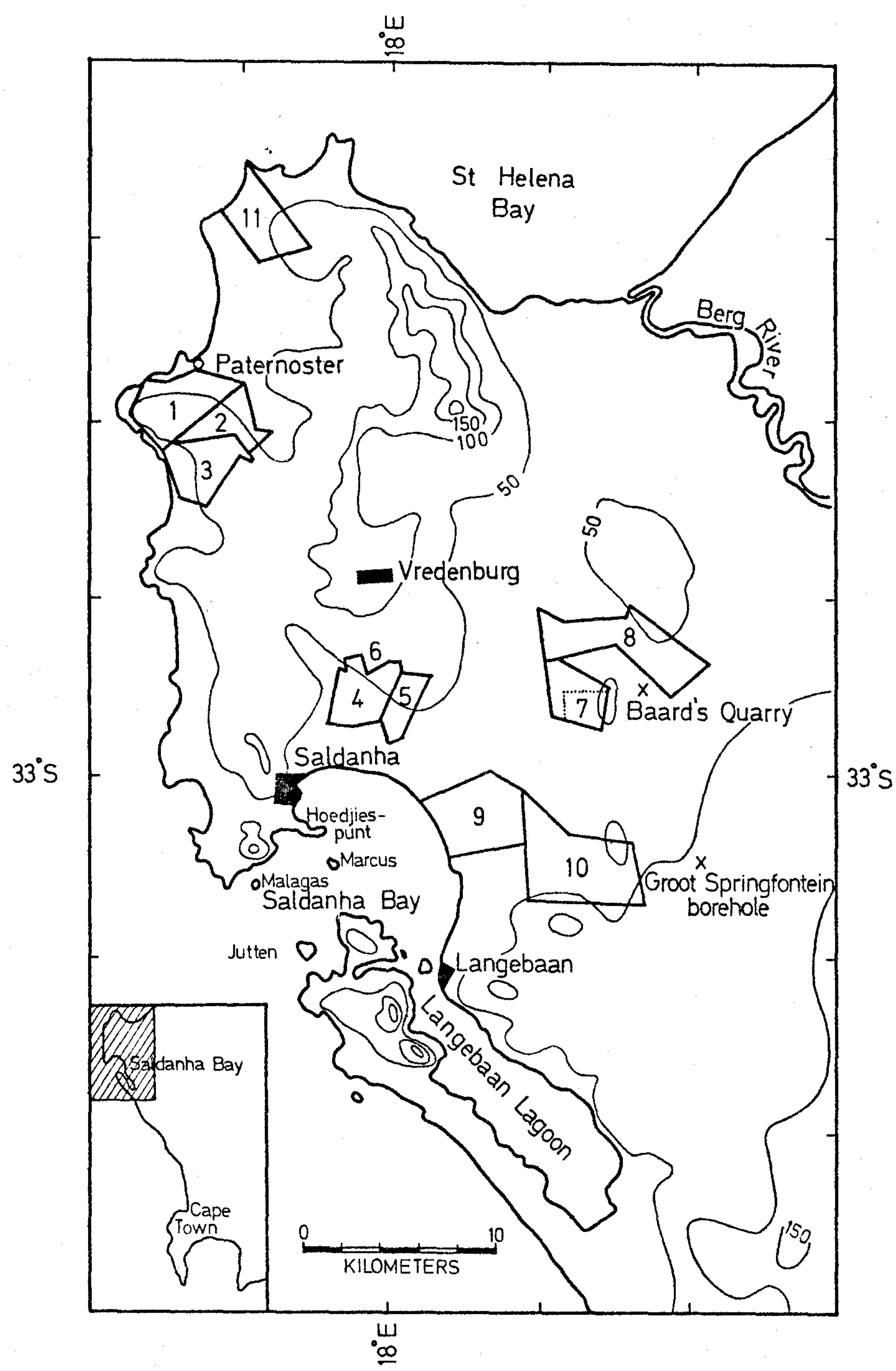


Figure 3.2 Locality map for Varswater Formation sediments. 1, Pelgrimsrus; 2, Noodhulp; 3, Besterskraal; 4, Sandheuwel; 5, Langlaagte 6, Witteklip; 7, Langeberg; 8, Muishondsfontein; 9, Tiekosklip; 10, Waschklip; 11, Duyker Eiland.

B. Quartzose Sand Member

1. Estuarine facies

Present only on Langeberg. Thickness about 2m. Moderate yellowish brown (10YR 5/4) to pale orange (10YR 8/2) sandy silt. Poorly sorted. Mud fraction 30–40 per cent. Negligible phosphate. Upper limit marked by phosphatic sandstone layer. Large numbers of terrestrial vertebrate and occasional marine animal remains. Maximum elevation 30m a.s.l. Grades laterally into carbonaceous sands and peat with *Podocarpus* and grass pollens.

2. Fluvial facies

Maximum thickness about 7m. Layers of coarse sand intercalated with fine sand. Moderately sorted quartzose sands. Very occasional phosphorite fragments. No faunal element.

C. Gravel Member

Thickness about 0,5m on Langeberg, 1,5–2m on Sandheuwel. Consolidated quartzose sands, frequently phosphate mineralised. Associated with these are well-rounded, discoidal beach cobbles and gravelly-sand layers. Casts of marine molluscs of warm water affinity. On Sandheuwel a coarse gravelly-sand and coquina of mollusc and barnacle shell fragments; ostracods and foraminifera common (9:14); echinoid spines. Maximum elevation on Langeberg 30m, Sandheuwel 24m.

D. Kaolinitic Clay

Maximum thickness 6m. Grey-black carbonaceous clay, pyrite inclusions, intercalated pyritic sands. A few pollens; *Podocarpus* and grass pollens recorded. It overlies fine quartzose sand of unknown origin.

E. Distribution of the Varswater Formation

The Varswater Formation is a "classic" marine transgressive complex that is preserved mainly in an embayment or erosional basin lying between Saldanha and Langebaanweg, although exploratory drilling has revealed isolated pockets at Paternoster and Duyker Eiland (Figure 3.2) which face the Atlantic Ocean.

The oldest sediments of the Varswater Formation are black-grey carbonaceous kaolinitic clays, overlying a quartzose sand. (This quartzose sand is known only from borehole logs, and, in fact, could be of Miocene-Early Pliocene age). The kaolinitic clay horizon is probably diachronous. A brown-black carbonaceous clay is encountered in the basin in the granite floor on Langlaagte close to sea level. The clay is first encountered on Langeberg at 10m a.s.l. where it is horizontally disposed (slope 1:360), and extends to the foot of the first rise (Figure 5.3)(to be discussed). A further sub-horizontal clay horizon migrated across the terrace area with a further rise of sea-level to about 30m a.s.l. On Langlaagte (Figure 3.2) the clay horizon was encountered in a deep depression in the granite floor close to sea level. South of a WSW-ENE granitic ridge which divides the farm Waschklip, boreholes have intersected the clay horizon preserved in bedrock depressions.

Within the New Varswater Quarry the Gravel Member, consisting of gravelly-sand, discoidal phosphatic sandstone beach cobbles, and lenses of fine quartzose sand with mollusc moulds, lies directly on the Miocene phosphatic sandstone. Boreholes show the gravel horizon with shark teeth to be widespread along the outer edge of the terrace. On Sandheuwel a coarse coquina with complete oyster valves is encountered at 24m a.s.l. (borehole S'58). At Duyker Eiland this horizon crops out at 30m a.s.l.

The Quartzose Sand Member, with its fluvial and estuarine facies, is restricted to Langeberg. Fluvial sediments were encountered to the north-east (boreholes M6-15-N6-15 and N4-76-04), while estuarine-type sediments were encountered within the floor the New Varswater Quarry. Recent excavations have demonstrated that the estuarine facies consists of fine-grained quartzose sands with an extensive fossil mammal assemblage, and that these sands grade laterally, in a southerly direction, into carbonaceous sands and peat.

Within the New Varswater Quarry the Pelletal Phosphorite Member overlies the estuarine sediments, displaying a typical onlap aspect. The thickest part of the Pelletal Phosphorite Member occurs on Langeberg where it reaches a thickness of 25m. On Langeberg this member is encountered up to 53-54m a.s.l.

On Waschklip the Pelletal Phosphorite Member is encountered at 7-10m a.s.l. The configuration of this member in the Witteklip-Sandheuwel area (Figure 3.2) is controlled by the granite bedrock contours (Figure 5.2). On Tiekosklip the succession commences with a gravelly sand at 15m below sea level, and is followed by an upward-fining sequence. Although pelletal phosphoritic sands were encountered during exploration in the Paternoster and Duyker Eiland areas, concentrations were generally too low to be of economic importance.

The Varswater Formation is overlain by calcareous shelly-quartzose sands of aeolian origin. They are thought to be of Pleistocene age and will be discussed fully along with other Pleistocene sediments. An interesting feature of these aeolian sands are the "phoscrete" horizons. These are pedogenic horizons that are phosphate cemented by phosphate that has been derived from the Tertiary phosphorites. Also to be discussed in more detail later are the Baards Quarry Pleistocene channel sediments which have also been mineralised by solutions derived from the Tertiary phosphorites. Hendey (pers. comm.) recognises two fossil faunas from the channel sediments. The older of these, which has been reworked, he believes is closely related to Pliocene fossils on Langeberg.

CHAPTER 4

THE SALDANHA FORMATION AND PRE-SALDANHA NEOGENE SEDIMENTS

I. INTRODUCTION

Occurrences of phosphorite were first reported from the Agulhas Bank by the Challenger Expedition of 1873-1876 (Murray & Renard 1891), while Collet (1905) described phosphorite dredged between the Agulhas Bank and 25°E longitude. More recent studies of the offshore phosphorites include those of Parker (1971), Parker and Siesser (1972), and Summerhayes (1973). Tankard (1974b and in press a) has described the petrology of offshore phosphorites and phosphatic sandstones from the Saldanha-Langebaanweg area, which were informally referred to as the "basal bed". Brief mention has also been made of similar deposits north of Hondeklipbaai (Tankard 1974a). All of these occurrences appear to be of shallow marine origin, and of Miocene age. Lithologically they are phosphatic sandstones, phosphorites, or microsphorites. It was proposed that the occurrences at Hoedjiespunt (Saldanha), Langebaanweg, Ysterplaat (Cape Town), and Hondeklipbaai should be grouped together as the "Saldanha Formation", a new lithostratigraphic name. Furthermore, this name should also encompass the phosphatic sandstones and phosphorites of the South African continental shelf.

II. PRE-SALDANHA DEPOSITS ON LANGEBERG

In the first half of 1975 Chemfos Limited sank two boreholes through the phosphatic sandstone which forms the floor of the New Varswater Quarry at Langebaanweg in order to ascertain the depth of bedrock. Strongly flowing water stopped boring operations after only 20m had been cored. Although the depth to bedrock is still unknown, a very interesting sedimentary sequence was encountered (Figure 4.1). This sedimentary succession must be the most complete and most complex Neogene sequence so far recorded in southern Africa. The preservation of unconsolidated Miocene deposits below the Saldanha Formation may be attributed to the

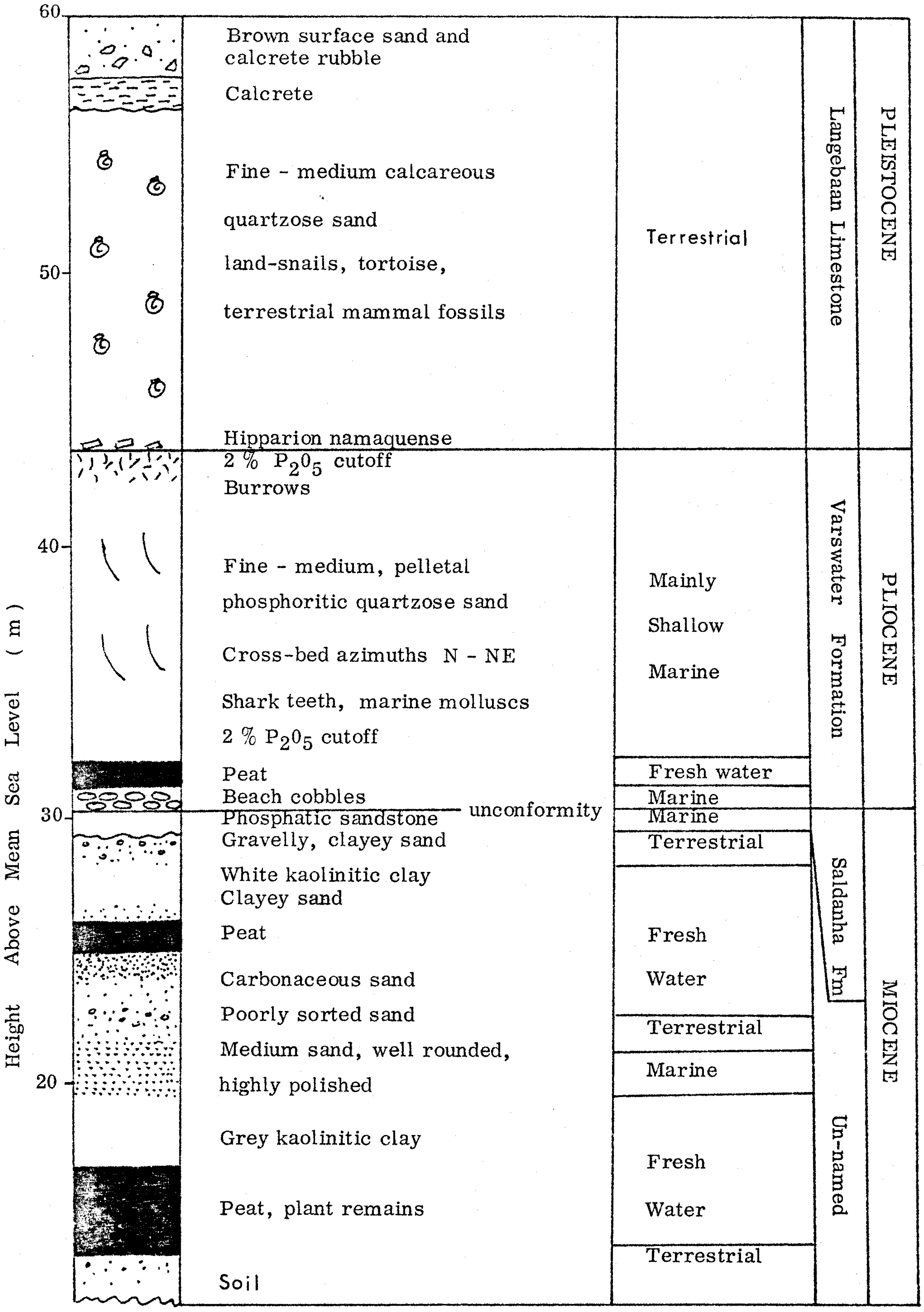


Figure 4.1. Composite section through Neogene sediments on Langeberg drawn from borehole log data and quarry exposures.

fact that phosphatisation may be a rapid process, and once the overlying phosphatic sandstone had formed the underlying strata would be largely protected from possible erosion.

The outstanding feature of these pre-Saldanha sediments are the two sedimentary cycles, each of which is dominated by peat and overlying clay. The cycles are separated by a marine unit. Generally, the sequence is suggestive of slow transgression.

The oldest sediments encountered in these two boreholes consist of brown, poorly sorted, loamy, quartzose sand, typical of a regolith deposit. The first peat horizon overlies this soil. The peat is 3m thick and contains macro-plant remains (i.e. a peat sensu stricto). The upper boundary of the peat is gradational through a grey-black clay with carbonaceous material into a pure grey "kaolinitic" clay.

Two metres of medium grained beach sand overlie the clay. The sand is well sorted and the quartz grains well rounded and highly polished (a wet lustre); typical of a beach sediment.

The sedimentary cycle is again repeated: poorly sorted sand through carbonaceous sand, peat, clay, and finally another poorly sorted gravelly, clayey sand which immediately underlies the phosphatic sandstone.

It is likely that these deposits formed close to sea level. The peat probably accumulated at sea level, but was protected from the sea by a barrier. The upward gradation through the clay is indicative of a deepening freshwater environment. Either breaching of the barrier or tectonic lowering could account for the marine incursion. But the limited extent of the marine unit which overlies other unconsolidated sediments suggests a minor event of very short duration which is more readily explained by breaching of a barrier. Further development of the barrier permitted repetition of the cycle until another marine incursion took place. This second flooding was accompanied by phosphate mineralisation to form a phosphatic sandstone.

Professor E.M. van Zinderen Bakker is at present undertaking a detailed study of the peats for pollens.

III. GEOLOGICAL DESCRIPTION OF THE SALDANHA FORMATION

A. Stratotype

The stratotype is 1,2m thick on Hoedjiespunt. It is preserved in a basin on a granitic platform; the floor of the basin is 5,25m above mean sea level. The lower 0,7m of the stratotype consists of horizontally-bedded microsphorite. The upper 0,5m is a layer of granite boulders and cobbles supported in microsphorite which has been reworked. Bedding is not readily apparent in the upper layer. The unit consists in part of phosphatised micrococquina. The upper part of the stratotype represents more turbulent conditions in a shallow environment, probably due to regression, while the lower part is a quieter water deposit. Figure 4.2 is a detailed section, and Figure 4.3 is a photograph of the type-section. Thin-sections show that there is less than 1 per cent quartz in the microsphorite. High P_2O_5 concentrations in whole rock analyses, 34 and 35 per cent, show that the microsphorite is nearly pure apatite. Although there is considerable coquina and many foraminiferal tests, none of the forminifera are identifiable. Gastropod moulds, Fissurella and Patella, occasional bivalve moulds and bryozoan remains testify further to the marine origin of the microsphorite.

The lower boundary at Hoedjiespunt is taken as the contact between the microsphorite and Cape Granite. The upper boundary is the sharp interface between the microsphorite and the overlying Pleistocene shelly limestone. In some places there is a suggestion of secondary phosphate mineralisation of the shelly limestone at the contact.

B. Reference Stratotype : Langeberg

Phosphatic sandstone is exposed in two sumps in the New Varswater Quarry. It is overlain unconformably by the Pliocene Varswater Formation (Tankard 1974a). Tilting of the phosphatic sandstone to the southwest took place in Late Miocene-Early Pliocene times, that is, before deposition of the Varswater Formation. Transgression of the Pliocene sea truncated the phosphatic sandstone at 30m above sea level. Underlying the sandstone is an unconsolidated silty-clay of unknown origin. The

Figure 4.2 Detailed section of the Saldanha Formation stratotype at Hoedjiespunt, Saldanha

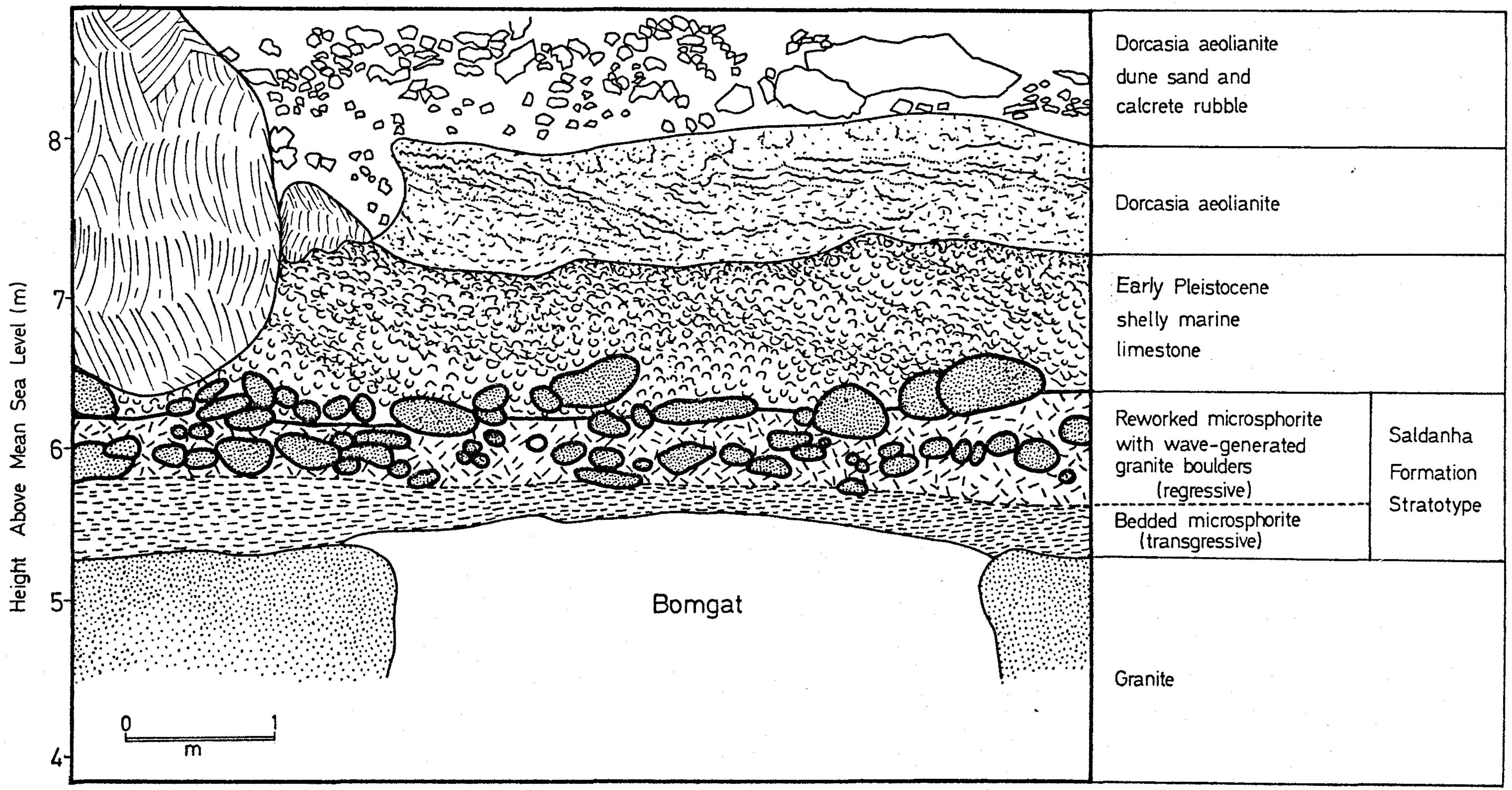


Figure 4.3 Photograph of type-section at the Bomgat, Hoedjiespunt.

Figure 4.4 Scour and fill structure of the phosphatic sandstone on Langeberg. Scour channel occupied by intraformational conglomerate.

phosphatic sandstone is 1 to 1,5m thick. The rock is brown, while the undulating surface is polished and pitted by differential erosion, and burrowed by marine animals. In places it is conglomeratic, containing clasts which are less phosphatic, or occasionally more phosphatic, than the host rock. This intraformational conglomerate material frequently fills in hollows in the phosphatic sandstone (Figure 4.4). Some occurrences show several erosion levels within the phosphatic sandstone, thus demonstrating repeated periods of phosphate mineralisation and erosion.

Thin-section analysis shows the phosphatic sandstone to be a fine-grained packstone with a matrix of microsphorite and finely-divided argillaceous and organic material. Heavy minerals, mainly ilmenite, are present in only trace amounts, while rare microsphorite pellets are found. In some places the phosphatic sandstone is coarse-grained and contains an abundance of shark teeth and bone fragments.

C. Other Occurrences of the Saldanha Formation

1. Ysterplaat

In 1973 foundation excavations by Paramount Construction Company revealed 75cm of low-grade Tertiary phosphatic sandstone. The exposure has since been back-filled. Underlying the sandstone are deeply weathered, steeply dipping Malmesbury shales. The surface of the Malmesbury Group is 8,6m above sea level. One hundred metres west of the excavation there is a large shallow quarry. The east wall of the quarry consists of rounded pebble-grade fluvial sediments, presumably of Pleistocene age but certainly younger than the phosphatic sandstone.

The phosphatic sandstone consists of a series of interdigitated lenses of fine, medium and coarse sands, which may be broadly subdivided into three units:

(i) The lowest 10 to 20cm consists of lithified medium to coarse sands with an abundance of marine mollusc shell moulds. The upper surface of this unit appears to be eroded.

(ii) The second unit consists of 28 to 30cm of iron-stained gravelly-sands which contain vertebrate bone remains; notably whale and penguin (Simpson 1973).

(iii) The topmost 20cm consists of medium to fine iron-stained sands with whale and penguin bone fragments and shark teeth.

The middle and upper units show an increasing influx of coarse detrital material. These sediments of the Saldanha Formation are overlain by 16cm of grey surface sand.

The median diameters of the sands range from 0,92 Ø to 3,03 Ø (10 samples); the populations are bimodally distributed. Eight of the ten samples analysed were positively skewed (range: 0,19 to 0,17), and the other two negatively skewed (-0,18; -0,41). Scanning electron microscope examination reveals surface textures typical of deposition in a high-energy beach environment. Figure 4.5 shows randomly oriented impact V-shaped patterns, and Figure 4.6 shows mechanical breakage patterns on a quartz grain. According to Krinsley and Takahashi (1962) and Krinsley and Margolis (1969) littoral textures are characterised by V-shaped indentations and conchoidal breakage patterns.

The fossil fauna is summarised in Table 4.1. Simpson (1973) named the penguin *Palaeospheniscus huxleyorum*. The shark teeth all belong to the Miocene-Pliocene species *Oxyrhyna hastalis*. The mollusc fauna includes *Glycymeris borgesi* which is known only from the Neogene. *Dosinia lupinus*, *Lutraria lutraria*, *Scissodesma spengleri* and *Cardium cf. edgari* are known from the Alexandria Formation. *Dosinia*, *Lutraria* and *Scissodesma* are still living.

That Miocene marine sediments are more widespread in this area is shown by the derived fossils occasionally found. In the South African Museum collection there is a single unrolled shark tooth of the species *Megaselachus megalodon* which was found in the original Ysterplaat quarry excavations. During storms the higher amplitude waves tear fossils from submerged deposits just offshore from Milnerton which are below normal wave erosion base (about 10m). These include *Megaselachus megalodon*, *Oxyrhyna hastalis*, and a Late Tertiary Gomphotherium (Q.B. Hendey, pers. comm.).

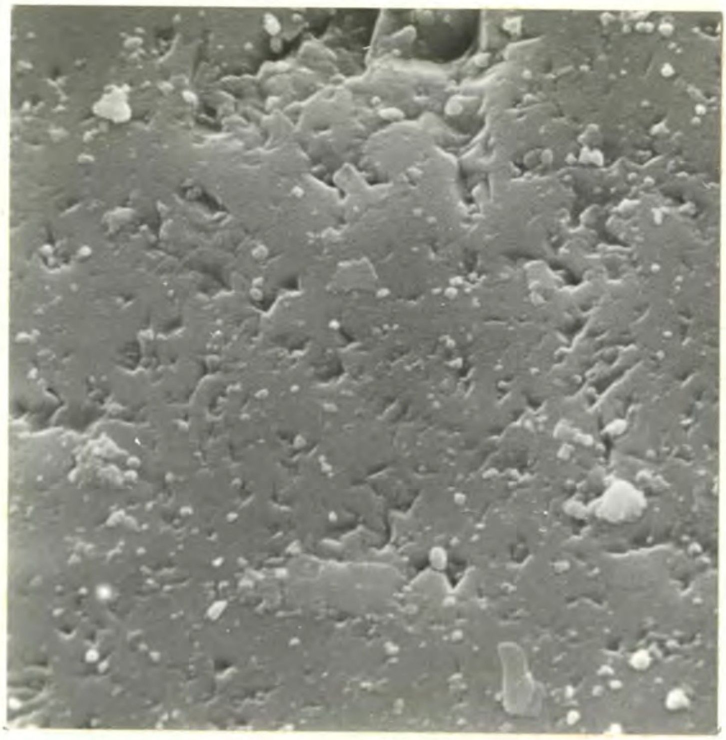


Figure 4.5. Scanning electron photograph of quartz grain surface from Ysterplaat showing randomly oriented V-shaped notches which characterise high-energy littoral environments (X 5000)

Figure 4.6 Scanning electron photomicrograph of quartz grain surface from Ysterplaat showing conchoidal breakage patterns (X 1000)

TABLE 4.1

List of Fossils in the Saldanha Formation

	Hoedjiespunt	Langeberg	Ysterplaat
FORAMINIFERA	X		
GASTROPODA			
Patella sp.	X		
Fissurella sp.	X		
Bullia sp.			X
Thais sp.			X
BIVALVIA			
Glycymeris borgesi			X
Cardium cf edgari			X
Scissodesma spengleri			X
Lutraria lutraria			X
Tellina sp.			X
Donax serra			X
"Tivela" sp.			X
Pitar sp.			X
Dosinia lupinus			X
BRYOZOA			X
ECHINOIDEA			X
CHONDRICHTHYES			X
Megaselachus megalodon		X	X
Oxyrhyna hastalis			X
AVES			
?Palaeospheniscus huxleyorum			X*
Palaeospheniscus sp.			X*

Footnote: * Simpson 1973

2. Hondeklipbaai

"Lower E stage" strata in this area consist of calcarenite, coquina, and phosphatic siltstone (A.J. Carrington, pers. comm.) Thin-section examination shows that the phosphatic siltstone originated by phosphatisation of a limestone. The quartz grains are well-rounded, while in the matrix there is evidence of finely bladed overgrowths developed syntaxially about intraclasts. There are occasional phosphate haloes about the intraclasts. Shell debris with crystalline overgrowths is common. Other occurrences of phosphatic siltstones are similar to the phosphatic sandstones seen at Langebaanweg, and are also conglomeratic in part.

In the Hondeklipbaai area the Saldanha Formation is preserved as bedrock-hollow infills, with a maximum thickness of 1m.

3. Off-shore

"Well-consolidated rocks are typically composed of intact and fragmented sand-sized microfossils (forming 40-65% of the rocks) and macrofossil fragments (1-10%) set in a collophane/micrite matrix. The ferruginous varieties are distinguished by their intimately mixed goethite, collophane, and micrite cements and the general absence of macrofossils. Silt-sized angular quartz and feldspar may be present in accessory (1-5%) amounts. Collophane is the apatite mineral identified in thin-section; XRD studies indicate that this mineral is francolite, a carbonate fluorapatite ..... The cementing material in the iron-poor varieties is an intimate mixture of collophane and micrite ....." (Parker & Siesser 1972).

IV. THE BREDASDORP FORMATION

In the quarry 4 km northwest of Saldanha a massive marine limestone, more than 3m thick, is overlain by high-angle cross-bedded aeolianites (Figure 4.7). The interface is at 56,4m above sea level. The marine

Figure 4.7 Bredasdorp Formation limestone exposed in quarry at Saldanha shows marine limestone overlain by high-angle cross-bedded aeolianite.

Figure 4.8 Thin-section of marine limestone shown in previous figure. It is a coquinite lime packstone. The components are quartz (Q), mollusc fragments (M), echinoderm fragments (E), and foraminifera (F). Crossed nicols. (X 24).

calcarenite is generally a carbonate packstone (i.e. the grains are arranged in a self-supporting framework). The elongate shell grains are frequently subparallel aligned, and in other places randomly aligned (Figure 4.8). The detrital quartz, shell grains, and intraclasts which are common are set in a sparry calcite matrix. The shell grains, which are re-crystallized to sparry calcite, often have a micrite or microspar border. Loose packing (grain contact about 1) is evident in some parts of the limestone. Common to the entire exposure are recrystallized foraminiferal tests, echinoid spines, and bryozoan remains, which are not uncommonly corroded and which form the nuclei of syntaxial overgrowths. Pellets are rare. The quartz grains, usually less than 20 per cent, are round to well-rounded, and sometimes corroded. Occasional patches of collophane are observed.

The interface between the aeolian and marine components at 56,4m a.s.l. may represent the present elevation of the shoreline associated with the Saldanha Formation phosphorites. Du Toit (1954) equated this calcarenite with the Alexandria Formation of the south coast.

There are two other occurrences of marine limestone in the environs of Saldanha Bay which are lithologically similar to the Saldanha Quarry limestone. In a cutting along the shore of Smitswinkel Bay there is a massive marine limestone. But in addition to detrital quartz there is also some oligoclase, derived from the granite substrate. In a shallow quarry at 9,5m above sea level just north of Langebaan this limestone underlies an Early Pleistocene beach deposit.

The petrography of this limestone at Saldanha has been described in great detail by Siesser (1970, 1972). He notes that intraclasts are common, and that they are bimodally sized. The larger intraclasts contain recognisable fragments of molluscs, echinoderms, foraminifera, ooliths, smaller intraclasts, glauconite, and terrigenous material. Ooliths are generally taken to represent precipitation of carbonate in a shallow, agitated environment (Siesser 1970). Tankard (in press a) has suggested that the marine limestone at Saldanha is a near shore deposit and that the interface between the marine and the aeolian components (Figure 4.7) represents the highest extent of the Miocene sea.

Figure 4.9 Bredasdorp limestone overlying Table Mountain sandstone at Die Kelders.

Figure 4.10 Conglomeratic zone at base of Bredasdorp Formation, Die Kelders.

Figure 4.11 Thin-section of Neogene limestone at Die Kelders, showing cavity filling drusy calcite which increases in crystal size away from the walls. Crossed nicols. (X 24).

Another outcrop of Bredasdorp limestone forms a low coastal cliff on a platform on Peninsula sandstone (Figure 4.9) at Die Kelders. The lower part of the limestone is a subangular, quartz cemented, breccioid horizon (Figure 4.10). The breccioid material is predominantly Cape Supergroup rock. But its subangular nature is not typically marine, and it would seem that the material could have originated by rapid transgression. Its origin is far from clear. Lenses of coquina occupy depressions. Above these units the limestone shows heavy mineral banding in the lower part, but is otherwise structureless. The total absence of high-angle cross-bedding would argue against a dune accumulation.

In thin-section the Bredasdorp limestone at Die Kelders may be described as a quartzose lime-wackestone, using the classification of Dunham (1962). The detrital quartz component is bimodal. The dominant mode is fine-grained (average about 2,3 Ø), and the subordinate mode very coarse-grained (average about -1 Ø). Generally the grains are subangular and corrosion is common. Skeletal material constitutes less than 5 per cent of the rock, and foraminifera are present but scarce. The matrix is microspar (recrystallised micrite: Folk 1965), while cavity-filling drusy calcite mosaic (see Bathurst 1958) is common (Figure 4.11). Generally, the drusy mosaic increases in crystal size away from the walls (Figure 4.11).

V. AGE OF THE SALDANHA FORMATION

The occurrence of microsphorite and phosphatic sandstone at Hoedjiespunt, Langebaanweg (Langeberg) and Ysterplaat respectively, are largely barren of identifiable fossil remains. At Ysterplaat the Saldanha Formation contains an abundant conservative endemic mollusc fauna which cannot be used for dating. The incorporated penguin bones are of greater value. There are several lines of evidence to suggest a Miocene age.

From one borehole, X14, on Langeberg, a very rolled shark tooth, the Miocene Megaselachus megalodon, came from the gravels at the base of the Varswater Formation. Its rolled nature suggests that it was derived from older deposits. Hendey has found (pers. comm.) a Hipparion primigenium tooth in the gravels at the base of the Varswater Formation which is rolled and obviously derived. It indicates a Miocene age. Two distinct species of penguin have been identified from the Ysterplaat occurrence (Simpson 1973). Simpson named one of these species Palaeospheniscus huxleyorum and notes that the nearest comparison is with Palaeospheniscus material from the Patagonia Formation of southern Argentina. The age of this formation is not precisely fixed, but it is of approximately Early Miocene age. Simpson suggests that the Ysterplaat beds must be pre-Pliocene. Shark teeth of a single species, Oxyrhyna hastalis, were also found in these sediments. This species is recorded from Miocene and Pliocene deposits in various parts of the world. The mollusc fauna is typically conservative endemic and the assemblage is very similar to that from the Miocene "Lower E stage" of the Namaqualand coast (A. J. Carrington, pers. comm.) and the fauna of the Alexandria Formation described by Newton (1913). Particularly interesting is Glycymeris borgesi (described by King 1953, as G. austroafricana) which is known only from Neogene deposits, and Cardium cf. edgari known from the Alexandria Formation. Other diagnostic fossils have been recorded from the Ysterplaat and Milnerton areas, but not in situ. In the South African Museum collection there is a single Megaselachus megalodon tooth which came from a quarry at the Ysterplaat airfield. The only quarry at the airfield is situated a mere 100m from the fossiliferous deposits already described. The fresh appearance of the tooth would suggest that it was in fact derived from these deposits. Furthermore, fresh specimens of this species are frequently thrown up onto the beaches at Milnerton during heavy seas, implying the presence of Miocene marine sediments just below normal wave base. Oxyrhyna hastalis and a Late Tertiary Gomphotherium are also thrown up onto the Milnerton beach during storms.

Other inferences concerning the age of the Saldanha Formation assume that phosphorite formation is dependent upon a unique set of environmental factors (Tankard 1974a). Phosphorite formation is dependent upon increasing temperature as cold phosphate-rich water upwells to the surface. Summerhayes (1970) has remarked on the limited evidence for contemporaneous formation of apatite on the sea floor, although Baturin (1971) does find some evidence. The widespread occurrence of bedded phosphorites in the Upper Cretaceous, Eocene and Miocene, suggest that those were the times of optimum conditions. The widespread deposition of phosphorite during the Late Tertiary is attributed to the warmer climate at that time. In fact all documented autochthonous Neogene phosphorites are of Miocene age:

Monterey Formation	California	Middle Miocene
Arcadia Formation	Florida	Middle Miocene
Pungo River Formation	North Carolina	Lower to Middle Miocene
Riecito Formation	Venezuela	Middle Miocene
Sechura desert deposits	Peru	Miocene

(Sheldon 1964; Trueman 1971; World Survey of Phosphate Deposits published by the British Sulphur Corporation Limited, London).

Phosphatic rocks on the continental shelf of the Cape Province have been assigned to the Upper Tertiary by Parker and Simpson (1972), while Dingle (1974) more recently has suggested two phases of formation of phosphorite - Late Eocene and Late Miocene/Early Pliocene. Haughton (1956) examined a phosphatised fauna dredged in the vicinity of the Cape Peninsula. On the basis of the nautiloid, Aturia lotzi (Böhm) which occurs also in Miocene outliers in the Bogelfels area, Haughton concluded that much of the phosphorite is of Lower Miocene age. On foraminiferal evidence (Sherbonina sp., Late Eocene to Miocene; Globigerinoides spp., Early Miocene to Recent; Nummulites heterostegina, Early Miocene) Rogers (1974) suggests an Early to Middle Miocene age for the overlying marine limestones, which would imply a minimum age for the phosphorite. Thus, all available evidence suggests that the Saldanha Formation is of Middle Miocene (Langhian) age.

VI. CORRELATION

Widespread Late Cretaceous, Eocene and Miocene marine transgressions are believed to have resulted from changes in elevation of mid-oceanic ridges due to changes in spreading rates (Russell 1968; Hallam 1971; Frerichs & Shive 1971; Flemming & Roberts 1973; Rona 1973). These transgressions have left a record in South Africa. More particularly Miocene sediments have been recorded around the coast from Mocambique to Namaqualand.

Du Toit (1954) mentions Miocene marine sediments in Mocambique. Soares and da Silva (1970) equate the marine Santaca Formation with the Uloa sequence of Zululand, having a similar fauna. King (1953) suggested that the Pecten bed at Uloa was of Lower Miocene age and that it was overlain by a Pliocene calcarenite. Frankel (1968) disputes that these two horizons are separated by an unconformity and suggests a Middle Miocene age for the entire sequence, an age now confirmed by Stapleton (pers. comm.). Ruddock (1968) and Bourdon and Magnier (1969) have also recorded Miocene sediments in the Algoa Bay area which are partly included in the Alexandria Formation. The Alexandria and Bredasdorp Formations appear to be partly of Miocene age. A.J. Carrington (pers. comm.) believes that authigenic phosphorites occurring along the Namaqualand coast, on foraminiferal evidence, are possibly of Late Miocene age. The mollusc fauna from Ysterplaat (Table 4.1) is very similar to that of the calcarenite which includes Glycymeris borgesi, Lutraria lutraria, Scissodesma spengleri, the same form of Donax serra, etc. Like the Hoedjiespunt microsphorite, the Namaqualand "Lower E stage" sediments are preserved in depressions in the bedrock floor. Lower to Middle Miocene phosphatic sandstones occur extensively on the continental shelf (Rogers 1974). (Although Cox (1939) described Glycymeris borgesi (Cox) as G.africana, he later renamed it borgesi (Cox 1946) since G. africana Cox was a secondary homonym for the South African Cretaceous G. africana (Griesbach). King (1953) independently recognised that G. africana Cox was a homonym for G. africana (Griesbach) and renamed the species G. austroafricana).

VII. ORIGIN OF THE SALDANHA FORMATION SEDIMENTS

The phosphorite deposits of the Saldanha Formation bear comparison with the Israel phosphorite deposits. The latter form part of an Upper Cretaceous to Eocene phosphogenic province stretching from Morocco through north Africa to the Middle East and Turkey (Trueman 1971). In Israel a deeper water facies is characterised by oolitic microsphorite while a shallower water facies is characterised by bone beds containing some pellets and cemented by a collophane-mud matrix. The deeper water facies resembles the Hoedjiespunt microsphorite, and the shallow water facies is comparable to the Langeberg and Ysterplaat occurrences.

Deposition of the Saldanha Formation resulted from a marine transgression in the Middle Miocene. This transgression complex is recorded at many places along the South African coast. The sediments are characterised by:

(i) a shallow water phosphatic rock facies at Langeberg, and

(ii) a quieter water microsphorite facies at Hoedjiespunt.

Tankard (1974a) has discussed the origin of the phosphorites, and has pointed out that bedded marine phosphorite of Late Tertiary age is found today in the warm climates between the 40th parallels in areas adjacent to divergent upwelling of nutrient-rich waters. The precipitation of the phosphate is dependent upon an increase in temperature as the upwelling water reaches the surface, and low rates of supply of terriginous detritus.

CHAPTER 5

THE VARSWATER FORMATION : DEPOSITIONAL ENVIRONMENTS

I. INTRODUCTION

The Varswater Formation has attracted interest over the last two decades because of economic quantities of phosphate, present as pelletal phosphorite, which is used as a fertiliser. Mining operations commenced in 1953. Secondly, since mammal fossils were first found in these deposits in 1958 (Singer & Hooijer 1958) subsequent research has shown the mammal fauna to be one of the richest occurrences of its type in Africa. Extensive mining and exploration programmes have provided the material and data upon which this study is based. At Langebaanweg only open-cast mining is carried out, while exploration has been performed using a Selby coring technique.

The aim of this chapter is to assemble the various types of data to present a general palaeoenvironmental study. In the next chapter detailed textural analyses will be presented. Distribution of the various lithologic units was presented in Chapter 3.

All altitudes are recorded as height above mean sea level. The tidal range on the open coast (St Helena Bay) is 1,8m. For the New Varswater Mine (Langeberg) collar elevations of the boreholes were usually available. In the few cases where this was not so, collar elevations have been interpolated from a Chemfos Limited surface contour map with a 1,5m contour interval. Borehole collar elevations for the other prospected areas were interpolated from 15m contour interval maps. The terrain is gently undulating.

II. PALAEOBATHYMETRY

Structure contour maps (Figure 5.1 and 5.2) have been drawn from borehole log data from over 120 boreholes on Langeberg and a further 95 boreholes on Witteklip-Sandheuwel-Langlaagte. For the construction of the structure contour maps either the lower 2 per cent P_2O_5 cutoff or

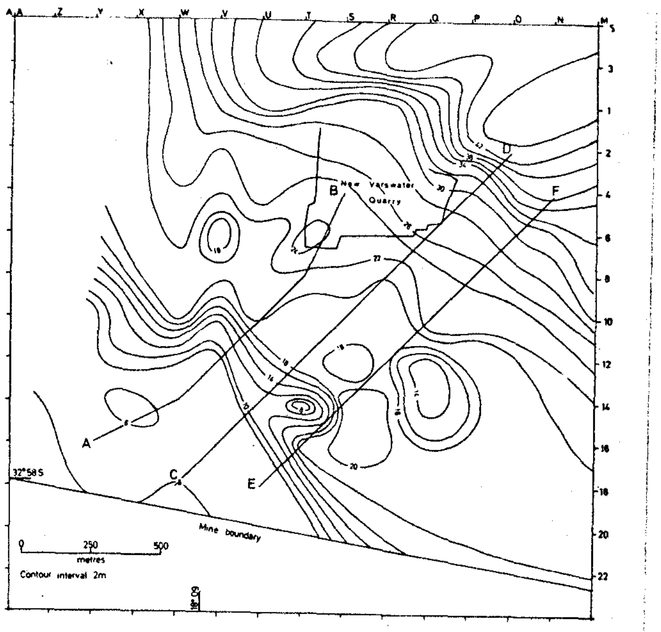


Figure 5.1 Structure contour map of the base of the Pelletal Phosphorite Member, Langeberg. (A-B, Fig. 2; C-D, Fig 5a; E-F, Fig. 5b).

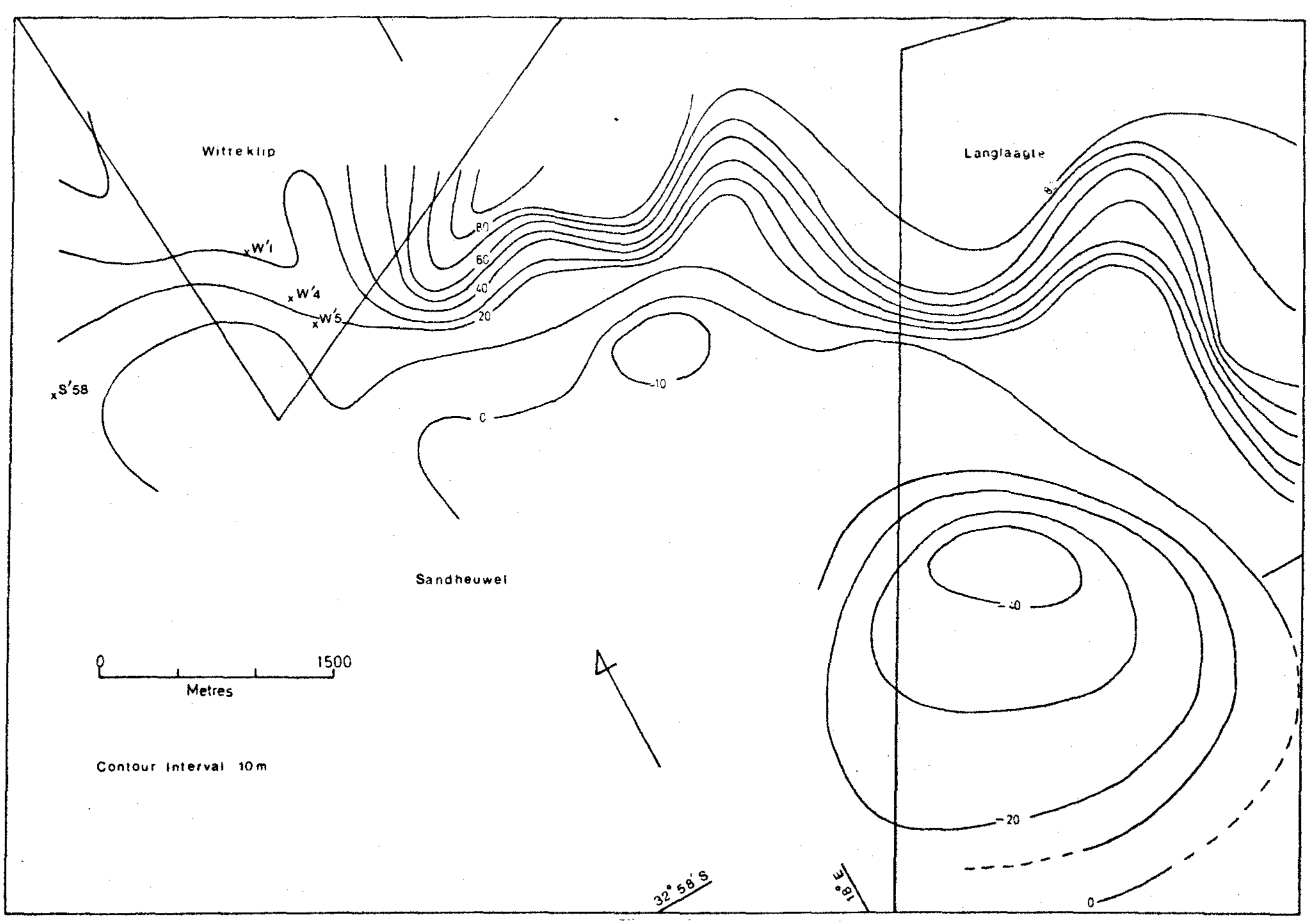


Figure 5.2 Structure contour map of the base of the Varswater Formation on Witteklip-Sandheuwel-Langlaagte.

bedrock intersections (either Saldanha Formation or granite) were used. On Langeberg this structure contour map represents the base of the Pelletal Phosphorite Member and is of importance in the discussion of the textural analyses of the sand. But in view of the fact that the underlying units of the Varswater Formation are generally thin, the structure contour map can be used as an approximation of the basal configuration of the formation. This is borne out in Figure 5.3 which illustrates the lithology and structure of the Varswater Formation on the farm Langeberg, drawn from borehole logs. (All boreholes referred to in this study may be located as grid-intersects in Figure 5.1, unless otherwise stated).

The dominant features on Langeberg (Figure 5.1) are two parallel northwest-southeast trending steps, with a slope of 3-5^{o}, separated by a broad platform of 1^{o} slope which dips to the southwest. The central platform, which shall henceforth be referred to simply as "the platform", rises gently from 18m a.s.l. in the southwest to 30m a.s.l. in the northeast. The step behind the platform, rising from 30m a.s.l., may represent a marine cut cliff or notch. Recent mining operations (August 1975) in the vicinity of T2 have revealed a 1m thick phosphatic sandstone which in general appearance and in thin-section relates most closely to the Miocene phosphatic sandstone. The area between the T2 exposure and the truncated quarry floor exposure would then represent incision by the Pliocene stillstand at 30m a.s.l. (It is hoped that eventually mining operations will produce a deep enough section beneath the T2 sandstone to verify this interpretation). Along the outer edge of the platform is a breaker-bar which has had a marked effect on pelletal phosphorite concentration. Maximum pelletal phosphorite concentrations (18-20 per cent P_2O_5) are found on the platform on the leeward side of the breaker-bar (Figure 5.4). An interesting feature of the second step is an inflection of the contours which suggests a valley profile (Figure 5.1). If this is a valley it is in keeping with Hendey's (1970a) view that a river had once flowed in that direction supplying the freshwater necessary to explain the concentration of vertebrate remains. The highest concentration of fossils lies at the entrance of this valley. Coarse fluviatile sands have been encountered 300-400m to the east.

The basal structure on the Witteklip-Sandgeuwel-Langlaagte prospect

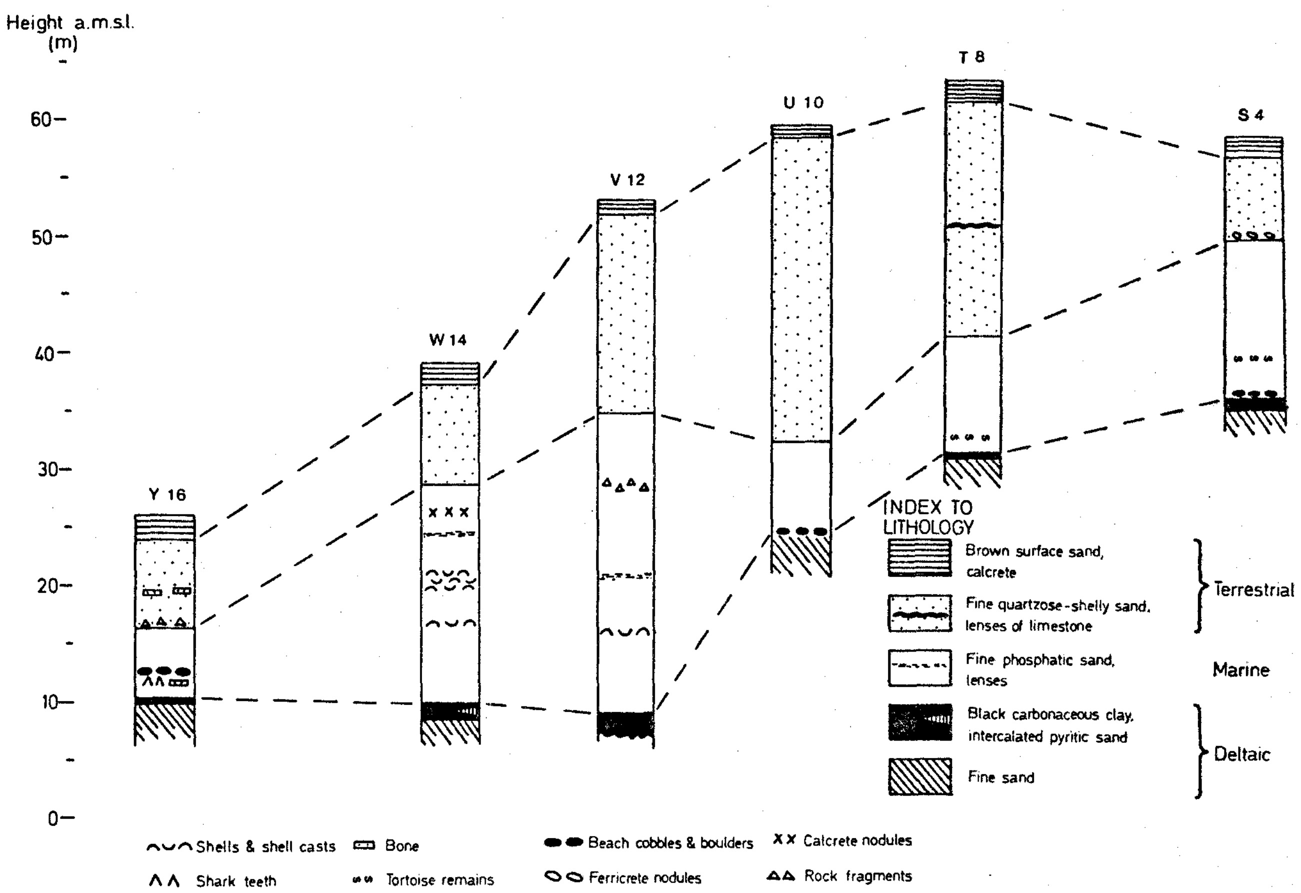


Figure 5.3 Cross-section of typical lithology on Langeberg measured along line A – B in Figure 5.1. The northern part includes part of the west wall of the quarry which does not contain all the sediment types described.

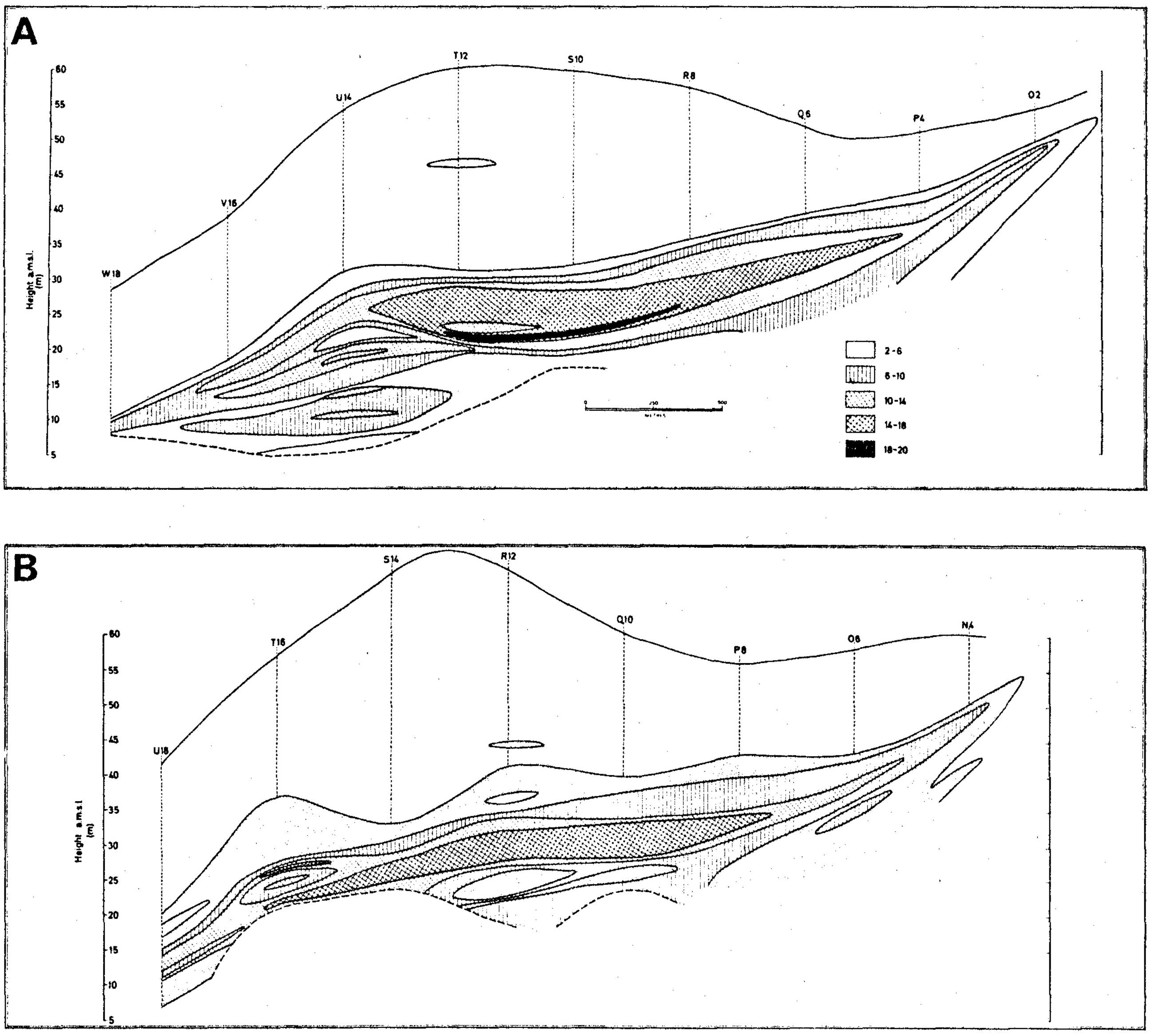


Figure 5.4 Vertical distribution of P_2O_5 (chemically measured at intervals of 1,07m in each borehole). 4A (top) is line C-D in Figure 5.1; 5b (bottom) is line E-F. Note breaker-bar at platform edge.

(Figure 5.2) is marked by a platform rising northwards to the foot of a granite escarpment. A major feature of this platform is a basin on Langlaagte which sinks to 40m below sea level. This basin has had no effect on phosphorite concentration. Iron staining and deep weathering of the granite floor are indicative of subaerial weathering.

Although no structure contour maps were drawn for the other prospects, an insight has been gained from borehole log interpretation. The bedrock on Waschklip is predominantly weathered granite and sericite schist. On the northern part of the farm bedrock lies generally at 20m a.s.l. The homestead lies on the western extremity of a WSW-ENE trending granite ridge. South of this ridge the granite floor forms a shelf at 10-20m a.s.l.

On Tiekosklip the bedrock is again predominantly granitic and the bedrock surface is in places 15m below sea level.

Forty-two boreholes sunk at Paternoster suggest a featureless granitic platform planed at 35-40m a.s.l.

III. DEPOSITIONAL ENVIRONMENTS

Figure 4.1 demonstrates the relationship of the Varswater Formation to the underlying Saldanha Formation. The relationships of the various units of the Varswater Formation to each other are illustrated in Figure 5.5. The Varswater and Saldanha Formations are separated by an angular unconformity. Within the New Varswater Quarry the phosphatic sandstone of the Saldanha Formation is truncated at 30m a.s.l., and the surface of the formation now slopes to the southwest. Erosion of the phosphatic sandstone has produced a fluted and polished surface. A patina could also imply subaerial weathering before deposition of the Varswater Formation. Tankard (in press a) has suggested that the Miocene Saldanha Formation can be correlated with a world-wide eustatic transgression at that stage. But since that Miocene sea level peak the sea has been generally regressive. It was suggested that deposition of the Varswater Formation took place during a secondary transgression, within a eustatically falling sea, induced by tilting.

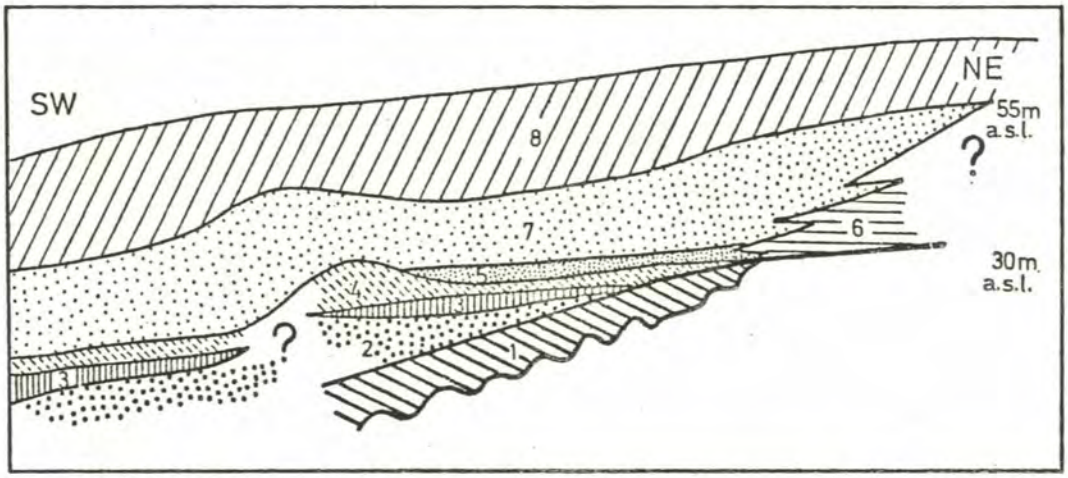


Figure 5.5 Idealised cross-section through the Neogene sequence on Langeberg. 1 = phosphatic sandstone (Saldanha Formation); 2 = quartzose sands; 3 = kaolinitic clay and peat; 4 = beach gravels and sands; 5 = estuarine sediments with peats; 6 = fluvial sediments; 7 = pelletal phosphoritic sands: 8 = Pleistocene dune sands.

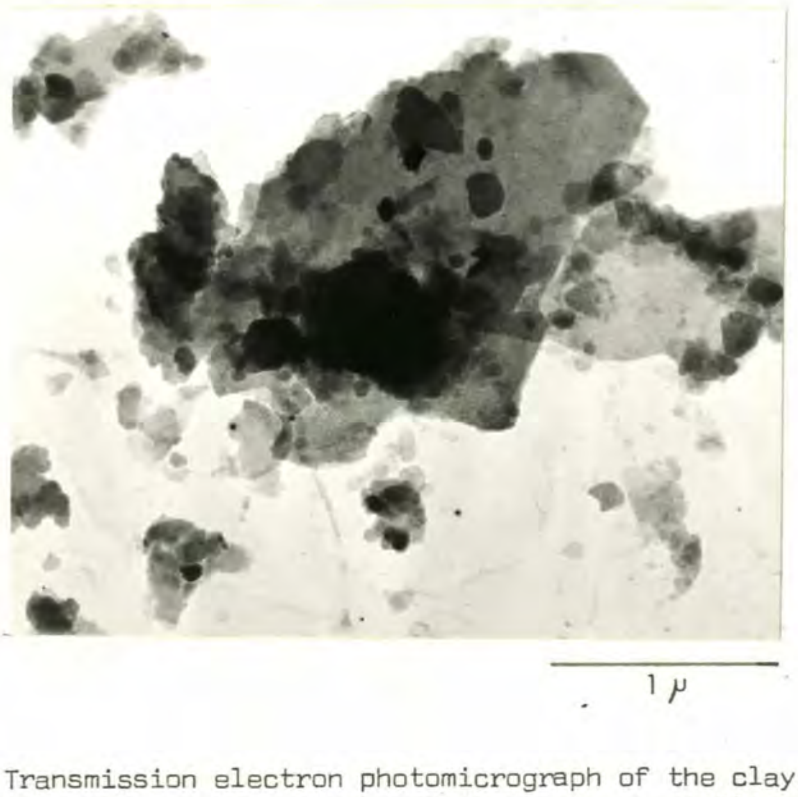


Figure 5.6 Transmission electron photomicrograph of the clay, suggesting that the mineral is poorly crystallised kaolinite.

The Varswater cyclothem is a transgressive complex. Discharge of a river into this transgressive sea in the vicinity of Langebaanweg created a series of transitional environments. Environments identified include: high energy marine, estuarine, fluvial and (?) terrestrial, together with the inevitable myriad microenvironments.

A. Kaolinitic Clay

Underlying the black kaolinitic clay on Langeberg is a little known quartzose sand which could, presumably, be of Late Miocene or Early Pliocene age. Pollen analyses have shown that the clay contains a poor pollen assemblage, but *Podocarpus* and grass pollens were recorded (Prof. E.M. van Zinderen Bakker, in litt.). It is suggested that the kaolinitic clay formed in a deltaic marsh environment, because it is up to 6m thick, it overlies fine quartzose sands, is associated with fluvial sediments, and it has a slope of 1:360 (equivalent to that of the lower reaches of many South African rivers). The kaolinitic nature of the clay, determined by transmission electron microscopy as poorly crystallised kaolinite (Figure 5.6), and the grass pollens are suggestive of freshwater conditions, while pyritic inclusions suggest an anaerobic environment. The kaolinitic clay is frequently overlain by consolidated to semi-consolidated quartzose sands with moulds of marine molluscs, demonstrating the transgression of the sea across the "delta". As the sea transgressed, so the marsh environment moved ahead of it until it abutted on the first step (Figure 5.3). Boreholes intersected the clay at -15m on Tiekosklip, while a brown-black carbonaceous clay is encountered in the basin in the granite platform close to sea level on Langlaagte

B. Gravel Member

Following submergence of the clay horizon, the marine sands and gravel of the Gravel Member were deposited. The sands are rich in shell casts of marine molluscs of warm water affinity. Within the New Varswater Quarry the Gravel Member rests directly on the Miocene phosphatic sandstone at 30m a.s.l. Here truncation and re-working of the phosphatic sandstone near 30m a.s.l. has produced a gravelly-sand of rounded quartz grains and phosphatic sandstone fragments with rolled and unrolled shark teeth of fifteen species (Hendey pers. comm.) and shell fragments. Well-rounded wave generated discoidal cobbles of phosphatic sandstone lie directly upon the Saldanha Formation floor (Figure 5.7). Boreholes show the gravel horizon with shark teeth to be widespread along the outer edge of the platform. Associated with the gravel and beachcobble horizon, or lying directly on the phosphatic sandstone floor within the quarry is a thin (0,5-2m) semi-consolidated silty-sand with shell moulds. In places it is phosphate mineralised. In borehole W14 at 13m a.s.l. a sandstone with many shell moulds, especially Pitar sp., and including mytilids and barnacles, overlies the clay deposits. In borehole U14 at 15m a.s.l. this horizon contains Patella sp. and a mytilid. Within the New Varswater Quarry the shelly sands occur at 30m a.s.l., i.e. the same elevation as the truncation of the Saldanha Formation. Most of the molluscs from the quarry area have been described by Kensley (1972). They are mostly rock-dwelling forms, except Donax serra and Bullia sp., which are sand-dwellers. Furthermore, the intertidal to infratidal nature of the fauna indicates a littoral environment, i.e. temporary stillstand of the sea at that altitude.

On Sandheuwel a coarse shelly sand with complete oyster valves and an ostracod and foraminifera fauna is encountered at 24m a.s.l. (borehole S'58), while at Duyker Eiland this horison crops out at 30m a.s.l. The writer has identified the oysters Ostrea atherstonei and Striostrea margaritacea from Sandheuwel and Striostrea cf. margaritacea from Duyker Eiland.

A stillstand of the sea at 30m a.s.l. appears to have been a prominent event in the history of the Varswater Formation, and the Mollusca summarised in Table 5.1, suggest water temperatures about 5°C warmer than today at the time of deposition. Cellana capensis is found today north of East London, while Turbo sarmaticus occurs eastward from False Bay, as do Barbatia obliquata, Ostrea atherstonei and Striostrea margaritacea.

C. Quartzose Sand Member

Sediments of the Quartzose Sand Member are indicative of two facies: an

Figure 5.7 Wave-generated discoidal cobbles in the Gravel Member exposed in the floor of the New Varswater Quarry at 30m a.s.l. The cobbles are lying on Miocene phosphatic sandstone.

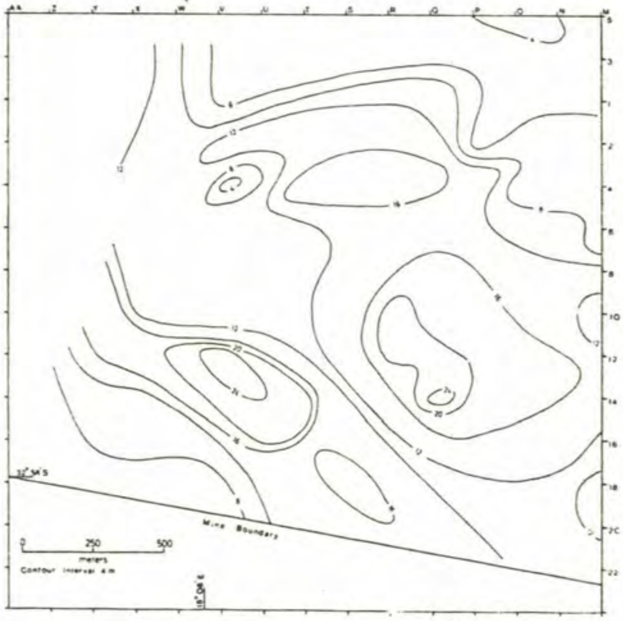


Figure 5.8 Isopach map of the Pelletal Phosphorite Member, Langeberg.

TABLE 5.1

List of Invertebrate Fossils in the Varswater Formation.

	Langeberg	Waschklip	Sandheuwel	Duyker Eiland	Ref
FORAMINIFERA					
Lagena hexagona			X		3
Lagena sp.			X		3
Ammonia beccarii	X	X	X		2,3
Elphidium advenum	X	X	X		2,3
PORIFERA				X	3
GASTROPODA					
Cellana capensis	X				1
Cellana sp.	X				1
Patella granularis	X				1
Patella sp.				X	3
Haliotis saldanhae	X				1
Haliotis sp.	X				1
Diodora parviforata	X				1
Oxystele trigina	X				1
Turbo sarmaticus	X				1
Tricolia neritina	X				1
Bullia sp.	X				1
Littorina cf. *saxitilis*	X				3
Ocenebra scrobiculata	X				1
Thais dubia	X				1
"*Clavatula*" sp.	X				1
"*Crassispira*"sp.	X				1
"*Turris*" sp	X				1
?*Pyramidella* sp.	X				1
Turbonilla kraussi	X				1
Siphonaria sp.	X				1
BIVALVIA					
Barbatia obliquata	X				
? *Perna* sp.	X				
Ostrea atherstonei			X		
Striostrea cf. *margaritacea*			X	X	
Pitar sp.	X				
Donax serra	X				
Donax sp.	X				
OSTRACODA					
Caudites sp.			X		
Bradleya sp.			X		
CIRRIPEDIA					
Barnacle fragments	X				
ECHINOIDEA					
Parechinus angulosus	X				

(1 = Kensley 1972; 2 = Visser & Schoch 1973; 3= Tankard 1974, this st

estuarine facies and a fluviatile facies. Hendey (1970a, 1973) to explain the rich mammal deposits in the floor of the New Varswater Quarry, postulated that a river flowing into the sea at that point created freshwater estuarine conditions. Most of the mammal remains have come from a horizon 2,3m thick lying directly upon the Gravel Member. The top of these estuarine sediments is capped by a thin phosphatic sandstone band. Whereas bones in the gravelly-sands of the Gravel Member are usually rolled, the bones from the estuarine deposits are not. The skeletons are never articulated, although jaws with teeth still intact are numerous. That no preferred orientation is shown by any of these elements rules out dispersion by any strong unidirectional currents. Boreholes have proved bone to be widespread on Langeberg, but it would appear that the really fossil-rich sediments are concentrated within the quarry area.

The faunal associations suggest that Hendey's Bed 2 is an estuarine facies. The occurrence of seal and penguin remains show the close proximity to a shoreline, while catfish and otter remains are probably indicative of estuarine conditions. The mammal fauna contains animals from various environments brought together by the common need for water. Hendey (1973) visualises an environment of riverine woodland flanked by grasslands.

Hendey (1974, tables 5,88) has summarised the mammal fauna from these sediments, which includes marine, freshwater and terrestrial species, the latter of which predominates. Over 70 mammalian species are recorded. "Small mammals are common and include insectivores, rodents, a lagomorph and small viverrids. Medium-sized herbivores are Hipparion, Nyanzachoerus and a variety of bovids. Large herbivores are Mammuthus subplanifrons (Maglio & Hendey 1970), a gomphothere, Ceratotherium praecox (Hooijer 1972), a sivathere and Giraffa. The larger carnivores include a machairodont, a viverrid and hyaenids, which are described elsewhere in this report. Also represented is a seal ....." Non-mammalian vertebrates include a land tortoise Chersina (the most common species in these deposits), sharks, bony fish, snakes, lizards, frogs and birds.

The presence of a breaker-bar along the outer edge of the platform has already been mentioned. Cross-sections through the breaker-bar (Figure 5,4) were drawn for the Pelletal Phosphorite Member. But the configuration of the platform and the presence of estuarine sediments with a close association of terrestrial and marine animal remains, suggest that a breaker-bar and barrier-beach complex had already developed at the time of deposition of the Quartzose Sand Member. River-flow in the north and northeast produced estuarine conditions in the lee of the bar. But periodic breaching of the bar by the sea mixed the terrestrial and marine animal remains.

Recent mining operations (1974-75) have extended the quarry face in a southerly direction and have shown that the clean estuarine quartzose sands grade laterally into a carbonaceous sand and peat deposit. This carbonaceous deposit is 0,5-1,0m thick and includes lenses of sand accretion. Vertebrate fossils from the carbonaceous sands are typical of the clean quartzose sands already discussed, and the land tortoise is still the most common; the bones are usually fragmentary. But of particular significance are the still articulated distal extremities of a sivathere (giraffe) standing vertically (Hendey, pers. comm). It seems likely that the giraffe was trapped in the marsh while drinking. It has also been found that the fossils include waterbirds, and these are absent from the neighbouring quartzose sands. Bird egg-shell has also been found. Molluscs from the silty sand are well preserved and include marine and freshwater species.

The carbonaceous sands and peat also contain a rich pollen spectrum (Prof. E.M. van Zinderen Bakker, pers. comm) which includes the following:

Dominant unidentified type	approx. 92 per cent
Gramineae (grass)	approx. 3 per cent
Restionaceae (reed)	approx. 1,5 per cent
<u>Cliffortia</u> (bush)	approx. 1,5 per cent
<u>Podocarpus</u> (tree)	< 1 per cent
<u>Myrica</u> (dry area shrub)	< 1 per cent
Chenopodiaceae-type (salt flat bush)	< 1 per cent
Compositae (flower)	< 1 per cent
Other unidentified types	approx. 2 per cent

Professor Bakker notes that the morphology of the dominant unidentified pollen type is comparable to that of Cuscuta (a parasite) and Elephantorrhiza (a dry area plant) and also certain of the Aizoaceae (dry area plant).

Hendey (1973, 1974) has postulated that the mammal remains found in the estuarine sediments suggest that a river once flowed there. In boreholes M6-15-N6-15 and N4-76-04 there are layers of coarse to medium sands intercalated with fine sands. These sands, unfortunately, contain no faunal element. That they are of fluviatile origin is suggested by their coarse nature, and by scanning electron microscopic examination which shows a texture similar to that of river sands illustrated by Krinsley and Donahue (1968; figure 1). The grains show some chemical etching, but also relatively small subaqueous V-patterns suggesting vigorous conditions.

D. Pelletal Phosphorite Member

A major stillstand of the sea which has been recorded on Langeberg, Sandheuwel and Duyker Eiland, resulted in accumulation of littoral marine and estuarine sediments. This period of deposition was followed by a final transgression which carried the shoreline to 50-55m a.s.l. and resulted in accumulation of the Pelletal Phosphorite Member. The transgression displays a typical onlap aspect. The Pelletal Phosphorite Member is composed essentially of fine-grained quartzose sands with varying amounts of pelletal phosphorite. Chemfos Limited is currently mining the pelletal phosphorite from this member.

The isopach map (Figure 5.8) of this Pelletal Phosphorite Member shows that the thickest parts are either banked against the lower step where it reaches a thickness of 24m, or over a depression in the platform where a thickness of 25m is attained. To the northeast the horizon pinches out at 53-54m a.s.l. The general trend of the body parallels that of the basal structure (Figure 5.1). In Figure 5.9 the highest P_2O_5 values for each borehole have been mapped. It is immediately apparent (compare with Figure 5.1) that the highest concentrations of phosphorite are found on the platform, particularly in the depressions.

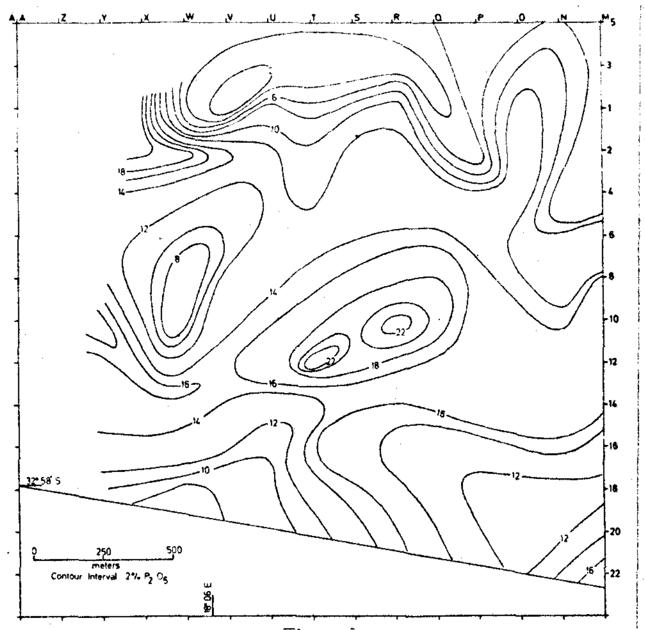


Figure 5.9 Horizontal distribution of highest P_2O_5 values, Langeberg.

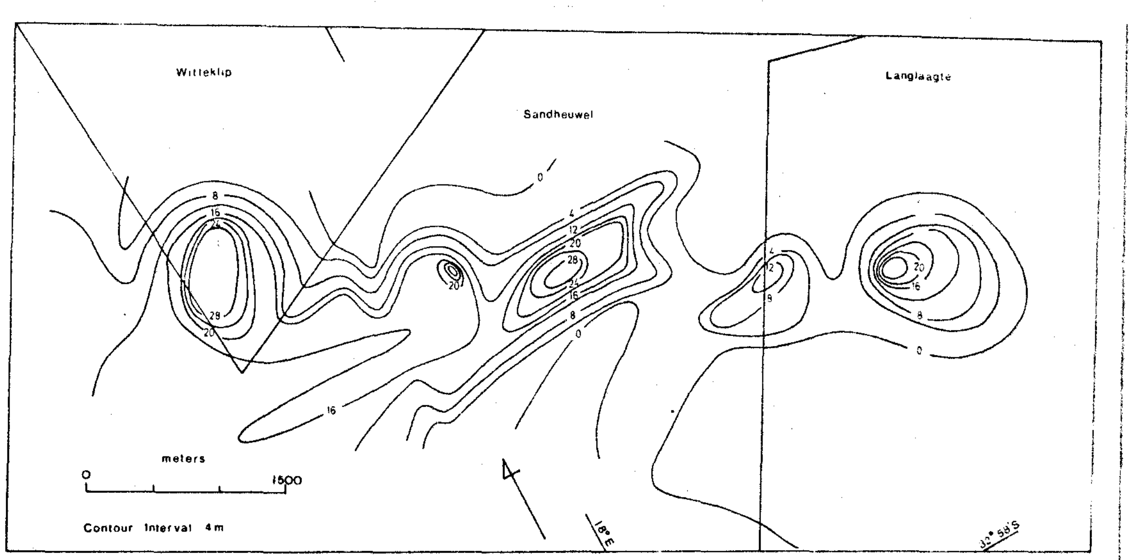


Figure 5.10 Isopach map of the Pelletal Phosphorite Member, Witteklip-Sandheuwel-Langlaagte.

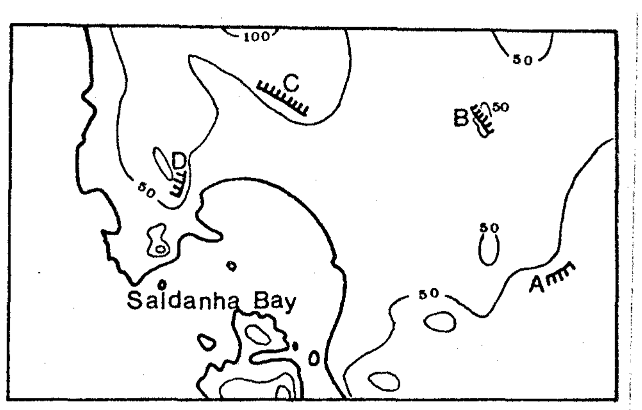


Figure 5.11 Distribution of shorelines in the Langebaanweg basin. A,B,C are Pliocene shorelines and represent the peak of the transgression (50–55m a.s.l.). D is a Miocene shoreline.

Two cross-sections (Figure 5.4) illustrate this relationship clearly. A very important feature of these cross-sections is that at 20-25m a.s.l. a breaker-bar runs along the outer edge of the platform following the general strike of the basal structure. The pelletal phosphorite is a shallow water detrital deposit with the maximum phosphorite concentrations on the shallow platform between the breaker-bar and the beach zone.

Shark teeth, mollusc shells and shell casts occur throughout the phosphorite horizon, and have been concentrated by brief erosional episodes. Also included in this sediment are phosphatised foraminiferal fragments (mainly Elphidium sp.), echinoid spines, minute fish teeth and coprolites. Occasional tortoise carapace fragments testify to the nearness of the shoreline. In the New Varswater Quarry the sediments are largely structureless, although several observed cross-bed sets have north and northeasterly azimuths in the direction of the shoreline. This is typical of a shallow marine transgressive model. Burrow structures have been observed at the top of this member, and mark a sediment water interface.

On Waschklip the Pelletal Phosphorite Member is encountered at 7 to 10m a.s.l., where large quantities of shell fragments, shark teeth, phosphatic pebbles, and rounded beach gravel have been cored. The configuration of this member in the Witteklip-Sandheuwel area is controlled by the granite bedrock contours (Figure 5.10). On Tiekosklip the succession commences with a gravelly sand at 15m below sea level, and is followed by an upward-fining sequence. Pelletal phosphorite concentrations are low at Paternoster and Duyker Eiland. Sediments from Duyker Eiland include some very well sorted, well rounded fine sands typical of the beach zone with an abundance of tetraxon sponge spicules.

Following the temporary stillstand at 30m, final transgression carried the sea to its furthest inland extent at 50-55m. Profiling of P_2O_5 values on boreholes shows a pinching-out of the Pelletal Phosphorite Member on Langeberg at 53-54m a.s.l., on Witteklip and Paternoster 54m a.s.l., and at Duyker Eiland 47-50m a.s.l. On the farm Groot Springfontein, east of Waschklip, a water borehole struck a marine horizon at 50m a.s.l. Nodular phosphate, phosphatised bone, shark teech and mollusc shells were recovered from this borehole. According to the owner of the farm the shells included the "common white fishing mussel", which in the southwestern Cape is Donax serra and is also known to occur in the Varswater Formation (Table 5.1). Donax serra today lives in the intertidal zone, and confirms that a hypothetical Varswater Formation shoreline at 50m a.s.l. is a good approximation. The distribution of this shoreline around the Saldanha-Langebaanweg erosional basin is shown in Figure 5.11). It is unlikely that this shoreline represents the true height of sea level in the Pliocene since the west coast has been epeirogenically unstable since the Miocene and through the Early and Middle Pleistocene (to be discussed in detail later).

IV. AGE OF THE VARSWATER FORMATION

The rich fossil vertebrate fauna from the Varswater Formation includes no extant species of larger mammals (Hendey 1974). Hendey has defined a Langebaanian mammal age based on the most characteristic fossils. These include: Prionodelphis, Agriotherium, Percrocuta, Machairodus, Enhydriodon, Mammathus subplanifrons, Nyanzachoerus, and Ceratotherium praecox. Comparison of the larger mammals with the radiometrically dated east African faunas indicates a Pliocene age (Hendey 1970b, 1974; Hooijer 1972). An apparent age of 4 million years is inferred.

Dr S.P. Applegate of Los Angeles has examined a large assemblage of shark teeth from the area for the writer. Odontaspis acutissima, apparently the first record of this species from South Africa, was found in the Waschklip and Langeberg deposits. Also recorded from Langeberg are Megaselachus sulcidens and Carcharhinus cf. melanopterus, and a myliobatid spine. O. acutissima ranges in age from Oligocene to Pliocene. C.cf. melanopterus is a Recent species which appears to have no published fossil record. Applegate concludes that the assemblage is probably of Late Tertiary age. Visser and Schoch (1973) suggest on foraminiferal evidence (Elphidium advenum and Ammonia beccarii) that these deposits are of Pleistocene to sub-Recent age, but the evidence compiled by Hendey, Hooijer and Applegate is more substantial. Wolff et al (1973) also, and independently, suggest a Pliocene age for the Varswater cyclothem.

At the base of the calcareous dune sands, and immediately overlying the Varswater Formation on Langeberg, are teeth of Hipparion namaquense (Hendey, pers. comm.) which demonstrates the Pleistocene age of the dune sands, and suggests that the upper part of the Varswater Formation could be of Late Pliocene age. A derived Miocene Hipparion primigenium is found at the base of the Varswater Formation, while the Varswater Formation itself has a heterogeneous hipparion distinct from these other two forms.

CHAPTER 6

SIZE ANALYSIS AND HYDRAULIC EQUIVALENTS : VARSWATER FORMATION

I. INTRODUCTION

The lack of exposure precludes the use of primary sedimentary structures in the stratigraphic analysis. Apart from a very limited exposure in the quarry, stratigraphic reconstruction is based largely on textural analyses of borehole samples. This chapter reports the results of textural analyses of the sand fraction ($> 63\mu$) of 153 samples from 54 boreholes, 52 of these on Langeberg and the other two (W'1 and W'5) on Witteklip. In addition analyses were also performed on 18 pelletal phosphorite concentrates and the quartz fractions of these sediments.

The use of grain-size analysis in geology has been common practice for some time, but only in recent years has there been any concerted attempt to relate size distribution characteristics to the depositional processes and environment (Folk & Ward 1957; Mason & Folk 1958; Friedman 1961; Duane 1964; Visher 1969). These authors have found the sorting (standard deviation), skewness and kurtosis parameters to be the most diagnostic of environment. On the basis of skewness it was found on Mustang Island, Texas, that beach sands were negatively skewed and dune sands positively skewed (Mason & Folk 1958). One of the problems of size distribution interpretation is that the same processes occurring within a number of environments result in similar textural responses (Visher 1969). It is the aim of this chapter to describe the textural characteristics of the sediments and to show how the characteristics of the Pelletal Phosphorite Member sands on Langeberg are related to the environment of deposition. It will also be shown that deviations from hydraulic equivalence between the pelletal phosphorite and the quartz fractions are explained by the depositional environment.

II. METHODS

All sediment samples were obtained from the Chemfos Limited exploration programme. A Selby coring technique was used, driving a 1,07m long cylinder into the sediment. Sediment samples were thus collected vertically every 1,07m, and sample depths reported are class-marks of 1,07m class-intervals.

Grain-size measurements were made on the sand fraction ($> 63 \mu$) by conventional dry-sieving methods using half-phi intervals. The formulae of Folk and Ward (1957) were used because the samples include the extremes of distribution, a region not covered by the more conservative measures (e.g. Inman 1952). A pipette method was used to measure the clay-silt sizes where the sand fraction was less than 95 per cent of the whole sample. The statistics were determined from graphs on arithmetic probability graph paper. (A more comprehensive discussion is presented in the appendix).

Pelletal Phosphorite concentrates were obtained by heavy liquid (bromoform) separation from the detrital quartz.

Rock colour codes refer to the Rock Colour Chart (Rock Color Chart Committee, 1951).

III. KAOLINITIC CLAY

Determination of the clay group was done from X-ray diffraction data. But this same method is not sufficient for identification of the specific kaolinitic mineral. The basal reflections at 7,10 (001) and 3,57 (002) are diagnostic of the kaolinite group. Comparison of a transmission electron microphotograph of this clay (Figure 5.6) with an illustration in Grim (1968, figure 6-4) suggests that the clay mineral is poorly crystallised kaolinite. In one T.E.M. photograph there is also a trace of a prolate clay form, possibly halloysite. The clay horizons contain layers of pyrite sand and gravel, suggesting a fluctuating environment.

IV. GRAVEL MEMBER

After tilting the Miocene phosphatic sandstone (the basal bed) was truncated during a Pliocene transgression. Reworking of the phosphatic sandstone has resulted in accumulation of well-rounded pebbles and cobbles (Figure 5.7), which are frequently associated with a gravelly-sand, or in other places with semi-consolidated to consolidated fine to very fine quartzose sand with mollusc moulds. Texturally the sand is moderately sorted, negatively skewed with a mean size 3,10 Ø. The sands are of littoral origin, nearness to the shoreline being demonstrated by the close association of land and marine animals, by the shallow water nature of the molluscs (Kensley 1972) and by the fact that this horizon is directly overlain by estuarine sediments.

On Sandheuwel the mean grain-size is 2,54 Ø. The sediment is polymodal, the coarser fraction being composed of two populations, subangular and rounded grains. Subrounded ilmenite grains are present in the silt fraction. The coquina is typically poorly sorted.

V. QUARTZOSE SAND MEMBER

A. Estuarine Facies

Within the New Varswater Quarry estuarine sediments are present as a thin layer with abundant terrestrial vertebrate remains. The sediment is almost free of phosphate and has a mottled yellowish brown (10 YR 5/4) to very pale orange (10YR 8/2) appearance. It consists of sandy silt and silty sand, mean grain size 4,0 - 4,4 Ø, with 35-43 per cent mud.

B. Fluvial Facies

In borehole M6-15-N6-15 (i.e. 15m from M6 towards N6, then 15m south) in particular there are layers of coarse to medium sands (M_z = 0,94, 1,60 Ø) intercalated with fine sands, all of these being moderately sorted quartzose sands. These sands are bimodally or polymodally distributed. Generally the sediments of N4-76-04 are a moderately sorted fine sand with the mean size of the coarsest horizons being 1,63 Ø. The sand is essentially quartzose and iron stained.

Scanning electron microscopic examination of the surfaces revealed textures identical to those of river sands illustrated by Krinsley and Donahue (1968, figure 1). Small subaqueous V-patterns suggest vigorous conditions. Furthermore, the sands are coarse compared to the average marine littoral sands of the rest of the formation.

VI. PELLETAL PHOSPHORITE MEMBER

The largest number of size analyses was performed on the sediments of the Pelletal Phosphorite Member since this is the most extensive member of the Varswater Formation. Being of economic importance samples from this member are the most readily available. Each of several boreholes was sampled at regular intervals and grain-size distribution computed (Figure 6.10). Those samples chosen to give an areal distribution of the parameters were selected from the level of the highest P_2O_5 concentrations.

The choice of the datum used in the sample selection may appear to be somewhat arbitrary, but there was no other datum available. It may be argued that the highest P_2O_5 values arise from concentration of the pelletal phosphorite due either to low accretion rates of detrital quartz, or erosion and winnowing of the lighter quartz fraction (phosphorite is denser than quartz). If winnowing of the quartz fraction had taken place then the highest P_2O_5 values should always be associated with the coarsest parts of the succession. Figure 6.10 shows that this is in fact not the case. Furthermore, the datum could be diachronous and discontinuous and may represent a set of horizons unrelated in time and space. Figure 5.4 shows that by choosing the sediment sample with the highest P_2O_5 value for any single borehole a nearly continuous datum has been selected, particularly on the platform between the breaker-bar and the beach (i.e. from T12 to O 2 in Figure 5.4A). The writer realises that of the 52 boreholes sampled on Langeberg it is highly likely that in places the datum may be diachronous. But in selection of samples the sediment above and below each sample had been qualitatively examined to assure that the sample is representative of the pelletal phosphorite horizon. A further lack of precise control is reflected in the coring technique, where sediment mixing is known to take place as the 1m long cylinder is driven into the sediment. These limitations should be borne in mind when reading the following discussion. The distribution maps of the various statistics show definite trends which are supported by the palaeobathymetry and the discussion of hydraulic equivalents.

The Pelletal Phosphorite Member is transgressive across the other units already described. That it is transgressive is demonstrated by the fact that it overlies freshwater estuarine and fluviatile sediments, while at its base it has also incorporated some of the smaller vertebrate elements from the estuarine facies, together with lumps of clay torn from that same horizon. In boreholes S12 and Q12 there is a gradual coarsening-upward sequence (Figure 6.10). The highest grades of pelletal phosphorite are concentrated on the platform (Figure 5.4). These high grades reflect the close proximity and exposure of the basal bed which was the source of the pelletal material.

The sediments are generally moderately sorted phosphatic-quartzose sands. Almost all the samples are polymodal, consisting usually of sand and silt with very occasional gravel or clay. The main constituent of the sediment is fine sand, the medium and coarse sand fractions forming only a very small part. Particles larger than 2 Ø are scarce. Heavy minerals are concentrated in the fractions smaller than 3,5 Ø and more particularly smaller than 4 Ø. Ilmenite constitutes about 95-98 per cent of the heavy mineral population, garnet making up the balance. Zircon is rare. The ilmenite and garnet are subangular to subrounded, while the zircons occur as pale pink or grey to colourless euhedral crystals or occasionally as rounded grains. On Langeberg the highest concentration of heavy minerals (Q12) was 0,7 per cent, while on Witteklip the highest concentration (W'5) was 1,0 per cent.

The areal distribution of the phi mean, inclusive standard deviation (dispersion), skewness and kurtosis using the graphic formulae of Folk and Ward (1957) has been plotted (Figures 6.1 - 4). Similar distribution maps have been drawn for the pelletal phosphorite fractions of eighteen samples (Figures 6.5 - 7).

Comparing the phi mean distribution (Figure 6.1) with the structure contour map (Figure 5.1) reveals several striking features. It is immediately apparent that the coarsest sediments overlie the platform between the breaker-bar at the platform edge and the beach zone to the northeast, with the coarsest sizes nearest the breaker-bar. South of the platform edge the sediments become progressively finer grained. Figure 5.1 shows that this must have been an area of deepening water offshore. The phi mean distribution map for the pelletal phosphorite (Figure 6.5) is considerably less complex. There is a progressive coarsening of the pelletal phosphorite in a southerly direction towards the edge of the platform. There is a localised area from S12 to T14 where the phosphorite grain-size becomes finer, and this corresponds to an identical trend in the whole sediment (Figure 6.1). On Langeberg the mean grain-size for the pelletal phosphorite is 2,02 Ø (18 samples) with a range 1,8 to 2,3 Ø. On Witteklip the pelletal phosphorite tends to be finer grained, mean 2,27 Ø (4 samples) but with a wider range, 1,5 to 2,7 Ø. The fact that the pelletal phosphorite has a narrow size range, and if it is accepted that it has reacted to the environment of deposition in the same way as the quartz, would explain the low concentration south of the platform where the quartz is finest grained.

The phosphatic-quartzose sands are generally moderately sorted (Figure 6.2). There is a general improvement in sorting south of the platform edge where the mean grain size decreases. The pelletal phosphorite is moderately sorted (Figure 6.6) but better sorted than the associated sands. They become progressively more poorly sorted in a south-easterly direction which is opposite to the trend for the whole sediment. From U6 to S10 there is an area of well-sorted pelletal phosphorite which coincides with the better sorted part of the whole sediment.

Zones of skewness for the sediment are related to the topography (compare

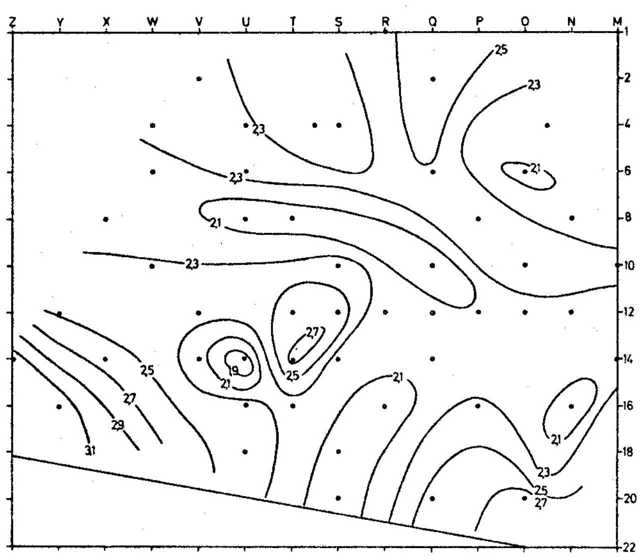


Figure 6.1 Phi mean distribution map, Langeberg.

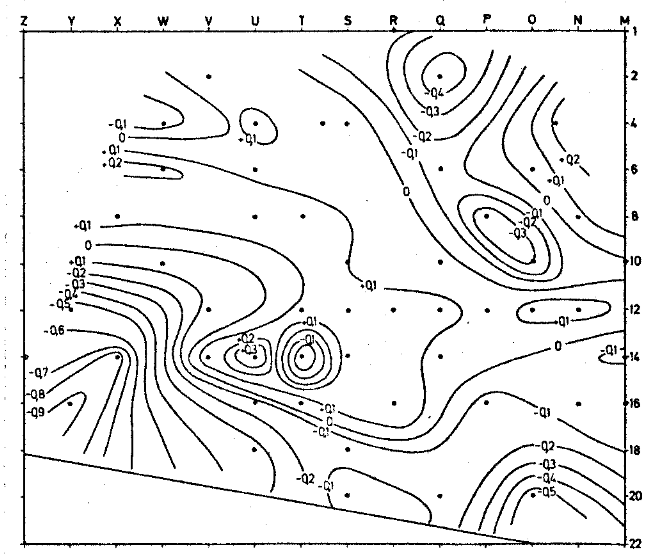


Figure 6.2 Skewness distribution map, Langeberg.

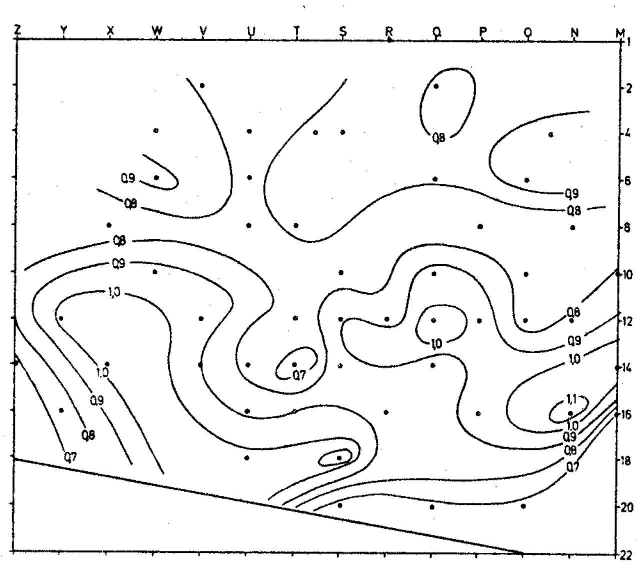


Figure 6.3 Inclusive standard deviation distribution map, Langeberg.

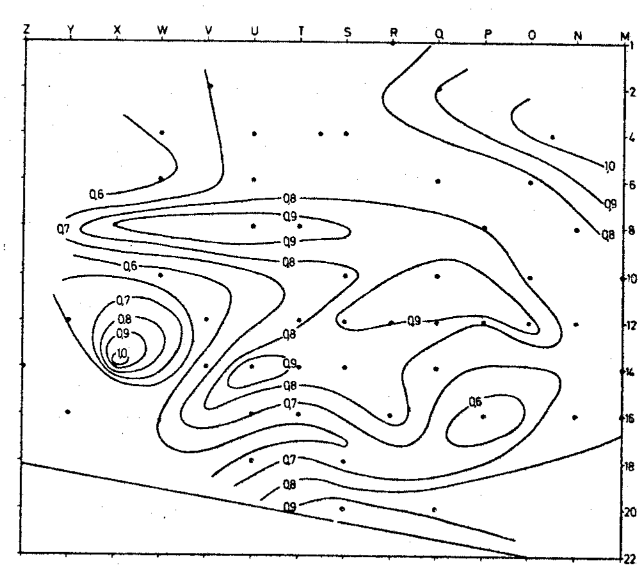


Figure 6.4 Kurtosis distribution map, Langeberg.

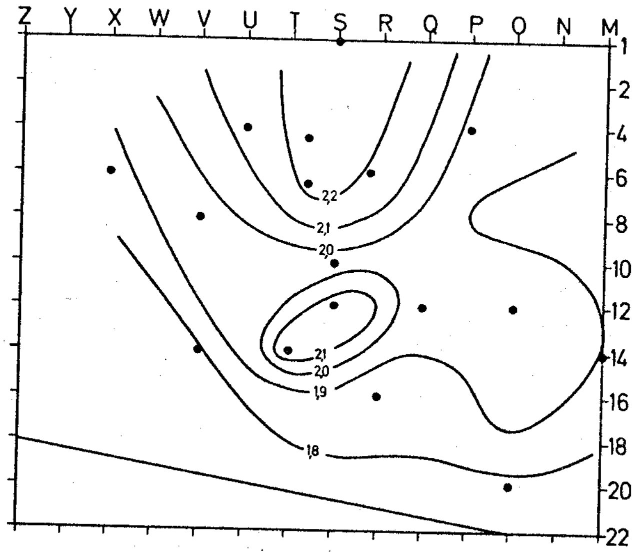


Figure 6.5 Phi mean distribution map for pelletal phosphorite component, Langeberg.

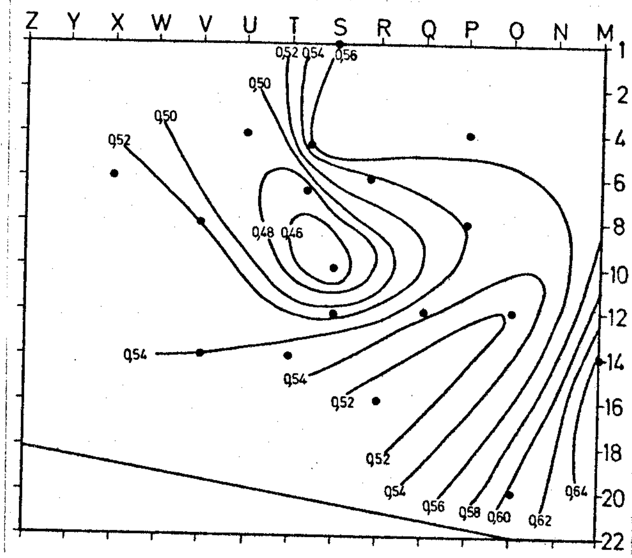


Figure 6.6 Inclusive standard deviation map for pelletal phosphorite component, Langeberg.

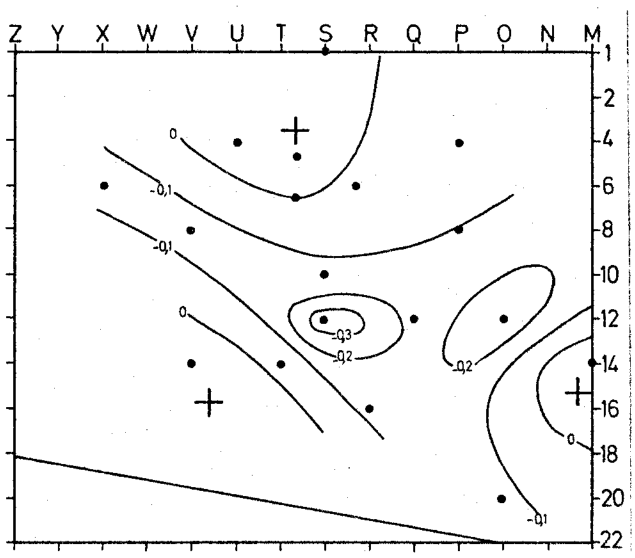


Figure 6.7 Skewness distribution map for pelletal phosphorite component, Langeberg.

Figures 6.3 and 5.1). An area of positive skewness, i.e. the tail towards the fine fraction, stretches from northwest to southeast along the platform. South of the platform skewness becomes markedly negative, i.e. the coarser fraction relative to the fine fraction is dominant. There is a prominent zone of negative skewness centred around Q2 and extending southeast to M10. This zone is seen to lie along the inner step, as the other negatively skewed zone lies along the outer step. At Langeberg negative skewness is associated with the finest sediments. This is probably the result of winnowing of fine sediments that had previously crossed the platform area. Only the finer grained sediments were crossing the platform to the slightly deeper offshore area where conditions prevented deposition of the finest particles. A beach sand, on the other hand, would owe its negative skewness to winnowing of the fines from a lag deposit (Friedman 1961). The skewness distribution map for the pelletal phosphorite fraction (Figure 6.7) contrasts with that of the entire sediment. In the southwest the pelletal material tends to be positively skewed. Likewise, it is positively skewed in the north and east. Whereas the platform sediments are positively skewed, the phosphorite component there is negatively skewed. In Figure 6.8 skewness of the whole sediment has been plotted against mean grain-size. Positive skewness is associated with an average grain-size greater than 2,3 $\emptyset$. As the proportion of very fine sand and silt relative to the fine sand increases, so the skewness values decrease. Zero skewness is associated with subequal amounts of fine sand and very fine sand and silt. South of the platform edge and breaker-bar where the sediments are markedly negatively skewed, they are also finer grained. Fine sand overlying the platform is associated with positive skewness. Duane (1964) considers that negative skewness is produced by winnowing of the fine particles (high energy) while positive skewness is the result of deposition of fine material in a sheltered environment (low energy). Interpreting the skewness distribution in this way would suggest that the shoal area along the outer edge of the platform was a site of erosion (winnowing) under relatively low energy conditions while the platform itself was a sheltered depositional area.

It will be seen from Figure 6.4 that the sediments are generally

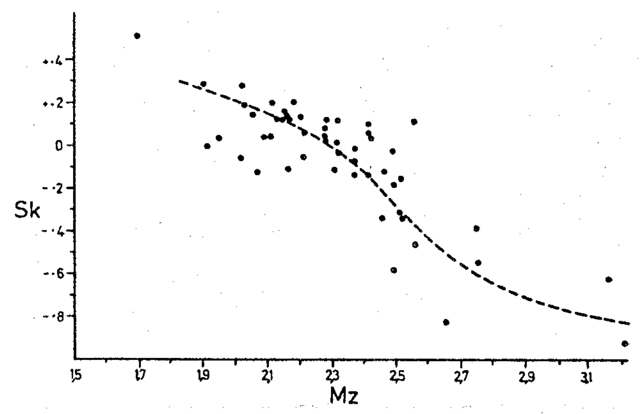


Figure 6.8 Plot of skewness against mean.

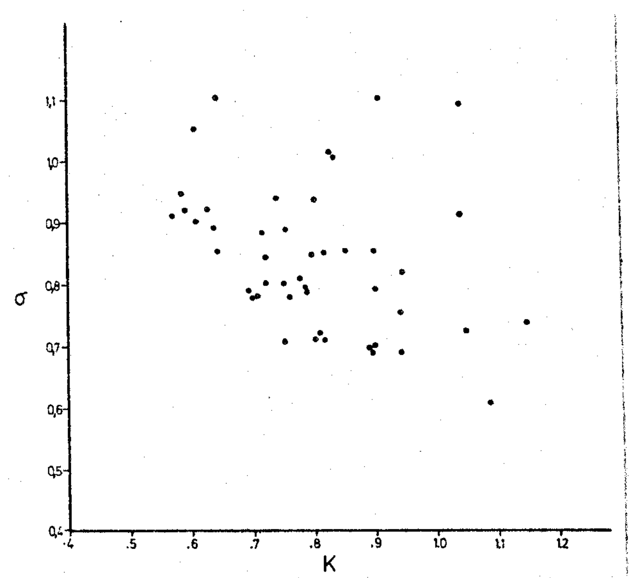


Figure 6.9 Plot of standard deviation against kurtosis.

platykurtic, with kurtosis values ranging from about 0,6 to 1. Very seldom are there sediments with kurtosis greater than unity. The low kurtosis values indicate that the sediments are on the whole better sorted in the tails than in the central part (25 Ø - 75 Ø). Folk and Ward (1957) have found a relationship between skewness and kurtosis, but in this study no definite relationship is apparent because the sediments are spread over a very small size range, although Figure 6.10 shows that skewness and kurtosis have responded similarly to grain size variation. Worst sorting is associated with equal amounts of the two modes (Folk & Ward 1957). Figure 6.9 shows that with the improvement in sorting, i.e. the progressive dominance of one mode, kurtosis values increase.

Grading analyses were performed on regularly spaced samples within individual boreholes (V14, S12, Q12, N4-76-04, M6-15-N6-15, W'1, W'5). The results of these analyses are shown in Figure 6.10. Perhaps the most interesting feature is the skewness. In boreholes V14, S12 and W'1, the skewness is predominantly positive, suggesting low energy conditions. In Q12, M6-16-N6-15 and W'5 there is an alternation between negative and positive skewness. Such an alternation can be attributed to fluctuating energy levels; negative skewness resulting from erosion, positive skewness from deposition. It will be seen from Figure 6.10 that the standard deviation, skewness and kurtosis have responded similarly to changes of grain size.

Visher (1969) has based his analyses of sediments on a recognition of sub-populations within individual grain-size distributions. He relates each lognormal sub-population to different modes of sediment transport and deposition. In this way he recognises three modes: suspension, saltation and surface creep. Of all the grading-analyses performed in this study, two basic types of grain-size distribution predominate with a gradation between them. These two basic types are shown in Figues 6.11 and 6.12. In Figure 6.11 the curves contain three populations separated by inflexion points at 3,00 Ø and 1,25 Ø. Following Visher, the curve between 3,00 and 1,25 Ø represents the saltation population. The curve suggests that the saltation population may itself be composed of two sub-populations. The curve coarser than 1,25 Ø

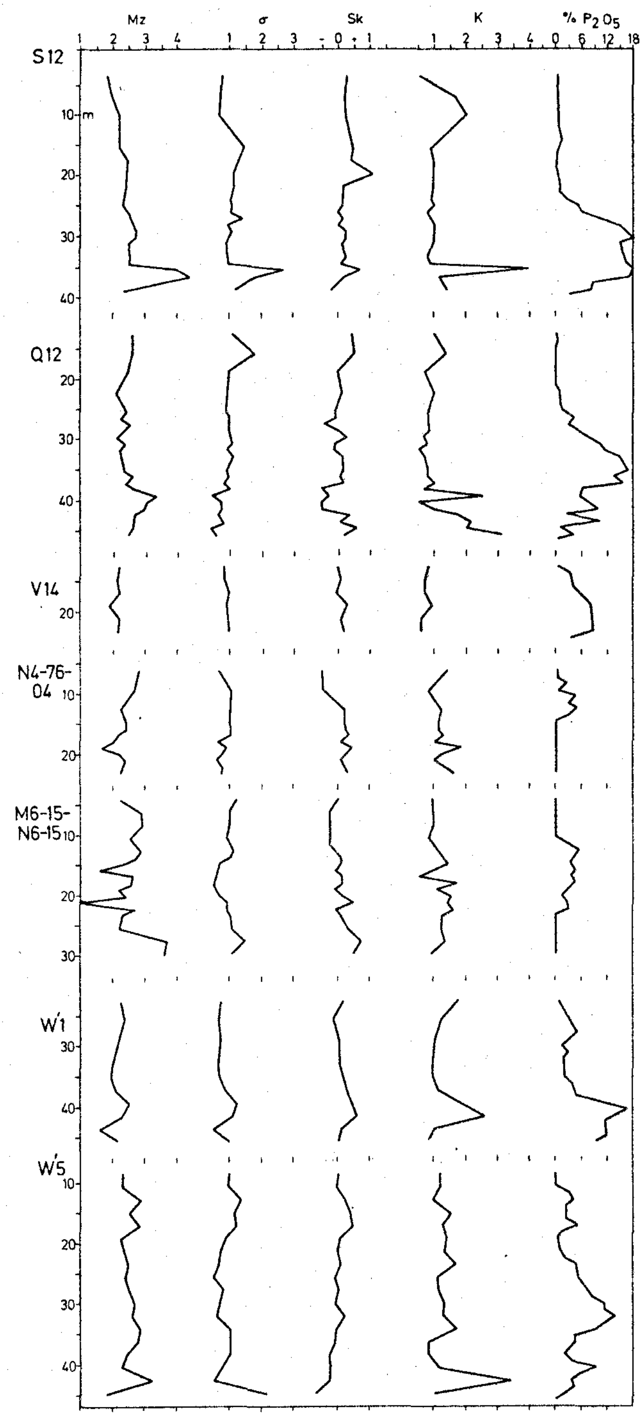


Figure 6.10 Variation of mean, standard deviation, skewness, and kurtosis, and P_2O_5 in individual boreholes, measured every 1,07m.

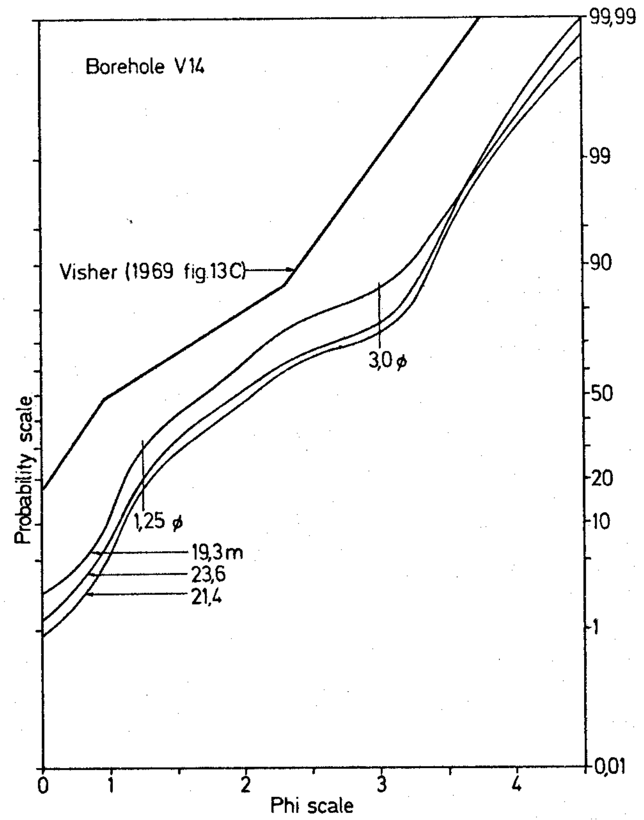


Figure 6.11 Examples of sediment grading curves suggesting shoaling conditions.

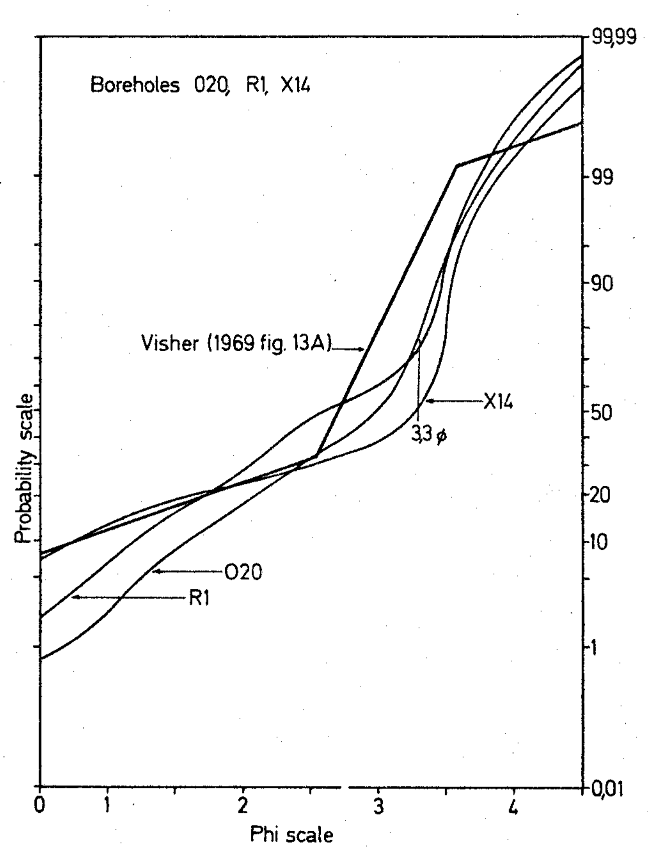


Figure 6.12 Examples of sediment grading curves from sheltered positions.

would be due to the surface creep population and that finer than 3,00 Ø is attributed to suspension. The saltation population is poorly sorted and covers a wide range of grain-size (1,75 Ø), while the surface creep and suspension populations are better sorted. Also included in Figure 6.11 is a curve that Visher found to be typical of a shoal area. If one locates the position of samples producing such curves on the structure contour map (Figure 5.1) it is found that they lie to the south of the platform edge, or are concentrated west of the "S"-drill line. This is extremely significant for it has already been demonstrated (Figure 6.3) that the same area is predominantly negatively skewed. The structure contour map shows that these points are concentrated either on a shoal area at the platform edge or on the more exposed western part. Visher (1969) notes that the fine saltation population suggests winnowing by wave action, while the poorly sorted intermediate population suggests "dumping from a highly turbulent graded suspension-traction carpet, and the coarse population suggests bedload transport by a strong current". All samples with such grain-size distributions are concentrated in an area where such conditions could quite conceivably have operated.

At the opposite extreme are the curves of Figure 6.12, again compared with a diagnostic curve of Visher's. Again it is found that there is an inflexion point at 3,3 Ø between the suspension population and the well-sorted saltation population. The characteristics of such a size distribution are thought by Visher to be typical of deposition by oscillation waves. This would be consistent with the location of the sampled boreholes.

It was stated earlier that the sediments of the Pelletal Phosphorite Member are essentially bimodal, or even polymodal. In Figure 6.13 the grain-size distribution of the whole sediment and the pelletal phosphorite and quartz fractions in a particular sample are compared by means of histograms. It will be seen that the pelletal phosphorite is essentially coarser grained than the quartz fraction. The pelletal phosphorite grain-size distribution is unimodal and nearly lognormal. D'Anglejan (1967) argues that these are original properties of the

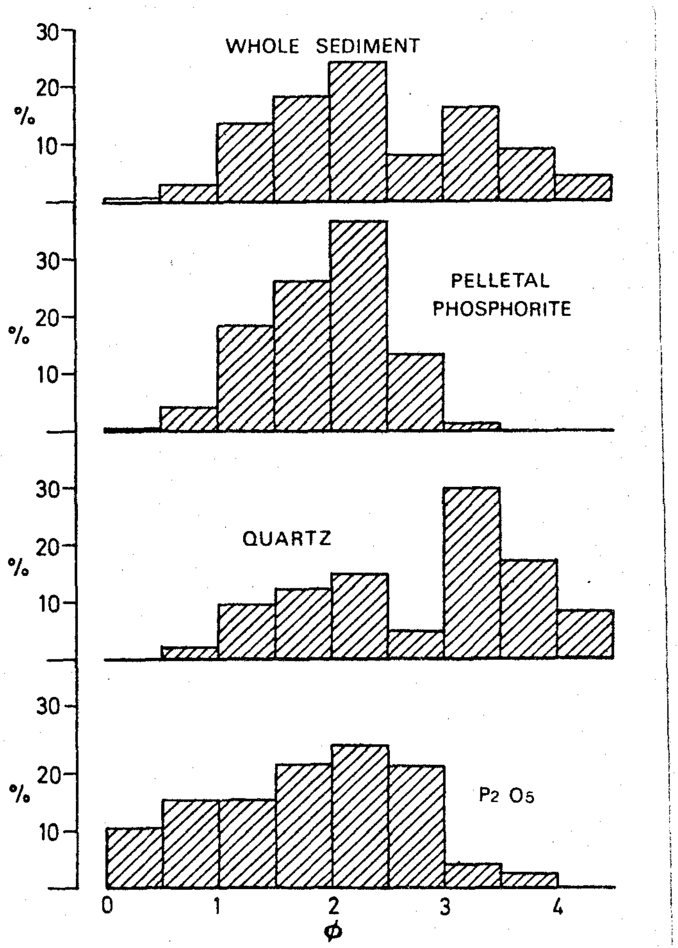


Figure 6.13 Histograms derived from grading analyses of a particular sample, showing how the individual pelletal phosphorite and quartz components contribute to the sediment as a whole. The size distribution of the pellets is also compared with P_2O_5 values for each size grade.

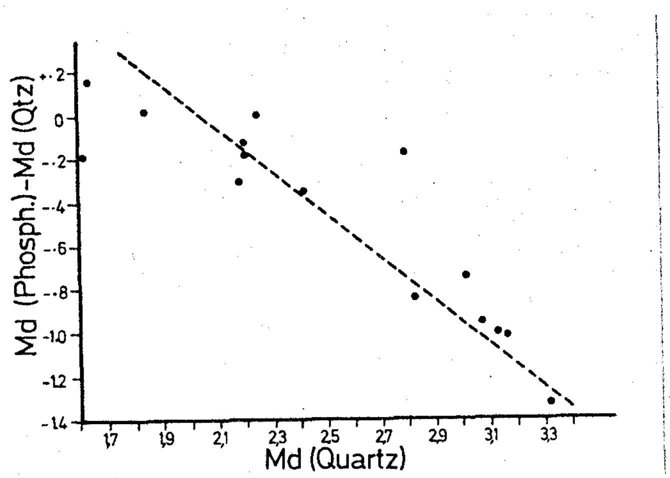


Figure 6.14 Plot of median grain-size differences between quartz and pelletal phosphorite against the median of quartz.

phosphorite rather than a result of sorting, as there are no silt-size grains of apatite. The quartz fraction on the other hand is bimodally distributed, with the principal mode at 3 to 3,5 Ø and having a subsidiary mode at 2 to 2,5 Ø. The subsidiary mode for the quartz coincides with the pelletal phosphorite mode. The presence of the pelletal phosphorite in this sediment sample (45,5 per cent apatite) has resulted in the principal mode being situated between 2 and 2,5 Ø. This relationship holds good for nearly all the samples. To understand fully the cause of this bimodality within the quartz fraction the surface textures of the grains were examined in detail. If it is assumed that the bimodality of the quartz population has resulted from different sources of supply of material, then it might be expected that the roundness of the grains would differ. But the unimodality of the phosphorite suggests just one generation for that mineral. Examination of the roundness of the individual quartz grains (following Powers 1953) has shown that grains coarser than 2,0 Ø are mixtures of rounded to well-rounded and angular to subangular populations. The rounded grains are usually polished or only lightly frosted, although extensive frosting has been noted. Finer than 2,0 Ø the grains are predominantly angular to subangular. Rounded grains are easily inherited and it is not the most rounded grains, but the most angular grains that are the true index of the amount of rounding that has taken place at that particular site (Folk & Ward 1957). In mixtures of rounded and angular grains Folk and Ward believe that the rounded grains are almost always reworked from an older deposit. The fracturing of many of the well-rounded grains in this deposit supports this idea.

Whereas the quartz grains are best rounded in the coarse fractions, the pelletal phosphorite shows a reversal of this behaviour. Coarser than about 1,5 Ø the pellets are present as subangular plates, but roundness and sphericity improve with diminishing size. At 2,5 Ø the pellets are nearly all well-rounded and polished. This reversal of behaviour is explained by its soft (hardness of apatite is 4,5 to 5) but brittle nature, the larger particles tending to fracture and remain subangular. Finer than 2,0 Ø the phosphorite abrades more rapidly than the quartz. The spheroidal and discoidal shape of the pellets results in their behaving differently under turbulent conditions than the coarser plates. The sub-spherical pellets are probably in hydraulic equilibrium with the coarser platey pellets. The high degree of rounding and sorting of the pellets is characteristic of deposition in a littoral environment. Bushinsky (1964) argues that the majority of pelletal phosphorites were found in water shallower than 100m.

VII. HYDRAULIC EQUIVALENTS.

The two major constituents of the Pelletal Phosphorite Member sediments are detrital quartz and pelletal phosphorite, while the major heavy-mineral component, ilmenite, is usually less than 1 per cent. Garnet and rarer zircon make up only about 3 per cent of the heavy-mineral fraction. Grading analyses showed that the ilmenite was concentrated in the very fine sand or silt grades. An attempt was consequently made to test for hydraulic equivalence between the quartz, pelletal phosphorite and ilmenite grains.

The association of different detrital minerals within a deposit immediately suggests a similar response to the same hydraulic conditions. Rittenhouse (1943) defined the concept of hydraulic equivalence: "Whatever the hydraulic conditions may be that permit the deposition of a grain of particular physical properties, these conditions will also permit deposition of the other grains of equivalent hydraulic value". Thus the difference in grain-size between ilmenite, pelletal phosphorite and quartz occurring together may be expected to be compensated for by their difference in density. The grains deposited from suspension under uniform conditions should have the same settling velocities. Shape may be expected to be an important factor affecting settling velocities, but for these three minerals shapes are very similar.

According to Stokes' law particle size is inversely proportional to particle density. The settling velocity of a particle is dependent upon the resistance of the settling medium. For particles smaller than 0,14mm (2,85 $\emptyset$) the resistance is dependent upon fluid viscosity (Rubey 1933). Particles larger than this are controlled by the impact law (Newton's law of resistance). Rubey shows that the settling velocity of small grains is described by Stokes' law, while grains larger than 0,14mm follow an impact law.

McIntyre (1959) has presented a formula that relates the theoretical differences in size to differences in mineral density:

$$\phi_A - \phi_B = \frac{1}{X} \left(\log_2 D_A - \log_2 D_B\right)$$

where ϕ_A and ϕ_B = phi mean (or median) for minerals A and B,

D_A and D_B = density of these two minerals in water.

When X = 1, the equation is the impact law hydraulic equivalence formula,

X = 2, the equation is the Stokes' law hydraulic equivalence formula.

Theoretical grain-size differences were calculated from the above equation using mineral densities of 2,655 for quartz (A), 2,91 for phosphorite (B), 4,70 for ilmenite (C).

	X = 1	X = 2
$\phi_C - \phi_A$	1,16	0,58
$\phi_C - \phi_B$	0,96	0,48
$\phi_B - \phi_A$	0,20	0,10

Comparison between the mineral sizes has been recorded as median-diameters solely for the ease of measuring. The median grain-size and grain-size differences are presented in Table 6.1.

TABLE 6.1

Comparison of median grain-size differences between Quartz, Pelletal Phosphorite and Ilmenite (phi units)

Sample		Depth	Quartz A	Pelletal Phosphorite B	Ilmenite C [a]	Quartz Saltation popn. D	B – A	B – D	C – A	C – B
Witteklip	W'5	30,0	2,75		< 4				> 1,25	
	W'5	32,1	2,80	2,63	< 4	2,25	–0,17	0,08	> 1,20	> 1,37
	W'5	36,9	2,96		> 4				> 1,04	
	W'5	41,2	2,25	2,32	3,65	2,20	0,07	0,12	1,40	1,33
	W'1	43,8	1,72	1,43	3,50	1,60	–0,19	–0,17	1,88	2,07
Langeberg	M14	18,6	2,18	1,88		2,10	–0,30	–0,22		
	O12	32,9	2,20	2,08		1,90	–0,12	0,18		
	O20	44,7	3,13	1,92		2,75	–1,21	–0,83		
	P 8	21,9	2,82	1,92		2,20	–0,86	–0,24		
	Q12	30,0	1,99		3,92				1,93	
	Q12	35,3	2,22	2,03	< 4	2,00	–0,19	0,03	> 1,78	> 1,97
	R16	33,2	1,85	1,88		1,80	0,03	0,08		
	S10	36,9	3,32	1,98		2,70	–1,34	–0,72		
	S12	33,2	2,62		< 4				> 1,38	
	S12	36,4	3,02	2,27		2,05	–0,75	0,22		
	S12	39,6	3,13	2,10		2,10	–1,03	0,00		
	T14	36,0	3,07	2,13		2,90	–0,94	–0,77		
	U 4	15,5	2,43	2,08		2,22	–0,35	–0,14		
	V14	19,3	1,63	1,80	< 4	2,60	0,17	0,20	> 2,37	> 2,20

[a] Precise values for the ilmenite have not been measured in all cases since it is obvious that this mineral is not hydraulically equivalent to the quartz or pelletal phosphorite.

These results show that never is the ilmenite hydraulically equivalent to either the quartz or the phosphorite. Comparison of the grain-size distributions of the pelletal phosphorite with that of the associated quartz shows that although the phosphorite is the denser mineral it is, in most cases, also the coarser grained. There may be several reasons for the large deviation of the quartz and ilmenite from hydraulic equivalence.

(i) Theoretical hydraulic equivalences are only valid when the reference mineral (quartz) and the mineral being tested (ilmenite) have the same shape (McIntyre 1959). Here shape differences are thought to be minimal.

(ii) There may also be a lack of coarser size fractions of the heavy-mineral. In this respect the ilmenite from Witteklip where the Varswater Formation lies upon granite is coarser than the ilmenite from Langeberg where it has probably been derived from the Saldanha Formation. It is quite feasible that the size distributions of the heavy and light minerals are inherited from older sediments.

(iii) Testing for hydraulic equivalence has been performed by sieve-analysis which possibly does not accurately predict the original settling velocities which are the only true measure of hydraulic equivalence.

(iv) A very real source of error is undoubtedly the sampling. McIntyre has demonstrated that hydraulic equivalents should only be computed for mineral species from the same lamina, and he argues against bulk sampling. Unfortunately, the use of the 1,07m tube used here for coring has made it impossible to sample individual laminae. In many borehole samples mixing of the coarse and fine laminae has taken place. Such samples have been avoided where possible.

(v) However, the most likely explanation for the large discrepancy lies in the application of the laws governing the settling velocities of the minerals. The settling velocity of the ilmenite, being smaller than 2,85 ∅ is described by Stokes' law while the larger quartz is governed by the impact law.

It was shown earlier that the different sub-populations recognised in the grain-size distribution curves could be attributed to different modes of sediment transport. It was also shown that there is a consistent inflexion point at 3,0 to 3,8 ∅ separating the suspension from the saltation populations. This inflexion point represents the junction of the Stokes' law and the impact law formulae , the theoretical junction point being calculated at 2,85 ∅ (Rubey 1933). The size distribution of the ilmenite should thus be compared with the size distribution of the suspension population of the quartz. But the measurements show that there is still a divergence from hydraulic equivalence between these two populations, although the divergence has decreased. As most of the samples from Langeberg come from an area that was once sheltered by a breaker-bar, it is likely that the low energy environment prevented efficient selective sorting. Decrease in the energy results in a decreasing capacity to suspend <u>all</u> the sizes so that a perfect selection of grains by settling velocity is not even theoretically possible. Furthermore, once deposited from suspension the grains would be moved on the bottom by traction and would hence be sorted into groups determined by ease of sliding or rolling, and sorting would thus not be related to settling velocity.

McIntyre (1959) believes that movement in the foreshore zone is mainly in suspension, but as velocity drops transportation by traction predominates. The drop in velocity is accompanied by deposition of finer and finer grains of the heavy mineral. Once deposited the smaller ilmenite would be more difficult to entrain than the larger hydraulically equivalent quartz.

The effect of environment can be tested by examining the departure from hydraulic equivalence between the quartz and the pelletal phosphorite. The settling velocities of both these minerals are governed by the impact law. It will be seen in Figure 6.14 that as the individual quartz samples become finer grained, there is a tendency for the pelletal phosphorite to become increasingly larger than the associated quartz. Of the twelve samples analysed on Langeberg only two (V14 and R16) showed the phosphorite to be finer grained than the quartz and to approach hydraulic equivalence. Both boreholes V14 and R16 lie on the edge of the platform where sediment analyses have shown more turbulent conditions to have operated. The hydraulic equivalents for these two samples approach the theoretical impact hydraulic equivalence value (0,20). Overlying the platform, and sheltered by the breaker-bar, low energy conditions have permitted the settling of the suspension load which has decreased the median diameter of the quartz grains. Prevention of deposition of the suspension load at the platform edge has resulted in a closer approach to hydraulic equivalence. If the median diameter of the quartz saltation population is measured and compared with the pelletal phosphorite median grain-size ($\emptyset_B$-$\emptyset_D$) there is a tendency to approach hydraulic equivalence (Table 6.1), although only 6 of the 15 samples are hydraulically equivalent. The departure of the phosphorite and quartz from hydraulic equivalence reflects the nearness of the source of the pelletal phosphorite. Another limiting factor is the composition of the pelletal phosphorite fraction. It has already been shown that the pelletal phosphorite has a unimodal distribution, but that the coarser subfraction is platey and the dominant subfraction pelletal. It seems that these two parts could be hydraulically equivalent,and that combining them would possibly bring the entire pelletal population closer to hydraulic equivalence with the quartz.

It is interesting to compare the phosphorite occurrence with that of the uraninite from the Witwatersrand banket. Nel (1958) considered that the uraninite had been precipitated within the pore-spaces of the sediment. This has resulted in minute rounded particles of uraninite being the same size as the sand grains. Koen (1961) agrees that this hypothesis explains the oval shape of the uraninite grains and their concentration in the banket. But he deems it unlikely that uraninite grains so formed would consistently show exactly the required size-distribution to indicate hydraulic equivalence with the banket fractions. Koen concludes that the "uraninite nodules, which formed in the muddy floors of vast marshes or shallow lakes, were churned up by wave action initiating a new cycle of sedimentation and during the process, became redistributed, mixed with other sedimentary material, sorted and eventually deposited together with sand and gravel". This short description agrees very favourably with the suspected origin of the pelletal phosphorites, except that the phosphorite was almost certainly precipitated within the pore-spaces of a marine sand with organic matter behaving as a catalyst, and as an inhibitor to carbonate precipitation. At a later stage reworking produced the pelletal phosphorite.

It may be helpful at this stage to summarise the preceding discussion. The coarsest sediments of the Pelletal Phosphorite Member on Langeberg are found overlying the platform between the breaker-bar at the platform edge and the beach zone to the northeast. Only the finer grained population was crossing the platform to the deeper, offshore area at the platform edge. The sheltered environment on the platform has permitted the settling of the mud fraction and produced positive skewness. The platform edge, on the other hand, was a shoal area where erosion (winnowing) under relatively high energy conditions resulted in negative skewness. Deviations from hydraulic equivalence between the pelletal phosphorite and quartz are explained partly by location of the samples. Hydraulic equivalence is approached at the platform edge where turbulent conditions operated. Maximum departures are found on the sheltered platform area where a lower flow regime could not suspend all the grains so that selection by settling velocity alone would take place.

(Size statistics are tabulated in the appendix).

CHAPTER 7

DIAGENETIC SURFACE TEXTURES ON QUARTZ GRAINS FROM THE VARSWATER FORMATION

I. INTRODUCTION

During transportation, accretion, and compaction detrital grains may be mechanically abraded and chemically altered (Krinsley & Takahashi 1962). A voluminous literature discusses the consequent surface textures (e.g. Krinsley & Donahue 1968a; Coch & Krinsley 1971; Krinsley & Margolis 1971; Doornkamp & Krinsley 1971; Margolis & Krinsley 1971; Blackwelder & Pilkey 1972; Tankard 1974d). Many electron microscope studies have been performed on the relationship of surface textures to sedimentary environments, but relatively few have concentrated on diagenetic features. Waugh (1970), Pittman (1972) and Tankard and Krinsley (1974) have described solution textures and multiple overgrowths on quartz.

In the previous chapter conventional grain-size analysis was used as a means of interpreting the depositional environment. The purpose of this chapter is to present the post-depositional history of the major mineral component (quartz) in a highly energetic chemical environment. In all samples examined, approximately the same conditions of solution and precipitation prevailed. The main characteristics to be described are:

(i) crystallographically controlled solution surfaces,

(ii) solution surfaces independent of internal symmetry,

(iii) reprecipitation textures.

All samples used in this study were collected, examined and interpreted by the writer who benefited from discussions with Professor D.H. Krinsley of New York University. The results were published in a joint paper (Tankard & Krinsley 1974).

Methods used in this study are described in Appendix 2. Only the 0,7-1,0mm size range was used.

II. SEDIMENT CHEMISTRY

The considerable extent of post-depositional diagenetic textures preserved on the quartz grains can be related to the chemical energy of the environment. The chemical characteristics of the sediments are summarised in Table 7.1. The pH range is 6,9 to 8,9 while the average pH of the pelletal phosphoritic sands and the overlying Pleistocene aeolian sands are similar, 8,2 and 8,1 respectively.

TABLE 7.1

Chemical Characteristics of the Sediments

Bore-hole	Depth m	pH	P_2O_5 %	Na (ppm)	K (ppm)	Ca (ppm)	Mg (ppm)	Conductivity micro-ohms
N6	22	6,9	–	800	30	360	180	2500
Q12	18	8,0	Trace	80	40	1540	40	330
S12	15	8,7	0,7	10	20	1100	20	85
S12	26	8,7	0,9	50	20	1540	40	150
S12	35	8,0	17,9	400	40	860	110	650
S18	38	8,4	6,1	10	20	280	30	150
T16	29	7,8	5,4	140	20	330	50	410
U14	29	8,3	12,2	50	20	240	20	160
V14	13	8,9	1,8	40	40	840	30	135
W'1	45	8,3	12,0	20	20	260	20	170
W'5	27	8,0	6,2	360	20	440	40	650
Y16	10	7,7	4,9	30	40	290	40	180

Analysts: Fedmis (Pty) Limited
P_2O_5 – Chemfos Limited

Krauskopf (1959) believes that quartz becomes appreciably soluble above a pH of 9. The pH values shown in Table 7.1 reflect only the magnitude of the present pH range and do not imply pH values at the time of diagenesis. The potassium and magnesium values are moderate. Conductivity is a measure of the total soluble salts present and, as Table 7.1 suggests, appears to be closely related to sodium content.

Permeability is a measure of the relative ease of liquid flow through a sediment; measurements ranged from 0,01-0,39 darcies. The very low permeability is attributed to the high mud fraction of the samples (2,6 to 10,4 per cent). Permeability is also affected by particle size and degree of sorting. In this respect the fine nature of the sand and the moderate to poor sorting accounts for the low permeability rates. Low permeability suggests that the quartz grains are continuously subjected to corrosive alkaline solutions which move slowly through the pore spaces.

III. RESULTS

A. Crystallographic Control of Chemical Etching

Chemical solution of quartz grains will tend to follow planes of least resistance, and the internal structure of the quartz may manifest itself in strongly parallel alignment of the textures. Slow solution at high pH results in aligned V-shaped etch patterns (Krinsley & Donahue 1968a; Doornkamp & Krinsley 1971). This texture is demonstrated in Figure 7.1, where etch triangles have developed on the trigonal face of the quartz.

Chemical etching along the rhombohedral cleavage (Figure 7.2) produces an almost rectangular "grid" texture ($rr' = 85^{o}$). It is clear that both these triangular and rhombohedral forms are composite in structure, consisting of a series of progressively smaller triangles or smaller rhombohedrons within one another.

Solution along parallel cleavage planes may produce a series of steps parallel to a cleavage direction (Figure 7.3) or, if the etching actually penetrates the cleavage plates, a deep cavernous texture

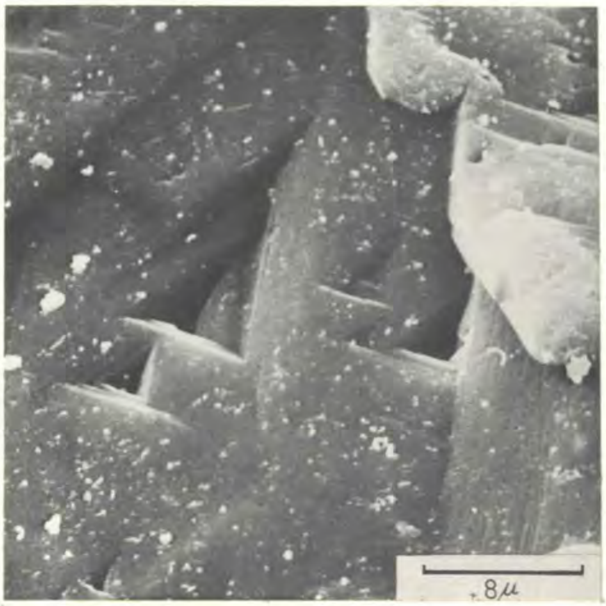


Figure 7.1 Quartz grain surface with depressed triangles the result of chemical etching. Note the triangles within triangles.

Figure 7.2 Quartz grain surface showing rectangular "grid" pattern with mechanical V-shaped patterns on protruding portions of surface.

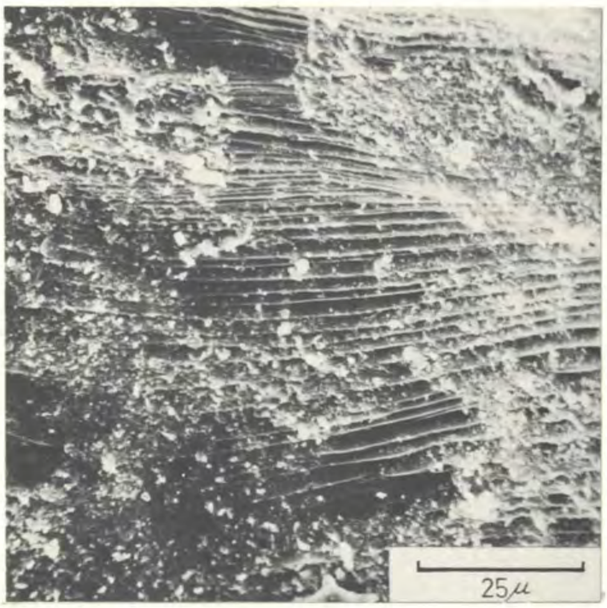


Figure 7.3 Quartz grain surface showing a series of possible cleavage plates, probably modified by solution. The irregular bits of debris resting on the plates are probably due to intense chemical activity.

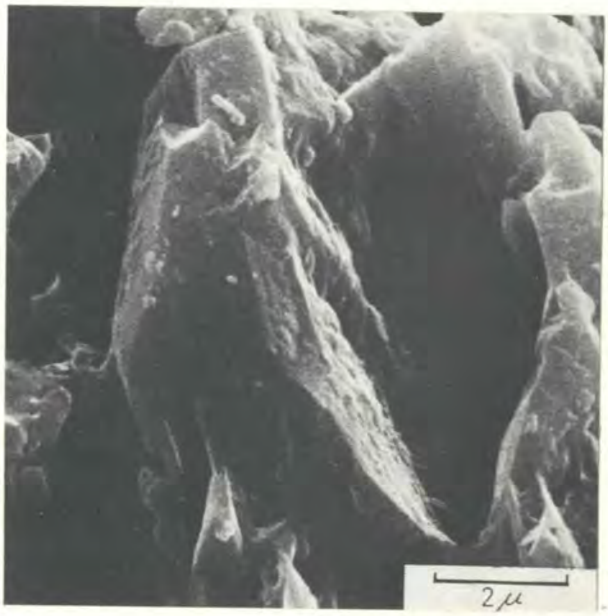


Figure 7.4 Quartz grain surface with cavernous structure, probably due to etching. A series of cleavage plates are exposed here.

results, which is still controlled by the internal symmetry of the quartz (Figure 7.4).

Biederman (1962), Margolis (1968, figure 11), and Krinsley and Donahue (1968a, plate 2), recognised similar features to that of Figure 7.1. These authors attributed the oriented V-shaped markings to chemical solution in low-energy beaches. The Figure 7.1 sample is from a marine horizon in the Varswater Formation. Deposition in an embayment suggests that low-energy conditions prevailed. However, also observed were crystallographically controlled etch triangles on quartz grains from the Middle Stone Age sediments of Die Kelders I identical to an illustration in Krinsley and Margolis (1969, figure 4). It is estimated that the shoreline was 16km away when deposition of the Middle Stone Age sediments occurred (Tankard 1974d). Unless this grain was inherited from an older deposit it would appear that this type of texture could arise also from post-depositional etching, and is not necessarily diagnostic of a marine environment.

B. Solution Surfaces Independent of Internal Symmetry

Chemical solution of a quartz grain will obviously take advantage of any natural weakness, apart from the cleavage directions. In Figure 7.5 low density areas within a single quartz grain (dark patches) are revealed by low energy secondary electrons. Chemical etching has taken advantage of these low density areas and it will be observed that this is where the solution pits lie. This particular grain shows extensive diagenesis, perhaps even a previous history of diagenesis.

Krinsley and Donahue (1968b) believe that solution surfaces develop through three stages. In the first stage solution results in rounded forms, such as the background on the Figure 7.5 specimen, but with continued solution a series of irregular pits develops. The final stage, extreme solution, results in what they describe as resembling "lapies developed in limestone terrain". In Figure 7.6 the last of these stages is observed: an initial flat solution surface has been extensively pitted. It can also be seen at the top of the photograph,

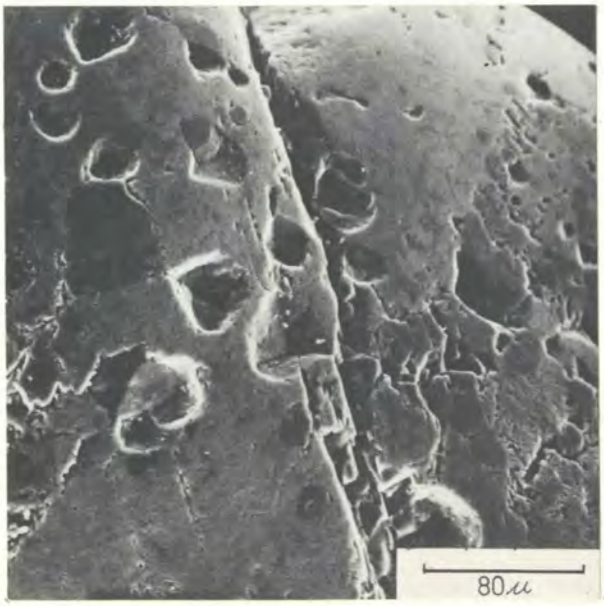


Figure 7.5 Portion of quartz grain surface showing intense rounding due to solution and localised depressions where advanced solution has occurred. The fine lines and pits are also due to solution.

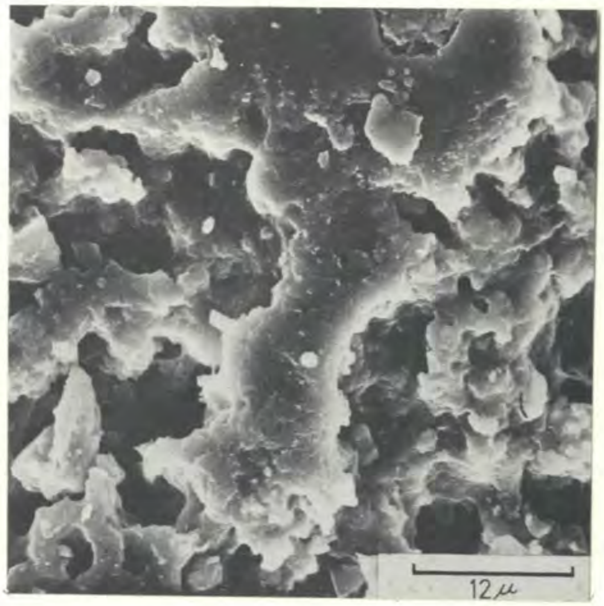


Figure 7.6 Quartz grain surface with intense localised solution, creating "solution holes" resembling lapies in limestone terrain.

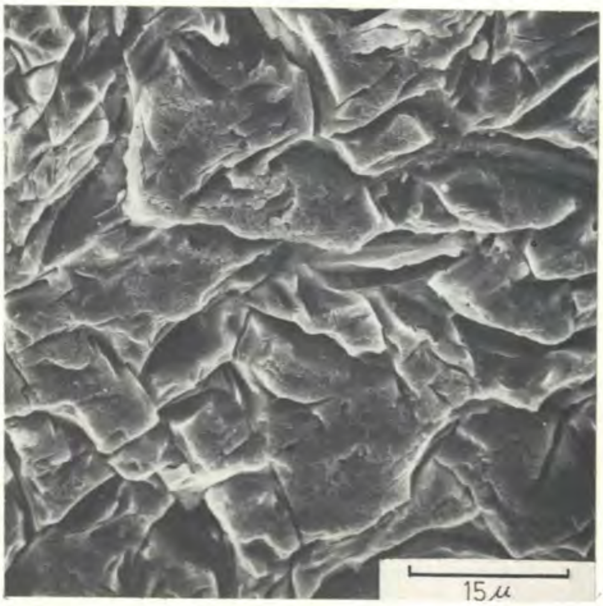


Figure 7.7 Quartz grain surface with network of etch lines; all mechanical features have been eliminated.

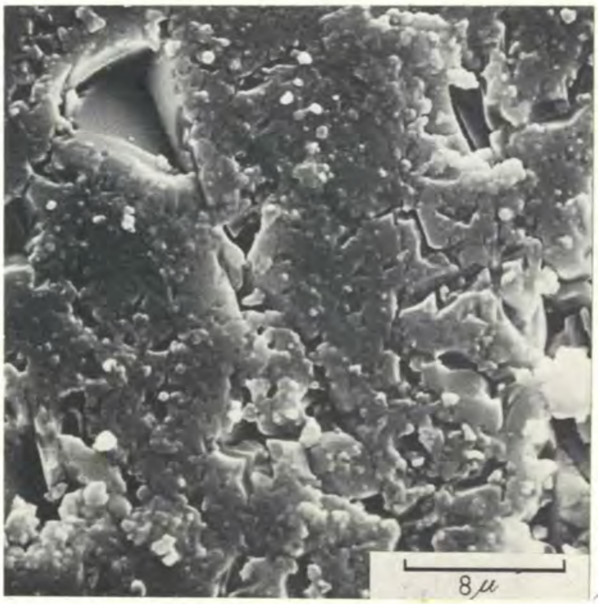


Figure 7.8 Quartz grain surface with post-depositional chemical action. The regular depressions may be the result of resurrection after burial followed by mechanical action.

that fracturing has taken place, probably because of expansion of cleavage plates by means of chemical action. Figure 7.7 shows a different type of diagenesis: furrows forming a branching network. With continued action the larger furrows expand to consume many smaller furrows.

Figure 7.8 shows post-depositional chemical action with considerable energy available in the environment. The irregularly pitted surface formed before the breakage blocks; note that block edges have been relatively unaffected by diagenesis. Pitting of the surface shown in Figure 7.9 again demonstrates the effect of a high energy chemical environment. Cleavage plates are also shown. Doornkamp and Krinsley (1971, figure 10) illustrate a very similar texture from a tropical environment (Uganda). Figure 7.10 shows pitting of a quartz grain; cleavage plates have developed within the pit. The surface of the grain is smooth, as is the surface of the Figure 7.5 grain. Rounding of the cleavage plates within the pit suggests that the smoothing is the result of solution and not abrasion.

C. Reprecipitation Textures

Besides solution features, reprecipitation of silica is frequently encountered. In Figure 7.11 intense reprecipitation has followed a period of solution. This has produced a series of stacks, arches and caves. Krinsley and Doornkamp (1973, plate 22) illustrate a very similar texture.

The reprecipitation of silica shown in Figure 7.11 is in optical continuity with the original grain structure. Sorby (1880) recognised that overgrowths on a host mineral of the same species are in optical continuity with the host, i.e. the overgrowth forms a syntaxial rim.

IV. DISCUSSION

It has been demonstrated that marked diagenesis can occur in geologically

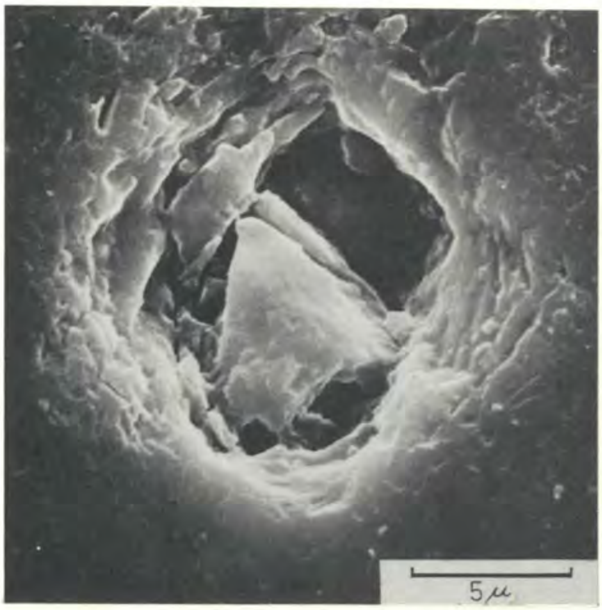


Figure 7.9 Quartz grain surface showing chemical action; cleavage plates are visible in the depression.

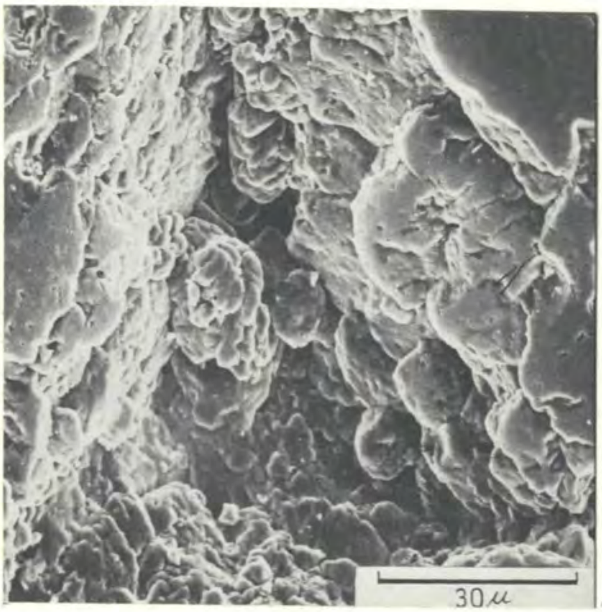


Figure 7.10 Etch pit with cleavage plates; rounded surfaces show a second cycle of solution.

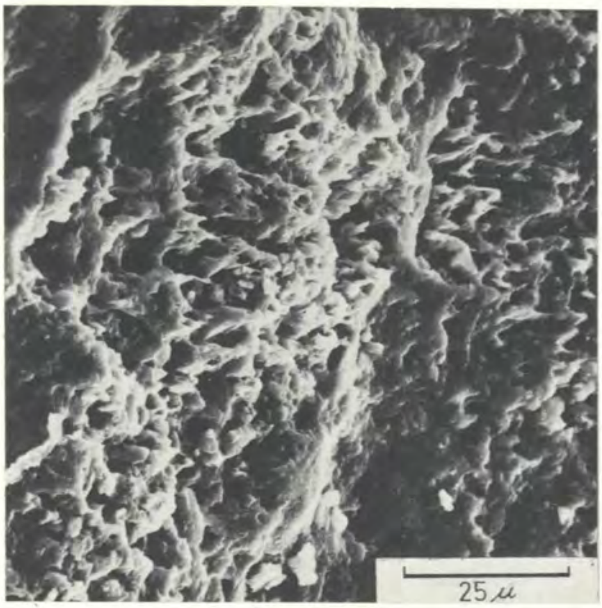


Figure 7.11 Quartz grain surface showing stacks, arches and caves caused by intense chemical solution and precipitation.

young sediments, and may obliterate mechanical textures which are diagnostic of environment. Surface textures of sand grains from the Varswater Formation and pH measurements suggest a post-depositional environment of high chemical energy.

Margolis (1968) has produced triangular and rhombohedral etch patterns (similar to those of Figures 7.1 and 7.2) by etching with low concentrations of hydrofluoric acid and sodium hydroxide, implying a chemical origin for these textures. Solution of quartz grain surfaces in the Varswater Formation has taken place in an alkaline environment. Krinsley and Margolis (1969, figure 4) illustrate chemical etch triangles which they thought to be diagnostic of low-energy littoral conditions. An identical texture was found in a Late Pleistocene cave deposit which was formed when the sea had receded due to advance of the Würm ice sheets. This example can be used to sound a warning about over-hasty application of this technique to interpret environment of deposition. It is possible to produce identical textures by diagenesis and by mechanical abrasion, and only combinations of surface textures should be used to detail environment of sedimentation (Tankard 1974c).

CHAPTER 8

PETROLOGY OF THE PHOSPHORITE AND ALUMINIUM PHOSPHATE ROCK

I. INTRODUCTION

Although submarine phosphorites were first recorded on the Agulhas Bank by the Challenger Expedition of 1873-1876 (Murray & Renard 1891), relatively few studies have been made of South African phosphorites. Collet (1905) described phosphorite dredged between the Agulhas Bank and 25°E longitude. He recorded large slabs of phosphorite, and also conglomeratic varieties. Cayeux (1934) attributed the morphology of the phosphorite nodules to erosion. This line of research was continued by Parker (1971; 1975), Parker and Siesser (1972), Parker and Simpson (1972), and Summerhayes et al (1972). Summerhayes (1973) points out that phosphorite is one of the major rock types of the Agulhas Bank. Research on emerged phosphorite deposits has not been extensive. Haughton (1932) presented chemical data for the Langebaanweg phosphorite, while Frankel (1943) examined the phosphorite by means of X-ray diffraction. The only complete chemical, petrological, and X-ray diffraction studies are those of Tankard (1974b, c; in press a), upon which this study is based. A voluminous literature discusses phosphorite deposits from other parts of the world (e.g. Sheldon 1964; D'Anglejan 1967; McKelvey 1967; Rooney & Kerr 1967; Tooms et al 1969; Trueman 1971).

Bedded phosphorites originate by shallow marine precipitation of collophane in the voids of the sediment, and close to the sediment-water interface. Pelletal phosphorite is the result of erosion of bedded phosphorite and microsphorite, so that the pellets become detrital components of the new deposit. The present day occurrences of phosphorite on land may be attributed to tectonic and eustatic causes.

On the granite hills north and south of Saldanha Bay there are a further two varieties of phosphate rock. Both probably originated by leaching of guano deposits by ground water and the subsequent alteration of the granite bedrock to form aluminium and aluminium iron phosphates. The first variety, the primary phosphate rock, is an insular deposit produced by interaction of the original guano and bedrock. Weathering of this insular deposit has produced the second variety, lateritic phosphate. The phosphate minerals encountered in insular phosphate rocks are mainly crandallite, millisite, wavellite, and members and polymorphs of the variscite-strengite (barrandite) group (Trueman 1971). In contrast the mineral most commonly found in marine phosphorites is a carbonate fluorapatite. Exploitation of lateritic phosphates is restricted to Senegal (Trueman 1971) and Konstabelkop (Tankard 1974b) (Figure 3.2). The aluminium phosphates of the Saldanha area have been discussed by du Toit (1917), Hutchinson (1950), Visser and Schoch (1973) and Tankard (1974b).

II. DISCUSSION OF TERMS USED

Phosphorite

"Phosphorite" is the term generally used to describe marine sedimentary deposits which contain more than 18 per cent P_2O_5, approximating to 50 per cent apatite (Bushinsky 1966). The apatite mineral is generally a carbonate-fluorapatite (Altschuler et al 1958).

It should be noted that other workers prefer a broader definition of the term "phosphorite". For instance, Parker and Siesser (1972) describe as phosphorites those phosphatic sandstones having an average P_2O_5 content of only 15 per cent.

Collophane

The term "collophane" is generally used in petrologic descriptions to describe the isotropic apatite. The use of this term does not imply specific chemical composition. "Collophane" is a useful term that encompasses all the cryptocrystalline carbonate apatites.

Microsphorite

Riggs and Freas (1965) proposed the term "microsphorite" to describe non-pelletal phosphorite in Florida. The term is the phosphorite equivalent of micrite in carbonate sediment nomenclature. Tankard (1974b) and Dingle (1974) have explained some phosphorites as phosphatised micrite. Trueman (1971) outlines microsphorite as follows:

(i) It is an orthochem, i.e. collophane precipitated in situ.

(ii) It is a microcrystalline collophane mud.

(iii) It shows no evidence of transportation.

(iv) It occurs either as cement binding clastic material, or as discrete laminae or beds, or as inorganic controlled structures (moulds, faecal pellets, etc.).

(v) Microsphorite beds exist as discrete conformable stratigraphic units and are continuous.

(vi) The contact with underlying sediment is often gradational, while the upper contact is sharp and marked by an indurated upper bored surface.

Pellets

The term "pellets" is used in a descriptive sense with no genetic connotations. Neither is "pellet" restricted to particles smaller than 0,15mm as defined by Folk (1962). Phosphorite pellets are generally sand-size or smaller, ovoidal, ellipsoidal or occasionally discoidal, structureless particles.

Basal Bed

For convenience the term "basal bed" will be used synonymously with Miocene phosphatic sandstone (Tankard 1974b, 1974c).

III. PETROLOGY OF THE PHOSPHORITES

A detailed optical study of the phosphorite pellets and phosphatic sandstone is largely precluded because of the submicroscopic size (0,25–4 μ) of the apatite mineral and the admixture of argillaceous, carbonaceous and ferruginous impurities. The pellets may all be classified as true phosphorite, having a P_2O_5 content in excess of 18 per cent. But this does not necessarily hold true for all the phosphatic sandstone specimens which will be described along with the true phosphorite.

A. Pelletal Phosphorite

The mean grain size of the pellets is coarser than that of the associated quartz grains in the sediment. The mean pellet size on Langeberg is 2,02 Ø (range 1,8–2,3 Ø), on Witteklip they are finer grained, 2,27 Ø (range 1,5–2,7 Ø). North of Hondeklipbaai the pellets are typically coarser (0,5–1,5 Ø). The amount of pelletal phosphorite in silt-size grades is negligible. The pelletal particles are present in three distinct forms. Those of biogenous origin are present only in trace amounts in the coarser fractions. There is a relationship between pellet size and their gross morphology. In the coarser fractions (2,0 to –0,5 Ø) the pellets are of platey or irregular appearance, while below 2,0 Ø ellipsoidal structureless pellets predominate, with some discoidal pellets.

The increase in degree of roundness with diminishing size displayed by the pelletal phosphorite is opposite to the trend of the quartz grains in the same sediment where roundness improves with increasing grain-size. This reversal of behaviour of the pellets is readily explained by their soft (hardness of apatite is 4,5 to 5) but brittle nature. The larger particles tend to fracture and remain subangular. When finer than 2 Ø, the phosphorite abrades more rapidly than the quartz. The spheroidal shape of the pellets results in their behaving differently from the coarser plates under turbulent conditions. The subspherical pellets are probably in hydraulic equilibrium with the coarser platey pellets. The high degree of rounding and sorting of the pellets is characteristic of deposition in a littoral environment. The high degree of sorting of the pellets contrasts markedly with that of the associated clastic constituents.

Generally the edges of the coarser platey pellets show some signs of wear while many of the grains have a conchoidal fracture, and often a striated surface. Undoubtedly many of these grains are actually phosphatised mollusc shell fragments but such an origin does not account for all of them as d'Anglejan (1967) suggested for his phosphorite. The pellets range in colour from pale yellow to orange to deep red. They are usually of fresh appearance and often translucent. Frequently they have white blemishes due to finely divided clay material, or are mottled black by organic carbon. Generally these grains are of lighter and more yellow colouration than the smaller ovoidal pellets. Frankel (1943) analysed the white portion of nodules from Langebaanweg as montmorillonite.

The ovoidal pellets constitute more than 90 per cent of all pellets in the 1,5–2,5 Ø fraction. They are regular, often ellipsoidal and even discoidal in shape (Figure 8.1). Rod-shaped, but nevertheless well-rounded pellets are also found. Rooney and Kerr (1967) have described similar particles with a groove running down the length of the rod which they thought were probably minute bones, although they concede the possibility of faecal pellets in their sediments. Arakawa (1971) illustrates some faecal pellets of invertebrates which are very similar to those found in the Varswater Formation. Very rare are some well-rounded pellets composed of aggregates of smaller pellets which are phosphate cemented. The ellipsoidal pellets are usually brown, orange or red coloured, but black and green pellets have also been noted. Towards the base of borehole W'5 (Witteklip) the pellets are mostly black due to large amounts of organic carbon (Figure 8.5). Black mottling and white blemishes are common in most of the ellipsoidal pellets.

Included with the biogenous particles are translucent and opaque phosphatised echinoid spines, foraminifera, minute fish teeth, bryozoa

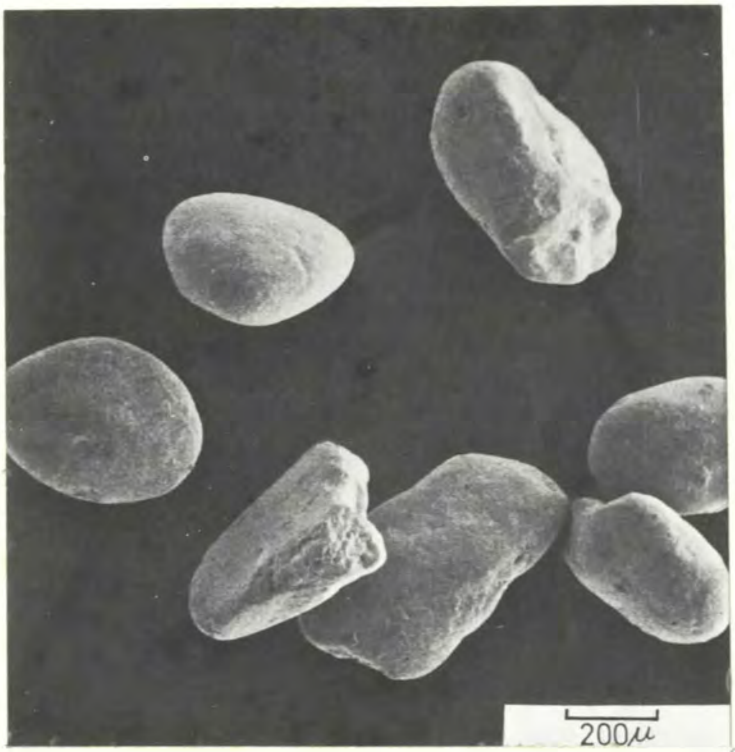


Figure 8.1 Morphology of typical phosphorite pellets.

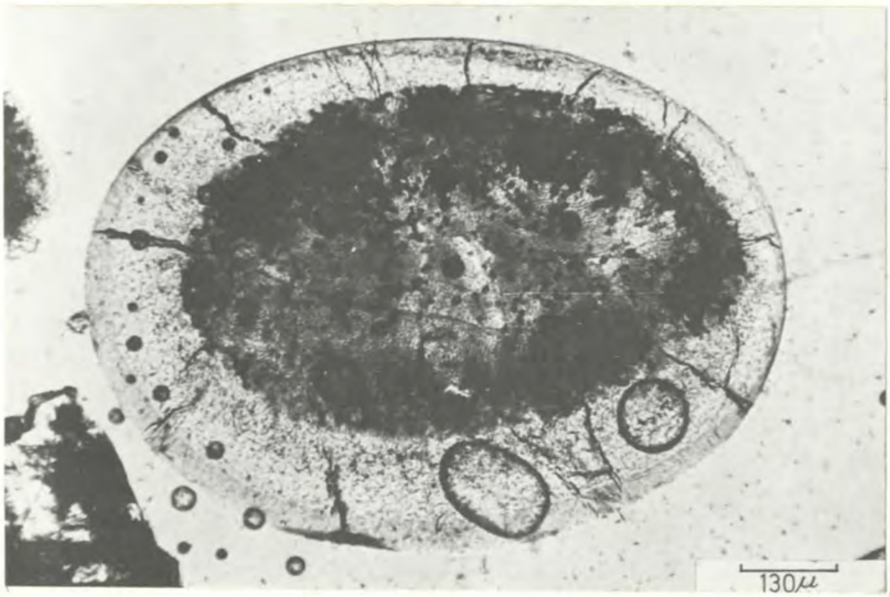


Figure 8.2 Structure of pellet that has resulted from the migration of impurities away from the rim.

and faecal pellets. The echinoid spines generally have rounded extremities and the foraminifera (predominantly Elphidium sp.) also show considerable wear. Some of these pellets resemble pellets illustrated by Manheim et al (1975, figures 4, 5, 6) which have been shown to originate by advanced phosphatisation of foraminiferal tests, particularly Bolivina, Uvigerina and Cassidulina. The faecal pellets are cylindrical and usually have a groove down their length.

Because of the cryptocrystalline state of the apatite, the pellets are difficult to study in thin-section. And the poorly crystalline state results in low intensity peaks by X-ray diffractometry. In thin-section the pellets are yellowish brown and isotropic to very slightly anisotropic. Some pellets have an outer rim of anisotropic apatite.

The pellets are usually structureless, although a structure may occur after apatite has grown about an older pellet, or the growth of the apatite crystals has pushed aside the carbon and argillaceous inclusions (Figure 8.2). Frequently the pellets contain subangular silt-size quartz grains randomly scattered throughout the groundmass (Figure 8.3). These frequently give the impression of a nucleus. But the presence of organic carbon and argillaceous material within pellets suggests precipitation within the pore-spaces of a sediment, the silt-size quartz being also present in these pore spaces. (Attempts to determine the organic carbon content by direct combustion first in an atmosphere of nitrogen and again in an atmosphere of oxygen, produced inconclusive results. However, hydrochloric acid leaching of the apatite left behind a black fibrous organic carbon-like residue which was rapidly oxidized by peroxide). In only one partially corroded pellet was a well developed oolitic structure present showing concentric shells of radiating anisotropic francolite. Figure 8.3 shows the occurrence of a well rounded grain that is predominantly quartz, but includes one portion of phosphorite. Although the phosphorite formed originally as an autochthonous mineral within the pore-spaces of an older formation, in the latest cycle of erosion the phosphorite has become associated with the detrital components of the sediment and has been abraded to the typical pelletal form. Figure 8.4 demonstrates typical pellet form. It also demonstrates the

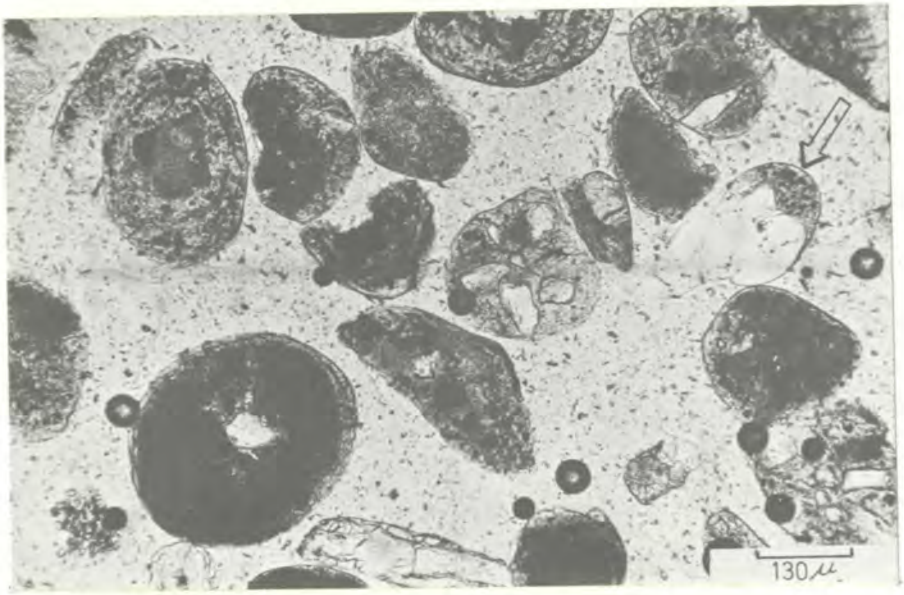


Figure 8.3 Pellets with quartz grain inclusions.
Arrow indicates a well-rounded grain composed of part phosphorite and part quartz.

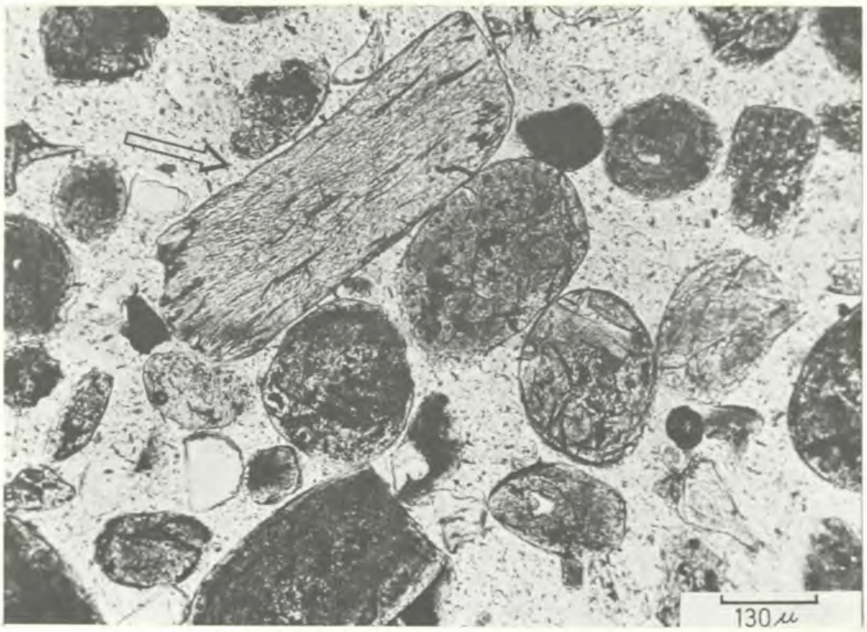


Figure 8.4 Typical structure of pelletal phosphorite.
Arrow indicates biogenous particle.

ubiquitous nature of organic carbon in the pellets, and shows a pellet of biogenous origin. At the base of W'5 (Witteklip) the pellets are saturated with organic carbon (Figure 8.5).

Many of the pellets contain small laths (Figure 8.6) possibly of organic material, but it is impossible to identify these.

Phosphorite pellets with oxidized rims are common. Similar features led Rooney and Kerr (1967) to suspect extensive reworking. The ellipsoidal pellets have had a longer period for oxidation of the organic constituents to take place (their shape implies a longer period of abrasion), whereas the more angular pellets have a fresher appearance due to their comparative youthfulness. The pellets owe their colour to disseminated aggregates of carbonaceous, argillaceous and ferruginous impurities. The iron oxide (Fe_2O_3) content of many pellets is as high as 3,4 per cent. Opaline silica is frequently present within the pellets, and in other cases has formed a shell about the pellet from within which the apatite has been dissolved leaving a fragile empty shell. D'Anglejan (1967) records a similar phenomenon. The coarser platey pellets have in many cases a texture typical of shell debris. Some sections strongly resemble young bone or cartilage (Figure 8.7) and tooth dentine (Figure 8.8). It was found that some of the pellets on Witteklip have a texture very similar to that of the Hoedjiespunt microsphorite. The Langeberg pellets likewise have a texture very similar to that of the matrix of the underlying phosphatic sandstone.

Figure 8.9 shows the minute dimensions of the crystallites in the phosphorite. The texture of the fracture surface of the pellets is further illustrated in Figures 8.10 and 8.11.

B. Phosphate Rock

Phosphatic sandstones have resulted from the precipitation of collophane within the pore-spaces of a quartzose sand. At Hoedjiespunt the microsphorite horizin, 1,5m thick, contains less than 1 per cent quartz. Bushinsky (1966) defines a phosphorite as rock containing ubiquitous nature of organic carbon in the pellets, and shows a pellet of biogenous origin. At the base of W'5 (Witteklip) the pellets are saturated with organic carbon (Figure 8.5).

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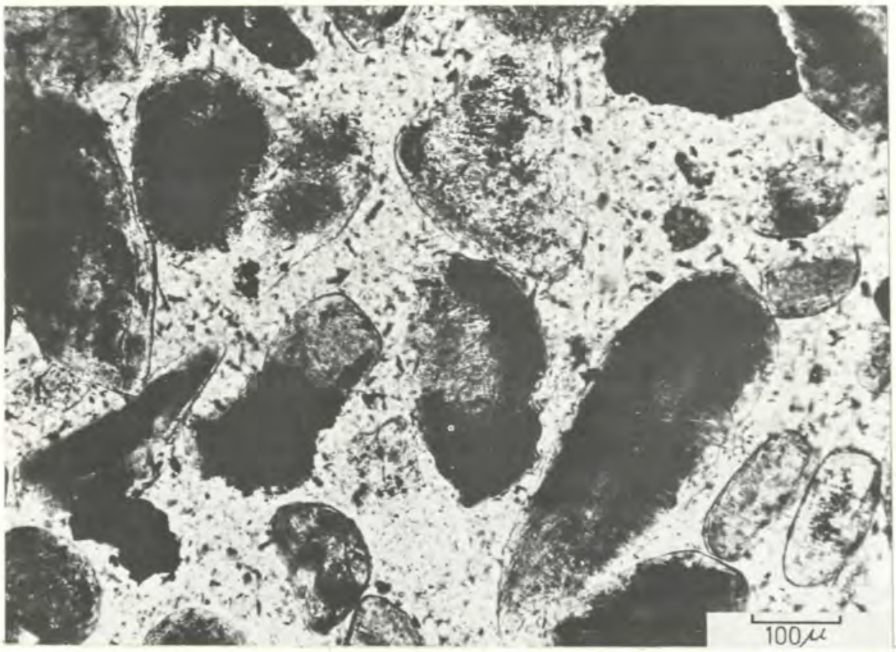


Figure 8.5 Pellets coloured black due to carbonaceous material.

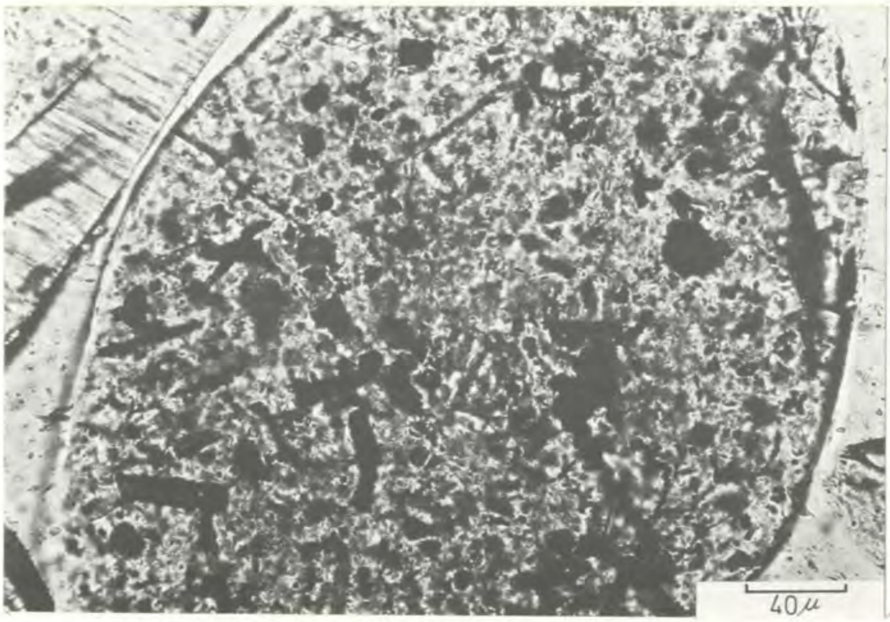


Figure 8.6 Organic material present as small laths.

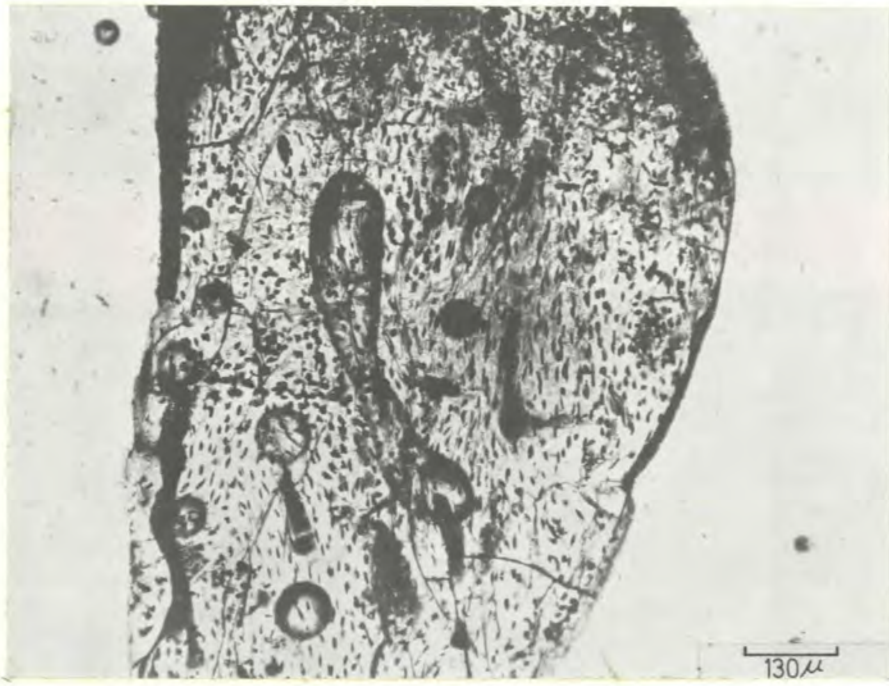


Figure 8.7 Pellet structure resembling young bone or cartilage.

Figure 8.8 Structure similar to tooth dentine.

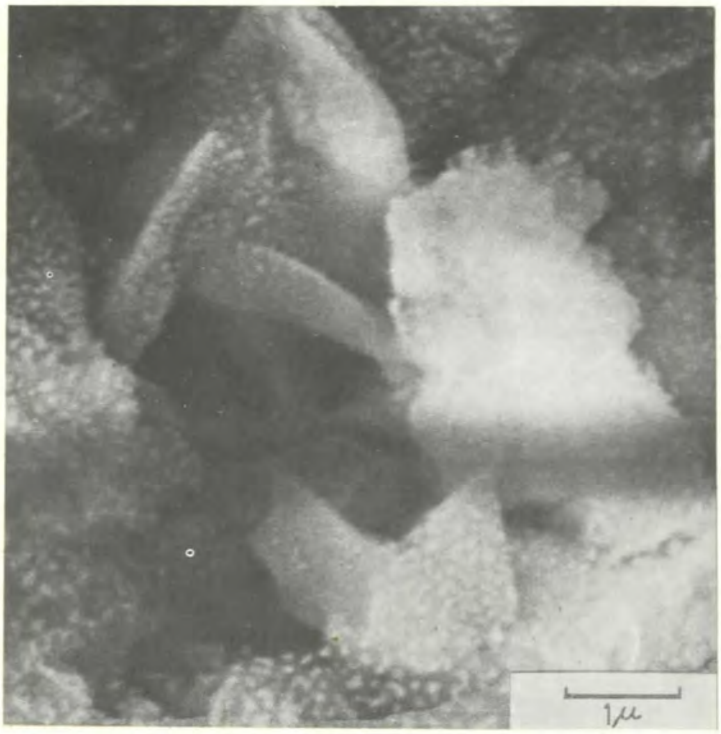


Figure 8.9 Scanning electron photomicrograph showing apatite crystals in phosphorite pellet.

Figure 8.10 Scanning electron photomicrograph showing structure of pellet fracture surface.

more than 18 per cent P_2O_5 (approximately 50 per cent apatite). In only 38,1 per cent of the phosphatic rocks of this study did the P_2O_5 concentration exceed 18 per cent. The highest P_2O_5 value is 27,2 per cent. The average value for the basal bed is 14,9 per cent P_2O_5. Parker and Siesser (1972) report an average of 15 per cent P_2O_5 in the continental margin phosphate rocks.

1. <u>Miocene Phosphorite - Saldanha Formation</u>

The rock is brown with a surface that is polished and undulating or pitted by differential erosion. Numerous burrows of marine animals show that the rock still displays its original surface. Essentially the rock of this horizon consists of fine sand size quartz embedded in a matrix of finely divided argillaceous and organic material that is collophane cemented. Frankel (1943) has identified the argillaceous material as montmorillonite. The detrital quartz fraction constitutes from 20 to 95 per cent of the rock. It is a mixed grain fraction (in the sense of varying textures of the quartz grains), generally poorly sorted, and frequently shows considerable iron staining. Fe_2O_3 percentage ranges from 0,3 to 6,5.

Near the southwestern corner of the New Varswater Quarry the basal bed is markedly conglomeratic, and in places brecciated (Figures 8.12 and 8.13). Intraformational conglomerates frequently occupy erosional channels (scour and fill) (Figure 4.4). Fractures in the larger non-phosphatic sandstone clasts are penetrated by collophane. The clasts are usually rounded and sometimes bent while in a semilithified state, and some have been almost completely phosphate mineralised. Such conglomerates indicate that lithification was disrupted by erosion and redeposition (Blatt <u>et al</u> 1972). Very similar conglomeratic phosphorites have been described from the continental shelf (Parker 1971, 1975).

Most of the onshore Saldanha Formation phosphatic sandstones may be described as medium to fine grained collophane packstones. In the majority of cases the collophane has undoubtedly originated by direct

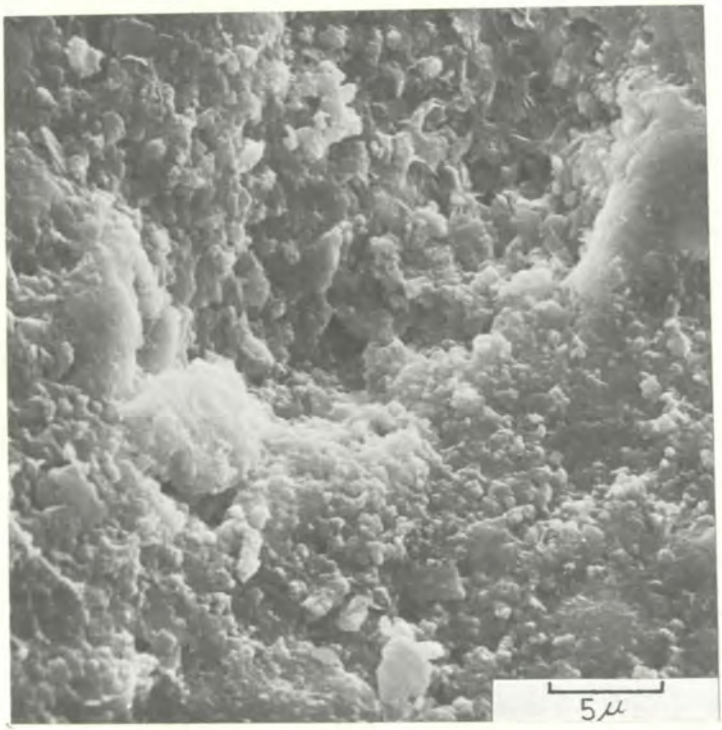


Figure 8.11 Scanning electron photomicrograph showing typical structure of pellet fracture surface.

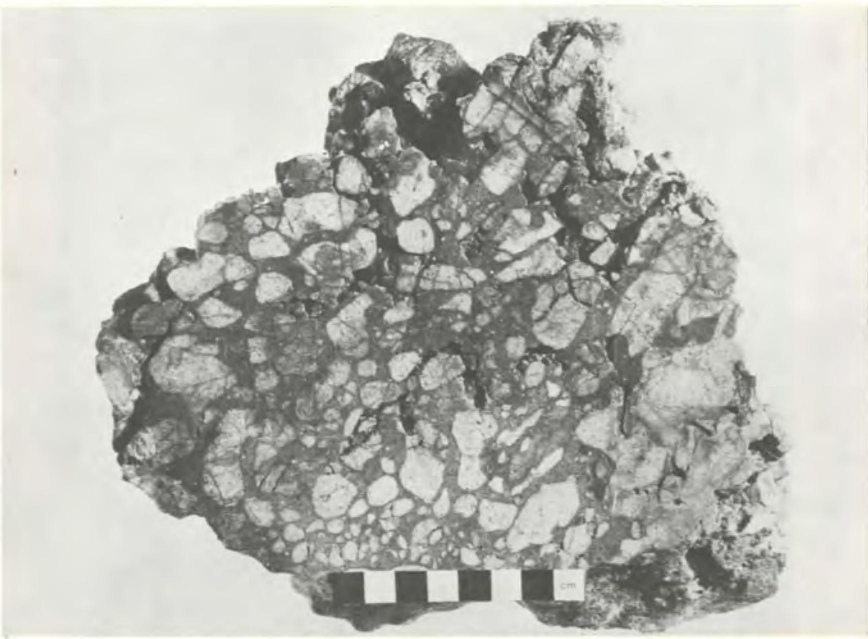


Figure 8.12 Intraformational conglomerate typical of parts of the Miocene phosphatic sandstone.

Figure 8.13 Preferential phosphate mineralisation of Miocene sediments.

Figure 8.14 Concentric structures in phosphatic sandstone produced either by algae or diagenesis.

precipitation of phosphate within the pore-spaces of a quartz sand from phosphate-rich waters. But in a few cases precipitation of the phosphate has possibly been induced by algae present in the sediment (Tankard 1974b). Figure 8.14 shows a structure that has developed in the form of concentric shells of different mineral composition. In Figure 8.15 which is an enlargement of part of the previous figure, it is shown that this structure has developed independently of the basic sediment texture, and is possibly diagenetic. The concentric shells reflect zones of enrichment by collophane and sometimes iron oxide. Iron oxide (? goethite) would appear to be ubiquitous in these phosphate rocks, but it could be younger than the phosphate component in many cases.

Heavy minerals, mainly ilmenite, are present only in trace amounts. Bone fragments, which are common, are always completely phosphatised. Shell debris is much less common but, nevertheless, still observed. A single shark tooth was found still embedded in reworked phosphatic sandstone, and in a very coarse-grained sample shark teeth were abundant.

Packing of sediments is easiest described by the packing proximity which is the ratio of the number of grain-to-grain contacts to the total number of grains observed in a thin-section traverse (Kahn 1956). The packing proximity for the phosphatic sandstone is occasionally as high as 50 per cent (i.e. 5 contacts along a traverse of 10 grains). But an average value of 10-20 per cent is more diagnostic. This low packing proximity is a result of the high mud content, and in some cases due to apatite crystal growth pushing the quartz grains aside. The quartz fraction is a mixed grain fraction, the largest grain-size being about 0,4mm and well rounded. The quartz fraction is largely negatively skewed and poorly sorted with a mean grain-size of about 3 ϕ. Many of these well rounded grains have been broken in the last cycle of erosion, the rounded aspect having been inherited from a previous cycle. Corrosion of quartz grains has also occurred. An interesting feature of the phosphate rock is the frequency of pellets of phosphorite with much silt-size quartz in the matrix (Figure 8.16) and which constitute 3-5 per cent of the sandstone. Some of the

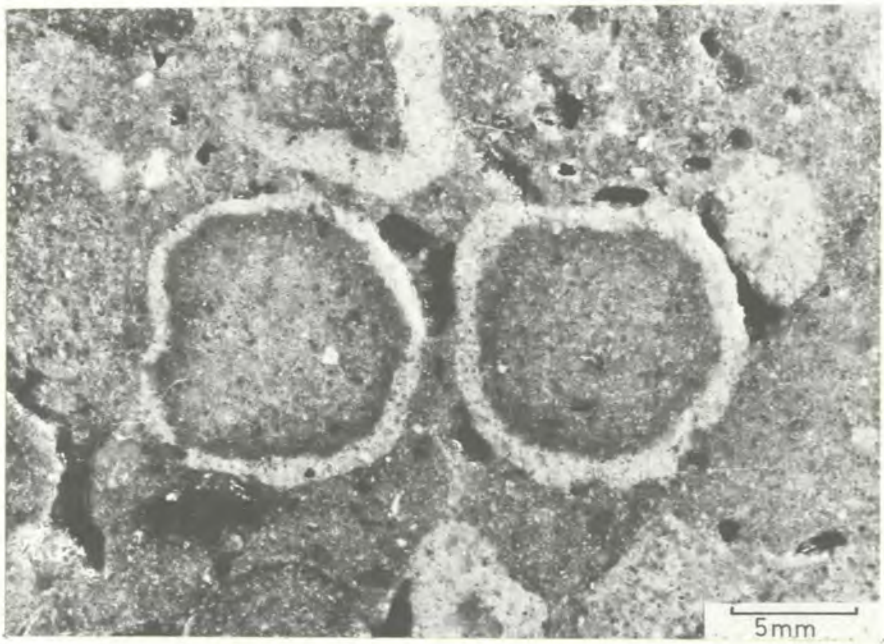


Figure 8.15 Enlarged section of previous figure showing that the concentric structure has developed independently of sediment texture.

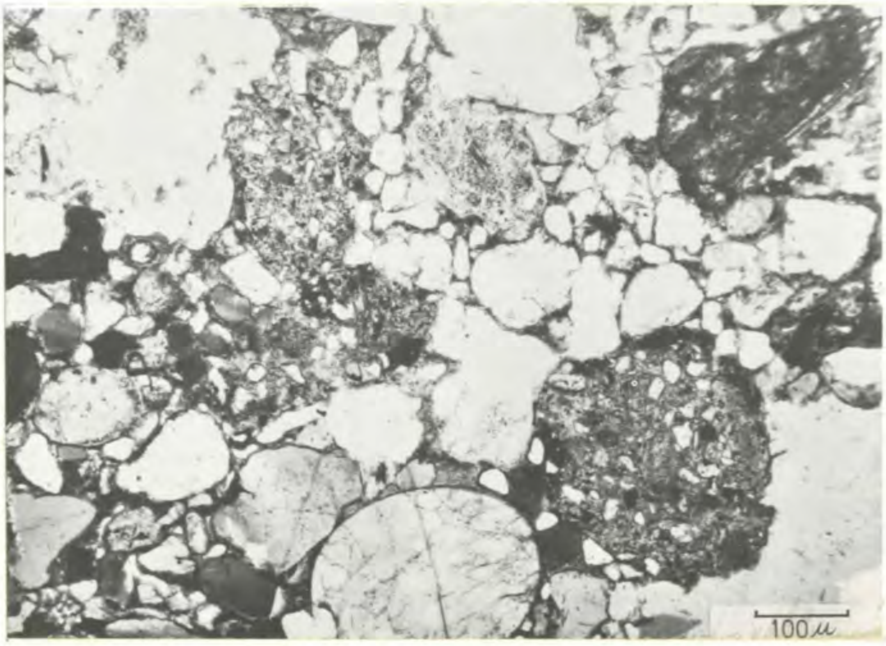


Figure 8.16 Phosphorite pellets containing quartz grains in phosphatic sandstone of basal bed.

included phosphorite pellets are, like the quartz, corroded. Sometimes pore-spaces are lined with drusy francolite. Generally the matrix is yellow brown under plane-polarized light. It consists often of a rim of anisotropic francolite grown from the walls of the voids, the remaining space being filled with isotropic collophane. Precipitation of the apatite has taken place in the pore-spaces of the sediment, and Figure 8.17 illustrates typical drusy francolite. Calcite was seldom observed.

The wave-generated beach cobbles from the Gravel Member of the Varswater Formation were derived by marine erosion during the Pliocene transgression. Thin-section examination shows that they are identical to their parent phosphatic sandstone of the Saldanha Formation in every respect. They consist of about 90 per cent quartz with a packing proximity 10–30 per cent. The matrix, like the basal bed, is composed of isotropic or slightly anisotropic collophane.

The Hoedjiespunt exposure is a microsphorite sensu stricto. It contains 3,6 per cent SiO_2, 1,8 per cent Al_2O_3, 0,3 per cent Fe_2O_3, and 34–36 per cent P_2O_5 (Table 8.2). The iron oxide content of the Hoedjiespunt microsphorite is considerably lower than that of the Langeberg basal bed. Thin-section examination and X-ray diffraction data confirm that little of the SiO_2 is present as free quartz, in fact less than 1 per cent. Even the clay mineral that accounts for the Al_2O_3 and the balance of the SiO_2 is present in small amounts. The low fluorine content in sample 33 suggests that the apatite mineral is in part dahllite, but sample 34 has a fluorine content more characteristic of francolite. Both dahllite and francolite are members of an isomorphous series. Like the francolite of Langeberg, it is isotropic to very slightly anisotropic under crossed nicols.

The microsphorite is preserved in a bedrock depression within the granite platform. The lower part is bedded while the upper part contains more phosphatised micrococuina, wave-generated granite boulders, and reworked and rounded fragments of microsphorite from the lower unit.

Shell debris, bryozoan remains and foraminifera tests, including a

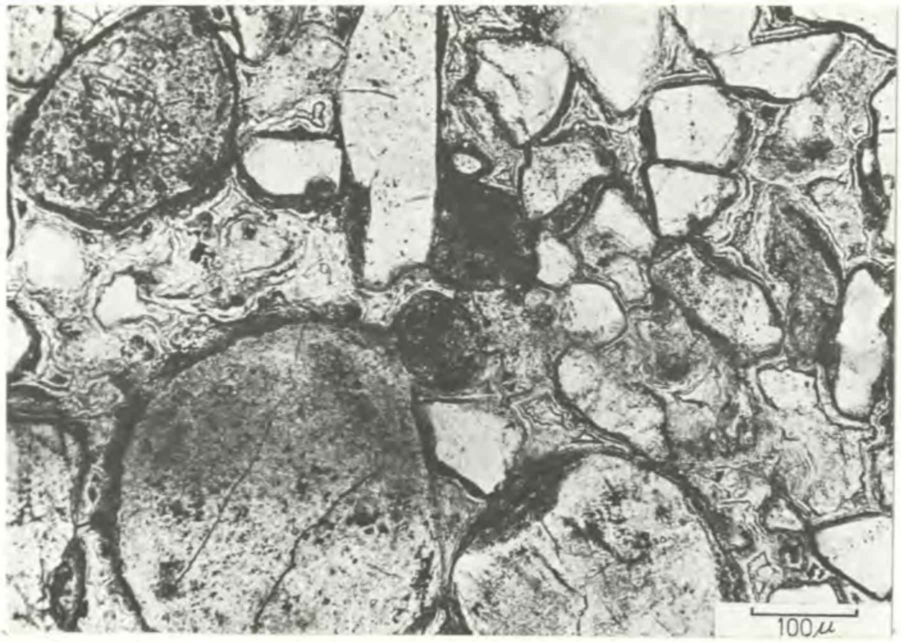


Figure 8.17 Drusy francolite grown from walls of the voids in the phosphatic sandstone.

Figure 8.18 Hoedjiespunt microsphorite showing foraminifera and shell debris.

planktonic element, are set in microsphorite. Figure 8.18 shows typical phosphatised microcoquina. Reworking of the earlier lithified microsphorite already mentioned, may take the form of cleavage flakes of phosphorite which sometimes resemble micro-mud-flake breccia, (Figure 8.19). It certainly is indicative of rapid lithification and reworking, while further collophane precipitation is shown in Figure 8.19 as infillings of cavities in the collophane matrix. Oolitic texture (Figure 8.20) is common in the Hoedjiespunt microsphorite, and is identical to oolitic phosphorite from the Meade Peak Mine, Utah (Trueman 1971, plate 7A). The wave-generated granite boulders at Hoedjiespunt, the reworked microsphorite and microsphorite-flake-breccia, the bioclastic material, and the oolitic textures all suggest a very shallow depositional environment. Moulds of Fissurella and Patella show the close proximity to the intertidal zone.

The phosphatic sandstones of Miocene age at Ysterplaat have also been examined. In thin-section the coarser quartz grains of the lower unit are seen to be rounded to well-rounded, while the finer grains are subround. Fracturing of the grains is common. The closely packed detrital quartz is set in a microsphorite matrix. Anisotropic francolite has formed about the quartz grains, while the remaining voids are filled with isotropic collophane. Occasional pellets were observed ($< 1\%$), together with a few phosphatised foraminiferal tests and phosphatised bone fragments. Two phosphate determinations show 8,0 per cent and 9,2 per cent P_2O_5 respectively (analyst: Chemfos Limited).

2. Phosphate Rock of the Varswater Formation

Precipitation of isotropic collophane within the pore-spaces of the quartzose sands of the Varswater Formation has been induced by local concentrations of organic matter. This has resulted in the development of thin lenses and concretions of phosphatic sandstone. Figure 8.21 shows how a tortoise bone has acted as a nucleus for phosphate deposition which has followed the contours of the bone. Although the bone is completely phosphatised, the maximum phosphate deposition outside the bone is slightly separated from the bone. The following

Figure 8.21 Tortoise bone that has acted as a nucleating centre for phosphate precipitation.

Figure 8.22 Typical phosphatic sandstone lens structure from the Varswater Formation.

Figure 8.23. Collophane mudstone.

TABLE 8.1

Comparison of P_2O_5 Percentage in Phosphate Rock with that of surrounding Phosphoritic Sands, Pelletal Phosphorite Member.

Borehole	Depth	P_2O_5 %	
		Rock	Sand
M6-100-N6	15,0m	13,6	12,1
"	16,0	14,2	3,9
"	17,2	12,0	3,0
"	23,6	5,2	3,1
"	24,6	11,8	1,3
S12	41,7	18,5	16,8
"	42,8	11,6; 19,6	8,7
"	43,8	10,7	3,1
Q12	36,4	21,0	13,6
"	43,8	13,3	10,4
W 1	43,8	13,8	11,1
"	44,9	12,3	12,0
W 5	38,5	6,4	1,8
"	43,8	11,0	4,6

Analyst: Chemfos Limited

C. Mineralization

Petrographic evidence suggests that at least three distinct periods of post-depositional phosphatisation have occurred. The conglomeratic phosphate rock or packstone of the Langeberg basal bed shows two periods of mineralisation. The first phase is represented by fragments of an older phosphate rock as well as occasional phosphorite pellets. These older components were reworked, and the new sediment lithified in the second phase of phosphatisation. In the basal bed phosphate precipitation has resulted in a phosphate packstone only one metre thick. Phosphate mineralisation must have taken place fairly rapidly after deposition, and must have been nearly contemporaneous with sedimentation, because it depended upon the close proximity of upwelling of phosphate-rich water (to be discussed later). Phosphatisation appears to be nearly uniform through much of the rock of the basal bed, although contact with phosphate rich water has increased the degree of phosphatisation in a thin layer at the surface. The overlying phosphatic packstone cobbles of the Gravel Member are identical to the basal bed and have been derived by erosion of that bed which suggests that the phosphatisation must have preceded the last phase of erosion. On Langeberg phosphate mineralisation has taken place within the pore-spaces of a Miocene quartz sediment, mineralising the argillaceous material already there. Organic matter within the argillaceous material has probably behaved as a catalyst. On Langeberg there is little evidence that this mineralisation has proceeded via a lime replacement mechanism and it is certainly not a phosphatised calcrete as suggested by Wolff et al (1973). I have suggested that locally the phosphate precipitation could possibly have been induced by algae. If this were so the algae would have behaved in a similar way to the other organic matter. On Langeberg the phosphorite pellets of the Varswater Formation have a very similar aspect to that of the matrix of the basal bed, although some do show concentric layering. In these rare cases the concentric structure is probably the result of apatite growth about an initial pellet. There is no evidence to suggest growth of the pellets about nuclei, although silt-size quartz inclusions give the impression of nucleation. Thin-section analyses show that the quartz was present in the matrix of the original phosphorite as was the organic material. Summerhayes (1970) argues that if the francolite in the original phosphatic rock has grown from the walls of the voids, these layers may act as lines of structural weakness. Disintegration along such lines leads to the formation of pellets of collophane that contain impurities but have a clear collophane margin.

On Langeberg the third phase of phosphate mineralisation is marked by the appearance of phosphatic sandstone lenses and concretions in the

Varswater Formation. The estuarine sediments are capped by a thin phosphatic sandstone horizon. In the Pelletal Phosphorite Member the phosphatic sandstone lenses and concretions include pelletal phosphorite in the same abundance as the surrounding sediments. In this same member there are phosphatised foraminiferal tests, indicating a lime replacement mechanism while the presence of phosphatised cartilage or young bone (Figure 8.7) shows just how rapid phosphate mineralisation may be. Shell material, shark teeth, bone fragments, echinoid spines and foraminiferal tests are all well phosphatised. On Hoedjiespunt the bedded microsphorite, by coincidence, also demonstrates three phases of phosphate mineralsation. The first two phases are demonstrated by phosphatisation of a microsphorite-flake breccia (Figure 8.19), and in a third phase of mineralisation fractures within this phosphatised wackestone have been the sites of phosphate deposition. Foraminiferal tests and microcoquina have all been completely phosphatised.

Within the Varswater Formation deposition appears to be confined to deposition of fine sand material. Frequent erosional episodes have resulted in concentration of shark teeth and mollusc shells into definite horizons. Widespread reworking is attested by the broken detrital grains and shell debris and worn foraminiferal tests. All evidence suggests low rates of sedimentation.

D. Chemical Composition of the Phosphorite

The apatite mineral of all the pelletal phosphorites and the Langeberg basal bed is francolite, while the apatite of the Hoedjiespunt microsphorite ranges through dahllite to francolite. Francolite is the name applied to an apatite containing appreciable CO_2 and more than 1 per cent fluorine, whereas the name dahllite has been applied to apatite containing abundant CO_2 but less than 1 per cent fluorine (McConnell 1938).

Table 8.2 shows the major-element geochemistry for the pelletal phosphoritic sands, pelletal phosphorite concentrates, phosphatic sandstone, phosphatic aeolianite from Darling, phosphatised bone, and microsphorite. In the next chapter these values have been re-calculated

Figure 8.19 Microsphorite flake-breccia.

Figure 8.20 Oolitic texture in Hoedjiespunt microsphorite.

hypothesis of origin of this type of texture is suggested:

(i) the bone acts as a nucleating centre for precipitation of phosphate to form a nodule;

(ii) the apatite of the bone, which is dahllite (Carlström 1955), is not in equilibrium with the surrounding francolite in the nodule and phosphate is withdrawn from the surrounding matrix in converting the bone apatite to francolite. Tankard (1974c) has shown how the original bone apatite takes up more fluorine from the phosphate-rich environment.

A typical phosphatic sandstone from the Varswater Formation is shown in Figure 8.22. The detrital components of the phosphatic pack-stones are always similar to that of the surrounding sediment, and they include phosphorite pellets. Cementation has taken place by precipitation of francolite about the grains to give a radially disposed cryptocrystalline francolite with the pore-spaces filled with clear collophane or mud. (The collophane disseminated through the matrix of the phosphatic sandstone can truly be referred to as microsphorite, as can the collophane component of the pelletal phosphorite already discussed). Only occasionally is drusy quartz found. In some of the packstones a crudely graded bedding has been observed. The colour of these phosphatic packstones ranges from pale brown to black. Samples from borehole S12 (42,8m) were analysed. The brown variety contained 11,6 per cent P_2O_5 and the black variety 19,6 per cent P_2O_5. Where phosphate rock has formed as lenses or concretions within the Pelletal Phosphorite Member, P_2O_5 concentration is always substantially higher than that of the surrounding phosphoritic sands (Table 8.1). Towards the base of the Varswater Formation a collophane mudstone is frequently encountered (Figure 8.23). It contains 5 to 10 per cent detrital quartz set in a fine phosphatised argillaceous matrix.

TABLE 8.2

MAJOR ELEMENT GEOCHEMISTRY

Varswater Formation																	
Pelletal Phosphorite Concentrates																	
Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
SiO_2																	4,13
Al_2O_3																	1,73
Fe_2O_3	1,74	1,87	1,79	1,83	1,24	1,56	1,73	1,34	1,48	2,15	1,72	1,61	1,45	1,67	3,15	3,43	1,77
CaO	46,76	39,97	39,23	44,58	49,01	43,69	47,81	49,38	47,53	47,67	47,11	47,95	48,09	48,51	49,07	47,11	46,97
MgO	1,64	5,83	6,67	2,83	0,04	3,49	0,50	0,16	0,91	1,01	0,30	0,30	0,34	0,20	0,20	0,81	0,28
Na_2O	0,70	0,75	0,70	0,69	0,73	0,76	0,71	0,74	0,76	0,77	0,76	0,76	0,77	0,77	0,75	0,77	0,68
K_2O																	0,26
H_2O+																	2,43
H_2O-																	0,44
P_2O_5	34,15	34,12	33,98	34,66	35,55	33,87	33,72	34,55	34,51	33,19	32,67	33,88	34,06	34,35	34,01	32,27	33,16
CO_2	4,32	4,04	4,19	4,05	4,02	4,38	4,62	4,03	3,96	4,73	4,91	4,65	4,37	3,95	3,85	3,83	3,81
F	3,43	3,40	3,27	3,87	3,33	3,30	3,67	3,92	4,37	4,00	3,30	3,38	3,33	2,85	2,95	2,92	3,44
SO_3	0,90	0,78	0,80	0,69	0,92	0,83	0,78	1,05	0,73	0,80	0,90	0,84	0,91	0,96	0,96	1,62	0,17
Total	93,64	90,76	90,63	93,20	94,84	91,88	93,56	95,17	94,25	94,32	91,67	93,37	93,32	93,26	94,94	92,76	99,27

	Varswater Formation										Saldanha Formation			Phosp. Aeo-lianite	Bone	Saldanha Formation	
	Phos. sst.	Pelletal Phosph. Sands							Phosph. sst.		Phosph. sst.					Microsphorite	
Sample	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34
SiO_2		10,30	44,80	56,30	50,24	72,26	67,24	80,88	54,56	20,86	59,92	35,16	63,08	29,78	1,72	3,60	
Al_2O_3		1,25	1,72	1,76	1,72	3,27	1,93	2,29	1,11	0,62	0,94	2,70	0,66	1,38	0,60	1,83	
Fe_2O_3	0,80	1,54	1,29	1,16	1,27	2,19	0,93	0,96	0,71	0,51	0,63	2,02	0,36	4,23	0,16	0,27	
CaO	23,97	44,36	25,74	19,57	22,99	9,59	13,91	6,62	21,65	40,55	19,35	29,72	18,06	31,63	50,36	46,04	49,59
MgO	0,30	2,54	2,02	1,57	1,76	0,75	0,96	0,55	0,76	0,88	0,99	1,04	0,93	0,75	1,16	2,32	0,47
Na_2O	0,45	0,61	0,55	0,46	0,56	0,21	0,32	0,28	0,37	0,36	0,22	0,36	0,17	0,26	0,44	1,27	1,00
K_2O		0,41	0,79	0,81	0,76	0,57	0,87	0,88	0,29	0,09	0,32	0,18	0,12	0,01	0,01	0,06	0,05
H_2O^+																	2,24
H_2O^-		0,77	0,70	0,61	0,71	0,61	0,42	0,33	1,42	1,48	0,72	0,91	0,45	1,46	1,79	2,55	2,54
P_2O_5	18,28	31,19	17,79	14,02	16,24	7,54	10,50	5,37	16,00	28,47	14,52	21,93	13,59	23,47	36,19	33,97	35,39
CO_2	1,97	3,02	1,87	1,59	1,77	0,83	1,06	0,53	1,13	1,92	0,75	1,84	0,85	2,02	2,50	3,28	4,08
F	1,50	2,00	1,50	1,30	1,40	0,72	1,10	0,43	1,20	2,00	1,30	1,60	1,20	1,70	1,50	0,57	1,59
SO_3	0,39	0,34	0,24	0,18	0,20	0,06	0,16	0,14	0,06	0,06	0,16	0,22	0,08	0,06	0,12	0,44	0,80
Total	47,66	98,33	99,09	99,39	99,69	98,62	99,45	99,31	99,28	97,82	99,87	97,75	99,58	96,77	96,59	96,35	97,75

(Note: Sample locations as for Table 9.1)

Analysts: 1–18, 34 – General Superintendence Co.
19–33 – Anglo American Research Labs.

to an impurity free basis so as to characterise the apatite mineral. Sample 33 has only 3,6 per cent SiO_2 which is partially present as free quartz, or bound up with the Al_2O_3 and Fe_2O_3, as a clay mineral. The low fluorine content in this sample (0,57 per cent) defines the major apatite component as dahllite (less than 1 per cent F). The inorganic component of bone is dahllite, but sample 32 shows enrichment in fluorine due to post-depositional diagenesis. In this respect the sample 34 microsphorite with a similar concentration of fluorine (1,59 per cent) has been similarly affected. The bone sample, both microsphorite samples and the Pleistocene aeolianite sample are all characterised by low amounts of K_2O. (The geochemistry of the apatite will be discussed fully in a later chapter). It is worth noting how closely the major element geochemistry of the pelletal phosphorite approaches that of the Saldanha Formation phosphatic sandstone and microsphorite. This indicates a genetic relationship and strengthens the argument that the pellets are detrital particles derived by erosion of these older rocks.

IV. ALUMINIUM PHOSPHATES

The aluminium phosphates will be discussed only briefly as they are not directly related to the marine phosphorites, although both are related to the upwelling of phosphorous-rich waters. These aluminium phosphates have been described in some detail by du Toit (1917), Hutchinson (1950) and Tankard (1974b).

The aluminium phosphate distribution map (Figure 8.24) shows how outcrops of the phosphate rock occur sporadically about granite hills north and south of Saldanha Bay. They are not confined to any particular altitude. The degree of phosphatisation is very variable over the area and even varies within any particular hand specimen. On Konstabelkop the insular phosphate rock is hard and of moderate yellowish brown (10YR 5/4) coloration, while the phosphatised clay of slickensides is grey - brown (5YR 3/2). North of Saldanha Bay the insular phosphate rock has a green hue. On Konstabelkop an exploration pit in a (?) marine terrace at 152m a.s.l. shows rounded boulders lying on

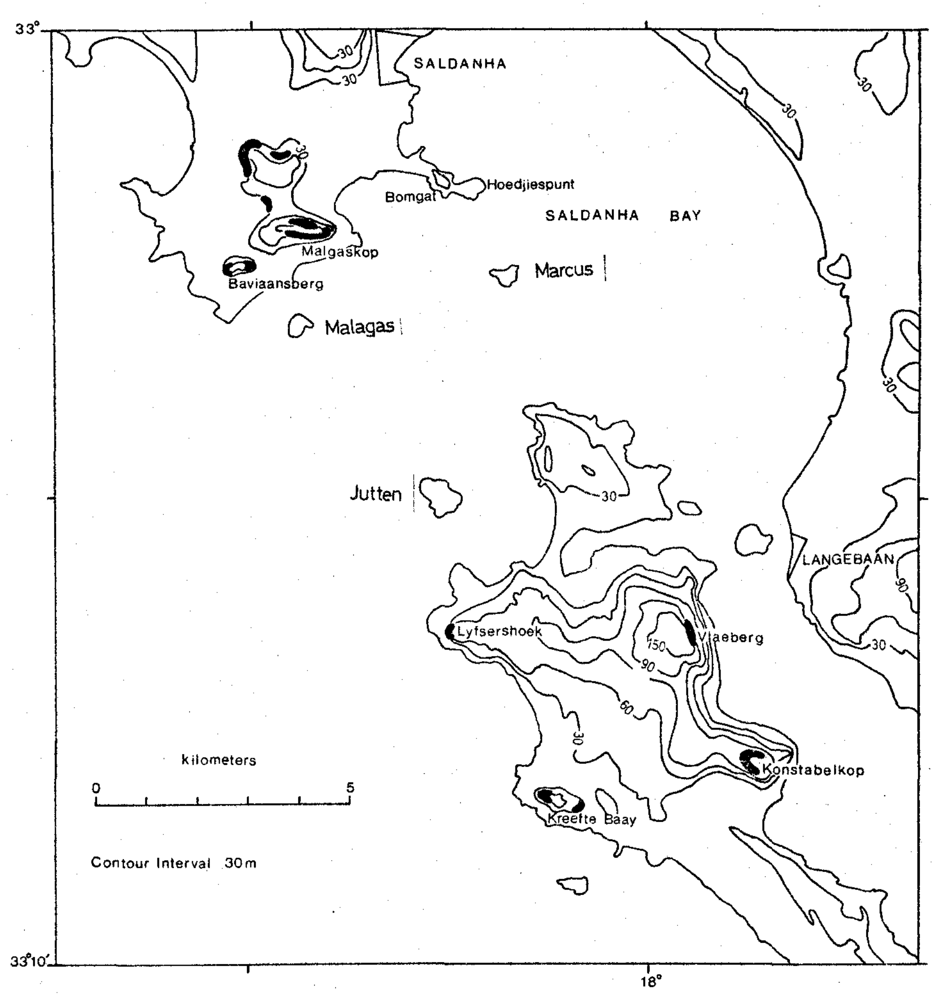


Figure 8.24 Distribution map for aluminium phosphate rocks.

weathered porphyry. Both the underlying porphyry and the rounded boulders have been phosphatised, but in the case of the boulders the degree of phosphatisation is greatest near the surface, suggesting post-depositional phosphatisation and demonstrating that the phosphatising solutions were exotic.

In the main quarry on Konstabelkop the highest concentration of P_2O_5 is close to the surface, the concentration decreasing irregularly with depth. The maximum depth ranges from 1,5m to 12 m, the base being highly irregular. At the surface weathering has produced lateritic phosphate.

A feature of the Konstabelkop insular phosphatic rock is the existence of numerous slickensides along which phosphatisation is extensive. In thin-section it is seen that the ground-mass is a fine clay material which has been phosphatised. The phosphatised parts of the clay are isotropic while slight anisotropism is evident at the contact of the phosphatised and unphosphatised clay. Larger quartz grains are generally fractured and have fissures filled with the same isotropic phosphate mineral. The faulting appears to have followed the phosphate mineralisation since the texture of the rock shows the drag effect of movement along the fault plane. The clay minerals must originally have developed along joint planes in the granite. A detailed thin-section study was carried out and reported by du Toit (1917).

The altitude of the phosphate deposits is very variable (as shown in Figure 8.24) and cannot be related to any particular sea level as Visser and Schoch (1973) maintain. On Baviaansberg within the grounds of the Naval Academy (SAS SALDANHA) the phosphates occur at 45m a.s.l. and have imparted a green colouration to the granites. On Malgaskop the phosphates range from 40 to 100m a.s.l. The lower limits of these phosphate deposits are not significant as they have originated by percolating solutions.

Partial chemical analyses of the phosphate rock are listed in Table 8.3.

TABLE 8.3

Partial chemical analyses of the aluminium phosphate rock

	Phosphatized porphyry	Phosphatized porphyry	Yellow-sandy regolith	Phosphatized limestone
	%	%	%	%
SiO_2	59,20	65,40	10,72	10,70
Al_2O_3	8,32	6,78	31,11	5,66
Fe_2O_3	4,02	5,29	7,14	0,43
CaO	0,10	-	0,42	40,90
K_2O	0,60	0,48	2,41	0,60
P_2O_5	12,14	11,68	18,31	0,92
CO_2	-	-	-	31,40
F	Trace	Trace	Trace	Trace

Analyst: A.E. & C.I. Limited. Results supplied by Mr Botha of the Konstabelkop Mine.

Several significant features are apparent:

(i) the Fe_2O_3 content is high on the granite areas, but insignificant on the phosphatised limestone;

(ii) Al_2O_3 is highest in the yellow sandy regolith (31,11 per cent). Also in this horizon the Fe_2O_3 value is highest (7,14 per cent).

(iii) Fluorine is always present in only trace quantities.

The yellow sandy regolith has resulted from supergene alteration of the underlying phosphatised granite to produce the superficial lateritic zone. Under normal soil forming processes weak solutions from the leaching of the rock evaporate and the least soluble components precipitate first. These include hydroxides of iron and aluminium, silica and carbonates. Further solution leaves behind the aluminium and iron hydroxides in an insoluble state.

X-ray diffraction analyses were carried out on samples from the phosphatised granite and from the enriched layers associated with slickensides. The only phosphate mineral identified from these two types was variscite. The strongest measured X-ray lines are: 5,37 (6) - 4,79 (4) - 4,27 (10) - 3,90 (3) - 3,32 (6) - 3,05 (4) - 2,86 (3) - 2,70 (6) - 2,46 (2).

Du Toit (1917) suggested that the average composition of the Konstabelkop phosphate is close to that of barrandite while minute yellowish crystals may be referred to wavellite or variscite. In the samples examined the X-ray lines characteristic of barrandite and wavellite were absent. Barrandite is an intermediate mineral in the variscite ($AlPO_4.2H_2O$) - strengite ($FePO_4.2H_2O$) group, i.e. it is an aluminium iron phosphate mineral. The high iron content shown in Table 8.3 also suggests that barrandite could be encountered. None of the other phosphate minerals identified by Harrington et al (1966) and Altschuler et al (1956) from the United States have been encountered, fiz. metavariscite, crandallite, wavellite and millisite.

CHAPTER 9

GEOCHEMISTRY OF THE APATITE IN THE PHOSPHORITE

I. INTRODUCTION

The Pliocene Varswater Formation is characterised by economic concentrations of pelletal phosphorite. The pellets are an allogenic component of this formation, having been derived by erosion of the underlying Miocene basal bed of the Saldanha Formation where the authigenic apatite occurs in the microsphorite matrix. This chapter summarises the chemical composition of the pelletal phosphorite of the Varswater Formation in the Langebaanweg area (the farm Langeberg) and the Saldanha area (the farm Witteklip). Also discussed is the authigenic apatite of the basal bed on Langeberg and the microsphorite bed at Hoedjiespunt. The results presented in this chapter have largely been published (Tankard 1974c).

The microcrystalline sedimentary apatites differ in composition from fluorapatite, $Ca_{10}(PO_4)_6F_2$, because of extensive substitution of carbonate and fluorine for phosphate and of other metals for calcium (Smith & Lehr 1966). McConnell (1938) subdivides the carbonate-apatite according to fluorine content. Dahllite, a carbonate-hydroxyapatite, has less than 1 per cent fluorine, while francolite has more than 1 per cent fluorine. Isomorphous substitution is common among the natural apatites. In its substitution for phosphate the carbonate ion is accompanied by fluorine. This has the effect not only of maintaining charge balance, but also of retaining the original tetrahedral co-ordination of the PO_4 group (Gulbrandsen et al 1966). Although it is now generally accepted that the carbonate ion substitutes for the phosphate ion, the high CO_2 content of phosphorites has also been attributed to "amorphous" calcite present as an impurity and to carbonate ions adsorbed to the crystallite surfaces. This aspect will be discussed further. A voluminous literature discusses carbonate-apatite, the major component of phosphorite (for instance: Altschuler et al 1953; Gulbrandsen 1969; Lehr et al 1967; McClellan & Lehr 1969; McConnell 1952;

Rooney & Kerr 1967; Smith & Lehr 1966; Tooms et al 1969; Whippo & Murowchick 1967).

II. METHODS

A. X-ray Methods

Powder diffraction data were obtained using a General Electric XRD-3 diffractometer. Iron-filtered CoK_α radiation was used at 25 kv/10 ma. Samples were scanned at $0{,}5^\circ$ 2θ per minute with a chart speed of 150 cm/hour. Quartz was used as a standard.

The lattice constants a and c were calculated from the measured (300) and (002) reflections. The interplaner spacings (d_{hkl}) are related to the cell parameters, a and c, by the following formula:

$$d_{hkl} = \frac{1}{\sqrt{\frac{4(h^2+hk+l^2)}{3a^2} + \frac{l^2}{c^2}}}$$

B. Infrared Spectral Analysis

Infrared spectra were obtained using a KBr disc technique on a double-beam spectrophotometer.

C. Chemical Methods

Analyses were made of rock samples and of pelletal phosphorite. Pelletal phosphorite samples 1 to 17 were concentrated by heavy liquid separation, and leached of free carbonates. The heavy liquid separation resulted in concentrates of pelletal phosphorite in excess of 90 per cent in all cases. Pelletal phosphorite samples 19 to 25 were not concentrated, but analysed as phosphoritic-quartzose sands. Analytical methods used were as follows:

	Samples 1-18, 34	Samples 19-33
CaO; MgO	Volumetric (MgO in sample 34 A.A.S.)	Spectrophotometric
P_2O_5	Volumetric	Volumetric
Na_2O; K_2O	Flame photometric	Spectrophotometric
SiO_2; SO_3; CO_2; H_2O^+; H_2O^-	Gravimetric	Gravimetric
F	Spectrophotometric (distillation)	Fluoride electrode

The majority of elements in samples 19 to 33 were checked by X-ray fluorescence analyses.

III. CHEMICAL CHARACTERISATION

The chemical composition of the carbonate-apatite can be approximated by the content of CaO, P_2O_5, CO_2 and F, while Na_2O, MgO, and SO_3 are minor constituents. The carbonate-apatite differs considerably in composition from pure fluorapatite in that a carbonate ion replaces a phosphate ion in the lattice, while the vacant oxygen site may be occupied by fluorine (Smith & Lehr 1966). The results of chemical analyses of pelletal phosphorite and rock phosphorite are shown in Table 9.1, while the range of constituents is shown in Table 9.2.

TABLE 9.1

CHEMICAL COMPOSITION

Composition wt. % (recalculated to impurity free basis)

No.	CaO	MgO	Na_2O	P_2O_5	CO_2	SO_3	F	F/P_2O_5	CaO/P_2O_5
1	51,69	1,81	0,77	37,75	4,78	0,99	3,79	0,10	1,37
2	45,70	6,67	0,86	39,01	4,62	0,89	3,89	0,10	1,17
3	44,85	7,62	0,80	38,85	4,79	0,91	3,74	0,10	1,15
4	49,68	3,15	0,77	38,62	4,51	0,77	4,31	0,11	1,29
5	53,16	0,04	0,79	38,56	4,36	1,00	3,61	0,09	1,38
6	49,13	3,92	0,85	38,09	4,93	0,93	3,71	0,10	1,29
7	52,96	0,55	0,79	37,35	5,14	0,86	4,07	0,11	1,42
8	53,57	0,17	0,80	37,48	4,37	1,14	4,25	0,11	1,43
9	52,27	1,00	0,84	37,95	4,36	0,80	4,81	0,13	1,38
10	52,67	1,12	0,85	36,68	5,23	0,88	4,42	0,12	1,44
11	53,20	0,34	0,86	36,89	5,54	1,02	3,73	0,10	1,44
12	53,08	0,33	0,84	37,50	5,15	0,93	3,74	0,10	1,42
13	53,16	0,38	0,85	37,65	4,83	1,01	3,68	0,10	1,41
14	53,67	0,22	0,85	38,00	4,37	1,06	3,15	0,08	1,41
15	54,19	0,22	0,83	37,56	4,25	1,06	3,26	0,09	1,44
16	53,47	0,92	0,87	36,63	4,35	1,84	3,31	0,09	1,46
17	53,95	0,32	0,78	38,09	4,38	0,19	3,95	0,10	1,42
18	51,84	0,65	0,97	39,54	4,26	0,84	3,24	0,08	1,31
19	53,30	3,05	0,73	37,48	3,63	0,41	2,40	0,06	1,42
20	52,44	4,12	1,12	36,25	3,81	0,49	3,06	0,08	1,45
21	51,31	4,12	1,21	36,76	4,17	0,47	3,41	0,09	1,41
22	51,86	3,97	1,26	36,63	3,99	0,45	3,16	0,09	1,42
23	49,43	3,86	1,08	38,87	4,28	0,31	3,71	0,10	1,27
24	50,48	3,48	1,16	38,11	3,85	0,58	3,99	0,10	1,32
25	48,18	4,00	2,04	39,08	3,86	1,02	3,13	0,08	1,23
26	53,25	1,87	0,91	39,35	2,78	0,15	2,95	0,07	1,35
27	55,23	1,20	0,49	38,79	2,62	0,08	2,72	0,07	1,42
28	52,67	2,69	0,60	39,52	2,04	0,44	3,54	0,09	1,33
29	53,03	1,86	0,64	39,13	3,28	0,39	2,86	0,07	1,36
30	52,55	2,71	0,49	39,54	2,47	0,23	3,49	0,09	1,33
31	53,46	1,27	0,44	39,67	3,41	0,10	2,87	0,07	1,35
32	54,84	1,26	0,48	39,41	2,72	0,13	1,63	0,04	1,39
33	52,38	2,64	1,44	38,65	3,73	0,50	0,65	0,02	1,36
34	52,11	0,49	1,06	37,19	4,29	0,84	1,67	0,04	1,40

Analysts: 1–18, 34 General Superintendence Co. Limited
19–33 Anglo American Research Laboratories

TABLE 9.1 (Sample Localities)

Samples 1-17, 19-25 Pelletal phosphorite from Varswater Formation; 18, 28-30 phosphatic sandstone (Miocene basal bed); 26, 27 phosphatic sandstone from Varswater Formation; 32 phosphatised bone; 33,34 microsphorite. Sample localities (borehole, depth) : (L) = Langeberg, (W) = Witteklip. 1. O20, 45m (L); 2. M14, 20m (L); 3. O12, 33m (L); 4. S1, ? (L); 5. V14, 19m (L); 6. R16, 44m (L); 7. P4, 13m (L); 8. X6, 11 m (L); 9. U4, 16m (L); 10. T14 36m (L); 11. S10, 37m (L); 12. P8,22m (L); 13. W'5, 32m (W); 14. V8, 20m (L); 15. W'1, 44m (W); 16. W'5, 40m (W); 17. R6, ? (L); 18. Quarry (L); 19. R6, ? (L); 20. Q12, 37m (L); 21. S12, 33m (L); 22. S12, 36m (L); 23. W'1, ? (W); 24. Q12, 31m (L); 25. W'4, ? (W); 26. Quarry (L); 27. Quarry (L); 28. Quarry (L); 29. Quarry (L); 30. Quarry (L); 31. Darling area; 32. Quarry (L); 33. Hoedjoespunt; 34. Hoedjiespunt.

TABLE 9.2

Range and Average Composition of the Pelletal Phosphorite (Samples 1-17)

Constituent	Range (wt.%)	Average (wt.%)
CaO	44,85 - 54,19	51,79
MgO	0,44 - 7,62	1,69
Na_2O	0,77 - 0,87	0,82
P_2O_5	36,63 - 39,01	37,80
CO_2	4,25 - 5,54	4,70
SO_3	0,19 - 1,84	0,96
F	3,15 - 4,81	3,85
F/P_2O_5	0,08 - 0,13	0,10
CaO/P_2O_5	1,15 - 1,46	1,37
Specific gravity	2,90 - 2,95	2,91

These analyses have all been recalculated to an impurity-free basis and reflect primarily the composition of the apatite. The SiO_2, Al_2O_3 and K_2O content cannot be assigned to the apatite structure. In the rock phosphorite the SiO_2 is present largely as free quartz with a minor amount assigned to a clay mineral. Minute quantities of quartz are an ubiquitous component of the pelletal phosphorite, but there is also abundant argillaceous material present. Frankel (1943) noted that the clay mineral present in the Langebaanweg phosphorite is essentially montmorillonite. Three analyses are here of interest, numbers 32, 33 and 34 (Table 9.1). In sample 32, a portion of phosphatised rhinoceros rib bone, the SiO_2 content was 1,72 per cent, Al_2O_3 0,60 per cent and $K_2O < 0{,}01$ per cent. The inorganic component of calcified bone tissue is carbonate-hydroxyapatite (Carlström 1955). Samples 33 and 34 were taken from the microsphorite lens at Hoedjiespunt. Again it is found that SiO_2, Al_2O_3 and K_2O content is low: 3,60 per cent, 1,83 per cent and $< 0{,}06$ per cent respectively.

Inspection of Table 9.1 shows that all samples contain an appreciable amount of carbonate. Sample 33 alone contains less than 1 per cent fluorine. By McConnell's (1938) definition samples 1 to 32, and 34 are francolite and sample 33 is dahllite. This division is also borne out by X-ray data. The range and average ratios of F/P_2O_5 and CaO/P_2O_5 (Table 9.2) are within the range of apatite composition. The ideal F/P_2O_5 ratio for a non-carbonate apatite is 0,089 (Rooney & Kerr, 1967). Parker and Siesser (1972) report a very high F/P_2O_5 ratio (0,143) for phosphorite samples from the South African continental shelf. They also report unusually high CO_2 content, average 5,7 per cent, for their continental shelf phosphorites.

The structural formulae for the apatite of the pelletal phosphorite and basal bed, and the method of formula calculation are shown in Table 9.3.

TABLE 9.3

Formula Calculations on Average of Analyses 10, 11, 12,32

	Wt.%	Mol. Ratios	Ionic Ratios	No. of Ions	Distr. of Ions	+ Charges	− Charges	Redistr. ions: 10 in Ca posns. 6 in P posns.
CaO	52,52	0,937	0,937	9,27	Ca=9,27	18,54		9,29
MgO	1,98	0,050	0,050	0,49	Mg=0,49	0,98		0,49
Na_2O	0,68	0,011	0,011	0,11	Na=0,22	0,22		0,22
					9,98			10,00
CO_2	3,01	0,068	0,137	1,35	C = 0,68	2,72		0,65
P_2O_5	39,43	0,278	1,389	13,75	P = 5,50	27,50		5,29
SO_3	0,48	0,006	0,018	0,18	S = 0,06	0,18		0,06
					6,24			6,00
F	3,28	0,173	0,173	1,71	F = 1,71		1,71	
	101,38							
−O≡F	1,38							
	100,00							
O					O = 24,29		48,58	
					26,00	50,14	50,29	

Calculated Structural Formulae

1. $(Ca_{9,31}Mg_{,42}Na_{,27})\ (C_{,25})(P_{4,91}C_{,73}S_{,11}O_{24,02})(F_{1,98})$
2. $(Ca_{9,29}Mg_{,49}Na_{,22})\ (C_{,16})(P_{5,29}C_{,49}S_{,06}O_{24,29})(F_{1,71})$
3. $(Ca_{9,52}Mg_{,13}Na_{,35})\ (C_{,23})(P_{4,98}C_{,69}S_{,10}(OH)_{1,34}O_{22,66})(F_{,84}(OH)_{1,16})$

1 = Pelletal phoshorite apatite

2 = Miocene phosphatic sandstone apatite

3 = Hoedjiespunt apatite

The contents of the unit-cell have been recalculated to an even 10 cations in the Ca positions and 6 cations in the P positions. Although sulphur has been included it never forms a significant part of the unit cell. Comparing the structural formulae with that of fluorapatite, $Ca_{10}(PO_4)_6F_2$, it is immediately apparent that the francolite is markedly Ca deficient and slightly P deficient. Whereas most francolite analyses show excess of fluorine over that required in the fluorapatite formula , the western Cape francolites have slightly less. In Table 9.2 it will be seen tht there is a wide range in composition of the pelletal phosphorite, particularly with respect to CaO, MgO and SO_3. The other constituents, however, are relatively constant. The MgO content is much higher than would normally be found in pure carbonate-fluorapatite. Rooney and Kerr (1967) have also found this to be the case with their North Carolina pelletal phosphorite. They suggest that Mg could either substitute for Ca or be present as an impurity.

Gulbrandsen (1970) has found a relationship between CO_2 content of apatite and regional facies. In the Varswater Formation on Langeberg maximum CO_2 content is centred around S10 with values decreasing fairly regularly outwards to form a lobate NW-SE trending body (Figure 9.1), which parallels the gross structure of the platform area on which it lies (see Tankard 1974a).

IV. X-RAY DIFFRACTION DATA

Mehmel (1930)and Náráy-Szabó (1930) independently determined the structure of fluorapatite. The hexagonal unit-cell has the dimensions a = 9,37 Å and c = 6,88 Å (Whippo & Murowchick 1967). But the difference in characteristic d-spacings between carbonate and non-carbonate apatite is small. In Table 9.4 the measured interplanar spacings for pelletal phosphorite, bone apatite, and Richtersveld francolite are tabulated and compared with North Caroline pelletal phosphorite (Rooney & Kerr 1967).

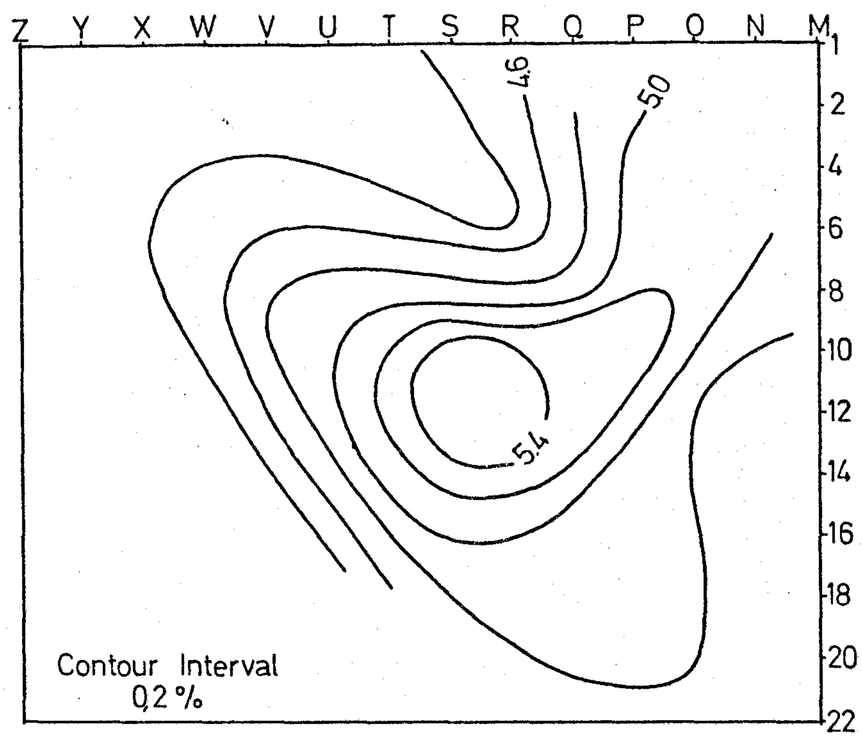


Figure 9.1 Distribution map showing CO_2 contents of phosphorite pellets on Langeberg.

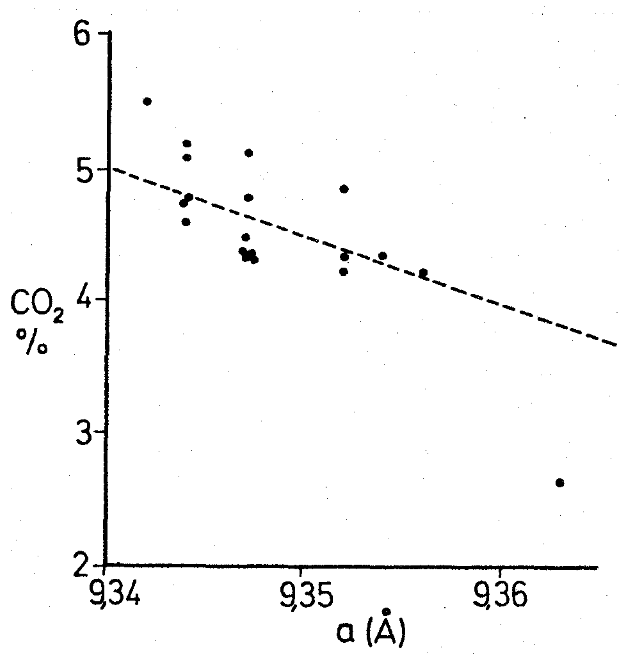


Figure 9.2 Relationship between CO_2 content and a-cell dimension of francolite.

TABLE 9.4

Francolite X-ray diffraction Data
(Fe-filtered Cobalt Radiation)

hkl	N.Carolina Pellets d(Å)	Richtersveld Encrustation d(Å)	Langebaanweg Pellets d(Å)	Langebaanweg Bone d(Å)	Intensity I
100	8,12	8,12	8,12	8,15	M
200		4,04			M
111		3,87			M
002	3,45	3,46	3,44	3,44	VS
102	3,16	3,18	3,16	3,17	M
120, 210	3,06	3,06	3,06	3,06	M
121, 211	2,79	2,79	2,79	2,79	VS
112	2,78				
300	2,70	2,70	2,70	2,70	S
202	2,62	2,63	2,62	2,62	S
301		2,52			W
122, 212	2,29	2,28	2,28	2,36(?	M
130, 310	2,24	2,24	2,24	2,23	S
131, 311	2,13	2,13	2,13	2,13	M
113	2,06	2,06	2,06	2,06	M
123 β		2,02	2,03		W
203	1,996	1,995	1,998	1,992	M
222	1,932	1,932	1,932	1,924	S
132, 312	1,875	1,877	1,878	1,879	M
123, 213	1,835	1,838	1,833	1,830	S
231, 321	1,789	1,789	1,782	1,791	M
140, 410	1,763	1,763	1,763	1,763	M
402	1,738	1,743	1,742	1,746	M
004	1,723	1,724	1,720	1,721	M
232, 322			1,634	1,629	W
133, 313			1,607		W
a	9,34	9,354	9,344	9,363	
c	6,89	6,836	6,879	6,884	
c/a	0,7377	0,7308	0,7362	0,7352	

The Richtersveld sample comes from the Wondergat at Annisfontein where 1,2m of pure, clear carbonate-apatite overlies the Kaigas limestone. The francolite forms a thick botryoidal crust on the altered limestone and slate (de Villiers & Söhnge 1959). The Richtersveld francolite gives well developed diffraction peaks. The cryptocrystalline nature of the pelletal phosphorite, on the other hand, results in poorly resolved diffraction peaks. But in all cases the diffraction patterns are typically apatitic.

McClellan and Lehr (1969) and Gulbrandsen (1970) have attempted to correlate the degree of substitution of carbonate for phosphate with changes in the lattice dimensions. In this study it was found that the range of the a-cell dimensions (9,34 to 9,42 Å) was far greater than the range of the c-cell dimensions (6,88 to 6,91 Å) (Table 9.5).

TABLE 9.5

Unit Cell Dimensions (Å)

No.	a	c	c/a
1	9,344	6,879	0,7362
2	9,344	6,879	0,7362
3	9,344	6,885	0,7369
4	9,347	6,884	0,7365
5	9,352	6,884	0,7361
6	9,352	6,879	0,7356
7	9,344	6,876	0,7359
8	9,347	6,876	0,7356
9	9,347	6,876	0,7356
10	9,344	6,884	0,7367
11	9,342	6,884	0,7369
12	9,347	6,884	0,7365
13	9,347	6,884	0,7365
14	9,347	6,876	0,7356
15	9,352	6,876	0,7352
16	9,347	6,888	0,7369
17	9,354	6,884	0,7358
18	9,356	6,876	0,7349
32	9,363	6,884	0,7352
33	9,423	6,908	0,7333

It would be expected that the various substitutions would have a greater effect in the a-cell dimension. Figure 9.2 shows that with increasing CO_2 content there is a tendency for the a-cell dimension to decrease. This relationship is expressed as follows:

$$Y = -0,02x + 9,44$$

where Y = CO_2 content and x = a-cell dimension. The results of correlation tests were disappointing, probably due to the effect of non-additive substitutions. However, although the points show considerable scatter they do suggest a trend, i.e. that CO_2 content is inversely proportional to the a-cell dimension.

Although the interplanar spacings of the francolite are very similar to those of fluorapatite, it appears that the a-cell dimension is sufficient to distinguish between the mineral species. As a result of isomorphous substitution francolite shows contraction along the a-cell dimension, and the range 9,34 to 9,36 Å is smaller than the 9,37 Å of fluorapatite quoted by Whippo and Murowchick (1967). The crystal chemistry shows the Hoedjiespunt apatite to range from dahllite through francolite and measurements show that it has an a-cell dimension of 9,42 Å. Other reported a-cell dimensions of dahllite are: 9,45 Å (McConnell 1960); 9,39 Å to 9,41 Å (McConnell 1938); 9,42 Å (Carlström 1955); 9,41 Å (Deer,Howie and Zussman 1962).

It was mentioned earlier that although it is now generally accepted that the carbonate is an intimate part of the apatite, there have been suggestions that it could be present as a separate $CaCO_3$ phase. If attributed to calcium carbonate the range of CO_2 shown in Table 9.2 would equate to a range of 9,7 to 12,6 per cent $CaCO_3$. Ames (1959) found experimentally that the apatite can contain up to 10 per cent carbonate, above 10 per cent a separate carbonate phase can be expected. But more recently Legeros et al (1967) have prepared synthetic carbonate-apatite containing 16,5 per cent CO_2 while natural carbonate-apatite containing 16 per cent CO_2 has been reported from the Moroccan continental shelf (Eldefield et al 1972). McConnell and Gruner (1940) believe that a separate $CaCO_3$ phase would be detectable at 10 per cent concentration, Carlström (1955) believes it to be detectable at 1 to 2 per cent concentration. In none of the diffractograms obtained in this study was there any indication of characteristic calcium carbonate reflections. It must thus be assumed that the carbonate of the apatites from the south-western Cape is a structural part of the mineral.

V. INFRARED SPECTRAL ANALYSIS

Table 9.6 lists the characteristic infrared absorption bands of francolite, calcite and aragonite. A characteristic of apatite is a large absorption peak at 9,60 to 9,66 μ which is apparently caused by P – O antisymmetric stretch (V_3 mode) while the band at 10,38 μ is the result of P – O symmetric stretch (V_1 mode) (Adler 1964). The presence of a

TABLE 9.6

Characteristic Infrared Absorption Bands (μ)

Band	Francolite[a]	Francolite[b]	Francolite & calcite	Calcite	Aragonite[c]
C – O	–	–	–	5,57	5,61
P – O	–	6,17	–	–	–
C – O	6,88	6,90	–	–	6,81
C – O	7,00	7,02	7,00	7,12	–
P – O	9,15	–	–	–	–
C – O	–	–	–	–	9,23
P – O	9,60	9,60	9,66	–	–
P – O	10,37	10,38	–	–	–
C – O	11,42	–	–	11,46	–
C – O	11,56	11,59	11,59	11,80	11,70
P – O	–	12,56	12,56	–	–
P – O	–	12,90	12,90	–	–
C – O	–	–	–	14,06	14,04
C – O	–	14,50	14,50	–	14,31

a Colorado (Rooney & Kerr 1967)

b Langebaanweg

c (Adler & Kerr 1962)

very weak V_1 peak is indicative of the lowering of the symmetry of the PO_4^{3-} group, these vibrations being forbidden the full tetrahedral symmetry. The presence of carbonate ion gives rise to similar peaks at 6,90 μ and 7,00 to 7,02 μ. Adler and Kerr (1963) find that calcite has only one peak in this region, presumably the 7,12 μ band in Table 9.6. But it is interesting that aragonite has a band at 6,81 μ. Obviously the carbonate in the francolite behaves differently to that in either calcite or aragonite, and indicates that there is no separate $CaCO_3$ phase. Neither are the C – O bands at 11,59 μ and 14,50 μ in the francolite matched by corresponding bands in either calcite or aragonite. According to Gulbrandsen et al (1966) the band at 11,59 μ in the phosphorite represents a shift from the corresponding position of 11,46 μ in the calcite.

Rooney and Kerr (1967) found that spectral analysis of finely ground and citrate leached francolite left the 7 μ doublet essentialy unchanged, which precluded the possibility of calcite being a separate phase. This was also confirmed for the Langebaanweg phosphorite. Finely ground pellets from the Varswater Formation were treated with hot, dilute acetic acid. This treatment had no effect on the characteristic francolite absorption bands.

VI. DISCUSSION

Chemical and X-ray data show that the phosphorite at Hoedjiespunt is partly dahllite (less than 1 per cent fluorine and a-cell dimension 9,42 Å), but that it probably also forms an isomorphous series with francolite since in one sample the fluorine content was 1,67 per cent (sample 34). The basal bed on Langeberg contains a carbonate-apatite in the matrix but it is defined on fluorine content (> 1 per cent F) and a-cell dimension (9,34 to 9,36 Å) as francolite. Whereas the apatite of the basal bed is authigenic, reworking by a Pliocene transgressive sea gave rise to the pelletal phosphorite of the Varswater Formation (Tankard 1974b). The pelletal phosphorite is identical to the basal bed phosphorite chemically and structurally.

Several apatites form a continuous series of solid solutions and cell dimension can be expected to vary with variation in composition. Figure 9.2 shows that there is a tendency for the a-cell dimension to decrease with increasing carbonate substitution for phosphate. This substitution, however, had little effect on the c-cell dimension. The a-cell dimension distinguishes carbonate-fluorapatite from other apatites (Altschuler _et al_ 1953; Silverman _et al_ 1952). Increasing carbonate substitution causes a contraction of the a-cell dimension from that of pure fluorapatite (a = 9,36 Å). Dahllite on the other hand is marked by a larger a-cell dimension.

At the base of the Varswater Formation in estuarine sediments there are extensive mammal fossil remains. In fresh bone the mineral of the inorganic part is dahllite, a carbonate-hydroxyapatite (Carlström 1955), and is characteristically low in fluorine content. At Langebaanweg post-burial diagenesis has affected all these remains. The lengthy association of these fossils with a phosphate rich environment has caused enrichment in fluorine and over a length of time the bone apatite has changed from dahllite to francolite (F = 1,61 per cent; a-cell dimension = 9,36 Å).

To account for the CO_2 content of carbonate-apatite Hendricks (1952) postulated that the carbonate was present outside the lattice as adsorbed or "amorphous" carbonate. Gruner and McConnell (1937) hypothesised that the carbonate ion was an intimate part of the apatite lattice. Borneman-Starinkevitch and Belov (1955, quoted in Smith & Lehr 1966) explained that the substitution of the planar CO_3^{2-} group for the tetrahedral PO_4^{3-} group leaves a vacant oxygen site which is then filled by fluorine, thus preserving electroneutrality. Although the role of the carbonate is not completely resolved, it is now generally agreed that the carbonate group substitutes for the phosphate group (McConnell 1952; Altschuler _et al_ 1958; Ames 1959; Altschuler _et al_ 1953; Silverman _et al_ 1952). This conclusion has been confirmed in this study.

It was shown that if the CO_2 content was attributed to a free carbonate phase it would equate to 9,7 to 12,6 per cent $CaCO_3$, but X-ray diffraction has failed to reveal any extraneous carbonate. It was also shown that increasing CO_2 content results in contraction of the a-cell dimension, which could only occur if the carbonate were part of the apatite lattice. Infrared spectral analysis showed the presence of CO_3^{2-} but in an environment different to that of calcite or aragonite. Furthermore, acid leaching of finely ground phosphorite failed to remove the carbonate.

In view of the wide range of substitutions possible in apatite, and in view of the marine origin of the phosphorite, one would expect the chemical composition to be related to depositional environment. The cations most readily available as replacements for calcium are Na, K and Mg (Lehr et al 1967). The phosphatic sandstone of the basal bed on Langeberg is situated now at about 30m above sea level but the actual beach zone is probably now at 56m above sea level. On the other hand the Hoedjiespunt microsphorite represents a deeper water facies in which the free quartz and clay mineral form a very small portion. This microsphorite is now situated at only 5m to 6m above sea level and probably formed at a depth of 50m in a sheltered environment. This difference in environment probably explains why at Hoedjiespunt the apatite is partly dahllite, but at Langeberg it is francolite.

CHAPTER 10

GENESIS OF THE PHOSPHORITE AND ALUMINIUM PHOSPHATE

I. PHOSPHORITE

A. Introduction

Since very few (if any) phosphorites are known to be forming at present the processes that form phosphorite can only be inferred from geologically young deposits. Many authors have remarked on the close relationship between phosphoritic deposits and areas of active upwelling of nutrient-rich water (McKelvey 1959, 1963; McKelvey *et al* 1953; Sheldon 1964; Tooms *et al* 1969).

Geologically young phosphorite is commonly found in sediments adjacent to areas of modern oceanic upwelling. These are mainly on the west coasts of continents but also to a limited degree on other coasts. Such areas of active upwelling and phosphorite occurrence lie between the 40th parallel (Sheldon 1964) e.g. S.W. Cape Province, Morocco, S. America, California. Ancient phosphorites on the other hand are found at higher latitudes, their present distribution being the result of lateral displacement of continents. Pevear (1966) has shown that upwelling is certainly not a prerequisite for phosphate enrichment. He has suggested an estuarine origin for the phosphate of the Phosphoria Formation of the Atlantic coastal plain of the U.S.A. Such an estuarine environment would be an area of high biologic productivity. Although an estuarine origin has also been proposed for some of the Agulhas Bank phosphorites (Parker 1975), the evidence is not conclusive.

Marine apatites could be formed in many different environmental settings, but the optimum conditions for apatite formation that would account for the major phosphate deposits concern us most. Most accounts of phosphorite genesis follow Kazakov's (1937) model of direct inorganic precipitation of marine apatite. This model attributes phosphorite formation to direct precipitation of apatite from upwelled phosphate-rich waters.

Sea water is saturated with respect to calcium, and this calcium is common to both the $CaCO_3$ and apatite phases. It is important, therefore, that the phosphate is supplied in such quantities that apatite becomes the solid phase in equilibrium with sea water, rather than $CaCO_3$ or in combination with $CaCO_3$ (Gulbrandsen 1969).

Brongersma-Sanders (1957) has related marine mass mortalities, and phosphorite deposition to regions of upwelling. Accumulation and dissolution of organic matter in the oceans at depths below 500m concentrate phosphorus, and with upwelling of this phosphorus-rich water caused by trade-winds blowing the surface water offshore, large populations of organisms arise and precipitation of apatite occurs. The contribution from decaying organic materials is the most important factor controlling phosphorite formation (Burnett 1974). Burnett notes that diatoms contain large quantities of phosphorus. The factors favouring phosphorite formation are summarised in Figure 10.1. Precipitation of the phosphate occurs when the amount of phosphate supplied exceeds the saturation value of sea water. Precipitation of the apatite from such saturated water is accelerated by increasing temperature and pH (Sheldon 1964). Such conditions are the result of solar heating of cold upwelled water over the shallow continental shelf. Increase in temperature results in a decrease in partial pressure of CO_2 with a consequent rise in pH. Rising temperature and pH lowers the solubility of apatite (Kramer 1964). Rising temperature and pH are also conditions favouring the precipitation of calcium carbonate (Gulbrandsen 1969). But one situation that would favour apatite precipitation would be that of high organic content of sediments. Organic material acts as an inhibiting agent to calcium carbonate precipitation, but as a catalyst to apatite formation (Burnett 1974). Such conditions exist in the reducing pore waters of sediments close to the sediment-water interface which may even be supersaturated with respect to calcium carbonate.

Factors such as changes in organic productivity, intensity of upwelling, changes in ocean currents and continental runoff, possibly have a

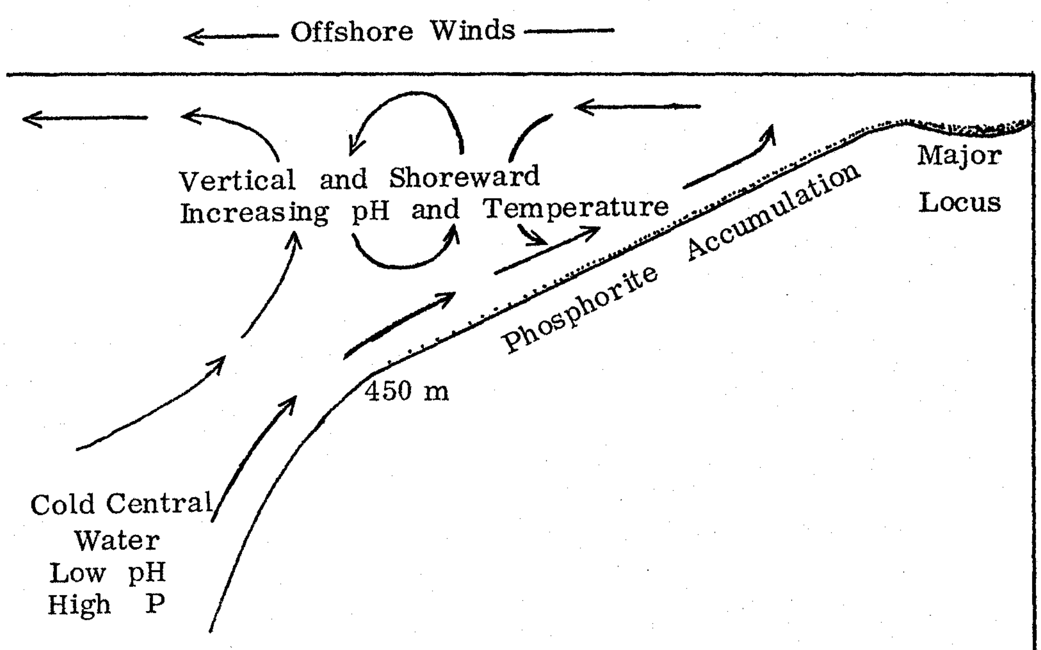


Figure 10.1 Diagrammatic illustration of optimum conditions for formation of phosphorite.

significant influence on the depositional environment (Burnett 1974). Tooms et al (1969) have found a close association between ancient phosphorites and arid areas. These areas, adjacent to upwelling water masses, generally display low rates of sedimentation. They have also suggested that carbonate apatite forms within the pore waters of sediments where phosphate is more concentrated than in bottom waters. Bushinsky (1966) and d'Anglejan (1967) suggest that phosphorites have formed in water shallower than 100m, while Parker and Simpson (1972) have found the greatest concentration of phosphate nodules on the Agulhas Bank between 100 and 140m.

B. Phosphorites of the Saldanha Area

The phosphorites of the Langebaanweg-Saldanha area have much in common with other phosphorite deposits. These deposits occur onshore adjacent to a belt of active upwelling of nutrient rich water. These conditions are illustrated in Figure 10.1 where the shelf is steep and the shelf-break deep. The upwelling is generated by strong offshore winds, related to the anticyclone at 30^{o}S, driving the surface water offshore. This belt of cold upwelled water is an effective barrier against the moisture laden winds from the central Atlantic, causing precipitation offshore. The climate on the adjacent coastlands is consequently arid (average annual rainfall about 260mm) and perennial runoff is restricted to the Berg, Olifants and Orange Rivers. These rivers are contributing only small quantities of terrigenous sediments to the shelf at the present-day (Dingle 1973b) while the Cape Submarine Canyon probably acts as a sediment drain to movement of sediment (Tankard in press b). Sedimentation rates would be relatively slow along the west coast, and dilution of the phosphorite would have been minimal.

The marine apatite is formed authigenically in the Miocene sands. Although these phosphatic sandstones have been truncated at 30m a.s.l. on Langeberg, it has been inferred that littoral limestone at 56 m a.s.l. north of Saldanha represents the furthest inland extent of that sea (Tankard 1974a). A shallow marine genesis of the Hoedjiespunt microsphorite is suggested by the oolitic texture, bioclastic material, microsphorite-flake breccia, wave-generated granite boulders, reworking, and Fissurella and Patella fossils. But the Hoedjiespunt microsphorite could still be a deeper water facies relative to the Langeberg phosphatic sandstone where the microsphorite forms the matric of a clastic sediment.

A common characteristic of phosphorite accumulations is their location on one side of a basin where deep phosphate-rich waters are upwelling adjacent to a shallow shelf (Blatt et al 1972). Although uneconomic concentrations of phosphorite do occur on the open Atlantic coast between Saldanha Bay and St Helena Bay, the major locus of accumulation (Figure 10.1) is found in the erosional basin between Langebaanweg and Saldanha. The nearly ubiquitous presence of organic matter within the pellets suggests high productivity and a common origin for the organic matter and the phosphorite. At the base of the Varswater Formation on Witteklip the pelletal phosphorite is black due to excessive quantities of carbonaceous material (Figure 8.5). The locus of major apatite formation is also that of maximum organic-matter production (Gulbrandsen 1969). The Saldanha-Langebaanweg erosional basin would have been such a locus (Figure 10.1), where sheltered conditions would act as a trap. Likewise, on the Agulhas Bank the only significant local concentration of phosphate other than that provided by the erosion of Tertiary phosphorite, is in sediments enriched in non-skeletal organic matter (Summerhayes 1973).

In the Miocene basal bed on Langeberg the phosphorite occurs as an interstitial component of a quartzose sandstone (packstone). Frequently the phosphorite occurs as drusy-encrustations about the quartz grains indicating crystal growth from the grain surfaces, while the remaining voids are filled with clear collophane and argillaceous and organic material. The structureless aspect of the phosphorite pellets of the Varswater Formation, the rare occurrence of nuclei and the mixing of collophane with organic, argillaceous and ferruginous material suggests precipitation within the interstices of the sediment, an origin identical to that of the basal bed from which it was derived. Ames (1959) found that the replacement of calcium carbonate was possibly the only way in which the carbonate fluorapatites could form. Rooney & Kerr (1967) have suggested that the North Carolina phosphorite probably originated by replacement of calcareous matter and by chemical precipitation under reducing conditions in a large shallow lagoon or estuary with a restricted circulation. Pevear (1966) suggests that the reason why phosphorite is not forming today in Georgian estuaries is because virtually no calcium carbonate is forming at present.

The only evidence for a replacement mechanism within the phosphatic sandstones of the Saldanha Formation consists of very occasional phosphatised foraminiferal tests and echinoid spines. The radial growth of francolite crystals about quartz grains, and the euhedral crystals of francolite (Figure 8.9) are more consistent with a direct precipitation mechanism. Tooms et al (1969) have shown that the synthesis of carbonate apatite from solution means that the prior existence of calcium carbonate is not a prerequisite to formation of this mineral.

The Hoedjiespunt microsphorite consists of less than 1 per cent quartz, and the apatite mineral ranges in composition from dahllite through francolite. Oolites are common in the microsphorite. It is conceivable that the microsphorite arose by phosphate mineralisation of a micrite and bioclastic material (biosparite). Certainly much of the microsphorite has arisen from replacement, but not necessarily all of it. The elemental ratios (CaO/P_2O_5 and F/P_2O_5) for the Hoedjiespunt microsphorite and the Langeberg matrix are close to those predicted for pure apatite, leaving no residual calcium carbonate to suggest an original carbonate phase. It was shown in the previous chapter that the carbonate in these deposits is an integral part of the apatite.

The close proximity of the Langeberg and Hoedjiespunt phosphorites throws further light on the origin of phosphorite. Burnett (1974) and Manheim et al (1975) have noted that wherever youthful phosphorite was encountered on the sea floor off Peru and Chile, the bottom waters were deficient in dissolved oxygen. These low oxygen values resulted from the high organic productivity off those coasts. The composition of pore water in bottom sediments is extremely sensitive to slight changes in the oxygen content of the sea water, and phosphate content is high where dissolved oxygen is lowest (Sholkovitz 1973). Burnett (op.cit.) predicts that such conditions are most likely to be found within pore spaces of sediments. The Langeberg deposits confirm this type of origin, where the precipitation of apatite has taken place in the pore spaces close to the sediment-water interface. The high organic carbon content of the matrix suggests low oxygen values at time of formation, and this also inhibited formation of calcium carbonate. But the Hoedjiespunt microsphorite does not have a clastic framework, and the bioclastic and oolitic material suggests a very shallow depositional environment. The low detrital quartz and clay content are suggestive of a sheltered environment with minimal circulation. The apatite on Hoedjiespunt is preserved in a depression in the granite platform. It is suggested that this was a shallow area in the Miocene, and the sheltering was such that water circulation and oxygenation were minimal and the environmental conditions equated with those of the pore waters of anoxic sediments. Low oxygen values at Hoedjiespunt would have been maintained because of its location on the seaward edge of an embayment adjacent to active upwelling where organic productivity was high.

Summerhayes (1970, 1973) has noted the limited evidence for contemporaneous formation of phosphorite. Very few submarine phosphorites are of Recent age. McKelvey et al (1953) found no bedded phosphorites younger than Late Tertiary. The environmental conditions most favourable to phosphorite formation occurred in the warmer mid-Tertiary seas (Tooms et al 1969).

Because phosphate deposition is favoured by periods of warmer seas (Burnett 1974) there is only limited evidence for contemporaneous formation of phosphorite (Goldberg & Parker 1960; d'Anglejan 1968; Baturin 1970, 1971; Summerhayes et al 1972; Summerhayes 1973; Burnett 1974). Baturin (1971) and Summerhayes (1973) discuss possible present-day formation of phosphorite along the South African shelf. There is evidence that phosphorite is forming at present off Peru (Manheim et al 1975), but even there it is limited to replacement of foraminiferal tests. There is no evidence for substantial present-day formation of bedded phosphorites from any of the continental shelves.

Haughton (1932) attributed the phosphorites of Langebaanweg to alteration of a calcareous deposit by phosphatic solutions. He thought the aluminium phosphate and the phosphorite had a similar phosphate source, in that both were derived from organic masses such as guano. Haughton saw three possible ways in which the phosphate could have arrived in the sediments at Langebaanweg:

(i) the phosphate was derived from the guano deposits about Vredenburg by percolating waters;

(ii) the area in question could be covered by continuously replenished guano-dust blown in by west winds;

(iii) the guano was deposited on islands at Langebaanweg, but he notes that no evidence of the guano can be found in the area.

If the phosphate was transported by groundwater from Vredenburg, it must be remembered that the Langeberg deposits are 13km from that area. Phosphorite deposits are also found along the Namaqualand coast where the topography is not conducive to guano accumulation. Thin-section analysis has already confirmed a marine origin.

Certainly on the granite hills north and south of Saldanha Bay percolating phosphate solutions have produced aluminium phosphates, the mineral produced depending entirely upon the rock being phosphatised, e.g. on Konstabelkop the phosphatisation of granite and limestone. On Witteklip-Sandheuwel the phosphorites are banked against the same granite masses but the phosphate is a carbonate fluorapatite just as at Langeberg. Similarly the phosphorite at Hoedjiespunt is a carbonate apatite, although it rests on a granite shelf which it has phosphatised. It has been shown that the basal bed at Langeberg contains the phosphorite as an interstitial material and that there is no evidence of calcium carbonate replacement. Chemically the pelletal phosphorite of the Varswater Formation and the aluminium phosphates of the granite terrain are totally different (compare Tables 8.2 and8.3). The phosphorite characteristically has a higher F content, while the aluminium phosphates contain higher concentrations of Al_2O_3 and Fe_2O_3. If these deposits had a similar origin then surely the pelletal phosphorites should include an aluminium-rich component. Frankel (1943) has noted the small percentage of iron and aluminium phosphates. Frankel suggests that the sources for the fluorine must be found in either the sea or the granitic areas. The latter option must be discarded since the aluminium phosphates on the granitic rocks are notably poor in fluorine and the phosphate mineral is variscite. Frankel concludes that phosphatic solutions in percolating through these deposits "converted the limestone and calcareous nodules into (probably) hydroxy-apatite and the argillaceous rock types into aluminium and iron phosphates". Marine phosphorites are generally oversaturated with fluorine, and insular and lateritic phosphate deposits undersaturated with respect to fluorine.

Parker (1975) has attempted to reconstruct the depositional environments for Agulhas Bank phosphorites. He describes the intraformational conglomeratic nature of the rock which shows no transportation. He points out that the textural features resemble poorly sorted mudflow conglomerates, and believes that they could be largely attributed to density currents. It is unfortunate that he complicates the model by invoking an unlikely sequence of eustatic sea level oscillations. East of Cape Agulhas the conglomerates are located on a flat and shallow shelf where, he suggests, they could best be explained as having an estuarine origin since the area would be favourable to the development of mudflows and turbidity currents. An estuarine origin is not necessarily correct as it is known that the Agulhas Current does have a considerable effect on that shelf where immense submarine dune fields are forming at present (B. Flemming, pers. comm).

Various authors have attributed the origin of pelletal phosphorites to four processes:

(i) Replacement of $CaCO_3$ pellets;

(ii) excretion;

(iii) inorganic accretion, and

(iv) mechanical erosion of phosphorite beds (Trueman 1971).

Trueman also notes that there is evidence that some pellets have formed by all these processes. In the Varswater Formation no evidence was found of oolitic pellets, although oolitic structure was noted in the Hoedjiespunt phosphorite. The common inclusion of finely divided argillaceous and organic material in the pellets as in the phosphatic sandstone matrix, and sedimentological and chemical data suggest that the pellets must have originated mainly from mechanical erosion of the Saldanha Formation. Summerhayes (1973) says that the bulk of the phosphorite in the unconsolidated sediments of the Agulhas Bank must also have had a detrital origin, and was eroded from outcrops of Tertiary phosphorite during Late Tertiary and Pleistocene regressions. In Morocco there are very extensive deposits of emerged Eocene pelletal phosphorite while offshore there are rock phosphorites. Trueman (1971) points out that the offshore rock phosphorite is older than the pelletal phosphorite and that the microsphorite matrix represents the original source of the pellets which were formed by mechanical reworking. Bushinsky (1966) suggests that pelletal phosphorite may form by phosphatisation of faecal pellets. Some, but very few, of the pellets from the Varswater Formation are morphologically identical to faecal pellets. If the majority had originated as faecal pellets, it seems unlikely that they could have survived the littoral energy conditions of the Pelletal Phosphorite Member. The warm mid-Miocene seas were a time of phosphorite genesis, and the Saldanha Formation is characterised by authigenic phosphorite which always contains fewer than 5 per cent pellets. If the Varswater Formation pellets were of faecal origin, it would be difficult to explain how they survived in a high energy environment, and how large scale and selective phosphatisation took place in the Pliocene. Thin-section, mechanical and chemical analysis show a very close relationship between the Pliocene pelletal phosphorite and the Miocene bedded phosphorite. The Miocene rocks are the source of the Pliocene pellets which have formed by mechanical erosion of the former.

C. Summary

Evidence suggests that a lime-replacement mechanism is not essential for phosphorite formation, although phosphatised bioclastic material at Hoedjiespunt and foraminiferal tests in the Varswater Formation illustrate that such replacement does take place. The main requirements favouring inorganic precipitation of marine apatite are:

(i) There must be strong and persistent upwelling of water with high inorganic phosphate content.

(ii) High pH.

(iii) High temperature.

(iv) An oxygen minimum zone.

(v) Suitable nucleating sites, e.g. voids of sediments.

(vi) Supply of organic detritus to inhibit calcium carbonate precipitation, and facilitate phosphorite formation.

(vii) Low supply of detrital sediment so that the incipient phosphorite is not unduly diluted. Such conditions are found adjacent to arid regions.

II. ALUMINIUM PHOSPHATE

The aluminium phosphate on the granite hills north and south of Saldanha Bay is very different from the marine phosphorites of the Varswater Formation, although it, too, can be related indirectly to upwelling phenomena. The upwelling of nutrient rich water inaugurates a food chain upon which great colonies of seabirds thrive. The largest guano deposits are found associated with such areas (Hutchinson 1950). On the Cape West coast the most important guano birds are Morus capensis (Cape gannet), Spheniscus demersus (Jackass penguin) and Phalacrocorax capensis (Cape cormorant). Sea bird guano is a richly nitrogenous phosphorous material, the phosphate concentration increasing as the more soluble nitrates are leached out. Flack (1916) reports the total nitrogen and P_2O_5 contents of mixed guano from Malagas, Marcus and Jutten Islands (Figure 8.24) as 10,94 per cent and 13,80 per cent respectively. CaO for this sample was 12,49 per cent. 4,11 per cent of the P_2O_5 was water soluble and the rest acid soluble. Fresh guano is characteristically rich in nitrogenous phosphorus, but with leaching of soluble nitrates the P_2O_5 concentration increases. Low pH waters transport the phosphate from the guano to the bedrock where it reacts to form new minerals. Thus reaction with limestone has produced calcium phosphate and reaction with clay minerals formed along joint planes in the granite on Konstablekop has produced the aluminium phosphates. Whereas the phosphorites have high carbonate content and are oversaturated with fluorine, these are dependent entirely upon the type of rock that is being phosphatised in the case of the aluminium phosphates. Although the original source of the phosphate can be attributed to guano accumulations with confidence, direct evidence of a guano deposit has been destroyed by weathering. The formation of phosphorite, on the other hand, appears to be largely due to precipitation from phosphate-rich sea water. At Hoedjiespunt this has led to formation of microsphorite. While the precipitation of this phosphorite is temperature and pH dependent, the aluminium phosphates appear to be dependent upon large sea bird colonies. Upwelling of nutrient rich waters could at present sustain vast seabird colonies but there is little evidence of phosphorite deposition at present. Thus although the two phosphate deposits are dependent upon the upwelling, there is no reason why the phosphorites and aluminium phosphates should be related in time. It is more likely that the aluminium phosphates are derived from leaching of guano deposits which have accumulated over a considerable length of time. Visser and Schoch (1973) believe that these phosphates can be correlated with particular sea levels, but if they owe their origins to seabird colonies no such relationship seems likely.

According to Harrington et al (1966) aluminium phosphates result from the solution and transportation of phosphate minerals from phosphatic limestone or guano by low-pH waters in hot humid regions. Hutchinson (1950) and Harrington et al (1966) have shown that the largest guano deposits are found in the vicinity of the same regions of upwelling of cold phosphorus-rich water as the marine phosphorites (high pH control). Harrington et al infer that where the bedrock consists of silicates the phosphate will form a variety of aluminium silicate minerals such as variscite or metavariscite and crandallite. Du Toit (1917) found that the average composition of the Konstabelkop phosphate is close to that of barrandite while minute yellowish crystals may be referred to wavellite or variscite. Altschuler et al (1956) have found that the upper part of the Pliocene Bone Valley Formation, Florida, has been altered to aluminium phosphate in a zone averaging about 2m in thickness. This alteration has taken place by weathering and ground-water which has produced a progressive change in mineralogy with depth. They found that the top of the zone is characterised by the aluminium phosphate wavellite while the middle zone is characterised by the calcium aluminium phosphates crandallite and millisite. Both crandallite and millisite are virtually isotropic. The Bone Valley phosphates are characterised by higher CaO and F than the Konstabelkop phosphates, but then the original phosphate was a marine phosphorite in which higher CaO and F would be expected. On Konstablekop the only phosphate mineral definitely identified is variscite, while chemical data (Table 8.3) suggest the possibility of barrandite.

III. CONCLUSIONS

Bedded marine phosphorites of Late Tertiary age are found today in warm climates between the 40th parallels in areas adjacent to divergent upwelling of nutrient-rich waters. Precipitation of the phosphate is dependent upon an increase in temperature as the upwelling water reaches the surface and low rates of supply of terrigenous detritus. At Langeberg a consolidated phosphatic sandstone of Miocene age, has formed by precipitation of the phosphate (francolite) in the voids of a marine sand. The francolite is associated with finely divided argillaceous, ferruginous, and carbonaceous material. On Hoedjiespunt on the other hand the bedded apatite contains a total of only 3,6 per cent SiO_2, most of this being present in a clay mineral. The Hoedjiespunt microsphorite is characterised by oolitic and bioclastic texture and the apatite is defined by the concentration of fluorine and carbonate as dahllite in part, but ranges through francolite. At both Hoedjiespunt and Langeberg sedimentation rates were very low in a sheltered environment. In the phosphatic sandstone at Langeberg the francolite has grown from the quartz grain surfaces, the voids being finally filled with isotropic collophane. The basal bed shows scour and fill structures with intraformational conglomerate infill.

Deposition of the Miocene basal bed was followed by a period of emergence. Then in the Pliocene the basal bed was partially reworked by a further transgression, and the Varswater Formation deposited. Mechanical erosion of the basal bed liberated the matrix material which then became associated with the detrital component of the Varswater Formation as pelletal phosphorite. In thin-section the pelletal phosphorite is very similar to the authigenic microsphorite matrix of the basal bed in texture. It contains much argillaceous, ferruginous and carbonaceous material as well as silt-size quartz particles. Also included with the pellets are biogenic remains, foraminiferal tests, minute fish teeth, etc. The highest organic carbon content is found at the base of the Pelletal Phosphorite Member on the seaward side of the basin, and organic carbon content decreases shoreward.

The phosphorite deposits are briefly compared with the aluminium phosphates which occur on the granite hills north and south of Saldanha Bay. Whereas the phosphorite has formed as a marine precipitate, the aluminium phosphates probably originated by leaching of guano deposits and subsequent reaction of these solutions with the bedrock to form insular phosphate rock. Subsequent supergene alteration has produced a superficial lateritic zone.

CHAPTER 11

NEOGENE OF THE SOUTH-WESTERN CAPE: SYNTHESIS AND DISCUSSION

I. MIOCENE

Bedded marine phosphorites of Neogene age are found today in warm climates between the 40th parallels in areas adjacent to divergent upwelling of nutrient-rich waters. Phosphatisation may be a rapid process, and at Langeberg very rapid lithification has formed a thin (approximately 1m) bed of phosphatic sandstone which has protected a complex suite of sediments of terrestrial and fresh-water origin. Two cyclothems dominated by peats and their overlying kaolinitic clays are separated by a high-energy beach sand. All of these sediments have formed close to sea level and their cyclic nature is probably indicative of a fluctuating environmental setting, for instance marine breaching of a barrier and flooding of the freshwater deposits, rather than tectonic downwarping. Undoubtedly considerable compaction of the peats and clays has taken place.

The Saldanha Formation, consisting of bedded phosphorites, originated by authigenic precipitation of apatite in the pore waters of anoxic sediments, conditions which would inhibit precipitation of $CaCO_3$. The phosphate originated in the Central Water where it accumulated at a depth of 500m (Brongersma-Sanders 1957), and was brought to the surface by active upwelling. Phosphorites have generally formed in water shallower than 100m (Bushinsky 1966; d'Anglejan 1967), while suggested depths of formation in the Saldanha-Langebaanweg area are shallower than 50m. Riggs and Freas (1965) have shown that the Florida microsphorite formed in very shallow water. The phosphorite deposits of Israel form part of the major upper Cretaceous to Eocene phosphogenic province stretching from Morocco through North Africa to Turkey (Trueman 1971). In Israel there is a deeper water facies characterised by oolitic microsphorite with a dominantly calcareous matrix, and a shallower water facies characterised by bone beds containing some pellets and cemented by a microsphorite matrix. This account of Trueman's also fits the Hoedjiespunt-Langeberg outcrops.

The Saldanha Formation is the result of a major Miocene marine transgression. Other marine sediments of Miocene age along the South African coast with which the Saldanha Formation is correlated include those of Zululand (King 1953; Frankel 1960) and the eastern Cape (Ruddock 1968; Bourdon & Magnier 1969). This worldwide Miocene transgression has been attributed to changes in elevation of mid-oceanic ridges due to changes in spreading rates (Russell 1968; Hallam 1971; Frerichs & Shive 1971; Flemming & Roberts 1973; Rona 1973).

Optimum conditions for precipitation of apatite existed in the warm Late Tertiary seas (Tooms et al 1969) when extensive phosphorite deposits were formed.

Flemming (1944) found that during the Miocene New Zealand lay wholly within the subtropical zone but during the Late Miocene to Middle Pliocene polar faunas expanded. Kennett (1972) has recorded from deep-sea cores from the Southern Ocean an increase in planktonic foraminiferal diversity and a reduction in amount of ice-rafted debris in Early and Middle Miocene time indicating a warming trend that ended in the late Middle or early Late Miocene. Ingle (1967) and Jenkins (1967) have suggested that Late Miocene through Middle Pliocene water temperatures were as cool as those of the Pleistocene. The Late Miocene-Early Pliocene Antarctic ice sheet was much thicker than at present, and glaciation has been continuous since then (Shackleton & Kennett 1974).

One can hazard a few guesses at Mid-Miocene climatic conditions in the southwestern Cape based on the following facts:

(i) phosphorite formation took place in warmer mid-Miocene seas;

(ii) although eastern Antarctica was extensively glaciated at that time, the major ice sheet only developed in the Middle to Late Miocene;

(iii) apatite precipitation to form phosphorite is the direct result of upwelling of nutrient-rich Central Water.

Upwelling is the direct result of offshore winds blowing the warm surface water offshore and this wind system is related to the South Atlantic anticyclone (Tankard, in press b) which is presently situated at 30^{o}S in summer. The position of the anticyclone in space and time would vary with solar radiation. The warmer Mid-Miocene climate would have moved the intertropical convergence and the South Atlantic anticyclone southward so that upwelling off the Saldanha coastline may have been more intense than at present. A very interesting possibility bearing on phosphorite genesis, and which has not hitherto been suggested, is that the upwelling may have become a year-round phenomenon. This certainly would enhance the chance of phosphorite formation. The phosphate would still be derived from Central Water which originated in the Southern Ocean. But if the Antarctic ice sheet was less extensive than at present it would seem likely that the upwelled water would be warmer than the present Central Water. This would have been further heated by solar radiation so that the pH of the surface water may have been higher than today, although the water was not necessarily more saline. But the organic productivity would probably have been at the same level as at present. Together all these factors would facilitate the formation of phosphorite.

The Saldanha Formation sediments in the Saldanha-Langebaanweg and Ysterplaat areas are shallow marine deposits. On Langeberg this is evidenced by scour and fill structures, with the fill material consisting of intraformational conglomerate. The basal bed is truncated at 30m a.s.l., so that there is no indication of the actual shoreline. The Hoedjiespunt microsphorite consists of a lower bedded unit which has become reworked in a higher flow regime. Reworking, oolitic texture, bioclastic material, and <u>Fissurella</u> and <u>Patella</u> moulds show the close proximity of a shoreline. It is believed that these deposits can be equated with a 56m shoreline in the Saldanha area. North of the town, and exposed in a quarry, is a 100m section through a contact of massive marine limestone and overlying high-angle cross-bedded aeolianite (Bredasdorp Formation). The interface between the two units, where they are banked against the granite hills is at 56,4 m a.s.l. The Saldanha Formation at Ysterplaat contains a conservative endemic mollusc fauna at 9m a.s.l. The mollusc fauna which is similar to that of the Alexandria Formation described by Newton (1913), suggests a sandy beach facies.

Other occurrences of the Saldanha Formation phosphorites are found onshore north of Hondeklipbaai, where they are preserved in bedrock depressions; and offshore, particularly between the Agulhas Bank and longitude 25°E. Dingle (1974) suggests that there were two phases of Tertiary phosphorite formation on the Agulhas Bank, Late Eocene and Late Miocene-Early Pliocene or Late Pliocene. Furthermore, he attributes these phases of phosphorite formation to regressive sea level movements. The onshore phosphorites are of Middle Miocene to early Late Miocene age, and it is known that this was a period of worldwide marine transgression and warmer climates. Marine limestones directly overlying the Agulhas Bank phosphorites have recently been dated as Middle Miocene (Rogers 1974), which suggests contemporaneous formation of the Agulhas Bank category a and b phosphorites and the onshore phosphorites.

Northwards to South West Africa, and eastwards along the Cape south coast, Tertiary marine deposits are recorded at higher elevations relative to those of the southwestern Cape. This leads one to suspect that during the time of higher Late Tertiary sea levels, the Swartland between Elands Bay and the Cape Peninsula was above sea level, and subsequent sagging has reduced the Tertiary Sandveld deposits to their low elevations. The two most likely explanations for planation of the Swartland would be marine planation or pediplanation. But the total absence of marine sediments on the Swartland platform suggests it is an ancient feature and probably pre-Miocene since the Miocene shoreline is 250m below the Swartland platform.

II. PLIOCENE

The Varswater Formation is a "classic" transgressive complex; the cyclothem starts with a freshwater kaolinitic clay and passes upwards through marine, estuarine and fluvial, to a dominant marine unit. On Langeberg the cyclothem displays a complex relationship of facies change over very small areas and, presumably, in a short time range.

In the new Varswater Quarry the Varswater Formation lies unconformably upon the Saldanha Formation. The unconformity is an erosional one attributed to tilting and marine transgression. The tilting took place in Late Miocene or Early Pliocene. Wolff et al (1973) include the Miocene phosphatic sandstone, which they believe to be phosphatised calcrete, with the Varswater Formation, but the South African Committee for Stratigraphy, Tertiary-Quaternary Working Group, has accepted it as a separate formation.

With transgression of the Pliocene sea the kaolinitic clay ecozone migrated landward and was overlain by littoral deposits as transgression progressed. Truncation of the basal bed at 30m a.s.l. was associated with a temporary stillstand of the sea which, on a shallow mesotidal coast, allowed a barrier-beach complex to develop along the outer edge of the platform. Fluvial inflow from the north and northeast developed a complex suite of estuarine sediments in the lee of the barrier. These sediments include clean quartzose sands which grade laterally into carbonaceous sands and peats. The extensive mammal fauna is associated with the estuarine sediments. Final transgression reworked some of these lower units and carried the shoreline to 50-55m above present sea level.

The 30m shoreline is represented at Langeberg, Sandheuwel and Duyker Eiland. The sea must have remained stable for a considerable period of time since it allowed a considerable amount (1m) of peat to accumulate on the lee of the barrier-beach on Langeberg. The mollusc fauna from the 30m level includes a thermophilic element which suggests water temperatures approximately $5^{o}C$ warmer than today. These species are: Cellana capensis, Turbo sarmaticus, Barbatia obliquata, Ostrea atherstonei, and Striostrea cf. margaritacea. All except the Striostrea come from deposits within the Saldanha-Langebaanweg basin, and occur in great numbers. The sheltered and shallow nature of the basin would, in part, account for the warm conditions where the thermophilic molluscs could have survived as relicts. But the Duyker Eiland site with Striostrea faces the adjacent ocean. Oyster valves are very common there, but the very thick valves may indicate conditions that were not quite optimum.

Like the Miocene phosphorites, these thermophilic molluscs enable one to guess at the Pliocene hydroclimate. Miocene phosphorites do occur on the Duyker Eiland and Paternoster platform, but the absence of Pliocene phosphorite suggests that water temperatures were cooler. Striostrea margaritacea occurs extensively in Namaqualand Early Pleistocene sediments adjacent to the open coast (Haughton's Ostrea prismatica). There the valves are in no way stunted or overthickened, and suggest optimum conditions for survival. S. margaritacea requires minimum summer temperatures in excess of 24^{o}C (Korringa 1956). Tankard (1975a) suggested that with warmer Early Pleistocene climates the consequent southward movement of the Intertropical Convergence and anticyclone centre brought tropical conditions down as far south as the Olifants River. One can argue that in the Pliocene these conditions had operated even further south so that the Saldanha-Elands Bay area was transitional. S. margaritacea was thus able to form reproducing communities on the open coast, but it was existing at the extreme of its distribution. Hence the stunted and overthickened valves. Upwelling of cold Central Water would still have taken place, and the Central Water would have been as cold as at present since the Antarctic ice sheet was more extensive than the present ice sheet (Shackleton & Kennett 1974). In the Late Pliocene rapid melting of the Antarctic ice sheet brought it to the present volume, and this was accompanied by a rise of sea level (Tankard, in press c). Bandy (1967) and Bandy and Wilcoxon (1970) found that a Late Pliocene rising sea level was associated with expansion of tropical marine faunas.

It is with the Pliocene mammal fossils of the estuarine facies of the Quartzose Sand Member that Cenozoic research at the South African Museum has been mainly concerned. Hendey (1974) called this horizon the "estuarine faunal unit 1". Marine fossils are rarely found in this unit. Regarding the concentration of the bones Visser and Schoch (1973) suggest that the bones were washed across a beach into the sea by rain, and that with withdrawal of the sea the bones were concentrated by the wind and buried beneath shifting dunes. Butzer (1973a), on the other hand, has suggested longshore drift as the concentrating mechanism. The distribution of the fossils and the fact that bones from individual animals are often closely grouped is against any reworking or concentrating agent. Not many of the bones show signs of transportation (Hendey,pers. comm.) while there is evidence to suggest subaerial accumulation on banks or bars of an estuary. Many of the bones have been chewed by carnivores, while others still show evidence of burning. The most compelling evidence for subaerial accumulation is that suggested by coprolites. Coprolites do not lend themselves to transportation, while groups of coprolites and some with plant impressions on the surface suggest _in situ_ occurrences (Hendy, op. cit.).

A very strong argument against Visser and Schoch (1973) and Butzer (1973a) is the lateral facies change. Over a very short distance these quartzose sands grade into carbonaceous sands, with peat. The fossils from the peat are generally less complete and land tortoise remains common, suggesting brief periods of high energy conditions. But it is envisaged that the peat deposits may have represented a waterhole for the mammals. This is certainly supported by the finding of the distal extremities of a sivathere fore-and hind-limbs, still articulated, and vertically disposed (Hendey, pers. comm.). It is easy to imagine the animal being caught in the mud while drinking, and unable to extricate itself. Waterbirds are common, and egg-shells with colour preserved have been found. The mollusc fauna includes well-preserved forms with a freshwater species common.

Professor E.M. van Zinderen Bakker kindly examined the peats for pollens. A preliminary examination showed a rich pollen, but the usefulness of the spectrum was reduced by the fact that 92 per cent of the sporomorphae belong to one unidentified taxon. Grasses (Gramineae) constituted approximately 3 per cent of the spectrum, a reed (Restionaceae) and a bush (_Cliffortia_) each account for about 1,5 per cent. The inclusion of grasses is very significant as the high-crowned teeth of _Ceratotherium_ and _Hipparion_ are indicative of grasslands (Hendey 1972). Giraffe remains suggest the presence of trees. Hendey (1973) visualises an environment characterised by riverine woodland flanked by grasslands. This suggests strong seasonality of rainfall, with most of the rain falling in the summer months. The present environment is characterised by Mediterranean Macchia vegetation and winter rainfall. Professor Bakker (pers. comm.) notes that the "morphology of the dominant unidentified pollen type is comparable to that of Cuscuta and Elphantorrhiza and also certain members of the Aizoaceae, but because of small differences it could not be definitely assigned to any of these". The most likely plant would appear to be Aizoceae (mesembrianthemum) since that is a dry-area plant and very common on the Sandveld today. Professor Bakker's report continues: "The great abundance of this form (dominant) makes interpretation of the assemblage very difficult. It is possible that the plant which produced it is not important for the understanding of the whole flora. The possibility exists, especially with a great dominance like this, that it is an autochthonous form which does not represent the pollen rain, or even that a flower of the particular plant was present in the sample. In this case the percentage of other types like the grasses, Restionaceae and Cliffortia, would in reality form a much greater part of the spectrum .....

"The presence of Restionaceae and Cliffortia, which are typical of the Cape Flora and winter rainfall, suggests that this flora was already established at the time of sedimentation. It is interesting that only about half of the taxa which were distinguished could be identified. This could be the result of extinction, which would suggest that the flora was still very different from the present one.

"The Gramineae, Restionaceae, Cliffortia, Myrica, Compositae and Chenopodiaceae suggest a treeless, probably coastal, vegetation. The only positively identified tree pollen is that of Podocarpus. As is often known to be the case with Podocarpus, its pollen could have been wind-transported over a great distance from the mountains. The data is, however, inadequate to definitely state that the vegetation was treeless. There is, for instance, no way to prove that the unidentified types do not belong to trees.

"It is difficult to know whether there was more grass present than now. The value of about 3 per cent does not suggest this. If the dominant form is removed from the spectrum, the grasses would form about 30 per cent. This is still considerably less than the average percentage of more than 60 per cent from the grassveld of the Orange Free State".

A recent paper by Wolff et al (1973) is unduly critical of Hendey's (1970a, 1970b) interpretation of the geology of the Varswater Formation. They write that it is "unfortunate that attempts to date the Langebaanweg mammalian assemblage have been made in the light of assumed maximum heights for former marine sediments in the Langebaanweg area in relation to Pleistocene strandlines elsewhere in South Africa ..... and even the 'classical' Mediterranean sea levels of Zeuner (1959)". This criticism is very true, and Hendey (1973), in a paper ignored by Wolff et al (1973, published June 1975), has admitted these faults and given a more objective interpretation of the geology. But Wolff et al are also guilty of very subjective reasoning. In their paper (Wolff et al 1973) they question Hendey's taxonomic work. They then attempt to show that the Baard's Quarry and New Varswater Quarry deposits are related and represent merely facies changes. These authors had not seen the Baard's Quarry geology and have to base their entire argument on faunal lists from the two sites published by Hendey. In their list of species (table 1) they go on to list the faunas together so that the reader is unable to compare the faunas. Hendey (pers. comm.) suggests that the original faunal lists now need updating as the result of continuing taxonomic work. Bearing in mind that the New Varswater Quarry contains over 70 identified mammalian species, there are only two species in common with the Baard's Quarry fauna, although the two sites are only 2500m apart. These two species are Mammuthus subplanifrons and Mesembriportax acrea, both of which have a wide time range. Equus at Baard's Quarry shows that it definitely post-dates the Varswater Formation (Hendey pers. comm.).

There is evidence that fluvial deposits of similar age occur in the Hondeklipbaai area. Hooijer (1972) mentions an isolated rolled tooth of Ceratotherium praecox of Pliocene age that was found in poorly sorted fluviatile gravel at 18m a.s.l. These gravels overlie an Early Pleistocene marine sand, and it appears that the tooth was reworked from earlier deposits.

Deposition in the estuarine-fluvial phase at Langeberg ended with the final transgression, possibly induced by renewed tilting and subsidence, which took the shoreline to 50-55m a.s.l. Reworking of earlier deposits took place. The pellets of the Pelletal Phosphorite Member are detrital grains which originated in the microsphorite matrix of the Miocene phosphatic sandstone. The pellets were formed by mechanical reworking. They are denser than the quartz sand component (S.G. 2,91 vs. 2,655) and are always better sorted, i.e. spread over a narrower size range. But in spite of it being the denser mineral it usually has the slightly larger grain size. In testing for hydraulic equivalence this apparent inversion is explained. Grain-size distribution data show the quartz fraction to be composed of three populations separated by inflexion points at 1,25 $\emptyset$ and 3,00 $\emptyset$. Between these two inflexion points lies the saltation population whose hydraulic behaviour is governed by the same Impact-Law (Newton's law of resistance) as is the pelletal phosphorite. Comparison of the median-size of the saltation population with that of the pelletal phosphorite shows that the two minerals approach hydraulic equivalence.

Generally the low flow regime over the platform and in the shelter of the breaker-bar was insufficient to suspend all the grain sizes, so that only a part of each grain-size range has been selected by settling velocity. Bottom traction has probably modified the size-distributions, while post-depositional changes have been brought about by addition of smaller grains by percolating water. Along the edge of the platform, and the exposed western part of the platform more turbulent conditions operated, so that hydraulic equivalence is approached. The structure contour map (Figure 5.1) shows that the area southwest of the platform was low lying and must have been a deeper water area, and indeed the sediments become finer grained in that direction. Sediments in this area are better sorted than in other areas, indicating more active winnowing conditions. Although these sediments are finer grained, the skewness values show the coarser fractions of the sediment to be dominant over the finer fractions. All these facts would indicate a low flow regime at the platform edge. That the pelletal phosphorite is detrital is indicated by the similar median grain diameters of the quartz and pelletal phosphorite and similar sorting. The reversal of skewness values, compare Figures 6.3 and 6.7, would possibly be explained by the narrow size range of the phosphorite; very few phosphorite pellets are smaller than 3,5 Ø. The size distribution of the pelletal phosphorite has been shown to be related to the detrital part of the sediment and must have been influenced by the same mechanical processes. Whereas the source area for the pelletal phosphorite was very close, the polymodality of the quartz suggests more than one generation for that mineral. Similar grain-size relationships between quartz and pelletal phosphorite have been noted by D'Anglejan (1967) and Summerhayes (1970). The average phosphorite grain-size on Langeberg (c.2 Ø) is somewhat coarser than that found by D'Anglejan, but the size of the pellets is related to degree of exposure of the coast. On the exposed Namaqualand coast typical pelletal phosphorites north of Hondeklipbaai have a range of 0,5 to 1,5 Ø, as opposed to the embayment environment of the Langeberg deposits.

Hendey (1972) has subdivided the Pelletal Phosphorite Member into beds 3a and 3b on palaeontological grounds. Terrestrial and marine aquatic vertebrates are concentrated in Bed 3a, but become progressively less common upwards. Bed 3b is largely unfossiliferous, the only fossils tending to be of a smaller size. The larger mammals so characteristic of the underlying estuarine sediments are either not recorded from Bed 3a or are very fragmentary. Whereas Hendey reads paleaontological significance into this relationship, it is here thought to be a graded deposit. The difference in size is attributed to mechanical processes and is not of palaeontological significance. The fossils of Beds 3a and 3b have been derived from the underlying estuarine sediments and marine fossils added during the final transgression. The smaller size of fossils on ascending the sediments reflects the energy of the eroding medium. Their fragmentary nature also indicates reworking. Large lumps (remanié) of clay have also been torn from the lower horizons and included in the Pelletal Phosphorite Member.

Butzer (1973a, published 1975) has described, from the lower part of the Pelletal Phosphorite Member (Bed 3a), "a number of lenticles of compact, laminated, gray to very dark gray (10YR 3-6/1), clay with sandy pockets ..... Sorting is poor in a bimodal distribution with a secondary grade maximum near 175 μ . These lenticles are seldom over 1 cm thick and possibly include rootlet or worm structures". Butzer's cumulative curve (figure 1) shows a broad clay peak as characterising Bed 3a. His description is misleading since what he apparently sampled was a block of kaolinitic clay reworked from underlying strata. The lower part of the Pelletal Phosphorite Member does not contain clay lenticles in quantities sufficient to characterise the member.

Butzer has analysed 12 samples from the Varswater Formation and 2 from the modern Saldanha beach. Comparison of these samples suggests to him similar depositional environments. But it must be emphasized that he used the 37, 63, 210, 595 and 2000 μ sieves. This sieve spacing is not sufficient to give an environmental interpretation. Furthermore, his statistics are calculated from the Trask (1930) formulae, which make use of the 25 and 75 percentiles, i.e. they avoid the environment sensitive extremes of distribution. The Trask sorting coefficient, for instance, has only a 37 per cent efficiency rating, compared to 79 per cent for the Folk and Ward (1957) formula.

Pelletal phosphoritic sands also occur on the Namaqualand coast, north of Hondeklipbaai, where they overlie Miocene phosphatic sandstone and siltstone (A.J. Carrington, pers. comm.). There the Miocene sediments have been truncated at 36m a.s.l. Pelletal phosphorite supplied by Mr. A.J. Carrington comes from the "Upper E-stage" (mining terminology) which he considers to be of latest Pliocene age. It has strong West African faunal affinities and reflects water temperatures considerably warmer than present. He envisages an arid climate. The final altitude of the "Upper E-stage" is not known. The widespread occurrence of the pelletal phosphorites and the presence of warm-water molluscs suggests a possible correlation with the Varswater Formation.

III. SUMMARY

The sediments of the Saldanha and Varswater Formations are transgressive complexes associated with climates that were warmer than today, and the Miocene warmer than the Pliocene. Onshore deposits of Miocene age are extensive around South Africa, and are a worldwide phenomenon. These Miocene deposits can be attributed to eustatic sea level movement consequent upon elevation of the mid-ocean ridges (Flemming & Roberts 1973). Since the Miocene the sea has generally been regressive. But on the west coast of the Cape Province a secondary transgression, in the Pliocene, has resulted from tilting so that transgression has occurred within a regressive phase of sea level. It is impossible to separate the tectonic from the eustatic components.

CHAPTER 12

REVIEW OF OTHER SOUTHERN AFRICAN MARINE NEOGENE DEPOSITS

I. MOZAMBIQUE

Tertiary sediments in Mozambique include bedded marine Eocene limestone at Salamanga 32km inland (du Toit 1954; King 1972) and Nummulitic limestones, 180m thick, which form plateaux north and west of Beira. The foraminifera include several species of Nummulites and suggest a Middle Eocene (Lutetian) age (du Toit, op. cit.). In the south of Mozambique calcareous sandstones have been described as the Santaca Formation (Soares & Da Silva 1970). Mollusc fossils in the Santaca Formation include Glycymeris borgesi (= G. austroafricana), Pecten sapolwanaensis, Aequipecten uloa, and Amusium umfolozianum. These fossils correlate the Santaca Formation with the Pecten bed at Uloa. Elevation of the Neogene shoreline at 26°45'S is 60m a.s.l.

II. ZULULAND

A. Stratigraphic Setting

Although exposures of Neogene sediments are best developed where river incision by the White Umfolozi River has occurred, marine Neogene sediments probably underlie much of the Zululand (including KwaZulu) coastal plain.

Ever since the first discovery of Megaselachus teeth at Uloa in 1951, the lithology and stratigraphic correlations of scattered outliers of marine Neogene sediments in Zululand have led to an unfortunate controversy (King 1953, 1966, 1970, 1972; Frankel 1960a; 1966, 1968). Although the contentious sequences at Uloa and Sapolwana are "classic" regressive complexes, complications have arisen due to unnecessary debate about precise age and the relationship of these Neogene sediments and "unconformities" to land surfaces. King (1966) believes that continuations of "unconformities" within the marine sediments are represented by an elevated pair of land surfaces. Frankel (1968) suggests that if the pre-Miocene unconformity in the Lower Umfolozi area, which according to King contains reworked laterite, and the inland erosion surface had once been continuous, then originally both were probably nearly horizontal. But he finds it impossible to accept that the subhorizontal unconformity could correlate with a land surface at 600m elevation and only 50km to the west. However, the structure of coastal Natal is dominated by step-faulted tilted blocks, and not monoclinal tilting (Maud 1961) so that if King is correct in his correlation a mechanism does exist to explain both the rapidly varying elevations and the low dip of the pre-Miocene unconformity. The debate is unnecessary since the concepts involved are probably unprovable.

A composite and idealized stratigraphic column is shown in Figure 12.1. Most of the Zululand Neogene deposits lie unconformably upon Cretaceous (Maestrichtian) silty mudstones which dip eastward at 1-2^{o}. The Miocene-Cretaceous unconformity is irregular, but shows no uniform seaward dip in the Uloa area (Frankel 1966). This unconformity is also essentially horizontal in the Richards Bay area, with no discernible landward rise in 5km (Maud & Orr, in press). Maud and Orr show that the higher elevation of the unconformity in the Uloa and St Lucia areas compared with the Richards Bay area indicates a southerly dip of 1,5m/km. The unconformity is a marine abrasion platform, and the presence of reworked phosphorites suggests that it originated in the Eocene, but was probably remodelled in the Neogene.

The base of the Neogene strata at Uloa, 18m a.s.l., consists of a brown gravel bed 30-100 cm thick. Frankel (1966) named this the ferruginous nodule bed. The gravels are poorly sorted and include reworked ammonites, silicified Cretaceous wood, Tertiary shark teeth, bone, phosphatised nodules and Tertiary glauconitic sandstone pebbles. Frankel believes the phosphatised nodules originated in the Cretaceous and Eocene. P_2O_5 determinations were 16,2 per cent, 18,1 per cent and

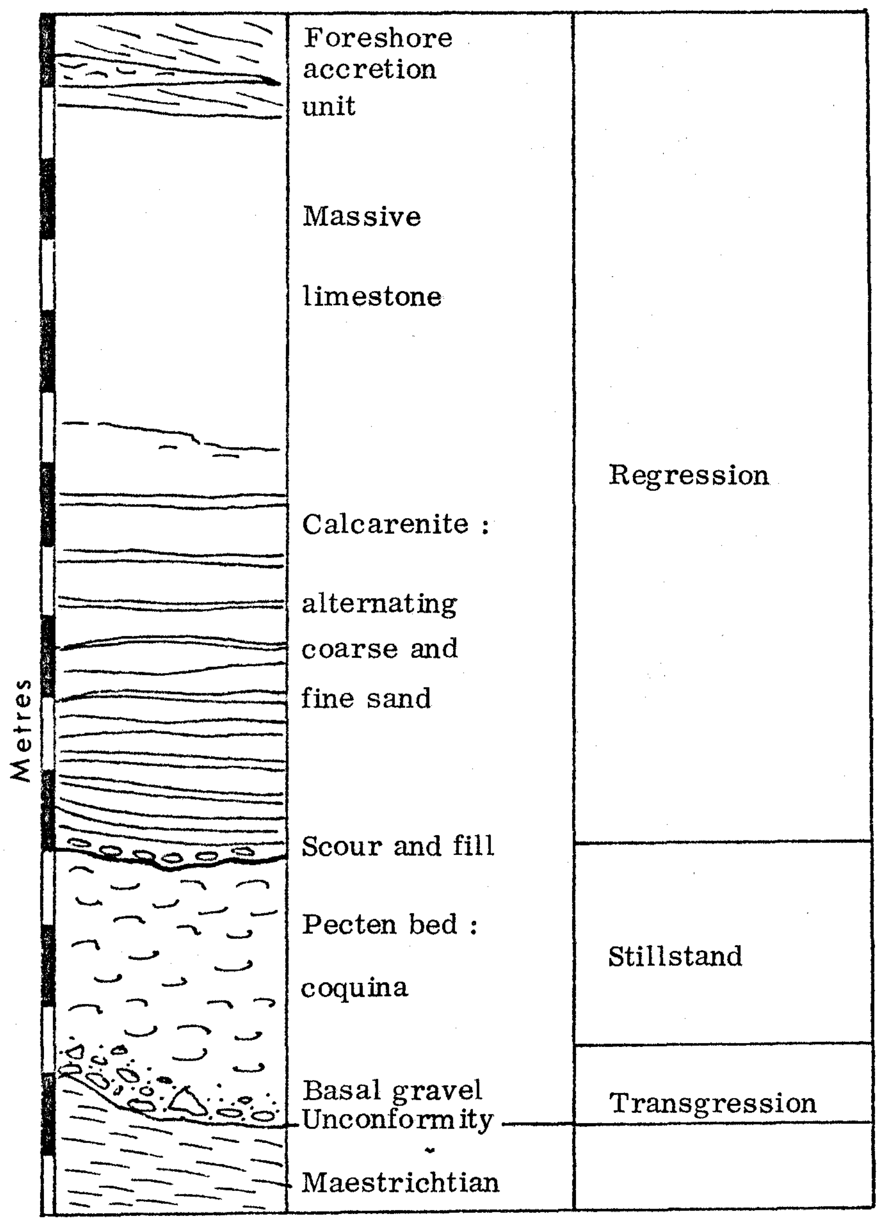


Figure 12.1 Composite and idealized stratigraphic column for Uloa and Sapolwana, Zululand.

24,44 per cent. Together with a CaO determination of 45,1 per cent, a francolite containing phosphorite is suggested. A radiometric date of 55 m.y. on the glauconite suggests the earlier existence of Lower Eocene marine sediments. The gravel bed has been ferruginised by percolating groundwater, and does not contain reworked laterite.

The basal gravel at Uloa grades vertically into the Pecten bed which is 4,6 m thick. This bed is characterised by Aequipecten uloa, although other invertebrate fossils are also abundant. These fossils include coral, crinoid ossicles, echinoid fragments, bryozoans, foraminifera, and shark teeth. King (1953) discusses the palaeontology in detail. The Pecten bed is a coquina in which fine to medium, subround quartz constitutes less than 5 per cent of the total (Frankel 1966). Sparry calcite fills the interstices. The bivalves are usually disarticulated and show a horizontal stratification. The unbroken valves suggest low energy conditions (lower flow regime).

At Uloa the Pecten bed is overlain by 6,7m of well-bedded, flaggy, fine to coarse grained, foraminiferal calcarenites with low-angle cross-stratification. The interface between the Pecten bed and the calcarenites is marked by water-rounded discoidal cobbles. The contact between the two units is an erosional surface, typical scour and fill, which has resulted from an increasing flow regime during a regression. But this does not imply an unconformity as King and King (1959) suggest. McCarthy (in Frankel 1968), on the other hand, believes that the irregularities in the stratified sequence are due to sagging of the strata where removal of carbonates has caused local subsidence. It seems unlikely that solution of a coquina would produce sagging on this scale, because solution would affect the matrix initially while the macrofossils form a self-supporting framework.

The best exposure of the calcarenites is at Sapolwana where alternating coarse and fine calcarenites form fairly continuous flaggy layers 5-15cm thick (Figure 12.2). The thicker layers are well cemented and alternate with less coherent coarser intercolations which are up to 30cm thick in places (Frankel 1966). This type of sequence is typical

Figure 12.2 Pecten bed overlain by flaggy calcarenite at Sapolwana.

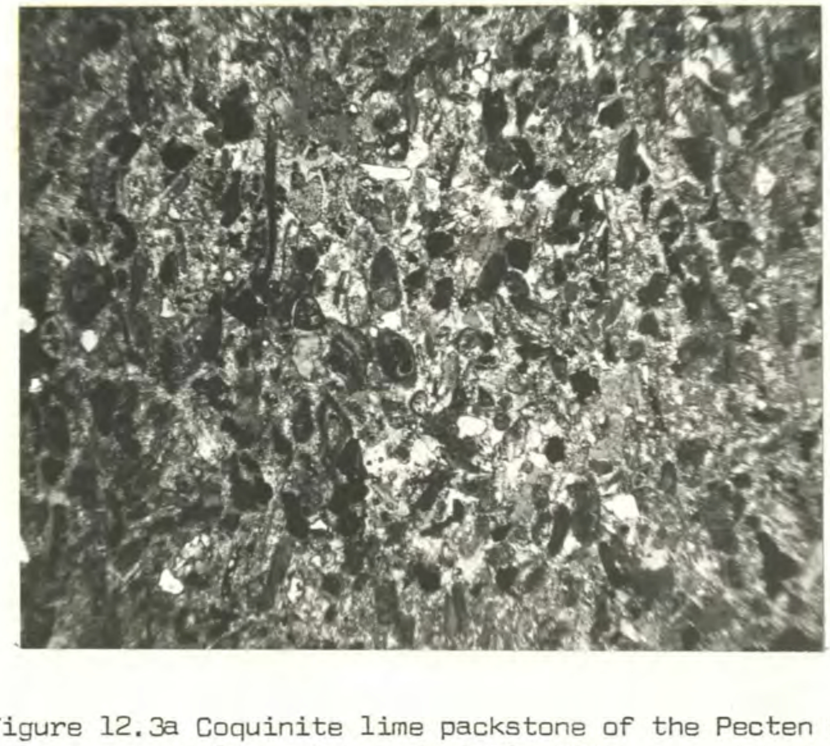


Figure 12.3a Coquinite lime packstone of the Pecten bed, Uloa. Crossed nicols X24.

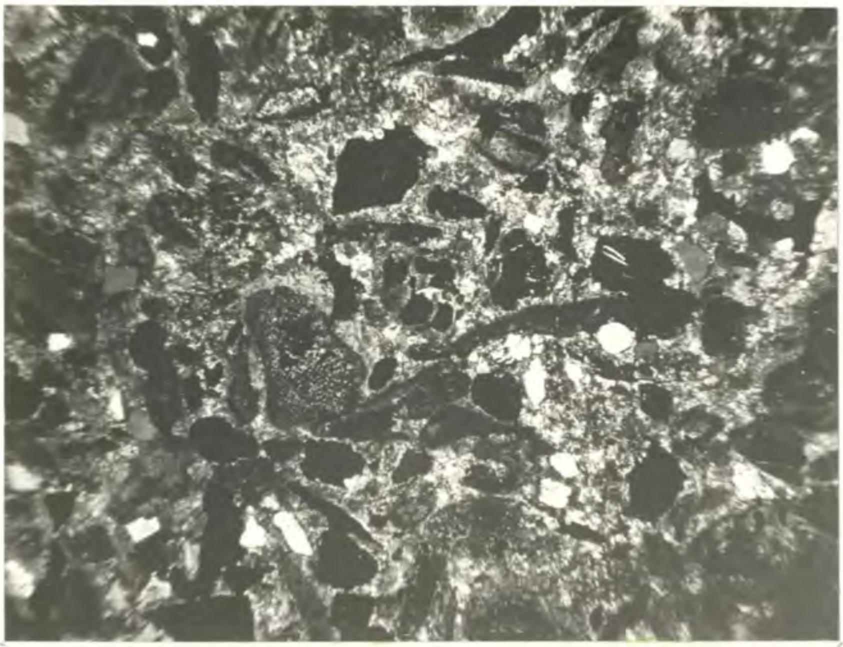


Figure 12.3b Coquinite lime packstone of the calcarenite at Sapolwana. Crossed nicols. X24.

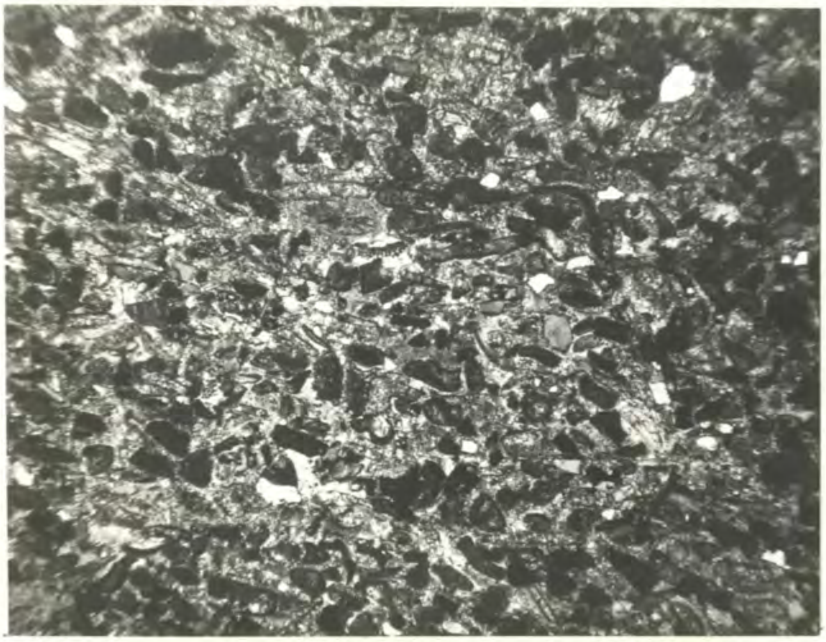


Figure 12.3c Coquinite lime packstone of the foreshore accretion unit at Sapolwana. Crossed nicols. X24.

of a shallow tidal to subtidal environment, and the coarsening-upward beds (Figure 12.2) demonstrate continuous regression. With continued regression the flaggy calcarenites at Sapolwana grade upwards into massive limestones (Figure 12.1), while the top of the succession is capped by a foreshore accretion unit with subordinate Pecten-bearing lenses.

Figure 12.3 illustrates the texture of Pecten bed matrix, calcarenite, and the foreshore accretion limestone. Note the similar textures of the Pecten bed and foreshore accretion unit, and the coarser calcarenite.

Neogene sediments in the Richards Bay area are only 5–6m thick and comprise a lower yellowish brown coarse coquina which is overlain by an upper sandy calcarenite horizon. High-angle cross-stratification in the upper part of the calcarenite shows transition from subaqueous to aeolian deposition (Maud & Orr, in press).

The Neogene deposits at Umkwelane Hill, 13km west of Uloa, consist of a basal conglomeratic zone, 1,5m thick, at 46m a.s.l. The conglomerate is overlain by fine to medium calcareous sandstone and limestone. Subround quartz averages medium grade. The bioclastic material includes an abundance of benthonic foraminifera and mollusc shell fragments (Frankel 1966). High-angle cross-stratification at the top is indiciative of aeolian deposition.

B. Depositional Environments and Correlation

The deposits described above are a typical shallow marine regressive sequence, but lithological correlation between them is not as clear-cut as Frankel (1968) believes. He has commented on the close lithological similarily between the calcarenites at Warner's Drain, Sapolwana, Uloa, Peaston and Umkwelane Hill, and the fact that there is a basal conglomerate both at Umkwelane Hill and Uloa.

The Pecten bed at Uloa, Sapolwana and Richards Bay with the disarticulated but nevertheless, unbroken, Aequipecten valves suggests deposition in a shallow offshore environment. Pectenids are characteristically shallow infratidal molluscs. The upward-coarsening calcarenite at Sapolwana and the overlying foreshore accretion unit suggest a regressive sea. If the Umkwelane Hill deposit is broadly time-equivalent, then an intertidal depositional environment is suggested. This is borne out by the foraminifera. Whereas the foraminiferal assemblage at Uloa includes numerous planktonic as well as benthonic forms which suggest an offshore facies with free access to the sea, the exclusively benthonic foraminiferal assemblage at Umkwelane Hill (Frankel 1968) suggests a sheltered nearshore facies (e.g. lagoonal). The foraminifera from both the Pecten bed and the calcarenite are very similar (Dr Albani, quoted in Frankel 1968), so that the time interval between these two units is small. On sedimentological and palaeontological grounds the Pecten bed can differ only slightly from the overlying calcarenite in age and both are the product of a single and continuous regressional episode. Although there are no sedimentological grounds for placing the Umkwelane Hill deposits into the same episode, similarities in the foraminiferal assemblages do suggest a correlation. Differences in these assemblages would appear to be environmental.

C. Age of the Zululand Neogene Deposits

Accepting these Neogene sediments as being the product of a single regressive cycle implies that they are contemporaneous. But it is the age of these deposits that has been most contested. King (1953) has attempted to date the deposits as Early Miocene by means of the mollusc fossils, even though he recognises the very high proportion of extant species, and the fact that the rate of organic evolution is slow in tropical waters. Molluscs are notoriously conservative and can seldom be used for precise dating. Biesiot (1957) has suggested an Early Miocene age on the basis of foraminifera he examined. But his determinations were made on benthonic forms which are the least reliable. Subsequent investigators (quoted in Frankel 1960a, 1960b, 1966, 1968) have also attempted to date these deposits by means of foraminifera and have used planktonic forms to suggest a Late Miocene age. Several points should be stressed however. Most of these determinations, it would appear, have been made from thin-section examination. This method is highly unreliable when used on Cenozoic foraminifera. Furthermore, recrystallisation of the rock is extensive. Most of the specific identifications are cited as sp.aff., which would hardly make them valid for age determination.

The only three firm determinations cited by Frankel (1960c) were Anomalina aegyptica (Lower Eocene or Palaeocene), Elphidium parri (Miocene) and Triloculina angularis (Aquitanian to Recent). The genus Elphidium is rare before the Miocene. More recently Orbulina universa has been recorded (Frankel 1969). O.universa is usually taken as post-Burdigalian, except in tropical Burdigalian rocks. King (1953) and Biesiot (1957) have both shown the tropical character of the Zululand Neogene fauna, so that even O.universa does not necessarily contradict a Burdigalian age.

A very important study is that of Dr R.P. Stapleton who has examined and identified 13 planktonic species which were separated from the rock by crushing (pers. comm.). According to Dr Stapleton the assemblage is restricted to the N-17 planktonic foraminiferal zone, and the age assigned depends upon the position of the Miocene-Pliocene boundary. Since this boundary is taken as 6 m.y. in this thesis, following Berggren (1971), the N-17 zone would be late Late Miocene. Maud and Orr (in press) have also suggested a Late Miocene age for the Richards Bay coquina and calcarenite, rather than an Early Miocene age, but concede that the fauna cannot be older than Middle Miocene.

Shark teeth, also somewhat unreliable as a means of dating, have been used by Davies (1964) and Applegate (1970). Applegate writes that there is no reason why the fauna, all from the coquina, should not have a Late Miocene age, but that it could certainly not be Early Miocene. Davies suggested a Middle Miocene age. The Megaselachus megalodon is the large form which would agree with a Miocene age.

On palaeontological grounds the coquina and calcarenite, including the Umkwelane Hill calcarenite, are all broadly time-equivalent. A Miocene age younger than Early Miocene is now becoming generally accepted. The nearly horizontal nature of the Miocene-Cretaceous unconformity shows that warping has not taken place in the Tertiary, and Frankel (1966) and Maud (1968) could find little evidence for substantial epeirogenic movement. Clearly these sediments must be attributed to eustatic sea level movement. There are several reasons why this eustatic movement must be of Middle to early Late Miocene age:

(i) this coincides with an ocean floor spreading discontinuity at 10 m.y. (Flemming & Roberts 1973);

(ii) the development of the Antarctic ice sheet to its maximum extent in the Late Miocene (Kennett et al 1974) would have caused a significant fall of sea level;

(iii) the warm climates of the Middle Miocene were conducive to phosphorite formation and expansion of tropical marine faunas (Bandy 1967; Bandy & Wilcoxon 1970; Tankard in press a).

The Zululand Neogene sediments can then be attributed to a sea level peak in the Middle to early Late Miocene, with a general regression through the Pliocene.

III. NATAL

The geology of the Natal coastlands is dominated by a sequence of aeolianites which extend to 100m below sea level (Maud 1968). Although Krige (1932), King (1962) and McCarthy (1967) have attributed these sediments to a single depositional cycle, Kent (1938), Parr (1958) and Maud (1968) have recognised several aeolian depositional cycles.

Frankel (1964) described calcareous sandstones at Burman Bush with a terrestrial gastropod assemblage. Foraminifera in the Burman Bush calcareous rocks are of shallow marine origin, and allegedly Pliocene-Pleistocene age (Frankel 1966). But these assemblages differ greatly from the older assemblages of the Uloa and Umkwelane Hill calcarenites. King (1966) and McCarthy (1967) have dated the Natal aeolianites as Pliocene on the foraminiferal evidence. But dating rocks of Late Tertiary and post-Tertiary age by means of foraminifera is unreliable since the forms could have evolved little in this time (Maud 1968). Maud's interpretation of these aeolianites in terms of Pleistocene sea level movements is undoubtedly correct. The Pliocene foraminifera (if they are really Pliocene) would have been blown inland from marine sediments exposed on the continental shelf as the sea regressed.

IV. EASTERN CAPE

The marine Cenozoic geology of the eastern Cape differs greatly from the Zululand geology, and lends itself to a different approach.

Detailed studies have been made of two quarries situated on an east-west ridge on Needs Camp Farm between King Williams Town and East London, at an altitude of 365m a.s.l. The lithology has been most recently described by Lock (1973). The lower quarry, 17m below the upper quarry, was thought to be of Danian age by Chapman (1916) and Upper Senonian to Maestrichtian by McGowran and Moore (1971). The upper quarry has usually been accepted as of Tertiary age (see Lock 1973, for discussion), although King (1973) has reinterpreted the palaeontology to suggest that these deposits are also of Cretaceous age. Lock refutes this claim.

The fossils from the upper quarry include (Lock 1973): *Voluta* sp., *Isognomon* cf. *gaudichaudi*, *Ostrea atherstonei*, *Glycymeris 'pilosa'*, *'Panopea' gurgitis*, and *Lima* sp. The *Glycymeris*, if it is the same species that Newton (1913) recorded from the Alexandria Formation, would be *G. borgesi* (Cox) rather than *G. pilosa* (Linn). *G. borgesi* is the extinct form and *pilosa* the extant Mediterranean form. Both of these species apparently occur at Uloa (Professor L.C. King, pers. comm.). Although Mollusca are generally conservative and unreliable as a means of dating, the assemblage would suggest a Neogene rather than a Palaeogene or Quaternary age.

Neogene features of the area between the Swartkops and Sundays Rivers have been studied in detail by Ruddock (1968) who has attempted to establish the Cenozoic diastrophic and eustatic histories. Cenozoic littoral deposits lie on a dissected and stepped platform. Ruddock follows Engelbrecht et al (1962) in restricting the term Alexandria Formation to the marine crystalline and sandy limestones and pebble beds of Tertiary age. The Alexandria Formation averages 3-9m thick. Discontinuous conglomerates occur at the base. The upper part of the limestones are coarsely porous and less recrystallised. Kinkelbos silts overlie the Alexandria limestones in places and form a wedge thickening from about 6m at the seaward margin to 37m at the landward margin. These silts have been attributed to shallow marine and lagoonal depositional environments, and their upper surface has been remodelled by Quaternary sea level fluctuation. The upper surface of the Alexandria Formation is frequently erosional, while a boulder horizon sometimes separates it from the Kinkelbos silts.

The marine component of the Alexandria Formation overlies a composite seaward-sloping surface (Ruddock 1968, 1972). The inland margin of this surface is situated at 305m a.s.l. in the Bathurst area, and at 284m near Uitenhage. Ruddock describes the Salt Pan escarpment which separates the Grassridge platform above from the Coega platform below. The Grassridge platform is also composite, consisting of an older outer part (slope 1:100) which was truncated by a sea which formed the younger inner part (slope 1:200).

There is very little evidence available to date these various units. Certainly comparison with the Zululand Cenozoic deposits would be insufficient to date the complex diastrophic history. In fact this diastrophic history is more closely parallel to that of the western Cape. Chapman (1930) dated the oldest limestones at Birbury (elevation 204m) as Upper Eocene. But Bourdon and Magnier (1969) were able to distinguish two micro-faunas: a reworked Lower-Middle Eocene one and a well-preserved Miocene one. Furthermore, the Eocene fauna is more truly marine than the Miocene one. Haughton (1925) mentions a Megaselachus megalodon tooth from Birbury. Middle Eocene-Middle Miocene shark teeth were identified by Dr S.P. Applegate (quoted in King 1972b). The writer collected shark teeth from the Birbury site in 1971, and submitted these to Dr Applegate. He reported (in litt. 26.X.71) that the "material from Birbury near Bathurst turned out to be Eocene. In this fauna is Otodus obliquus. Also there is Odontaspis macrota and probably Odontaspis cuspidata. There may be also several species of Odontaspis in this collection but more and hopefully better material will be needed to confirm this, particularly of the small teeth would be valuable. Professor King sent me similar material from Bathurst which I determined to be Eocene in age".

The Birbury deposits overlie the inner Grassridge platform, and would appear to date the original planation as Eocene. The fresh Miocene micro-fauna indicates the reworking of the original Eocene strata and truncation of the inner part of the Grassridge platform when the sea finally reached the 284m level. This interpretation is in agreement with that already suggested by Ruddock (1968, 1972). King (1972b) has shown a Pliocene age for some of the marine limestones on this platform.

Most of the molluscs described are from limestones overlying the Coega platform. Newton (1913) lists the following species (his designation in brackets where the name has been changed):

GASTROPODA

Bullia annulata Lamarck
Melapium patersonae Bullen Newton
Pirenella stowi Bullen Newton
Voluta africana Reeve

BIVALVIA

Cardium edgari Bullen Newton
Glycymeris borgesi (Cox) (= G. pilosa).

Isognomon gaudichaudi (Orbigny)(= Melina gaudichaudi)
Macoma orbicularis Sowerby
Notocallista schwarzi (Bullen Newton)(= Chamelea schwarzi = C. rogersi)
Ostrea atherstonei Bullen Newton (including O.redhousiensis)
Scissodesma spengleri Linn. (= Schizodesma spengleri)
Tellina sp. (= T. perna)
Tivela baini Bullen Newton
Venus verrucosa Linn.

Engelbrecht et al (1962) also list many of these species, but their species list includes several superceded generic names and many invalidated and superceded synonyms for the species names. The entire collection would have to be re-examined before the list could be used. Although several of the species listed by Newton are still extant, the assemblage as a whole can be taken to indicate a Neogene age. The following Neogene zone fossils could be suggested: Cardium edgari; Glycymeris borgesi; Isognomon gaudichaudi; Notocallista schwarzi and Tivela baini. Newton was careful to separate the Neogene fossils from the very different Quaternary fossils in his report.

King (1972b) has listed foraminifera derived from the Alexandria Formation which were identified by Dr L.R. Nolten of Royal Dutch Shell. Dr Nolton reports that the three markers Ammonia ammoniformis, Ammonia italica and Florilus victoriense indicate a probable Pliocene age, but notes that of these three only A. italica is common. This confirms the earlier interpretations of Ruddock (1968, 1972).

The dates Eocene, Miocene and Pliocene serve as a means of labelling the diastrophic history. Miocene tilting caused a transgression which carried the shoreline to the inner edge of the Grassridge platform, while the Salt Pan escarpment is probably a Pliocene feature created by further tilting and incision. Ruddock (1968, 1972) suggests that tilting has taken place about an axis situated about 33km inland. He documented the three episodes of seaward tilting and transgressive cycles as: Early Tertiary, Miocene-Pliocene and Late Pliocene.

V. MOSSEL BAY - BREDASDORP

Considerably less detail exists for the area between Bredasdorp and Mossel Bay. But even here the Bredasdorp Formation, which reaches a maximum thickness of 140-150m, has been recognised as a regressive sequence (Haughton et al 1937). The Bredasdorp Formation rests on a platform formed across rocks of the Table Mountain and Bokkeveld Groups. This platform dips to the south in the Riversdale district with a slope 1:160 (Wybergh 1919).

Silcretes cap remnants of a landsurface inland of the Bredasdorp Formation, and King (1963) realised the possibility of dating this surface by examining the conglomerates at the base of the formation. He was able to state (p. 264) that "pebbles of silcrete occur in the conglomerate present at several places beneath the limestone ..... The age of the limestone may thus determine whether the silcrete capped remnants belong to the early or late Tertiary cycle which here are crossing each other into the coastal depositional sequence ....." In a later publication King (1972) reinterpreted the data. "The basal limestones do not contain any such abundance of silcrete or ironstone fragments as might be expected if they rested upon the older Tertiary landscape (c f. Miocene of Uloa). Instead, all the field indications are that the limestones rest upon a marine modified continuation of the lower cyclic surface of presumed Miocene development. This then requires that the limestones are younger, perhaps of Pliocene age".

The Bredasdorp Formation consists of a discontinuous conglomeratic zone which averages about 1m thick at Die Kelders (chapter 4) and in the Riversdale area, although it may locally be as much as 6m thick (Wybergh 1919). Overlying the conglomerates are crystalline limestones and coquina, the petrography of which has been described by Siesser (1970). In the Riversdale area the crystalline limestones are frequently separated from the conglomerates by soft reddish sands (Wybergh op. cit.). These sands may be an alteration product of the limestones similar to an occurrence on the Cape Peninsula where Late Pleistocene aeolianites overlie reddish sands.

Little is known about the depositional history of the Bredasdorp Formation. According to Wybergh, the formation is largely aeolian, while Spies et al (1963) concede that the lower part must be marine. Two measured stratigraphic sections at Bredasdorp and Still Bay (Siesser 1972, figure 2) show only 2-3m of shelly limestone which contain the remains of molluscs, foraminifera, and shark teeth. This limestone reaches an elevation of approximately 93m in the Bredasdorp area. King (1972) gives the maximum inland extent as 106m a.s.l.

In the De Hoop valley ($34^{0}24'$S; $20^{0}19'$E) the marine limestones are 8m thick. They are conglomeratic at the base, but display a fining-upward sequence through gravels and coquina (Figure 12.4). The gravels and coquina contain oyster and gastropod fossils. The top of the marine unit is marked by cross-bedding typical of a shoreface situation, and well rounded discoidal pebbles. High-angle cross-bedded aeolianites overlie the marine unit. These aeolianites form a prominent east-west ridge, possibly marking an old shoreline, between Bredasdorp and the Breë River mouth.

The mollusc fauna (Wybergh 1919; Barnard 1962; Spies et al 1963 and De Villiers et al 1964) provides the only basis on which to hazard a guess at the age of the Bredasdorp Formation. Wybergh has suggested a Pleistocene age. The Mollusca include: Bullia annulata, Melapium sp., Pirenella sp., Voluta africana, Glycymeris "pilosa" (probably borgesi), Ostrea atherstonei, Notocallista schwarzi, Striostrea margaritacea and Tivela baini. This assemblage is identical in all respects to that of the Alexandria Formation described by Bullen Newton (1913) and Spies et al (1963) are correct in assigning the marine strata to the Neogene.

VI. SOUTH WEST AFRICA

Tertiary sediments occur as isolated pockets preserved in bedrock depressions between 40 and 160m a.s.l. north and south of Lüderitz-bucht (Haughton 1963). Kaiser (1926) described the geology in detail

Figure 12.4 Marine limestone with basal conglomerate exposed in the De Hoop valley, Bredasdorp district.

Figure 12.5 The Buntfeldschuh escarpment (27°35'S; 15°35'E) which terminates the Namib plateau. The plateau is overlain by Eocene volcanics.

and suggested a Middle to Late Eocene age on the basis of the faunas at Bogenfels and Buntfeldschuh. (This extensive fauna has been described in the same volume by Böhm). Both Kaiser (1926) and Greenman (1969) describe these deposits as being overlain by phonolitic lavas which erupted 35,7 m.y. ago (ZSA 53). Haughton (1963) has suggested that the cephalopod Aturia lotzi and most of the other molluscs are of Miocene age, and that in situ Eocene strata are preserved at higher elevations. The writer briefly examined these deposits in February 1975. At Buntfeldschuh the strata lie upon a wave-cut platform at approximately 120m a.s.l. The platform is overlain by gravels containing shark teeth, followed by 15m of marl with large concretions (lower part of Figure 12.5), then 5m of marl. Overlying the marls are horizontally bedded fine-grained quartzose sands (Figure 12.5). Planar cross-bedding cosets (Figure 12.6) are suggestive of shallow marine deposition. The marine strata are overlain by cross-bedded aeolianites (Figure 12.7). Shark teeth from the marine strata have been submitted to Dr S.P. Applegate for dating.

At Bogenfels marine molluscs occur in shallow marine sands and clays. Most common at the site examined were Ostrea subradiosa and Turritella kaiseri. The interesting aspect about this assemblage is that the oysters are "fresh" and still have a calcareous shell. But they occur amongst an older fauna typified by silicified and corroded shells. It would seem that Haughton (1963) was correct to suggest Miocene reworking of Eocene deposits. The Buntfeldschuh deposits are a coarsening upward regressive sequence and probably of Eocene age.

VII. CONTINENTAL SHELF

Cenozoic sedimentation on the continental shelf is concentrated in two large basins, the Orange Basin off the west coast and the Agulhas Basin off the south coast (Dingle 1973c). Sedimentation commenced with an Eocene transgression, while a major regression and erosion started in the Late Eocene and lasted throughout the Oligocene and into

Figure 12.6 Planar cross-bedding cosets in the Buntfeldschuh Tertiary succession suggestive of shallow marine deposition.

Figure 12.7 Aeolianites overlying the Buntfeldschuh marine deposits.

the Early Miocene (Dingle 1973c, 1974; Siesser *et al* 1974). Seismic records confirm this major hiatus. Basalt and trachyte plugs (Alphard Banks) have been radiometrically dated as 58 m.y. old.

The Tertiary rocks are broadly divisible into the Cape St Blaize Group (Palaeogene) and the Agulhas Group (Neogene) which are separated by the major mid-Tertiary hiatus. The Cape St Blaize Group (Upper Palaeocene to Upper Eocene) comprises calcareous clays, siltstones, sandstones, glauconitic-quartzose limestone, and glauconitic siltstones and sandstones (Dingle 1973a; Siesser *et al* 1974). The Agulhas Group (Early Miocene-Pliocene), on the other hand, comprises mainly foraminiferal-bryozoan limestones, and a great abundance of phosphorites. The Agulhas Group was deposited in moderately deep water. Pelletal phosphorite has formed by Late Cenozoic erosion. Generally the Palaeogene strata are thicker than the Neogene strata: in the Orange Basin 800m and 300m respectively, and in the Agulhas Basin 500m and 400m. The Neogene deposits of the west coast differ from those in the south coast in having a prominent intra-group unconformity (Dingle 1973a).

Several boreholes which were sunk into the Mozambique Basin from the *Glomar Challenger* in mid-1972 (JOIDES DSDP leg 25 - Geotimes November 1972) confirm the above history. At site 248 an unconformity spans the entire Oligocene from latest Eocene to earliest Miocene.

At site 249 two unconformities were recognised: Maestrichtian-Middle Miocene, and Late Miocene and Early Pleistocene.

In fact, the only strata allegedly of Oligocene age are *Pecten* bearing shelly limestones recorded in borehole J(c)-1 24km east of Stanger on the Natal coast (du Toit & Leith 1974). These limestones contain reworked Eocene sediments. Du Toit and Leith equate the unconformity beneath the Oligocene strata with that below the Pecten bed at Uloa, and with the Oligocene hiatus in the JOIDES DSDP boreholes.

CHAPTER 13

PLEISTOCENE HISTORY AND COASTAL MORPHOLOGY OF THE YSTERFONTEIN - ELANDS BAY AREA

I. INTRODUCTION

In the Pleistocene sea level fluctuated in sympathy with the repeated waxing and waning of northern hemisphere ice sheets, and palaeogeographic studies show that coastal lowlands all over the world have been subjected to periods of alternating submergence and emergence. In South Africa the Pleistocene history of the coast is related to such a sea level history.

Previous literature on the Pleistocene marine history of the coastal areas of the southwestern Cape is limited to a few texts, most of which are of a geomorphic nature. Krige (1927) published the first major study. He found evidence for two general emergences: shorelines at 15-18m he called the Major Emergence, and those at 6-9m the Minor Emergence which was thought to be of Recent occurrence. In the Saldanha area he identified the Major Emergence at 6-12m and proposed that the area had been downwarped. Haughton (1931) briefly described the geology and palaeontology. Mabbutt (1956, 1957) claimed marine terraces at 91m, 45-60m, 20m, 13m, 9m, and 7,5m a.s.l. (above mean sea level). All of the terraces above 9m he identified as marine terraces on the basis of flattish surfaces. The most detailed geological approach to the problem has been that of Parker (1968). Davies (1973) found evidence of transgressions to 9m a.s.l. (Eem) and 1,5m a.s.l. (Recent).

This chapter describes the results of an investigation into the Pleistocene history of the coastal area between Elands Bay and Ysterfontein (Figure 13.1). All altitudes are recorded as height above mean sea level (a.s.l.). Tidal amplitudes are shown in Figure 13.1. Mean wave height is 3m, and the highest recorded waves 9,5m (Pomeroy 1965). Ocean swell is predominantly southwesterly.

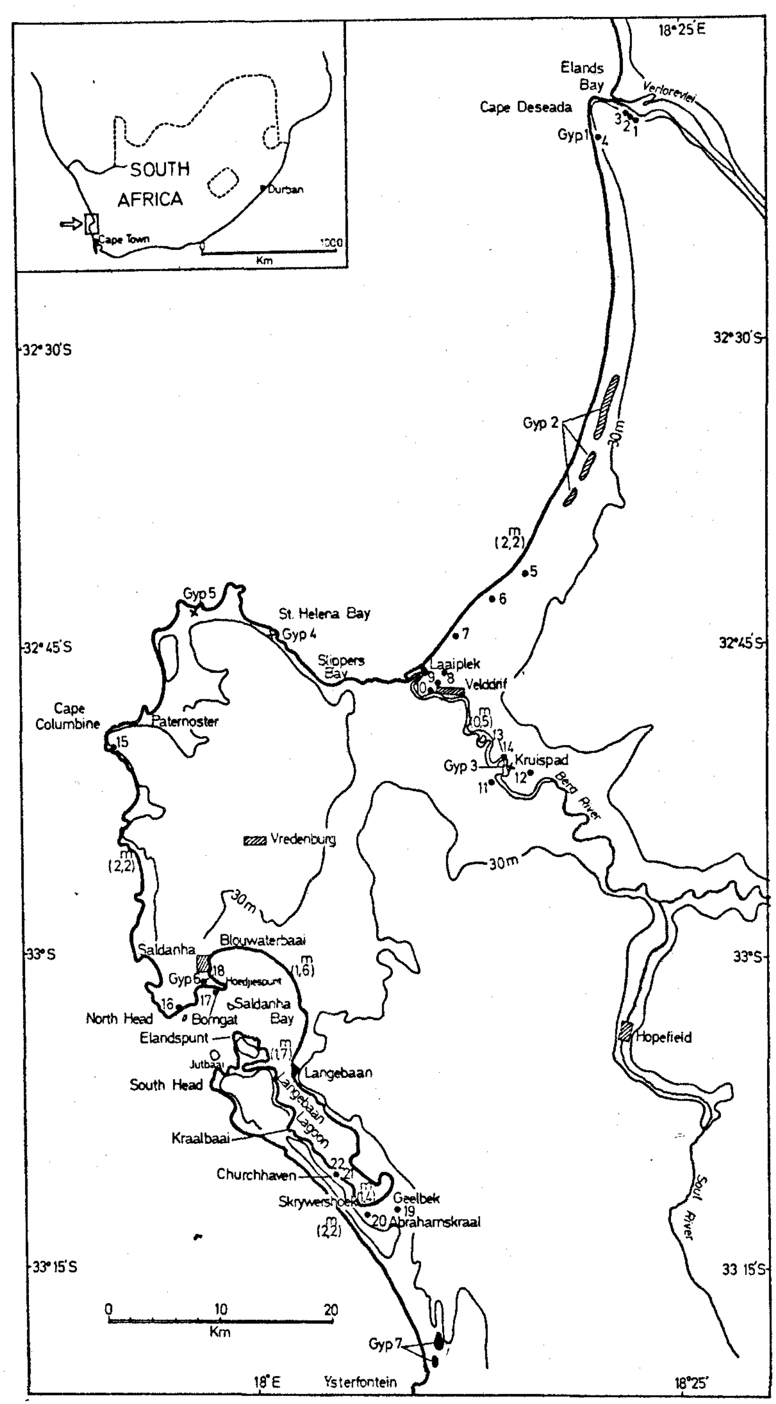


Figure 13.1 Locality map showing sampling sites (solid circles). Tidal range is shown in brackets. Evaporite deposits are referred to as "Gyp".

II. EARLY PLEISTOCENE MARINE DEPOSITS

Deposits of presumed Early Pleistocene age occur sporadically around Saldanha Bay: on either side of the Hoedjiespunt peninsula, Elandspunt, and 2,1 km north-east of Langebaan (Figure 13.1). The underlying platform on Hoedjiespunt is uneven, ranging in elevation from 6,2m to 8,9m a.s.l. At the Bomgat on Hoedjiespunt peninsula the Early Pleistocene deposit is separated from the granite floor by a lens of Miocene microsphorite which occupies a bedrock depression. The contact between the microsphorite and the limestone is an erosional one. On the South Head, at Elandspunt, the eroded uneven surface of the granite is at 7m a.s.l. The Early Pleistocene deposit north of Langebaan rests on the surface of an older marine limestone which is probably of Neogene age (not an aeolianite as thought by Davies 1973). Here the contact is at 9,5m a.s.l. A terrace at the same altitude may indicate a greater lateral extent of the Early Pleistocene deposits in the Langebaan area.

The relationship of the Early Pleistocene marine limestone at the Bomgat to the underlying microsphorite and the overlying Langebaan limestone is shown in Figure 13.2. The limestone contains wave-generated granite boulders at the base of 1,5m of coquina. The matrix consists of micrite which indicates a sheltered environment. Although articulated Perna perna would also indicate a sheltered environment, much fragmented debris is indicative of periods of higher energy marine conditions. Parker (1968) found that calcareous material constituted more than 95 per cent of the limestone.

The base of the marine limestone on the northern side of Hoedjiespunt peninsula (behind the Sea Harvest factory) is characterised by wave-generated granite boulders. The limestone consists partly of unconsolidated shell deposits, and partly of coquina and microcoquina. The marine limestone is overlain by approximately 20m of aeolian limestone. The microcoquina is a medium-grained skeletal lime grainstone with a drusy calcite cement. Detrital shell grains form in excess of 95 per cent of the rock and the remainder is quartz.

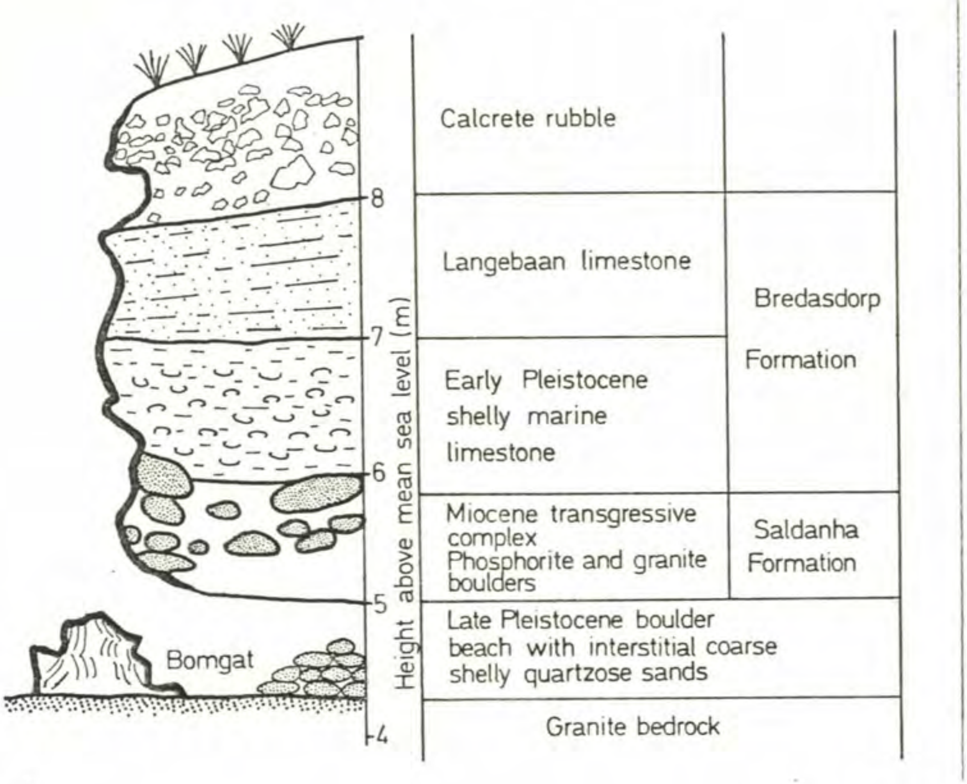


Figure 13.2 Geological section at the Bomgat at Hoedjiespunt.

Figure 13.3 Shelly limestone on the granite platform at Elandspunt.

The quartz and shell grains are generally well rounded but corroded. The shell grains include foraminifera and echinoid spine fragments. Although the limestone rests on a granitic platform, thin-sections show a total absence of felspar grains. This is characteristic of a high-energy beach environment.

Wave-generated granite boulders occur at the base of the coquina at Elandspunt. The coquina consists of much disarticulated and fragmented bivalve fragments, and less fragmented gastropod remains. The coquina, shown in Figure 13.3, is very different from Late Pleistocene deposits (Figure 13.9).

The Early Pleistocene mollusc fauna is shown in Table 13.1. The nature of the outcrop prevented a shell count being made at the Bomgat and Elandspunt sites. The fauna as a whole is indicative of shallow-water, mainly intertidal, conditions. Although Bullia annulata is commonly dredged today, it is sometimes found intertidally. All of the species of Patella are intertidal, except P. tabularis which is infratidal. Littorina knysnaensis is found most abundantly above HWS (high water spring). The barnacle Balanus amphitrite which still encrusts the limestone bedrock at the Langebaan site, is indicative of the infratidal zone.

TABLE 13.1

Early Pleistocene Invertebrate Fauna

	Bomgat	Sea Harvest	Elands-punt	Langebaan
GASTROPODA				
Haliotis midae Linnaeus		<1		<1
Fissurella robusta (Haughton)	X	30	X	2
Helcion cf. pruinosus (Krauss)	X	3		
Patella argenvillei (Krauss)		2		

TABLE 13.1 (Continued)

	Bomgat	Sea Harvest	Elands-punt	Langebaan
GASTROPODA (Continued)				
P. barbara Linnaeus	X	7		2
P. cochlear Born		2		
P. granatina Linnaeus	X			
P. granularis Linnaeus		1		
P. oculus Born		< 1		
P. tabularis Krauss		< 1		
P. variabilis Krauss		< 1		
Gibbula rosea (Gmelin)				< 1
Turbo cidaris Gmelin	X	19		2
Oxystele trigina (Chemnitz)	X	< 1		
Littorina knysnaensis Philippi				1
"Rissoa" capensis Sowerby		X		
Solariella undata Sowerby				< 1
Cerithidea bifurcata Kilburn & Tankard				7
Crepidula sp.	X	18	X	26
Thais cingulata (Linnaeus)	X			
Purpura praecingulata (Haughton)				< 1
Burnupena papyracea cincta (Röding)	X	2		3
Triumphis dilemma Kilburn & Tankard			X	1
Bullia annulata (Lamarck)				3
B. digitalis Meuschen				4
B. laevissima (Gmelin)				1
Nassarius capensis (Dunker)				1
N. scopularcus Barnard				2
N. speciosus Adams				11
Fusus sp.				< 1
Peristernia nassatula (Lamarck)				1
Marginella capensis Krauss				< 1
Marginella piperata Hinds				2

TABLE 13.1 (Continued)

	Bomgat	Sea Harvest	Elands-punt	Langebaan
GASTROPODA (Continued)				
Marginella sp.				<1
Cythara amplexa (Gould)				<1
Conus mozambicus mozambicus Hwass				1
BIVALVIA				
Aulacomya ater (Molina)		<1		2
Perna perna (Linnaeus)	X	3		2
Gryphaea sp.		<1		
Ostrea atherstonei Newton				<1
Mysella convexa (Gould)				1
Tellimya trigona Barnard				1
Carditella capensis Smith				<1
Thecalia concamerata Bruguière		<1		<1
Eucrassatella sp.				<1
Lutraria lutraria (Linnaeus)				2
Tivela tomlini Haughton		1		<1
Petricola prava Kilburn & Tankard		8		14
CIRRIPEDIA				
Balanus amphitrite Darwin		<1		4
Total individuals	-	172	-	220

The Hoedjiespunt molluscs indicate, in general, the close proximity of a rocky shore. But whereas the Bomgat assemblage is characteristed by abundant *Perna perna*, the Sea Harvest assemblage has fewer bivalves and abounds in patellids. Although rock-dwelling forms are common at the Langebaan site, *Patella* is generally absent. Sand-dwelling forms predominate there. Furthermore, species such as Cerithidea bifurcata and Thecalia concamerata are indicative of sheltered mud flats. The Elandspunt assemblage tabulated by Parker (1968) could not have come from the Early Pleistocene limestone.

Assigning these deposits to any age with confidence is difficult. Parker (1968) reports two ^{14}C dates for the Elandspunt site: 41 100 B.P. (Pta - 097), and greater than 49 500 B.P. (Pta - 098). Besides the fact that the ^{14}C technique is unreliable on carbonate of this age, there is also a distinct possibility that the molluscs were not collected in situ from the limestone. The number of extinct molluscs in the fauna seems to preclude a very young age. These include Fissurella robusta, Cerithidea bifurcata, Crepidula sp., Purpura praecingulata, Triumphis dilemma, Tivela tomlini, and Petricola prava. The Crepidula possibly originated from a C. porcellana-type ancestor (Kilburn & Tankard 1975). Stratigraphically the fauna resembles the 45-50m transgression complex fauna of the Namaqualand coast. It has in common Fissurella robusta, Purpura praecingulata, Triumphis dilemma, the large form (20cm) of Perna perna, and Petricola prava. The absence of Donax haughtoni and Striostrea margaritacea, 45-50m transgression complex zone fossils, is attributed to the higher energy environment and colder water conditions in the Saldanha area.

Carrington and Kensley (1969) have assigned an Early Pleistocene age to the Namaqualand 45-50m transgression complex. This is based on altimetric correlation with the Moroccan succession, the number of extinct species, the low degree of lithification, and the fresh appearance of the shells. They also note that this was the last warm water fauna.

III. THE 13m SHORELINE

At the southern entrance of the Elandsberg tunnel at Cape Deseada there is a prominent horizon of large wave-generated boulders banked against the foot of the cliff at 13m a.s.l. Another horizon of well rounded cobbles and boulders occurs beneath talus material on the southern shore of Verlorevlei at 13m a.s.l. No fossils are associated with these two horizons. This shoreline is pre-Late Pleistocene, since it is considerably higher than the Late Pleistocene sea level datum at 6m. Possibly it correlates with the 10m Early Pleistocene deposits of the Saldanha area. If younger than that, then the most likely correlation would be with the Namaqualand 17-21m transgression complex which Carrington and Kensley (1969) assign to the Middle Pleistocene.

IV. LATE PLEISTOCENE MARINE HISTORY

The geomorphology and mollusc fossils indicate several depositional environments. An open-coast facies includes rocky shores and exposed sandy shores. An estuarine-lagoonal facies is identified in the vicinity of Saldanha Bay-Langebaan Lagoon, Berg River and Verlorevlei. At Velddrif (Figures 13.1 and 13.8) lagoonal deposits are situated between breaker-bars of the open-coast facies. The mollusc fauna of the open-coast facies includes rock and sand-dwelling forms which still inhabit the adjacent coast. But the estuarine-lagoonal facies regularly contains several species which today live in tropical waters to the north.

Description of the marine deposits between Ysterfontein and Elands Bay will be discussed according to the dominant depositional environment which also coincides closely with a regional approach. The mollusc fauna has been described elsewhere (Tankard 1975a).

A. Open-Coast Facies : Rocky Shores

Formation of a platform results from waves breaking in shallow water and the movement of detritus across the rock surface. All bottom loss of energy takes place shoreward of the surf base which Dietz (1963) defines as the greatest depth where the waves begin to peak appreciably during storms. The depth of vigorous abrasion is thus 1,5 times the wave height. The abrasion platform is a function of the energy generated by the average waves. The depth of the abrasion base off the Cape Peninsula is 15m (B.W. Flemming in press). Wave energy is reflected from slopes greater than 30^{o} so that erosion is slow, while maximum erosion takes place on slopes less than 15^{o} (Flemming 1965). Examples of this are, firstly, the Chapmans Peak coast of the Cape Peninsula where a very steep slope and consequent reflection of wave energy is associated with little erosion. Secondly, in the Saldanha area slopes are low and vigorous abrasion has resulted in a well-defined platform.

In the Saldanha area mean wave height is 3m (Pomeroy 1965), suggesting that most vigorous abrasion will take place shallower than 4,5m. Davies (1973) attempts to identify Holocene platforms which are related to sea levels at 5,4m, 3,9m and 1,8m a.s.l. The platform on the seaward side of the Vredenburg pluton varies in altitude up to 10,7m (Visser & Schoch 1973), so that the relief is about 7m. It is likely that the platform was eroded during a single stationary sea level, and it is highly unlikely that short-term oscillations of sea level would have left a permanent record on a platform of low relief.

Table Mountain sandstone forms a natural groyne at Cape Deseada where structurally controlled abrasion has formed a narrow platform. The outer margin of the platform is 4,7m a.s.l., and the inner margin 6,5m a.s.l.

The most widespread platform is that bordering the Vredenburg pluton, broken only by the deep entrance to Saldanha Bay. At Hoedjiespunt bedded Miocene phosphorite lies in a bedrock depression with a contact at 5,25m a.s.l. This suggests that the platform may be an old feature that originated in the Miocene. Tankard and Schweitzer (1974) have shown that a similar planation surface at 7-8m a.s.l. at Die Kelders on the southern Cape coast is overlain by Neogene limestone. But Early Pleistocene limestones in the Saldanha area suggest that there the platform has been remodelled in the Pleistocene, and that it is a composite feature. "Fresh" stacks on the North Head platform possibly originated in the Late Pleistocene. The present sea does not appear to be cutting a notch or platform at mean sea level. Instead the rocky shore plunges vertically about 3-4m before flattening out onto a sandy bottom.

Parker (1968) has mentioned two beach ramparts which parallel the present coast on the North Head and South Head platforms. Both ramparts, or storm beaches, are composed of well rounded granite boulders with a matrix of coarse and comminuted shell. Slight imbrication of the boulders suggests that storm waves have played little part in forming these ridges. The outer rampart on North Head is a composite feature. It rises from the shoreline with a concave surface to a vegetated crest 7m a.s.l. It thus has a fossil as well as a modern component. The second rampart (Figure 13.4) peaks at 12,1 m a.s.l. and is approximately 40-50m behind the lower and outer rampart. The flat between them is underlain by coquina, and shelly sand typical of a foreshore accretion unit abuts against the inner rampart. Pans have developed behind the ramparts.

The crests of these ramparts, like breaker-bars, probably relate to MHWS, which at St Helena Bay is 0,75m above MSL. The two ramparts thus reflect sea levels at 6,25m and 11,35m a.s.l. The 7m rampart would most likely be of last interglacial age since it agrees with the Late Pleistocene datum, but the 12m rampart would be older, and of unknown age.

Shell deposits are extensive on the platform and mollusc assemblages from sites at Paternoster and North Head will be discussed in the next chapter (Table 14.1). Parker (1968) has listed the molluscs from other sites. A horizontally bedded shelly sand is exposed below the Paternoster lighthouse. The bedding takes the form of layers of complete shells separated by layers of comminuted shell. Whole shells constitute 25-30 per cent of the organic remains. Gastropods constitute 96 per cent of the assemblage. Most species are rock-dwelling intertidal forms. Only Patella compressa of the six patellids is infratidal. Conus scitulus and Cypraea algoensis are common.

Figure 13.4 Excavation through inner storm beach on the North Head platform. Elevation of the crest is 12m a.s.l.

Figure 13.5 Rhythmically and wavy laminated evaporite deposit at Cape Deseada.

The Paternoster deposits are typical of the marine sediments covering the platform. Rock-dwelling forms are always in excess of 90 per cent, and gastropods predominate. On the North Head platform at the entrance to Saldanha Bay occasional ($<1\%$) and stunted Ostrea atherstonei valves occur. There are also seal bones. Barnacles are common. Due to maximum exposure to the swell this part of the platform has a great amount of comminuted shell ($>95\%$). Beach rounded boulders and cobbles are common, as well as fragments of coquina.

These shell beds regularly occur up to 6m a.s.l. This and the outer rampart suggest a Late Pleistocene shoreline dominated by a 6,25m sea level.

Several localities have features which suggest lower sea levels than that at 6,25m a.s.l. Inside the Bomgat (Figure 13.2) there is an horizon of bedded wave generated granite boulders 0,7m thick resting on the cave floor at 3,5m a.s.l. The molluscs are typically rock-dwelling, open-coast forms. The Bomgat is an extension of a gully through the granite platform through which waves surge high above mean tide level. The cave deposit was probably formed above high tide level and would have been related to a sea level below 2m a.s.l. The fresh appearance of the molluscs and the cave setting permit the possibility that the beach could be a modern feature, related to exceptionally high seas.

There is an intertidal occurrence of beachrock at the northern end of Jutbaai. It is a medium-grained skeletal lime grainstone. The lath-shaped, well rounded shell detritus constitutes 95 per cent of the rock. The absence of felspar is characteristic of a beach environment. The shell and echinoid spine detritus shows a preferred orientation, and is drusy calcite cemented. Coquina is preserved at two further localities amongst beach boulders at 1,6m a.s.l. and 3,9m a.s.l.

B. Open-Coast Facies : Exposed Sandy Shores

Waves tend to break further from the shore with decreasing gradient so that there is also a decreasing amount of energy available for shoreline erosion (Flemming 1965). With loss of energy in crossing the shallow floor, coarse stirred-up particles are deposited and these lead to formation of breaker-bars (Holmes 1965). Since the slope of the sea floor adjacent to the Berg River is only 1 in 400, one would expect accretion units to develop. But this area is relatively sediment-starved since perennial runoff is restricted to the Berg River which contributes only a small quantity of sediment, and the Cape submarine canyon possibly acts as a sediment drain.

Late Pleistocene sandy shore deposits are most prominent between Slippers Bay (west of the Berg River mouth) and Cape Deseada. The present shoreline forms a smooth curve. At Cape Deseada, where the coastal plain is narrowest, aeolian sands tend to form a concave profile. Further south the coastal plain becomes more complex.

1. Evaporite Deposits

Gypsum deposits are frequently encountered on the flats behind the coastal dunes (Figure 13.1). Generally, these evaporite deposits are less than 6m in altitude and are believed to be mainly of last interglacial age. The evaporite deposits north of Dwarskersbos (Figure 13.1, Gyp 2) and Ysterfontein (Gyp 7) are probably still forming today since they are associated with salinas. (A salina is a modern salt pan on an arid coast). Visser and Schoch (1973) record several $CaSO_4 12H_2O$ determinations on these deposits. The gypsum content of the evaporite deposits usually exceeds 80 per cent. The gypsum usually occurs in uneven masses of fine grained texture. Elevations above sea level are as follows: Cape Deseada (Gyp 1) 4,5–5m; north of Dwarskersbos (Gyp 2) 3,5–4,5m; Kruispad (Gyp 3) 3,5m; St Helena Bay area (Gyp 4 and 5) 3m; Naval Academy (Gyp 6) 2m; Ysterfontein (Gyp 7) 2m.

The Cape Deseada occurrence (Gyp 1) was the only one examined in detail. Here the coastal plain is backed by 180m high sandstone cliffs. The sedimentary succession of the terrace from bottom to top is as follows:

(i) shelly calcareous sands overlying the sandstone platform;

(ii) rhythmically laminated (2mm units) carbonate-gypsum-halite units (Figure 13.5);

(iii) a thin layer of saline mud;

(iv) modern dune sands which also form extensive coastal dunes.

The shelly calcareous quartzose sands at the base are attributed to a transgression to 6-7m a.s.l. The mollusc assemblage is indicative of an intertidal sandy beach, except for Choromytilus meridionalis, which is a rock-dweller. Solen capensis suggests a sheltered environment which, adjacent to the present high energy beach, could only have prevailed in a lagoonal environment on the leeward side of a breaker-bar. An extensive emerged bar which once formed a barrier-beach follows the St Helena Bay coastline. In the Cape Deseada region a salina must have developed behind the bar. Flooding of the salina occurred at spring tides. Precipitation from the stranded brine led to formation of the evaporite. The evaporite is now about 0,75m thick and would have required a seawater column of 45m for its formation (by extrapolation from statistics quoted by Hsü 1972).

The evaporite contains a microfauna suggestive of a complex history of formation. The microfauna is summarised in Figure 13.6 and Table 13.2. The foraminifera, which constitute only 5 per cent of the assemblage, are all abraded. In contrast, the ostracodes (95 per cent of the assemblage) are thin shelled, show no signs of abrasion, and are frequently still articulated. All growth stages are also encountered. Candona sp., Pionocypris assimilus, and Cypridopsis are all fresh-water species. Aurila dayii suggests a more saline environment as found in estuaries and lagoons. Cyprideis cf. limbocostata is an inhabitant of brackish-water, salt lagoons and marsh environments. Species of Cyprideis have both smooth forms and nodose forms, but the number of nodose dimorphs decreases with decreasing salinity (Benson 1961). The smooth tests of Cyprideis

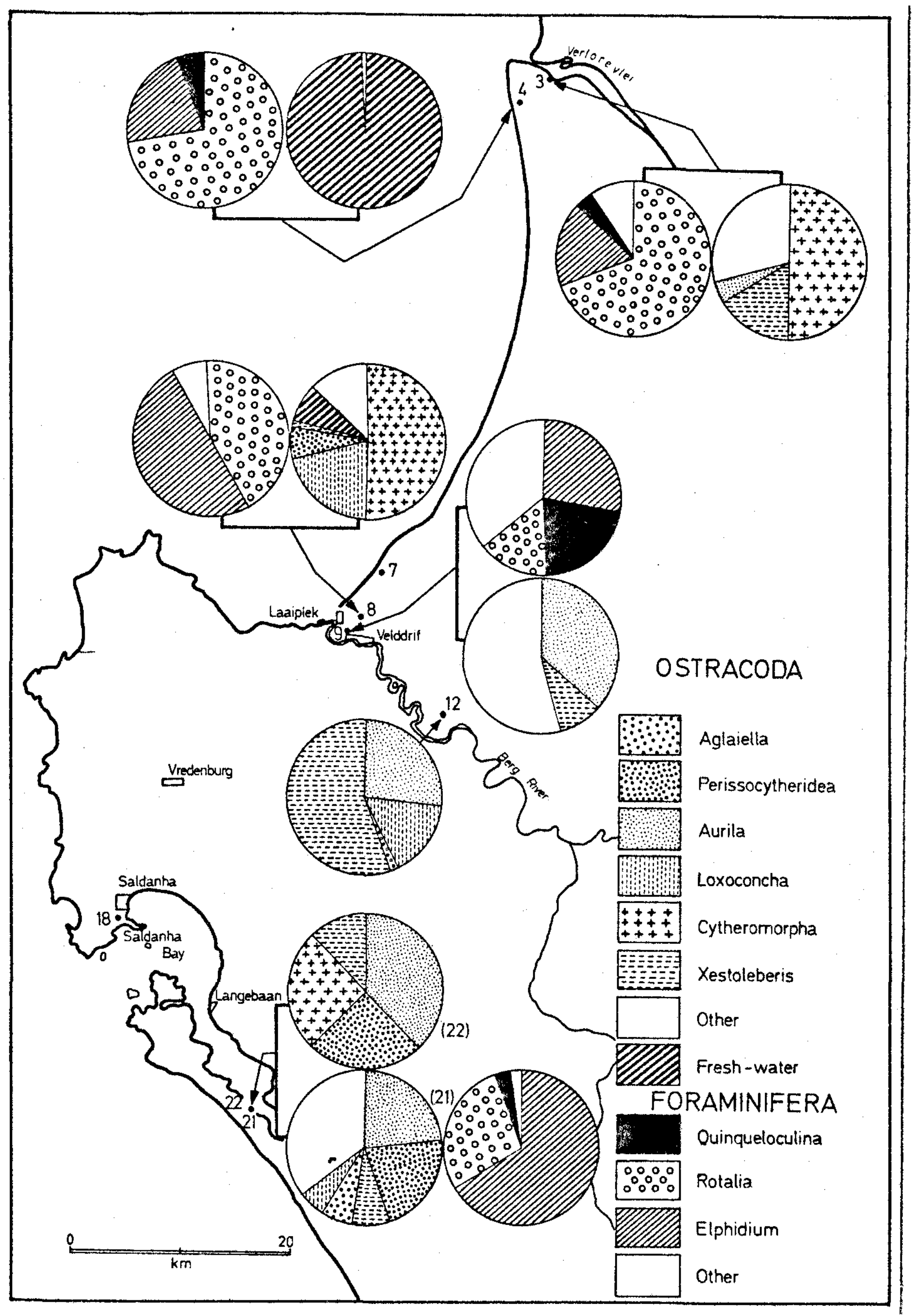


Figure 13.6 Distribution of ostracode and foraminifera assemblages.

TABLE 13.2

LATE PLEISTOCENE MICROFAUNA (%)

Sites	3	4	7	8	9	12	18	21	22
FORAMINIFERA									
Quinqueloculina spp.	3	6			21		X	X	X
Sigmoilina sp.									X
Frondicularia sp.	<1								
Lagena hexagona (Williamson)	<1				3	X		X	
Lagena semistriata Williamson	7			3					
Lagena cf. orbignyana (Seguenza)				<1					
Lagena spp.					2	X	X		
Bolivina variabilis (Williamson)	<1				23				
Bulimina sp.					7				
Rotalia beccarii (Linnaeus)	69	73	X	46	15	X	X	X	
Elphidium alvarezianum (d'Orbigny)	17				6				
Elphidium macellum (Fichtel & Moll)	1		X		21				
Elphidium spp.		21		50		X	X	X	
Cibicides cf. pseudoungeriana (Cushman)	<1								X
Cibicides spp.						X			
Virgulina cf. advena Cushman					2				
OSTRACODA									
Eucypris sp.				1					
Paracypretta ampullacea Sars				7					
Cypridopsis ochracea Sars		69							
Pionocypris assimilis Sars		25							
Candona sp.		2							
Paracypris westfordensis Benson & Maddocks				5					
Aglaiella railbridgensis Benson & Maddocks						1		7	
Cytheretta knysnaensis Benson & Maddocks					X				
Cyprideis cf. limbocostata Hartmann		2							
Perissocytheridea estuaria Benson & Maddocks				6				6	
Cytherura sp.				<1					
Hemicytherura parvifossata Hartmann					X				
Bairdia cf villosa Brady								1	
Aurila dayii Benson & Maddocks	4	1			X	27		22	38

TABLE 13.2 (Continued)

	3	4	7	8	9	12	18	21	22
OSTRAPODA (Continued)									
Caudites knysnaensis Hartmann					X				
Procythereis sp.								4	
Urocythereis sp.		1		<1				12	
Loxoconcha parameridionalis Benson & Maddocks	<1			<1			X	20	25
L. peterseni Hartmann				21		16		3	
Cytheromorpha sp.	49			49				7	25
Bradleya sp.	8			7				10	
Xestoleberis capensis G.W.Müller	17			1	X	56	X	8	12
Cytherella punctata Brady	21			1					

(X = identified, not counted)

cf. limbocostata confirm the fresh-water nature of the assemblage.

Since fresh-water ostracodes are scarce in waters more saline than $2^{o}/oo$ (Benson 1961), the Cape Deseada assemblage must reflect low salinities considering the abundance of these ostracodes. But gypsum will only precipitate from a brine at salinity of about 117 $^{o}/oo$, 3,35 times that of normal sea-water. The apparent contradiction between the ostracode evidence and the gypsum evidence suggests a complex history of sea-water inflow, evaporation to dryness, and fresh-water inflow (Figure 13.7).

Krumbein and Sloss (1963) discuss the origins and associations of evaporite deposits. Two conditions are necessary for the genesis of evaporites: a warm and arid climate with little fresh-water runoff, and a restricted body of sea water. Excessive evaporation at high temperature and little dilution by inflow of fresh-water raises salinity above that of the open sea, and ultimately leads to precipitation of salt. Calcium carbonate is the first to precipitate when evaporation halves the volume of water. Gypsum is next to precipitate at about $25^{o}C$, followed by halite after 85 per cent of the gypsum has precipitated.

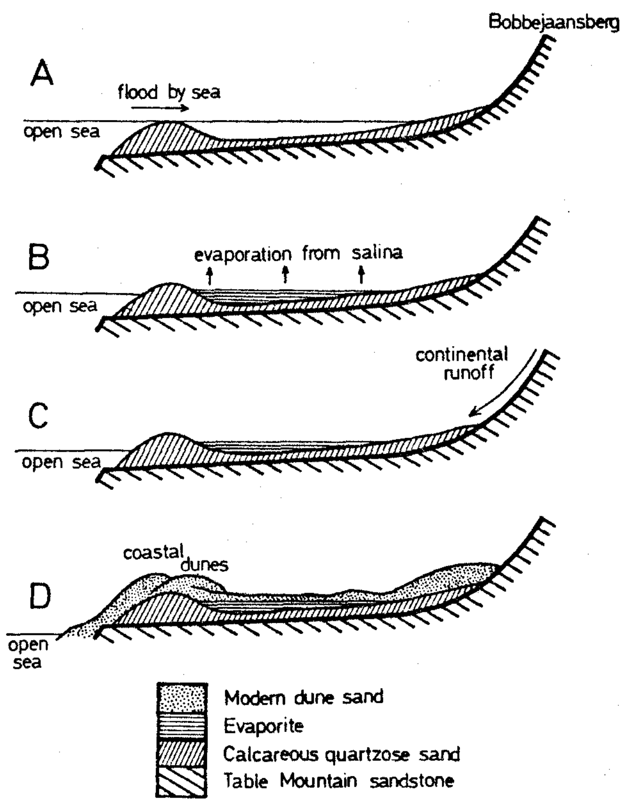


Figure 13.7 Summary of history of evaporite accumulation at Cape Deseada.

At Cape Deseada a restricted lagoon has developed behind an emerged breaker-bar in a warm and arid climate. The mollusc fauna of this transgression suggests a hydroclimate warmer than today (next chapter). Present-day rainfall is less than 200 mm/year.

Microscopic examination of evaporite specimens shows that each cycle begins with carbonate precipitation, and is followed by gypsum and frequently by halite. But the absence of halite in some of the units suggests inflow of more sea-water, probably tidal, before precipitation of gypsum from the previous brine had been completed. The precipitation cycle would be terminated at that stage and a new cycle initiated. Since each carbonate-gypsum-halite unit is, on average, about 2mm thick, it would have taken 300 to 400 tidal floodings of the salina (Figure 13.7 A-B) to accumulate the 0,75m of evaporite. The broken foraminifera tests suggest that they have come from the neighbouring high-energy beaches. Complete evaporation of the brine was occasionally followed by inflow of fresh-water, runoff from the sandstone hills, to produce fresh-water pans (Figure 13.7). This is suggested by the ostracode fauna of fresh-water affinity (salinity less than $2^{0}/oo$). The articulated valves and completeness of the thin-shelled carapaces suggest little agitation and the abundance of juvenile forms suggests that the population was probably suddenly exterminated, possibly by another tidal flooding of sea-water.

It is possible that the unique conditions of a saline water-fresh water cycle could also be satisfied by movement of sea water through the shelly breaker-bar, rather than over the top during high tides as suggested above. The sea-water would at some stage meet and dip beneath ground water (density control). Saline or fresh water conditions in the salina area would then depend entirely upon advance or retreat of the mixing zone due to ground water fluctuations. But this explanation would not necessarily account for the rhythmic sedimentation illustrated in Figure 13.5.

2. Barrier-Beach Coast

The entire length of the St Helena Bay coastline is marked by an emerged breaker-bar which, at some stage, must have formed a barrier-beach. Dune sands now accentuate the crest of this breaker-bar, and only south of Dwarskersbos are there any good exposures.

Northeast of Dwarskersbos (2,5km) surface exposures show the terrace behind the bar to consist of shell beds with entire valves in a predominantly detrital shell deposit with pebbles of Malmesbury rock. The shells are not evenly distributed throughout the exposure. Frequently shells of one species, e.g. Aulacomya ater, may occur together. The mollusc assemblage is characteristic of an intertidal sandy shore.

The assemblage is dominated by Crepidula capensis (46,7 per cent), Tellina trilatera (12 per cent) and Venerupis senegalensis (12 per cent). The predominance of undamaged valves of T. trilatera, a thin-shelled species, suggests relatively low energy conditions such as are presently found on the adjacent shore.

5,5km southwest of this site the mollusc assemblage is very different. Crepidula capensis is less dominant, Tellina trilatera is totally absent, and Lutraria lutraria and Argobuccinum argus make up nearly 70 per cent of the assemblage. Many barnacles, patellids, and Choromytilus meridionalis suggest a partly rocky shoreline. There are also many angular Malmesbury rock fragments. The Malmesbury platform is at 4m a.s.l.

The Malmesbury platform continues west of the Berg River, and is exposed behind the Varkvlei homestead where it is associated with rounded boulders at 6,7–7m a.s.l. (Visser & Schoch 1973). A deep excavation west of the homestead penetrated 4m of horizontally bedded shell deposits with little detrital quartz before reaching the Malmesbury platform at 1m a.s.l. The platform is mantled with wave-generated boulders. The shell bed rises gently inland to an elevation of 6m a.s.l.

Two breaker-bars are well developed in the Velddrif area (Figure 13.8). They are composed mainly of shell material with little detrital quartz, suggesting low sedimentation rates. Longitudinal excavations show

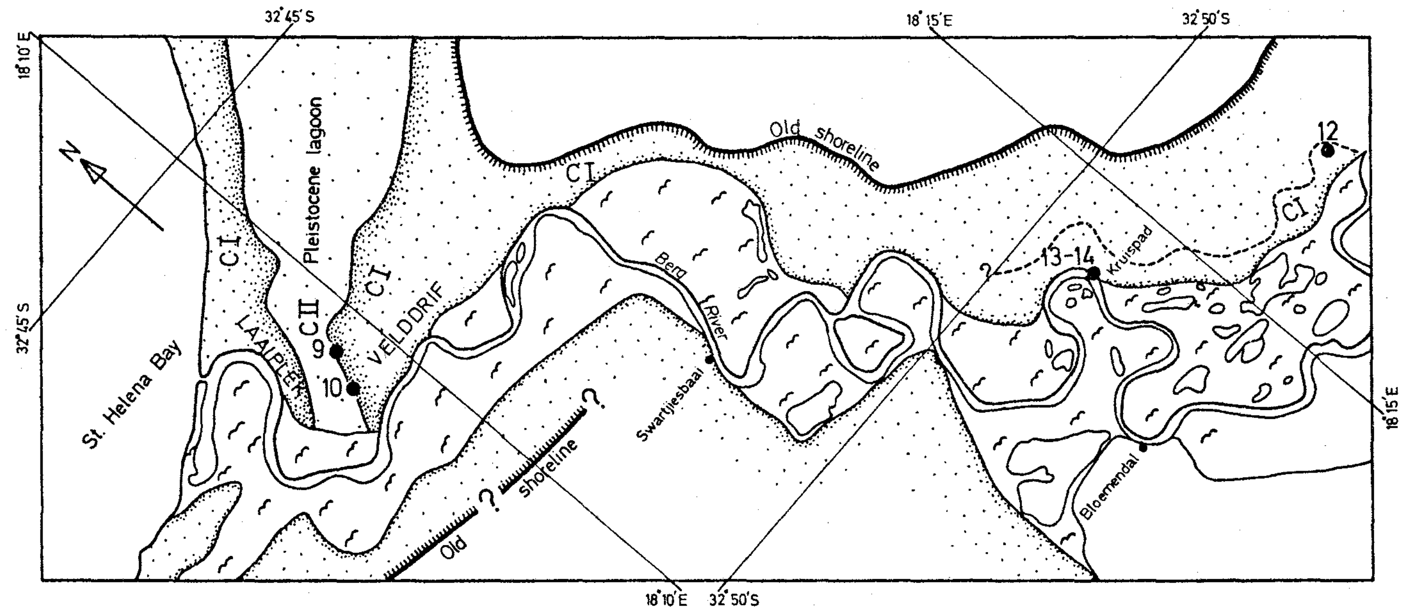


Figure 13.8 Late Pleistocene marine features in the Velddrif area. Note the two emerged bars which parallel the present coast. The broken line on Kruispad indicates the furthest inland extent of last interglacial shell beds.

subhorizontal bedding with alternating beds of whole shells and fragmented shells.

The mollusc assemblage of the outer bar north of Laaiplek is dominated by Crepidula capensis, and Venerupis senegalensis. The top 0,5m of the bar has a maximum of whole shells, and Venerupis senegalensis has a high articulation ratio (6,2). This horizon is underlain by 1,1m of comminuted shell. Scissodesma spengleri is common. Its present geographic range is False Bay to Algoa Bay (Barnard 1964), suggesting slightly warmer conditions in the Late Pleistocene. The assemblage suggests an intertidal sandy beach. Figure 13.9 shows the concentration of shell debris and articulated shells constituting the bar.

The seaward side of the second bar at Velddrif (site 9) is constructed of four basic units (Figures 13.10 and 13.11). The lowest unit is at least 0,5m thick (base concealed) and consists of horizontally bedded sand and shell debris. From 2,5m a.s.l. to 3,6m a.s.l. whole bivalves increase in proportion. The disarticulated valves are convex-up and have a preferred long-axis direction 220°. Then follows 1,4m of fine quartzose sand with several shell bands containing articulated Perna perna. Other bivalves are convex-up and have a preferred long-axis orientation 130°. The fourth unit from 5m–7m a.s.l., consists of a cross-bedded partially cemented shell deposit, with cross-bed azimuths 180°–220°.

Wave-energy appears to have been at a maximum in the lower unit where shell debris is broken down and finer quartz sand winnowed out. Incipient bar development to the west reduced the wave-energy and resulted in a greater amount of unfragmented shell. But the very thick shells of Venerupis senegalensis reflect the high energy conditions nearby. Further growth of the breaker-bar afforded a greater degree of shelter. Energy on the leeward side was insufficient to winnow away the fine quartz sand. Fragile Perna perna and Phaxas pellucidus lived in the sand. There are generally more juveniles in these sediments, and Venerupis senegalensis is thinner shelled. There was only enough energy to turn the shells over, but not enough energy to

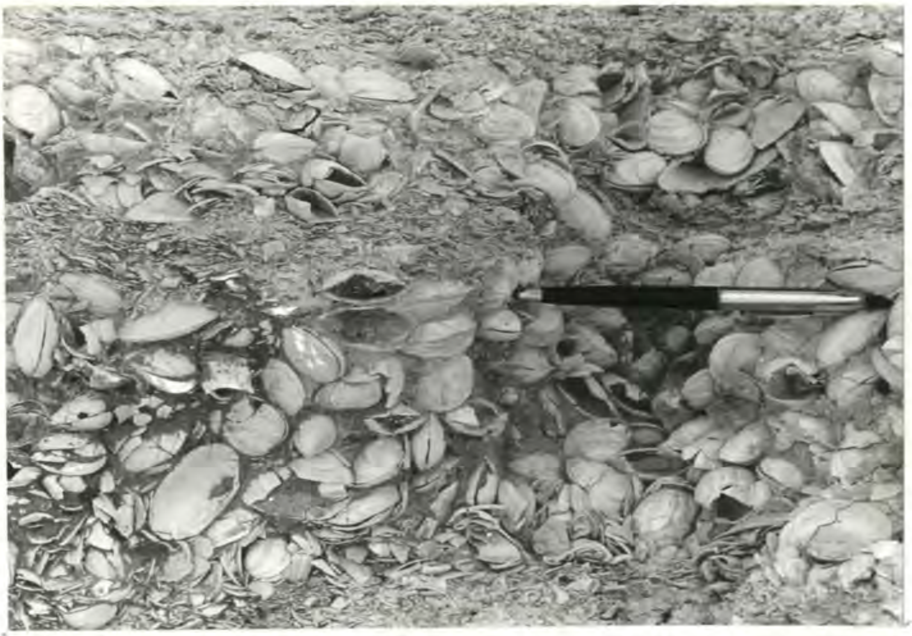


Figure 13.9 Shell accumulation which typifies the bar deposits. North of Laaiplek.

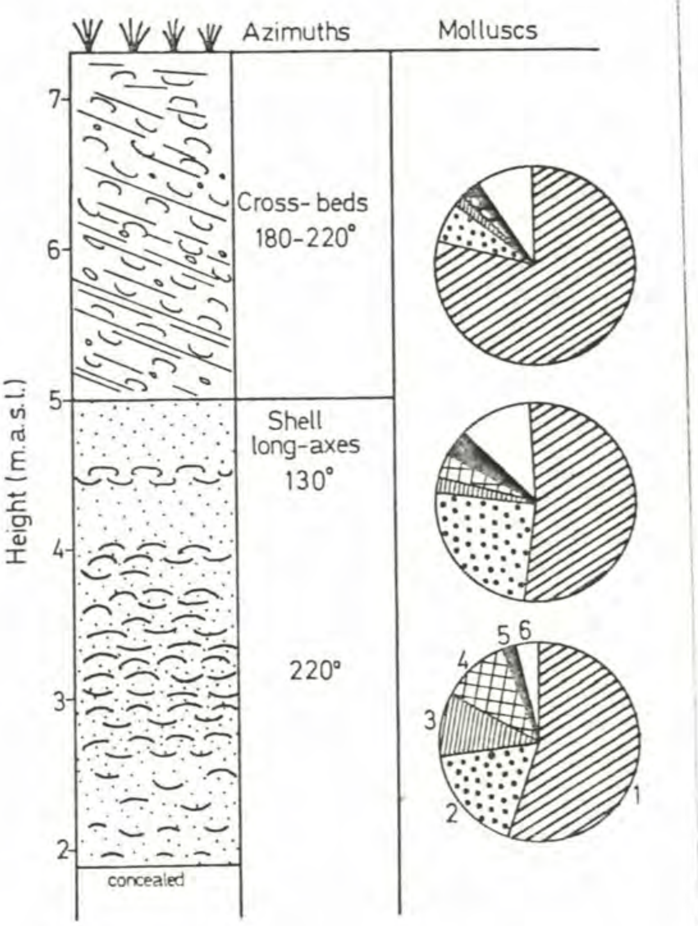


Figure 13.10 Measured section through the inner bar and wash-over fan at Veldrif. 1 = Venerupis senegalensis; 2 = Crepidula capensis; 3 = Bullia laevissima; 4 = Lutraria lutraria; 5 = Perna perna; 6 = others.

Figure 13.11 Bar exposure at Velddrif.

Figure 13.12 Convex-up bivalves in the quartzose sand unit of the inner bar. Note articulated Perna perna (lower left). Velddrif.

fragment them (Figure 13.12). Finally, the bar migrated shoreward over the lower units. Cross-bedding frequently dips 10°, but dips in the uppermost part of the bar are shallow and characteristic of a washover fan.

The fine quartz sand reflecting sheltered conditions was not necessarily brought across the top of the bar, but could have arisen from longitudinal transport on the leeward side and could have derived from tidal channels. Another excavation in the bar south of Velddrif shows small scale cross-beds suggestive of a tidal inlet.

Figure 13.9 shows the typical composition of the mollusc assemblage. Sand-dwelling forms predominate, and one of these, *Venerupis senegalensis*, always exceeds 50 per cent, and in the upper cross-bedded unit 81 per cent. Thin-shelled *Perna perna* are commonest in the sandy unit. There are also burrows in the sandy unit, possibly *Callianassa*.

The crest of the bar (7m a.s.l.) is a washover fan which probably relates to MHWS and can be correlated to a last interglacial sea level of 6,25m a.s.l.

C. Estuarine - Lagoonal Facies

Shorelines of the open-coast facies parallel the present shoreline very closely. Besides Saldanha Bay and its southerly offshoot Langebaan Lagoon, the Late Pleistocene sea extended up the Berg River valley and along Verlorevlei to form two extensive estuaries. Whereas the open-coast facies is continuous, the estuarine-lagoonal facies is laterally discontinuous.

1. Verlorevlei

Verlorevlei is situated in a northwest trending drowned valley which was formed along a fault plane by the ancestral Papkuils, Antonies, and

Kruis Rivers which now feed into the head of the vlei. Fossiliferous sediments are exposed on the steeper southern bank. The present vlei is 16km long, and a maximum of 1,5km wide. The present mouth of the vlei is marked by a bar of Palaeozoic sandstone at 1m a.s.l. A slight rise of sea level, as in the last interglacial, would effect considerable environmental changes, although the low bar at the mouth would still ensure a low energy regime. Recent bridge foundations showed that the deepest channel, and hence a previous mouth, was further to the north.

Geological sheet 3118 C/3218 A shows a sand covered terrace on the north bank to 60m a.s.l. A thick sand cover buries any evidence that may indicate a marine origin.

Late Pleistocene fossiliferous deposits (sites 1-3) are typically of a saltwater estuarine environment. Maximum thickness (site 3) is 2,2m where shelly sands lie unconformably upon Palaeozoic siltstones. The highest contact is at 5m a.s.l. where Ostrea stentina was found still adhering to the rock. The lowest 0,5m of the shelly sand consists of complete shells and valves, fragmented shell, and fine-grained quartzose sand. From 1,5m to 2,35m a.s.l. is a reworked horizon where the original deposit is diluted with coarse, iron-stained sand. From 2,35m to 3,8m a.s.l. the deposit consists of poorly sorted colluvium with scattered shell debris.

The mollusc assemblage (Table 14.1) consists of an infaunal bivalve element and a minor epifaunal gastropod element. The assemblage is dominated by Dosinia lupinus (50 per cent) and contains several thermophilic species such as Tellina madagascariensis, Loripes liratula, Macoma tricostata, Venerupis dura, etc. The assemblage suggests an estuarine sandy substrate. But Argobuccinum argus, Burnupena papyracea, Oxystele variegata, Patella spp., Ostrea stentina, indicate that the shore was in part rocky. The low numbers of rock-dwelling forms show that they were derived from a neighbouring environment, which may not have been extensive.

The ratio of foraminifera to ostracodes (Figure 13.6, Table 13.2) is

9 to 1, indicating an open exit to the sea. Cytheromorpha sp., which constitutes 49 per cent of the ostracode assemblage, can tolerate brackish water. Aurila dayii, Loxoconcha parameridionalis, Xestoleberis capensis, and Cytherella punctata, are today found in estuarine environments (Benson & Maddocks 1964; Hartmann 1974). They indicate saline water and a sandy substrate. Xestoleberis and Loxoconcha indicate the presence of marine grasses such as Zostera. The foraminifera are dominated by Rotalia beccarii (69 per cent) and Elphidium alvarezianum (17 per cent). Rotalia beccarii thrives in brackish water (Loeblich & Tappan 1964).

Estuarine animals generally colonise the area from mid-tide to low-tide (Emery & Stevenson 1957). Ostrea stentina, which encrusts the rock at 5m a.s.l., is always found infratidally. These deposits would thus indicate a mean sea level in the region of 6m a.s.l. or more.

The molluscs constitute a mixed life and indigenous death assemblage. Although some winnowing of juveniles may have taken place, the organisms appear to have lived in one broad environment reflected in the sediments (see next chapter).

2. Berg River

The Berg River has not developed a net of tributaries over the sandveld where infiltration capacity is high. In its lower reaches it meanders and gives rise to the usual features associated with meandering streams, e.g. oxbow lakes, meander scrolls. The last interglacial climatic peak was probably associated with aridity so that the Berg River did not flow as strongly as today, and a last interglacial transgression to 6–7m a.s.l. would have extended saline conditions far up the valley.

Fossiliferous deposits containing marine molluscs are encountered 15km up the Berg River on the farm Kruispad (Figure 13.8). The terrace contains evidence for still-stands of the sea at about 6m a.s.l. and 3,5m a.s.l. The furthest inland extent of the terrace (Figure 13.8) has been drawn from aerial photographs. The inner edge of the terrace is covered with dune sand and calcrete so that it is impossible to evaluate the significance of this shoreline. Mr D.S. Melck of Kruispad has determined the furthest inland extent of the shell deposits (Figure 13.8) by hand-augering.

A shallow excavation southeast of the homestead and 15km from the Berg River mouth, revealed a shell bed beneath 1m of dune sand. The surface of the shell bed is 5,1m a.s.l. The mollusc assemblage is typically estuarine, and contains several tropical west African species: Nuculana bicuspidata, Loripes liratula, Leporimetis hanleyi, Venerupis dura, and Panopea glycymeris. Mr S. Kannemeyer of the South African Museum identified the otolith of the sea eel Tachysurus fossor. Excavation shows Panopea glycymeris to be always articulated and in life orientation, and on the same plane 0,6–0,7m from the surface of the shell deposit, i.e. a siphon-length from the surface. Living Panopea glycymeris in the Mediterranean is always infratidal, and never extends into the intertidal zone. This suggests a sea level in the vicinity of 6m a.s.l. or higher.

The microfauna (Figure 13.6, Table 13.2) consists almost entirely of ostracodes, and the dominant species are Aurila dayii (27 per cent), Loxoconcha peterseni (16 per cent), and Xestoleberis capensis (56 per cent). They confirm a shallow estuarine-type environment. Loxoconcha and Xestoleberis suggest an abundance of Zostera.

The sediment is fine-grained quartzose sand with 8–10 per cent shell debris. Except for Panopea glycymeris few of the bivalves are articulated. Panopea is a deep burrower and would have been protected from disturbance. A dark layer (hydrotroilite) with corroded pollens is situated 0,25m from the surface of the shell bed. The pollens include many Gramineae, Chenopodiaceae, Cyperacae, Compositae (Professer E.M. van Zinderen Bakker, pers. comm.). The Gramineae (grasses) and the Compositae are very varied. Chenopodiaceae is a small plant that thrives in a salt marsh environment in relatively dry climates. Cyperaceae (sedge) indicates the presence of water. These pollens suggest that for a time the site was isolated from open circulation. Hydrotroilite ($FeS.nH_2O$) suggests reducing conditions which have resulted from a restricted circulation.

An excavation to a depth of 10m just inland of this site shows that the shell bed does not continue landward.

There is evidence in the river bank a few hundred metres west of the homestead of two separate transgressions (sites 13 and 14). The oldest unit consists of a semi-consolidated shelly sand with rounded quartz pebbles, 1m thick, with a surface 4,5m a.s.l. The mollusc assemblage which is leached, is characterised by an abundance of Mactra glabrata. Solen capensis occurs articulated. Loripes liratula is the only thermophilic mollusc. This bed grades upwards through silty loam to a dune sand. The shelly deposit is most likely a continuation of the 6m transgression complex.

A shell bed dominated by 'fresh' Dosinia lupinus is banked against this older unit up to 3,2m a.s.l., but extending below river level. At river level there is an extensive oyster reef 50m in length consisting of Ostrea algoensis. Nearer the homestead gypsum deposits indicate former saline conditions with little dilution by fresh water. The younger shell bed reflects a sea level in the vicinity of 3,5m a.s.l.

Other fossiliferous deposits are exposed on the south bank of the Berg River at Bloemendal (site 11) at 3-4m a.s.l. and near the bridge at Velddrif. The Bloemendal assemblage, which contains the thermophilic species Loripes liratula and Leporimetis hanleyi is suggestive of a shallow, sheltered environment. The bridge assemblage is a mixture of sand-dwelling and rock-dwelling forms: Patella sp., Choromytilus meridionalis, Dosinia lupinus, and Venerupis senegalensis. Well rounded cobbles are suggestive of a surf beach.

Withdrawal of the 6,25m sea left the barrier-beach and bar complex at Velddrif emerged, while a lagoon developed in the area between the bars. Aerial photography (Figure 13.13) shows that this area is an old lagoon floor with microrelief of shallow channels and sand-bars. Several pits into these fine-grained lagoonal sediments enabled the collecting of mollusc fossils (site 8). All the forms were small.

Figure 13.13 Aerial photograph showing relict Late Pleistocene lagoon floor between emerged bars at Velddrif and Laaiplek; trend north-east. (Aerial photograph reproduced under Government Printer's Copyright Authority 4603 of 2 November 1971).

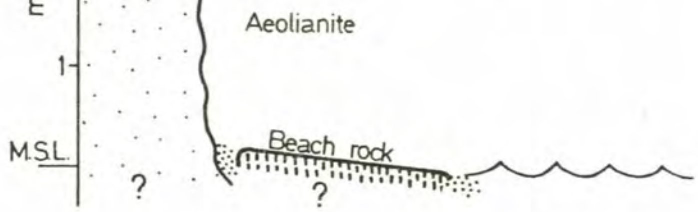


Figure 13.14 Typical section through Blouwaterbaai limestones.

"Rissoa" constitutes about 90 per cent of the assemblage. Nassarius kraussianus is abundant in weed-beds at low tide on muddy sands (Day 1969). Other fossils include crab (Brachyura) chelae, echinoid spines, bryozoan remains, sponge spicules, ostracodes and foraminifera.

The ratio of foraminifera to ostracodes is 34 to 1, indicating free access to the sea. The foraminiferal assemblage (Figure 13.6, Table 13.2) is dominated by Elphidium sp. (50 per cent) and Rotalia beccarii (46 per cent). Snider and Curran (1974) have found large numbers of Rotalia beccarii and Elphidium spp. to characterise the lagoonal environment. The large number of tests per unit volume of sediment indicate slow sedimentation rates. Carter (1951) has found that sorting governs the size of the specimens, so that the size of the foraminiferal tests would be similar to the sediment grain size. According to Krasheninnikov (1960, quoted in Loeblich & Tappan 1964) large numbers of Elphidium indicate mobile water. The tests of the Velddrif (site 8) specimens show no wear or evidence of transportation.

In comparison with other west coast sites the tests of Rotalia and Elphidium are small, and are present in large numbers. Phleger (1960) has found that under optimum conditions, and in large living populations, small size may be taken to indicate unusually favourable conditions and rapid reproduction.

Cytheromorpha sp. constitutes 49 per cent of the ostracodes, Loxoconcha peterseni 21 per cent, and Perissocytheridea estuaria 6 per cent. But 8 per cent of the assemblage comprises freshwater forms: Eucypris sp. (1 per cent) and Paracypretta ampullacea (7 per cent), suggesting periodic influxes of freshwater. Cytheromorpha can tolerate brackish water. Loxoconcha and Xestoleberis prefer saline water (Benson 1961) and suggest the presence of marine grasses.

These fossiliferous deposits between the bars are at 1,5m a.s.l. and must relate to a transgression about 2m a.s.l.

3. Saldanha Bay

Saldanha Bay, the only major inlet on this coast today, has a deep entrance (42m) which shoals to 2m at the mouth of Langebaan Lagoon. The bay probably originated by differential erosion of the different granite types and Malmesbury shale. (Drilling operations show the presence of extensive shale north and north-east of the shoreline). There is no evidence of faulting nor of an old river valley.

On the landward side of the Hoedjiespunt peninsula a low plain below 10m connects Noordbaai with Smitswinkelbaai. Fossiliferous deposits show that this area was probably flooded by the last interglacial sea. The surface of the shelly sands is only 1m a.s.l. (site 18). The dominant molluscs are Oxystele variegata (15,7 per cent), Clionella sinuata (19 per cent) and Tellimya trigona (26,1 per cent). The assemblage is composed of rock-dwelling and sand-dwelling forms, and indicates a shallow environment.

Limestones form a low cliff along the Blouwaterbaai shore (Figure 13.14). The oldest unit is an indurated aeolianite which is capped with calcrete. At 2,8m a.s.l. the calcrete is overlain by 0,6m of shelly marine limestone which contains angular cobbles of Malmesbury hornfels. The molluscs from this marine unit include rock-dwelling forms such as Patella spp., Burnupena papyracea, etc., and sand-dwelling forms such as Turritella capensis, Bullia digitalis, etc. They suggest an intertidal deposit. The youngest unit is a thin pavement of beachrock in the present intertidal zone, and reaches nearly to the foot of the cliff. Beachrock is formed in the intertidal zone.

4. Langebaan Lagoon

Extensive shell beds along the southern and western shores of the lagoon show that the lagoon is a feature which existed at least by the Late Pleistocene. The lagoon is a north-westerly trending body 15km long and a maximum of 4km wide whose present channels are maintained by strong currents generated by the summer south-easterly winds. Where the eastern shore is controlled by granite topography the last interglacial and present shorelines probably coincide. But extensive shell beds on the southern shore on the farms Geelbek, Abrahamskraal, and Skrywershoek, demonstrate a more southerly extension in the Eem.

The lagoon is separated from the Atlantic Ocean by a long ridge that connects Ysterfontein with the South Head granites. One can only surmise that it developed as a prograding spit and tombolo or a barrier-beach complex. Whatever its origin, it was reinforced by dune sand accumulation. The present lagoon is the result of Flandrian flooding of a pre-existing dune landscape.

The Geelbek deposits are a lagoon floor shelly limestone accumulation reaching an elevation of 2m a.s.l. and overlain by calcrete. The coquina includes quartz porphyry cobbles. Many of the bivalves are still articulated. The mollusc assemblage is indicative of a shallow, sandy substrate, and contains several thermophilic molluscs (Table 14.1), for example Loripes liratula, Tellina madagascariensis, Venerupis dura.

At Churchhaven a shelly limestone has been locally planed at 1,2m a.s.l. (Figure 13.15). Thin-section examination shows the limestone to consist of 90 per cent quartz and 10 per cent detrital shell with drusy calcite cement. All the detritus is fine-grained and well rounded, and both quartz and shell are extensively corroded. Most of the shell grains are molluscan, with only occasional echinoid spines and foraminiferal tests. Figure 13.16 which shows the growth of calcite scalenohedra on the inside of an ostracode carapace (Urocythereis sp.), demonstrates the extensiveness of diagenesis in these deposits.

A layer of mollusc shells contains many articulated shells and unbroken valves. They are predominantly shallow water, sand-dwelling forms. Aurila dayii and Aglaiella railbridgensis are common ostracodes (Figure 13.6, Table 13.2). Ichnofossils include crab-burrows (Figure 13.17), and some 2-3cm diameter vertical burrows, probably Ophiomorpha which is attributed to the marine decapod Callianassa.

Figure 13.15 Planation platform on marine limestone at Churchhaven.

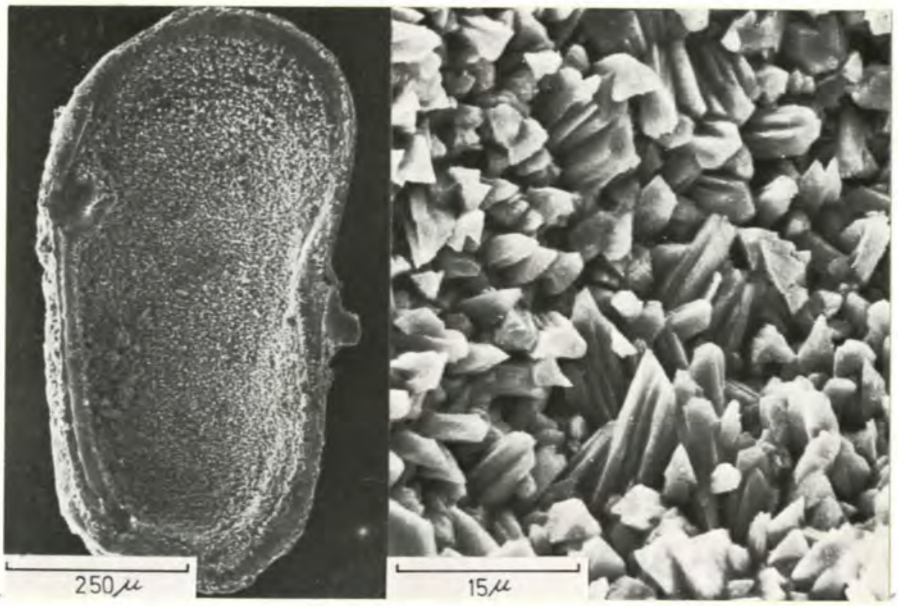


Figure 13.16 Calcite scalenohedra in inner surface of ostracode carapace demonstrates extensive diagenesis in the Churchhaven limestone.

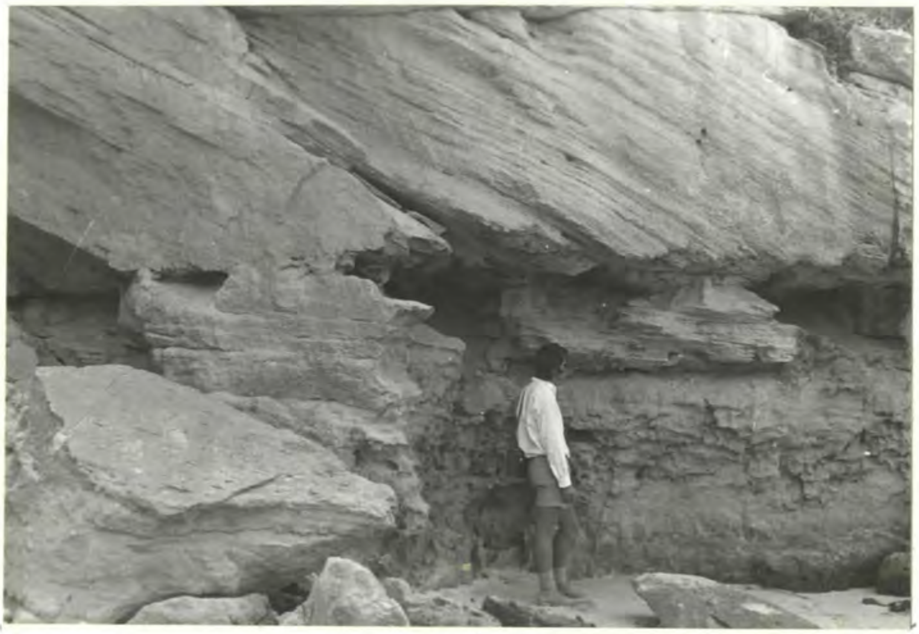
According to Mr B.W. Flemming (pers. comm.) the erosional platform at Churchhaven is part of an extensive platform which he has traced for 100m off Kraalbaai. At Kraalbaai there is a contact with the marine limestone and overlying aeolianite at 2,5m a.s.l. (Figure 13.18) with bivalve remains at the contact. Ophiomorpha is common in the marine limestones at this site. A very interesting trace-fossil that indicates that the top of the marine limestone at Kraalbaai is partly non-erosional is the series of footprints of a terrestrial mammal, probably the strandwolf Hyaena brunnea (Figure 13.19). The preservation of these footprints, the depth of the print, and the detail of the print, suggest a wet surface. Ridges of sand on one side of each footprint demonstrate that the animal was walking on a slope. Together with ripplemarks on the same plane this suggests a wet beach.

That the platform was eroded by another transgression is shown by the sedimentary succession at Churchhaven (Figure 13.20). The lower marine limestone platform is overlain by dune sands in which at about 4m a.s.l. there is an outwash deposit. The outwash consists of cobbles derived from the lower limestone along with mollusc shells. Excavation showed that the platform extends beneath the outwash. The outwash deposit is composed of 90 per cent cobbles and 10 per cent shells (Parker 1968). The shells are mainly disarticulated.

Parker (1968) recognised two distinct units at Churchhaven,which he separated according to lithology and the mollusc fossils. He recognised that the lower unit was not a shoreline feature, but a lagoon floor deposit. He believed the outwash deposit to be a 4,3m shoreline. Davies (1973), on the other hand, recognised only one bed into which a 1,5m transgression incised.

The outwash deposit occurs at varying altitudes in this region, that at 4m a.s.l.being the highest exposure. Pie diagrams comparing the mollusc and ostracode assemblages from the two horizons (Figures 13.6 and 13.20) confirm Parker's interpretation. But the field evidence shows that this shelly conglomerate is not an in situ marine unit, but rather a colluvial deposit that has gravitated down a dune surface. It must therefore reflect a shoreline higher than 4m a.s.l. Although

Figure 13.17 Crab-burrows associated with marine limestone at Churchhaven.



. Figure 13.18 Contact of marine limestone and aeolianite at Kraalbaai.

Figure 13.19 Mammalian footprints at the surface of the marine limestone at Kraalbaai, and detail. Note claw marks in photograph on the right.

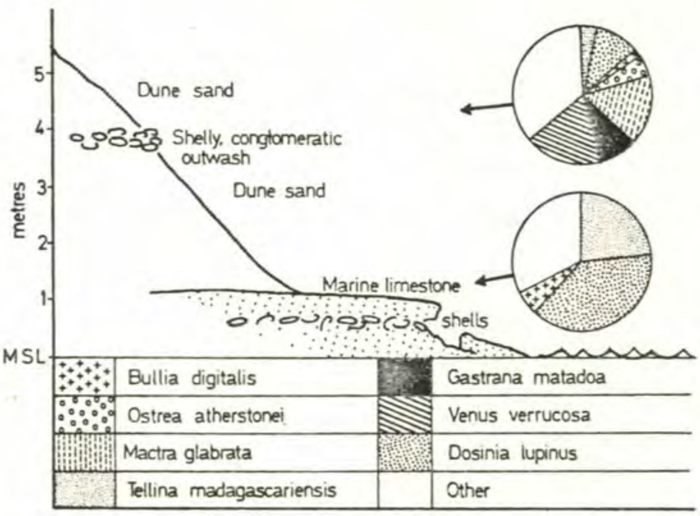


Figure 13.20 Sedimentary succession at Churchhaven showing lithology and mollusc distribution.

the higher beach material (outwash) may well contain fossils derived from the lower limestone unit, its mollusc assemblage does appear to be distinct. Furthermore, the shells are fresher.

The conglomeratic colluvium at Churchhaven implies a more vigorous abrasion than occurs along the shores of the present lagoon. Two explanations can be suggested. The first assumes that the peninsula separating the lagoon from the ocean owes its present extent to dune accumulation during the last glacial lowering of sea level. (Evidence will be cited shortly to substantiate this claim). The width of the tombolo or barrier-beach would have been considerably less than at present. The ratio between the present lagoon width to barrier width is 2:1, whereas the world average is 6:1 (Tanner 1960). A sudden rise of sea level, as suggested at Churchhaven, could have partially breached the barrier, or pushed it back, and reworked the lower limestone. The mollusc and ostracode fauna in the colluvium demonstrates that sheltered lagoonal conditions did exist at some stage during this sea level rise.

An alternative explanation could be that the breach of the barrier took place further north where Kraalbaai is connected with the ocean by a low area. But it is doubtful that this would have created the vigorous abrasion at Churchhaven. Besides, this gap may be the last remaining evidence of more extensive breaching or flooding of the original barrier.

For several years oyster deposits (Ostrea atherstonei) have been commercially dredged from the floor of the lagoon where there are reputed to be reserves of 3 million metric tons. East of Kraalbaai they form a layer 1-3m thick 5,5m beneath the lagoon surface. Oysters generally settle on a hard substrate and it is possible that initial colonisation took place on the limestone platform. Ostrea atherstonei valves are common in the Geelbek and Skrywershoek deposits as well as among the colluvium at Churchhaven, but rare in the older limestone. The lagoon floor oyster deposits would most likely correlate with these last interglacial deposits which in turn are correlated with a climatic peak at that time (Tankard 1975).

D. Age of the Deposits

Assessing the age of these deposits depends upon the assumption that they are primarily of glacio-eustatic origin, and that they can be regionally correlated. Despite the evidence for substantial down-warping in the Pleistocene, there is good reason to believe that tectonism has been minimal in the last 100 ka or so, and that shore-lines below +7m can be correlated.

Table 13.3 summarises the available ^{14}C dates pertaining to former sea levels between present sea level and +7m between Muizenberg (False Bay) and the Berg River. All the dates except Pta-461 and G r N-5878 are greater than 20 ka.

TABLE 13.3

Radiocarbon Determinations on Samples from the Area between False Bay and the Berg River

Sample No.	Locality	Altitude (m)	Material Dated	Date (B.P.)
Pta - 094	Kreeftebaai	3 - 5	shell	40 200 ± 130
- 095	Luisterhoek	3 - 5	shell	48 200 + 2600 - 2200
- 096	Churchhaven	1,5	shell	48 500 + 3600 - 2900
- 097	Elandspunt	6,5	shell	41 100 ± 1200
- 098	Elandspunt	6,5	shell	>49 500
- 461	Jutten Point	6 - 7	shell	2 070 ± 50
- 794	Kruispad	5,1	shell	43 700 + 4300 - 2700
- 796	Kruispad	5,1	shell	39 000 ± 1860
GrN - 5803	Melkbos	4	shell	43 200 + 2000 - 1500
- 5878	Langebaan Lagoon	-1 (?)	shell	6410 ± 45
I - 8372	Milnerton	1,5	shell	33 750 ± 1780
-	Muizenberg	Intertidal	Beach-rock	25 860 + 1040 - 1190
-	Muizenberg	Intertidal	Beach-rock	25 430 + 1050 - 1210

(References: Parker 1968; Vogel 1970; Vogel & Marais 1971; Davies 1973; Pta-794 and Pta-796, Dr Vogel pers. comm. 30.1.73; I-8372, Dr B. Kensley pers. comm; Muizenberg beachrock, Siesser 1974).

Mörner (1971) discusses some of the difficulties of dating shell material older than 15ka. Miniscule contamination of shells older than 20ka will give completely spurius dates (Fairbridge 1971). Glacio-climatic evidence suggests that any interstadial glacio-eustatic sea levels close to present sea level or higher are unlikely (Fairbridge 1971; Mörner 1971; Thom 1973). All of the dates shown in Table 13.3 greater than 20ka would be pre-last glacial.

Only two dates (Pta-461 and GrN 5878) are comparatively young. The date of 2070 B.P. (Pta-461) was derived from Patella argenvillei which appears to have been surface collected from a heavily vegetated rampart at 6-7m a.s.l. (Davies 1973). Even if reliable this date would hardly date the rampart; Davies mentions the "high throw" of waves along this coast in connection with the ramparts. The other date (6410 B.P.) was derived from an Ostrea atherstonei valve dredged from the southern end of Langebaan Lagoon. This date would appear to be too young because sea level would only have recovered by 5,5 ka (Godwin et al 1958). Furthermore, 3 million metric tons of shells implies a lengthy period of accumulation and optimum conditions for reproduction which would be better satisfied in the last interglacial. Flandrian transgression sediments and their mollusc fossils at an equivalent depth below present sea level show that water temperatures in Saldanha Bay in the mid-Holocene were no warmer than at present.

Available ^{14}C measurements indicate that the marine deposits are older than the range of this technique, and suggest a pre-Weichselian age. The mollusc fauna indicates an Eem age. Composition of the fauna compared with older Pleistocene faunas suggests a comparatively young age. Abundance of thermophilic forms correlates best with the warm isotopic substage 5e peak of deep-sea cores (Shackleton 1969) which occurred at 100-120 ka. Chappell (1974a) points out that the widely recognised shoreline between 2 and 7m a.s.l. in areas remote from plate boundaries, and dated as 120-130 ka, is becoming widely used as a Late Pleistocene sea level datum.

The higher shoreline up to 6,25m a.s.l. would thus date to about 120 ka, although the lower shorelines nearer present sea level may be younger. In Die Kelders cave on the south coast wave-generated boulders are preserved up to 2m a.s.l. in an erosional strike-passage in steeply dipping Table Mountain sandstone. Occupation of the cave by Middle Stone Age people which followed soon after withdrawal of the sea, serves to date the boulder bed as pre-Weichselian, but not much older. Since the boulder bed would have arisen from wave surges through the passage, it probably reflects a sea level close to the present and would correlate with the intertidal beachrock at Muizenberg described by Siesser (1974). An immediately pre-Weichselian age, about 80 ka, is suggested for these lower units.

V. RECENT

Figure 13.21 summarises the stratigraphy revealed in an excavation 100m from the Saldanha Bay shoreline at 18°E. Cross-bedded units of variable direction in medium and fine-grained sands with articulated Perna perna, and suggestive of shoreface sedimentation, form the lowest bed from -4m to -3,4m. This passes upwards into horizontally laminated fine sand and coarse sand with Perna perna, a typical fining-upward beach sequence. There is a gradual coarsening of the sediment to -1,4m suggestive of a temporary regression. Shells are more fragmented, and the bivalves disarticulated. The assemblage is an intertidal sand-dwelling one. A period of pedogenesis resulted in calcrete formation. The calcrete is again overlain by a fining-upward shelly sand due to a final transgression.

The molluscs include: Turritella capensis, Burnupena papyracea, Bullia digitalis, B. laevissima, Clionella sinuata, Petricola bicolor, Perna perna, Scissodesma spengleri, Lutraria lutraria, Donax serra, and Venerupis senegalensis.

Drilling operations on the coastal flats adjacent to Saldanha Bay have shown the widespread occurrence of shelly sands and shelly limestones, with intercalated calcrete horizons, below sea level datum. Furthermore, several boreholes into the bay floor north of Hoedjiespunt penetrated a peat at -22m which was preserved beneath calcrete (Mr A. de la Cruz, pers. comm.).

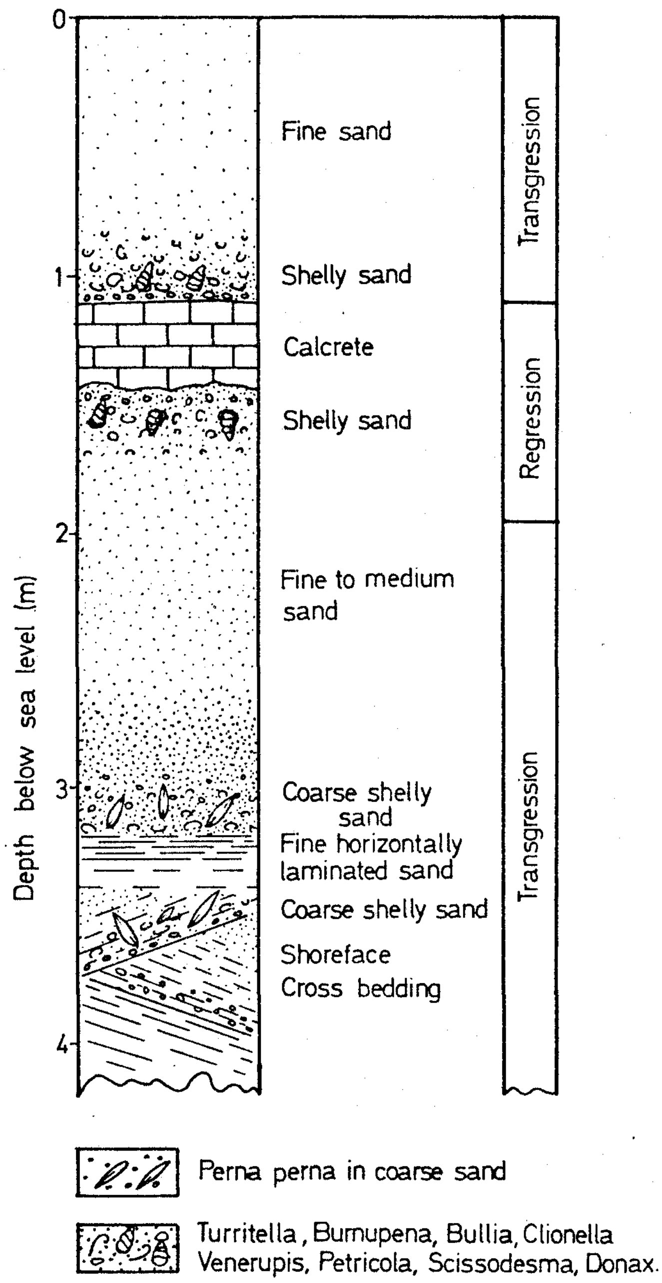


Figure 13.21 Flandrian transgression sedimentary succession at Saldanha. Exposure in a pit dug by ISCOR as part of the harbour works.

VI. FLUVIAL SEDIMENTS

At several localities from Cape Town to Verlorevlei there is evidence of considerable fluvial activity in the past. These sediments are presumably of Pleistocene age, although extensive deltaic sediments on the continental shelf between Dassen and Robben Islands have been assigned to the Lower Tertiary (Dingle 1971).

A. Baards Quarry

Uneconomic phosphate deposits occurring on Muishondsfontein are totally different from the phosphates of the Varswater Formation. Most exploration was done by trenching, the trenches having long since been back-filled. Most of my conclusions are based on Chemfos records, although there are samples in the collection of the South African Museum from Baards Quarry. Cut into greenish white clayey sands are channels of clayey sand with phosphatic sandstone (Figure 13.22). The depth of the channels is up to 1,8m. Boné and Singer (1965) describe the phosphatic sandstones as "phoscrete", but this term has the disadvantage of implying a duricrust-type origin. Some excavations have shown that the phosphatic sandstone occurs in places as a thin depressed bed 12–20m thick. Boné and Singer have described the phosphatic sandstone at Baards Quarry on Muishondsfontein as a "hard compact mass of phosphatised sand consisting of sand grains cemented and partially corroded by an amorphous calcium phosphate cement ..... The phoscrete and nodular phosphate sands represent replacement of older shelly sands by phosphate solutions, the phosphate of which is probably derived from guano". Thin-section examination of phosphatic sandstone from Baards Quarry (Museum collection) has shown no trace of phosphatised shell material. The phosphatic sandstone is composed of rounded, poorly sorted, quartz grains ranging in size from 0,4 to 0,04mm set in a matrix of finely divided argillaceous material, with organic carbon, the whole of the matrix being phosphatised. Many of the quartz grains bear phosphatic haloes. The sand has a packing proximity 10–30 per cent. The excavations by Chemfos Limited on

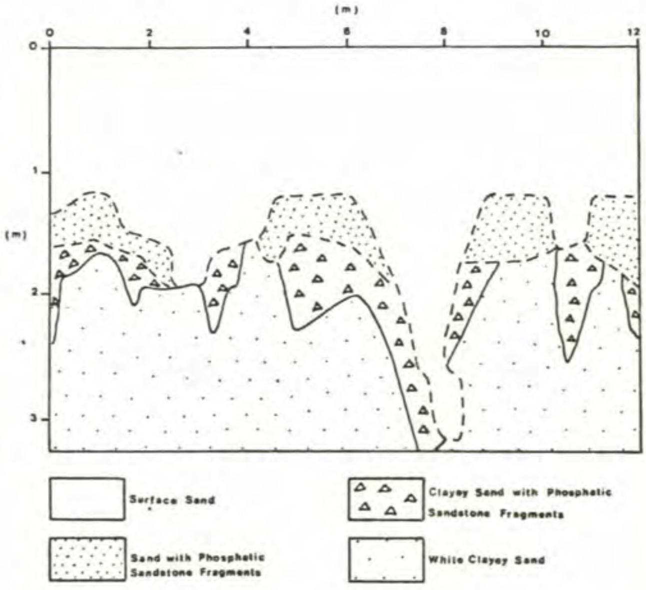


Figure 13.22 Lithology of Early Pleistocene fluvial deposits on Muishondsfontein.

Figure 13.23 Relict fluvial sediments at the head of Verlorevlei.

Muishondsfontein show the phosphates to occur as channel deposits which are north-south orientated.

The Baards Quarry fossils come from two distinct horizons. The younger occur in the surface sands (Figure 13,22) which are unconsolidated sands with duricrusts. The older assemblage comes from the clayey channel sands. These assemblages are distinctive in appearance and preservation. According to Hendey (pers. comm.) both units contain teeth of Equus which appear to represent two separate species. Hipparion spp. teeth occur only in the channel sediments. All the Baards Quarry equids apparently postdate those from the Varswater Formation where only Hipparion is represented, this Hipparion being less evolved than the Baards Quarry Hipparion. The Baards Quarry channel sediments have only two species in common with the Varswater Formation: Mammuthus subplanifrons and Mesembriportax acrea, both of which have a wide time range. Hendey suggests that the Baards Quarry channel sediments are of Makapanian age (Lower Pleistocene) and the surface sand assemblage younger (Cornelian or Florisian).

The Muishondsfontein channel deposits were mineralised by phosphate drawn from the Varswater Formation. Nearly spherical well-rounded cobbles and boulders of riverworn quartz porphyry are frequently found on the surface in this area, while riverworn cobbles can be identified in the Baards Quarry material. The Muishondsfontein deposits occur at 30-35m a.s.l.

C. Other Fluviatile Deposits

The planed surface of the Malmesbury Group at 30m a.s.l. east of the Sout River is an old river terrace and is mantled with angular gravel (the farms Maatjesfontein and Hamburg in the Hopefield area).

At Ysterplaat (Cape Town) an exposure of cross-bedded, rounded quartz gravel is suggestive of a point-bar deposit, while opposing dips laterally separated through the section suggest a meandering stream. Another exposure through relict fluvial sediments occurs at Redlinghuys

(Figure 13.23), at the eastern extremity of Verlorevlei (26km from the mouth). Here at least 18m of river sediment occupy an old channel with a thalweg less than 6m (concealed) above vlei level. The sediment is predominantly fine-grained, iron-stained, quartzose sand, with lenticular, gravelly, channel lag deposits which show some imbrication of the pebbles.

Extensive fossiliferous deposits demonstrate that the ancestral Berg River valley cutting was pre-Eem. Consolidated fluvial gravels and bog-iron-ore are exposed in the bank at Swartjiesbaai. Weichselian lowering of sea level to -130m lowered the controlling base level of the Berg River. Visser and Schoch (1973) mention four boreholes for the Velddrif bridge foundations which intersected at least 4,3m of conglomerate beneath a 5m cover of shelly sands which probably originated during the Flandrian rise of sea level.

VII. AEOLIANITES

The coastal plain is covered with a veneer of dune sands and aeolianites. Du Toit (1917) referred to the aeolianites as *Dorcasia* limestone, while Visser and Schoch prefer the name Langebaan limestone since the common fossil land-snail shell is *Trigonephrus globulus* and not *Dorcasia.* These authors present a lengthy discussion of the aeolianites.

Several exposures suggest that the aeolianites may range in age back to the Neogene. In a quarry 4km north of Saldanha high-angle cross-bedded aeolianite overlies marine Bredasdorp Formation rock of probable Miocene age. Behind the President Jetty on Saldanha Bay a well lithified, fine-grained aeolianite underlies younger aeolianites.

Overlying the Varswater Formation is a shelly-quartzose sand of aeolian origin. Shell fragments of the land-snail *Trigonephrus* are common. These sands are usually capped with a brown sandy soil overlying calcrete. On the southern part of Langeberg these sands are 41m thick, and in the Witteklip-Sandheuwel area 15-20m thick. In this latter area they have a regular horizon of ferricrete at a depth of about

8m indicating wetter conditions. Post-depositional diagenesis has produced a layer of aeolian phosphatic sandstone ("phoscrete") on Witteklip-Sandheuwel and Langeberg.

On Langeberg Tankard (1974a) has taken the contact of the Pleistocene dune sands and the marine Varswater Formation as the 2 per cent P_2O_5 cutoff, but Wolff et al (1973) have actually recorded this interface as an unconformity. In the vicinity of the New Varswater Quarry the top of the Varswater Formation is marked in places by burrow structures, while teeth of Hipparion namaquense show not only the terrestrial environment but also confirm a Pleistocene age for the dune sands.

Grading analyses are shown in Figure 6.10 where it will be observed that there is a tendency towards positive skewness. The quartz grains are generally rounded to well-rounded,and frosting is common. Calcrete rock fragments brought to the surface during coring illustrate the widespread occurrence of this rock. At various levels the quartz grains are iron stained.

Also present in some horizons are multigranular aggregates. These are rounded aggregates of fine grained quartz cemented with lime. Some of these have been rounded. Also frequently present with these aggregates are rounded calcareous fragments.

An interesting feature of these sands is the microfauna they contain. Abundant foraminifera represent two genera, Anomalina and Elphidium (5:1). Often the foraminifera are in a fresh state but frequently they are worn by wind abrasion. All of these are unphosphatised. Fragments of echinoid spines on the other hand are frequently phosphatised. It is interesting that in the Varswater Formation the majority of foraminifera identified were Elphidium. Near the top of M6-15-N6-15 the -1,5 Ø fraction is composed entirely of rounded shell fragments. In all boreholes the heavy mineral content of the aeolian sands is always considerably less than in the Varswater Formation.

The fact that the microfauna is almost entirely unphosphatised, save for a portion of the echinoid spines, and that the foraminifera are essentially fresh Anomalina as opposed to the weathered Elphidium tests in the underlying Pelletal Phosphorite Member of the Varswater Formation suggests that these sands were blown inland subsequent to the regression of the sea in which the Varswater Formation was deposited. The abundance of land-snail shell fragments (Trigonephrus sp.) along with a marine fauna demonstrates the close proximity of a shoreline.

Most of the aeolianites appear to be much younger, and probably accumulated during the last glacial lowering of sea level when vast tracts of unvegetated sand lay exposed on the emerging sea floor. Several occurrences of interbedded outwash suggest that the climate was not necessarily arid. Cross-bedding azimuths at Kraalbaai (mean 360^o) show that a southerly wind was prevalent. Rogers and du Toit (1909) describe a borehole at Paternoster which penetrated sandy limestone containing land-snail shells and tortoise bones to 21m below sea level. Tankard and Schweitzer (1974) have described the former seaward extension of dunes in Die Kelders area, and also mention in situ Middle Stone Age artefacts in the outwash deposits which shows that there the aeolianites must be Weichselian or younger Furthermore, limestone blocks in the cave show that lithification of the aeolianites occurred prior to about 6000 B.P. Similar inter-bedded outwash deposits containing Middle Stone Age artefacts occur as far north as Saldanha. A single ^{14}C assay on ostrich egg shell from a midden in the limestones behind the Sea Harvest factory gave an age greater than 40 000 years (UW 282).

An interesting feature of the coastal limestones is the solution-pipe cavities which have formed by karst weathering. At Churchhaven and Velddrif they penetrate the last interglacial marine limestones and shell beds. The pipes are 0,3-0,8m in diameter and vertical. At Churchhaven they are marked by a 3cm thick lithified crust. All of the pipes are filled with non-calcareous terrestrial terra rosa type sediment. Blackburn et al (1965) described an identical situation from Australia and suggested that they arose from solutional weathering and the filling of the voids with sediment by simple gravitation. The terra rosa sediment together with the karst weathering and colluvial deposits suggests a wetter climate in the southwestern Cape in the last glacial age.

Extensive Late Stone Age shell middens overlie the Langebaan limestone on the North and South Heads. On the South Head the midden is in part calcretised.

Aeolian phosphatic sandstones are also common. Besides the limited "phoscrete" horizons within the Pleistocene aeolian sands, there are larger bodies of aeolian phosphatic sandstone. North of Elands Bay a north-south trending ribbon of aeolianite, apatite cemented, lies banked against the western face of Table Mountain sandstone hills (Figure 13.24). The aeolianites are extensively dune-bedded and lie exposed between 120 and 150 m a.s.l. In the Darling-Mamre area similar aeolianites are known to exist, and high P_2O_5 values on the geological sheet demonstrate their extent. X-ray diffraction data show that the cementing material is a carbonate fluorapatite. The aeolianites contain much fine shell-debris and are, in fact, very similar to the "phoscrete" already mentioned. The shell material has been phosphate replaced.

VIII. PROPOSED NEW LITHOSTRATIGRAPHIC NAMES

The Bredasdorp Formation comprises a variety of limestone types of marine and aeolian origin. It includes marine limestone with conglomerate and coquina horizons, and beds consisting mainly of whole shells. The major part of the formation consists of dune sand and "shell grit and locally grades into pure shell-beds; it has been calcified into rocks varying from crumbly calcareous sandstone to hard, crystalline limestone ..... Solution channels are often encountered in the limestone" (Spies *et al* 1963). Lithologically this description describes the marine and aeolian limestones and shell beds of the Saldanha area. Although locally the limestones may be indistinguishable lithologically, it is usually possible to separate the aeolian and marine components. Two members of the Bredasdorp Formation are here defined:

(i) Velddrif Member: the type-section is the bar exposure at Velddrif (Figures 13.10 and 13.11). Lithologically it

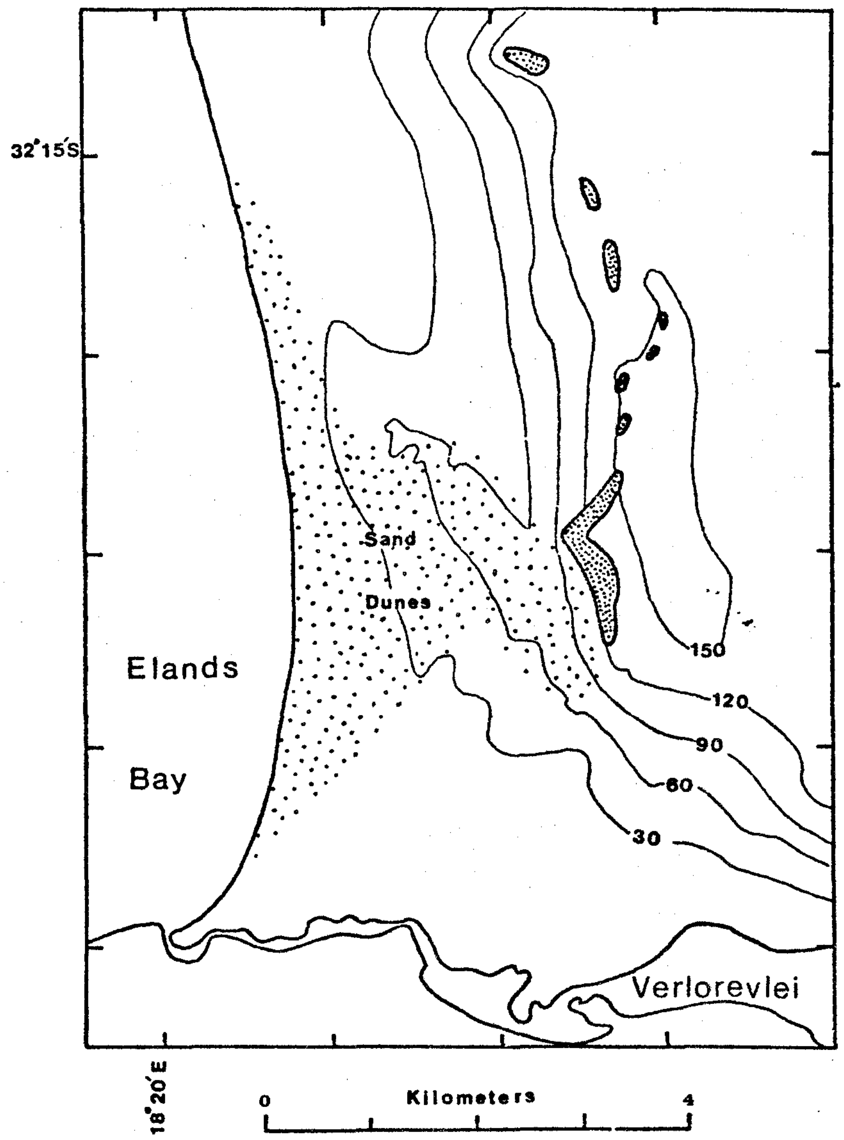


Figure 13.24 Ribbons of phosphatised dune sands ('phoscrete') at Elands Bay where they are banked against sandstone hills.

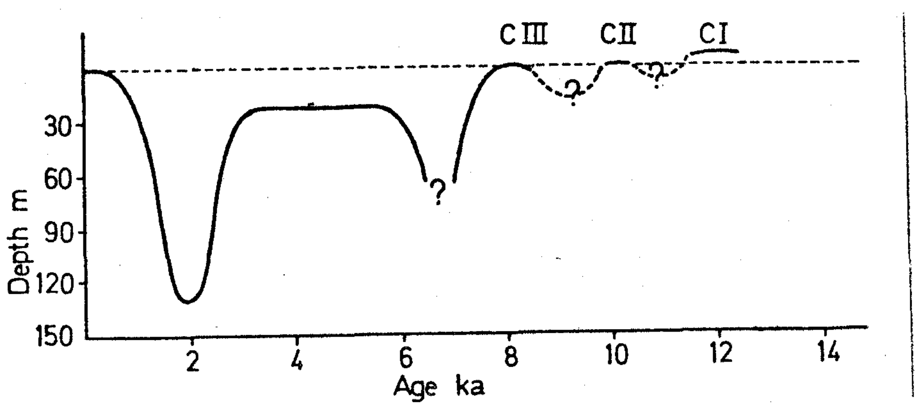


Figure 13.25 History of sea level oscillation over the last 120 ka shown in graph form. CI = 6,3m a.s.l; CII = 2 - 3,5m a.s.l.; CIII = 0 m a.s.l.

consists of unconsolidated beds of shell and comminuted shell, locally cemented, and with little quartz. But it ranges through limestones with shell layers to coquina, for example, Churchhaven.

(ii) Langebaan Limestone Member: type-section on Hoedjiespunt behind the Sea Harvest factory where it is over 20m thick. It consists of fine to medium-grained calcareous sand, loess, and limestone with fine shell detritus and complete shells of the land-snail Trigonephrus globulus. This land-snail is distinctive enough to feature as a lithological component. The member includes outwash horizons and Middle Stone Age middens.

IX. CONCLUSIONS AND SYNTHESIS

Tilting and warping since the Miocene have displaced marine units from their original elevations, so that the equivalent of the Namaqualand 45-50m transgression complex (Carrington & Kensley 1969) is situated at only 10m a.s.l. in the Saldanha area. Sediments equivalent to the Namaqualand 17-21m transgression complex are possibly submerged in the Saldanha area. But there is good reason to believe that warping has been minimal in the Late Pleistocene.

Evidence for stillstands of the sea in the last interglacial is not always unequivocal. There is no difficulty in recognising shorelines, but it is frequently difficult to discriminate between major stillstands and lesser features. In summary, data from the Elands Bay-Ysterfontein area identifies three stillstands of the sea in the last interglacial, with the shoreline at 6,3m a.s.l. being a major feature. These stillstands are at 6,3m a.s.l. (CI), 2-3,5m a.s.l. (CII), and sea level (CIII).

The platform on the seaward edge of the Vredenburg pluton was carved partly in the Neogene, remodelled in the Early Pleistocene but owes its present relief to Late Pleistocene abrasion. Relief of the platform is such that it could have been carved by a single stationary sea level. The outer ramparts on the North Head and South Head platforms were formed by a sea level in the region of 6,3m a.s.l. Shell beds on the platform agree with this interpretation. Similarly, shell beds on the Malmesbury platform west of the Berg River mouth are related to a major stillstand at 6,3m a.s.l. as are the extensive barrier-beach and breaker-bar complex of the St Helena Bay coastline. That there are few indications of lower last interglacial shorelines along the open coast suggests that this stillstand was a major one.

Shell deposits in the former Verlorevlei and Berg River estuaries have maximum elevations of 5m a.s.l. Infratidal molluscs still in their living positions suggest a mean sea level at least 6m a.s.l. But the fauna in general, which in estuaries would inhabit the mid-tide to low-tide zone (Emery & Stevenson 1957), shows that this sea level could not have been much above 6m and thus agrees with the CI transgression.

Evidence for a transgression to 2-3,5m (CII) consists of a shell bed which is banked against older CI sediments on Kruispad, the lagoon sediments between the emerged bars at Velddrif, Blouwaterbaai sediments, Geelbek tidal flat sediments at 2m a.s.l. and the Kraalbaai aeolianite - marine limestone contact at 2,5m a.s.l.

A boulder bed in the Bomgat on the Hoedjiespunt peninsula obviously owes much of its present elevation to waves surging up the passage through the granite in front of the cave. Considering that the elevation of this boulder bed is only 3,5m a.s.l. it could be attributed to very heavy surf during high tides. The 'fresh' appearance of the shells which still have their colour preserved would suggest a Recent age, and the elevation is certainly not excessive given the physical setting of the cave. But it must also be borne in mind that this beach could date to the last interglacial CII or the CIII transgressions. Samples have been submitted for radiocarbon dating even though ground-water contamination seems likely. This site illustrates some of the difficulties encountered in raised shoreline studies.

At four sites there is evidence for a sea level in the last interglacial coinciding with the present beach (CIII). This includes beachrock at Muizenberg (Siesser 1974), Blouwaterbaai and Jutbaai. Wave-generated boulders beneath Middle Stone Age sediments in Die Kelders cave are attributed to waves surging into the cave from about the present position of sea level, or even lower.

These results are summarised in a graph of Late Pleistocene eustatic sea level changes (Figure 13.25). That portion of the curve from about 47ka to the present is based on ^{14}C dates of submerged material, and has been fully discussed in the first chapter. The last interglacial part of the curve agrees well with the shorelines on Mallorca (Butzer & Cuerda 1962). It is suggested that the CI shoreline correlates with their Tyrrhenian (T) IIa, CII with T IIb, and CIII with T III. According to Butzer and Cuerda, it is the T II shoreline (5-10m) which contains thermophilic molluscs, and Tankard (1975) has suggested correlation of the CI thermophilic fauna with that sea level. Broecker et al (1968) have dated three Eem transgressions on Barbados: BIII (6m a.s.l) at 122 ka, BII at 103ka, and BI at 82ka, and Chappell (1974b) has dated three sea level peaks on New Guinea at 120 ka, 100 ka and 80 ka. The thermophilic mollusc content of the C I and C II shorelines suggests the distinct possibility that the three last interglacial peaks of the southwestern Cape may all correlate with the warm peak at 120 ka, and that the two lower shorelines would then represent only temporary halts in a regression from the 6,3m level.

It is not known whether the C II and C III shorelines represent temporary stillstands in regression from the major C I stillstand, or whether they are, in fact, transgressive and separated by regressions.

Withdrawal of the sea with the build-up of high latitude glaciers left vast, unvegetated tracts of sand exposed on the emerging shelves, and these were blown into dune fields by wind systems which may have increased in intensity with the onset of glaciation. But in the southwestern Cape they do not signify an arid climate. On the contrary, there is evidence that precipitation was at times higher than today.

There is no evidence to suggest Holocene sea levels higher than the present.

CHAPTER 14

LATE PLEISTOCENE MARINE PALAEOECOLOGY AND PALAEOCLIMATOLOGY

I. INTRODUCTION

Isolated fossiliferous, marine sediments occur on the wave-cut platforms and in the sheltered Late Pleistocene embayments from Elands Bay on the west coast to Knysna on the south coast. The fossils occur in unconsolidated quartzose shelly quartzose sands. The invertebrate fauna is essentially modern in composition and comprises some 150 species which today live in shallow-water environments. Only three of the mollusc species from the west coast deposits are not known to be living today. In so far as most of these taxa are still living, and are inhabitants of shallow water, they provide ideal material for a palaeoecological study.

The Late Pleistocene fauna of the west coast is broadly divisible into two distinct, but contemporaneous, ecologic zones. These are, firstly, a cool-water, open-coast facies characterised by rocky shore and sandy beach assemblages, and secondly, a warm-water, sheltered embayment facies (estuaries and lagoons). Whereas the open-coast facies is laterally continuous, the sheltered embayment facies is restricted in distribution. The open-coast facies is characterised by molluscs which commonly inhabit the present coast. A striking feature of the estuarine-lagoonal facies is the association of extant and extra-limital thermophilic species. ("Thermophilic" and "extralimital" imply species which occur outside their normal spawning range). Examination of their present latitudinal ranges indicates that a significantly warmer hydroclimate prevailed when those species were common along the south-western Cape coast. In attempting to reconstruct the palaeoenvironment, a detailed examination of a south coast assemblage is necessary. The fossiliferous deposits at Knysna are therefore included in this study.

Late Pleistocene faunas of open-coast and sheltered-embayment aspects from southern California are similarly distinctive. Nearly all the dominant species also inhabit the adjacent coast. But the sheltered embayments also contain a high proportion of thermophilic molluscs which are found far north of their present-day geographic range end-points. Several authors have used these anomalies as a key to Late Pleistocene climatic interpretation(Valentine 1955, 1957, 1961; Valentine & Meade 1961; Addicott & Emerson 1959; Emerson & Chase 1959; Kern 1971). Valentine (1955) explained the diverse nature of the fauna by changes in intensity of the oceanic circulation and upwelling, while the water of the embayments was heated by increased solar radiation. A recent alternative explanation suggests that the larvae of tropical molluscs were transported into the cooler areas by periodic local and temporary current changes, and that they became only temporary, non-breeding, members of the community (Zinsmeister 1974).

The south-western Cape Province, with its extensive Late Pleistocene fossiliferous deposits, and its complex present-day ocean current systems, is well placed to make a contribution to the knowledge of Late Pleistocene hydroclimates. The purpose of this chapter is to describe the mollusc fauna, particularly between Ysterfontein and Elands Bay, and to examine their palaeoenvironmental significance.

II. METHODS

The absolute density of fossil specimens of each molluscan species at any particular site is difficult to determine because of the sampling problems inherent on the size of the shell. For instance 'Rissoa' capensis is a small gastropod (usually less than 3mm) and would frequently number in the hundreds from just 100 g of sediment, while the larger (approximately 200mm) Panopea glycymeris would occur at approximately 1m intervals. Bearing in mind that the present study is a palaeoecological one, and that the larger molluscs have been better studied with respect to taxonomy and ecology, emphasis has in all cases been placed on the macro-molluscs. A further source of error arises from the differential fragmentation of shells. Bivalve shells are more easily broken than gastropod shells, and in the living assemblage they may have been far more abundant than the faunal list suggests. Table 14.1 should be taken only as an approximation to the original community structure.

The procedures adopted in drafting table 14.1 are as follows:

(i) It was desirable to sample as small an area as possible to obtain not only the absolute density of each species, but also to obtain restricted samples for size-frequency analyses. In most cases a quadrat size of 1 square metre has been found adequate, although quadrat size may have to be adjusted up or down depending upon the relative abundance of specimens.

(ii) For a shell to be counted as an individual it must be nearly complete, or so nearly so that the remainder could not be identified and counted separately. Left and right valves of each bivalve species were counted separately, and the highest count taken as the total number of individuals of that species in that quadrat.

(iii) Each species has been recorded in Table 14.1 as percentage frequencies in the sample where more than 50 individuals were counted. Where a total of less than 50 individuals were counted they are recorded in Table 14.1 as "X".

No attempt has been made to relate species to sediment texture since every mollusc at some time must have been living among already dead and fragmented shells. The substrate would thus consist of quartzose sand and bioclastic material ranging in size from complete shells to finely comminuted fragments.

TABLE 14.1

Distribution and Abundance (%) of Late Pleistocene Mollusca

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	*Lagoon Floor*	*Knysna*
GASTROPODA																								
Amblychilepas scutella (Gmelin)												< 1						< 1						< 1
Fissurella aperta (Sowerby)																					< 1			
F. mutabilis Sowerby												< 1	×					1,3	1,0					
Helcion pectunculus (Gmelin)															< 1	< 1								
Patella argenvillei Krauss															7,8	7,2								
P. barbara Linnaeus	2,0														7,4	< 1	×							
P. cochlear Born															2,0	2,6	×							
P. compressa Linnaeus									1,0						3,9	1,3		< 1						
P. granatina Linnaeus			< 1						1,0	×					8,3	2,0								
P. granularis Linnaeus															4,9	10,5								
P. longicostata Lamarck																		< 1						
P. miniata Born						×			< 1						< 1									
Cantharidus suarezensis fultoni (Sowerby)																								9,8
Gibbula cicer (Menke)												< 1												
Minolia sp.																								< 1
Oxystele trigina (Chemnitz)			< 1																		< 1			
O. variegata (Anton)			1,4													< 1		15,7		9,4	2,7	2,4		
Monilea alexandria (Tomlin)																								1,8
Solariella sp.																						1,0		
Pseudostomatella orbiculata (A. Adams)																								< 1
Turbo cidaris Gmelin															< 1	< 1								
T. sarmaticus Linnaeus																								< 1
Littorina knysnaensis Philippi			< 1													2,0		4,6						< 1
Alvania alfredensis Bartsch																		< 1						
Coriandria cf. *halia* (Bartsch)																					< 1			
'*Rissoa*' *algoensis* Thiele								×																
'*Rissoa*' cf. *capensis* Sowerby			×	×				×			×	×						×			×	×		×
Turritella capensis Krauss								×	< 1		18,0								2,0			1,7		2,6
T. carinifera Lamarck																			18,4	7,1		6,6		
T. sanguinea Reeve												10,3	×											
Cerithium kochi Philippi																								3,7
C. scabridum rufonodulosum E. A. Smith																								16,3
Diala infrasulcata Sowerby																								< 1
Alaba pinnae Bartsch																								< 1
Calyptraea chinensis (Linnaeus)							×		< 1					×				< 1				1,0		
Crepidula aculeata (Gmelin)					1,3							< 1	×	×		3,3			< 1			1,0		
C. porcellana Lamarck						×		×	< 1			< 1		×	1,5	< 1	×			1,2	1,9			< 1
C. capensis praerugulosa Kilburn & Tankard			< 1	×	46,7				18,1			< 1			10,8									
Cypraea algoensis Gray																< 1								
Natica genuana Reeve									< 1		4,0	< 1							1,0	12,9				< 1

TABLE 14.1 (Continued)

Argobuccinum argus (Gmelin)	< 1		2,7				< 1						5,4	6,5	×	< 1						
Cymatium dolarium (Linnaeus)	< 1																					
Thais cingulata (Linnaeus)				×									1,5	3,3								
T. dubia (Krauss)																< 1			< 1			< 1
T. squamosa (Lamarck)			1,3	×			< 1	×		< 1		×	12,3	19,6	×	1,3						
Burnupena papyracea papyracea (Röding)					×		< 1							35,9	×	7,8		10,6	< 1			
B.p. cincta (Bruguière)	< 1									< 1			19,1				2,0		1,0	3,3		
B. digitalis Meuschen			8,0	×	×	×	1,0	×					< 1			1,3	< 1	1,2	5,5	1,0		
B. laevissima (Gmelin)			5,3	×	×		10,5									< 1	< 1					
Nassarius analogicus Sowerby		×					< 1															
N. kochianus (Dunker)										< 1												
N. kraussianus (Dunker)						×																1,6
N. plicatellus (A. Adams)																	4,1	2,3	5,1	4,1		
N. scopularcus Barnard	4,7	×					< 1	×	30,0	11,5	×											
N. speciosus A. Adams							< 1									< 1	1,0		< 1	1,0		
Fasciolaria lugubris Reeve							< 1							< 1		< 1	< 1	3,5	< 1	2,5		
Marginella capensis Krauss		×				×	< 1	×		5,8	×					2,0	3,0	3,5		1,0		
M. piperata Hinds													< 1	< 1								
Mitra aerumnosa Melvill														< 1								
Vexillum capense (Reeve)														< 1								
Cythara amplexa (Gould)										< 1						< 1						
Clionella sinuata (Born)													1,0	< 1		19,0	3,0	10,6	2,3	4,1		
Conus scitulus algoensis Sowerby													6,4	1,3	×							
Syrnola aganea (Bartsch)																						< 1
Turbonilla kraussi Clessin										< 1												
Turbonilla sp.																						< 1
Pupa daviesi Kilburn & Tankard																						< 1
Ringicula turtoni Bartsch																						< 1
Philine aperta (Linnaeus)																						< 1
Cylichna tubulosa Gould										< 1												
Bullaria ampulla (Linnaeus)																						1,0
Atys cylindrica Sowerby																						< 1
Siphonaria aspera Krauss													1,5	1,3		5,2						
S. capensis Quoy & Gaimard	< 1																1,0		1,6	1,7		
BIVALVIA																						
Nuculana bicuspidata (Gould)										3,2												
Aulacomya ater (Molina)	< 1		1,3	×	×								2,9	< 1						1,0		
Perna perna (Linnaeus)		×	1,3	×			1,0	×								1,3						
Choromytilus meridionalis (Krauss)				×											×				< 1			
Chlamys tinctus (Reeve)																						< 1
Pecten sulcicostata Sowerby																						< 1
Lima fragilis (Chemnitz)																						< 1
L. rotunda Sowerby																						< 1
Ostrea atherstonei Newton														< 1			5,1	7,1	< 1	5,0	> 99	
O. stentina Ranson	2,7																					
O. algoensis Sowerby										9,6	×											< 1

TABLE 14.1 (Continued)

BIVALVIA (*contd.*)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	*Lagoon Floor*	*Knysna*
Loripes liratula (Sowerby)	9,0	20,0	26,4								8,0	15,4	×	×					20,4	14,1	< 1	4,1		57,5
Anodontia edentula (Linnaeus)																								1,4
Felania diaphana (Gmelin)																								< 1
F. subradiata Sowerby																					1,0			
Erycina subradiata (Gould)								×																
Lasaea rubra Montagu			< 1																		1,0	1,0		
Mysella convexa (Gould)			< 1															< 1						
Tellimya trigona Barnard												< 1						26,1						
Tellimya sp.								×																
Cardita sp.												< 1												
Parvicardium turtoni (Sowerby)																								< 1
Mactra glabrata Linnaeus														×					4,1			11,4		
M. ovalina Lamarck																			1,0	1,2	2,3	1,7		1,2
Scissodesma spengleri (Linnaeus)						×	×																	
Lutraria lutraria (Linnaeus)					2,7	×	×		11,4	×		1,3												
Lutraria sp.																								< 1
Solen capensis Fischer			1,4	×					< 1		2,0	1,9		×				1,3	< 1	1,2	< 1			< 1
Phaxas pellucidus (Pennant)									< 1															
Tellina ponsonbyi Sowerby																					< 1			
T. madagascariensis Gmelin	6,0	10,0	8,1						< 1										10,2	9,4	23,4	3,3		< 1
T. trilatera Gmelin					12,0														< 1					
T. gilchristi Sowerby												< 1						< 1			< 1			
Macoma crawfordi Sowerby				×														< 1						
M. litoralis (Krauss)																								< 1
M. tricostata (Römer)	2,0	3,0	< 1																					
Gastrana matadoa (Gmelin)			4,1											×					1,0			9,1	< 1	< 1
G. fibrosa Kilburn & Tankard																		1,3						
Leporimetis hanleyi (Dunker)											2,0	1,0								< 1				< 1
Donax serra (Chemnitz)					2,7	×			1,0										< 1	< 1	3,1			
D. sanctuarium Kilburn & Tankard																					1,0			
Psammotellina capensis Sowerby				×				×			34,0										1,0			
Theora alfredensis Bartsch												< 1										1,0		< 1
Venus verrucosa Linnaeus			< 1		2,7				< 1	×									16,3	3,5		18,4		< 1
Tivela tomlini Haughton																					1,7			
Dosinia lupinus Linnaeus	81,0	67,0	50,0	×							2,0	18,6	×						1,0		39,1	9,9		< 1
D. hepatica (Lamarck)																								< 1
Venerupis senegalensis Gmelin			< 1	×	12,0	×			55,0	×		12,8	×	×	1,0			3,3			3,5	1,7		1,0
V. dura Gmelin			< 1									6,4	×						4,1	1,2				< 1
Panopea glycymeris (Born)												1,3												
Total number individuals	100	100	148		75				105		50	156			204	153		153	98	85	256	121		492
% thermophilic species	60,0	75,0	26,1								22,2	25,8	20,0	33,0				6,9	26,9	31,6	22,6	25,9		12,8
% thermophilic individuals	17,0	33,0	41,9								20,0	42,9						—	45,9	32,9	28,1	35,5		60,4

III. PALAEOECOLOGY

In this chapter past extensions of tropical and subtropical mollusc geographic ranges are used as a basis for interpreting Late Pleistocene palaeotemperature changes. These inferences are based only on fossils of still extant species, and the assumption (Durham 1950) that stenothermal organisms are in general more critically limited by minimum temperatures than by maximum temperatures. The validity of such palaeotemperature inferences depends on the fossils being preserved in the sediments in which they once lived.

In describing the relationship between the fossils, after death, and the sedimentary environment, we define the following types of fossil assemblages:

(i) Life assemblage: disturbance after death negligible (Hallam 1960).

(ii) Death assemblage:

(a) Indigenous : organic remains disturbed after death but not transported very far (Hallam 1960).

(b) Transported : organic remains introduced from a neighbouring contemporaneous or older environment.

(iii) Mixed assemblage: this comprises any combinations of the above possibilities and is the general case (Hallam 1960). The status of this assemblage is clarified by describing it, for instance, as a mixed life and indigenous death assemblage. In the present chapter this is the commonest case.

A. Population Dynamics

Boucot (1953), Olson (1957), Craig and Hallam (1963), and Craig and

Oertel (1966) have attempted to discriminate between life and indigenous death assemblages on the one hand and transported death assemblages on the other, by using size-frequency distributions. According to Boucot an indigenous death assemblage is characterised by a positively skewed distribution (large number of small forms), while negative skewness or a normal distribution characterises the transported death assemblage. He suggested that negative skewness resulted from winnowing of the smaller shells. Craig and Oertel (1966) question this and maintain that the shape of the size-frequency distribution in a fossil population depends principally upon the growth-rate and mortality-rate of the relevant species. In this way negative skewness could arise from a decreasing growth-rate with constant mortality-rate which would concentrate the older age-classes in a few size-classes. Craig and Oertel suggest the following options:

Growth-rate	Mortality-rate	Size-frequency distribution
decreasing	constant	negative skewness
constant	decreasing	positive skewness
constant	increasing	flattening of curve, possibly negative skewness
constant	constant	mirror image of living population by dead population.

Growth-rate and mortality-rate complement each other when one decreases and the other increases, but cancel each other if both increase or decrease.

Higher mortality-rates which favour large populations may result from a fluctuating environment (Valentine 1971). Mortality is affected by nutrients, temperature, and salinity changes. In general, invertebrates have higher mortality-rates in the early stages of life, but the rate may be lower in some species than others (Craig & Oertel 1966). Environmental conditions in estuaries and lagoons would be expected to fluctuate widely and rapidly. They would be expected to show a variable salinity range due to evaporation and influx of fresh water, and a high diurnal temperature range. Furthermore, seasonal upwelling of cold water along the Cape west coast leads to marked instability of the environment on the open coast too.

Although growth rate is an important factor in this type of study, it is one of the attributes about which there is little information. There is evidence, however, that most bivalves maintain a slightly decreasing, but nearly linear, growth-rate throughout life (Craig & Oertel 1966). This has been shown to be the case for Cardium edule, Tapes japonica, Dosinia exolata, and Venus striatula (Kristensen 1959; Wilbur & Yonge 1964).

The most likely effect of the unstable environmental characteristics of the west and south coasts of the Cape would be a high mortality-rate among the juvenile molluscs. Assuming a constant growth-rate for the bivalves, the interplay of mortality-rate and growth-rate should lead to positively skewed size-frequency distribution for a life assemblage, or an indigenous death assemblage.

Besides the effect of growth-rate and mortality-rate on the shape of the histogram, the assemblage may be affected by post-death mechanical change such as sorting or winnowing by currents (Boucot 1953) and selective fragmentation, and solution of the smaller or thinner shells. Experience with the west coast fossils shows that crushing and fracturing is of primary importance and affects the bivalves more than the gastropods.

The large number of fragmented shells in both the open-coast facies and estuarine-lagoonal facies sediments suggests death assemblages that have undergone considerable modification by wave-action. Whereas the open-coast bivalves are usually disarticulated, those in the estuarine-lagoonal sediments show a high degree of articulation. These sheltered embayment sites contain epifaunal and infaunal molluscs, some of which are preserved in their living positions. Size-frequency distributions for some of the bivalves are shown in Figure 14.1. Since the warm-water element inhabited the Late Pleistocene lagoons and estuaries, samples from those environments have been analysed in most detail.

A single sample from a known high-energy open-coast site was examined in detail. Figure 14.1A shows the size-frequency distribution for Venerupis senegalensis from the Velddrif site. The field-setting suggests a transported death assemblage in which vigorous wave-action piled shell debris up to form a breaker-bar. In these deposits the thinner shells have generally been fragmented. The V. senegalensis population is composed of thick-shelled forms. Its estuarine ecomorph, on the other hand, has a thinner shell and is of more constant morphology. The Velddrif V. senegalensis shows marked negative skewness (−0,82). It has an articulation ratio of less than 0,05. Articulation ratio is defined as the ratio of complete shells/$\frac{1}{2}$ (RV + LV).

Size-frequency distributions of bivalves from the estuarine-lagoonal facies differ from the pattern of the Velddrif example. Tellina madagascariensis from Verlorevlei (Figure 14.1C) and Churchhaven (Figure 14.1H) tends to have a flattened histogram. The Verlorevlei Tellina has a size-peak at 64–66 mm, and the Churchhaven species at 56–60 mm. While the narrower size range of the Verlorevlei material (50–86 mm) suggests a transported death assemblage, the greater range of the Churchhaven material (16–74 mm; more juveniles) suggests an indigenous death assemblage. That this argument can be misleading is shown by their articulation ratios of 1,46 and 0,47 respectively. At both of these localities T. madagascariensis is associated with comminuted shells and disarticulated valves of other species. T. madagascariensis at Verlorevlei was observed in a nearly horizontal attitude as were the other bivalves. But whereas most bivalves burrow with the shell vertical, Tellina burrows rapidly and settles in a horizontal position. For this reason the posterior end of the shell is strongly flexed to the right so as to broaden the radius of curvature of the siphons to minimise the current flow constriction.

In a small quadrat at Verlorevlei 29 articulated shells were observed. Of these 26 had the right valve uppermost, i.e. flexure upwards, and

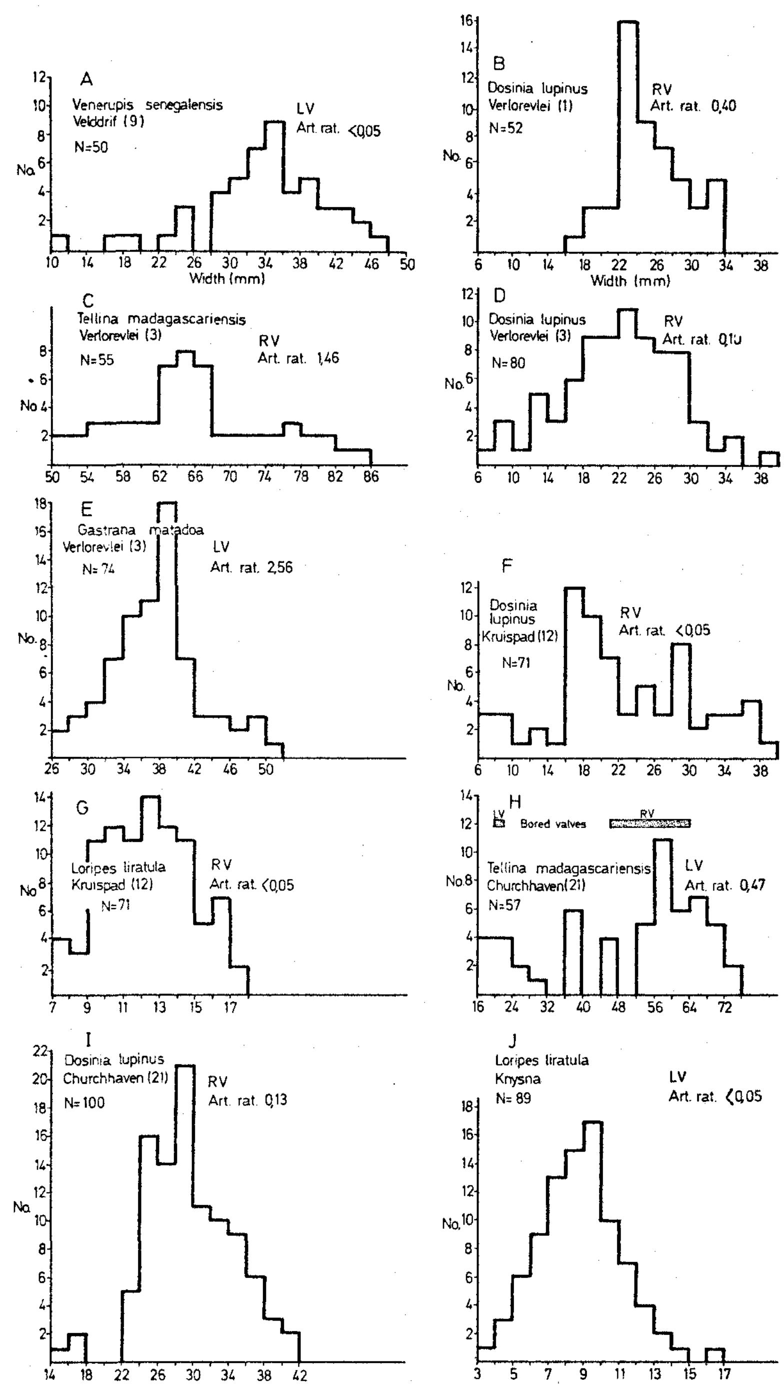


Figure 14.1 Size-frequency distribution of selected bivalve shells.

with the posterior end slightly raised. Many of the T. madagascariensis shells still had the ligamental material attached across both valves as a powdery residue. Clearly these animals must have died in their life's positions where the weight of sediment prevented opening of the shells after relaxation of the adductor muscles. If they had opened after death and subsequently closed again due to increased sediment load, the ligament would have broken.

At Verlorevlei the T. madagascariensis (site 3) size-frequency distribution can be compared with those of Gastrana matadoa and Dosinia lupinus (Figures 14.1 B, D, E). Dosinia lupinus at site 1 (Figure 14.1 B) has a positively skewed (+ 0,37) histogram with an articulation ratio of 0,40. The field occurrence suggests that reworking of the sediment has taken place, although the shells are still situated close to their original life habitat. Although the articulation ratio is high, relative displacement of each valve is common. At site 3 the D. lupinus histogram is bell-shaped (Figure 14.1 D). Here the articulation ratio is only 1,10. The bell-shaped distribution suggests less winnowing of the small specimens. Both D. lupinus populations peak at 22-24mm. From site 1 to site 3 at Verlorevlei there is a tendency for a greater spread of size-ranges of D. lupinus: 16-34mm at site 1 and 6-40mm at site 3. Conceivably the histogram for D. lupinus (site 1) was originally bell-shaped but winnowing may have removed the smaller sizes. This is borne out to an extent by the increase in relative proportions of Loripes liratula in the assemblage, from 9 per cent at site 1 to 26,4 per cent at site 3. The size-range of L. liratula 3-17mm (Figure 14.1 G, J) coincides with the juvenile fraction of D. lupinus. Because of the close similarity in shape of D. lupinus and L. liratula, equal-size specimens of each species would be expected to be hydraulically equivalent.

The size-frequency distribution of Dosinia lupinus at Churchhaven has a tail towards the larger specimens (Figure 14.1 I) suggesting winnowing. It has a narrow size-range, 14-42mm and peaks at 28-30mm. The low articulation ratio (0,13) suggests reworking.

At Kruispad, the Dosinia lupinus population again has a wide size-range, 6–40mm (Figure 14.1 F), but an articulation ratio less than 0,05. Furthermore, the histogram has two peaks, at 16–18mm and 28–40mm. The bimodal distribution could be explained by extinction of one living population, followed by fresh recruitment when environmental conditions improved. The site is 15km up the Berg River and could possibly have been influenced by sudden influxes of fresh water. But Loripes liratula at this site appears to represent a single population: this histogram is bell-shaped and peaks at 12–13mm. Its low articulation ratio (< 0,05) may be meaningless since this species does not have prominent dentition, and seldom in this study were articulated valves encountered. For example, Loripes liratula from Knysna has an articulation ratio less than 0,05 although juveniles predominate. Craig (1967) found that of five species he examined, only Divaricella quadrisulcata tended toward a normal distribution. (Divaricella and Loripes are both Lucinidae).

Gastrana matadoa (Figure 14.1 E) at Verlorevlei is positively skewed and has a high articulation ratio (2,56). This high articulation ratio, compared with that of Dosinia lupinus from the same site, could be due to the more robust dentition of the former, but the greater proportion of juveniles of G. matadoa suggests it probably is closer to a life assemblage.

In general, the sign of skewness alone as used by Boucot (1953) is insufficient to distinguish between indigenous and transported death assemblages. The field-setting shows the Velddrif breaker-bar deposit to contain a transported death assemblage. It has strong negative skewness and a low articulation ratio. This tail towards the left has been produced by winnowing of the finer fractions, and effectively displacing the mode towards the coarser size-grades. The other size-frequency distributions are suggestive of indigenous death assemblages which have in most cases been slightly reworked. In Figure 14.1 B the lack of juveniles of Dosinia lupinus coincides with low numbers of the hydraulically equivalent Loripes liratula.

Other indications that the shells are found in the sediment in which they once lived include relatively high articulation ratios of the larger shells. Smaller shells, e.g. Loripes liratula, tend to have weaker hinge attachments, while juveniles of other species may have been subjected to predation. Tellina madagascariensis at Verlorevlei appears to be in a life orientation, and also has the ligamental material still attached. At Kruispad a bed contains Panopea glycymeris still in the life orientation. The shells were all articulated, all posterior end upwards, and all on the same horizontal plane, i.e. they had all burrowed a siphon-length below the sediment-water interface. At most sites, left and right valves were present in equal proportions.

In conclusion, the field-setting shows that open-coast assemblages are all transported death assemblages, but transportation has been only local. The faunas of the sheltered environments are mixed life and indigenous death assemblages. Taken as a whole the fauna of the open-coast facies and the estuarine-lagoonal facies are consistent with the inferred Late Pleistocene environments.

Predation

Many of the bivalve shells are punctured by countersunk borings, 1–2mm in diameter, made by predatory gastropods. The bored bivalves include: Dosinia lupinus, Tellina madagascariensis, Gastrana matadoa, Venerupis senegalensis, Venerupis dura and Ostrea algoensis. Crepidula capensis and Natica genuana were the most frequently bored gastropods.

At Churchhaven 17 per cent of Dosinia lupinus and 17 per cent of Tellina madagascariensis were bored. In the case of the former an equal number of right and left valves were bored, as would be expected in view of the fact that it burrows in a vertical position. Analysis of predation of T. madagascariensis reveals a very different pattern. Successfully bored valves fell into two size-classes (Figure 14.1 H). The first group comprised juveniles with a size range 20–22 mm and the second group adults with a size range 46–64 mm. Only right valves of adults were bored, which is to be expected because this species lives buried in sediment in a horizontal position with right valve uppermost. Craig (1967) found that the right valve of Tellina radiata was also preferentially bored. The second group of T. madagascariensis comprised juveniles in which left valves were preferentially bored, although a few right valves were also bored. The preferential boring of juvenile left valves could be the result of the burrowing characteristics. Tellina is a rapid burrower (Stanley 1970). Depth of burrowing would be controlled by length of the siphons, and for this reason the juveniles would presumably live at shallower depths than the adults and would possibly burrow more slowly. Subsequent current scour would possibly reach only the juveniles and flip them over to leave them left valve uppermost, hydrodynamically the most stable position. This also suggests that, besides predation, the juvenile bivalves would be more susceptible to environmental changes since they live closer to the surface.

Most of the bored bivalve shells have only one hole, which is not surprising since only one puncture is necessary to kill the animal. In a few cases two holes per shell were encountered. As far as Tellina madagascariensis is concerned the borings occur most frequently in the antero-dorsal half of the shell (Figure 14.2). It was also observed that all bored specimens of T. madagascariensis at Churchhaven consisted of separate valves, while four bored articulated shells of Dosinia lupinus were observed.

Of the gastropods, Crepidula capensis was the most extensively bored. Again there was usually only one hole per shell, although up to three were recorded. The reason why Crepidula capensis is so susceptible to predation is because it lives exposed at the surface where it is an easy prey. Unlike the bivalves, Crepidula does not appear to have any preferred area for boring (Figure 14.2). Crepidula capensis from the shell beds north of Laaiplek have smaller bored holes (less than 1 mm) than other C. capensis or the bivalves.

It is difficult to isolate the species that would have been the predator, although it seems probable that the borings were made by a gastropod. The most likely predator would probably be the buccinid

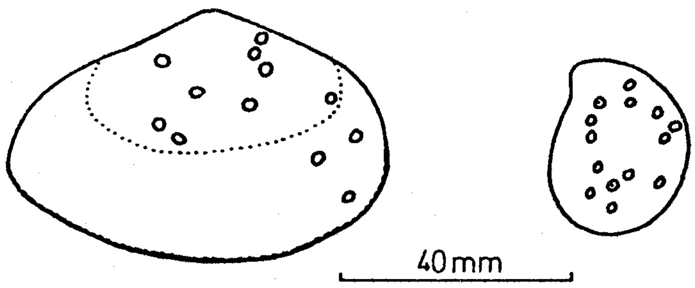


Figure 14.2 Predation : distribution of borings on Tellina madagascariensis (left), Crepidula capensis (right).

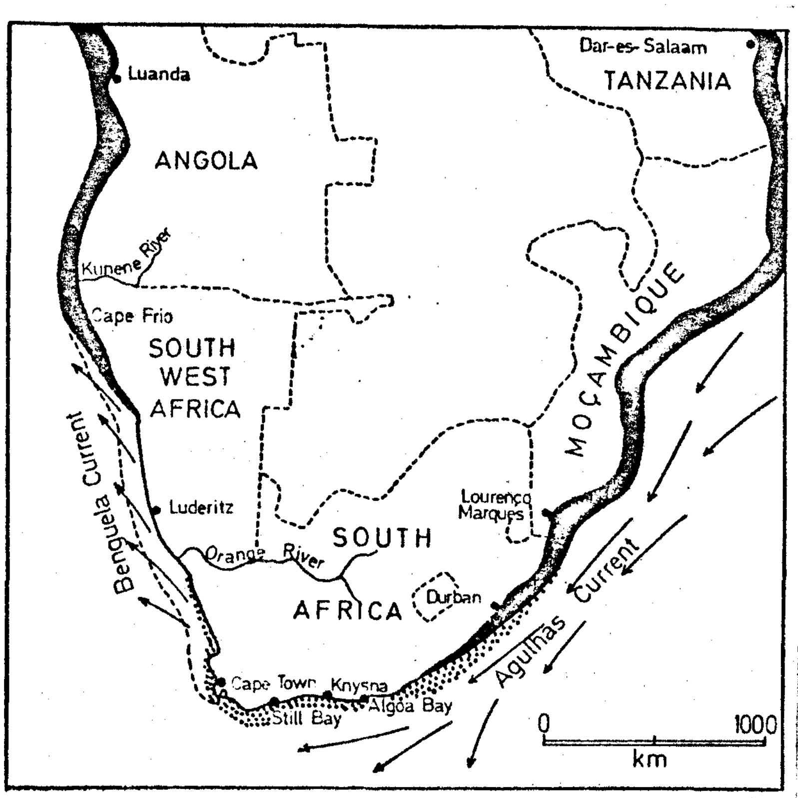


Figure 14.3 Ocean current systems and distribution of present-day invertebrate range zones. Black = tropical; stippled = warm-temperate; area bounded by broken line on west coast = cold water fauna.

Burnupena papyracea. Where many bivalves were found to be bored at Churchhaven, *B. papyracea* was very prominent. At Verlorevlei there was a total absence of *B. papyracea* and no sign of predation on bivalves.

Teratological specimens

Teratological specimens are those that are outside the normal range of variation of a species (Ager 1963). At most sites an occasional aberrant form was observed, and it was found generally that *Gastrana matadoa* was the most susceptible to damage during life. This was very apparent at Verlorevlei (site 3) where 27 per cent of *G. matadoa* were in some way deformed, compared with less than 2 per cent for *Tellina madagascariensis*. Perhaps this reflects the degree of adaptation of *G. matadoa* to its niche, and that the environment at Verlorevlei in particular did not favour this species.

Depth

Regional geomorphic analysis suggests that the depth of water reflected in the strata of all open-coast and sheltered-embayment sites examined could not have exceeded 5 m. The Late Pleistocene estuarine-lagoonal facies is dominated by intertidal deposits, and the open-coast facies by beach deposits. At Velddrif there is a good exposure of a breaker-bar with a wash-over fan, the top of which is probably a close approximation to high water spring tide. At Churchhaven and Kraalbaai *Callianassa* burrows and crab-burrows are still preserved, indicating an intertidal or shallow subtidal environment. The ostracode fauna, too, is indicative of intertidal conditions.

Very few of the 124 species listed in Table 14.1 suggest water depths greater than 25 m. Most of these species live today in intertidal and the uppermost sublittoral zones. Less than 4 per cent of the fauna lives today in water deeper than 25 m.

Of the open-coast assemblages Nassarius speciosus is today usually dredged, although it is sometimes found intertidally. No more than a few isolated shells were found. Erycina subradiata has been encountered at 26 m (Barnard 1964). Other species have a wide depth range although they are common in the intertidal zones. Nearly all species of Patella encountered at open-coast sites are intertidal. Only P. miniata, P. tabularis, P. compressa are infratidal (Branch 1971). Other intertidal molluscs include: Littorina knysnaensis (above HWN), Marginella capensis (low tide down), Bullia digitalis (follows the tide), B. laevissima, Thais cingulata, Burnupena papyracea (below mid-tide), Perna perna (mid-tide), Aulacomya ater (mid-tide to low tide on rocky shores), Choromytilus meridionalis, Dosinia lupinus, Donax serra (borrows in surf beaches below mid-tide), Venus verrucosa (found on the surface at mid-tide). Depth ranges are given by Day (1969).

Of the estuarine-lagoonal facies molluscs Ostrea algoensis prefers depths of 25-200 m, Tellina ponsonbyi prefers off-shore waters to depths of 95 m, and Theora alfredensis has been dredged from depths below 70 m (Barnard 1964). Ostrea algoensis and Theora alfredensis are common in this facies. Only one Tellina ponsonbyi valve was found at Churchhaven, suggesting that this specimen strayed into shallow water.

Most of the sheltered embayment molluscs prefer very shallow or intertidal water. These include the gastropods, Patella spp., Littorina knysnaensis (HWN), Turritella capensis (quiet water shallower than 3 m), Oxystele variegata, Burnupena papyracea, Nassarius kraussianus (mud-banks and weed-beds of estuaries); the bivalves Loripes liratula (muddy sand at low tide), Solen capensis (muddy sand of estuaries), Venus verrucosa, Choromytilus meridionalis, Perna perna, Aulacomya ater, Tellina madagascariensis (extensive infratidally), Dosinia lupinus, Donax serra and Panopea glycymeris.

None of the constituents of the open-coast or estuarine-lagoonal facies required water depths greater than about 25 m. Modern bathymetric ranges for these species suggest maximum Late Pleistocene water depths at these sites of the order of 0–5 m. This is consistent with the geomorphic evidence.

Temperature

Since the primary purpose of this chapter is a discussion of a southerly migration of tropical and subtropical mollusc species in the later interglacial, the most important criteria are those indicative of temperature. Sea surface temperature inferences may be made by comparing geographic ranges of living molluscan species with those of their fossil counterparts. Although the fossil fauna of the west coast is distinctly modern in character, the estuarine-lagoonal facies assemblages regularly contain several species whose present-day distribution is restricted to tropical and subtropical waters. The modern geographic range end-points of the thermophilic species that live further north on the west African coast are separated by 2 000 km from the fossil occurrences in the south-western Cape so that the fossil assemblages contain species that do not today live in association. Northward migration of the extralimital species was probably induced by deteriorating environmental conditions. Since the geographic ranges of recent molluscs seem to be determined mainly by temperature (Valentine 1955), it would appear that falling temperature was the compelling factor. It could also be argued that lower temperatures accompanied the advance of the Weichselian ice-sheets, in which cass falling sea level would have drained the Late Pleistocene estuaries and lagoons. But with reflooding of these environments in Recent time, the thermophilic molluscs did not return.

Mollusc assemblages in the estuarine-lagoonal facies of the west and south coasts contain a total of more than 20 species that do not sustain populations in those areas today. West coast sites contain 15 such species, mainly bivalves. These extralimital molluscs, their present-day geographic ranges, and minimum temperature tolerances, are shown in Table 14.2.

TABLE 14.2 Present-day Geographic Ranges, Temperature Minima, and Distribution of some Fossil Mollusca.
(The paper by Kilburn & Tankard is now published: 1975).

			Fossil Assemblages		
			W. Coast		S. Coast
	WARM WATER TAXA	*Minimum Temp. °C**	*Open*	*Estuaries, Lagoons*	*Knysna*
GASTROPODA					
(?)*Cantharidus suarezensis*	Durban—Tanzania (Kilburn & Tankard, in press)	19			×
Alvania alfredensis	Still Bay—Port Alfred (Kensley 1973)	14		×	
Cerithium kochi	Algoa Bay—Mozambique (Kensley 1973)	16			×
(?)*Cerithium scabridum*	Red Sea—Quirimba Is. (23,4°S; 40,7°E) (Kilburn & Tankard, in press)	24			×
Cypraea algoensis	Algoa Bay—Natal (Kensley 1973)	16	×		
Marginella piperata	Jeffreys Bay—Zululand (Kensley 1973)	14	×		
Atys cylindrica	Mozambique (Kensley 1973)	21			×
BIVALVIA					
Nuculana bicuspidata	Mauritania—Angola (Nicklès 1950)	17		×	
Ostrea atherstonei	Saldanha Bay—Bushmans River (Korringa 1956)	14		×	
O. stentina	Morocco—Congo Republic (Nicklès 1950)	17		×	
O. algoensis	False Bay—Delagoa Bay (Korringa 1956)	14		×	×
Loripes liratula	Mauritania—Angola (Kilburn & Tankard, in press)	17		×	×
Felania diaphana	Mauritania—Angola (Kilburn & Tankard, in press)	17			×
F. subradiata	Still Bay—Durban (Barnard 1964)	14		×	
Mactra ovalina	Durban—Delagoa Bay (Kilburn 1971*a*)	19		×	×
Scissodesma spengleri	False Bay (Barnard 1964)	14	×		
Tellina ponsonbyi	Still Bay—Zululand (Boss 1969)	14		×	
T. madagascariensis	Gabon—Baia dos Tigres (17°S) (Boss 1969)	17		×	×
Macoma tricostata	Angola (Kilburn & Tankard, in press)	17		×	
Gastrana matadoa	Mauritania—Ivory Coast (Nicklès 1950)	14		×	×
	Still Bay—Delagoa Bay (Barnard 1964, as *G. abildgaardiana*)				
Leporimetis hanleyi	Luanda (Kilburn & Tankard, in press)	17		×	×
Theora alfredensis	Algoa Bay—Zululand (Barnard 1964)	16		×	×
Venerupis dura	Morocco—Angola (Nicklès 1950)	17		×	×
Panopea glycymeris	Mediterranean—Senegal (Kensley 1974)	20		×	
	COLD WATER TAXA	*Temp. range*			
GASTROPODA					
Thais cingulata	Port Nolloth—False Bay (Kensley 1973)	14–17	×	×	
Burnupena papyracea papyracea	Paternoster—Hermanus (Kensley 1973)	14–17	×	×	
Nassarius plicatellus	Walvis Bay—Table Bay (Kensley 1973)	14–17		×	
N. scopularcus	Saldanha (Kensley 1973)	14–17	×	×	
Conus scitulus algoensis	Table Bay—Kommetjie (Kilburn 1971*b*)	13–17	×		
BIVALVIA					
Tellimya trigona	Lüderitz—Langebaan Lagoon (Barnard 1964)	12–16		×	
Lutraria lutraria	Lüderitz—False Bay (Barnard 1964)	12–15	×	×	

* References: Naval Oceanographic Office, Washington: Spec. Publ. SP-99
Sverdrup *et al.* 1942
Department Transport, Maritime Weather Office, Youngsfield, Cape Town: Sea Surface Temp. Charts

It is often difficult to assess the minimum termperature tolerance of living molluscs. Zinsmeister (1974) divides the normal biogeographic range of a taxon into a spawning range, and a non-spawning range. The modern geographic range of Ostrea atherstonei, from Saldanha to Bushmans River, implies a minimum temperature tolerance of about 13°C. But it appears that present temperatures in Saldanha Bay are too low for spawning (Korringa 1956). Spawning experiments in this area have confirmed this, and suggest that the oyster larvae may actually originate in an area of higher water temperature. It is possible that several species of molluscs extend their geographic range end-points by inhabiting local pockets of warm water such as estuaries. If this is the case then the temperature minima shown in Table 14.2 may actually be too low. It is not always clear from the literature whether the range end-points are open-coast, estuaries, or embayments. Other difficulties arise from taxonomic problems. For instance, Macoma ordinaria and M.crawfordi may be geographic variants of the Mediterranean M. cumana.

1. WEST COAST ESTUARINE-LAGOONAL FACIES

Because of physiographic changes at Verlorevlei and the lower reaches of the Berg River, present temperature ranges would be meaningless standards against which to measure Late Pleistocene changes. The present shoreline of Langebaan Lagoon has changed little since the last interglacial, and present temperatures at Churchhaven could be used as an approximation for the other sites as well. Day (1959) gives a surface temperature range for Churchhaven of 13,5°C to 37°C (HW) and 38,5°C (LW).

The fossil assemblage at Verlorevlei (site 3) is represented by 23 species of molluscs, 6 of which are indicative of warm water. These six thermophilic species constitute 42 per cent of the total individuals. The modern geographic range of Loripes liratula (26,4 per cent of the fauna) is Mauritania to Angola (Nicklés 1950), where its minimum temperature requirement would be 17-18°C. Ostrea stentina (2,7 per cent), Tellina madagascariensis (8,1 per cent), Macoma tricostata (< 1 per cent) and Venerupis dura (< 1 per cent) are also tropical west African species and could not tolerate temperatures below about 17°C. Gastrana matadoa (4,1 per cent) extends into water as cool as 14°C. Accepting that stenothermal organisms are more critically limited by minimum temperatures than by maximum temperatures (Dunham 1950), it would appear that the minimum temperature in this Late Pleistocene estuary must have been 17-18°C or at least 4°C warmer than present Churchhaven surface temperatures.

At Kruispad (site 12) 8 of the 31 species prefer warm water, and constitute 47,9 per cent of the total individuals. Of these Nuculana biscupidata (3,2) per cent, Loripes liratula (15,4 per cent) and Leporimetis hanleyi (1,0 per cent) have a minimum temperature requirement of 17-18°C. A significant constituent of this assemblage is Panopea glycymeris. Although it forms less than 1 per cent of the total individuals, it is nevertheless very common (because it is a large animal, about 200 mm, and is always found in life orientation, a shell count in a 1 m quadrat would be unlikely to yield more than one or two specimens). Panopea glycymeris lives in areas with a temperature range 20-27°C (Kensley 1974), although it could probably tolerate cooler water. This assemblage is thus indicative of a temperature minimum of about 18°C. At Bloemendal (site 11) Loripes liratula and Leporimetis hanleyi are common.

Three sites examined along the shore of Langebaan Lagoon yielded a significant proportion of thermophilic molluscs: Geelbek (site 19: 7 species in 26 constituting 46,0 per cent of the fauna), Skrywershoek (site 20: 6 species in 19 constituting 32,9 per cent of the fauna), and Churchhaven (site 21/22: 7 species in 31 constituting 28,1 per cent of the fauna in the lower unit; 7 species in 27 constituting 35,5 per cent of the fauna in the upper unit). The dominant thermophilic content of these sites is as follows:

			Churchhaven	
	Geelbek	Skrywershoek	Lower	Upper
	%	%	%	%
Ostrea atherstonei	5,1	7,1	1,0	5,0
Loripes liratula	20,4	14,1	1,0	4,1
Mactra ovalina	1,0	1,2	2,3	1,7
Tellina madagascariensis	10,2	9,4	23,4	3,3
Leporimetis hanleyi	-	<1,0	-	-
Venerupis dura	4,1	1,2	-	-

All of these species, except Ostrea atherstonei and Mactra ovalina, have a minimum temperature requirement of 17°C. Ostrea atherstonei is found living today in Saldanha Bay, but temperatures are apparently too low for spawning. It is possible that this species has a minimum temperature tolerance of 14°C. The floor of the present lagoon is underlain by about 3 million metric tons of O. atherstonei shells, with little detrital sediment, suggesting optimum temperatures for breeding, and hence probably greatly in excess of 14°C. Mactra ovalina has a minimum requirement of 19°C. A minimum temperature in the Late Pleistocene lagoon of about 18°C is indicated, about 4-5°C warmer than the present surface temperature minimum at Churchhaven.

2. SOUTH COAST ESTUARINE-LAGOONAL FACIES

A Late Pleistocene site at Knysna was the only one on the south coast examined in detail. But examination of material in collections of the South African Museum shows that very similar fossil faunas exist at Sedgefield, and Groot Brak and Klein Brak estuaries. The fossil assemblage at the Klein Brak estuary contains Panopea glycymeris, but their shells are, on average, much smaller than the Kruispad specimens (Kensley 1974).

The Knysna assemblage contains 47 molluscan species, of which 6 are extralimital and which constitute 60,4 per cent of the individuals. But these 6 species include the extinct subspecies Cantharidus suarezensis fultoni (9,8 per cent) and Cerithium scabridum *rufonodulosum* (16,3 per cent). Although the geographic range of the species is shown in Table 14.2, it is obviously not certain that the extinct subspecies had the same temperature tolerances. Their Recent relatives *Cantharidus suarezensis suarezensis* and *Cerithium scabridum* have minimum temperature requirements of 19°C and 24°C respectively. *Atys cylindrica* (< 1 per cent) has a minimum temperature requirement of 21°C. Of the bivalves *Felania diaphana* (3 per cent), *Tellina madagascariensis* (< 1 per cent), *Leporimetis hanleyi* (<1 per cent) and *Venerupis dura* (< 1 per cent), all suggest temperature minima of 17°C. *Loripes liratula* (57,5 per cent) also suggests a temperature minimum of 17°C, but the fact that it occurs in such great numbers, and the fact that all growth stages are present (Figure 14.1 J) suggest optimum conditions. Together the evidence suggests a temperature minimum in excess of 17°C for the Late Pleistocene estuary. Day *et al* (1952) give a temperature range at the railway bridge of 12-24°C.

3. WEST COAST OPEN-COAST FACIES

The present-day temperature range on the adjacent open coast is about 13-15°C (Shannon 1966). The fossil assemblage contains only three species which prefer warm water, but which never constitute more than 1 per cent of any assemblage: *Cypraea algoensis*, *Marginella piperata* and *Scissodesma spengleri*. The marked paucity of warm-water species indicates that Late Pleistocene nearshore water temperatures were not very different from the present.

Summary

Before discussing the palaeoclimatic significance of the extralimital molluscan species, it may be as well briefly to summarise the evidence.

1. All fossiliferous Late Pleistocene estuarine-lagoonal facies deposits contain extralimital species which usually

constitute more than 30 per cent of the individuals of each assemblage, and which indicate minimum water temperatures 4–6°C warmer than the present-day estuaries and lagoons, with minimum temperatures in excess of 18°C. Kanakoff & Emerson (1959) suggest temperatures in excess of 19°C for similar Californian occurrences.

2. All growth stages of the thermophilic molluscs are present (Figure 14.1). The importance of this is that it implies that temperatures were such that the normal spawning range of each mollusc is represented. Normal spawning range is characterised by a population which continually maintains its numbers (Zinsmeister 1974). The large number of individuals and presence of all growth stages show that the thermophilic taxa formed self-sustaining populations that were adequately adapted to the depositional environment.

3. The estuarine-lagoonal facies contains mixed life and indigenous death assemblages of molluscs. These molluscs obviously lived in environments reflected in the sediments and have suffered little post-death transportation.

4. The sediments and their mollusc fossils indicate water depths in general less than 5 m.

5. The fact that sediments with high proportions of extralimital molluscan species characterise most Late Pleistocene estuary and lagoonal situations, that these molluscs formed self-sustaining populations and inhabited very shallow water, suggests that the marine transgression to 6 m in the last interglacial was a major event during a climatic optimum.

Zinsmeister (1974) has suggested that similar well documented occurrences of thermophilic molluscs in Californian Late Pleistocene embayments represent only temporary, non-breeding members of the community. He believes that periodic local and temporary current changes introduced tropical mollusc larvae into areas of cooler water, and that these molluscs represent only temporary members of the community since temperatures would have been too low for them to maintain self-sustaining populations.

Zinsmeister's arguments stem from the fact that previous studies paid scant attention to the population dynamics, and consequently they fail to prove the presence of self-sustaining populations. Similar occurrences in the south - western Cape support the view that these faunas are the result of a warmer climate which must have affected the whole world. Age-wise these deposits are also similar and probably coincide with a high climatic peak (substage 5e) and sea level up to 7m higher than present at 120 ka (Shackleton 1969).

6. Configuration of the coastline was probably very important. Today a low Palaeozoic rock bar just above high-tide level keeps the sea out of Verlorevlei. A slight rise of sea level would create radical changes at Verlorevlei. Likewise the lower reaches of the Berg River formed a prominent estuary due to flooding by last interglacial high sea levels. Normally one would expect to find an overlap of geographic ranges of tropical and temperate species. In an overlap area the tropical species would be restricted to inshore, protected environments, and temperate species would live in the cooler open coast sites (Emerson 1956).

IV. PALAEOCLIMATIC INTERPRETATION

A. Review of Present Climate and Hydrology

The south-western Cape has a Mediterranean-type climate. Hot, dry summers are the result of the dominant anticyclones in those months, while depressions associated with westerly winds bring rainfall in the winter.

Hart and Currie (1960) have summarised the effect of the wind system on the climate of the south-western Cape Province. A subtropical high-pressure system is centred between 26° and 30°S. To the south this high-pressure system borders on the "westerlies" causing a steep pressure gradient. The south-easterly winds of the south-western Cape are the result of winds blowing anticyclonically around this high-pressure system. In summer the centre of the anticyclone lies at about 30°S and brings strong south-easterly winds to the south-western regions, but in winter it moves northwards to 26°S. The westerly wind system follows the anticyclone northward and the southern Cape is then frequented by depressions which bring rain from the south-west Atlantic, although the Namib Desert is still influenced by the Trade Wind belt.

The west coast of southern Africa, like the west coasts of other countries in similar latitudes, is characterised by a linear belt of centres of upwelling of cold subsurface water (Central Water) from Lüderitz to the Cape Peninsula (Figure 14.3). There is a north-south isotherm lineation with the coldest and least saline water near the coast (Shannon 1966). This upwelling, the Benguela Current, varies in intensity depending upon the wind system and local topography. The upwelling phenomenon arises from the displacement of surface water northwards and off-shore by the south-easterly wind system. Cooler subsurface water wells up to replace this warmer water. The result of this active upwelling is a complex system with tongues of cold water alternating with intrusions of warmer oceanic waters, and all divérging to the north-west (Bang 1971). Bang defines the Benguela Current as the area east of a belt of off-shore divergence within which the oceanic processes are dominated by short-term atmospheric interactions.

During the winter months when the anticyclone centre moves northward, weakening of the southerly wind component results in weakening of the upwelling system also. With less upwelling the surface waters over the inner part of the continental shelf are warmer by 2°C, but off-shore a drop in temperature tends to minimise this effect. With weakening of the Benguela Current system and the greater prominence of westerly and north-westerly winds, a southward-flowing inshore counter-current develops. This southward-flowing inshore counter-current is known as the Angola Current and it follows the decaying northern end of the Benguela Current. Shannon (1966) mentions a predominantly southward-flowing counter-current between Lambert's Bay and Cape Point which is present during all seasons, but most marked in winter. A southward-flowing counter-current carries "seeboontjies" from Angola and driftwood from the Orange River southwards (Wagner & Merensky 1928). Surface temperatures at the Orange mouth rise noticeably when the north-westerly winds blow.

The highest surface-water temperatures are encountered in the lagoons and estuaries which are protected from the effect of upwelling. The temperature range in the Langebaan Lagoon, 10-39^{o}C (Day 1959), contrasts markedly with that of the near-by open sea, 13-15^{o}C (Shannon 1966). During the summer months temperatures in Langebaan Lagoon are at a maximum, while those of the open sea drop.

The currents off the east and south coasts, the Mozambique Current and Agulhas Current, have been studied by Clowes (1950), Orren (1963, 1966) and Darbyshire (1964). Both of these currents are southward extensions of the great South Equatorial Current which flows westward across the Indian Ocean. At 26^{o}S the Mozambique Current is met by the southern branch of the South Equatorial Current, which is divided by Madagascar, and they combine to form the Agulhas Current. The Agulhas Current is finally deflected by the Agulhas Bank (Clowes 1950)(Figure 14.3).

Upwelling does take place off Cape Agulhas but it is a geostrophic upwelling, i.e. the upwelling varies with the velocity of the current. Here the Agulhas Current is weakest and Agulhas water retreats northwards in winter, thus allowing Central Water to reach the surface. In summer when the Agulhas Current is flowing at its strongest, it overrides the Central Water (Darbyshire 1964). Sometimes in summer the warm Agulhas Current moves away from the south coast and allows cold Central Water to replace it inshore, and within a day or two the temperature may fall by as much as 10^{o}C (Day 1963). Schell (1968) describes the occasional penetration of the Agulhas Current round the Cape into the South Atlantic.

B. Late Pleistocene Climate

It would be expected that the intensity, and probably position, of the anticyclones would vary with the solar radiation. A climatic optimum at 120 ka produced temperatures warmer than at any other time in the last 120 ka (Shackleton 1969). The anticyclone would have moved south of latitude 30°S (Van Zinderen Bakker 1967, figure 6). This would lead to the linear west coast upwelling belt moving south in response, and the winter-rainfall area would be more strictly restricted to the south-western Cape. Such a southward movement would be accompanied by more pronounced upwelling off the west coast south of the Olifants River than at present, and would operate over a longer period of time just as happens on the Namib Desert coast. Since the Central Water originated in the Southern Ocean, and since the Antarctic ice-sheet was essentially stable throughout the Pleistocene (Mercer 1968a), it is unlikely that the upwelling water would have been any colder than at present. This is confirmed by the fossil molluscs from the open-coast facies which suggest little temperature change. Addicott and Emerson (1959) and Valentine (1955) have also suggested intensified nearshore upwelling and a poleward expansion of isotherms to explain their mollusc faunas.

Eustatic rise of sea level would have changed the configuration of the coast by forming salt-water estuaries at Verlorevlei and the Berg River, as well as at numerous sites along the south coast. The present coast is not as embayed as the Late Pleistocene coast would have been. Thermally anomalous assemblages existed contemporaneously only in the vicinity of the protected bays where the increased solar radiation of the last interglacial heated the surface water considerably.

Periodic current changes permitted the introduction of tropical mollusc larvae into these pockets of warm water. The intensification of atmospheric circulation that led to more pronounced upwelling presumably also affected the inshore counter-current which possibly flowed more strongly than at present. Isaacs and Sette (1959) described anomalous wind fields in the Pacific area during 1957 and 1958 which so changed the oceanic circulation that tropical taxa were found far north of their expected ranges. This demonstrates that circulation changes as proposed in this chapter do happen at present, albeit less frequently and less pronounced than is here suggested.

On the east coast with intensification of the air circulation intensification of the present summer conditions would also be anticipated. Presumably the Agulhas Current flowed more strongly than today and continually overrode the Central Water, perhaps with frequent eddies of Agulhas water around the Cape Peninsula. This would limit the likelihood of sudden incursions of cold water, such as presently affect the Knysna estuary in summer and cause temperatures to drop between 10 and 15^{o}C (Korringa 1956).

These possible changes in the oceanic circulation during the hyperthermal period would have had far-reaching effects on the distribution of molluscs. Because of the low migratory ability of molluscs in the post-larval stage, their geographic distribution depends primarily on the dispersion of larvae. Distribution of the planktonic larvae depends upon ocean currents. About 85 per cent of tropical marine molluscs have a free-swimming pelagic larval stage which can exist, on average, three to four weeks before settling (Zinsmeister 1974). Dietrich (1935, quoted in Korringa 1956) gives the velocity of the Agulhas Current as 50 km per day. At this rate larvae could be transported 1 000 km before metamorphosis. Larvae can apparently withstand sudden changes of temperature; Korringa mentions a change from 25^{o}C to 2^{o}C which did not adversely affect oyster larvae.

A general southward movement of isotherms on the west and south coast is envisaged. This would have brought the tropical mollusc zone closer to the south-western Cape. Periodic expansions of the warm-water isotherms coupled with an intensified inshore counter-current on the west coast would have enabled tropical mollusc larvae to pass the cold Benguela barrier. If the larvae reached the sheltered estuaries and lagoons of the south-west coast, increased solar heating (substage 5e) would have allowed these molluscs to sustain their populations and persist there even after the open-coast thermal barrier became impassable again. Those thermophilic larvae that reached metamorphosis on the open coast may well have survived, but there is no evidence that they were able to sustain their populations. South coast estuarine facies show that larvae from the tropical west African coast were able to pass the Cape Peninsula; e.g. Tellina madagascariensis, Loripes liratula, Leporimetis hanleyi, Venerupis dura, Panopea glycymeris. The Agulhas Current was also able to transport Indo-Pacific mollusc larvae around the Peninsula to the west coast sites. In these sheltered embayments warm-water taxa survived as relicts.

V. DISCUSSION

The occurrence of thermophilic molluscs in Late Pleistocene sediments of the south-western Cape far beyond their present-day geographic range endpoints is not unique. Similar occurrences in California have been extensively studied, and the same conditions appear to have operated in the Miocene where the configuration of the coastline was very different from the present. The San Joaquin basin was a protected embayment in the Late Miocene, and there relict faunas persisted long after the temperate faunas had spread southward along the open coast (Addicott & Vedder 1963). The thermally anomalous Late Pleistocene molluscs also lived in shallow protected embayments (Addicott & Emerson 1959). The only significant point of difference between the two regions is that the open-coast facies of the Cape coast contains molluscs which nearly all live on the adjacent coast, while the Californian open-coast facies contain a fauna that reflects cooler water than the present. Valentine (1955) proposed that during the last interglacial of California upwelling was intensified, while at the same time the warmer water of the sheltered embayments was derived from increased solar radiation. He also suggested a general warming of the oceanic waters by increased solar heating.

The Late Pleistocene estuarine-lagoonal facies of the Cape Province are attributed to a eustatic rise of sea level which changed the configuration of the coast and produced a great number of estuaries where the effects of seasonal upwelling were excluded. A southward movement of the South Atlantic anticyclone at this time shortened the cold-water barrier on the western open coast by causing a southward shift of the Benguela Current, and possibly strengthened the inshore counter-current, so that thermophilic mollusc larvae were able to migrate southwards where they could sustain their populations in the solar-heated estuaries and lagoons. It would appear that the present temperatures are too low for breeding.

This history contrasts with that of the Early Pleistocene when sea temperatures on the open coast were much warmer than at present. The last of these warm-water open-coast episodes is associated with the 45-50 m transgression complex of the Namaqualand coast (Carrington & Kensley 1969). Striostrea margaritacea was common on the open coast, forming the so-called "oyster line" (Haughton 1931). This oyster requires a minimum water temperature of 25°C in summer (Korringa 1956). The 45-50m transgression complex sediments were probably pre-glacial Pleistocene. Such warm conditions must have been in response to the warmer conditions prevailing in the northern hemisphere, since the Antarctic ice-sheet is thought to have been stable throughout the Pleistocene (Mercer 1968a). Warming of the northern hemisphere would have moved the intertropical convergence farther south than even its Late Pleistocene position. This would have moved the South Atlantic anticyclone south, and with it the belt of upwelling, bringing tropical waters down the Namaqualand coast. But Striostrea margaritacea is totally absent from Early Pleistocene deposits in the Saldanha area, suggesting that cool water, and upwelling, were still dominant there, probably as far as the Olifants River.

If the model for a southward shift of the anticyclone system in the last interglacial is correct, then one would expect the opposite trend during the Weichselian. Lamb (1961) suggests that Ice Age circulation was marked by intensified circulation of the belt of westerlies, and greater mobility in the subtropical anticyclones which generate upwelling systems such as the Benguela Current. Van Zinderen Bakker (1967 and in press) presents evidence to show that during hypothermal periods the influence of the anticyclones and the

Benguela Current would have shifted northwards to the equatorial regions. Displacement of the westerlies towards lower latitudes would have expanded the winter-rainfall area of the southern Cape and would have caused a northerly shift of the Namib Desert. Tankard and Schweitzer (1974) have demonstrated, from Die Kelders cave sediments, that wetter conditions in the Weichselian coincided with a cooler period. Butzer (1973) has found a similar record in the Nelson Bay cave. A northward migration of the belt of westerlies would have resulted in a longer rainy season, if not year round precipitation (Tankard, in press d).

Finally, this chapter suggests a possible correlation of the last interglacial deposits of the Cape Province with the Eutyrrhenian of the Mediterranean. Although warping of the Eutyrrhenian shorelines is universal (Richards 1962), Bonifay and Mars (1959) have attempted to restore these shorelines to their original elevations. They attribute the Eutyrrhenian shoreline to a transgression to 2-3m a.s.l. Chronologically the Eutyrrhenian shoreline and last interglacial shorelines of the south-western Cape appear to be similar.

The *Strombus* fauna of the Eutyrrhenian deposits is characterised by *Strombus bubonius* and other species typical of Senegal and west Africa (Richards 1962) which suggest warmer hydroclimates than today. Like the warm water fauna of the south-western Cape, the *Strombus* fauna can also be attributed to poleward expansion of isotherms, and once in the Mediterranean the molluscs remained a relict fauna which survived in water heated by increased solar radiation.

CHAPTER 15

PLEISTOCENE HISTORY AND COASTAL MORPHOLOGY OF THE AREA BETWEEN THE CAPE PENINSULA AND DIE KELDERS

I. INTRODUCTION

As in the case of the west coast, previous studies of Pleistocene shoreline history of this part of the south coast are restricted to relatively few references: viz. Krige (1927), Mabbutt (1954), Gatehouse (1955), and Davies (1971, 1972). The most original of these studies is perhaps that of Krige. On this part of the coast his major emergence is situated at 18m a.s.l., and his minor emergence at 9m a.s.l. Gatehouse recognised shorelines at 90, 60, 29, 18 and 7,6m. Most of his sites have now been checked and will be commented on later. Davies recognised shorelines on the south coast at 60m (Cromerian), 50m, 30m, 18m, 9m, 6m, 3,5m and 1,5m. The last two he believes are Holocene.

The aims of this part of the project will be:

(i) to record heights of elevated Pleistocene shorelines;

(ii) to group these measured shorelines into equally spaced sets so as to be able to correlate them (if possible) and possibly determine the relative stability of the coastal margin;

(iii) to examine the influence of lithology on platform development.

II. COASTAL MORPHOLOGY

The Cape Peninsula geology is dominated by resistant ortho-quartzites overlying granite. Steeply plunging slopes on the Camps Bay-Hout Bay and Chapmans Peak coastlines result in nearly total reflection of wave energy so that little platform development is taking place. Elsewhere horizontally bedded sandstones lead to structurally controlled platforms. Furthermore, the differing degree of shoaling across these bedded sandstone platforms leads to variable altitudes of old storm beaches behind the modern beaches.

Rock structure has also had a pronounced effect on platform development east of Gordons Bay (Figure 15.1). For instance, at the Steenbras River mouth (5 km south-west of Gordons Bay) the sandstone dips away from the shore at approximately 40^{0}. Marine abrasion and undercutting have collapsed the strata so that platform development operates well below mean tide level. The outer edge of the platform here varies between 15 and 25 m, but averages 17-18 m a.s.l. Little beach material is preserved on the platform. Generally, strata dipping seaward, or sheared strata, are conducive to formation and preservation of lower terraces. Their preservation is the result of shoaling dissipating wave energy away from the shore. Figure 15.2 illustrates the effect of inland dipping strata on the outer edge of a platform.

The coastal platform is usually narrow and is backed by high mountain ranges with scree slopes graded to the platform level (Figure 15.3). Rivers have incised deeply into the platform during lower stands of sea level, and in Figure 15.3 such a drowned river valley is still clearly visible. The coastal platform appears to be a composite feature which originated in the Neogene. At Die Kelders Bredasdorp limestone of presumed Miocene age lies upon the sandstone platform at only 7-8 m a.s.l. (Figure 4.9). But it seems that this platform has been consistently remodelled in the Pleistocene. The platform appears to be stepped since its edges are usually encountered at predictable elevations, e.g. 7m,12-14m, 17-18m, 30m.

Other rock types result in sandy shores and estuary development. The Cape Flats are underlain by dune and beach sands lying upon a low level platform cut across weak Malmesbury rock. Tillites west of Kleinmond disintegrated easily under marine erosion to form wide sandy beaches. While the Bot River and its estuary have developed along

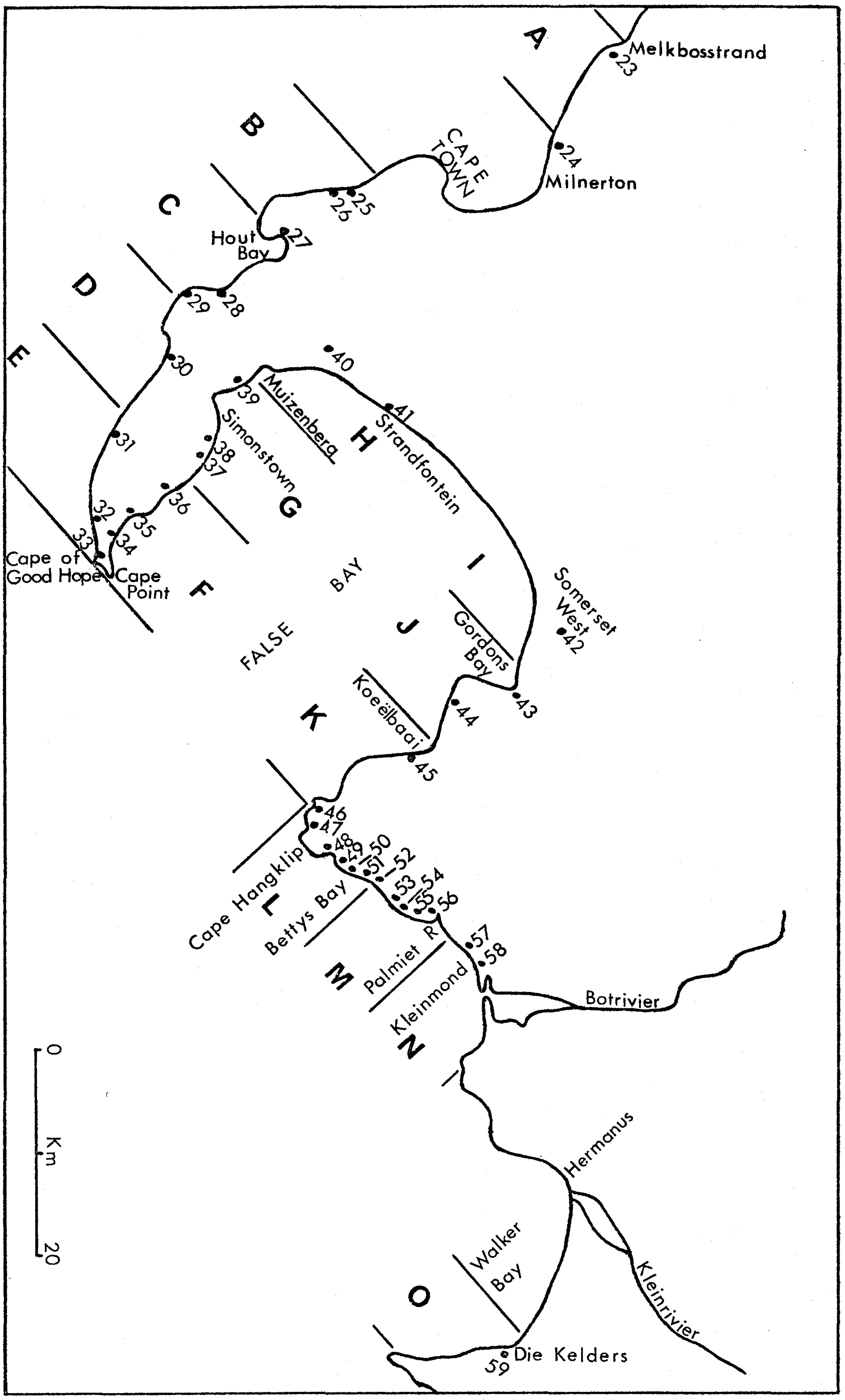


Figure 15.1 Locality map showing sampling sites (numbered) and sectors (lettered).

Figure 15.2 Platform cut across inland dipping quartzite; note absence of lower platform. Koëelbaai area.

Figure 15.3 Coastal platform on either side of Palmiet River. The Palmiet River valley extends below sea level. This photograph also shows the small amount of sedimentation that has taken place since the Flandrian flooding of the river valley. (Photograph by courtesy of Earldons (Pty) Ltd.).

synclinial Bokkeveld strata, the Klein River, its estuary, and the sandy extent of Walker Bay are the result of rapid erosion of a granite inlier.

III. ELEVATED SHORELINES

The results of a detailed survey of the coastal margin are summarised in Table 15.1. Most of these occurrences take the form of storm beaches which probably relate to MHWS. All elevations have been corrected to mean sea level (subtracting 0,7 m). In drawing up Table 15.1, the coastline has been divided into equally spaced sectors (A – O) so as to obtain a clearer impression of elevation trends. One consequence of this presentation is to show how beach development and elevation are functions of exposure. For instance, there are regular elevated shorelines at 6–7m a.s.l. (average 6,3 m) in sectors B–E and K–O, while a shoreline at this elevation is less pronounced along the western edge of False Bay (sectors F–H).

Table 15.1 shows significant variation in shoreline elevation. These variations reflect the problems in measuring old elevated sites.

TABLE 15.1

Summary of Elevated Shoreline Data (m)

Namaqualand	Saldanha Area	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O
2	0	1,5					1,3		0–1,3				1,3			0
5	2–3,5					2,6	2,3						3,1			
				4,5			4,6	4,3					4,0–4,7		4,6	4,6
7–8	6,3	6,9		6,6–7,1		5,8–6,3						6,1		6,3	6,2	6,3
	10,0						11,3					9,3		10,3	8,3	
	13,0															
					15,0			15,0						15,0		
17–21										20,0			20,0	20,0		
29–34			28,7								29,6–32,3	29,3–31,3				
45–50																
75–90																

Suggested Correlation

Namaqualand	Saldanha Area	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O
	0	1,5					1,3		0–1,3				1,3			0
2	2–3,5					2,6	2,3						3,1			
5				4,5			4,6	4,3					4,0–4,7		4,6	4,6
7–8	6,3	6,9		6,6–7,1		5,8–6,3						6,1		6,3	6,2	6,3
							11,3					9,3		10,3	8,3	
					15,0			15,0						15,0		
17–21	13,0									20,0			20,0	20,0		
29–34			28,7								29,6–32,3	29,3–31,3				
45–50	10,0															
75–90																

Problems encountered in measuring these elevations throw considerable doubt on previous claims (e.g. Davies 1971, 1972) that the low level shorelines occur at regularly spaced intervals. The first difficulty always encountered is the problem of recognising a fixed datum, usually mean sea level. How does one accurately identify mean sea level during a survey when the behaviour of the waves is controlled by degree of exposure and shoreface configuration? It was found in several cases that it was impossible to approach within 2 m of mean sea level even during calm seas.

Even if mean tide level is accurately approximated, what is the significance of the elevation measured? For instance, there is a prominent storm beach at Cape of Good Hope (site 33) which peaks at 6,2 m a.s.l (7,5m according to Davies 1972). The crest is vegetated, indicating that the storm beach is an old feature. The modern beach which rests on this ridge is composed of coarse shell detritus. A kelp line produced by storm waves is situated 2-3m above mean tide level. Although a washover fan situated behind the storm beach is also ancient, modern beach accretion is still taking place over this complex. Even though the crest elevation is corrected to MSL, it does not necessarily indicate a sea level as high as that.

At many sites (e.g. site 31) the old storm beach and the present boulder beach form a composite concave up profile, which displays effective size selection by the waves. Boulder-size material occupies the low tide level, and size decreases regularly to cobble size at the storm beach crest. Obviously size sorting is not merely a function of modern sea level, but is rather a product of repeated sea level oscillation.

A. The 30 m Shoreline

Beds composed of wave-generated beach boulders are exposed close to 30m a.s.l. at sites 25 (28,7 m), 44 (32,0 m) and 45 (30-32 m). Figure 15.4 illustrates a typical exposure of the 30 m shoreline above Koëelbaai, where shoreline retreat has exposed this old beach on the inner edge of a platform at the foot of the mountains, (i.e. it could not rise much

Figure 15.4 The 30 m beach exposed below talus in a road cut above Koëelbaai.

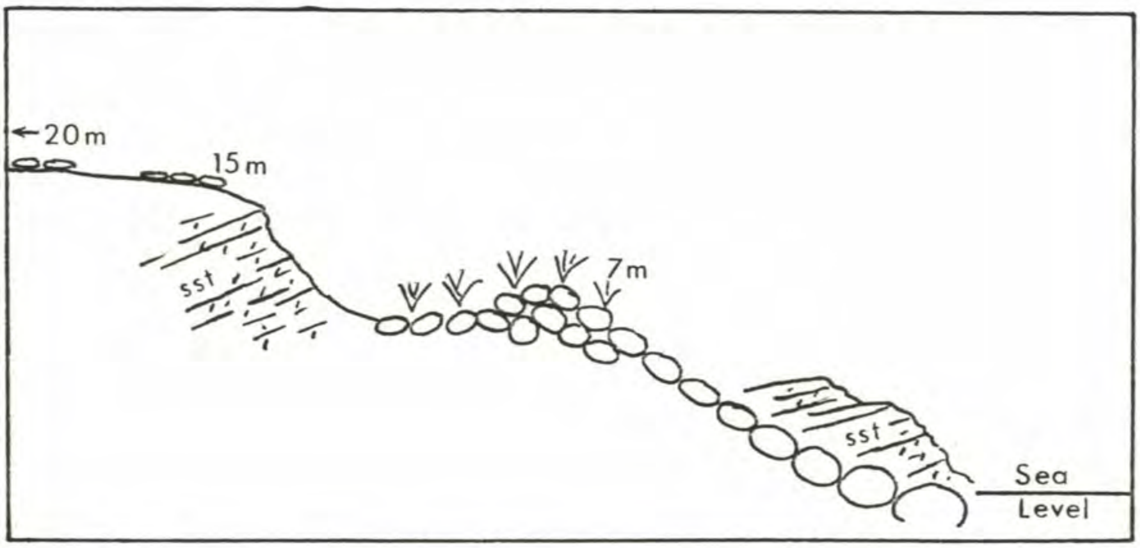


Figure 15.5 Illustration of storm beach development just west of the Palmiet River mouth. The inland dipping quartzite is sheared.

more above this elevation). The boulders are composed entirely of orthoquartzite. They are well rounded and have chattermarks. Maximum boulder size is 1 m. The bed is well sorted and contains no scree material, indicating little post-depositional surface creep.

On the Atlantic coast of the Cape Peninsula, this shoreline is exposed at 21 m at site 26, and 28,7 m at site 25. The beach boulders lie upon deeply weathered granite and vary considerably in elevation even over small distances, filling a platform of considerable relief.

B. Beaches between 8 and 20 m

Deposits of the 20 m shoreline occur sporadically on the coastal platform east of Gordons Bay (sites 47, 55, 56). Seldom do they form prominent horizons, suggesting that the 20 m sea level stillstand was minor compared with the 30 m event. Wave-generated cobbles and boulders occur on the platform west of the Palmiet River from 15,0 to 20,0 m a.s.l. The most widespread occurrence of the 20 m deposits is at Somerset West where they form a wide terrace protected on a re-entry in the mountainous terrain.

Even less significant are elevated shorelines at 15 m a.s.l. (sites 30, 37, 55), and 8,3-11,3 m a.s.l. (sites 36, 46, 54, 57) which form discontinuous horizons beneath talus deposits. Some of these low levels possibly also belong to the 15 m level. In some cases ridges of boulders pond vlei deposits. Extensive reaction with the acidic vlei waters has so completely corroded the boulders that all rounding has been removed.

C. Low Elevation Beaches

The low elevation littoral deposits include those up to about 7 m a.s.l. The best developed are those at 5,8-7,1 m and 4,0-4,6 m a.s.l. (Table 15.1). The average elevation of the higher set is 6,3 m a.s.l. (C I) which is equivalent to the 6,3 m shoreline of the west coast

(Chapter 13). The lower set (4,0–4,6 m shoreline) does not have a correlative on the west coast, and its development suggests that it was no more than a brief halt from the C I event. It has already been noted that the modern storm beach grades upwards into a vegetated component, and that the composite concave-up feature displays size sorting. Figure 15.5 shows a typical section through one of these storm beaches at the Palmiet River mouth. Average boulder size decreases from about 50 cm at low tide level to 15 cm at 3,5 m a.s.l. A recent excavation through the storm beach at site 50 shows imbricated cobbles up to 4,0 m a.s.l. Along other parts of the coast narrow sandy terraces with occasional shells fringe the sandstone platforms. The 6,2 m storm beach at Cape of Good Hope has already been cited as an example where modern beach accretion reaches its crest. In other areas (e.g. sites 28 and 32) 6,3 m ramparts are preserved on the inner edge of boulder terraces, while the outer edge of the terrace forms part of the modern beach complex.

Other beaches include those at 2–3,1 m a.s.l. which correlate with the 2–3,5 m shoreline (C II) of the Saldanha area, while a lower set at 0–1,5 m represents the C III shoreline. At Strandfontein this horizon takes the form of beach rock. Wave-generated boulders beneath the Middle Stone Age strata in Die Kelders I cave have been attributed to waves surging through the cave from a sea level approximately coincident with present datum (Tankard & Schweitzer 1974). The importance of the Die Kelders I horizon is that it occurs at the same elevation as the storm beach in front of the cave, and shows that the present storm beaches are possibly formed by repeated occupation of this level by interglacial seas.

A feature of all of these deposits, compared with those of the Saldanha area is the dearth of faunal remains. One site with an abundance of animal remains has recently been described by Kensley (in press). The site (Figure 15.1, site 24) is situated at approximately 1,5 m a.s.l. near the Milnerton lighthouse. Kensley identified 82 mollusc species in these deposits and noted that rock-dwelling forms predominated (59,3 per cent). Sand-dwelling and estuarine species formed 24,7 per cent and 8,6 per cent respectively. Kensley envisaged shell accumulation on a low energy beach, with the estuarine forms being contributed from flushing of nearby estuaries. The assemblage is dominated by Crepidula capensis praerugulosa (25,1 per cent). Twelve of the species are typical inhabitants of the warmer south coast.

In the vicinity of Melkbosstrand (site 23) Malmesbury rock forms a platform at -10 m. The platform is overlain first by a palaeosol and then by a coarse-grained shelly foreshore accretion unit up to 1,5 m a.s.l. The molluscs are modern in character, and usually very fragmented. The shell detritus, rounded cobbles, pebbles and the coarse sediment all indicate a high energy littoral origin, although it is thought that this deposit is contemporaneous with the Milnerton deposit described by Kensley. Furthermore, the palaeosol underlying these deposits is clear evidence that they are not merely temporary halts in the regression from the 6,3m (C I) shoreline. These occurrences suggest that the regression carried the shoreline below present sea level, and that the C II shoreline was the result of a new transgression.

The Cape Flats are underlain largely by Würm and Recent dune sands. Only on the western fringe north of Muizenberg (Sandvlei and Seekoevlei) is there evidence of shelly estuarine beds. These form a prominent terrace at 1,5 m a.s.l. The mollusc fauna is typically estuarine and includes species such as Bullia laevissima, Nassarius kraussianus, Turritella capensis, 'Rissoa' capensis, and Solen capensis. They suggest shallow water with a sandy substrate.

There is little evidence to suggest Holocene sea level higher than the present. But the Flandrian transgression was rapid (167 cm/century: Chapter 1). This led to drowning of the river valleys which were incised across the coastal platforms. Figure 15.3 illustrates the drowned valley of the Palmiet River, and shows how insignificant sedimentation has been since that flooding.

The upper reaches of the Bot River (30^{o} 16'S; 19^{o} 11'E) have a pronounced terrace underlain by immature but well-sorted and rounded, ferruginised river gravels at 67 m a.s.l. This terrace is extensive on both sides of the valley. A lower ferruginised gravel terrace is situated at 20-23 m a.s.l. nearer the mouth (34^{o} 22'S; 19^{o} 09'E). During the Eem the sea initially formed a deep funnel-shaped embayment at the mouth of the Bot River (Figure 15.6). But during the Holocene the shoreline has advanced by prograding spits with back filling of the area behind the spits. Later Stone Age midden material suggests accretion of 600 m in the last 2000 years.

IV. AGE AND CORRELATION

As far as the 30 m shoreline is concerned, there is hardly any means of dating it. But it is probably no younger than Early Pleistocene. Oxygen isotope studies of deep sea cores (Shackleton & Opdyke 1973, figure 9) show that glacio-eustatically sea level could not have exceeded present sea level by much in the last 800 ka. Shackleton and Opdyke argue that if the age of the 30 m shoreline was equivalent to their stage 9 (approximately 560 ka) it would have formed close to present sea level. To elevate such a beach to 30 m a.s.l., either by coastal uplift or tectono-eustatic regression, would ensure that the Eem (120 ka) shoreline at 6 m would today be recorded at +15 m. Since I am confident that this shoreline on the south coast is situated at 6 m a.s.l., it must be assumed that the 30 m event is of considerable antiquity.

It is possible to discuss the minimum age of the 20 m shoreline in more concrete terms. Acheulian handaxes and cleavers, and beach cobbles are at present being eroded from a black palaeosol at Cape Hangklip (Sampson 1974). Sampson could be correct in attributing this material to the 20 m shoreline. Furthermore, he describes the artefacts as Late Acheulian. Since Acheulian time stretches back to 700 ka (Klein 1974) from approximately 180 ka(Wendorf et al 1975), it would appear that the shoreline could be no younger than about 300 ka, and using the same arguments as for the 30 m sea level, it is suggested that this shoreline (20 m) possibly dates to the late Early to early Middle Pleistocene.

Davies (1971, 1972) cites numerous radiocarbon dates to suggest that the

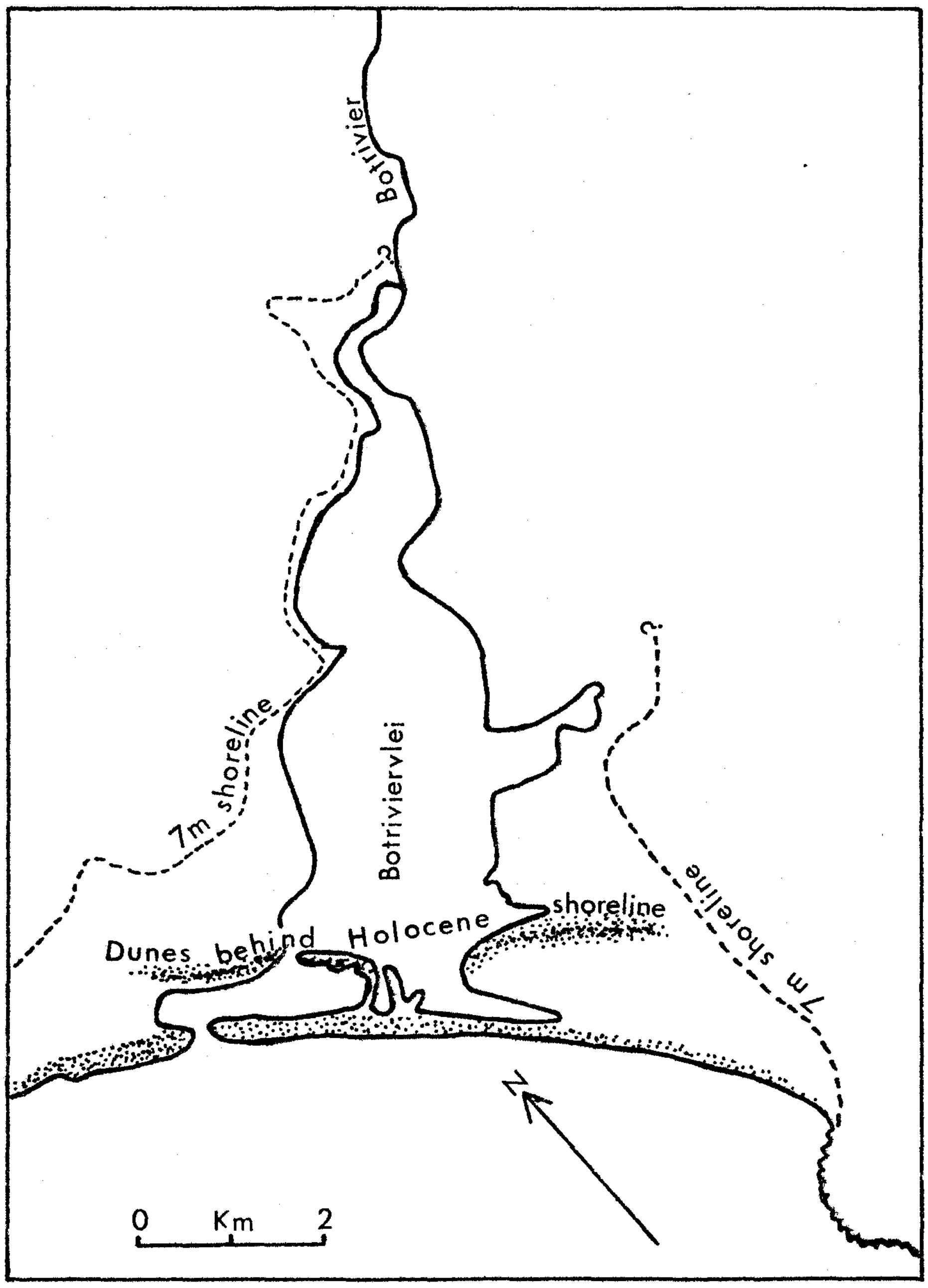


Figure 15.6 Botriviervlei showing last interglacial funnel shaped embayment. Modern accretion is shown by the inner dune ridges capping old spits.

6 m sea level of the southern and south-eastern Cape originated during the Würm interstadial. Detailed glacio-climatic studies, oxygen isotope studies, and reliably dated sequences prove the impossibility of his claim (discussed in detail in Chapter 1). All the dates quoted by Davies were derived from carbonate material which yielded ages older than 29 000 years. All of these dates should be taken as minima (see Chapter 1, and Davies 1973). The only young date (2300 years; Pta -445) derived from an Argobuccinum shell is used to suggest a Holocene age for the 5,7 m shoreline. Again this date may be meaningless since many of these storm beaches are capped with midden material and modern beach accretion. Two other dates (6870 B.P., Y-466; 1905 B.P., Y-467) from Sedgefield (Martin 1962) show a sea level close to present datum for this period. The dated beachrock at Muizenberg (25 860 B.P. and 25 430 B.P., Siesser 1974) and the molluscs at Milnerton (33 750 B.P.: I-8372, Kensley in press) give minimum ages which suggest that all beach deposits between present datum and 7 m are of last interglacial age. The wave-generated boulders in Die Kelders I, attributed to sea level equivalent to present datum, are overlain by Middle Stone Age sediments which confirm a last interglacial age.

The mollusc fossils discussed by Kensley (in press) show that the Milnerton beach deposits at 1,5 m a.s.l. are equivalent to the last interglacial deposits in the Elands Bay-Saldanha area. Furthermore, the Milnerton deposits contain a similar thermophilic fauna.

Table 15.1 shows a suggested correlation of shorelines in the Peninsula-Die Kelders area compared with Saldanha and Namaqualand sequences. There is no reliable data east of Die Kelders. It will be seen (Table 15.1) that the shoreline at 6,3m (average) provides a useful and fixed datum. But it should be pointed out that even though the coastline between Cape Town and Die Kelders has been examined fairly exhaustively, not enough sets of beaches at regular and small intervals were found to substantiate this correlation, even for distances shorter than 30 km.

V. AEOLIANITES

Aeolianite accumulations of Würm age occur along the coastal margin of the southern Cape. Cross-bed azimuths indicate deposition by southerly winds. The most impressive sequence is at Swartklip 8 km east of Muizenberg (36^{0} 06'S; 18^{0} 30'E) where they form a 25 m cliff above the beach. Professor K.W. Butzer is currently re-examining these aeolianites in detail. The only detailed geological description is that of Singer and Fuller (1962), while Hendey and Hendey (1968) have described the mammal fossils.

The cliff section consists of an alternating succession of coarse and medium-grained sands which are carbonate cemented, and several horizons of calcrete and ferruginous staining (palaeosols). Together these imply changing environmental conditions, while the palaeosols suggest periods of non-deposition during times of moister climate and possibly year-round precipitation (Tankard in press d, and Chapter 16). At Swartklip and Die Kelders the extension of the aeolianites below sea level implies formation during glacial times. Tankard and Schweitzer were able to show from Middle Stone Age artefacts in colluvial deposits that they formed during the Würm lowering of sea level. Furthermore, it was also shown that lithification had occurred prior to about 6000 B.P. (Chapter 16). The colluvial deposits were thought to have originated by movement of saturated regolith through a local karst terrain.

IV. DISCUSSION

A detailed examination has shown prominent terraces at 67 m a.s.l. (Bot River terrace), 30 m, 20 m, 6,3 m (C I), 4,4m (C IA), 2,3-3,1 m (C II) and 0-1,5 m (C III). It was suggested that the 30 m sea level could be no younger than Early Pleistocene, while the 20 m sea level could also possibly of late Early Pleistocene age. The shorelines from 6,3 m a.s.l. and lower would most likely be of Eem age, while the C II shoreline represents a transgression and not merely a halt in the regression from the C I level.

Perhaps the most valuable outcome of this part of the project is the negative aspect. The difficulties experienced in measuring any particular feature, its interpretation, problems experienced in dating, etc., cast considerable doubts on those studies which consistently find raised beaches at the 'right' elevations. For example, Davies (1972) assigns the estuarine beds with the Swartkops fauna to a 9 m sea level, when in fact most of these estuarine occurrences relate to a sea level close to 6 m a.s.l. It is not difficult to recognise raised beaches at practically any desired elevation. The difficulty lies in recognising the major stillstands of the sea and deriving a chronology. Recognition of the Late Pleistocene datum close to 6 m a.s.l. between Elands Bay and Die Kelders suggests comparative stability in the Late Pleistocene. Above this datum exposures are generally too far apart for adequate correlation without a mollusc fauna.

On the Peninsula considerable time was spent re-examining shoreline features described by other workers (Krige 1927, Gatehouse 1955, Davies 1972). In the case of Krige many of his higher shorelines are no more than structurally controlled terraces. But his work perhaps still remains the most original analysis, unconfused by any preconceived concepts. The value of his work lies in his recognition of sea level stillstands at 18 m and 6 m. He found that the major emergence shoreline rose from 15 m a.s.l. at Cape Town to 30 m a.s.l. at Infanta, before falling again to 13 to 15 m a.s.l. in the Mossel Bay area. The emergence seemed to Krige to be so ubiquitous as to suggest a lowering of sea level, with local subsidence of the coastal areas. But Krige's work certainly does not merit the general acceptance that has been bestowed upon it by South African geologists because of:

(i) incorrect identification of terraces; for example the Cape Flats, underlain by dune sand, are explained as a marine terrace, and

(ii) field evidence is accepted or rejected to suit his own prejudices; for example, a cave below Cape St Blaize lighthouse at 27,4 m a.s.l. is rejected as of marine origin because it is cut along the unconformity between

Peninsula sandstone and Enon conglomerate. Yet he does not similarly discard caves on the Robberg Peninsula which are cut along the interface between these same units. Finally,

(iii) Krige rejects other evidence of "strandlines" that do not coincide with his narrow constraints on the grounds of subsidence of that area.

These same objections can be raised to the more recent studies also. Gatehouse (1955) usually correctly identified shorelines up to 30 m a.s.l., but his reliability decreased with increasing elevation. Those shorelines above 30 m were either not re-discovered, or else they appear to be isolated midden remnants or structurally controlled terraces. Davies' (1972) account is very difficult to understand. He is of the opinion that the 6 m shoreline is a Würm interstadial feature, the 5,4 m shoreline a Holocene feature, and yet rejects the 7,5 m level recorded by others as an old and exhumed pre-Late Pleistocene feature. Davies incorrectly recognises terraces at 6 m and 1,5 m a.s.l. on granite promontories at Llandudno (1972, figure 22). He has taken no account of abrasion characteristics and the fact that platforms are generally graded to wave-base (see Chapter 13).

Finally, Davies (1970 - 1973) has extensively used archaeological implements as zone fossils. In the southern Cape only rarely are primitive tools found above the 60 m shoreline, but artefact abundance increases with decreasing altitude (Davies 1971). There appears to be little direct evidence of pre-Acheulian hominid occupation of the southern Cape (Klein 1974). Klein believes that the Acheulian of the southern Cape may span the Middle to Late Pleistocene time, from about 700 ka to 100 ka. He also suggests a possible span of the Middle Stone Age culture from 100 ka to about 40 ka.

To be able to use artefacts as zone fossils it is necessary to assume that they are older than the beach in which they were found, and that their inclusion is truly the work of the sea which deposited the host sediment. It is necessary that the artefacts show signs of marine abrasion. Objections against the use of artefacts are:

(i) little is known about the precise age of these artefacts;

(ii) if the artefacts are unworn, as all the illustrations in Davies' papers suggest, then they were not part of the original depositional environment;

(iv) if the artefacts are worn to suggest that they were an original inclusion, then it must be doubtful that they can be recognised. It would seem unlikely that vigorous abrasion on a cobble or boulder beach would always product just the required amount of abrasion;

(iv) I have only witnessed the collection of unworn artefacts from terrace surfaces and not *in situ*.

Against this background, and accepting that some artefacts may be genetically related to the sediments, I submit an alternative explanation for what must be the majority of *in situ* occurrences. The following argument is given in the hope that it will stimulate discussion of this problematical aspect.

Since the artefacts appear, from illustrations, to be remarkably fresh, and since they are found in such profusion, it would seem obvious that their occurrence is not genetically related to the sediments in which they are found and that they are, in fact, much younger than the terraces which they purport to date. It is my contention that Acheulian and Middle Stone Age peoples have found the flat remnants of terraces in an otherwise rugged terrain a very convenient place on which to live and scatter their artefacts in great profusion. The increasing quantity of artefacts with decreasing altitude (? or age) of the terraces merely suggests that the larger extent of the lower terraces would offer a more suitable site for habitation. The inclusion of the artefacts in a pseudo-stratigraphical context could have been the result of downward movement by erosion and undercutting, and the same processes of expansion and contraction that form the very conspicuous gravel lines seen in so many South African soil profiles. Downward movement would be arrested once the artefacts had come into contact with the firm gravel and boulder horizons. There is no mention in the literature whether the artefacts are an intimate part of the gravel or lie on the top of the gravels.

A comparison of the shoreline elevations in the area between the Cape Peninsula and Die Kelders with other recently published elevations for other parts of South Africa is presented in Table 15.2. The data from the Namaqualand coast, although appearing solely as a table in a taxonomic paper (Carrington & Kensley 1969) is based upon years of intensive mining operations in a small area north of Hondeklipbaai on the Namaqualand coast, and is probably the most reliable data available. The close agreement of these Namaqualand shorelines with those of the Algoa Bay area (Ruddock 1968) is interesting and confirms the apparent correctness of Ruddock's corrections for tilting.

TABLE 15.2

Comparisons of some published shoreline elevations for South Africa (m)

This Dissertation	Carrington & Kensley 1969	Ruddock 1968	Maud 1968	Davies 1970
			1,0	
0 -1,5			1,5	1,5
2,3-3,1	2		2,4	
4,0-4,7	5		4,5	3,5
5,8-6,9	7-8	6		6
8-11			8	9
			12	
15-20	17-21	18	18	18
		24		
29-32	29-34	30	33	30
				38
	45-50		45	48
		52		
		58		
		64		61
			70	73
	75-90	84		82
		91		
		106		
			115	110
				155
			170	

The most important aspect of Table 15.2 is the lack of agreement, even by two workers in the same province, Natal (Maud 1968, Davies 1970). Without precise stratigraphic control regional studies of high level shorelines prove to be no more than exercises in correlation. And since it is possible to find a raised beach at almost any desired elevation some sort of correlation is always possible. Apart from tabulating these various shoreline elevations (Table 15.2) there is little that can be said about them. It is not possible to suggest a chronology other than a relative one (i.e. the order of decreasing altitude is also that of decreasing age). Without a precise means of correlation and on the evidence available it is impossible to distinguish between eustatic effects and the effects of local tectonism.

CHAPTER 16

DIE KELDERS CAVE AND ITS PALAEOENVIRONMENTAL SIGNIFICANCE

I. INTRODUCTION

In recent years archaeology has become very much an interdisciplinary science. Descriptions of the cultural and dietary aspects of excavations, and cursory stratigraphic descriptions are now regarded as insufficient in any attempt to understand prehistoric man and his habitat. Geologists and archaeologists have now combined to extend the approach pioneered in Europe by Lais (1941). Techniques commonly used by sedimentologists to determine depositional environments have been successfully applied to cave sediments along the Cape south coast (Butzer 1973; Tankard 1974d, in press d; Tankard & Schweitzer 1974 and in press). These techniques include grain size analysis, surface textures of detrital grains (by electron microscopy) and sedimentary structures.

The importance of Die Kelders I is that it is situated in the winter rainfall area of South Africa, an understanding of which is important to an understanding of the climate of a greater part of South Africa. The cultural sequence in the cave ranges through the Middle Stone Age (MSA) and Later Stone Age (LSA). This represents potentially a time span of some 80 ka, i.e. the entire hypothermal period. In reality the MSA occupation may have been brief, perhaps as short as 25 000 years, and from 40 ka or 50 ka until 2000 B.P., the cave remained unoccupied. The MSA fauna includes seal and penguin, but no fish or mollusc remains, while the LSA is represented by a shell midden which contains vertebrate remains of a very different aspect to that of the MSA. The LSA vertebrate fauna includes sea birds, fish and seal.

The aim of this chapter is to discuss the changes in depositional environments occurring in Die Kelders I and environs, and to discuss the implications of these changes for our understanding of Late Quaternary environments.

Excavation of Die Kelders I was carried out from mid-1969 to early 1973 by Mr F.R. Schweitzer of the South African Museum.

II. GEOLOGICAL SETTING.

Die Kelders I is a north facing cave situated on Walker Bay 160 km south-east of Cape Town (34° 32,8'S; 19° 22,3'E) (Figure 16.1). The back of the cave is 8 m a.s.l. and 10 m inland from the high-watermark. The cave has formed along the unconformity between Palaeozoic Table Mountain sandstone (TMS) and Miocene marine limestone of the Bredasdorp Formation. The unconformity, which rises gently inland, was formed by planation during the Neogene transgression. The TMS dips in a northerly direction at 40° (Figure 16.2). Erosion along the strike of the sandstone produced a series of east-west passages which have acted as sluices for the movement of groundwater. The importance of the unconformity to an understanding of the cave depositional history lies in the fact that it forms an impermeable surface for the shoreward movement of groundwater from the mountains behind the coastal plain. This groundwater gives rise to an active spring near the cave, while sculpturing at the back of Die Kelders I shows that the cave too was a site of active groundwater movement in the past.

The frequent occurrence of caves situated at 7 to 10 m a.s.l. along the southern Cape coast suggests that they were formed during a period or periods of higher than present sea level. If the cave complex at Die Kelders was formed primarily by erosion by a high sea level (+6-8m), it is interesting that this also coincides with the height of the unconformity along this part of the coastline.

The cave complex originated from the fortuitous interaction of two dominant processes. Firstly, precipitation on the mountains behind the coastal plain results in surface run-off, and some water finds its way down to the impervious TMS and then flows down the north-westerly dipping unconformity. At about 120 ka there was a world-wide

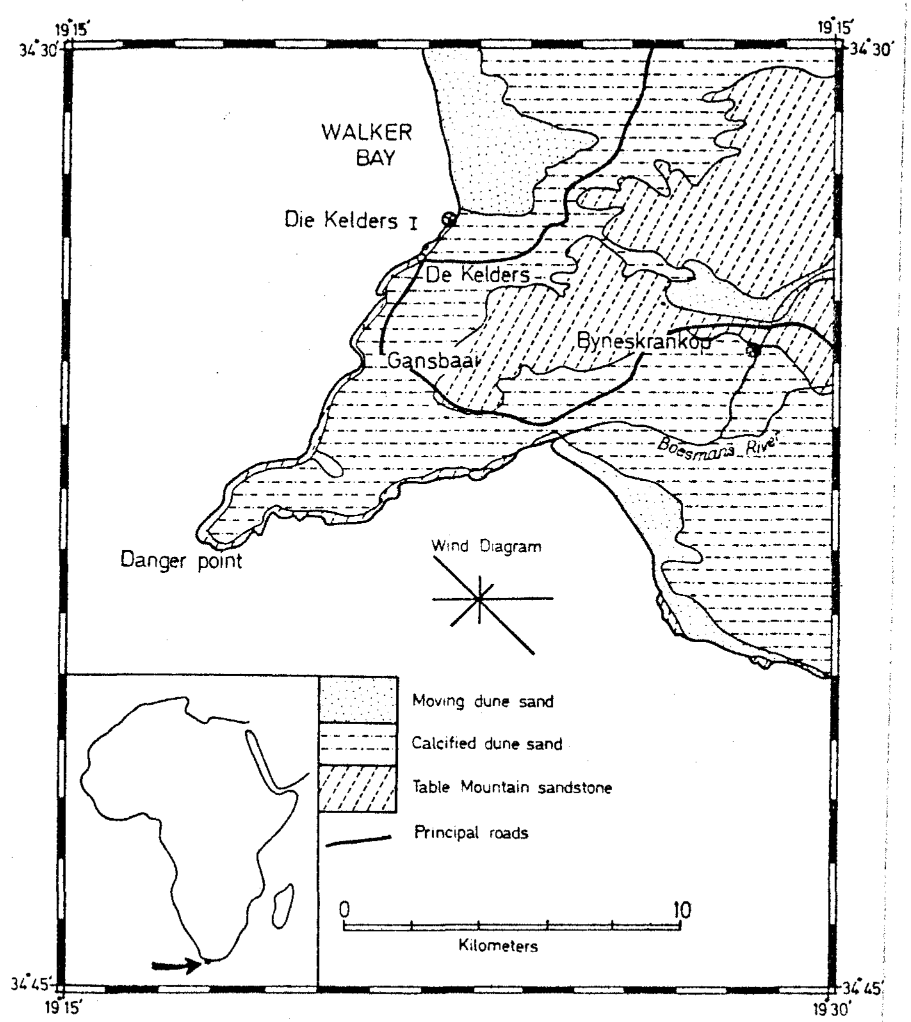


Figure 16.1 Locality map showing generalised geology and wind diagrams.

Figure 16.2 The Die Kelders cave complex. The excavated section is situated in the cave immediately behind the figures. Note the formation of the cave along an unconformity between steeply dipping sandstone and Bredasdorp limestone. Also note the northward dipping aeolianite above the cave (the lighter colour).

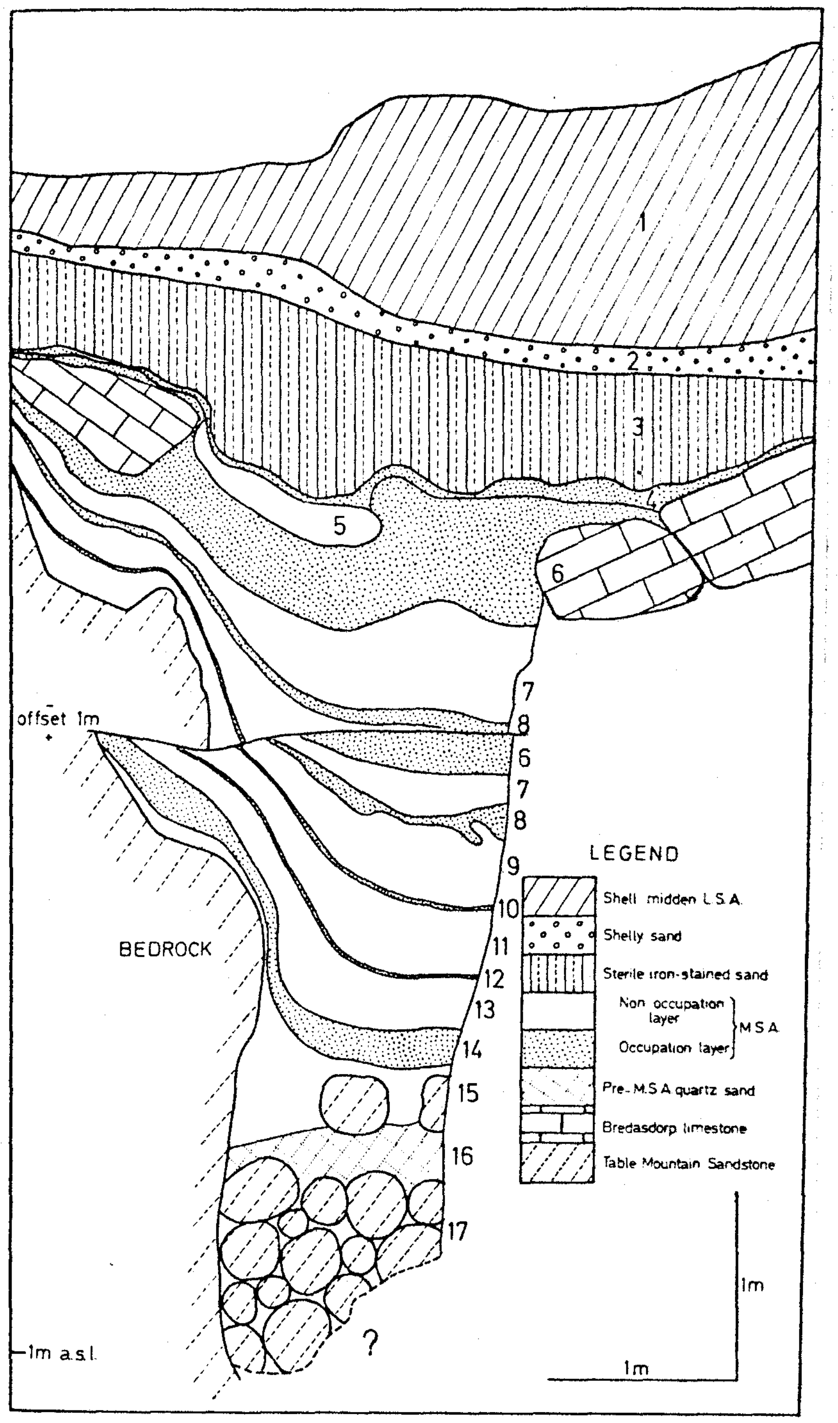


Figure 16.3 Cross-section of Die Kelders I stratigraphy showing numbered layers referred to in the text.

glacio-eustatic rise of sea level to 6-9 m above present sea level. The interaction of this higher sea level, the convenient occurrence of a plane of weakness (the TMS-Bredasdorp unconformity) in the coastal strata, and water flowing along this plane all contributed to formation of the cave complex. Enlargement of the cave has proceeded by chemical weathering of the limestone roof rock, frost weathering, and possibly even by collapse of the cave roof by a catastrophic earthquake shock (Tankard & Schweitzer 1974 and in press).

The second high sea level is marked by wave-generated beach boulders beneath the MSA succession. The altitude of this boulder bed (+2m) matches that of the beach rampart in front of the cave, suggesting that they were formed at the same time by a sea level coincident with present sea level. A minimum age for these beach boulders is fixed by the overlying MSA strata in the cave; MSA occupation appears to have followed soon after withdrawal of the sea. This suggests that the 0m sea level probably dates from late in the last interglacial.

Late Pleistocene aeolianites overlie the Bredasdorp limestone. Cross-bedding azimuths show that they were deposited by a southerly wind regime. Along the coast of Walker Bay the aeolianites obviously extend below present sea level and are now being cliffed by the present sea. The Walker Bay aeolianite cliffs, remnants of aeolianite on quartzite promontories, and truncated cross-beds above the cave complex (Figure 16.2) show that they were formed when sea level was lower and newly exposed sands on the continental shelf were blown inland by the southerly winds.

A poorly sorted outwash deposit containing MSA artefacts interbedded with the calcarenite occurs just north of the cave. The outwash deposit originated by movement of saturated regolith down a gentle slope. The calcarenites can be no older than the MSA. Rubble of the calcarenite forms a relict talus deposit immediately below the midden in Die Kelders I, suggesting that lithification had been completed before about 6000 B.P. Origin of the calcarenites during the last glacial while sea level was lower is thus confirmed.

III. GENERAL FEATURES OF THE CAVE DEPOSITS

A. Stratigraphy

Excavation at Die Kelders I has proved the existence of more than 7m of sediment preserved in an erosional strike-passage through the TMS. The stratigraphy of the cave deposits can be summarised as follows (from bottom to top, Figures 16.3 and 16.4):

(i) An incompressible horizon of quartzite "beach" boulders and coarse poorly sorted interstitial quartz sand containing weathered echinoid spines (layers 16 and 17).

(ii) Middle Stone Age sequence comprising alternating occupation and non-occupation layers. The occupation layers are compacted and characterised by cultural material and faunal remains. The non-occupation layers consist of fine to medium quartz sands (layers 4 through 15) and micro-fauna.

(iii) Sterile yellow sands, fine to medium grained, iron-stained (layer 3).

(iv) Shelly quartz sand, fine to medium grained, echinoid spines, foraminifera (layer 2). Talus material at the base.

(v) Late Stone Age shell middens with intercalated lenses of fine to medium calcareous sand (layer 1). Radiocarbon dates on charcoal samples range from 2020 B.P. to 1465 B.P. (Schweitzer 1970).

The history of the Die Kelders I deposits begins with accumulation of wave-generated "beach" boulders which arose from waves surging through the passage from a level approximately coincident with present sea

Figure 16.4 Die Kelders I excavation showing from top down: Later Stone Age shell midden; shelly quartzose sands with shell grains and foraminiferal and echinoderm remains; talus rubble; sterile yellow iron stained sands; massive roof rock boulders across the top of the Middle Stone Age succession.

Figure 16.5 Rhythmically bedded iron stained sands overlying roof rock collapse and underlying talus.

level. With the onset of glaciation in the high latitudes and the consequent lowering of sea level, the cave became available for occupation, and accumulation of the Middle Stone Age succession began.

The Middle Stone Age quartzose sediments (layers 4 through 15) have been modified by the addition of cultural debris (artefacts and bone), small rodent bones probably dropped by predators such as owls, spalled fragments of roof rock and humus. The mud-humus content reaches a maximum in occupation layers 6 and 14. The Middle Stone Age is represented in all by six occupations (layers 4, 6, 8, 10, 12 and 14) which are separated by sterile non-occupation layers (5, 7, 9, 11 and 13). The occupation layers are generally darker coloured than the non-occupation layers due to the higher mud and humus content, and characterised by artefacts and vertebrate remains. The occupation layers have a range of carbon values from 0,16 to 0,36 per cent, while the range in the non-occupation layers is 0,08 to 0,12 per cent (carbon determined by oxidation and titration). (Carbon in soils can be as much as 1 to 2 per cent). There is a total absence of marine shells suggesting that the shoreline was at that time some distance from the cave. But seal and penguin bones which are most abundant in the lowest occupation level do suggest that the sea was still exploited.

Occupation layer 6 near the top of the Middle Stone Age sequence is the most conspicuous, being 0,7 m thick. It is a dark yellowish brown colour (10YR 3.2-4.2: Rock Colour chart of the Rock Colour Committee 1970), but mottled red by hematite and yellow by limonite. The quartz grains in this horizon are extensively iron-stained. The underlying and older occupation layers are less distinct in terms of coloration and thickness, and considerable compaction has taken place due to leaching of the mud and humus. In layers 4, 10 and 12 leaching and compaction have resulted in the horizons being represented largely by artefacts and bone. In this respect it is envisaged that the strike-passage in which the cave sediments are preserved acted as a natural channel for drainage of groundwater and elutriation during accumulation of Middle Stone Age sediments.

An important component of the earliest MSA occupation layer and the succeeding non-occupation layer are flattish, angular, spalled flakes of roof rock. Unlike modern flakes of roof rock the edges are sharp and free of chemical weathering. Tankard and Schweitzer (1974) interpreted these angular fragments as the result of alternating freeze-thaw conditions. These éboulis secs diminish in quantity upwards through the MSA succession, until at the very top of the MSA, which is marked by a massive collapse of roof rock, they are totally absent. So extensive and massive are the tabular roof rock boulders that pneumatic drilling was required to remove them. It is thought that such an extensive collapse of roof rock could best be explained by a catastrophic earthquake shock. Frost action is an unlikely explanation because of the limited amount of small cryoclastic debris. But it should be noted that recent earthquakes (1969) have collapsed cliff rock along the present shore.

Extensive weathering of the collapsed roof rock boulders and leaching of the last occupation layer to the extent that it was reduced to a thin condensed layer demonstrate a lengthy period of non-deposition following the MSA occupation of the cave. The next cycle of deposition (layer 3) took place into standing water which was ponded between the back of the cave and the quartzite bar in the front. An alternating sequence of yellow medium bedded quartzose sands and laminae of red quartzose sands demonstrates rapid and rhythmic deposition. The broad yellow bands (? limonite stained) are thought to represent the wet part of the cycle, and the red bands (? hematite stained) a dry phase (Tankard & Schweitzer 1974). The only faunal remains consist of very occasional rodent bones in the red bands. Otherwise the sequence is sterile. There are no shell grains or éboulis secs. The top of this sterile unit is marked by a micro-soil horizon (Figure 16.6) which suggests prolonged stability of the cave environment.

By the time deposition was renewed a new environment prevailed. Shelly quartzose sands and talus rubble, with no iron staining, overlie the yellow sterile unit. The bioclastic component includes foraminiferal tests and echinoid spines. Complete echinoid spines demonstrate the nearness of the shoreline. Finally, the cave was reoccupied by Later Stone Age people who left behind an extensive shell midden.

B. Structure

Onset of high latitude glaciation caused a lowering of sea level which left the cave free for occupation by MSA people. But occupation of the cave was not immediate. Layers 16 and 17 at the base of the succession are incompressible sands and "beach"boulders. Figure 16.3 clearly illustrates how the Middle Stone Age succession consists of alternating occupation and non-occupation strata which dip towards the centre of the cave. On the east face (Figure 16.3) dips range from about 35° to nearly vertical against the quartzite wall rock, but flatten out to a subhorizontal attitude away from the wall rock. Sections of the south wall show that the general dip is 20 to 25° to the centre of the cave. It is most significant that the dips of 35° to 90° against the wall rock are greater than the angle-of-repose for clean, dry quartz sands.

Occupation layer 14 is more loamy in content than either the sediments above or below. But microscopic examination shows that the quartz grains from underlying layer 15 are humus coated. The vertical dip in the central part of layer 14 could only have arisen by compaction or removal of part of the underlying layer (15). Since it is difficult to account for a massive removal of quartz sand, it must be assumed that layer 15 at one stage contained an appreciable mud-humus content. Assuming that at the time of formation of layer 14, the occupants lived on an almost horizontal surface, the present configuration of the base of this layer suggests that the compaction of the underlying sediment (layer 15) has been of the order of 75 to 80 per cent. Compaction undoubtedly resulted from post-depositional leaching of organic matter, and possibly by elutriation of fine material from the sediment by groundwater using the strike-gully as a sluice. Upwards through the MSA succession the dip of each succeeding occupation layer becomes less and shows that compaction has operated throughout the history of the cave sediments, and is probably still operational. Compaction has been so extensive as to leave some occupation layers (4, 10 and 12) represented only by condensed layers of artefacts and bones which in layer 4 amounted to 60–80 per cent. But the occupation layers still have a higher loam content than the intervening non-occupation layers, as a result of which they are darker coloured.

Whereas the laminae in the sediments overlying the Middle Stone Age horizons are always sharply defined, the laminae in the Middle Stone Age succession are wavy and considerably less distinct, a natural response to compaction. Compaction has also resulted in the formation of numerous microfaults in the sediments (Figure 16.6).

The laminae of the overlying yellow sands are always sharply defined (Figure 16.5), showing a minimum effect of compaction of the underlying strata. The maximum dip is 20 to 25^{o} towards the back of the cave (southward) while the sediments are horizontally-bedded on the south wall. There is a general expansion of the laminae towards the back of the cave, illustrating clearly that the top of the Middle Stone Age succession formed a basin at the time of deposition.

The sterile sands dip towards the back of the cave, and individual laminae also expand in that direction, i.e. basinward. The highest elevation of the sterile sands is against the TMS bar at the cave mouth. Rhythmic bedding is typical of deposition into standing water. Broad bands of yellowish sands indicate rapid deposition into standing water. The narrow bands of red sands which separate the yellow layers represent drier periods. The only faunal component, small rodent bones, occurs along these dry-phase layers. The standing water into which deposition took place was ponded by the TMS bar across the cave entrance. The water probably originated by flow along the TMS-limestone unconformity. Numerous microfaults and micro-thrustfaults in layer 3 are the direct result of compaction due to dewatering.

Pink shelly quartz sands (layer 2) overlie the yellow sands. The interface is sharply demarcated, and sometimes erosional. Initially

Figure 16.6 Rhythmically bedded yellow sands. Note microfaulting due probably to continued compaction of underlying MSA strata and dewatering. Also note micro-soil at top of yellow sands.

Figure 16.7 West wall section showing talus rubble dipping towards back of cave.

deposition of the shelly quartz sand was into standing water, but in the main they represent dry-phase deposition. There are two important constituents of this horizon. The sand-grade calcareous component consists of well-rounded shell grains, echinoid spines, and some foraminifera. On a larger scale are angular blocks of aeolian calcarenite, derived from the calcarenites outside the cave and forming a talus deposit within the cave (Figure 16.7). Calcarenite rubble gravitated down the flank of a dune which had accumulated in front of the cave. The surface of the talus slope formed a basin in which sedimentation once again took place briefly into standing water. This bed includes some lime mud.

IV. TEXTURAL ANALYSES

A. Grain Size Distribution

Random sampling from the various levels of Die Kelders I sediments was carried out by Mr F.R. Schweitzer during the four years of excavation. The sampling was supplemented by the writer on several visits to the cave during 1973.

1. Mineralogy and Grain Shape

The major component of the cave sediments is quartz, with a heavy mineral fraction of less than one per cent consisting of ilmenite with minor amounts of garnet. The sediments in the Middle Stone Age horizons tend to have a significant humus content. Iron-staining of the quartz grain surfaces increases upwards through the stratigraphic column to the top of the yellow sands (layer 3).

The sediments overlying the yellow sands are composed of quartz sand and shell fragments. The $CaCO_3$ content of this bed (layer 2) ranges upwards from 53 per cent. The shell midden (layer 1) is composed very largely of whole and fragmented marine mollusc shells with ash and a very minor fraction of quartz sand.

The quartz grains range from angular to round with the majority subround. The sediments low in the stratigraphic column tend to be subangular to subround, with the very coarsest grains being angular. Generally the sphericity and degree of rounding increase with diminishing grain size. In layers 1 and 2 the quartz grains are better rounded than those of the underlying sediments. The quartz sands contain some euhedral quartz crystals, generally with well defined and polished facets, although frosting has occurred in some instances.

From the basal marine unit upwards through layer 3 the grain surfaces are lightly polished, with occasional grains displaying better polish or frosting. In layers 1 and 2 with the improvement in rounding there is also a significant increase in the degree of frosting, probably due to post-depositional diagenesis.

2. Cumulative Size Frequence Distribution Curve

The cumulative curve (Figure 16.8) depicts graphically the grain size variation in Die Kelders I sediments. The textural classes suggested by Cornwall (1958) have been adopted. The cumulative curve is semiquantitative with respect to layers 6 and 17 because in both these levels the exposure allowed only an estimate of boulder size. The blanket of collapsed roof rock at the top of layer 6 is not shown.

In layer 17 the "beach" boulders and cobbles constitute about 75 per cent of the material. The quartzite wave-generated boulders are water-worn round to subround. Again in layer 6 there is a high proportion of rubble and large tabular boulders of collapsed roof rock. The mud content throughout the column is generally less than 1 per cent, with two notable exceptions. Occupation layers 6 and 14 have a high mud-humus content and these correspond to dark colouration (10 YR 3.2-4.2).

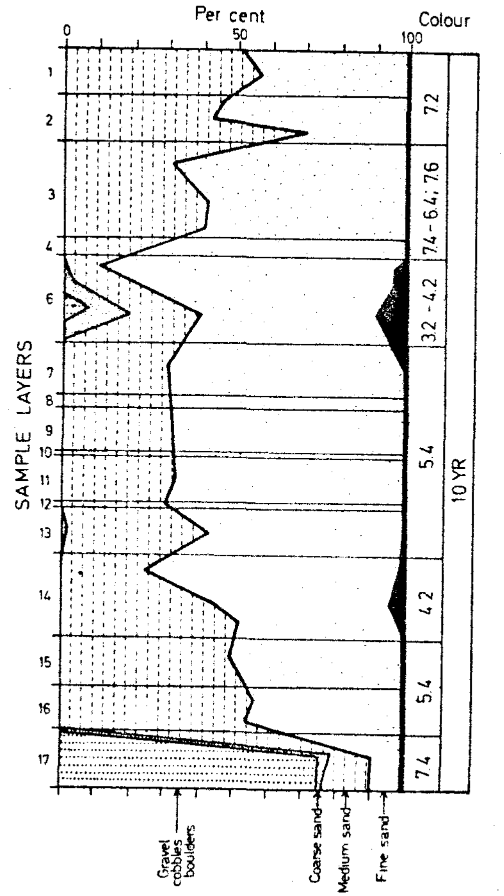


Figure 16. 8 Cumulative curve showing the broad constitution of the cave sediments.

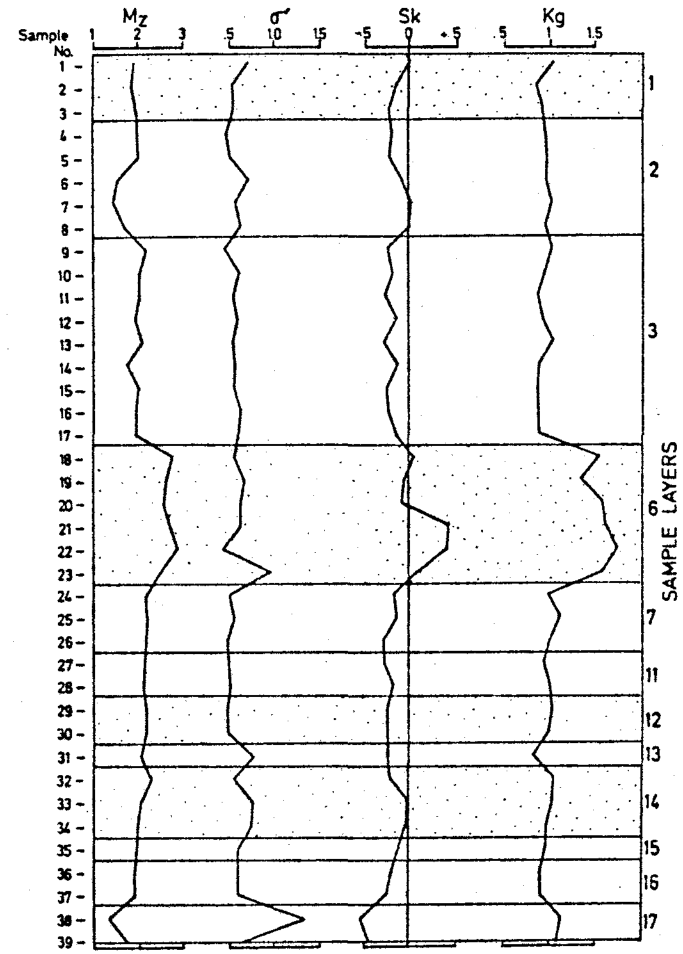


Figure 16.9 Variation in grain size statistics throughout the cave succession.

3. Character of the Sand Fraction

The sediments of Die Kelders I have so far been considered in totality, i.e. sediments transported into the cave from some outside source and which have subsequently been modified by roof rock debris and local loam formation. In this section the sand fraction will be examined quantitatively with a view to understanding the source of the sediment and, of course, the environment outside the cave at the time of sediment accumulation. The sieved fractions of the sediments were examined microscopically and macroscopically and all cryoclastic, lithic and bone material removed. The shell fraction in layers 1 and 2 has not been removed since the shell particles have, like the quartz sand, been transported into the cave from an outside source.

Grain-size distribution analysis of the sediment samples was done by sieving at half-phi intervals and computing the textural parameters from the graphic formulae of Folk and Ward (1957). (Grain size statistics are summarised in Appendix 1.).

(i) Mean

The mean sediment size virtually throughout the sediment column is close to 2 Ø (Figure 16.9). The range of mean grain size measurements is 1,38 to 2,79 Ø with an average value of 2,06 Ø and a standard deviation of the mean size 0,32 Ø. The sand is thus classified as fine sand, but close to the medium sand-fine sand boundary.

The coarsest sediment is the interstitial component from the boulder bed at the base of the succession. If coarseness of the sediment is a reflection of current strength, then a substantial current is indicated. This would be expected from the locality of the marine horizon where waves probably surged into the cave.

Coarser sand (ca 1,50 Ø) is again encountered in layer 2 due to the addition of much platey shell debris which occurs in the coarser grades, but which is probably hydraulically equivalent to the finer quartz. Decrease in grain size (average 2,61 ∅) in the Middle Stone Age succession is the result of admixture of a mud-humus fraction. In occupation layer 12 the mud fraction has been almost totally removed by leaching, with subsequent compaction, and the grain size is now marginally coarser than the sediment as a whole.

(ii) Standard Deviation

To describe the sorting of the cave sediments the classification of Folk (1966) is used. The plot of standard deviation (Figure 16.9) shows that the sediments average moderately well sorted (average 0,61; standard deviation of the standard deviation 0,15; distribution unimodal). In the basal marine unit the interstitial sand is moderately to very poorly sorted, suggesting an inefficient transporting medium. Sorting is somewhat variable throughout the Middle Stone Age succession, largely due to admixture of "fines". The yellow iron-stained sands (layer 3) contrast markedly with the Middle Stone Age sediments. Here the spread of sorting values is small (mean 0,57; standard deviation 0,07), where two-thirds of the samples have standard deviations between 0,50 and 0,64.

(iii) Skewness

Skewness measures the non-normality of a population. The Die Kelders I sediments are nearly all slightly negatively skewed, implying that the size frequency distribution has a tail towards the coarser grades. Generally skewness is between -0,10 and -0,25 (Figure 16.9) but with a range +0,42 to -0,53. The interstitial sands from the boulder bed (layer 17) are markedly negatively skewed (-0,43 and -0,53). This is undoubtedly the result of surges of sea water flushing out the "fines". Admixture of mud and humus in occupation layer 6 has produced positive skewness. While the negative skewness of layer 3 is due mainly to the absence of fine material, in layers 1 and 2 the negative skewness has been produced by admixture of platey shell material which occurs in the coarse grades.

The significance of the skewness values in these cave sediments is difficult to interpret. Folk and Ward (1957), Mason and Folk (1958), Friedman (1961) and Duane (1964) agree that beach sands are generally negatively skewed and dune sands positively skewed. In this respect the interstitial sand from the boulder bed can safely be interpreted as marine. Friedman (op cit) explains the negative skewness of beach sands as due to the operation of two opposing forces, the incoming wave and the outgoing wash, which effectively winnow out the fine material. Positive skewness of dune sands he attributes to unidirectional flow.

(iv) Kurtosis

Kurtosis compares the sorting in the "tails" with the sorting in the central part. Normal curves have Kg equal to one. The Die Kelders I sediments are predominantly platykurtic to mesokurtic (platykurtic = 0,67 - 0,90; mesokurtic = 0,90 - 1,11) with values close to one. The range of kurtosis values is 0,81 - 1,74. The largest deviation from the normal curve is in layer 6 where the sediment is leptokurtic (leptokurtic = 1,1 - 1,50): that is, better sorted in the central part than in the tails.

(v) Scatter Diagrams

By themselves the individual grain size parameters do not tell us much about the origin of the sediment, although they do define the sediments quantitatively. By plotting the grain size statistics against each other in 6 two-variant scatter diagrams their geological significance is revealed.

In Figure 16.10 the mean grain size for the Die Kelders I sediments is plotted against standard deviation (sorting). The sorting is

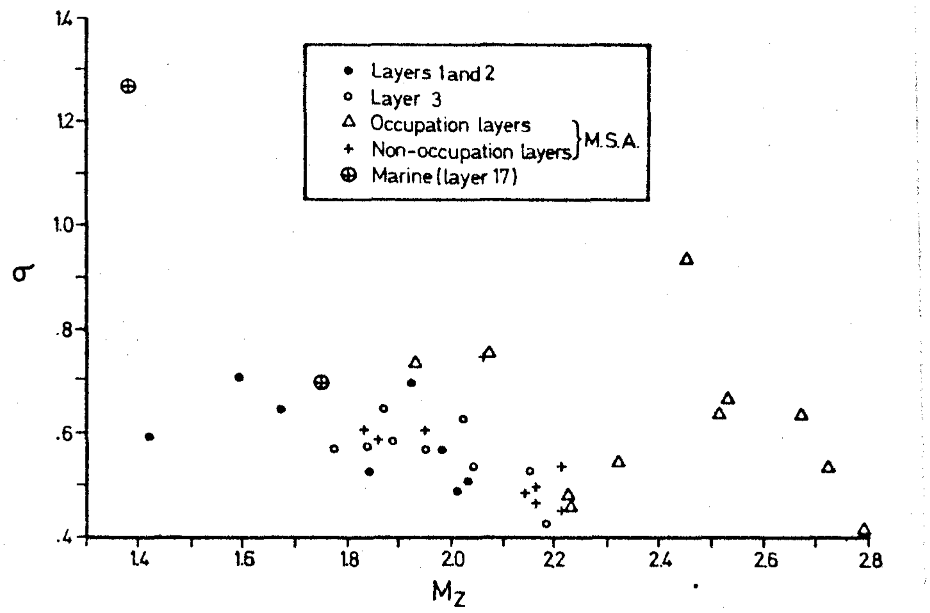


Figure 16.10 Scatter plot of mean against standard deviation.

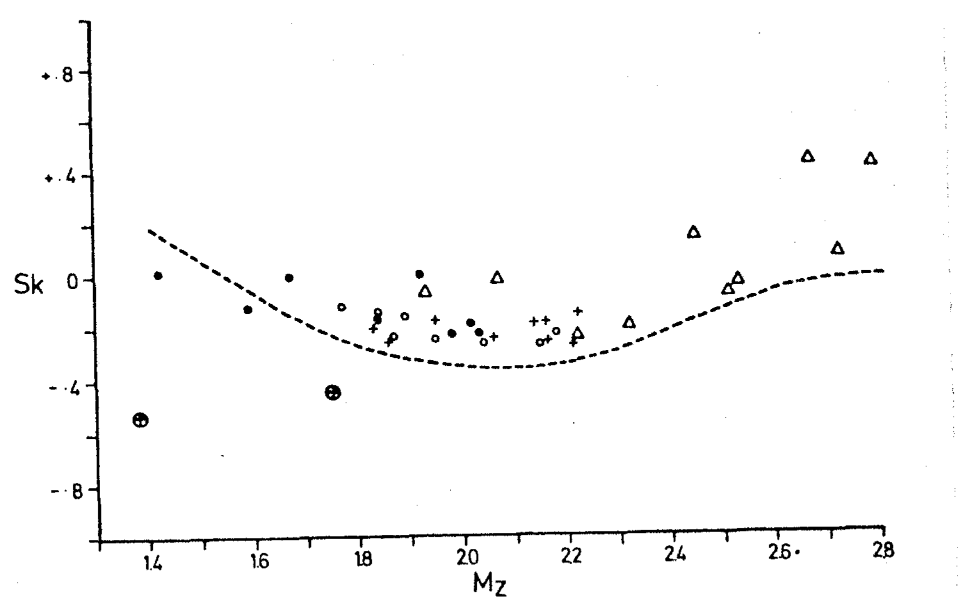


Figure 16. 11 Scatter plot of mean against skewness.

practically independent of mean grain size, although sorting is poorest with the coarse marine sand and the fine loamy sand from occupation layer 6. Best sorting is associated with the most prominent sediment mode.

In Figure 16.11 mean grain size is plotted against skewness. Apart from the two marine samples, the path followed by all other size-skewness statistics is arcuate. Coarser than 2,4 Ø the sediments are generally negatively skewed. The decrease in grain size below 2,4 Ø due to addition of a mud-humus mode to the sand mode is related to an increase in positive skewness.

It was mentioned earlier in a discussion of skewness that many authors have found dune sands to be generally negatively skewed and beach sands positively skewed. It was suggested that in the case of Die Kelders I sediments the sign of skewness was perhaps not necessarily diagnostic. If the sands had originated from dune accumulation, it is likely that they had been dumped over the lip of the cave, from the higher ground above, and rapid burial would not favour efficient sorting. Friedman (1961) plotted mean grain size against skewness and found an almost complete separation of the dune sand and beach sand statistic plots. The dividing line between these two fields computed by Friedman is shown in Figure 16.11. It is observed that the two undisputed marine sediments lie well within the beach sand area, while only two of the other 37 samples lie just within the beach sand realm. It is argued that a dune origin for these sands is highly likely. In a cave environment sorting would be negligible and the skewness values would be expected to be inherited. Positive skewness of the samples finer than 2,4 Ø is not inherited; it is due to the addition of small amounts (3-10 per cent) of mud-humus in the primary sand mode. It was also shown earlier that the Middle Stone Age succession has been compacted considerably, probably due to leaching of humus and elutriation of the silt-clay fraction. Negative skewness of many of the samples could possibly have arisen in such a fashion.

In Figure 16.12 mean grain size is plotted against kurtosis. It will be seen that two separate fields occur. For the bulk of the samples

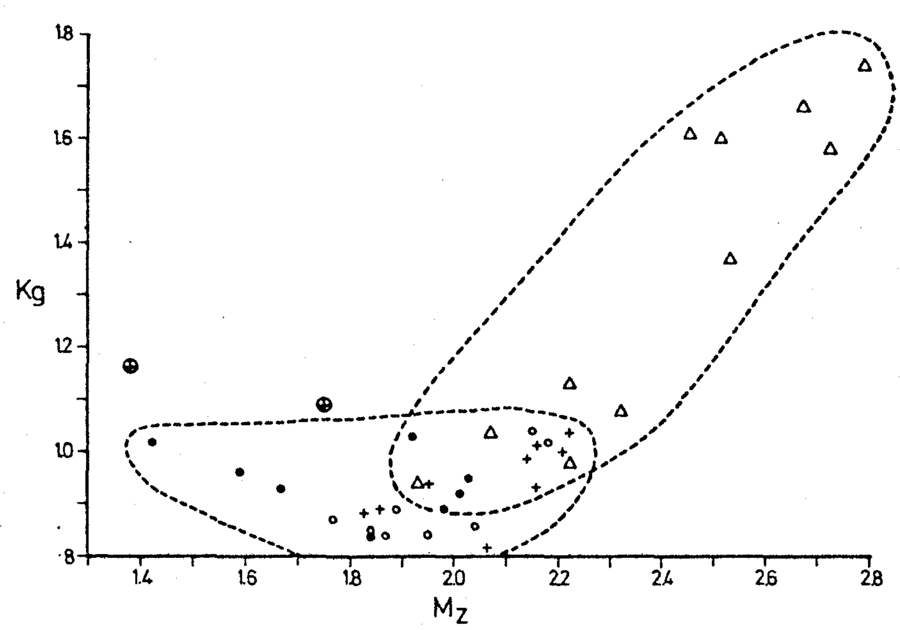


Figure 16.12 Scatter plot of mean against kurtosis.

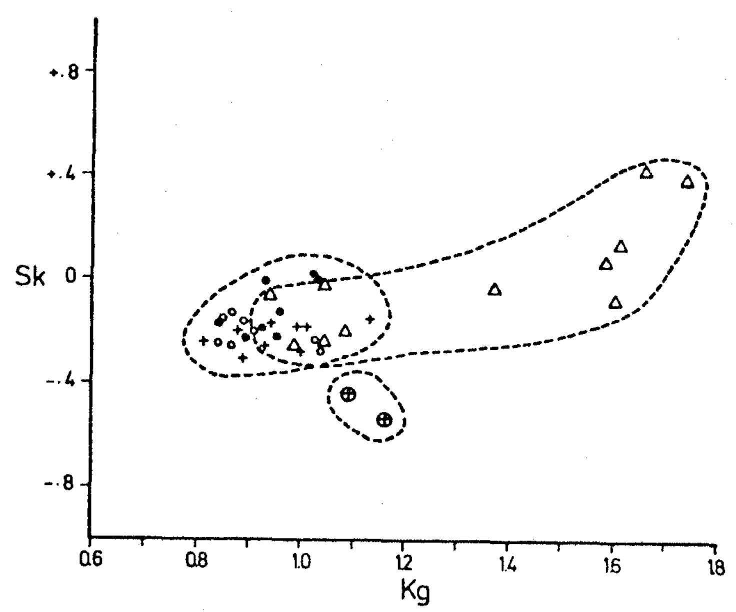


Figure 16.13 Scatter plot of skewness against kurtosis.

kurtosis values range from 0,80 to 1,05, showing that the sand mode gives a near normal curve. In the occupation layers the admixture of 3 to 10 per cent fine material has resulted in not only a decrease in mean grain size, but also a concomitant increase in kurtosis. This addition of a fine mode results in a poorer sorting in the tails when compared with the central part. The occupation layer sediments tend towards leptokurtosity ($kg > 1,5$). Of most significance is the fact that with progressive leaching and elutriation of the humus-mud fraction, the grain size statistics approach the field of the average "clean" sediment with kurtosis less than 1,05. In other words, a common origin for the sand fraction of these sediments is indicated.

The same trend is revealed in the other scatter diagrams; skewness versus kurtosis (Figure 16.13), skewness versus standard deviation (Figure 16.14), and standard deviation versus kurtosis (Figure 16.15). In all of these scatter plots it is observed that the marine sands at the base of the Die Kelders I succession are separated from the other sediment size statistics. For most of the sands the plots are very concentrated, suggesting a common origin. The only wide scatter of points occurs in the occupation layers where the fine mode ranges from 3 to 10 per cent. With progressive post-depositional removal of this fine mode, the sand fraction approaches the other tight cluster of points. In Figure 16.13 in particular high positive skewness due to admixture of the fine mode in the cave sediment correlates with increasing leptokurtosity as the sorting in the tails becomes progressively poorer compared with the central part.

B. Scanning Electron Microscopy

During transportation, deposition and compaction detrital grains may be mechanically abraded and chemically altered. A voluminous literature discusses the consequent surface textures, e.g. Krinsley & Takahashi (1962), Krinsley et al (1964), Krinsley & Donahue (1968a), Doornkamp & Krinsley (1971), Margolis & Krinsley (1971), Coch & Krinsley (1971), Krinsley & Margolis (1971), Blackwelder & Pilkey (1972), Tankard &

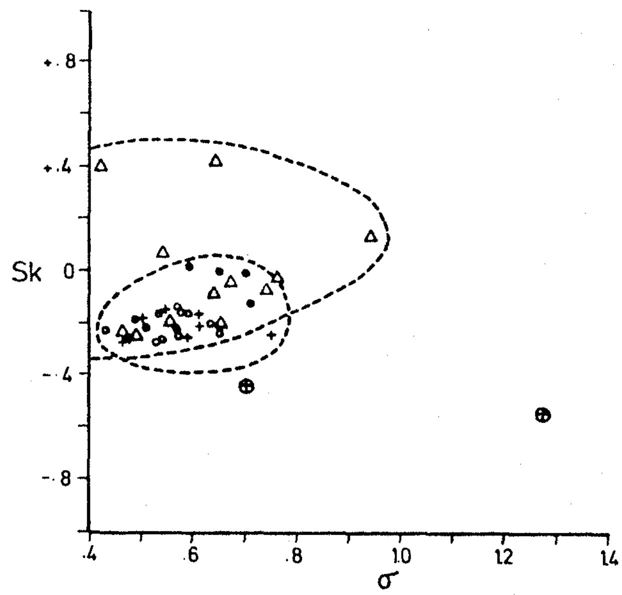


Figure 16.14 Scatter plot of skewness against standard deviation.

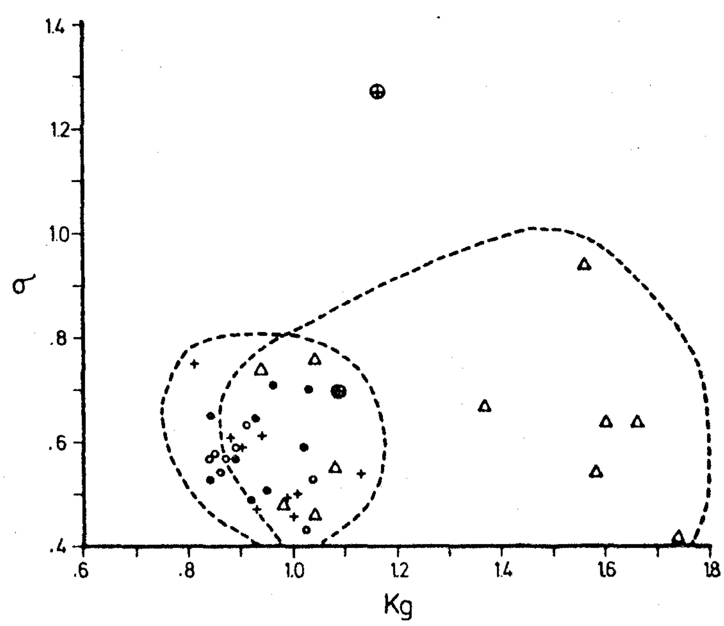


Figure 16.15 Scatter plot of standard deviation against kurtosis.

Krinsley (1974). Krinsley & Takahashi (1962) and Krinsley & Margolis (1969, 1971) have listed criteria whereby the various depositional environments can be recognised. Littoral textures are characterised by V-shaped indentations, straight or slightly curved grooves, and blocky conchoidal breakage patterns. A medium to low-energy beach environment produces en echelon arranged V-patterns, but these become randomly oriented as energy increases to high-energy surf conditions. Typical aeolian textures are meandering ridges, graded arcs, flat pitted areas and upturned plates. All of these textures may be modified by solution and precipitation to produce an undulating topography. These established criteria are the result of the analysis of many thousands of grains from modern unconsolidated deposits of known origin (Krinsley & Margolis 1971).

1. Results

Typical detrital grains observed in the Die Kelders I sedimentary succession are illustrated in Figure 16.16. The first grain illustrated (Figure 16.16A) is a rounded quartz grain from layer 2. Its roundness is inherited from a littoral environment. The well rounded shell grain (Figure 16.16B) is from the same sample. Its superior roundness is attributed to the softer nature and hence easier abrasion of the shell carbonate. Aeolian grains (e.g. Figure 16.16C) from layer 3 are characteristically more poorly rounded than the layer 2 and layer 1 quartz grains. But it is found that in degree of rounding and frosting they are identical to the Pleistocene aeolianites occurring on the high terrain above the cave. Post-depositional diagenesis results in solution and reprecipitation phenomena, and in Figure 16.16 D new crystal overgrowths on an older detrital grain are illustrated.

(i) Textures of littoral origin

A typical quartz grain from the sands which occur interstitially in the layer 17 "beach" boulders is illustrated in Figure 16.16 E. The

A Quartz grain rounded in littoral zone (layer 2).

B Shell grain rounded in littoral zone (layer 2).

C Dune quartz grain (layer 3).

D Development of new crystal faces on quartz grain.

E Beach environment: mechanical gouge marks (layer 17, quartz grain).

Figure 16.16 Scanning electron photomicrographs.

conchoidal gouge marks on the grain edge are probably the result of grain collisions caused by sea water surging through the cave during the last interglacial.

Some of the types of modification a quartz grain may undergo in a littoral or beach environment are illustrated in Figure 16.17. The V-shaped patterns (v) are the most common feature observed on the rounded edges of beach quartz grains at high magnification (X5000). They are triangular indentations with irregular edges. Randomly oriented V-notches as illustrated in this photograph are believed to be the result of gouging on impact of grains in a high-energy sub-aqueous environment (Krinsley & Donahue 1968a). Also shown in this photograph are blocky conchoidal breakage patterns (b) which characterise medium to high energy conditions. Also the product of a high energy environment are the straight grooves or scratches (g), which on this grain are 1 to 2 microns in length. According to Krinsley and Donahue the grooves are generally associated with non-oriented V-notches and are used as a criterion for wave action. These same workers believe that the grooves are caused by the scratching of the sand grain surface by the sharp edge of another grain.

Krinsley (1973) has found that mechanical V-shaped patterns and mechanical grooves decrease in number and the angles between the arms of the Vs increase with diminishing grain size, eventually to vanish at grain diameter of 300 microns.

Sand grains from low-energy beaches are characterised by <u>en echelon</u> or oriented V-shaped patterns which are invariably larger than the mechanical Vs described above. They result probably from chemical etching in sea water (Krinsley & Donahue 1968; Krinsley & Margolis 1969). The orientation is undoubtedly controlled by the lattice structure of the quartz.

High energy littoral textures appear on sand grains from layers 1, 2 and 17. No low energy features have been found. Layers 1 and 2 also contain well-rounded shell grains and echinoid spines and

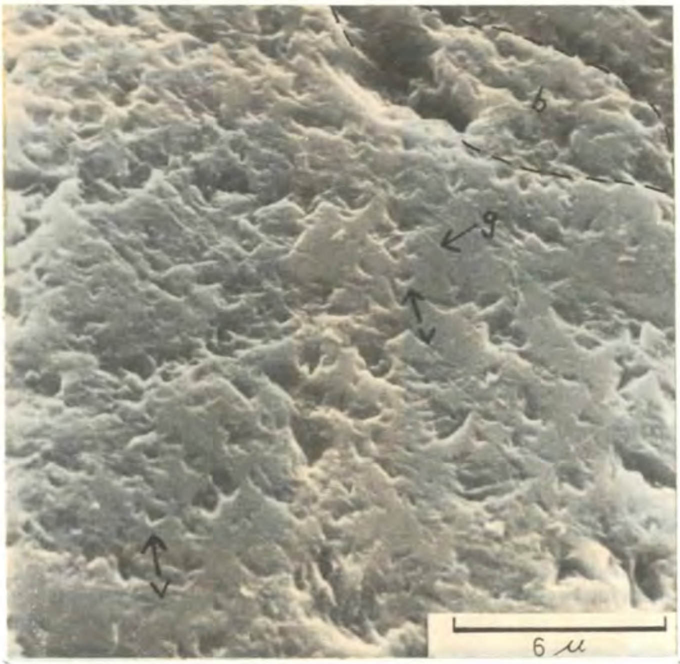


Figure 16.17 Beach environment: littoral V-shaped patterns (v), grooves (g), and blocky conchoidal breakage (b). Quartz grain from layer 2.

foraminaferal tests. The degree of rounding of these carbonate components is also suggestive of high energy wave conditions as found today in front of the cave.

(ii) Textures of Aeolian Origin

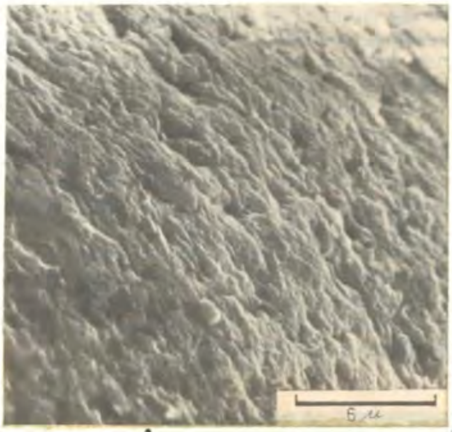
Surface textures of sand grains from layers 3 to 16 have for the most part been diagenetically altered. However, where mechanical textures are preserved they are suggestive of an aeolian origin, a conclusion already arrived at from extensive grading analyses of the sediments.

Krinsley and Donahue (1968) and Krinsley and Margolis (1969, 1971) list critera for recognition of dune sands. These include meandering ridges, graded arcs, flat pitted surfaces, and oriented fracture patterns. Only two of these textures were recognised on quartz grains from Die Kelders I.

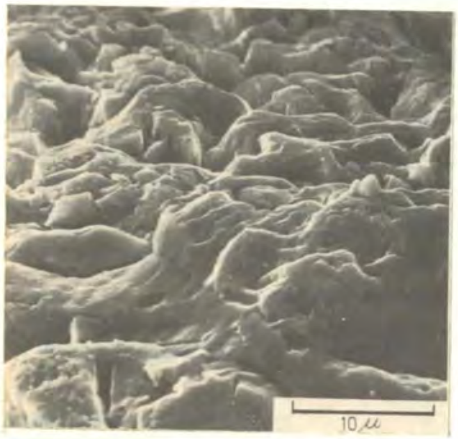
Figure 16.18 A illustrates a typical flat pitted surface. Krinsley and Donahue suggest that flat pitted surfaces gradually replace meandering ridges and graded arcs. But because of larger size and higher relief oriented fracture patterns (Figure 16.18B) would survive longer. No graded arcs or meandering ridges have been observed on Die Kelders samples. Solution has already subdued the original sharp edges of the oriented fractures shown in Figure 16.18 B. Krinsley and Margolis (1969, figure 9; 1971, figure 10) illustrate identical fracture patterns. They found oriented fracture patterns on samples of desert sands from various parts of the world.

(iii) Quartz Sand Diagenetic Textures

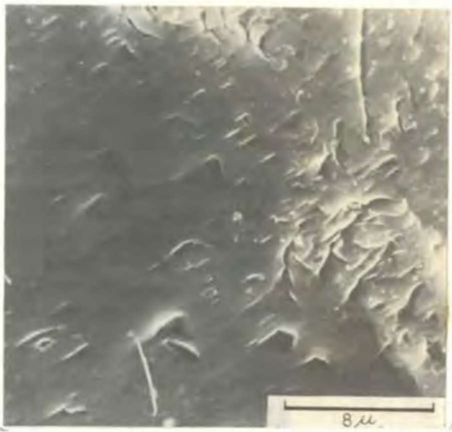
Throughout the entire Die Kelders I stratigraphic column diagenesis of the quartz grain surfaces is common, and has largely destroyed the original mechanical textures. Diagenesis is manifested in solutional features and reprecipitation features. Solution is commonly controlled



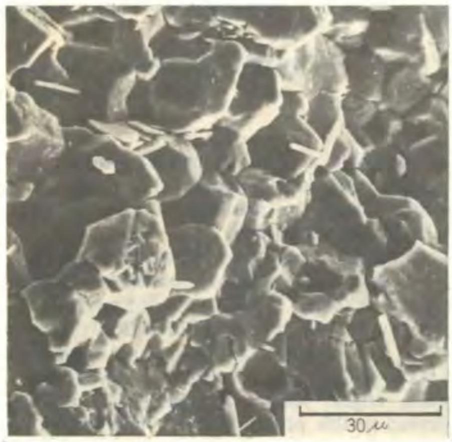
A. Flat pitted surface typical of dune grain. Quartz grain from layer 3.



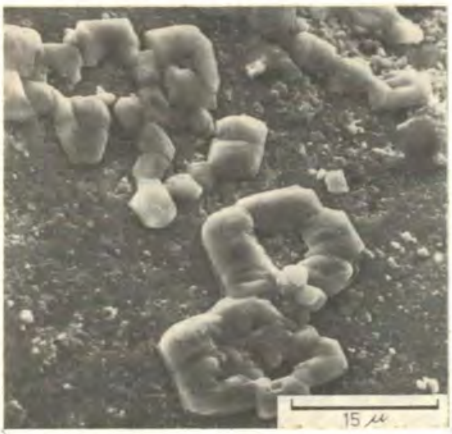
B Oriented fracture patterns as found on dune grains. Quartz grain from layer 3.



C. Diagenesis: smooth solution surface and crystallographically oriented triangles. Quartz grain.



D Diagenesis: incipient crustal growth on quartz fracture surface.



E Diagenesis: calcite overgrowths on shell grain.

F Diagenesis: globules on shell grain.

Figure 16.18 Scanning electron photomicrographs.

by the crystal lattice structure. New crystal development would be in optical continuity with the original grain structure. For a more comprehensive account of diagenetic features see Margolis and Krinsley (1971), Krinsley and Doornkamp (1973) and Tankard and Krinsley (1974).

Slow solution at high pH has produced en echelon arranged V-shaped etch patterns (Figure 16.18 C) which have developed on the trigonal quartz face. Note how much larger these triangles are than the littoral V-notches (Figure 16.17). The broad smooth surface shown in Figure 16.18 C is solutional.

Finally, Figure 16.18 D shows the development of incipient crystal growth on an original blocky disintegration surface. Figure 16.16 D shows crystal overgrowth almost entirely covering a quartz grain, while an irregular precipitation surface is preserved between the crystal faces.

(iv) Detrital Shell Grain Surface Textures

Detrital shell grains from layers 1 and 2 were hand-picked and examined by the SEM to determine if they would also bear mechanical textures. But in all cases only diagenetic textures were observed. The shape and well-rounded nature of the shell grains (Figure 16.16 B) is due to surface abrasion in a surf environment. On closer examination of the apparently featureless surface some very interesting diagenetic textures are revealed. Tankard (1974d) described five types of texture which were observed on Die Kelders shell grains.

The earliest phase of diagenesis must have occurred soon after the grains were blown into the cave from the nearby beach, or perhaps even in the littoral zone itself. The high polish observed with the optical-microscope is due to solution of the surface producing a smooth or very slightly pitted surface. A clear example is the background in Figure 16.18 E. In a highly energetic chemical environment overgrowths are common. Old carbonate from the Bredasdorp limestone of the cave roof and recycled carbonate from other corroded shell have been reprecipitated as overgrowths. Carbonate overgrowths on quartz grains were also observed.

In Figure 16.18 E rhombohedral calcite crystals have grown in chains over an initially smooth surface. The average crystal size is 5 to 6 microns. Precipitation may also give rise to extensive plate-like overgrowths (Tankard 1974d, figure 15).

Perhaps the most perplexing structures encountered are the numerous randomly distributed globular or spherical growths (Figure 16.18 F). Other specimens examined showed that many of these globules are surrounded by a furrow. These globules are possibly the result of very rapid precipitation of carbonate or could have formed due to bacterial activity. Slower precipitation would possibly form plate structures (Figure 16.18 E) and even slower precipitation would produce discrete crystal forms (Figure 16.18 E).

C. Summary

In general, then, it is seen that the grain size statistics for the sand fractions of all, save the marine sands, are very similar. Even sand from layers 1 and 2 which was blown into the cave from the nearby shoreline, conforms to the general pattern. The similar grain size statistics indicate a similar source for the sediments. Modification due to addition of a coarse shell mode or a fine mud-humus mode affects the skewness and kurtosis values most markedly.

While grading analyses show that Die Kelders I sediments are unimodally distributed, the uniform degree of rounding supports a single origin. Any subpopulations with significantly different rounding in any one sample occur to only a minor extent. Furthermore, the subround character of most of the sediment samples from layers 3 to 15 and their light polish are suggestive of an aeolian origin.

After the sea withdrew from the cave, the shoreline remained far distant for a long period. Its return is reflected in layers 1 and 2. Dune bedded aeolianites along this coast show that the dominant sand movement was from the south-east. The cave is north facing and moving dunes above the cave would dump sand down their advancing front to the foot of the cliff where there would be little opportunity for sorting as the sediment would be rapidly buried and reworking would be negligible. It was argued that the sign of skewness did not necessarily negate an aeolian origin. A plot of mean grain size against skewness confirmed the aeolian nature of the sands.

Detailed examination of quartz grain surfaces by scanning electron microscopy shows that textures diagnostic of sedimentary environment have largely been removed by post-depositional diagenesis in a highly active chemical environment (pH range 7,9 to 8,9). But textures characteristic of a high-energy littoral or beach environment were observed on quartz grains from layers 1, 2 and 17 (Figure 16.17). Although these sand grains were blown into the cave from a beach, the total absence of aeolian textures suggests that the distance of transport was not great. Furthermore, littoral textures are more easily impressed than aeolian features. The only environmental textures observed on quartz grains from layers 3 to 16 were suggestive of aeolian conditions. But the high degree of diagenesis has obliterated the finer textures. Detrital shell grains from layers 1 and 2 were found to contain only diagenetic textures.

The results of this SEM study were in broad agreement with those of conventional sediment analysis, and in many respects the technique is successful, particularly with respect to high energy marine environments. But the writer is not convinced that SEM studies alone are sufficient to detail aeolian environments since the diagnostic textures could also be the result of diagenesis.

V. RECONSTRUCTION OF THE PALAEOENVIRONMENT

The Die Kelders cave complex and other caves along the southern Cape coastline at similar altitudes were probably all formed by erosion by a sea 6 to 8 m higher than present at about 120 ka., although high sea levels of a previous interglacial may also have contributed.

At Die Kelders formation of the cave complex was facilitated by erosion along an unconformity. The interaction of this high sea with ground water moving along the unconformity was probably a very important factor.

A stillstand of the sea close to present datum, probably late in the last interglacial, is recorded by wave-generated boulders at the base of Die Kelders I succession.

In Figure 16.19 the history of Die Kelders I from the time of formation of the basal marine horizon to the final Late Stone Age occupation is illustrated, and Figure 16.20 correlates the environmental changes with the cave stratigraphy.

Regression of the last high interglacial sea level took place after 80 ka. Occupation of the cave was not immediate as is shown by the pre-occupation sediments. It is difficult to date the first occupation. Tankard and Schweitzer (1974) have suggested the possibility that angular limestone fragments in the first occupation layer might equate with the world-wide temperature depression which accompanied the early Würm glacial maximum at 75 ka-65ka (Shackleton and Opdyke 1973). Freezing temperatures are rare under the present climate, and the presence of éboulis secs indicates a significant lowering of temperatures during the initial MSA occupation. But frost-spalling also indicates the presence of moisture. The iron staining of the quartz sands confirms this. Although seal and penguin bones occur throughout the MSA sequence, they are most abundant in the earliest occupation layer but show a significant decrease in quantity upward (Dr R.G. Klein pers. comm.). This would suggest a nearby shoreline for the earliest MSA occupation and possibly a withdrawal of the sea soon afterwards. The éboulis secs may thus represent the first major drop in temperature which preceded the fall of sea level. Perhaps an explanation for the

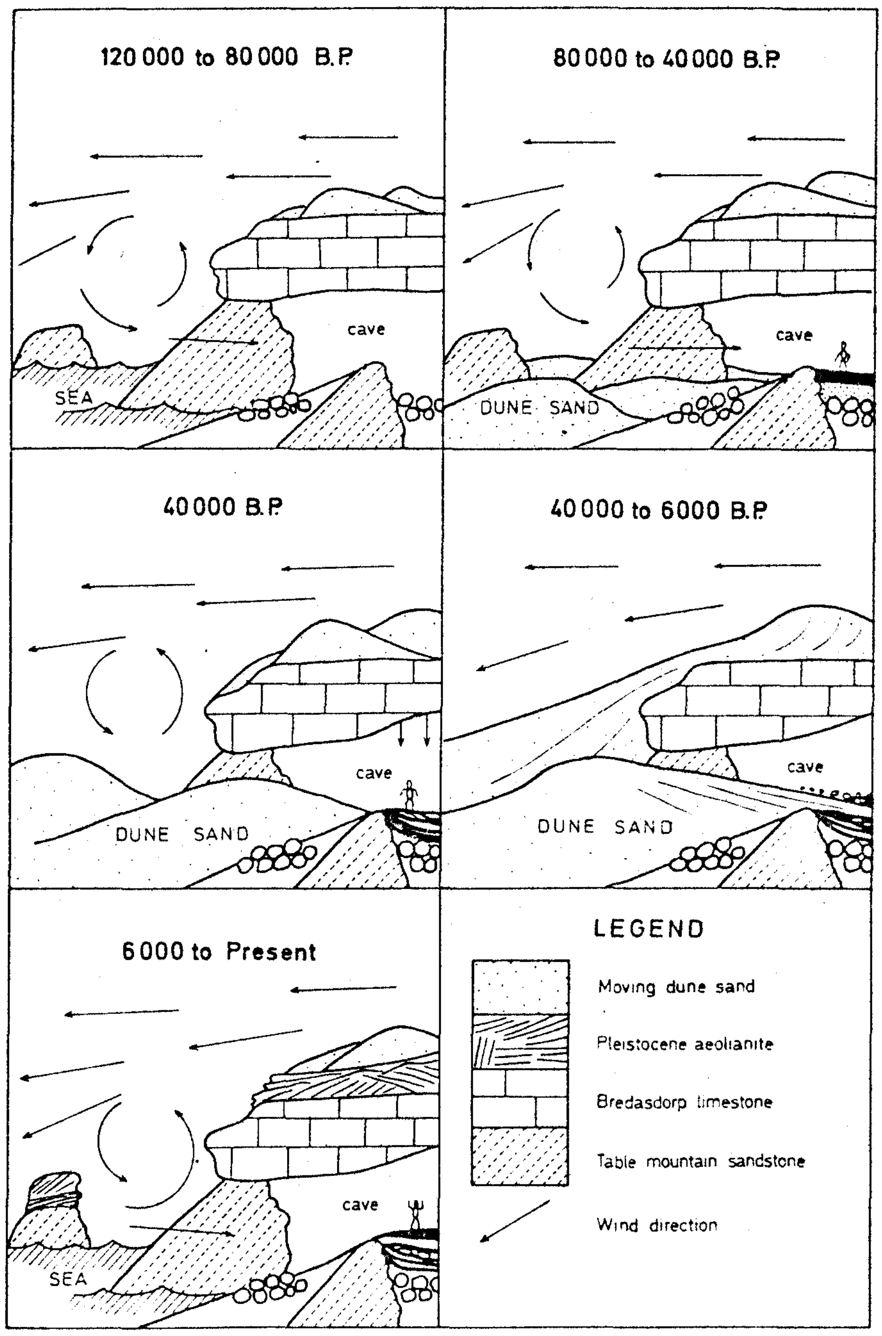


Figure 16.19 Cartoon summarising the evolution of Die Kelders I cave sediments from time of occupation of the cave by sea to the Later Stone Age midden deposit. Note how the southerly wind moves dune sand over the brink of the cliff, which results in burial of the cave from about 40 - 50 ka to about 6 ka. The dune field was finally removed by the Flandrian transgression, leaving only outliers of aeolianite on promontories and on stacks, besides that on high terrain.

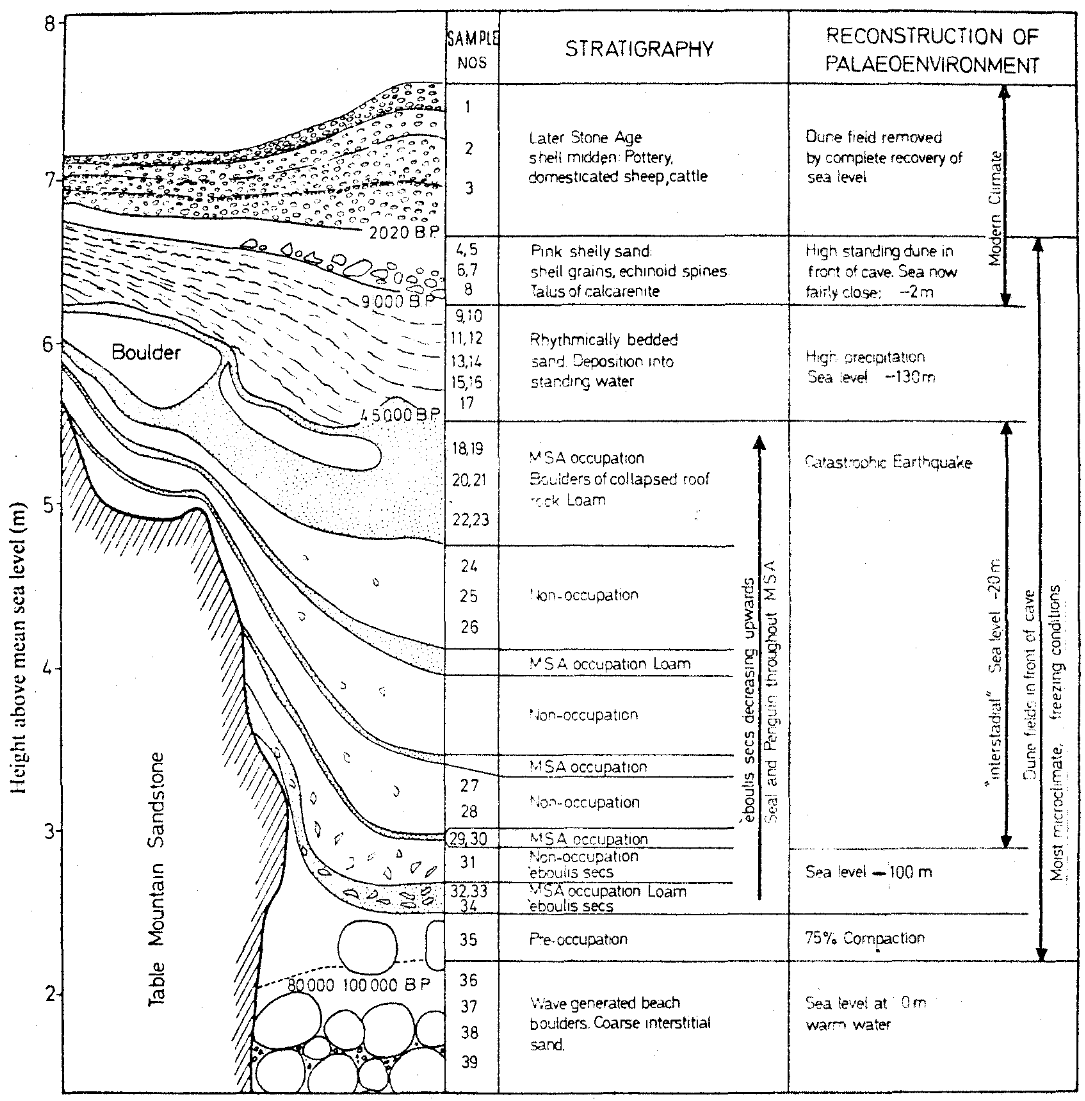


Figure 16.20 Schematic representation of the cave stratigraphy and the palaeoenvironmental reconstruction.

upward reduction of *éboulis sec* in strata which possibly dated to a glacial maximum and shoreline retreat, is that the initial freeze-thaw cycle removed the outer zone of limestone roof rock which owed its fissure development to the preceding interglacial. So initial frost-spalling and the seal and penguin bone distribution would suggest an early Würm age for the first MSA occupation, possibly about 75 ka.

Throughout the MSA occupation the cave environment was damp while textural analyses show the existence of a persistent dune field in front of the cave. Again it is difficult to estimate the cessation of MSA occupation. Klein (1974) has suggested that the MSA lasted until about 40 ka. It would indeed be most fortuitous if Die Kelders I occupation had lasted until the limit of MSA time. Instead it would be more probable that occupation had ceased sometime prior to this, say between 45 and 50 ka. And if the first occupation dated to, say 75 ka, a period of MSA occupation of about 25 000 years would be indicated.

An occupation hiatus follows the MSA occupation. The beginning of the hiatus is marked by solution of the tops of the roof rock boulders and complete leaching and elutriation of the youngest MSA unit. This was followed by a brief period (rapid sedimentation) when sedimentation took place into standing water. A precipitation maximum is indicated, and Tankard and Schweitzer (1974) have offered this as a partial explanation for the occupation hiatus. Butzer and Helgren (1972) record red podzolic soils, indicating a wetter climate, following the MSA of the Plettenberg Bay area. Nearer Die Kelders, at Rietvlei, radiocarbon dated pollen stratigraphy shows two wet intervals: 45-40,5 ka and 36,5 - 33 ka (Schalke 1973). It would seem reasonable to equate the rhythmically bedded sterile sands with these wet intervals.

The top of the sterile sands is marked by a micro-soil profile (Figure 16.6) which indicates a moist and stable environment with no erosion or deposition for a considerable length of time. Tankard and Schweitzer (in press) have suggested that with continual dumping of sand over the lip of the cave, which is north facing, complete or nearly complete burial isolated the cave. This would undoubtedly explain the period of stable microclimate in the cave associated with a period of non-deposition and non-erosion which led to soil formation on a small scale. This would also presuppose a drop in sea level, and shoreline retreat, so that a more extensive dune field could develop. The -25m shoreline which lasted until 25 000 B.P. (Chapter 1) would have been too close. A fall in sea level to -130m occurred at about 17 ka, while between 21 ka and 12 ka, sea level would have been low enough on the southern Cape coast for development of the dune field envisaged. It is thus suggested that burial took place during the upper Würm temperature minimum. Even though the cave moisture may still have been high, the stability of the closed cave system would account for the absence of cryoclastic debris at the top of the sterile sands.

Grading analyses and the use of scatter diagrams to present these results support this reconstruction. It was shown that the sediments from layers 1 to 16 are fine grained, but bordering on medium sand size. They are largely moderately well sorted and negatively skewed. If the sign of skewness was diagnostic the sediments should be interpreted as inherited from a beach environment, dune sands being generally positively skewed. Neither was kurtosis, by itself, found to be significant. But plotting all the grain size statistics as 6 two-variant scatter diagrams proved more illuminating. Essentially, the statistics separate into three broad groups in all the scatter diagrams. One group, the interstitial marine sands (layer 17) do not interest us immediately. The other two groups are overlapping and consist of samples from occupation and non-occupation layers. In the non-occupation samples the grain size statistics are tightly clustered in most of the scatter plots, suggesting a common origin. It was shown that the differences encountered in the occupation samples were due entirely to admixture of a fine mode, mud (silt and clay) and humus. With progressive leaching and elutriation of the humus and mud respectively, the occupation layer grain size statistics approach those of the non-occupation statistics more closely. Where the humus-mud mode has fallen to less than 1 per cent the statistics are the same.

Comparing mean grain size with skewness, Friedman (1961) found an almost complete separation between beach and dune sands. In plotting these two parameters (Figure 16.11) with Friedman's dividing line between dune and beach environments, Figure 16.11 shows that all but two of the sediments above the marine horizon fall on the dune side of Friedman's boundary, suggesting that all the sediments above layer 17 are of aeolian origin. Although these sediments had an aeolian origin the iron-stained sand of layer 3 was clearly deposited in ponded water. But sedimentation was rhythmical and fairly rapid. Narrow hematite-stained laminae with microfaunal remains suggests that the pond periodically dried out. Layer 2 was in part deposited in standing water. The lower part of this layer is a talus deposit comprising shelly sands and aeolianite rubble that gravitated down the flank of a high dune that stood at the cave entrance.

There is also other evidence to suggest a seaward extension of the dune belt during this period of lower sea level. Overlying promontories of quartzite along the coast are aeolianites. All along Walker Bay these aeolianites extend to below present sea level and are presently being cliffed by the sea. More specifically, above the cave the aeolianites extend to the edge of the cliff,and are found to contain Middle Stone Age artefacts very similar to those in Die Kelders I. A panoramic view of the aeolianite shows not only the original extent of the dune field, but also strongly suggests that the cave complex may have been completely buried at some stage. Microscopic examination shows that the quartz grains of Die Kelders I average subround and are very lightly polished. The grains from the aeolianite are also subround and have the same degree of polish. In contrast sand grains from the beach in front of the cave are all well rounded and polished.

Non-occupation of the cave from about 50 ka to 2 ka can be attributed to two basic causes. Firstly, we believe that for much of this period dampness of the cave probably discouraged occupation. The layer 3 sediments even demonstrate the presence of standing water. Secondly, regression of the sea during the Würm left behind vast quantities of sand exposed on the continental shelf. With increased wind activity, associated with glacial conditions in the high latitudes, a dune field built up above and in front of the cave. Not only would an active dune field discourage occupation, but dumping of sand over the lip of the cave would have built up a high dune there, and probably even have completely buried the cave at some stage. Sand falling over the brink of a cliff would enter a sand shadow where further transportation and sorting would be ineffective.

Reopening of the cave is shown by deposition of talus rubble during a dry period, in fact the first dry period encountered in the cave stratigraphy. The sands contain rounded shell grains, foraminiferal tests, and complete echinoid spines, indicating a nearby shoreline. Recovery of sea level would be required to remove most of the dune field, and expose the cave to renewed deposition. But sea level would still have been low enough to leave some dunes in front of the cave down which talus material, from the aeolianite overlying the cave roof rock, would migrate into the cave. The interstadial sea level at about -25 m had been sufficient to keep the cave open before. Recovery of the sea level following the 17 000 B.P. lowering was very rapid (167 cm/100 years, Chapter 1), and by 9 000 B.P. had reached -25 m, and by 6 000 B.P. was within a meter of present sea level. Reopening of the cave would most likely have occurred at 9 000-8 000 B.P. Occupation by LSA people only occurred at about 2 000 B.P. (Tankard & Schweitzer 1974).

To summarise (Figure 16.19): The entire MSA succession is dominated by moist conditions in the cave, while èboulis secs at the base of the succession indicate an early Würm temperature minimum. Standing water in the cave occurred probably during the wet intervals from 45 ka to 33 ka documented by Schalke (1973). Stable conditions in the cave due to burial beneath a dune field prevented frost-spalling during the upper Würm temperature minimum. Butzer (1973) estimated that average winter temperatures 10^{o}C lower than at present along the southern Cape coast were necessary to produce frost-shattering. The age estimates used in the previous discussion are reasoned guesses to establish a relative chronology.

VI. PALAEOCLIMATIC IMPLICATIONS

A. Hypothermal Climate

The position of the anticyclone varies with solar radiation, and Tankard (1975) has shown that last interglacial thermophilic molluscs of the southwestern Cape can be accounted for by a shift of the South Atlantic anticyclone during the climatic optimum at 120 ka. With cold climates that prevailed during the last hypothermal (glacial) period one would expect the opposite trend to take place. Lamb (1961) has suggested that ice age circulation was marked by intensified circulation of the belt of the westerlies, and greater mobility of the subtropical anticyclones. The climatic zones move towards the equator in hypothermal periods (Van Zinderen Bakker 1967 and in press). A consequence of the northward displacement of the climatic zones would be a northward movement of the cyclonic belt.

A northward movement of the belt of westerlies would have expanded the winter rainfall area of the southern Cape, and would have caused a migration of the Cape flora inland and along the eastern escarpment (van Zinderen Bakker in press). Relicts of this flora exist today in the Orange Free State. The arid belt which characterises the Karroo, and areas to the north, would have migrated northwards with the anti-cyclones which cause them.

The present seasonal variation of the south Atlantic anticyclone is between 26°S and 30°S (Hart & Currie 1960). Hypothermal climatic maps compiled by van Zinderen Bakker (1967, figures 4 and 5) show a variation between approximately 22°S and 25°S, a northward shift of the order of 4–5°. Since the present winter anticyclone centre is centred at 26°S, a summer position in the hypothermal at 25°S would ensure year round precipitation in the present winter rainfall area, as well as an expansion of the winter rainfall belt. But the area would still have received a winter rainfall maximum.

VII. EFFECT OF CLIMATIC MODIFICATIONS ON HUMAN OCCUPATIONS

A long period of non-occupation of Die Kelders I followed the MSA occupation, and lasted some 45 000 - 50 000 years. Tankard (in press d) and Tankard and Schweitzer (1974 and in press) have discussed in detail this break in occupation, which was attributed to standing water and burial of the cave beneath a dune field. But standing water and burial single Die Kelders out as a special case. Obviously the same conditions would not adequately account for occupation breaks following the MSA of other caves. Klein (1974) records that every southern Cape site with sufficient data features a post-MSA break in occupation, which generally ranges between 20 000 and 50 000 years. These sites include: Klasies River Mouth, Die Kelders I, Montagu Cave, Nelson Bay cave and Kangkara Cave. A failure of all these sites to provide a continuous sequence can hardly be fortuitous.

The following hypothesis may provide an explanation for the occupation breaks, and if accepted should indicate where one may look for occupations spanning the hiatus. Obviously the hiatus in occupation suggests that the entire coastal plain was an unsatisfactory habitat. In terms of structure and geography not every cave would contain standing water or be buried beneath dune fields. But northward movement of the belt of westerlies would ensure year round precipitation in the southern Cape. This alone would ensure that every cave site would be damper than today. The caves would also have been colder since this was also the time of lower world-wide temperatures, the upper Würm minimum. Colder conditions would result in lower evaporation rates. This and the higher rainfall is reflected in the relict pluvial lakes and the saturated regoliths which gave rise to colluvial deposits interbedded with aeolianites of the southwestern Cape. Dr Klein (pers. comm.) points out that the fauna from the coastal MSA sites also suggests a greatly increased precipitation, if not year round precipitation. Solecki and Leroi-Gourhan (1961) have also had to attribute a 14 000 year occupation hiatus in the Shanidar Cave of northern Iraq, a region with a Mediterranean-type climate, to cooler and wetter conditions.

At the same time as the Cape south coast, with its possible year round precipitation, became an unsatisfactory habitat, so northward migration of the winter rainfall belt would have brought increased rainfall to the Little Karroo and perhaps even the Great Karroo. But in these parts rainfall would have been seasonal. Expansion of the Cape flora would have provided a more luxurious vegetation which could have supported a denser animal population. Plentiful game and a more amenable climate would have enticed prehistoric man away from the coastal belt.

These climatic conditions and the consequent reduction in the human population along the coast suggest to the writer that any excavations of southern Cape coastal caves will reveal a significant period of non-occupation after the MSA. Sites that have been continuously occupied may therefore be sought on the fringes of the year round winter rainfall belt where rainfall would be seasonal. Such areas would be found inland of the coastal mountain ranges, or along the east coast, or perhaps further up the west coast. The validity of this argument is suggested by recent excavations at Boomplaas Cave in the Oudtshoorn district where occupation has been more or less continuous throughout the period of the coastal hiatus in occupation (Dr Klein pers. comm.).

CHAPTER 17

NAMAQUALAND COASTAL DEPOSITS

I. INTRODUCTION

A detailed account has already been given of the Tertiary sediments of the Langebaanweg-Saldanha and Cape Town areas. The evidence of phosphatic sandstone, pelletal phosphorite and fossils suggest that contemporaneous sediments occur along the Namaqualand coast. However, there have been relatively few published descriptions of the Namaqualand Tertiary geology. Since the discovery of diamonds in 1908-1909 Krige (1927) has discussed the raised beaches very superficially, Wagner and Merensky (1928) and Haughton (1931) described the general geology while Reuning (1931) hypothesised on the origin of the diamonds. The only recent discussion of the sediments, albeit superficial, is that of Hallam (1964).

An extensive diamond exploration project was initiated in the early 1960s by De Beers Consolidated Mines Limited (Namaqualand Venture). Between the Groen and Bitter Rivers prospecting was done by means of large diameter drills along lines perpendicular to the coast and spaced 2,44 km apart. Further north, on the farms from Zwart Lintjies Rivier to Noup (Figure 17.1), prospecting was done by means of large diameter drills, trenching and shafting. In all cases bedrock was intersected.

This chapter will discuss the geology of the Namaqualand coastal deposits in the areas Groen to Bitter Rivers, and Zwart Lintjies Rivier to Noup (Figure 17.1). The discussion will be superficial since the field work was carried out in the three months December 1965 to February 1966. Use will be made of personal records and an unpublished B.Sc. project report (Tankard 1966). Although geographically outside the scope of this thesis, it is included because it offers a very useful comparison with the Cenozoic sediments between Saldanha and Cape Town, and because it makes available a little more information on the all but unknown west coast Cenozoic geology. Mr

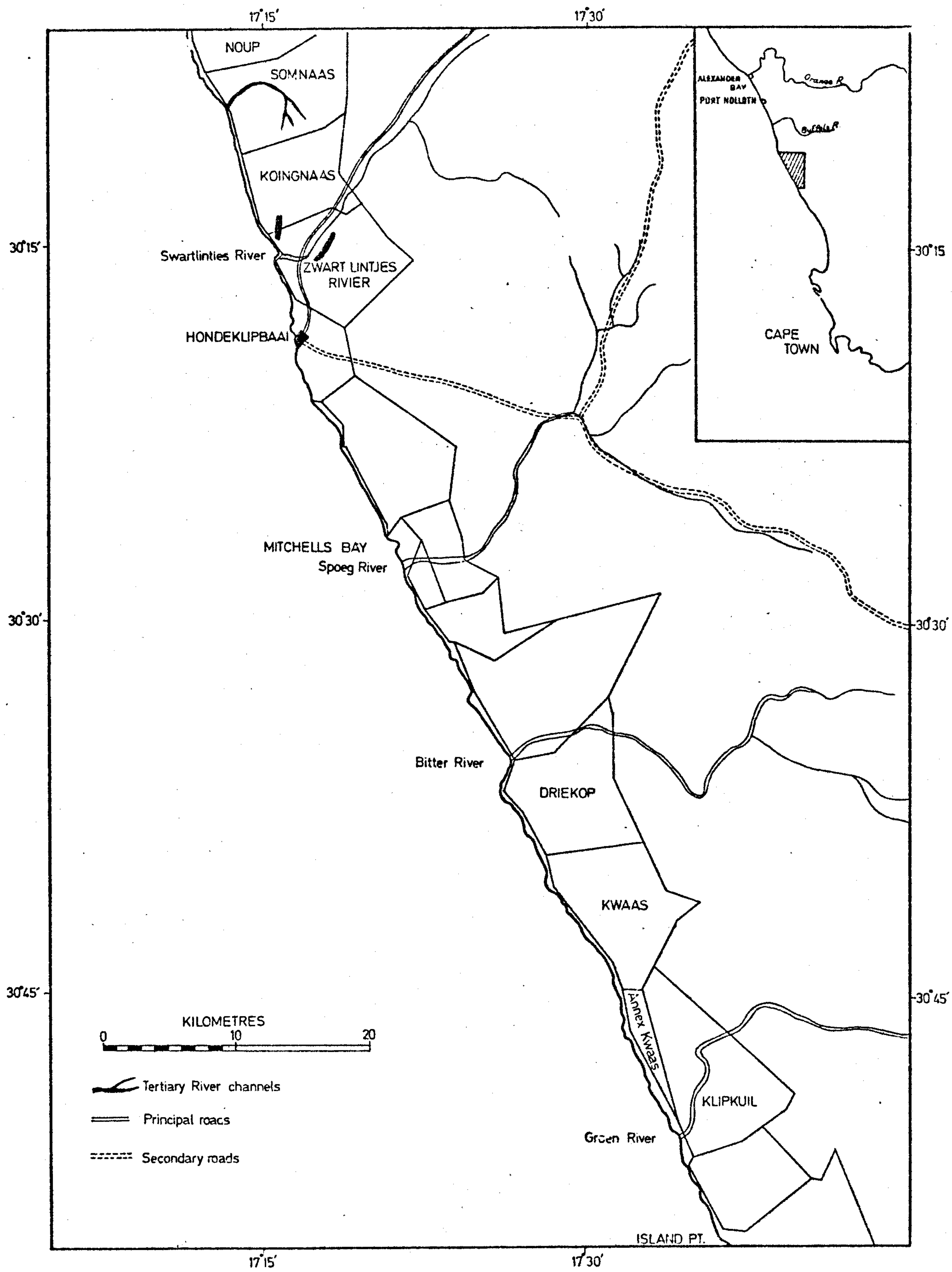


Figure 17.1 Locality map for Namaqualand prospects.

A.J. Carrington is currently examining the Namaqualand Cenozoic geology in detail.

II. PRE-MIOCENE GEOLOGY AND GEOMORPHOLOGY

The coastal sediments are underlain predominantly by Namaqualand granite-gneiss. The granite-gneiss is remarkably homogeneous, but becomes more leucocratic towards the Kheis System in the north. Ages of pegmatities in the granite-gneiss range from 1090 to 900 million years (Truswell 1970). The rocks are predominantly granitic biotite gneiss, often garnetiferous. The overlying Cenozoic sediments are in places rich in garnet and felspar derived from the granite-gneiss.

Intrusive dykes follow the structural trends of late Precambrian origin (NNW)(Kröner 1973). On the farm Dikdoorn on the Groen River an intrusive melilite basalt was dated at 38,5 m.y. (ZSA 56). No Tertiary sediments in the area are intersected by the intrusives. This extrusive and the Klinghardt volcanics clearly suggest instability of the Namaqualand coast at least in the Early Tertiary. Furthermore, the lines of volcanic activity are related to older structures in the crust (Kröner 1973).

Granite-gneiss is exposed along most of the coastline. Where sandy beaches are encountered they usually coincide with old river channels identified during prospecting (Figure 17.3). The lower reaches of the Groen, Bitter and Swartlintjies rivers, all ephemeral streams, are incised 45 to 90 m below the surrounding countryside, and the present channels run in flat, alluvial flood plains.

III. BEDROCK MORPHOLOGY

In the Groen River area extensive drilling has shown that the bedrock contours are fairly regularly spaced and are parallel to the present coastline. But they diverge northwards to form a broad embayment on the farm Kwaas (Figure 17.1). On the northern part of Kwaas and Driekop marine abrasion has developed a series of distinct platforms. The most prominent platforms, at 43 m, 20 to 21 m, 7,6 m and 4,6 m, probably represent major stillstands of sea level, although some of their present altitude may be attributed to tilting. The 4,6m and 7,6 m platforms are particularly well developed. The slope of the 43 m platform is 0,75 to 1^{o} towards the present coast. Generally, the platforms are broad and have undulating surfaces. Troughs and depressions in the platforms contain gravels.

Little is known about the bedrock morphology between the Bitter and Swartlintjies rivers. North of the Swartlintjies River, on the farm Koingnaas, there is a large embayment (Figure 17.2) that has had a profound effect on the formation of diamondiferous ore bodies (Tankard 1966).

IV. LATE TERTIARY AND PLEISTOCENE GEOLOGY

In the Koingnaas area the various Cenozoic stratigraphic units are separated by distinct unconformities (Tankard 1966; Carrington & Kensley 1969). Tertiary and Pleistocene sediments overlie the wave-cut platforms bordering the shoreline. They consist of conglomerates, gravels and sands which are richly fossiliferous. Generally the sediments occur up to 4 km from the coastline.

Haughton (1931) recognised two distinct mollusc faunas from the west coast. The older of these was a warm-water fauna characterised by *Chamelea krigei*, *Striostrea margaritacea* (Haughton's *Ostrea prismatica*) and *Donax haughtoni* (Haughton's *D. rogersi*). The younger marine sediments are characterised by a cold water fauna.

The stratigraphic units, each of which is separated by an unconformity, will be described in order of diminishing age.

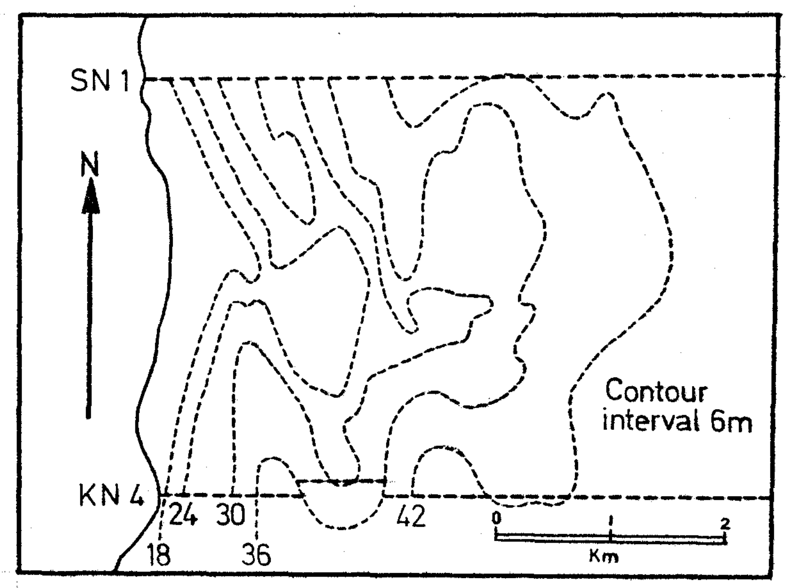


Figure 17.2 Bedrock configuration on Koingnaas
(after Tankard 1966).

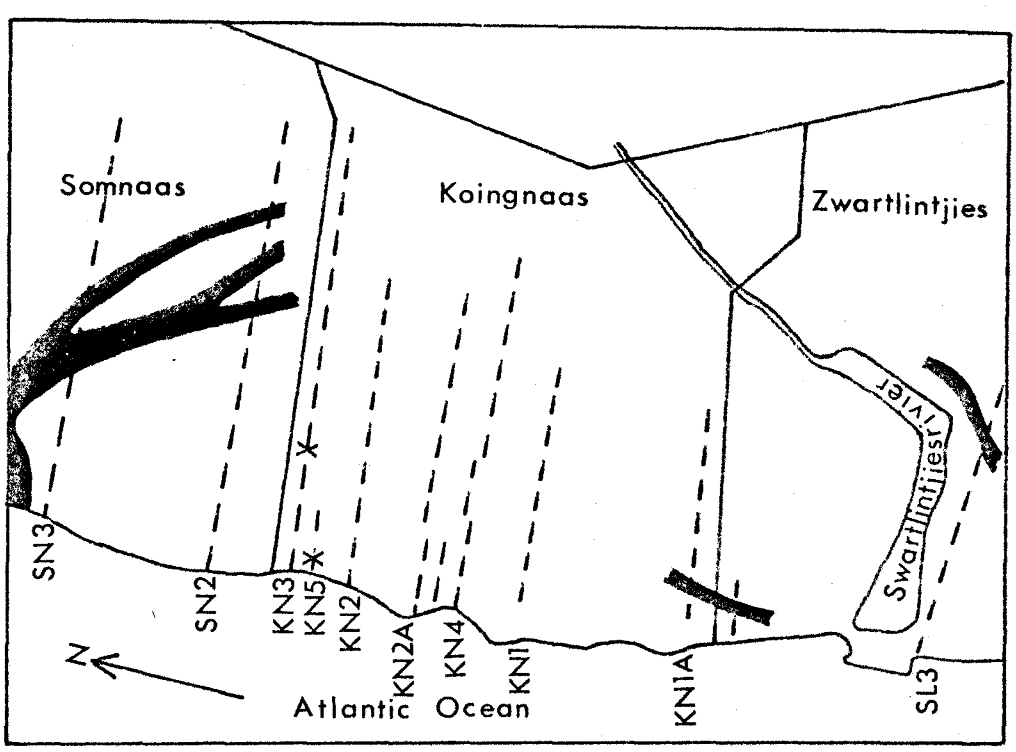


Figure 17.3 Location of prospecting lines (after Tankard 1966).

A. Tertiary

Preserved today as bedrock-hollow infills are a series of sediment types which are distinguished from younger sediments by high phosphate content. The pockets of "E stage" (Mine terminology) sediments represent the remains of former widespread Miocene and Pliocene strata that have survived Pleistocene transgressions.

1. "Lower E Stage"

"Lower E stage" strata consist of calcarenite, coquina, phosphatic siltstone, and a ferruginous sandstone. Thin-section study suggests that the phosphatic siltstone originated by phosphatisation of a micrite. The quartz grains are well rounded, while in the matrix there is evidence of finely bladed overgrowths developed syntaxially about intraclasts. Also present are finely crystalline microspar and medium to fine crystalline pseudospar which has replaced micrite (for explanations to the terminology see Folk 1965). There are occasional phosphate haloes about the intraclasts. The P_2O_5 content is low (1,3 per cent). Shell debris with crystalline overgrowths is common.

Maximum thickness is 1 m (Carrington & Kensley 1969) as is the Miocene basal bed at Langebaanweg. Since authigenic phosphate forms close to the water-sediment interface, a thin horizon is expected. Furthermore, the surface of much of the phosphatic siltstone is water-worn and polished, as at Langebaanweg, and a conglomeratic fabric is common. On foraminiferal evidence Carrington (in litt) suggests a Late Miocene age for the "Lower E stage", and argues that it is the lithological and chronological equivalent of the Saldanha Formation. Furthermore, the calcarenites contain a conservative endemic mollusc fauna similar to that described from the Miocene Ysterplaat deposits by Tankard (in press c) and that of the east coast Alexandria Formation described by Bullen Newton (1913). Typical molluscs are *Donax serra* (same form as at Ysterplaat), *Glycymeris borgesi*, *Lutraria lutraria*, *Scissodesma spengleri*, *Pitar* sp.

2. <u>"Middle E stage"</u>

The "Middle E stage" is marked by a phosphatic siltstone and is equivalent to the phosphatic sandstone/siltstone bed that marks the top of the Quartzose Sand Member of the Varswater Formation. This "stage is known to contain the vertebrate fossils <u>Prionodelphis capensis</u> and <u>Ceratotherium praecox</u>. Hooijer (1972) mentions an isolated rolled tooth of <u>C. praecox</u> of Late Pliocene age found in poorly sorted fluvial gravels of Pleistocene age at about 18m a.s.l. He assumed that the tooth would have been derived from sediments older than the fluvial gravels. Carrington (pers. comm.) believes that it was derived from "Middle E stage" sediments. Recently the fresh and unrolled tooth of a seal, <u>Prionodelphis</u> capensis, was found in fine quartzose sands in section 290, line KN3, on Kiongnaas. Carrington points out that the tooth is phosphatised, but that the fine quartzose sands do not contain even a vestige of phosphate. He suggests that the tooth was probably eroded from "Middle E stage" sediments by very low energy conditions.

Both <u>Ceratotherium praecox</u> and <u>Prionodelphis capensis</u> were taken to characterise the Varswater Formation (Hendey 1974). The genus <u>Prionodelphis</u> was previously known only from the Pliocene of Argentina. These fossils and the thin phosphatic siltstone suggest correlation of the "Middle E stage" with the Quartzose Sand Member of the Varswater Formation.

3. <u>"Upper E stage"</u>

On line KN 3 (Figure 17.3) on the northern part of Koingnaas between sections 276 and 291 the weathered gneiss surface dips gently to the east (inland). Overlying the weathered gneiss is 0,6 to 0,9 m of rounded quartz pebbles and gneiss pebbly sand, all of which is iron-stained. These sediments are essentially bedrock-depression infilling. Although all of the "E stage" is phosphatic to varying degrees, the "Upper E stage" is characterised by pelletal phosphorite

(Carrington pers. comm.). The pelletal phosphorite is somewhat coarser than the pelletal phosphorite of the Varswater Formation (0,5 to 1,5 Ø versus 1,8 to 2,3 Ø).

A very rich fossil fauna exists in the "Upper E stage" sediments. Abundant rounded or partially rounded bone fragments have been recorded. Phosphatic nodules and internal shell casts are common, as are shark and other fish teeth which show a high degree of polish. The mollusc fauna is west African in character and suggests water temperatures 5 to 8°C warmer than today (Carrington pers. comm.). The contemporary continental climate was arid. The terminal altitude of the "E stage" is not known with certainty. It has been recorded only up to 30 m a.s.l.

4. Fluvial Beds

Distribution of (?) Tertiary river channels is shown in Figure 17.3. On Somnaas the channels are confluent. In January 1966 a large diameter drill hole was sunk into the centre of river channel sediments that had been discovered by earlier prospecting. Once the channel sediments were encountered at a depth of about 25 m, the hole was drilled in 1,5 to 1,8m segments and each segment logged *in situ* by lowering a geologist down the drill hole. An annotated section is shown in Figure 17.4 (from Tankard 1966).

The fluvial bed on Koingnaas (KN1A Figure 17.4) is characterised by medium to fine quartzose sands with appreciable clay, broken by horizons which have a more clayey nature. At a depth of 28 m an horizon of blocks of brown, medium sandstone was encountered. The blocks gave the impression of a once continuous sandstone bed. From about 30 m massive, grey "channel clays" were encountered. Towards the base of this test hole proximity to bedrock was first shown by very weathered felspar. At this point heavy minerals were not encountered. With increasing depth weathered biotite and finally fresh biotite were found.

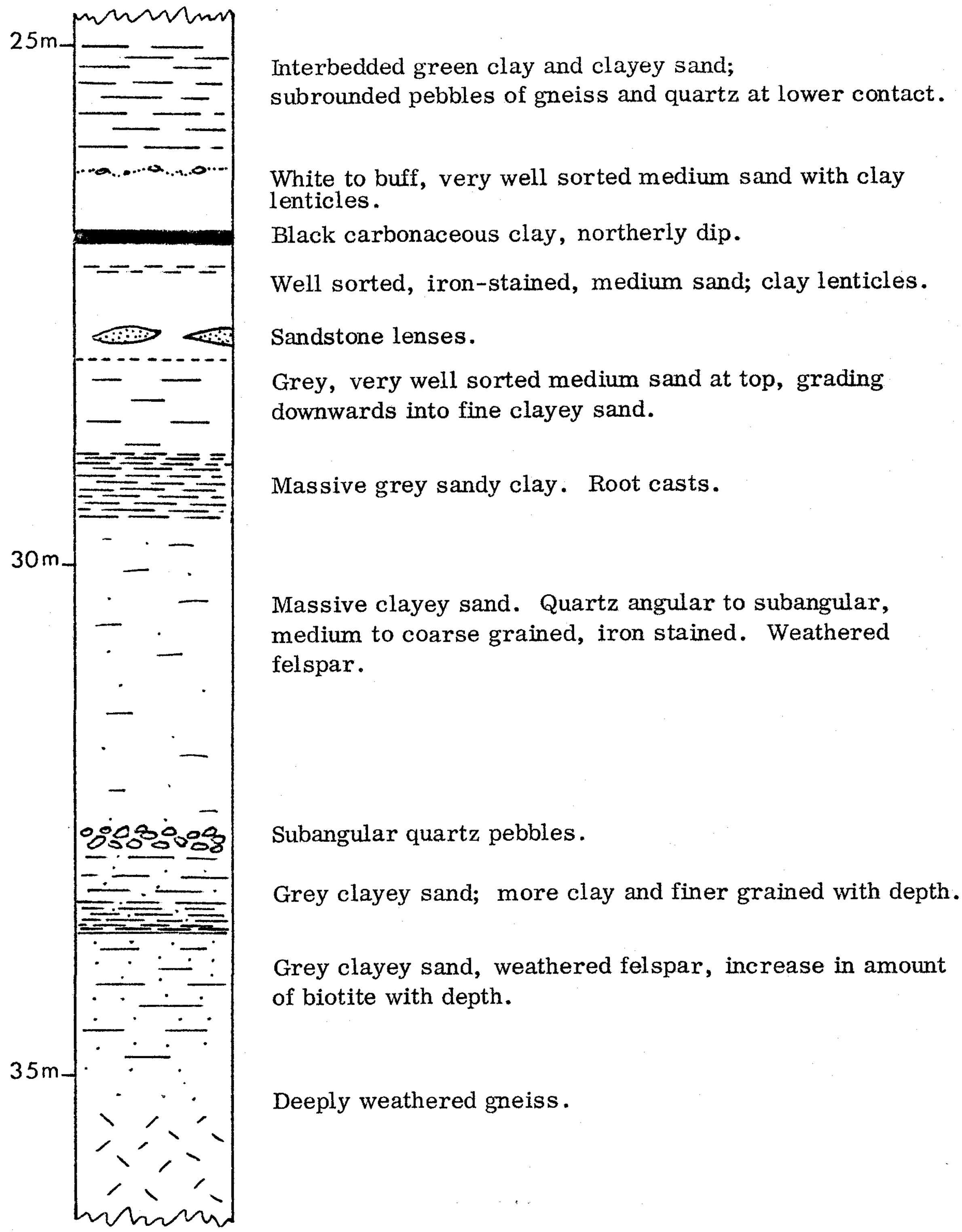


Figure 17.4 River channel sediments. Drill hole 158/9, KN1A.

On Driekop the fluvial sediments can be subdivided into three horizons:

(i) the basal unit is a carbonaceous sand composed of grey-black and brown fine to coarse sandy clay. It contains pebble and cobble size quartz. The colouration is due to the high carbon content. Well preserved wood and leaf impressions have been found in this horizon. When freshly excavated it has the smell of marsh sediments.

(ii) Overlying the carbonaceous sands is a complex horizon of interbedded clay, sandy clay and sand. The clay beds overlie the carbonaceous sand and are light grey to purple in colour.

(iii) The topmost unit is a coarse, angular clayey quartz sand. It is generally a grey colour but in places is red and contains ferruginous concretions.

The exact relationship of the fluvial beds to the "E stage" sediments is not clear since the latter are preserved as relict bedrock-pocket infills. They are possibly equivalent to the peat-like sediments within the Varswater Formation, but could also be much younger. The maximum thickness accorded the fluvial sediments is 20 m (Carrington & Kensley 1969).

B. Pleistocene

Carrington and Kensley (1969) have mentioned a series of transgressive complexes of Pleistocene age reaching to nearly 100 m above sea level. The highest beach deposit is situated at 98,3m a.s.l. in the north and sinks to 93 m southwards over a distance of 80 km. The transgressive complexes alluded to by Carrington and Kensley are as follows: 75-90m, 45-50m, 29-34m (thin, discontinuous), 17-21m, 7-8m, 5m, 2m. The 75 to 90m transgressive complex sediments are not known to the writer but they are said to be unfossiliferous, coarse, porous, marine sands.

1. 45 to 50 m Transgressive Complex

At the base of the 45 to 50 m transgressive complex beds on Koingnaas, or perhaps forming the top of the "E stage", is an horizon of water-worn cobbles and boulders which are partly cemented with siliceous-ferruginous cement. It has a highly polished upper surface. Overlying this indurated horizon is a gravel bed, 0,3m thick, which grades upwards into a fine to very fine pale green sand.

Similarly on Driekop this horizon commences with a gravel and grades upwards into a fine green sand. The colour of the sand when wet ranges from light yellow to olive green. The sand is mainly fine to very fine and is generally well sorted. There is a tendency for the sand to become coarser and more angular eastward, i.e. with shallowing of water. Interbedded clay bands frequently have heavy minerals associated with them. Below the clay the sands are finer grained than those above, a typical graded bedding sequence. Fluctuating conditions are suggested.

The fossil fauna from the 45 to 50 m transgressive complex on Koingnaas is very rich. Besides many "E stage" fossils it also contains coral remains. The sediments are characterised by Striostrea margaritacea, Chamelea krigei, Donax haughtoni, Petricola prava, Fissurella robusta and Triumphis dilemma. The Donax haughtoni has a more fragile shell than Donax rogersi and suggests that it would not tolerate turbulent conditions. This is supported by the fine nature of the sediment, while the more robust D. rogersi inhabited the coarser overlying sediments (Carrington & Kensley 1969). The 45 to 50 m transgressive complex, which is equivalent to Haughton's (1931) D-zone, contains a mollusc fauna typically of warm water affinity.

A typical section, situated on line KN 3 (Figure 17.3) is shown in Figure 17.5A.

2. 17 to 21 m Transgressive Complex

The 17 to 21 m transgressive complex is characterised by coarse iron-stained sands. This complex is equivalent to Haughton's B-Zone. That it is transgressive is demonstrated by the occurrence near the base of large rib and limb bones, small silicified limb bones, rodent teeth and land snail shells (Tankard 1966).

A calcareous cemented pebble horizon marks the unconformity between the 17-21 m and 45-50 m transgressive complexes. The pebbles are composed of gneiss and reworked conglomeratic material from the base of the 45-50 m transgressive complex sediments. The fine green sand of the latter is interbedded with lenses of coarse sand at the base of the 17 to 21 m transgressive complex. This is followed by very coarse, red-brown sand. Broadly speaking, this horizon is graded; coarse to very coarse at the base and becoming progressively finer grained upwards. The coarse sand is moderately well sorted.

Between the Groen and Bitter Rivers this complex is the most prominent horizon. Like the Koingnaas occurrence it displays a fining-upward sequence. The lower parts are a light coloured sand while the upper layers are more compact and orange-brown to green clayey sand. The basal parts are generally lime cemented. The finer grades are more angular. The top of this complex on Driekop has a high mud content ($>$10 per cent). It is capped with calcrete.

The fossils, which include limb and rib bones, rodent teeth and land snail shells, point to a terrestrial origin. But the poor sorting of the sand was found to be due to mixing of alternating coarse and fine cross-bedded layers which suggest fluctuating water currents. The sediments also contain comminuted shell. The zone-fossil would undoubtedly be *Donax rogersi*. *Fissurella robusta* and *Petricola*

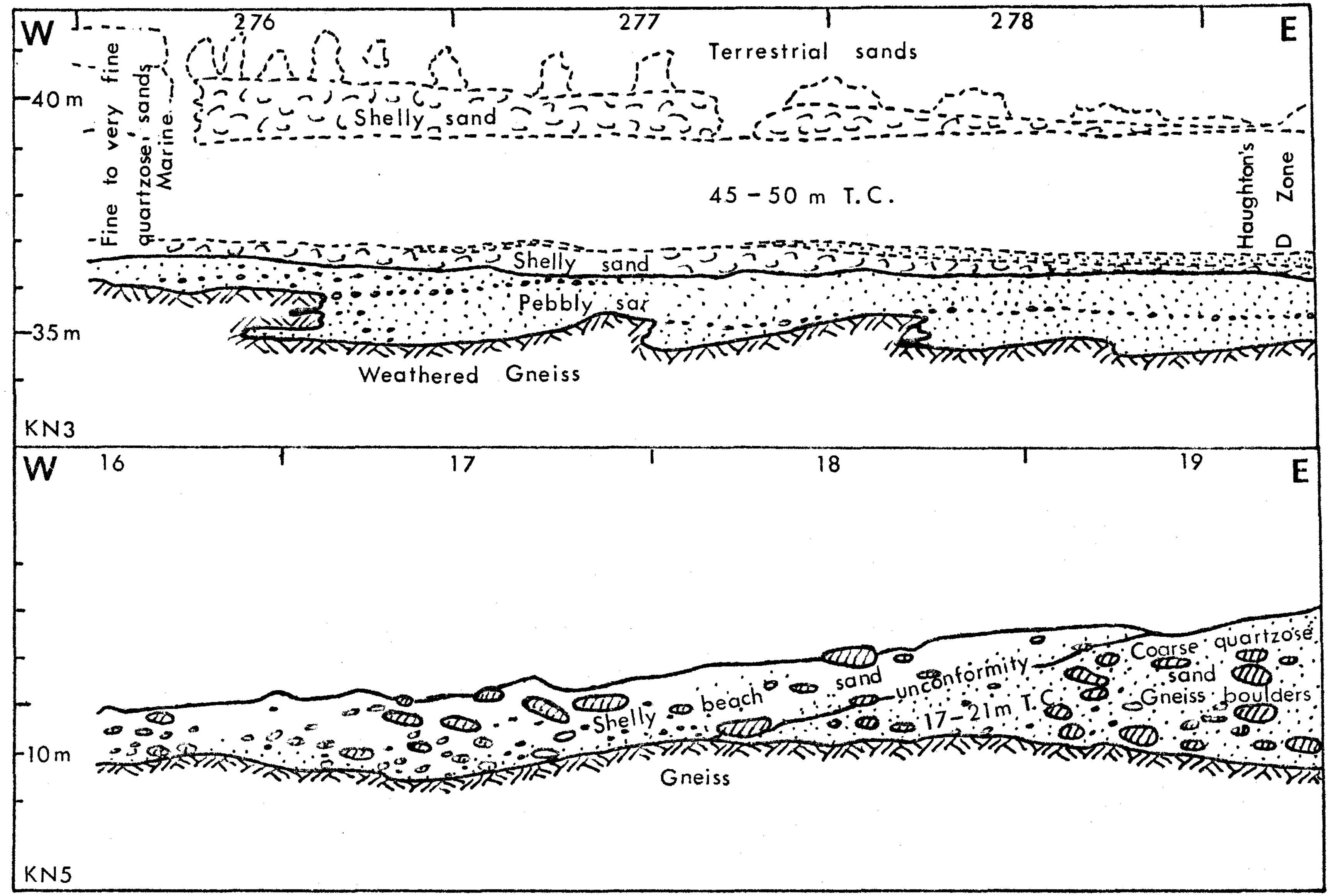


Figure 17.5 Typical lithology on Koingnaas (after Tankard 1966).

prava are also found.

A typical section, situated on line KN 5 (Figure 17.3) is shown in Figure 17.5 B. Figure 17.6 illustrates cross-bedding typical of a shoreface situation on KN 3. Trough-cross beds of variable direction indicate current reversals, and suggest a tidal origin.

3. Younger Pleistocene Marine Sediments

Marine sediments younger than the 17-21 transgressive complex were included by Haughton (1931) in his A-Zone. A typical section is shown in Figure 17.5 B. Note particularly the onlap aspect of this zone with sediments of the 17 to 21 m transgressive complex. The unconformity between the beds has a westerly dip. The unconformity is sharply defined by medium to coarse red-brown 17 to 21 m transgressive complex sediments in the east and highly calcareous shelly littoral deposits in the west. Mytilus and Patella are characteristic.

C. Summary

The earliest sediments encountered are Miocene and Pliocene phosphate-rich sediments preserved in bedrock-depressions. The Pleistocene witnessed a series of marine transgressions which reached a maximum altitude of 98 m. The mollusc faunas and phosphate horizons show that the 45 to 50 m transgressive complex sediments and older sediments may be classified as warm water marine sediments. The 17 to 21 m transgressive complex sediments and younger are typically cold water deposits.

Instability of the Namaqualand coast is hinted by the occurrence of Early Tertiary volcanics in South West Africa and the Groen River areas. Furthermore, the highest marine Pleistocene deposits in Namaqualand fall 5m in a southerly direction over 80 km (Carrington

Figure 17. 6 Cross-bedding produced by tidal reversals.
Exposure on line KN 3.

pers. comm.). Sediments of the 45 to 50 m transgressive complex in the Koingnaas area are encountered at 30 m a.s.l. in the Olifants River area where Donax haughtoni is found at that altitude (Davies 1973), i.e. a fall southward of 6 m in 80km. This supports the estimate derived from the highest Pleistocene deposits.

CHAPTER 18

SYNTHESIS OF THE LATE CENOZOIC OF SOUTH AFRICA

I. EUSTATIC HISTORY

A. Introduction

Sea level history since the Maestrichtian is one of general withdrawal of the sea from the continents. Superimposed upon this overall regression is an oscillatory pattern of transgressions and regressions. The vast Maestrichtian transgression has been unsurpassed in the Tertiary. The Cenozoic saga began with emergent continents, but two major eustatic transgressions occurred in the Middle Eocene and Miocene. Late Miocene, Pliocene and Quaternary time was marked by a progressive marine regression with superimposed glacio-eustatic oscillations of minor amplitude.

Glacially induced sea level oscillations could only be significant from the Miocene onwards, since the Antarctic continental ice sheet developed to its present thickness between the early Middle and early Late Miocene (Shackleton & Kennett 1974; see chapter 1 for a full discussion). If the Tertiary stratigraphic record suggests a synchronous cycle of sea level oscillations, then these eustatic changes of sea level must be the result of volume changes in the ocean basins (Hallam 1971).

The concept of plate tectonics and changes in ocean floor spreading rates is becoming generally accepted as an explanation for these Tertiary transgressive-regressive cycles. A relationship between increased spreading rates and mid-oceanic ridge uplift has been recognised (Russell 1968; Hallam 1971; Frerichs & Shive 1971; Flemming & Roberts 1973; Rona 1973). In particular the Neogene transgression has been attributed to ridge accretion consequent upon a spreading rate discontinuity at 10 m.y. (late Middle to early Late

Miocene) (Flemming & Roberts 1973).

B. Summary of South African Tertiary History

Criteria usually used by geologists for correlation of strata are not necessarily logical when applied to young rocks. Palaeoecological and sedimentological approaches (i.e. the facies concept) used in conjunction with an understanding of the chronostratigraphy and lithostratigraphy offer, perhaps, the best means of understanding the origin of Cenozoic strata, although there are instances where structural and geomorphic approaches have proved more rewarding (e.g. the eastern Cape).

The South African coastline originated in the Late Jurassic after the fragmentation of Gondwanaland (Dingle & Klinger 1971). Extensive Late Cretaceous marine strata along the coastal margins of Mozambique, Zululand, Natal, Cape Province and Angola demonstrate a vast transgression of that age (Cooper 1974, and others), while a Danian sea level rise has been documented at Richards Bay (Orr & Chapman 1974). Although world-wide the Middle Eocene transgression was much more extensive than that of the Miocene (Hallam 1963), Eocene sediments are comparatively rare in southern Africa. The only bedded Eocene strata are found south of Lüderitz (these deposits require re-examination) (Kaiser 1926) and at Salamanga in Mozambique (du Toit 1954; King 1972). South of latitude 28°S the only indication of Eocene strata is in the form of reworked particles at Uloa (Frankel 1966), Eocene *Nummulites* in aeolianites in Durban (King 1961), and Lower to Middle Eocene microfossils at Birbury in the eastern Cape (Bourdon & Magnier 1969). Shark teeth from Birbury suggest a Middle Eocene to Middle Miocene age (Dr S.P. Applegate pers. comm.).

The Neogene history along the entire South African coastline is remarkably similar, the only major variations being those of lithology and tectonic setting. Strata from Bredasdorp to Zululand are dominated by crystalline limestones, calcarenites, and coquinas. On the west coast the characteristic rock type is phosphorite, the phosphate equivalent of limestone. The Bredasdorp and Alexandria Formations were recognised by Haughton (1925) as regressive complexes, but complicated by seaward tilting in the Algoa Bay area (Ruddock 1968). At Bredasdorp the limestones are indicative of littoral deposition. The Zululand limestones are essentially of littoral (Umkwelane Hill) and shallow marine origin. The gravel beneath the Pecten bed indicates transgression; the Pecten bed itself formed in shallow offshore conditions during a marine stillstand; and the upward coarsening flaggy calcarenites followed by foreshore deposits indicate a progressive regression.

In the Saldanha area the same history prevailed except that there exists a better preserved record of lateral facies variation. In the Miocene these sediments include a basal regolith and two peat-clay dominated cyclothems which are separated by a thin marine horizon. This minor marine unit was attributed to breaching of a barrier-island which would flood an estuarine-vlei environment. The Saldanha Formation consists of bedded Miocene phosphorites. Tilting in the Pliocene (probably Late Pliocene) resulted in a secondary and irregular transgression which again permitted the formation of several facies of fluvial, estuarine, estuarine-vlei, lagoonal and littoral origin. Final transgression carried the shoreline to 50-55m a.s.l. Miocene and Pliocene strata also occur on the Namaqualand coast where they are preserved in bedrock depressions.

Since the origin of Neogene transgressive complexes is attributed to a eustatic sea level rise, they must all be broadly contemporaneous, except for local variations due to epeirogenesis. The age of the Zululand deposits was discussed in detail. The most detailed examination of planktonic foraminifera from Uloa indicates a late Late Miocene age, the N-17 planktonic foraminiferal zone (Dr R.P. Stapleton pers. comm.). Likewise Maud and Orr (in press) have suggested a Late Miocene (possibly Middle Miocene, but certainly not Early Miocene) age for the Richards Bay Pecten bed. They came to this conclusion on the evidence of both benthonic and planktonic foraminifera (Dr R.R.

Maud pers. comm.). The foraminiferal assemblages of the eastern Cape have not permitted dating more precise than that of Miocene series. In the southwestern Cape a Middle Miocene age has been based on occurrence of phosphorite, and environment sensitive material which last formed extensively in the Middle Miocene. Other lines of evidence do not permit age determination more precise than that of Miocene. Foraminifera from phosphorite on the Namaqualand coast suggest a Late Miocene age (A.J. Carrington pers. comm), and foraminifera from limestones overlying the continental shelf phosphorites a Middle Miocene age.

All of this data bearing on the age of the major Neogene transgression gives a limited range of Middle to Late Miocene. But if these geological features indicate a synchronous formation, then their age must conform to that of the world-wide eustatic transgression, and preferably to the Miocene climatic optimum which led to the formation of phosphorite. A late Middle to early Late Miocene age is here suggested which would agree with palaeoclimatic data, and the spreading rate discontinuity at 10 m.y.

C. Summary of the South African Quaternary Eustatic History

Although there have been several accounts of the Pleistocene eustatic history (Krige 1927; Gatehouse 1955; Maud 1968; Carrington & Kensley 1969; Davies 1970-1973), these accounts make little sense since they rely on preconceived notions (such as assuming a Pleistocene glacial chronology), long-distance correlation (such as between South Africa and Morocco), and they assume coastal stability. As Hey (1971) correctly points out, such studies are no more than exercises in correlation. There also appears to be a mistaken notion that interglacials are precisely dated. For instance, Davies (1970) writes that the 60 m shoreline is probably of Cromerian age, and he dates the Cromerian to the Bruhnes-Matuyama boundary (700 ka; Second SASQUA conference discussions). Shackleton and Opdyke (1973) have clearly shown that the Comerian complex stretches back from 350 ka to well into the Matuyama Epoch (see Table 1.1).

The soundest and most logical of these studies is that of Carrington and Kensley (1969) which is based upon a very extensive mining programme, and a thorough examination of lateral facies change, palaeoecology and palaentology. One of the answers this study does provide is that the Pleistocene eustatic history is not one of a single progressive regression since the various transgression complexes are separated by terrestrial sediments.

If the elevation of these shorelines were attributable solely to glacio-eustatic control, then it would be assumed that each higher beach going back in time would have resulted from a greater degree of ice melting. This in turn would imply that each older interglacial would be warmer. The geological record does not bear this out. It must therefore be assumed that the raised beaches are the result of a glacio-eustatic sea level oscillation superimposed upon a longer term regression.

The most precise data indicate that during the last interglacial there were at least three sea level peaks: C I at 6,3 m a.s.l. (120 ka), C II at 2-3,5 m a.s.l. and C III at 0 m. Advance of the Würm ice sheets lowered sea level along the South African coast, but sea level approached to within 25 m of present sea level during the interstadial. During the maximum glacial advance sea level fell to -130 m below present datum at 17 ka.

Table 18.1 tabulates an attempt at correlating the Late Cenozoic history of the Namaqualand coast (Haughton 1931; Carrington & Kensley 1969) with that of the Elands Bay-Saldanha area (this thesis).

TABLE 18.1

Correlation of Namaqualand and South-Western Cape Deposits

	S.W. Cape	Namaqualand	
		Carrington & Kensley 1969	Haughton 1931
Late Pleistocene	0 m 2-3,5 M 6,3 m	2 m T.C. 5 M T.C. 7-8 m T.C.	C - Zone A - Zone
Middle Pleistocene	13 m	17-21 m T.C.	B - Zone
Early Pleistocene	10 m	45-50 m T.C. 75-90 m T.C.	D - Zone
Pliocene	Pelletal phosph- oritic sands	"Upper E stage" *	
	Phosphatic sand- stone lens	"Middle E stage" *	
	Estuarine seds.		
Miocene	Phosphatic sand- stone, microsph- orite,limestone	"Lower E stage"*	
	Pre-Saldanha peats	Channel sediments	

* Mining terminology.

II. REINTERPRETATION OF THE TECTONIC FRAMEWORK OF THE COASTAL MARGIN OF SOUTH AFRICA

A. West Coast

Several lines of evidence indicate that the west coast of the Cape Province was more unstable than the south coast during the Cenozoic. These two areas lie on opposite sides of a regional swell, the Agulhas Arch. The continental margin of the broad and gently sloping shelf east of the Agulhas Arch is controlled by transform faulting, while the narrow and steeply sloping shelf to the west has a margin controlled by tensional faulting (Dingle & Scrutton 1974).

The shelf break east of the Agulhas Arch is at a normal depth, lying between 120 and 180 m, and 200 m at the southern tip of the Agulhas Bank (Dingle 1973a). West of the Agulhas Arch the nature of the shelf break is variable. West of the Cape Peninsula it has an average depth of 450 m, but shoaling to 200 m west of the Orange River (Dingle 1973b). The west coast shelf break is one of the deepest in the world (Shepard 1963). Simpson (1971) noted the variation in depth of the west coast shelf break and attributed it to differential warping of the continental margin. A double shelf break which appears in places off the west coast (Figure 18.1) and which was apparently in existence during late Lower Tertiary times (Dingle 1973b) suggests a long history of instability.

The Agulhas Arch is a NW-SE striking antiform coinciding with a structural trend (NNW) of Late Precambrian origin (Dingle 1973a). The arch is associated with igneous activity, with Early Tertiary intrusive dykes following its structural trend (Kröner 1973). Aegerine trachyte which forms intrusive plugs on the northeastern side of the Agulhas Arch (the Alphard Banks) has been dated at 58 m.y. (Palaeocene) by the K-Ar method (Siesser et al 1974).

On the farm Dikdoorn on the Groen River an intrusive melilite basalt has been dated at 38,5 million years (ZSA 56). In the Bogenfels area

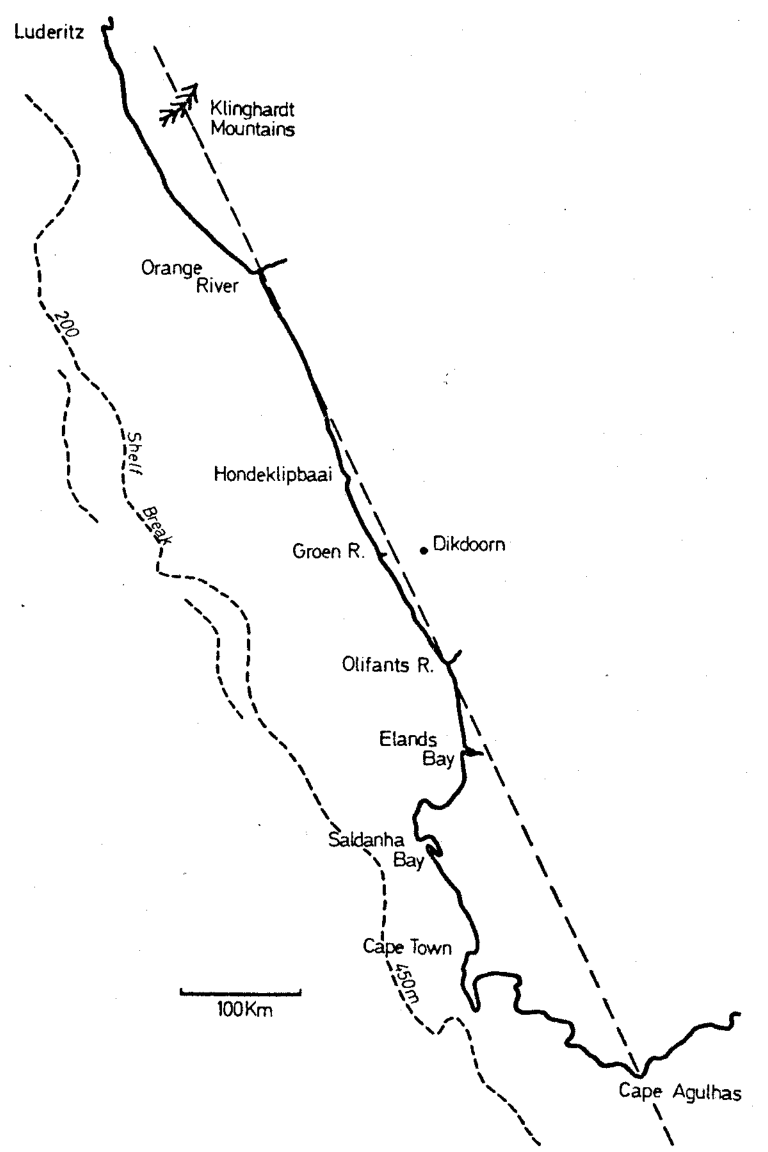


Figure 18.1 Coastline configuration and shelf break relative to a tilt axis from Agulhas Arch to the Klinghardt Mountains in South West Africa.

of South West Africa phonolitic lavas of the Klinghardt volcanism have been dated at 35,7 million years (ZSA 53).

Tilting of the area west of the arch apparently took place about an axis or "hinge line" which tended to follow NNW Precambrian structural lineaments further north and continued through the Agulhas Arch. The relative position of this inferred "hinge line" has been drawn in Figure 18.1 by connecting the Agulhas intrusives and the Klinghardt extrusives. The axis so defined passes near but west of the Dikdoorn instrusives. In reality there probably would not have been a simple well-defined "hinge line" but rather an axial-zone about which tilting and differential warping would have taken place.

It is suggested that tilting and warping west of the Agulhas axis have lowered the shelf break to abnormal depths. If the shelf west of the Agulhas Arch has tilted in relation to that east of it, it would be expected that the sedimentation to the west would be the more complex. Dingle (1971a, 1973a, 1973b) has shown just that. He has subdivided the Tertiary succession on both shelves by correlation of unconformities visible on seismic reflection profiles. A major intra-Tertiary unconformity separates Lower from Upper Tertiary on both shelves. In addition Dingle (1973b) found that the Upper Tertiary of the west shelf is further subdivisible by an unconformity into lower T3 beds (upper Middle Miocene) and upper T4 beds (Pliocene). Examination of onshore Neogene sediments between Cape Town and Saldanha has confirmed Dingle's subdivision of the shelf succession.

In the Saldanha-Langebaanweg area Middle Miocene phosphorites are overlain unconformably by the Pliocene Varswater Formation. Regression of the high Middle Miocene sea continued through the Late Miocene and Pliocene,and the Middle Miocene trangression complex sediments were for a time subaerially exposed. But tilting again dropped the area below sea level causing a local transgression during a time of world-wide regression. This second transgression resulted in erosion of the phosphorites and production of pelletal phosphorite from the liberated matrix. Pelletal phosphoritic sands are the dominant distinguishing feature of the Varswater Formation.

The maximum inland extent of the Miocene shoreline is at 56,4 m a.s.l. in the Saldanha area, where massive marine limestones are overlain by high-angle cross-bedded aeolianites. But northward in South West Africa and eastwards along the Cape south and east coasts, Tertiary marine strata are found at considerably higher elevations suggesting relative sagging and tilting of the Cape west coast since the Miocene. The highest Pleistocene marine sediments in the Saldanha area are now at 10 m a.s.l. The highest beaches in Namaqualand are at 98 m a.s.l. (Carrington & Kensley 1969). Furthermore, the mollusc fauna from the Saldanha 10 m beach correlates most closely with that of the 45–50 m transgression complex of the Namaqualand coast north of Hondeklip Bay. This fauna includes: Fissurella robusta, Purpura praecingulata, Triumphis dilemma, the largest Perna perna,and Petricola prava.

Tilting about the "hinge line" shown in Figure 18.1 readily explains the lower elevation of the Early Pleistocene shoreline in the Saldanha area. Between the Olifants River mouth and Cape Agulhas the coastline forms a "bulge", and the coastline between Cape Town and Saldanha is parallel to and furthest removed from the axis so that for any degree of tilting shorelines in this area would be found at lower elevations than those north of the Olifants River where the coastline and the axis almost coincide. Tilting alone would not affect elevation of the raised beaches north of the Olifants River mouth much, but differential warping and the local position of the axis certainly could displace them.

Dingle (1973b) has shown the existence of an outer shelf bar (T3) on the edge of the west coast continental shelf. Assuming that this bar formed close to sea level, it is possible to compute approximate rates of sinking since the end of the Miocene. Dingle calculated a rate of sinking 1m/27 000 years west of the Orange River, and 1m/1200 years west of the Olifants River. If the 10 m beach at Saldanha is equivalent to the 45–50 m transgression complex on the Namaqualand coast, and is of a late Early Pleistocene age, say 1 million years, an approximate and average rate of sinking since that time can be computed. The average rate of subsidence at a distance from the axis equivalent to that of the shelf edge west of the Olifants River is 1m/13 000 years. This agrees well with Dingle's estimate since the Miocene. The fact that last interglacial shorelines on the west coast show little variation from the Late Pleistocene 7 m datum implies that subsidence has either been episodic or has been operating at a decreasing rate. Little if any subsidence has occurred in the last 120-130 000 years.

Some data suggest that besides a westwart tilting and warping about the NNW striking "hinge line", there has also been a southerly dip of the axis. The southward deepening shelf break on the west coast has already been mentioned, and Figure 18.1 shows that the shelf break and the tilt axis are usually the same distance apart. Figure 18.1 also shows the tilt axis closely following the Namaqualand coastline so that the effect of tilting and warping should be minimal in that area. And yet the highest Pleistocene beach dips south from 98,3 m to 93 m over a distance of only 80 km (6,25 m per 100 km) (Mr A.J. Carrington pers. comm.). The 45-50 m transgression complex shoreline also dips to the south from Hondeklip Bay. Donax haughtoni is found in the Olifants River area at 30 m a.s.l. (dip 7,5 m per 100 km). The effect of tilting can also be demonstrated by comparing the three lowest Pleistocene shorelines: at Saldanha 6,3 m, 2-3,5 m and 0 m; on the Namaqualand coast 7-8 m, 5 m, 2 m (Carrington & Kensley 1969). In the case of these Saldanha shorelines it is impossible to separate the westward tilting component from the southward tilting component.

B. South and Southeast Coasts

A plot of Neogene shoreline elevation from Saldanha to Bathurst in the eastern Cape (Figure 18.2) suggests an interesting tectonic history (data derived from Ruddock 1972, Siesser 1972, and Tankard 1974a). This shoreline ascends regularly from 56 m a.s.l. at Saldanha to 305 m a.s.l. at Bathurst. As well as the tendency of the shoreline

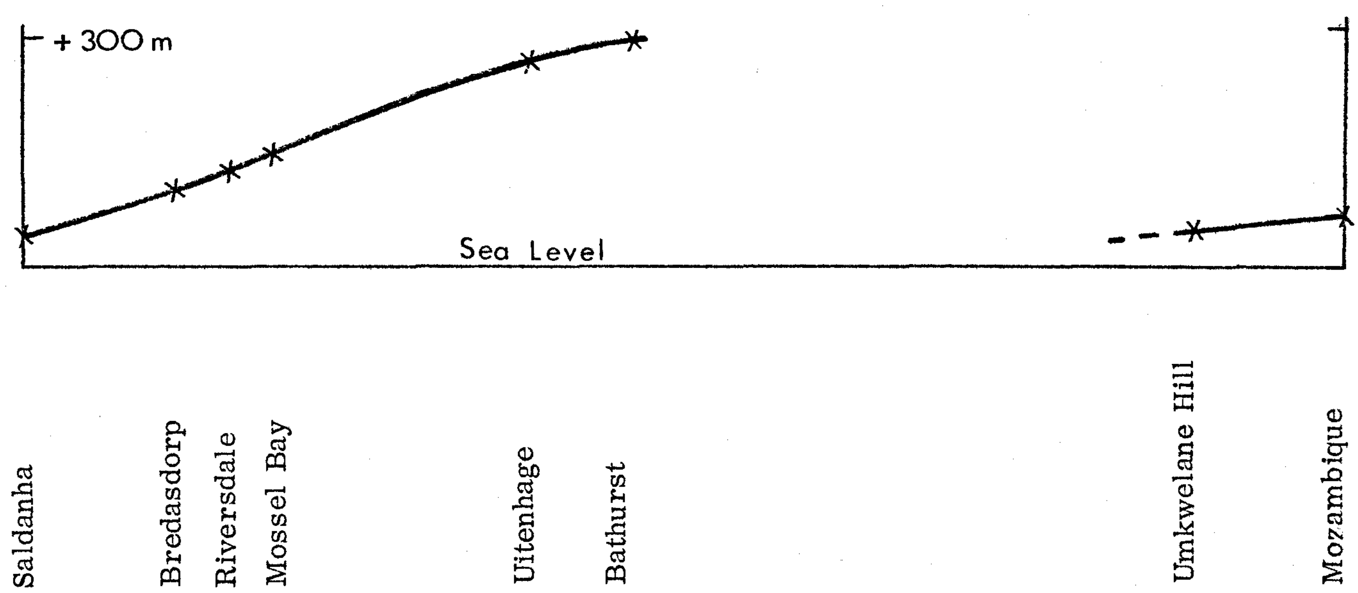


Figure 18.2 Plot of Neogene shoreline elevation from Saldanha to Mozambique showing an apparent offset between the eastern Cape and Natal.

to rise eastwards, there is a similar trend displayed by the modern shelf break (Figure 18.3). At the southern tip of Agulhas Bank the shelf break is found at a depth of 200 m (Scrutton & Dingle 1974). Off East London it is 120 m (normal shelf break depth) and off Port St Johns 100 m (Moir 1975). Although the shelf break dips regularly westwards, the Agulhas Arch has acted as a privotal point about which the shelf to the west of it has rotated, so that the depth of the shelf break increases rapidly from 200 m off Cape Agulhas to 450 m west of the Cape Peninsula (Figure 18.3).

Ruddock (1968, 1972) has documented considerable evidence to show that repeated tilting has shaped the coastal morphology, and that this tilting took place about an axis situated 33 km from the present shore-line. In Chapter 12 it was suggested that this could in effect have occurred during a single Neogene eustatic regressive cycle. Little is known of the offshore stratigraphy.

Westwards towards Bredasdorp the tectonic history appears to be extremely complex. Hales and Gough (1962) have mapped positive isostatic anomalies along the coast between 19,6^{o} E and 24,5^{o}E, and have attributed these positive anomalies partly to upwarping of the crust, and partly to uncompensated sediments deposited offshore. An effect of upwarping is seen in the deep incision of river valleys seaward. Taljaard (1948) described remnants of an old landsurface in the Swellendam-Heidelberg area, which is mantled by river gravels and duricrusts, and which dips inland. This northerly dip is due to upwarping and not pediplanation graded to river level. Evidence for this includes:

(i) northward tilting terraces;

(ii) southward decrease of size of these old river gravels, although the modern minor drainage is to the north;

(iii) major abandoned valleys carved by southward flowing rivers, now occupied by northward flowing misfit streams.

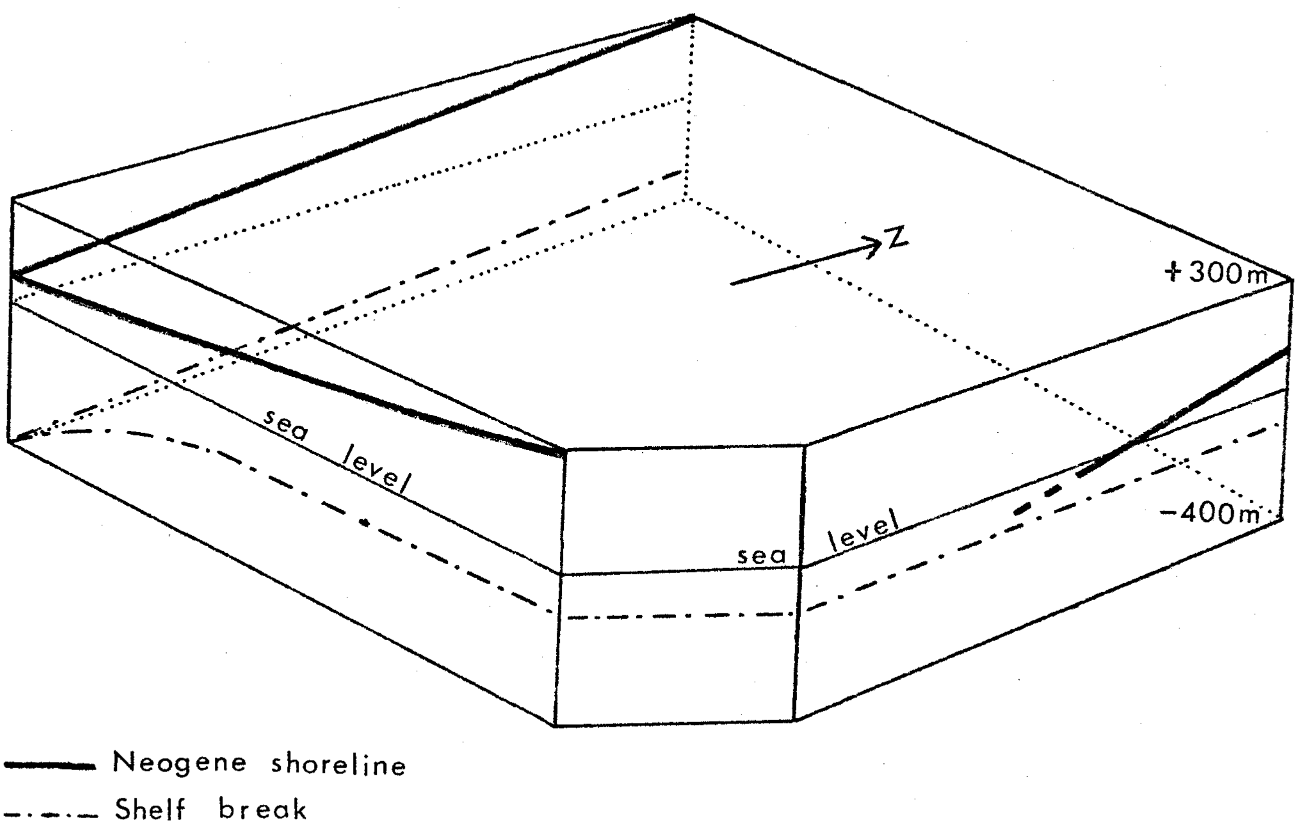


Figure 18.3 Block diagram showing the relationship of the Neogene shorelines and the modern shelf break to present sea level.

The upwarping would appear to have taken place mainly in the Early Tertiary, and not the Late Tertiary as maintained by Taljaard. Evidence for this is provided by the Neogene marine limestones which dip gently seaward (probably an original depositional dip). Unlike the eastern Cape and the west coast there is only a single Neogene regressive sequence in this area. This is matched offshore by a Tertiary stratigraphy which is subdivided into Palaeogene and Neogene units by a major unconformity. But a Miocene-Pliocene unconformity, which characterises the west coast shelf geology, is absent (Dingle 1973b). There is no known data to suggest seaward tilting in the Pliocene.

C. East Coast

Interpretation of the Neogene geology of the Zululand coastal plain has long been a contentious issue. King (1953, 1966) recognised two separate marine units (Miocene and Pliocene) which he ascribed to monoclinal tilting. Frankel (1960, 1966, 1968) and Maud and Orr (in press) recognised only one regressive sequence which was dated to Middle to Late Miocene. A re-examination of the Uloa and Sapolwana deposits (Chapter 12) has confirmed this latter view. And it has also been shown that the structure of coastal Natal is one of step-faulted tilted blocks and not monoclinal tilting (Maud 1961). A large positive isostatic anomaly follows the coastal margin from Zululand through Mozambique, but it is apparently entirely the result of thick uncompensated Cretaceous sediments which underlie the coastal plain (Hales & Gough 1962).

Figure 18.2 shows the Neogene shoreline dipping southward from Mozambique through Zululand, also mentioned by Maud and Orr (in press), which is parallel to an abnormally shallow but southward dipping shelf break (Figure 18.3). Off St Lucia the shelf break is at 60 m, off Durban 80 m,and off Port St Johns 100 m (Moir 1975). (The Neogene shoreline was drawn from data in Soares and da Silva 1970, and King 1972).

D. The Transkei Structure

A major structural anomaly separates the high Neogene shoreline of the Uitenhage-Bathurst area from the northern Natal-Zululand-Mozambique area (Figure 18.2). This apparent offset is suggestive of a major fold. The outstanding topographic features of this area are the high Drakensberg mountain ranges, deeply incised drainage patterns, a rugged cliff coastline (an emergent form), and an exceptionally straight coast which Kent (1938) attributed to faulting. Francheteau and le Pichon (1972) suggested that the continental slope originated by shear faulting between South Africa and the Falkland Plateau of South America. Moir (1975) has noted "that a southward extension of the Lebombo volcanic trend coincides approximately with the trend of the coast south of Durban, lending weight to the postulate of a major geo-suture at this location. Duncan (pers. comm.) has noted on ERTS photographs covering eastern South Africa that a prominent, gently arcuate, dyke-like feature, concave northwards, bisects the Stormberg basaltic outpourings in Lesotho on an east-west trend which is strongly aligned with the Durban fault system previously mentioned, and the trend of the continental margin flexure. This is suggestive of a line of structural weakness which might have controlled the location of the plane of separation where the Falkland Plateau pulled away from the African plate".

E. Synthesis of Structural Trends

To construct a tectonic model the following structural trends have to be taken into account:

(i) The variation in depth of the shelf break. The shelf break off the west coast is one of the deepest in the world (Simpson 1971). Off the southeast coast it is a a normal depth, and off the east coast abnormally shallow (Moir 1975).

(ii) Neogene shorelines rising northwards and eastwards from the southwestern corner of South Africa, with the highest point situated in the eastern Cape.

(iii) Miocene and Pliocene transgressive cycles on the west coast, and in the Algoa Bay area, while the south coast (Bredasdorp area) and east coast (Zululand) Neogene geology is the result of a single transgressive-regressive cycle. Shelf sediments on the west coast have an intra-Neogene unconformity, and this is lacking from the southern shelf.

(iv) Upwarping in the Bredasdorp area which is shown by the terrace morphology and drainage patterns, and by marked positive isostatic anomalies.

(v) Early Tertiary volcanic complexes south of Lüderitz (Klinghardt), Namaqualand coast (Dikdoorn) and the Alphard Banks.

(vi) Maximum shearing of Cape supergroup rocks in the Cape Agulhas area (continuation of an old trend) and the origin of the Agulhas Arch.

Simpson (1966) attempted to explain the variation in depth of the shelf break by means of tilting of the entire subcontinent about a north-south axis passing through Cape Agulhas. This would partly account for the variation, but would not explain the maximum depth (450m) west of the Cape Peninsula. Figure 18.3 shows that flexuring about the Agulhas Arch has taken place. In this sense there is an axis passing through the arch, but it is suggested that the axis follows the Precambrian structural trend, and passes through the volcanic complexes (Figure 18.1).

King (1972, figure 1) explained the elevated Neogene deposits of the Algoa Bay area as the result of two converging strike lines producing a plunging anticlinal fold. His axis trends north-west from Algoa

Bay. If this axis (Figure 18.4) were a major feature about which rotation of the subcontinent took place, rather than Simpson's (1966) north-south Agulhas axis, the variation in depth of the shelf break from deepest in the south-west, to shallowest in the north-east, would be readily explained. Uparching along this major axis would explain the high elevation of the Algoa Neogene sediments, and the plunging anticlinal fold across Lesotho and the Transkei.

It is possible to superimpose minor patterns upon this primary structure. In addition to the primary NNW-SSE axis, two other axes are drawn, one passing through the Agulhas Arch and the west coast Tertiary volcanics, the other through the south-eastern Cape. Rotation about these secondary axes has resulted in the complex transgressive histories of the west coast and the Algoa Bay coastal margin. In both of these areas land lies on the seaward side of the axes, hence seaward tilting.

Extension of the two secondary axes would make them intersect over the Agulhas Arch. This could explain the folding, faulting, and intrusion of basalt and trachyte plugs on the northeastern side of the Agulhas Arch (the Alphard Banks) mentioned by Siesser et al (1974), the elevation of the Arch, and the considerable shearing of the Cape Supergroup rocks in the Cape Agulhas area. Furthermore, since the coastal margin here lies landward of the converging axes, the area would be upwarped.

An interesting feature of the reconstruction (Figure 18.4) that is attempted is that the major axis passing through Algoa Bay, and the minor axis of the west coast are parallel. Besides the structural trends described it is possible that there could be a host of other trends determined by lines of old weaknesses of the continental block (see Fuller 1971).

Finally, besides explaining anomalous features of the local Neogene stratigraphy, this tectonic model serves as a clear warning that height above sea level of raised beaches, whether Tertiary or Quaternary,

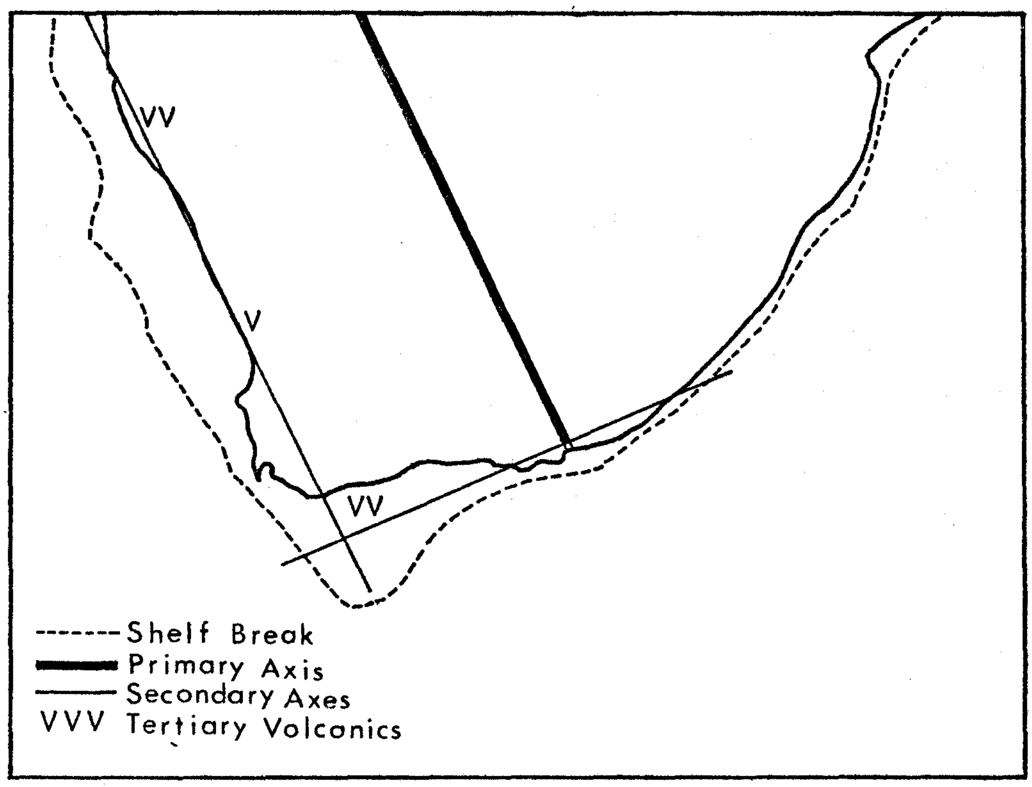


Figure 18. 4 Reconstruction of the tectonic setting for southern Africa.

is unreliable as a means of correlation, even on the same side of the arch, let alone across it.

III. EVOLUTION OF THE LATE CENOZOIC PALAEOENVIRONMENTS

A. Introduction

The physical, chemical, and biological components of a deposit are intimately related since they are the end products of processes which operated in a particular environment. These components, either individually or in association, may be used as indicators of a particular palaeoclimate or palaeoenvironment for a particular point in time. This project has been directed at the history of the coastal margin where evolution of the environment is a function of the development of the oceanographic framework.

Motion of the ocean surface is the result of wind stress so that oceanic circulation is generally aligned with the prevailing winds. In the Atlantic and Indian Oceans the ocean surface circulates anti-cyclonically around centres in the high pressure belt south of the equator (Lamb 1961). Interaction of the wind system blowing anticyclonically about the South Atlantic high pressure centre with the low pressure centre over the continents results in a southerly wind component over the south-western Cape (Chapter 14). These southerly winds blowing offshore on the west coast generate an upwelling system, the cold Benguela Current. Nutrient rich cold Central Water, which upwells from a depth of 500 m off the continental slope, originated in the Southern Ocean. It is the development of the upwelling system and the Southern Ocean (giving rise to cold Central Water) which has dominated the hydroclimate and environment of the coastal margin of the south-western Cape Province in the Late Cenozoic.

Van Zinderen Bakker (in press) has attempted to assess the age of the Namib desert biome by tracing the origin of upwelling of cold water along the adjacent shelf. He points out that it was a prerequisite of this system that the South Atlantic Ocean should have sufficient width and that it was not until the mid-Tertiary that the width was sufficient for the development of the wind and ocean current systems. But even when this width was attained, the circulation patterns would be greatly influenced by the Circum-Antarctic Current. Development of this current did not occur until the Middle to Late Oligocene (Kennett et al 1974).

Atmospheric pressure gradients are related to high latitude glaciation. Even if the South Atlantic Ocean existed in its present form by the mid-Tertiary, making an upwelling system along the west coast at that time feasible, the temperature of the upwelled water would still depend upon the temperature of the Southern Ocean (from which it was derived) and the existence of extensive Antarctic ice. Upwelling of warm pre-glacial water would not present an impenetrable barrier to moisture laden onshire winds, nor to migration of tropical and subtropical shallow marine animals.

The development of the Antarctic ice sheet has already been discussed in detail (Chapter 1). Kennett et al (1974) have summarised the Cenozoic sequence of major Antarctic and high latitude glacial and climatic events (table 1, p.1162). The Antarctic climate was temperate in the Eocene with glaciers restricted to high altitude. Major cooling and expansion of the ice cap continued through the Oligocene. The major eastern Antarctic ice cap developed in the Middle to early Late Miocene, and reached its maximum extent in the Late Miocene or Early Pliocene. Northern hemisphere ice sheets were in existence by the Late Pliocene. Since the retreat of the Antarctic ice sheet from its Late Miocene Early Pliocene maximum to its present extent in the Pliocene, there have been only minor fluctuations in the ice cover (Hayes et al 1973).

Against the background it is possible to hypothesise about the evolution of the climate of the South African coastal margin since the mid-Tertiary when the South Atlantic Ocean probably existed in its present form.

B. Miocene

Optimum conditions for precipitation of apatite on a scale that would account for the major world-wide phosphorite deposits of Neogene age occurred in the warm mid-Tertiary seas (Tooms *et al* 1969). Such phosphorites are found today in sediments adjacent to areas of modern oceanic upwelling (McKelvey 1963; Sheldon 1964). Precipitation of apatite from sea water that is saturated with respect to phosphorus is accelerated by increasing temperature and pH (Sheldon 1964). Solar radiation has only a minor effect on the surface temperature of the cold upwelled water since phosphorites are not forming extensively today. The most obvious solution to this dilemma is that the upwelling water was originally much warmer than it is today. This in turn implies a warmer Southern Ocean and, of course, less extensive Antarctic glaciation. Securely dated Neogene phosphorites mostly suggest a Middle Miocene age, i.e. they pre-date the development of the major ice sheets. The continuous existence of these ice sheets since then, and consequently colder upwelled water, would explain why there are no extensive authigenic phosphorites younger than the Miocene.

Pre-Saldanha Formation sediments on Langeberg include two extensive peat horizons with wood remains, lacustrine deposits, and soil horizons (Chapter 4). If the channel deposits on the west coast (Chapter 17) are contemporaneous, then it would be expected that a warmer Benguela Current did not act as a barrier to moisture laden winds, and that the west coast was considerably moister than today.

The distribution of modern mollusc faunas about the southern African coast is strongly provincial. Although there are overlaps, the faunal distribution defines four provinces (Figure 14.3): east coast tropical and subtropical; warm-temperate; cold; and west coast tropical and subtropical. The faunas from each of these provinces is distinctive, and there is a great diversity of species. The Neogene molluscan assemblages, on the other hand, are cosmopolitan and show a low degree of latitudinal faunal organisation. For instance, mollusc assemblages from Namaqualand to Algoa Bay are characterised by *Glycymeris borgesi*, Cardium edgari, Ostrea atherstonei, Scissodesma apengleri, Venus verrucosa, Lutraria lutraria, Dosinia lupinus, etc. A degree of provinciality is shown by the restriction of Pecten sapolwanaensis, Aequipecten uloa, and Amusium umfolozianum to the Zululand and Mozambique deposits, although these deposits have Glycymeris borgesi in common with the south and west coasts (King 1953 has commented on the tropical nature of the Uloa fauna). An identical situation has been noted for the Cenozoic deposits of California (Addicott 1970). Speciation leading to latitudinal diversity gradients is the direct result of horizontal zonation of the oceans which was consequent upon development of high latitude glaciation.

In summary, the evidence suggests a warmer hydroclimate about the entire South African coast at the time of the Miocene transgression. Furthermore, the environmental evidence tends to indicate that this transgression immediately predates the accumulation of the major ice sheets. This would imply a Middle to early Late Miocene age. And the cosmopolitan Mollusca at the base of the Alexandria Formation in the eastern Cape would agree best with a Miocene age.

Van Zinderen Bakker (in press) has correlated the development of Antarctic glaciation with a northward shift of the South Atlantic anticyclone and the Benguela Current to lower latitudes. I suggest that aridifying effects were also the result of colder upwelled water consequent upon development of the major eastern Antarctic ice sheet. These aridifying effects led to accumulation of the Kalahari strata during the Miocene (probably Late Miocene), Pliocene and Quaternary.

C. Pliocene

Deposits of Pliocene age are less widespread than those of Miocene age since the sea was generally regressive at that time. Pliocene faunas of the Algoa Bay area have not been described in detail, and it is not easy from the literature alone to separate these from the Quaternary faunas. On the west coast, however, the palaeontology is known in more detail.

In the Langebaanweg-Saldanha area sediments of the Varswater Formation are preserved in a Pliocene embayment. Several fossil molluscs suggest shallow marine temperatures 3-5°C warmer than today (Tankard 1974a). These molluscs include: Cellana capensis, Turbo sarmaticus, Barbatia obliquata and Ostrea atherstonei. Striostrea margaritacea occurs adjacent to the open coast of Duyker Eiland. S. margaritacea requires a minimum summer temperature of abou 25°C (Korringa 1956). However, their stunted valves suggest that they lived at the extremes of their distribution. The absence of authigenic phosphorite implies a hydroclimate colder than that of the Miocene, and yet S. margaritacea in deposits at Duyker Eiland implies warmer water than today, possibly with the anticyclone centre further south.

Peat and clay deposits from the Varswater Formation have yielded a pollen spectrum. The pollens indicate a flora not very different from the present flora (Prof. E.M. van Zinderen Bakker). This agrees with van Zinderen Bakker's claim (in press) that the Flora Capensis developed at the same time as the modern upwelling system, i.e. Late Miocene. But the spectrum does include Podocarpus and grass pollens. The large herbivore fossils suggest a luxuriant vegetation(Hendey 1974). Giraffe remains suggest the presence of trees, while the high-crowned teeth of Ceratotherium and Hipparion are indicative of grasslands. Hendey (1973) visualised an environment characterised by riverine woodland flanked by grasslands. This suggests strong seasonality of rainfall, with most of the rain falling in the summer months. Fluvial sediments and peats also suggest higher precipitation.

Ceratotherium also occurs in the Hondeklip Bay deposits (Hooijer 1972). Pliocene marine deposits from that area have strong west African faunal affinities and reflect water temperatures considerably warmer than today, possibly with an arid climate (Mr A.J. Carrington pers. comm.).

D. Pleistocene

1. Early-Middle Pleistocene

Haughton (1931) recognised two distinct mollusc faunas in the Namaqualand coastal deposits. Above 50 m the fauna was characterised by "Chamelea" krigei, Striostrea margaritacea (which Haughton described as Ostrea prismatica) and Donax "rogersi". Haughton attributed the widespread occurrence of oysters in these high beaches, the so called "oyster line", to changes in oceanic circulation. He suggested that the warm Agulhas Current by-passed the Cape Peninsula by means of the Cape strait, i.e. the submerged Cape Flats. The present restriction of the Agulhas Current is not due to emergence of the Cape Flats, but due to deflection by the Agulhas Bank (Chapter 14), although Agulhas water does occasionally penetrate the South Atlantic by rounding the Cape Peninsula (Schell 1968). Furthermore, Grindley (1969) has correctly pointed out that any flow through a shallow Cape strait would have been small compared with the volume of cold water washing the west coast, and could not have supported a warm water fauna north of Oranjemund ($28,7^{o}$S).

The last warm water open-coast episode on the Namaqualand coast was associated with the 45-50 m transgression complex (Carrington & Kensley 1969). The oyster Striostrea margaritacea from these deposits required a minimum water temperature of 25^{o}C in summer (Korringa 1956). The 45-50 m transgression complex sediments are apparently Early Pleistocene, i.e. pre-glacial Pleistocene. Such warm conditions must have been in response to the warmer conditions prevailing in the northern hemisphere, since the Antarctic ice sheet is thought to have been stable throughout the Pleistocene (Mercer 1968a). Warming of the northern hemisphere would have moved the intertropical convergence further south than its present location. This would have moved the South Atlantic anticyclone southwards and with it the belt of upwelling, bringing tropical waters down the Namaqualand coast. The absence of Striostrea margaritacea from Early Pleistocene deposits in the Saldanha area suggests that cool water, and upwelling, were still dominant there.

Sediments of the 17-21 m transgression complex are probably of early Middle Pleistocene age. According to Carrington and Kensley (1969) the Mollusca are predominantly cold water forms indicating a hydroclimate similar to the present. A difference between these Early and Middle Pleistocene deposits is that the fauna and fine-grained sediment of the 45–50 m transgression complex suggests a sheltered lagoonal environment, and the coarse sediment and the fauna of the 17–21 m transgression complex indicate a high energy open coast situation.

2. Late Pleistocene

Last interglacial deposits between Elands Bay on the west coast and Knysna on the south coast have been studied in considerable detail (Tankard 1975a and Chapter 14). It was found that these last interglacial faunas were characterised by the persistent occurrence of shallow-water tropical and subtropical molluscs far south of their present-day geographic ranges. These warm water taxa thrived and sustained their populations in the sheltered embayments. The warm water taxa on the west coast were characterised by *Nuculana biscupidata*, *Ostrea atherstonei*, *O. stentina*, *O. algoensis*, *Loripes liratula*, *Mactra ovalina*, *Tellina madagascariensis*, *Macoma tricostata*, *Leporimetis hanleyi*, *Venerupis dura* and *Panopea glycymeris*.

Without exception the warm water taxa survived in warm embayments, while the taxa from the adjacent open-coasts were similar to those found on the present open-coasts. A palaeoecological interpretation was based on a comparison of these fossils with their living representatives. It was suggested that the warm water element was related to the configuration of the coastline which had more protected embayments, estuaries, and lagoons than today due to flooding by a sea level 6–7 m higher than today (C I) at 120 ka. Since this was also a time of warmer climates (isotope substage 5e of deep-sea cores; Shackleton 1969; Shackleton & Opdyke 1973) it was suggested that increased solar-radiation would maintain high temperatures in the old estuaries and lagoons. The increased solar-radiation would also have had the effect of expanding the warm water isotherms, and moving the anticyclone centre, and with it the belt of upwelling, southward. Shortening of the cold water barrier would increase the possibility of mullusc larvae being able to reach the sheltered embayments. Extensive evaporites which are thought to be of last interglacial age indicate a warm arid climate.

With advance of Würm and Wisconsin ice sheets sea level fell well below present datum. But detailed examination of cave sediments at Die Kelders (Tankard & Schweitzer 1974 and in press; Chapter 16) and Nelson Bay (Butzer 1973) has demonstrated a colder and wetter climate along the coastal margin than today. A climatic model emerges whereby northward withdrawal of the anticyclone centre permitted the rain-bearing westerlies to sweep across the south coast throughout the year, albeit with a winter rainfall maximum (Tankard in press d). This northward expansion of the winter rainfall belt resulted in increased, but seasonal rainfall over the Little and Great Karoos. Relict patches of Flora Capensis still survive in the Orange Free State (van Zinderen Bakker 1967). These same conditions would explain the Late Pleistocene lake deposits at Alexandersfontein, Kimberley, which were recorded by Butzer et al (1973).

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APPENDIX 1

Grading Analyses

Grain-size determinations were made on the sand fraction ($> 63\mu$) by conventional dry-sieving methods (adequately described by Folk & Ward 1957). Approximately 50g of sample was disaggregated using a rubber stopper and porcelain mortar. Sieving was done for 20 minutes on a Retac 3D machine with 8 inch (20,3 cm) screens spaced at half-phi intervals. Each sample was weighed to 0,01g. Percentage of aggregates still left were estimated with binocular microscope and deducted from the raw material. All results were plotted on arithmetic probability paper. (The normal cumulative curve is straight when the arithmetic probability scale is used).

In computing the statistical parameters, the formulae of Folk and Ward (1957) were used because the samples frequently include the extremes of distribution, a region not covered by the more conservative measures (e.g. Trask 1930; Inman 1952). In those instances where the proportion of sand-size material was less than 95 per cent of the whole sample a pipette analysis at half-phi intervals was made.

It might be argued that the moment parameters of Friedman (1961) would give the most efficient results, but they are realistic only for unimodal distributions (Solohub & Klovan 1970). It has been shown in this project that unimodality is the exception rather than the rule. With the use of Folk and Ward's formulae, the spread between the 5 and 95 percentiles in a normal distribution is 3,3 standard deviations. Folk (1966) discusses the merits of the computer against that of the hand-drawn graph for computing the textural parameters, and describes the use of the computer as "hasty sloppiness". The most relevant disadvantages to the use of a computer are:

(i) cannot detect bimodality;

(ii) cannot detect experimental errors in weighing;

(iii) cannot see genetic relationships that may be brought out by actual inspection of the curves;

(iv) cannot compare the results with diagnostic curves such as those of Visher (1969).

The statistical formulae of Folk and Ward (1957) are as follows:

Phi Median (M_d) $= \emptyset 50$

Phi Mean Diameter (M_z) $= \dfrac{\emptyset 16 + \emptyset 50 + \emptyset 84}{3}$

Standard Deviation (σ) $= \dfrac{\emptyset 84 - \emptyset 16}{4} + \dfrac{\emptyset 95 - \emptyset 5}{6,6}$

Skewness (Sk) $= \dfrac{\emptyset 16 + \emptyset 84 - 2\emptyset 50}{2(\emptyset 84 - \emptyset 16)} + \dfrac{\emptyset 5 + \emptyset 95 - 2\emptyset 50}{2(\emptyset 95 - \emptyset 5)}$

Kurtosis (Kg) $= \dfrac{\emptyset 95 - \emptyset 5}{2,44(\emptyset 75 - \emptyset 25)}$

Besides the Mean (M_z), the average particle size may be expressed by either the median or the mode. The median is defined as the middle value ($\emptyset 50$). It is a very misleading value. For example, sample W'5 (44,9 m) has a median value of 2,65$\emptyset$ and mean 1,27$\emptyset$. In Table A.1 it will be seen that for positively skewed distributions the mean has a larger phi value than the median, the converse being true for negatively skewed populations. The difference between the median and mean may thus be used as a crude measure of skewness. The mode, the value of the variable which occurs most frequently, is certainly preferable to the median, but there exists no reliable formula for accurately determining the mode from graphs (Folk & Ward 1957). The present writer has found that the formula mode $= 3\emptyset 50 - 2M_z$ adequately describes the mode for skewness values close to zero, but as the distributions depart from the log-normal so this formula becomes unreliable.

The standard deviation is a measure of the dispersion or sorting of the sediment. For descriptive purposes the sorting classification of Folk and Ward (1957) has been adopted:

< 0,35	Very well sorted (VWS)
0,35 - 0,50	Well sorted (WS)
0,50 - 1,00	Moderately sorted (MS)
1,00 - 2,00	Poorly sorted (PS)
2,00 - 4,00	Very poorly sorted (VPS)
> 4,00	Extremely poorly sorted (EPS)

Skewness is a measure of the degree of symmetry of the population distribution. For a symmetrical distribution Sk = 0, while negative skewness or positive skewness "tails" towards either the coarser or finer population.

Kurtosis is a measure of peakedness of a distribution and depends on the relative concentration of observations in the vicinity of the mode, i.e. it is the "ratio of the sorting in the extremes of the distribution compared with the sorting in the central part" (Folk & Ward 1957). For a normal distribution K = 1 (mesokurtic). Excessively peaked samples, i.e. sorting better in the centre than the tails, are described as leptokurtic. Conversely a deficiently peaked distribution is described as platykurtic.

McCammon (1962) has discussed the efficiencies of parameters for describing mean size and sorting of sediments:

(i) <u>Mean</u>

		<u>Efficiency</u> (%)
Inman (1952)	(Ø16 + Ø84)/2	= 74
Trask (1930)	(Ø25 + Ø75)/2	= 81
Folk & Ward (1957)	(Ø16 + Ø50 + Ø84)/3	= 88
McCammon (1962)	(Ø10 + Ø30 + Ø50 + Ø70 + Ø90)/5	= 93
McCammon (1962)	(Ø5 + Ø15 ..... + Ø85 + Ø95)/10	= 97

(ii) Standard Deviation (sorting)

		Efficiency (%)
Trask (1930)	$(\phi 75 - \phi 25)/1,35$	= 37
Inman (1952)	$(\phi 84 - \phi 16)/2$	= 54
Folk & Ward (1957)	$(\phi 84 - \phi 16)/4 + (\phi 95 - \phi 5)/6,6$	= 79
McCammon (1962)	$(\phi 85 + \phi 95 - \phi 5 - \phi 15)/5,4$	= 79
McCammon (1962)	$(\phi 70 + \phi 80 + \phi 90 + \phi 97 - \phi 3$ $- \phi 10 - \phi 20 - \phi 30)/9,1$	= 87

TABLE A.1

Grain-size Statistics

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmetnite %
Langeberg	S12	8,6	0,7	1,80	1,84	0,77	0,27	0,48	MS	Trace
		11,8	0,9	2,00	2,00	0,75	0,21	1,66	MS	Trace
		15,0	0,7	2,19	2,22	0,72	0,26	2,00	MS	Trace
		20,4	0,9	1,90	2,27	1,43	0,47	1,35	PS	Trace
		22,5	0,5	2,12	2,47	1,22	0,40	0,91	PS	Trace
		24,6	0,7	2,38	2,46	1,11	1,09	0,99	PS	0,10
		26,8	1,6	2,34	2,43	1,12	0,21	1,02	PS	0,20
		28,9	2,6	2,33	2,35	1,02	0,08	0,95	PS	0,12
		30,0	4,9	2,30	2,33	1,06	0,08	0,99	PS	Trace
		31,0	6,3	2,52	2,49	1,01	-0,01	0,78	PS	Trace
		32,1	10,7	2,53	2,57	1,42	0,08	0,90	PS	Trace
		33,2	14,4	2,62	2,61	0,88	-0,02	0,97	MS	Trace
		34,2	15,9	2,58	2,73	0,99	0,21	1,02	MS	Trace
		35,3	17,9	2,56	2,67	0,92	0,18	0,99	MS	0,18
		36,4	14,5	2,40	2,46	0,84	0,08	0,95	MS	Trace
		38,5	15,8	2,30	2,54	0,93	0,29	0,82	MS	Trace
		39,6	16,8	2,45	2,53	0,89	0,12	0,92	MS	Trace
		40,6	18,1	2,74	4,00	2,66	0,69	3,93	VPS	Trace
		41,7	16,8	3,75	4,41	1,73	0,19	1,17	PS	0,40
		43,8	8,3	2,45	2,37	1,21	-0,23	1,38	PS	Trace
	Q12	12,9	0,5	2,38	2,62	1,10	0,44	0,99	PS	Trace
		16,1	0	2,11	2,61	1,78	0,54	1,37	PS	Trace
		19,3	0	2,43	2,45	0,99	0,00	0,73	MS	0,15
		22,5	0,9	2,02	2,09	0,94	0,11	1,02	MS	0,10
		25,7	2,1	2,48	2,42	0,84	-0,13	0,83	MS	0,20
		26,8	4,3	2,35	2,28	0,99	-0,08	0,78	MS	0,50
		27,8	3,3	2,90	2,56	0,97	-0,41	0,86	MS	0,30
		28,9	5,9	2,36	2,35	0,92	0,02	0,84	MS	0,08
		30,0	7,5	1,99	2,15	1,02	0,23	0,69	PS	0,24
		31,0	10,8	2,39	2,34	1,06	-0,05	0,80	PS	0,24
		32,1	11,7	2,18	2,22	0,91	-0,11	0,55	MS	0,30
		33,2	15,0	2,18	2,28	1,10	0,17	0,93	PS	0,18
		35,3	16,4	2,23	2,37	0,93	0,21	0,81	MS	0,12
		36,4	13,6	2,55	2,58	0,94	0,04	0,80	MS	0,18
		37,4	15,6	2,33	2,46	0,81	0,21	1,09	MS	0,08
		38,5	6,1	3,08	2,70	0,99	-0,46	0,70	MS	0,70
		39,6	5,7	3,38	3,31	0,46	-0,34	2,54	WS	0,40
		40,6	7,2	3,34	3,07	0,73	-0,54	0,51	MS	0,36
		41,7	9,6	3,28	3,03	0,74	-0,45	0,98	MS	0,20
		42,8	2,6	2,55	2,74	0,66	0,37	1,73	MS	0,06
		43,8	10,4	2,63	2,72	0,80	0,06	1,07	MS	0,16
		44,9	1,2	2,49	2,63	0,41	0,59	2,02	WS	Trace
		46,0	3,8	2,40	2,44	0,59	0,19	3,11	MS	Trace
	V14	12,9	1,8	2,25	2,27	0,85	0,00	0,81	MS	0,70
		15,0	4,1	2,08	2,17	0,85	0,13	0,71	MS	0,60
		17,2	5,0	2,20	2,20	0,97	-0,05	0,69	MS	1,20

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmenite %
Langeberg	V14	19,3	7,6	1,73	1,90	0,87	0,29	0,92	MS	0,30
		21,4	7,2	2,10	2,20	0,90	0,14	0,61	MS	0,40
		23,6	8,0	1,97	2,11	0,90	0,20	0,61	MS	0,50
	N4–76–04	6,5	0,7	3,30	3,02	0,69	–0,54	1,41	MS	0,18
		9,7	0,9	3,11	2,69	1,05	–0,53	0,81	PS	0,60
		12,9	4,4	2,18	2,23	1,00	0,25	1,20	PS	Trace
		15,0	0,5	2,35	2,40	1,03	0,19	1,14	PS	Trace
		16,1	Trace	2,33	2,39	1,05	0,23	1,15	PS	Trace
		17,2	0	1,98	2,11	1,06	0,31	1,27	PS	Trace
		18,2	0	1,95	1,98	0,64	0,13	1,00	MS	Trace
		19,3	0	1,50	1,63	0,88	0,42	1,82	MS	Trace
		20,4	0	2,08	2,15	0,75	0,28	1,22	MS	Trace
		21,4	0	2,32	2,32	0,57	0,07	0,96	MS	Trace
		22,5	0	2,23	2,26	0,74	0,23	1,35	MS	Trace
		23,6	0	2,12	2,14	0,73	0,27	1,58	MS	Trace
	M6–15–N6–15	4,3	Trace	2,26	2,29	1,20	0,04	0,95	PS	Trace
		6,5	Trace	3,10	2,87	1,03	–0,22	1,04	PS	0,10
		8,6	Trace	3,11	2,86	1,05	–0,26	1,00	PS	0,12
		10,7	0,5	2,78	2,56	0,95	–0,33	0,84	MS	0,26
		11,8	4,0	2,93	2,67	1,02	–0,27	1,03	PS	0,40
		12,9	5,3	3,01	2,87	1,09	–0,11	1,20	PS	0,35
		13,9	4,7	2,65	2,70	1,00	0,14	1,27	PS	0,08
		15,0	3,1	2,46	2,45	0,73	0,14	1,42	MS	0,04
		16,1	4,4	1,63	1,60	0,67	–0,05	0,92	MS	Trace
		17,2	3,5	2,66	2,64	0,58	0,13	0,52	MS	Trace
		18,2	3,7	2,56	2,56	0,54	0,12	1,75	MS	Trace
		19,3	2,3	2,38	2,19	0,59	–0,12	1,13	MS	Trace
		20,4	0,8	2,45	2,45	0,73	0,17	1,52	MS	Trace
		21,4	2,6	0,88	0,94	0,90	0,49	1,44	MS	Trace
		22,5	2,9	3,23	3,11	0,89	–0,15	1,55	MS	Trace
		23,6	Trace	2,70	2,67	1,06	0,07	1,23	PS	Trace
		25,7	0	2,45	2,62	1,07	0,30	1,21	PS	Trace
		27,8	0	3,45	4,12	1,47	0,69	1,30	PS	Trace
		30,0	0	3,70	4,03	0,94	0,51	0,91	MS	Trace
Witteklip	W'1	22,5	1,3	2,32	2,35	0,74	0,23	1,80	MS	Trace
		25,7	3,2	2,47	2,39	0,63	–0,16	1,23	MS	Trace
		28,9	2,9	2,20	2,18	0,77	0,06	1,11	MS	0,05
		33,2	1,3	1,98	1,98	0,67	0,05	1,00	MS	Trace
		35,3	1,7	1,89	1,94	0,76	0,20	1,06	MS	Trace
		37,4	4,0	2,00	2,10	0,91	0,28	1,16	MS	0,04
		39,6	10,6	2,16	2,54	1,30	0,46	1,79	PS	Trace
		41,7	13,9	2,12	2,43	1,12	0,59	2,51	PS	Trace
		43,8	11,1	1,58	1,60	0,51	0,17	1,08	MS	Trace
		46,0	9,3	2,33	2,37	1,05	0,08	0,86	PS	Trace
	W'5	8,6	0	2,24	2,36	1,02	0,00	1,20	PS	Trace
		10,8	Trace	2,40	2,35	0,96	–0,02	1,20	MS	Trace
		12,9	5,2	2,69	2,89	1,38	0,19	0,95	PS	Trace
		15,0	2,0	2,33	2,49	1,15	0,38	1,51	PS	Trace
		17,2	4,4	2,50	2,85	1,19	0,45	1,26	PS	Trace

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmenite %
Witteklip	W'5	19,3	0,6	2,33	2,32	0,90	0,08	1,43	MS	Trace
		21,4	1,2	2,40	2,36	0,75	-0,01	1,28	MS	Trace
		23,6	4,2	2,48	2,46	0,66	0,07	1,65	MS	Trace
		25,7	5,2	2,40	2,37	0,52	-0,11	1,13	MS	Trace
		27,8	7,0	2,53	2,49	0,81	0,03	1,17	MS	Trace
		30,0	10,8	2,75	2,68	0,68	-0,08	1,33	MS	1,20
		32,1	13,6	2,58	2,59	0,62	0,18	1,30	MS	0,60
		34,2	9,2	2,90	2,79	0,98	-0,04	1,67	MS	0,80
		36,4	4,4	2,96	2,74	0,98	-0,11	0,82	MS	1,20
		38,5	1,8	2,70	2,44	1,02	-0,26	0,82	PS	0,20
		40,6	9,6	2,23	2,32	0,75	0,20	1,16	MS	Trace
		42,8	3,8	3,31	3,26	0,48	-0,25	3,02	WS	0,40
		44,9	2,8	3,15	1,78	2,33	-0,77	0,98	VPS	0,20
Langeberg	M10	16,1	5,6	2,35	2,38	0,81	-0,01	0,73	MS	0,12
	M14	16,1	7,5	2,18	2,17	1,03	-0,12	0,83	PS	0,10
	M16	22,5	7,6	2,46	2,48	0,72	-0,03	0,81	MS	0,30
	N8	19,8	9,7	2,22	2,31	0,72	0,09	0,79	MS	0,25
	N12	27,1	13,9	2,04	2,13	0,87	0,11	0,72	MS	0,04
	N16	29,4	15,9	1,98	2,01	1,10	-0,05	0,91	PS	0,30
	O6	18,8	10,1	2,02	2,09	0,94	0,04	0,74	MS	0,05
	O10	29,4	12,4	2,75	2,60	0,78	-0,30	0,76	MS	0,05
	O12	34,2	20,9	2,05	2,15	0,74	0,16	1,15	MS	0,04
	O20	43,7	7,7	2,98	2,74	0,69	-0,53	0,95	MS	0,22
	P8	23,9	12,5	2,82	2,67	0,72	-0,34	0,80	MS	0,10
	P12	34,8	18,5	2,05	2,11	0,94	0,04	0,80	MS	0,05
	P16	42,9	12,7	2,35	2,31	0,95	-0,11	0,58	MS	Trace
	Q2	15,5		2,81	2,57	0,79	-0,46	0,90	MS	0,30
	Q6	15,4	10,8	2,49	2,47	0,81	-0,12	0,77	MS	0,30
	Q10	30,5	20,7	1,93	1,97	0,92	0,03	1,04	MS	0,04
	Q14	39,8	15,0	2,18	2,27	0,86	0,08	0,79	MS	Trace
	Q20	39,3		2,62	2,57	0,70	-0,17	0,88	MS	0,20
	R1	4,9	11,3	2,50	2,48	0,89	-0,19	0,76	MS	0,30
	R12	39,0	17,5	2,06	2,17	0,87	0,13	0,90	MS	0,25
	R16	43,3	8,8	1,90	2,03	0,86	0,19	0,87	MS	0,04
	S4	12,4	11,9	2,25	2,27	0,85	0,01	0,71	MS	0,50
	S10	34,8	16,6	2,33	2,42	0,72	0,10	0,81	MS	0,05
	S14	42,1	16,7	2,07	2,16	0,86	0,12	0,82	MS	0,10
	S18	38,0	7,8	2,20	2,07	1,11	-0,13	0,64	PS	0,50
	S20	30,7		1,98	1,92	0,61	-0,01	1,09	MS	0,04
	T8	23,0	12,1	1,90	2,03	0,83	0,24	0,95	MS	Trace
	T12	34,8	16,5	2,45	2,57	0,71	0,11	0,75	MS	0,16
	T14	35,4	12,2	2,88	2,74	0,69	-0,37	0,89	MS	0,06
	T16	29,4	5,4	2,15	2,21	0,90	0,07	0,64	MS	0,15
	U4	15,0	12,1	2,23	2,31	0,78	0,12	0,77	MS	0,03
	U6	25,9	12,4	2,36	2,41	0,80	0,04	0,74	MS	Trace
	U8	25,2	14,9	1,99	2,04	0,73	0,14	1,04	MS	0,12
	U14	29,4	12,2	1,44	1,69	0,76	0,51	0,94	MS	Trace
	U16	29,4	9,9	2,43	2,36	0,87	-0,14	0,70	MS	0,08

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmenite %
Langeberg	U18	27,3	7,9	2,64	2,45	1,02	−0,34	0,83	PS	Trace
	V2	11,2	10,0	2,33	2,41	0,79	0,07	0,70	MS	Trace
	V12	30,5	10,0	2,38	2,36	0,93	−0,08	0,59	MS	0,35
	W4	12,4	12,8	2,48	2,42	0,86	−0,14	0,64	MS	0,05
	W6	15,4	10,0	2,02	2,18	0,92	0,20	0,59	MS	0,08
	W10	20,9	6,8	2,32	2,31	0,92	−0,03	0,57	MS	0,08
	X8	13,4	11,5	2,20	2,27	0,70	0,11	0,90	MS	0,08
	X14	15,5	10,5	3,27	2,65	1,09	−0,81	1,04	PS	0,30
	Y12	15,5	9,7	2,95	2,49	1,06	−0,59	0,60	PS	0,35
	Y16	10,0	4,9	3,45	3,21	0,78	−0,93	0,44	MS	0,04
	Z14	11,8	4,5	3,26	3,17	0,69	−0,62	0,44	MS	0,50
Sandheuwel	S'58	23,5	3,8	3,03	2,54	1,32	−0,50	0,94	PS	Trace
Langeberg	S4-76-T4			2,15	2,20	0,81	0,26	1,32	MS	Trace
	Beach Gravel Member		0,0	3,10	2,98	0,63	−0,56	2,70	MS	Trace
	Basal Bed			3,53	2,58	1,11	−0,50	0,78	PS	Trace
PELLETAL PHOSPHORITE										
Langeberg	M14	18,6		1,88	1,90	0,65	0,09	0,91	MS	
	012	32,9		2,08	1,99	0,52	−0,28	0,94	MS	
	020	44,7		1,92	1,86	0,60	−0,09	0,94	MS	
	P4	13,4		2,02	1,98	0,57	−0,06	1,08	MS	
	P8	21,9		1,96	1,90	0,54	−0,10	1,09	MS	
	Q12	35,3		2,03	1,95	0,56	−0,16	0,94	MS	
	R6-30-S6			2,18	2,18	0,53	−0,02	1,21	MS	
	R16	33,2		1,88	1,81	0,51	−0,13	0,86	MS	
	S1			2,23	2,24	0,56	0,02	1,32	MS	
	S4-90-T4-60			2,23	2,25	0,56	0,05	1,20	MS	
	S6-106-T6-60			2,23	2,25	0,49	0,02	1,27	WS	
	S10	36,9		1,98	1,95	0,44	−0,12	0,88	WS	

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmenite %
Langeberg	S12	39,6		2,10	2,11	0,46	-0,06	1,15	WS	
	S12	36,4		2,27	2,16	0,52	-0,34	1,30	MS	
	T14	36,0		2,13	2,13	0,55	-0,04	1,26	MS	
	U4	15,5		2,08	2,12	0,47	0,07	1,44	WS	
	V8	20,9		2,03	1,96	0,52	-0,15	1,00	MS	
	V14	19,3		1,80	1,83	0,54	0,08	0,71	MS	
	X6	11,8		1,97	1,87	0,53	-0,15	1,03	MS	
Witteklip	W'1	43,8		1,43	1,48	0,48	0,15	0,88	WS	
	W'4			2,78	2,69	0,54	-0,19	0,90	MS	
	W'5	32,1		2,63	2,60	0,38	-0,10	0,80	WS	
	W'5	40,6		2,32	2,31	0,43	-0,08	1,25	WS	
QUARTZ										
Langeberg	M14	18,6		2,18						
	O12	32,9		2,20						
	O20	44,7		3,13						
	P8	21,9		2,82						
	Q12	30,0		1,99						
	Q12	35,3		2,22						
	R16	33,2		1,85						
	S10	36,9		3,32						
	S12	33,2		2,62						
	S12	36,4		3,02	2,73	0,91	-0,36	0,85	MS	
	S12	39,6		3,13	2,88	0,97	-0,34	0,90	MS	
	T14	36,0		3,07						
	U4	15,5		2,43						
	V14	19,3		1,63						
Witteklip	W'5	30,0		2,75						
	W'5	32,1		2,80						
	W'5	36,9		2,96						
	W'5	41,2		2,25						
	W'1	43,8		1,62						
ILMENITE										
Witteklip	W'5	41,2		3,65	3,60	0,42	-0,18	1,08	WS	
	W'1	43,8		3,50	3,38	0,56	-0,27	1,72	MS	

Locality	Sample Borehole	Depth m	P_2O_5 %	Md	Mz	σ	Sk	Kg	Sorting	Ilmenite %
Ysterplaat	YS 1		9,2	1,60	1,89	0,76	0,48	0,80	MS	
	YS 2			2,94	2,65	0,77	0,82	3,00	MS	
	YS 3			1,02	0,92	0,76	0,22	1,80	MS	
	YS 4		8,0	2,90	2,49	0,96	0,71	2,35	MS	
	YS 5			0,96	0,99	0,77	0,19	1,51	MS	
	YS 6			1,54	1,83	0,75	0,48	1,06	MS	
	YS 7			3,00	2,98	0,20	-0,41	2,61	VWS	
	YS 8			1,08	1,54	1,02	0,55	2,44	PS	
	YS 9			2,98	3,03	0,35	-0,18	2,53	WS	
	YS 10			1,55	1,91	0,78	0,53	0,67	MS	
Die Kelders	DK 1			1,95	1,92	0,70	0,00	1,03	MS	
	DK 2			1,90	1,84	0,53	-0,17	0,84	MS	
	DK 3			2,07	1,98	0,57	-0,22	0,89	MS	
	DK 4			2,07	2,01	0,49	-0,19	0,92	WS	
	DK 5			2,12	2,03	0,51	0,22	0,95	MS	
	DK 6			1,52	1,59	0,71	-0,12	0,96	MS	
	DK 7			1,39	1,42	0,59	0,02	1,02	MS	
	DK 8			1,66	1,67	0,65	0,00	0,93	MS	
	DK 9			2,25	2,18	0,43	-0,23	1,02	WS	
	DK 10			2,10	2,02	0,63	-0,20	0,91	MS	
	DK 11			2,13	2,04	0,54	-0,26	0,86	MS	
	DK 12			1,95	1,89	0,59	-0,16	0,89	MS	
	DK 13			2,25	2,15	0,53	-0,27	1,04	MS	
	DK 14			1,82	1,77	0,57	-0,13	0,87	MS	
	DK 15			2,05	1,95	0,57	-0,25	0,84	MS	
	DK 16			1,98	1,87	0,65	-0,24	0,84	MS	
	DK 17			1,91	1,84	0,58	-0,15	0,85	MS	
	DK 18			2,68	2,72	0,54	0,07	1,58	MS	
	DK 19			2,55	2,53	0,67	-0,04	1,37	MS	
	DK 20			2,56	2,51	0,64	-0,08	1,60	MS	
	DK 21			2,55	2,67	0,64	0,42	1,66	MS	
	DK 22			2,60	2,79	0,42	0,40	1,74	WS	
	DK 23			2,46	2,45	0,94	0,14	1,61	MS	
	DK 24			2,20	2,14	0,49	-0,19	0,99	WS	
	DK 25			2,29	2,22	0,54	-0,15	1,13	MS	
	DK 26			2,30	2,21	0,36	-0,27	1,00	WS	
	DK 27			2,25	2,16	0,47	-0,26	0,93	WS	
	DK 28			2,22	2,16	0,50	-0,18	1,01	WS	
	DK 29			2,30	2,22	0,46	-0,24	1,04	WS	
	DK 30			2,30	2,22	0,48	-0,25	0,98	WS	
	DK 31			2,20	2,06	0,75	-0,24	0,81	MS	
	DK 32			2,40	2,32	0,55	-0,20	1,08	MS	
	DK 33			2,13	2,07	0,76	-0,02	1,04	MS	
	DK 34			1,97	1,93	0,74	-0,07	0,94	MS	
	DK 35			2,02	1,95	0,61	-0,17	0,94	MS	
	DK 36			1,88	1,83	0,61	-0,20	0,88	MS	
	DK 37			1,95	1,86	0,59	-0,25	0,89	MS	
	DK 38			1,80	1,38	1,27	-0,53	1,16	PS	
	DK 39			1,85	1,75	0,70	-0,43	1,09	MS	

APPENDIX 2

Scanning Electron Microscopy

Each sample was sieved and the quartz grains soaked in dilute hydrochloric acid for 30 minutes to remove impurities. Shell grains were not treated in any way. The samples were then spread on a 1 cm diameter stub and coated with a vacuum evaporated gold-palladium alloy. Coating was done from two different angles, high and low, and the stub slowly rotated to ensure a uniform thickness of the alloy. The JEOL JSM U3 scanning electron microscope at Rhodes University was used throughout this study. For a more comprehensive account of sample preparation see Krinsley and Margolis (1969).

Much of the early work on quartz grain surface textures utilised the transmission electron microscope (TEM). But the TEM requires the use of replicas which are complicated and time consuming to make and which may be subject to distortion or to the introduction of artefacts. Furthermore, only half the grain is replicated and it is usually difficult to identify that part of the grain being examined.

Scanning electron microscopy (SEM) eliminates most of these problems. Numerous entire grains which are alloy coated on a stub permit the direct examination of each grain. Continuous magnification from about 40X to 60 000X or even 100 000X is possible. Although the TEM has a higher resolution than the SEM, 2,5Å against 25Å (Krinsley & Margolis 1971), the important textural features of sand grains are usually within the resolution capabilities of the latter. The SEM has the advantage of rapid processing of samples, ability to examine a large number of grains directly and greater depth of field.

APPENDIX 3

NOTES ON SOME OF THE MOLLUSC FOSSILS

The aim of this appendix is to present a few comments on some of the more interesting molluscan species encountered during this project. A taxonomic account would not only be voluminous, but would also be invalid since the International Commission on Zoological Nomenclature states (Article 9) that the mere deposit of a document (e.g. a thesis) in a library does not constitute a valid publication. For a detailed account of those species mentioned in this thesis the reader would be better served using the appropriate texts. A taxonomic treatment of many of the species mentioned below appears elsewhere (Kilburn & Tankard 1975). Several of these species appear to be extinct, and most of the others no longer live in South African waters.

CLASS GASTROPODA

Cantharidus suarezensis fultoni (Sowerby)

Figure A.1

This species differs from its modern relative, *C.s. suarezensis* in having fine and weaker spiral lirae, and in the oblique growth lines which always override the spiral sculpture. In Figure A.2 the ratio breadth/height for *fultoni* is compared with *suarezensis*. Kilburn and Tankard (1975) explained the differences even within a single sub-species as the result, possibly, of habitat. Figure A.2 shows that the comparatively broader *suarezensis* from Inhaca Island lives on broad bladed "sea grass", *Cymodocea*, while the narrower Durban Bay population inhabits the thin bladed *Zostera*. It can be seen (Figure A.2.A) that these same differences exist within the fossil populations (*fultoni*), and these differences could be used to indicate the type of vegetation that inhabited Late Pleistocene estuaries.

C.s. fultoni is restricted to last interglacial estuarine sediments

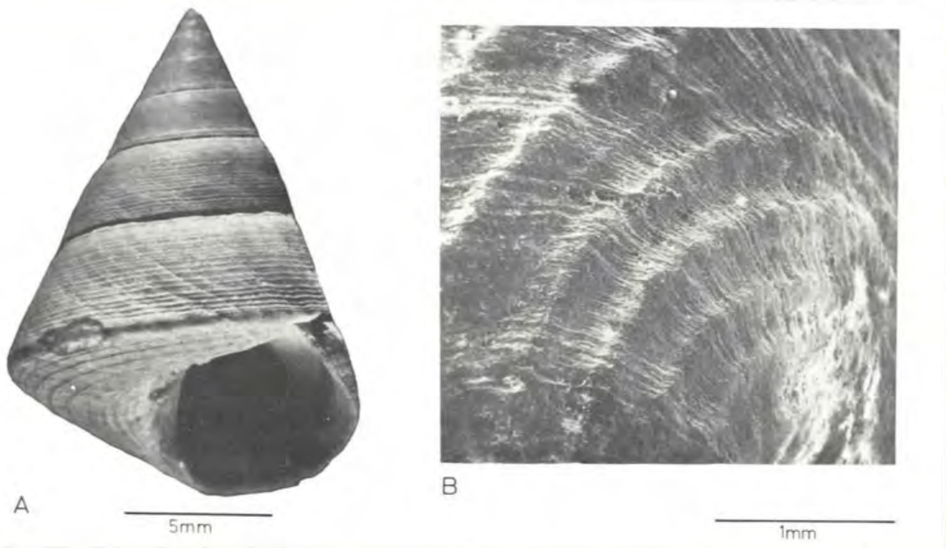
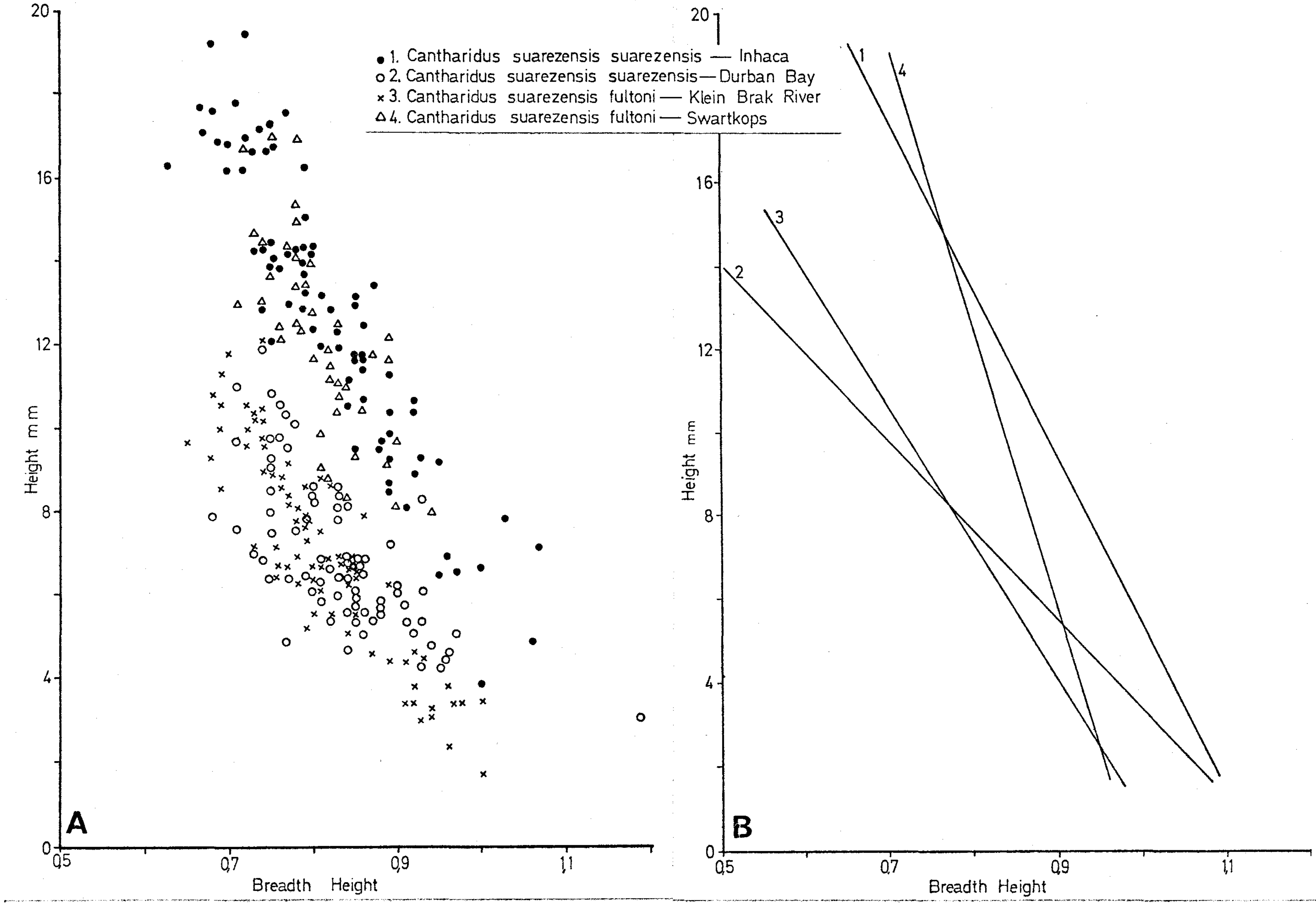


Figure A.1 Cantharidus suarezensis fultoni (Sowerby).
B. Scanning electron photomicrograph showing sculpture on the base. Swartkops river mouth.

Figure A.2 Comparison of height with ratio breadth/height for Cantharidus suarezensis fultoni and C.s. suarezensis
A. Scatter diagram. B. Calculated best-fit lines through the individual scatter diagram populations.



from Algoa Bay to Klein Brak River. The living C.s. suarezensis, on the other hand, is encountered only as far south as Durban.

Pseudostomatella orbiculata (A. Adams)
Figure A.3

This species only occurs fossil in South Africa from Knysna to Klein Brak River in last interglacial beds. Its living range is Mozambique to Ceylon (Kilburn & Tankard 1975). Barnard 1963 described Pleistocene specimens in detail.

Cerithidea (Cerithidea) bifurcata Kilburn & Tankard
Figure A.4

The closest ally to this species appears to be C. decollata (Linn.), a Recent Indo-Pacific species common in mangrove swamps in Natal, but living among salt marsh vegetation in estuaries as far west as the Gamtoos River. C. bifurcata may be the species described by Haughton (1931) as C. guinaicum Philippi (identified by J.R. le B. Tomlin) which today lives in the tropical west Atlantic. C. bifurcata was collected from the same site as Haughton's specimens: an Early Pleistocene shelly limestone 9,5m a.s.l., 1,5 km northeast of Langebaan in a quarry.

Cerithium scabridum rufonodulosum E.A. Smith
Figure A.5

C.s. rufonodulosum occurs only fossil in South Africa. Its living relative is C.s. scabridum which has a range Quirimba Island ($12,4^{o}$ S; $40,7^{o}$ E) to the Red Sea, Persian Gulf and India. Differences between rufonodulosum and scabridum are very small, and the only real difference is that the former has less prominent tubercles. C.s. rufonodulosum occurs in last interglacial beds from the Coega River mouth to Arniston.

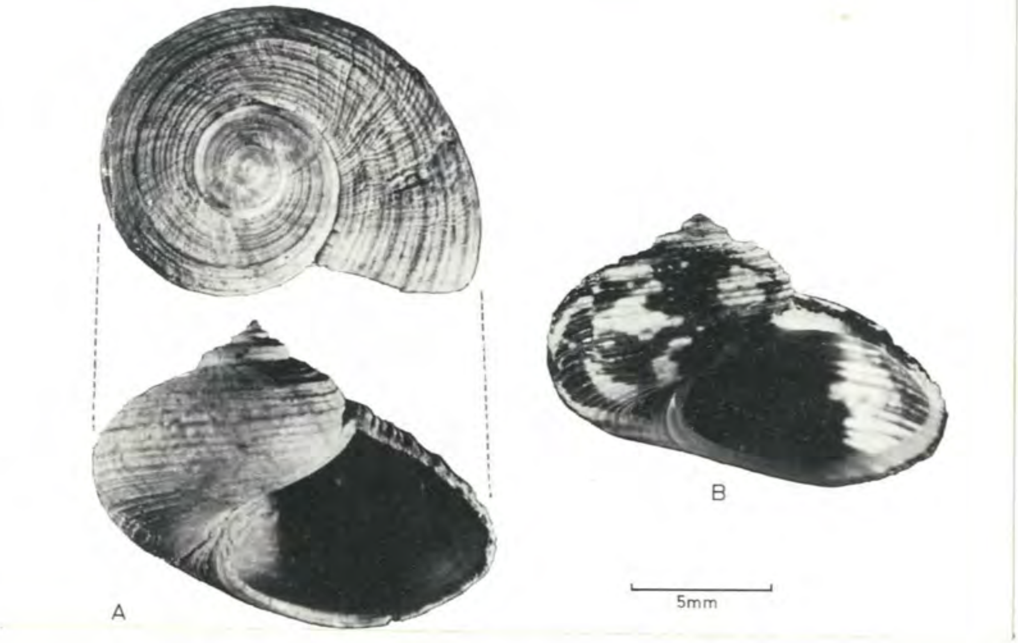


Figure A.3 Pseudostomatella orbiculata (A. Adams)
A. Fossil (Knysna). B. Living (Dar es Salaam).

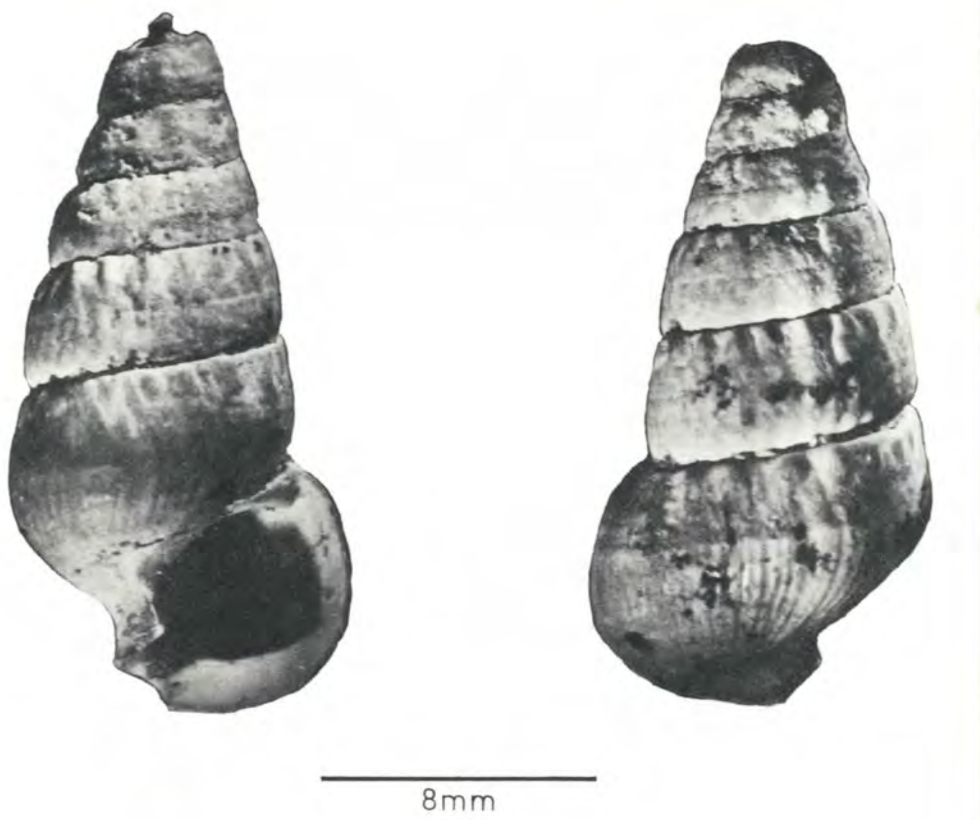


Figure A.4 Cerithidea bifurcata Kilburn & Tankard.
Early Pleistocene of Langebaan.

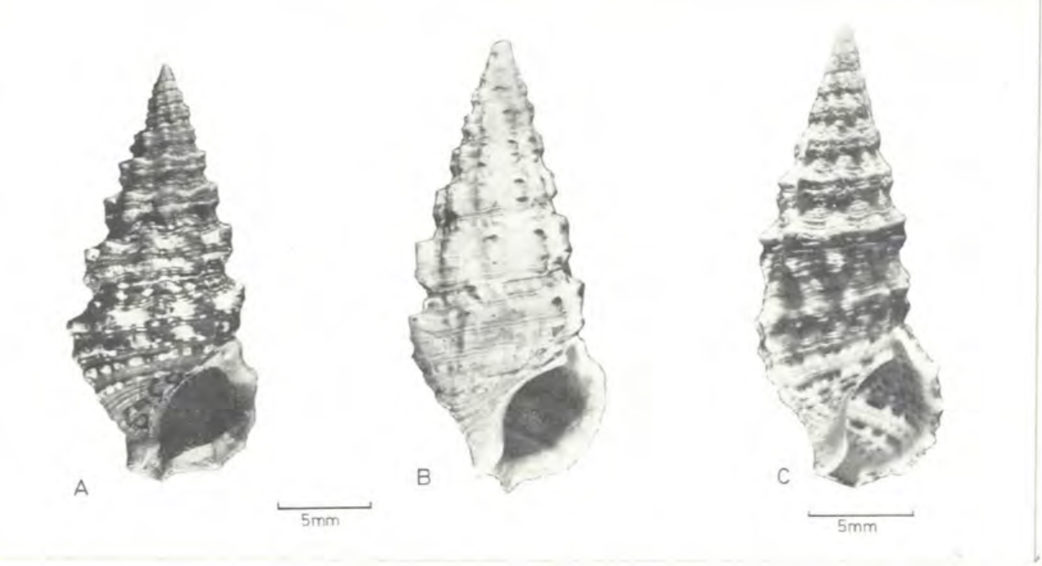


Figure A.5 A. Cerithium scabridum rufonodulosum E.A. Smith (Algoa Bay). B. C.s. rufonodulosum E.A.Smith (Sedgefield). C. C.s. scabridum Philippi (Gulf Oman).

Figure A.6 Crepidula capensis praerugulosa Kilburn & Tankard Velddrif.

Crepidula capensis praerugulosa Kilburn & Tankard
Figure A.6

This is a Pleistocene chronosubspecies of Crepidula capensis Quoy and Gaimard, and differs in its marginal apex and larger size, and in the absence of rugose sculpture. The regularly curved ventral margin of C.c. praerugulosa suggests that it lived attached to mussel shells, a common habitat of Crepidula porcellana Lamarck. Living C. capensis appears to live entirely on the undersides of rocks. One specimen of praerugulosa even shows a series of xenomorphic ridges, strongly suggesting attachment to the ribbed mussel Aulacomya ater (Molina).

C.c. praerugulosa is restricted to last interglacial beds, both estuarine-lagoonal facies and open-coast facies. Its known geographic range is Elands Bay to Milnerton lagoon (west coast).

A series of specimens from Early Pleistocene shelly limestone 1,5 km north of Langebaan is morphologically interesting, although preservation is poor. The apex in these specimens is more terminal than in typical C.c. praerugulosa, and the shells on the average more compressed, but the left side of the septum appears to be less lobate. This might be construed as indicating an origin from a porcellana-type ancestor.

Triumphis dilemma Kilburn & Tankard
Figure A.7

The affinities of Triumphis dilemma are not clear. In some respects it resembles species of the genus Cantharus Röding, 1798, but in apertural characters is closer to Triumphis Gray, 1857. The type area for this species is a quarry 1,5 km north of Langebaan at 9,5 m a.s.l. The deposit is of Early Pleistocene age. I have recently identified this species in the 45-50 m transgression complex fauna on the Namaqualand coast (Koingnaas, Hondeklip Bay).

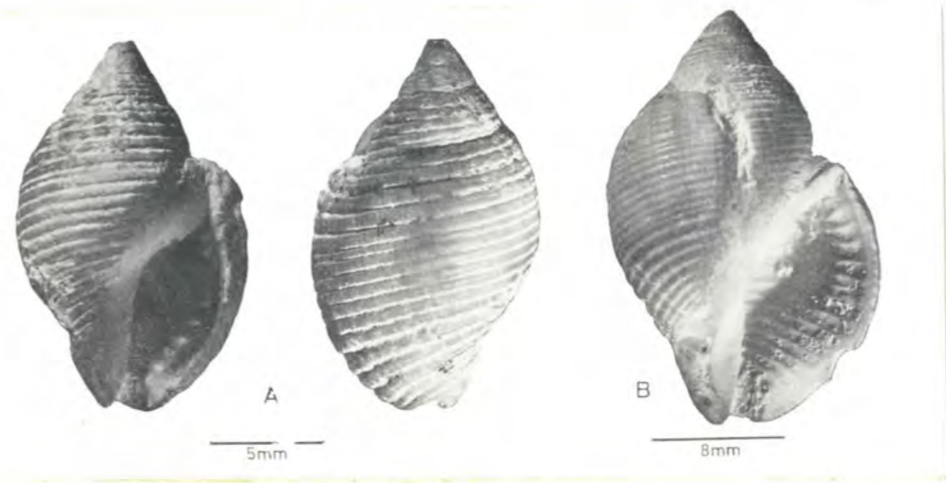


Figure A.7 Triumphis dilemma Kilburn & Tankard
Early Pleistocene, Langebaan.

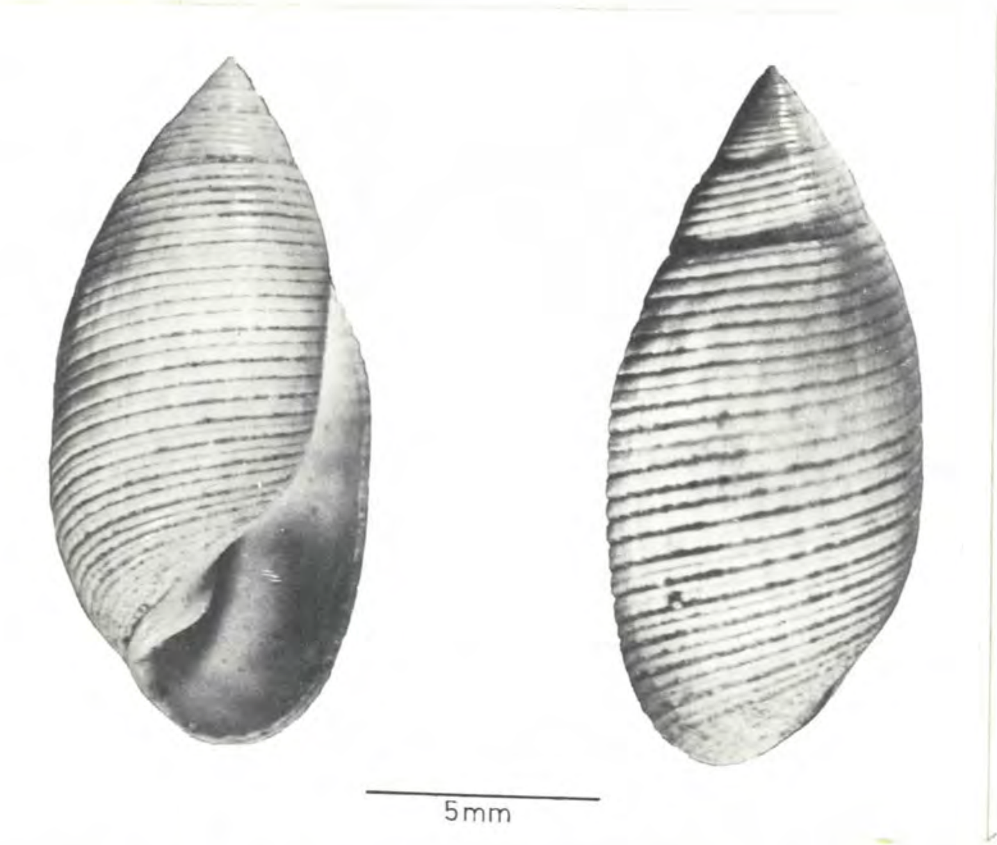


Figure A.8 Pupa daviesi Kilburn & Tankard. Knysna.

Pupa (Strigopupa) daviesi Kilburn & Tankard
Figure A.8

Pupa daviesi was initially reported from South Africa as the Philippine Pupa suturalis (A. Adams). The latter has a more conspicuously channelled suture. P. solidula (Linn.) which daviesi was originally referred to, is a globose species with very different columella folds and a rounded base. P. affinis which lives as far south as Durban is closely related to daviesi.

Pupa daviesi occurs in last interglacial beds from the Coega salt works to the Klein Brak River (Kilburn & Tankard 1975).

CLASS BIVALVIA

Ostreidae

Opinion number 338 of the International Commission on Zoological Nomenclature states that only the generic names, Ostrea and Gryphaea, are to be used for fossil oysters. This opinion arises because living oysters are classified according to their method of reproduction. But this decision would imply that a modern Striostrea valve washed up on a beach without its "soft parts" should be referred to Gryphaea, a change of generic name. Only the "living" names of the oysters have been used here.

There are three species of oyster indigenous to the Cape south coast: Striostrea margaritacea (oviparous), Ostrea algoensis ("weed oyster"), and Ostrea atherstonei (larviporous). Conchologically the various species of oyster may be separated, although any one species may show a great variety of shell shape.

Striostrea margaritacea (Lamarck)

This species has a pronounced umbonal cavity (right valve), and ventral displacement of the adductor muscle is due to the presence of a promyal passage. There are no chalky deposits in the right valve as there are in Ostrea atherstonei.

Haughton (1931) described a fossil oyster from the Namaqualand "oyster line". Mr J.R. le B. Tomlin, who examined these specimens at Haughton's request, stated that the oyster was similar to fresh Ostrea prismatica Grey from Durban. Korringa (1956) has referred these specimens to Crassostrea margaritacea (Lamarck), the common South African oyster which is prolific along the east coast. Stenzel (1971) has referred this east coast species to Striostrea which "differs from Crassostrea in its reniform adductor muscle imprints, chomata, nacreous and irridescent interior, very foliaceous shell structure and rudistiform growth pattern".

Ostrea atherstonei Newton

This species was originally described as a fossil from the Alexandria Formation (Newton 1913), but has since been found living on the south coast (Korringa 1956). It has no umbonal cavity, and has "chalky deposits" between the adductor muscle scar and the margin. The adductor scar is centrally placed and almost parallel with the hinge line.

Undoubtedly the most extensive beds of O. atherstonei are beneath the floor of Langebaan Lagoon where there are reputed to be 3 million metric tons. It thrives in a sheltered environment with fine to medium sand.

Nuculana (Leda) bicuspidata (Gould)
Figure A.9

Nicklés (1950) records this species living between Mauritania and Angola.

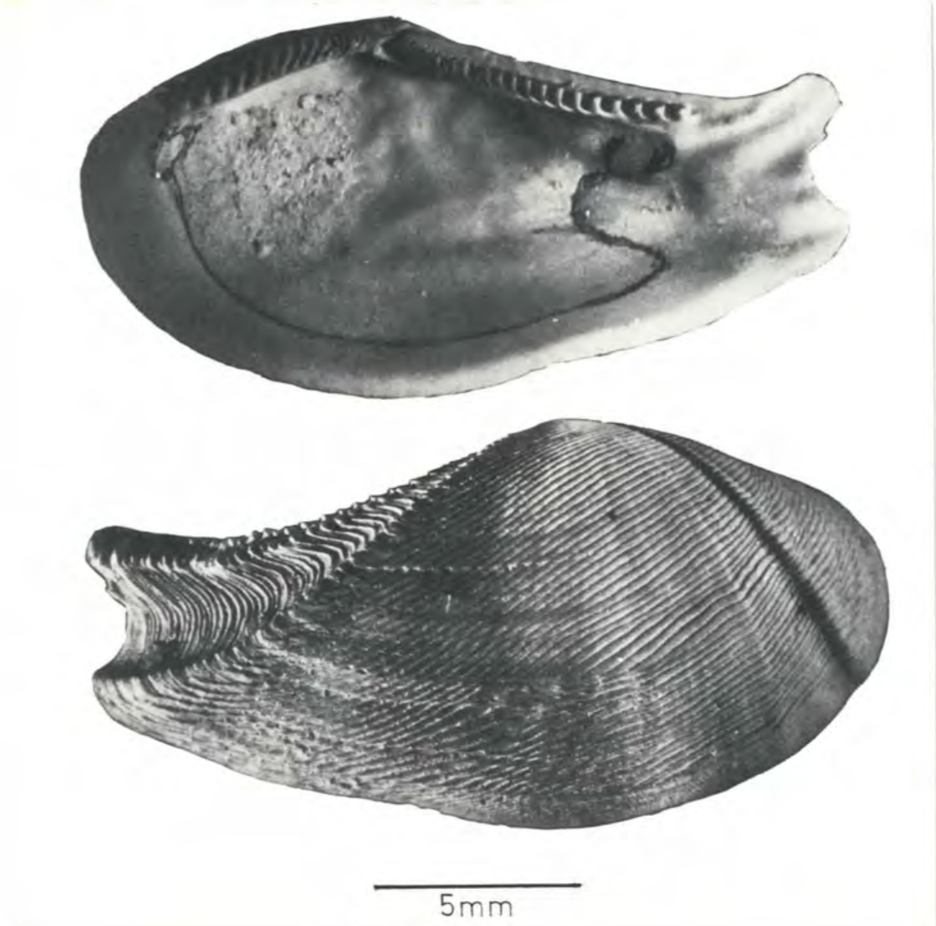


Figure A.9 Nuculana (Leda) bicuspidata (Gould). Kruispad.

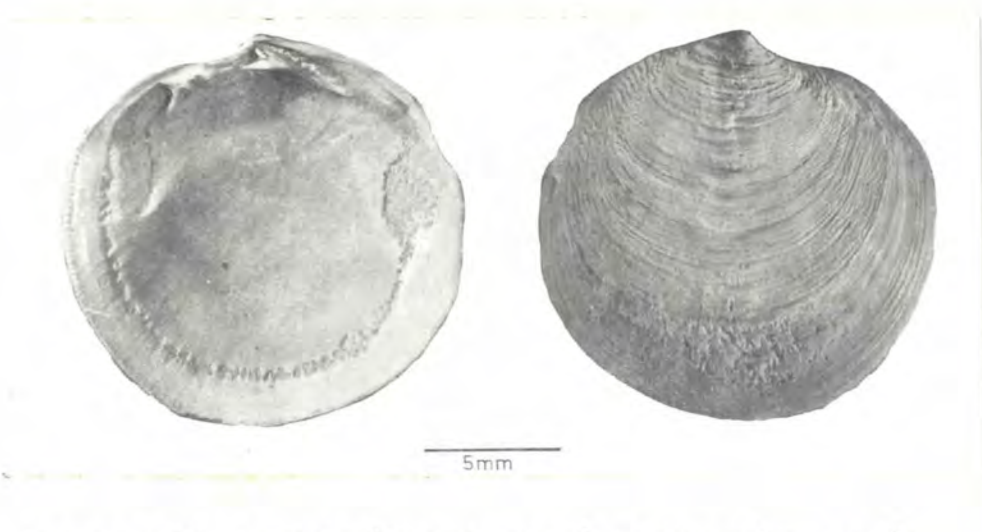


Figure A.10 Loripes (Microloripes) liratula (Sowerby). Kruispad.

I have recently examined a large collection of molluscs dredged from the South West African continental shelf. N. bicuspidata is very common in the very north which shows its distribution to be limited by the belt of upwelling. N. bicuspidata is common in last interglacial beds at Redhouse, Kruispad and Milnerton.

Glycymeris borgesi (Cox)

Although Cox (1939) described this species as G. africana, he later renamed it G. borgesi (Cox 1946), since G. africana Cox was a secondary homonym for the South African Cretaceous G. africana (Griesbach). King (1953) independently made this discovery and renamed G. africana Cox G. austroafricana (Cox).

Cox (1939) notes that borgesi is very close to the living Mediterranean pilosa. According to Professor King (pers. comm.) both pilosa and borgesi occur in the Uloa deposits, but were not separated in his 1953 report. Generally it appears that the taxodont teeth of pilosa are proportionately larger. Specimens from Ysterplaat (Miocene) are identical in all respects to borgesi from Mozambique (specimens sent by Dr da Silva), Uloa (specimens sent by Professor L.C. King), and the Zwartkops area (illustrated by Newton 1913). It is an extinct species apparently restricted to Neogene deposits.

Loripes (Microloripes) liratula (Sowerby)

Figure A.10

This species was described by Barnard (1964) as a Divaricella, but its oblique internal ligament shows it to be a Loripes. Furthermore, liratula could be a synonym of the Recent west African L. contrarius (Dunker).

Loripes liratula is common in the last interglacial beds from Swartkops River to Kruispad.

Felania diaphana (Gmelin)

Figure A.11

This Recent west African species is common in last interglacial beaches along the south Cape coast from the Swartkops River to Mossel Bay. The species at present is known to range from Mauritania to Angola (Nicklés 1950).

Cardium cf. edgari Newton

In shape and external sculpture specimens from the Miocene at Ysterplaat most closely resemble specimens illustrated by Newton (1913), and specimens of edgari in the South African Museum collection. Newton's specimens were from the Alexandria Formation.

Tellina (Eurytellina) madagascariensis Gmelin

Figure A.12

Boss (1969) showed that previous Recent records of Tellina madagascariensis were based on T. alfredensis Bartsch. He did, however, record the true West African madagascariensis from a raised beach at the Klein Brak River mouth. During the last interglacial this species (madagascariensis) was abundant along much of the south and west coasts of the Cape Province, from Redhouse to Verlorevlei, and in South West Africa at Cape Cross. After its extinction from the South African Coast due to climatic change, madagascariensis was replaced by alfredensis. T. madagascariensis is readily distinguishable from alfredensis by its larger size and deeper pallial sinus.

Macoma (Heteromacoma) tricostata (Römer)

Figure A.13

Specimens from Verlorevlei agree well with Recent valves from Angola,

Figure A.11 Felania diaphana (Gmelin). Mossel Bay.

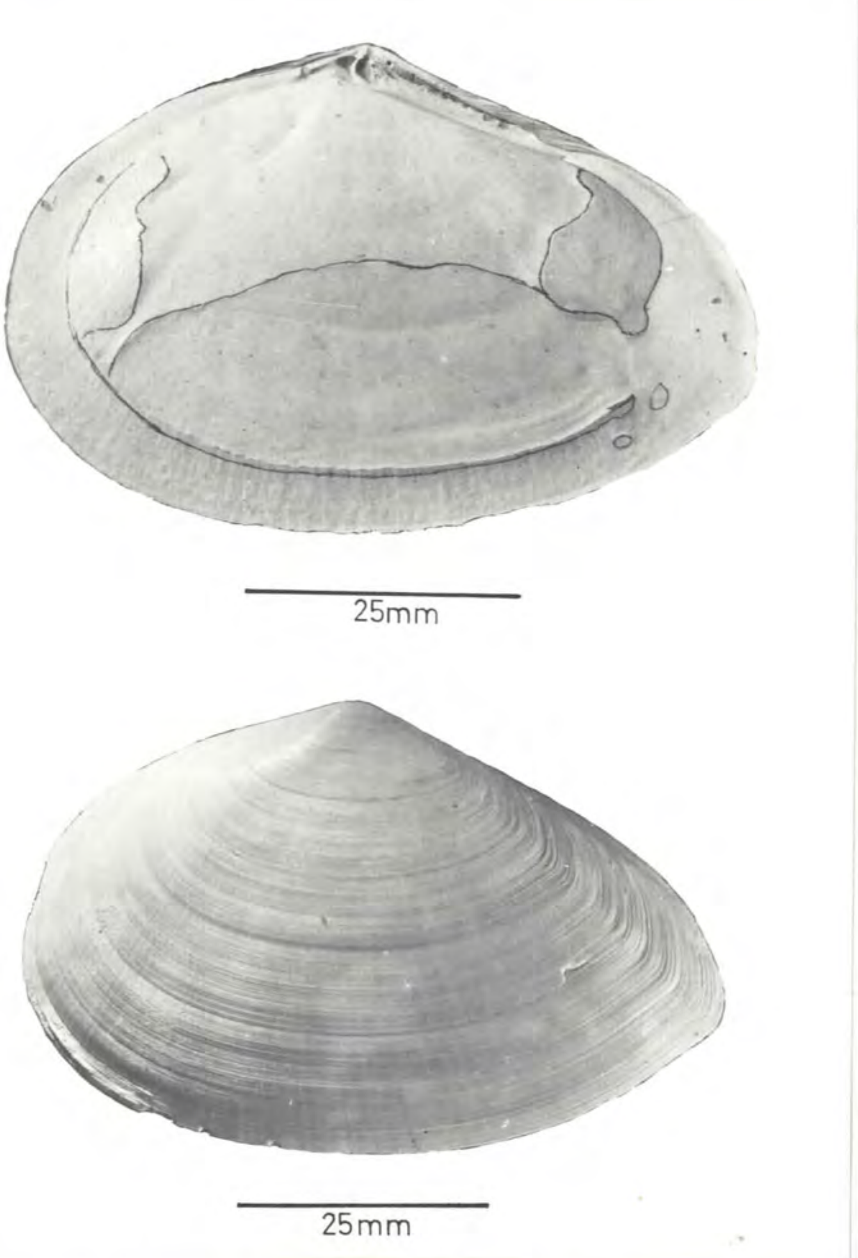


Figure A.12 Tellina (Eurytellina) madagascariensis Gmelin. Churchhaven.

except in being thicker. The Recent South African *M. litoralis* (Krauss) is a smaller, more compressed species with a lower umbo, stronger hinge-teeth and inconspicuous nymphs. *M. tricostata* has so far been identified only in the last interglacial deposits at Verlorevlei.

Gastrana fibrosa Kilburn & Tankard
Figure A.14

G. fibrosa closely approaches *G. rostrata* Carrington & Kensley from the Early Pleistocene of Namaqualand in size and sculpture, but differs in its non-rostrate posterior end and dentition. It is very different from living *matadoa*. *G. fibrosa* is found in last interglacial beds at Saldanha and Churchhaven.

Leporimetis (Leporimetis) hanleyi (Dunker)
Figure A.15

These specimens agree closely with material from Luanda. The species shows slight variation in shape. It is encountered in last interglacial beds from Redhouse to Klein Brak River.

Donax sanctuarium Kilburn & Tankard
Figure A.16

This species appears to be distinct from any other member of the genus. Its thin shell, weak dentition, and smooth inner ventral margins suggest that it is adapted to a sheltered environment. So far this species has been found only in the last interglacial beds at Churchhaven.

Pitar (Lamelliconcha) sp.
Figure A.17

The general shape, hinge detail (determined from silicone casts), and

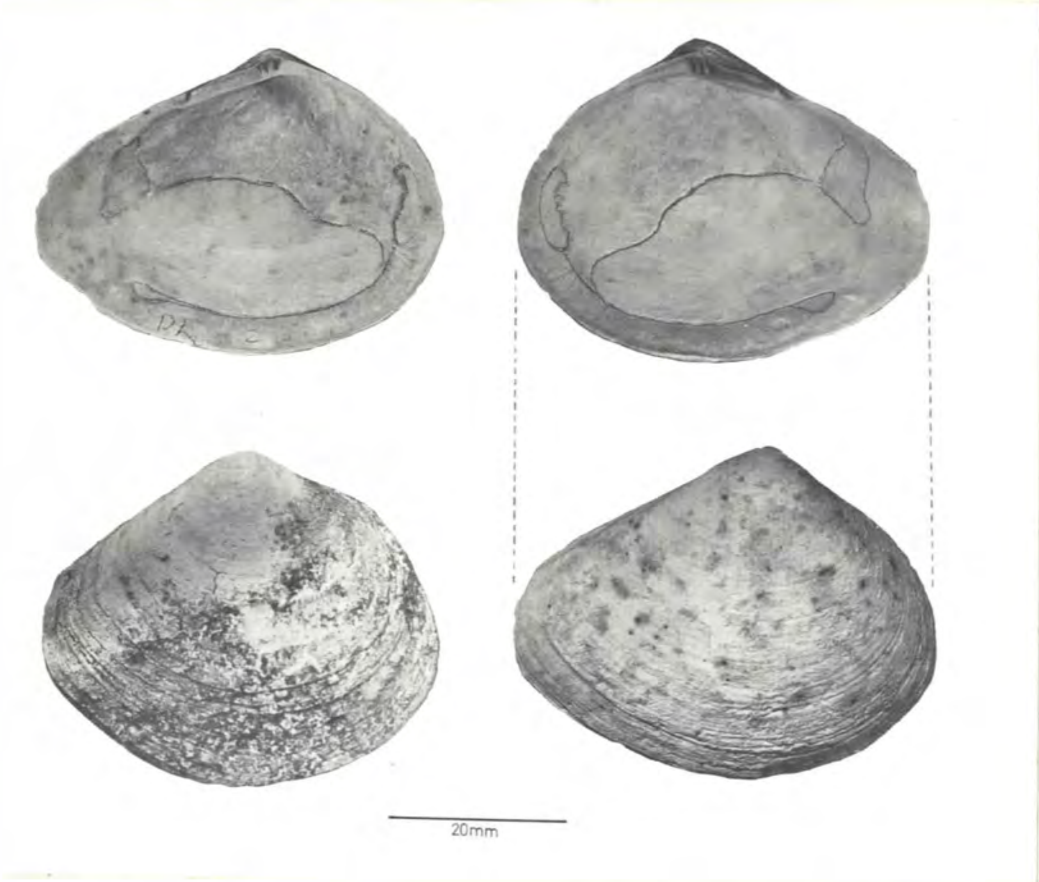


Figure A.13 Macoma (Heteromacoma) tricostata (Römer). Verlorevlei.

Figure A.14 Gastrana fibrosa Kilburn & Tankard. Saldanha.

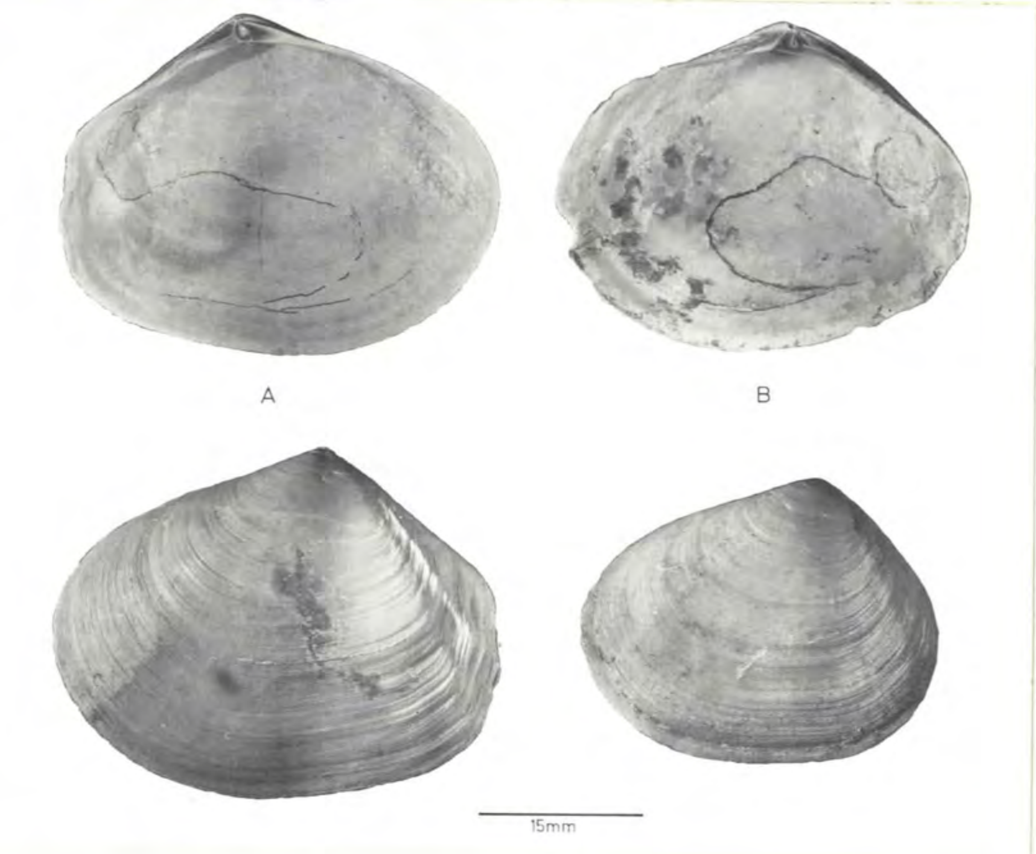


Figure A.15 Leporimetis (Leporimetis) hanleyi (Dunker).
A. Knysna. B. Redhouse.

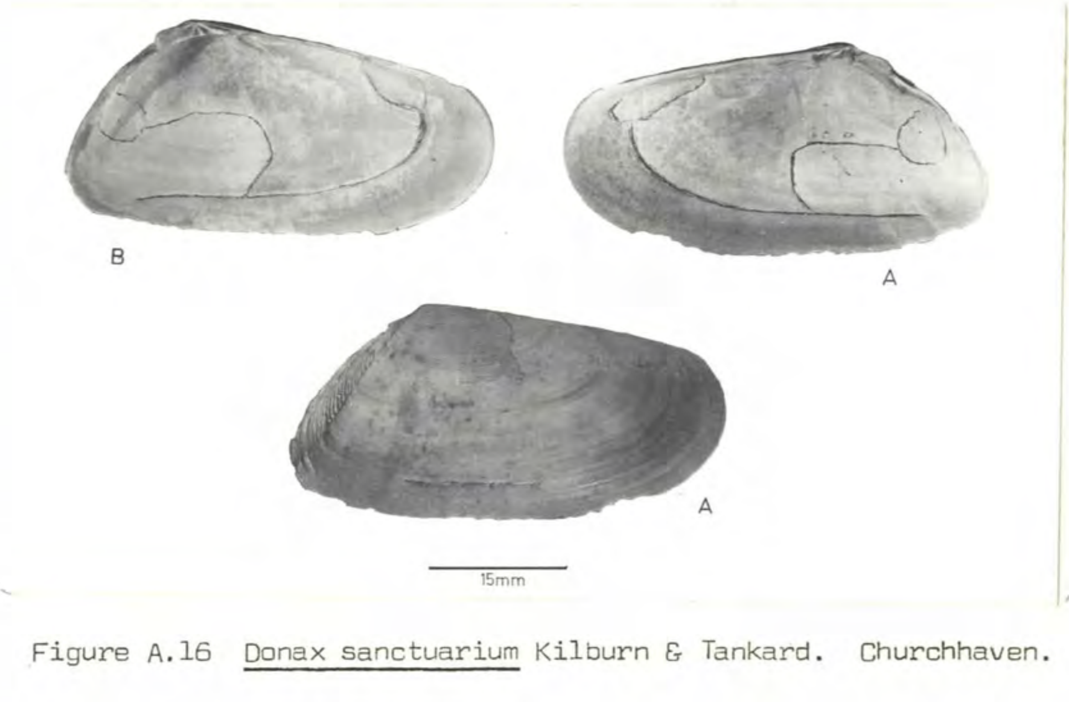


Figure A.16 Donax sanctuarium Kilburn & Tankard. Churchhaven.

configuration suggest similarities with P. madecassina of the Natal-Mozambique coast. A firm determination is precluded by the fact that only moulds are available from the Saldanha and Varswater Formations.

Notocallista schwarzi (Newton)

Newton (1913) described two new species of Chamelea, schwarzi and rogersi. Barnard (1962) was correct in synonymising these two species but erred in suggesting that they were based on worn Pitar madecassina. This Neogene species differs conspicuously from P. madecassina in its compressed valves, oblong-ovate shape, moderately curved umbo and weak umbonal ridge; also the anterior and median cardinals of the left valve only intersect at the very dorsal margin. P. madecassina is tumid, ovate-trigonal, posteriorly subrostrate with a strong umbonal ridge, umbo strongly prosogyrate, left anterior and median cardinals intersecting some distance below the dorsal margin (Kilburn & Tankard 1975). In dentition and wide, horizontal and pointed pallial sinus it most closely resembles Callista and Notocallista, but the sculpture of irregular concentric grooves and ridges is more like Notocallista.

Venerupis dura (Gmelin)
Figure A.18

This is a west African species living as far south as Luanda (Angola), but also abounding in last interglacial deposits from Redhouse to Verlorevlei. Mr C.P. Nuttall of the British Museum (Natural History) kindly examined specimens from Verlorevlei and Kruispad and found them comparable with "Callistotapes vetulus (Basterot) var. plioglabroides Sacco from the Pliocene (Astian) of Piedmont", Italy. V. rufiscensis Fischer-Piette & Metivier appears to be based merely on coarsely-ribbed examples of dura (Kilburn & Tankard 1975). V. dura at present ranges from Morocco to Angola (Nicklés 1950).

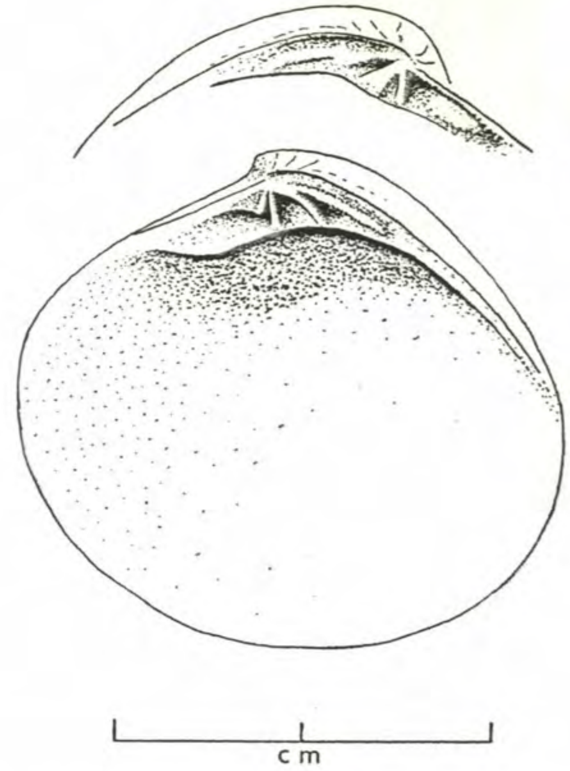


Figure A.17 Pitar (Lamelliconcha) sp. Miocene of Ysterplaat.

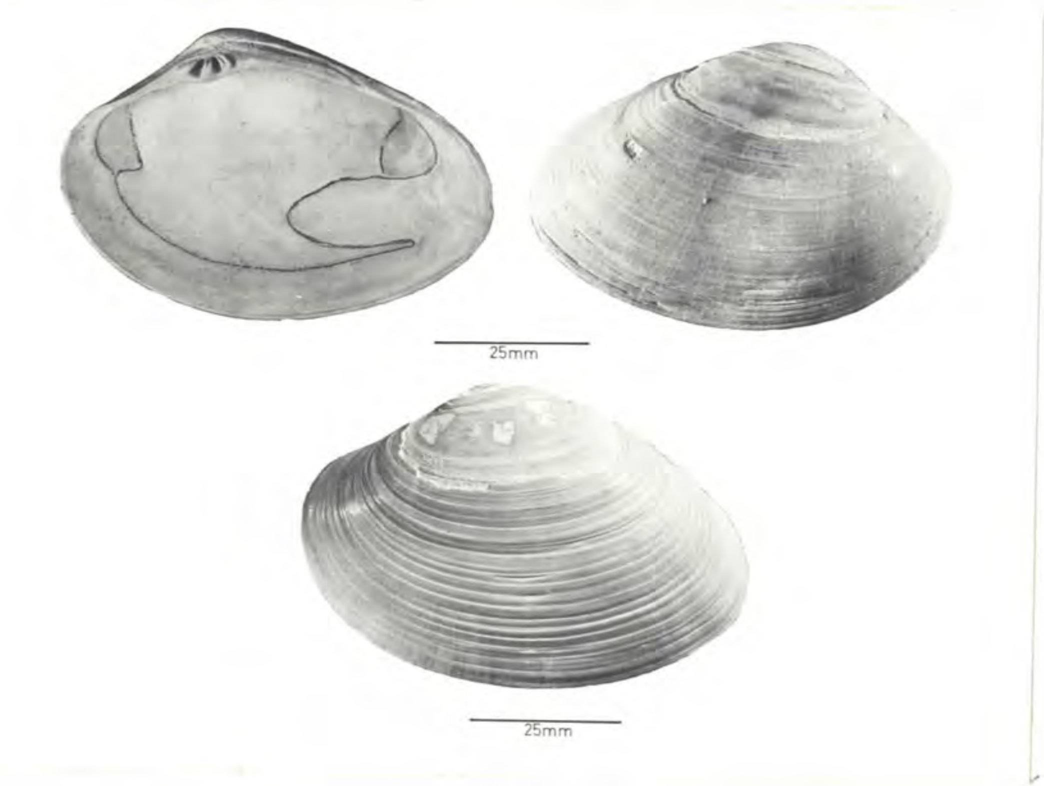


Figure A.18 Venerupis dura (Gmelin). Kruispad.

<u>Petricola (Claudiconcha) prava</u> Kilburn & Tankard

Figure A.19

The markedly inequivalve shell and large size distinguish <u>prava</u> from the three Recent South African species, <u>P. (Rupellaria) bicolor</u> Sowerby, <u>P. (Petricola) ponsonbyi</u> Sowerby, and <u>P. (p) divergens</u> (Gmelin). <u>P. prava</u> is common in the Early Pleistocene deposits at Langebaan and Saldanha.

<u>Panopea (Panopea) glycymeris</u> (Born)

Figure A.20

Specimens of <u>Panopea glycymeris</u> are common in the last interglacial deposits of Kruispad and Klein Brak River. In the past they have been referred to <u>P. aldrovandi</u> Gray, <u>P. natalensis</u> Woodward, and <u>P. dreyeri</u> van Hoepen. Its present day range is Mediterranean to Senegal. Kensley (1974) was wrong in describing the South African fossils as Plio-Pleistocene. They are all of Late Pleistocene age.

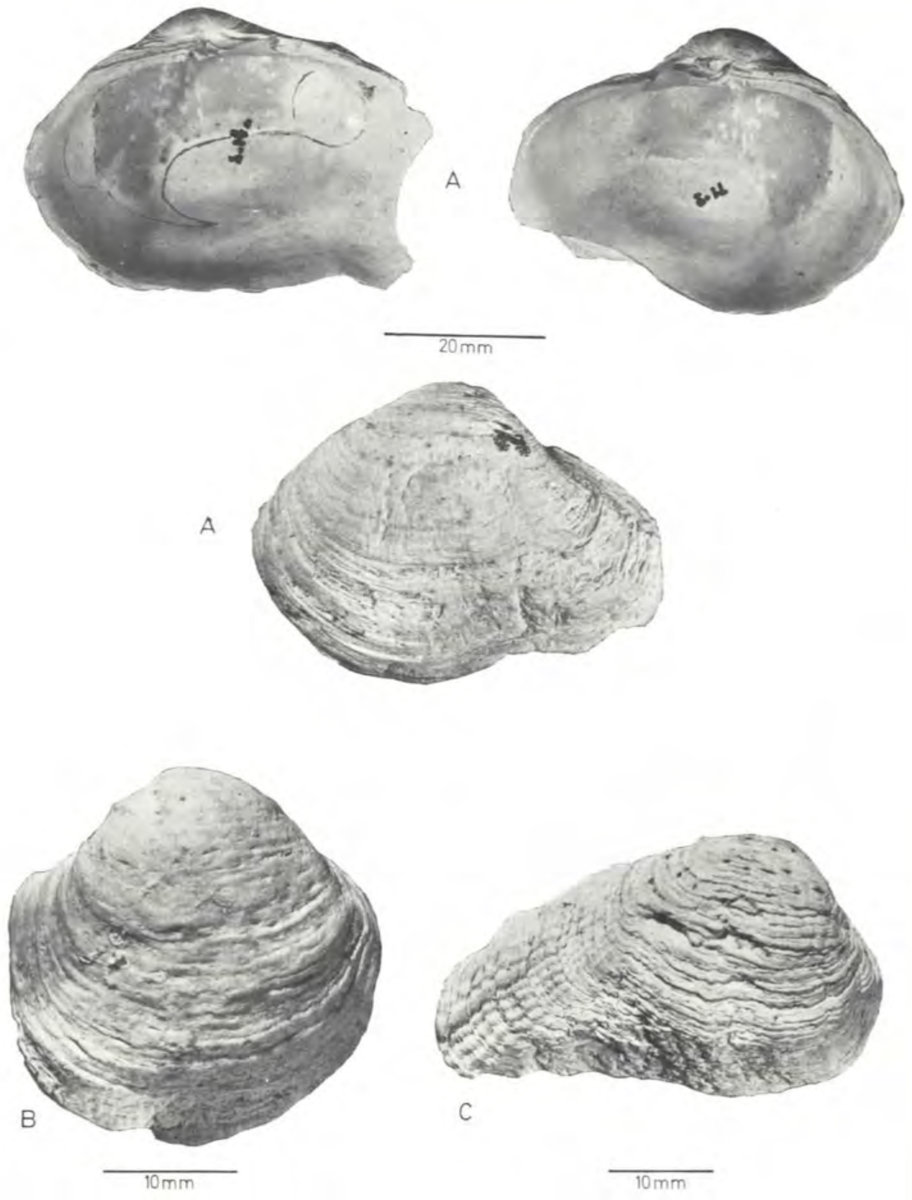


Figure A.19 Petricola (Claudiconcha) prava Kilburn & Tankard. Early Pleistocene of Langebaan.

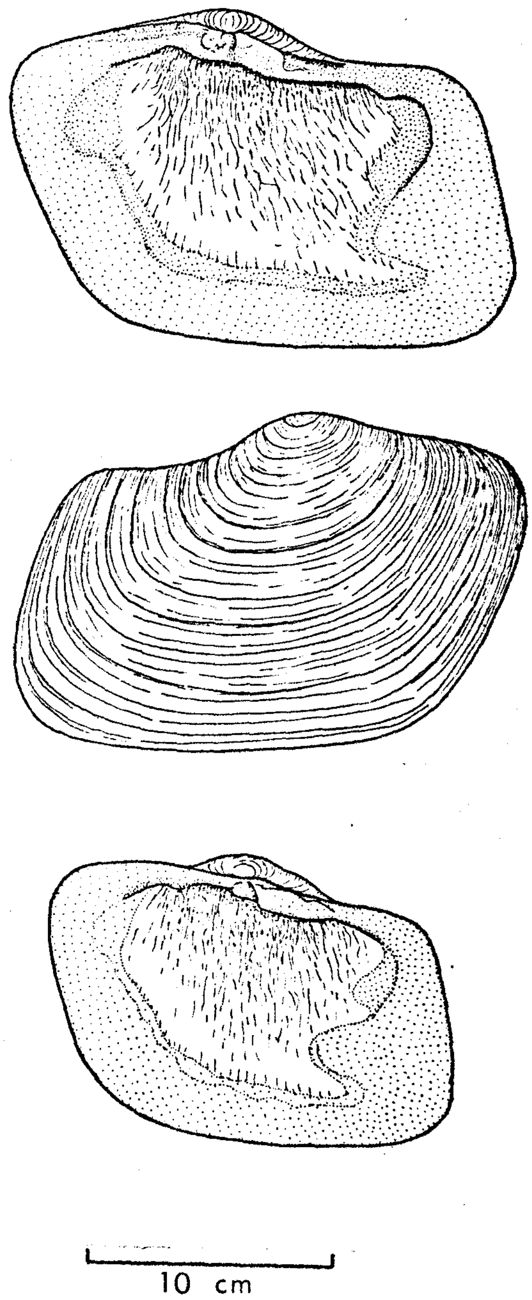


Figure A.20 Panopea (Panopea) glycymeris (Born). Kruispad.