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A geomorphology and Late Quaternary geology of the Vestfold Hills, Antarctica.

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A Geomorphology and Late Quaternary Geology of the Vestfold Hills, Antarctica

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ABSTRACT

Work on the geomorphology and Quaternary geology of the Vestfold Hills, Princess Elizabeth Land, Antarctica, is summarised. Structural, cryogenic and coastal landforms are described and an interpretation of evidence for late Quaternary events is offered.

Twelve types of periglacial landforms engendered by four agents are described. Periglacial landforms have developed on surfaces that have been ice-free for less than 5000 years. Measurement of expansion and contraction of sorted circles showed that their development is active in summer and stable in winter.

Since the late Pleistocene there have been at least two glacial cycles in the Vestfold Hills. Some of the present ice-free area was under the sea for an unknown duration ending about 25 000 years BP. From then until about 10 000 years BP ice covered the present ice-free area with a minimum thickness of 158 m. The period 6000-5000 years BP was one of milder climate with relative sea level higher than that at present so that many valleys were invaded by the sea in which there was abundant plant and animal life.

Glacio-isostatic uplift has caused regression and a fall of about 15 m in relative sea level during the last 6000 years. A series of young recessional ice-cored shear moraines reflect ice fluctuation that may correspond with Middle to Late Holocene ice advance and retreat documented in other parts of the world. Ice fronts are still in retreat.

Introduction

The Vestfold Hills (68°22′S to 68°40′S and 79°49′E to 78°33′E) occupy a relatively large (about 400 km^2) ice-free area on the eastern side of Prydz Bay, on the coast of Princess Elizabeth Land, Antarctica. The area forms a triangle bounded by the Sørsdal Glacier to the south, the slope of the continental ice sheet to the east, and the sea to the north-west. The physical geography and geology of the area have been described by several workers (Law 1959; Crohn 1959; McLeod 1963; Johnstone et al. 1973a; Burton and Campbell 1980; and Oliver et al. 1982). A reconnaissance survey of periglacial landforms can be found in Blandford (1975), but Quaternary strata and geomorphology have only recently attracted particular attention (Pickard and Adamson 1983a, b; Adamson and Pickard 1983; Zhang et al. 1983; Zhang, in press).

This report describes geomorphological and late Quaternary data collected by Zhang during 1981 and reviews further data provided by members of Australian National Antarctic Research Expeditions (ANARE) stationed at Davis, and other visitors to the area. The report was prepared while Zhang was a visiting scientist in the Department of Geography, Monash University, Clayton, Victoria. Most of the radiocarbon dates cited are from the work of the Radiocarbon Laboratory, Institute of Geography, Academia Sinica, Beijing, China and are prefixed with the Laboratory code ZDL. Unless otherwise stated all photographs used in this report were taken by Zhang.

The authors believe that the Vestfold Hills area will play an important part in the understanding of late Quaternary history of Antarctica and it is hoped that the work reported here will be a useful basis for future research. This report is the result of co-operation between Chinese and Australian scientists which the authors hope will be extended to cover further research programs in geomorphology and Quaternary studies.

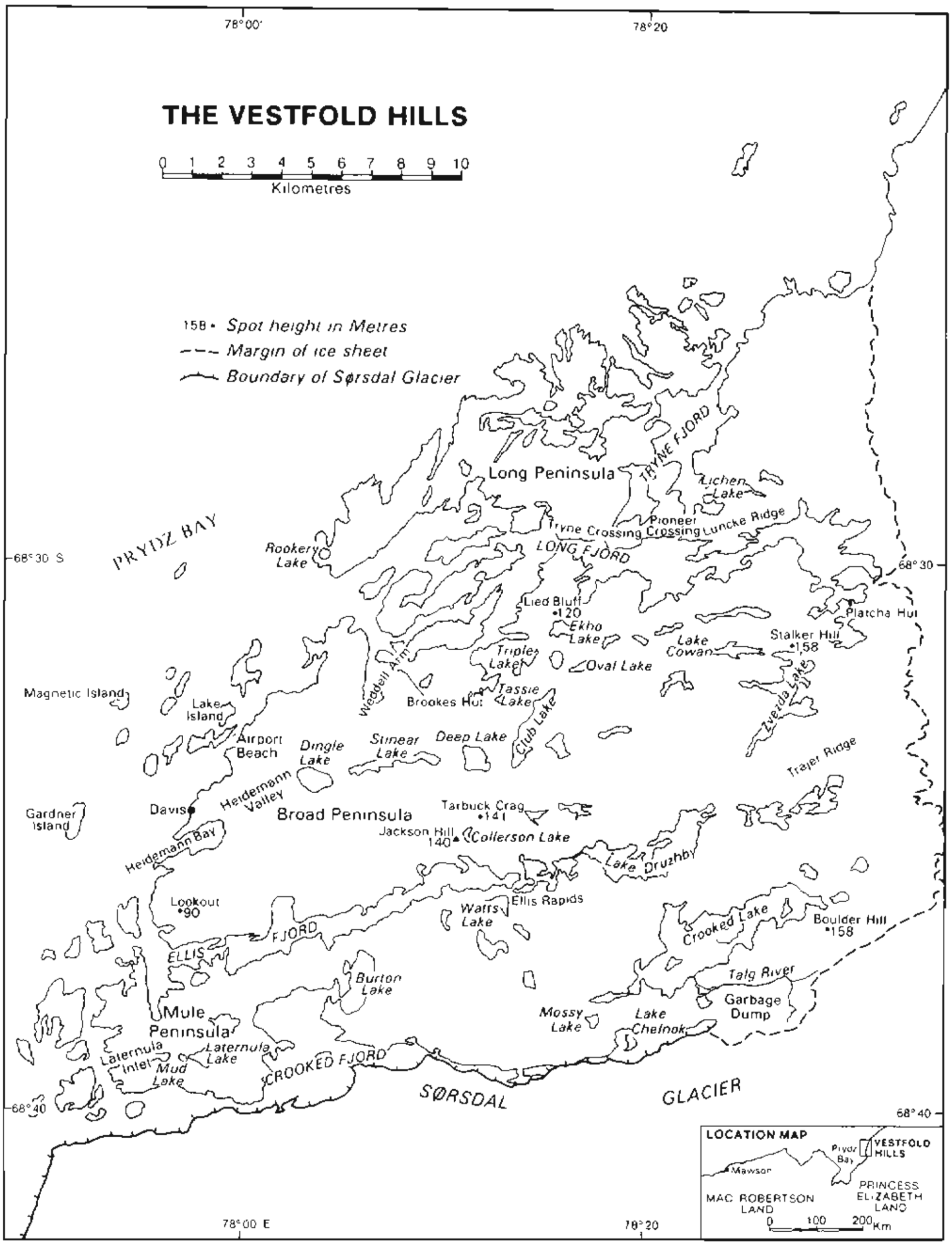

Figure 1. The Vestfold Hills showing the location of places named in the text. (Note that place names not yet in official use are enclosed in inverted commas in this map and throughout the text.)

Part A. Geomorphology

1. Background of the geomorphology

1.1 GENERAL RELIEF

The Vestfold Hills area is divided by two large east-west trending arms of the sea (Long Fjord in the north and Ellis Fjord in the south) into three parts: Long Peninsula in the north, Broad Peninsula, and Mule Peninsula on the southern boundary of which is Crooked Fjord*. Many islands occur along the coastline for up to 5 km offshore. They have a surface morphology similar to that of the mainland and some, Magnetic Island for example, appear to be partially submerged roches moutonnées.

The relief of the area is moderate, formed from ridges and hills rising to the maximum elevation of 158 m asl, and separated by many valleys nearly all of which display a preferred orientation of either east-north-east, north-east or north-west. It is later argued that this is a legacy of structural control as is the alignment of the fjords. Fossiliferous marine deposits are accumulating on the floor of the fjords and marine deposits of similar composition can be found in major valleys. That these valleys were formerly arms of the sea is confirmed by the raised beaches which can be traced along their sides to the coast. Approximately thirty basins in the floor of these former arms of the sea have trapped retreating sea water and these have become hypersaline due to evaporative losses exceeding meltwater input. Deep Lake is such a lake as it is over eight times more saline than the sea and its surface is now −56 m asl. The hypersaline lakes are listed in Tolstikov (1966). Most lakes, however, are fresh and approximately 700 lakes are mapped on the 1:50 000 map sheet 'Vestfold Hills' published by the Division of National Mapping, Canberra. Some of these lakes are in glacially eroded bedrock basins but many are at least partly dammed by glacial and/or periglacial deposits ranging in texture from large boulders to sand and silt. The largest freshwater body (Crooked Lake, 10 km^2) is close to the ice front. It is 9 km long and fed by meltwater from the Sørsdal Glacier. In contrast the hypersaline lakes lie further to the west where any meltwater that reaches the lakes is from snow patches only. Johnstone et al. (1973a) say, 'In general, there is an increase from the coastal towards the continental ice cap in the altitude, degree of relief and ruggedness of the country, and a transition from saline towards freshwater lakes'. Of the three peninsulas the mean altitude of Broad Peninsula is higher than that of the other two because of the large number of higher hills (more than 100 m asl) concentrated in its western part.

Glaciation, frost action (including wedging, thrusting and heaving) and wind action are major agents of current landform modification. However, as has been indicated by Law (1959), the landscape features owe much to the fact that the area was once

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*Designated as a fjord although the whole length of one of its shores coincides with the margin of the Sørsdal Glacier. It is worth noting that Wellman and Williams (1982) have interpreted geophysical data as indicating that bedrock lies at −750 m asl along the line of a trough occupied by the Sørsdal Glacier. In contrast the other two fjords are somewhat mis-named, being comparatively shallow.

covered by ice which advanced over the area during the Pleistocene. This probably occurred a number of times but the last and clearest event dates from about 30 000 and 10 000 years ago (Adamson and Pickard 1983; Zhang et al. 1983; Zhang, in press). Thus glaciated features appear almost everywhere in the area but the overall physiography is very greatly influenced by the lithology and structure. The geology of the Vestfold Hills is briefly summarised below.

The nature of current glacial ice formation processes, and the duration of sea ice and lake ice cover, as well as the type of periglacial regime are all a function of climate. This too is briefly summarised below.

1.2 GEOLOGY

Field work was concentrated on the geomorphology so that only brief observations were made of specific aspects of the pre-Quaternary geology (Zhang, in press). Thus the following summary relies heavily on the work of others (Oliver et al. 1982; McLeod et al. 1966; Crohn 1959; Collerson et al. 1983; and Parker et al. 1983).

The Vestfold Block is one of a number of discrete Archaean cratonic blocks now recognised on the East Antarctic Precambrian Shield (Collerson et al. 1983). According to Oliver et al. (1982) the Archaean gneiss complex in the Vestfold Hills is a variably deformed, layered sequence of diverse rock types comprising four principal map units. These units are:

(a) well-layered grey gneiss with intercalated, sharply bounded, mafic and felsic units and sporadic pods of metasedimentary rocks which exhibit a well-developed compositional layering on a scale of 1-20 cm caused by variation in the relative amounts of quartzo-feldspathic and mafic constituents, and accentuated locally by the intercalation of discrete mafic units;

(b) well-layered garnetiferous paragneiss with subsidiary mafic and submafic intercalated layers. The most common of the metasedimentary rocks is garnetiferous paragneiss. It typically consists of thin, elongate, lenticular layers of fine to medium grained quartz-feldspar-biotite-garnet-orthopyroxene gneiss intercalated with coarse grained (average grain size about 1 cm) quartz-feldspar-garnet gneiss;

(c) acid to intermediate orthogneiss with diffuse layering is medium to fine grained and ranged in composition from granitic to dioritic. The characteristic diffuse, discontinuous layering results from the elongation of mafic to submafic schlieren or variation in the texture of the quartzo-feldspathic component; and

(d) acid to intermediate, relatively homogeneous orthogneiss, with simple fabric, of which the dominant component is a medium grained, uniformly textured rock, weathered to buff colour, with 'streaky' foliation and downdip lineation of variable intensity. The foliation and lineation, where well developed, are due to parallelism of ellipsoidal quartz grains, feldspar prisms and ferromagnesian mineral clots 1-5 cm long.

A number of relatively large ultramafic bodies are included within the gneiss. Many are linear features which can be seen in aerial photographs, or consist now of a line of boudins or pods each 20-30 m in width. Within the metasedimentary rocks as well a number of smaller bodies of the same ultramafic composition occur as boudins.

There are no large pegmatite bodies in the Vestfold Hills but minor pegmatite development is widespread, mainly in the form of thin veins and patches. Gneissic rocks of a variety of compositions, from mafic to acid, serve as hosts for these pegmatites.

Rb-Sr geochronology by Arriens (1975) has indicated ages of 2500 million years for gneiss and 1400 million years for dykes from the Vestfold Hills. Previously, however, Harding and McLeod (1967) recorded a K-Ar age of 1000 million years for the same dykes. Several major swarms of tholeitic dykes are recognised by Collerson et al. (1983) as having intruded the Vestfold Block in Early and Middle Proterozoic time. Using Sr and Nd isotopic data these authors have documented at least two granulite-facies tectonothermal events of Archaean age (about 2800-3000 million years ago and 2400-2500 million years ago). These periods of granulite facies metamorphism are linked to a complex deformational history which involved 'at least four' (Parker et al. 1983) periods of deformation (see also Oliver et al. 1982).

As a result of this, when mapped on a regional scale, the boundaries parallel characteristic lineaments that are reflected in landscape patterns (see Section 2), especially in the alignment of the major valleys and fjords.

1.3 CLIMATE

The climate of the Vestfold Hills is documented from records and research since 1957 (Burton and Campbell 1980) although the period 1965 to 1968 lacks routine meteorological observations. The records illustrate the seasonal climate to be expected in high latitudes (Figure 2) but on average Davis station (68°35'S, 77°58'E), to which

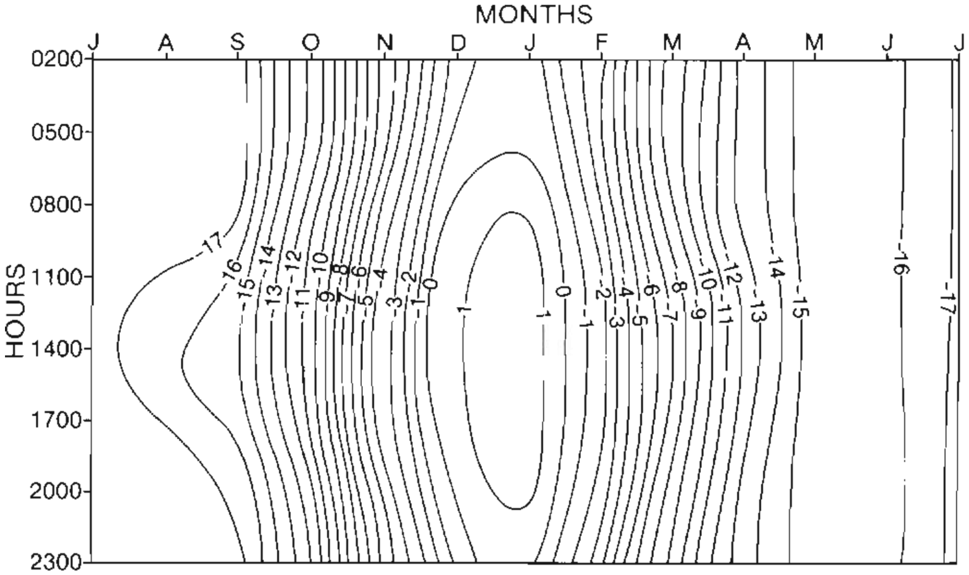


Figure 2. Thermoisopleths for Davis station using data for the periods 1957-1964 and 1969-1981. Isopleths are plotted from three-hourly observations.

the records relate, is warmer than other Antarctic stations at similar latitudes. Burton and Campbell attribute this to the 'rocky oasis' effect whereby albedo is lower and more solar energy is retained than for stations in closer proximity to the ice cap.

One of the most significant characteristics of the climate from a geomorphological point of view is that such climates promote seasonal fluctuation of soil temperature about the freezing point for pure water which brings ground freezing during winter and thaws during summer. Beneath this 'active layer' permafrost may occur. Thus seasonal ground heaving and frost sorting may take place in the regolith. During the summer there may be diurnal variation about the freezing point which may promote further frost sorting. In alpine periglacial climates (where permafrost is absent or discontinuous) there is a higher frequency of temperature oscillations about the freezing point than in high latitude continental climates. In these terms, the Vestfold Hills record averaging about 47 air-frost cycles per year (Figure 3) falls between that of other periglacial climates such as Yakutsk (continental location, 62°N) with 42 freeze/thaw days per year, and Spitsbergen (maritime location, 78°N), which has 63 freeze/thaw days per year (French 1976). However, explanation for the distribution of permafrost and seasonal active layers in the Vestfold Hills is complicated by the presence of saline ground water and so climatic indices alone are not a measure of the overall significance of periglacial processes (see Section 4.6).

Other geomorphologically relevant climatic elements include:

(a) mean temperatures (Burton and Campbell 1980, their Figure 3)

(b) extreme temperatures (Table 1)

(c) wind speed and direction (see Burton and Campbell 1980, Figures 7-10; Pickard 1982; and Zhang 1983 p. 480)

(d) precipitation and drifting snow (Burton and Campbell 1980 pp. 9 and 14; and Zhang 1983 p. 480)

(e) radiation and cloud cover (Burton and Campbell 1980 p. 3 and Appendix I)

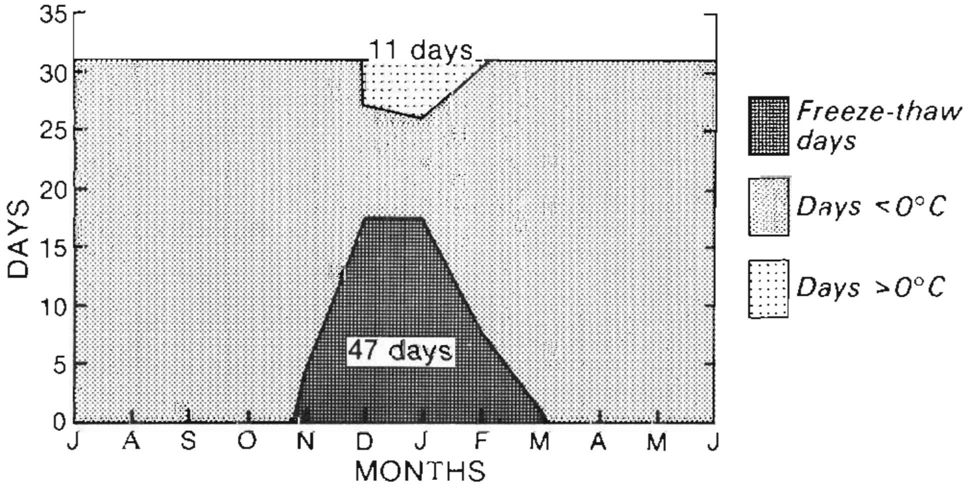


Figure 3. Average annual frequency and distribution of air-frost cycles at Davis station based on station records from the periods 1957-1964 and 1969-1981.

	MAXIMUM TEMPERATURE			MINIMUM TEMPERATURE		
YEAR	MONTH	DAY	°C	MONTH	DAY	°C
1957	12	19	6.7	5	20	−31.7
1958	12	11	6.7	7	27	−33.3
1959	12	23	7.2	9	05	−34.4
1960	12	19	6.1	7	04	−37.8
1961	1	15	6.1	7	28	−28.9
1962	12	06	6.7	8	02	−36.7
1963	12	29	9.4	5	06	−32.2
1964	1	29	6.7	5	23	−38.3
1969	12	08	7.0	8	19	−35.0
1970	12	29	7.3	7	20	−32.2
1971	1	20	9.3	7	21	−32.2
1972	1	05	10.0	8	04	−31.7
1973	1	14	8.7	7	10	−32.0
1974	1	07	13.0	5	23	−29.1
1975	1	17,18	10.0	9	14	−35.0
1976	12	18	10.0	7	18	−32.0
1977	1	03,05,26	10.0	7	03	−36.0
1978	12	26	11.0	8	11	−34.0
1979	12	16	10.0	7	28	−39.0
1980	12	21	10.0	6	25	−34.0
1981	1	14	6.0	4	24	−40.0

Table 1. Yearly extremes of maximum and minimum temperature (°C) at Davis.

2. Structural geomorphology

The influence of geological structure on landscape evolution is most obvious in active neotectonic regions despite the effects of erosion and deposition. But, as Wang (1980) has recently pointed out with reference to Chinese examples, some structural control of landscape evolution may be retained long after tectonic activity has ceased and most of the original tectonic relief has been subdued.

All structural landforms recognised in the Vestfold Hills have been to some extent degraded because of their position on part of the stable east Antarctic craton. There is no evidence for geomorphologically significant earth movement during the last 1000 million years or so apart from glacio-isostatic movement during the late Quaternary.

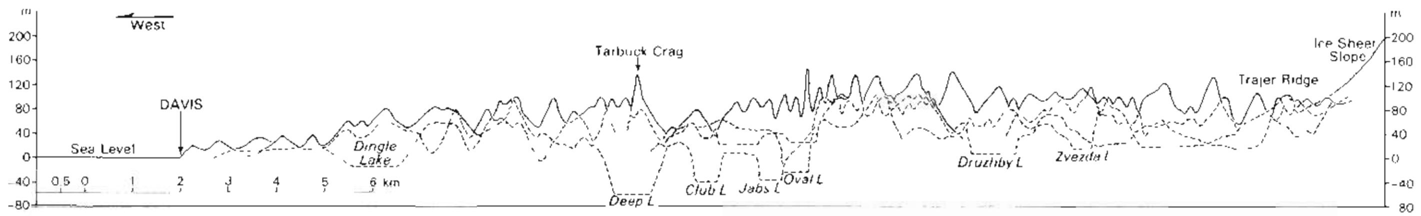
East Antarctica, including the Vestfold Hills, has been part of a supercontinent formed on a basement of Archaean rocks which themselves show evidence of having been partly formed from the products of weathering and deposition during a long history of cyclic denudation and orogenesis. The supercontinent, Gondwanaland, began to breakup during the Mesozoic and had begun the process of parting with Australia during the first quarter of the Cainozoic (Owen 1976). The last quarter of the Cainozoic brought the onset of glacial conditions that culminated in the Quaternary ice age. Apart from this time, and possibly that of earlier ice ages, the Vestfold Hills area for most of its history was directly or indirectly under the influence of non-cryogenic continental denudation. Palaeontological evidence suggests that vegetated Antarctic landscapes (e.g. those with Mesozoic and Tertiary cool-temperate beech forests) predominated (Truswell 1982). Thus lineaments that today control the alignment of valleys and major coastal inlets have persisted over a long period during which integrated drainage patterns appear, in some parts of Gondwanaland (including Antarctica) at least (Johnstone et al. 1973b), to have endured long enough to reduce large areas to form planation surfaces (*sensu* Adams 1975 p. 449).

2.1 PLANATION SURFACES

Crohn (1959) observes that, 'Some evidence that two old erosion surfaces are present in this area is afforded by the summits of the highest hills at about 400 feet above sea level and the tops of minor ridges at about 100 feet above sea level' (Crohn 1959, p. 43, and his Plate 11).

The two levels of 'old erosion surfaces' occur widely in the Vestfold Hills. The higher ground surface consists of smooth and flat-topped summits between 100 and 120 m asl and exhibits a gentle east to west decline from the edge of the ice cap to the coast (Figure 4 and Plate 1). This 'old erosion surface' was dissected by many valleys which are now floored by drift deposits of various depths and/or bedrock debris with some larger valleys containing fossiliferous marine sediments as relicts of times when they were arms of the sea.

The lower ground surface comprises gentle ridges of hills ranging 40-60 m asl dissected by a complex valley system.



PROJECTED PROFILES [WEST-EAST] FROM BROAD PENINSULA, VESTFOLD HILLS

Figure 4. Projected profiles, west to east, from Broad Peninsula. Accordant levels suggest the presence of planation surface remnants.

The 'old erosion surfaces' pre-date the Cainozoic glaciation which modified them so that no evidence, e.g. deep weathering mantles, remains from which their origin can be deduced in greater detail. They have survived inundation (Law 1959; Blandford 1975) to a depth of at least 160 m (Zhang et al. 1983) and probably more if the detailed history of other ice-free areas is any indication (e.g. Robin 1977; Hughes 1981).

There is the suggestion of a third and higher erosion surface in the accordance of a few of the highest summits, e.g. Stalker Hill and Boulder Hill (158 m asl). Establishing the overall slope of the former peneplain is rendered difficult by the effects of glacial isostasy, any differential in the nature of which has yet to be documented fully.

2.2 LINEAR SCARPS

Linear scarps controlled by foliation and/or fault lines occur widely in the Vestfold Hills (Figure 5 and Plate 2). The east-north-easterly trending fjords and linear basins conform with the regional foliation and in places they are congruent with faults of the same orientation. On the northern sides of those fjords scarps are up to several kilometres long with up to 100 m relief. For instance, inland from the head of Long Fjord, Luncke Ridge marks a fault line scarp which is identified by the silicified breccia along the steep, southward dipping fault plane (Plate 3). Again, an 8 km fault line scarp runs between Ellis Fjord near Lake Druzhby. It ranges between 60 m and 100 m in relief and is apparently controlled by a clockwise shearing fault zone which is evidenced by a horizontal displacement of about 250 m at about 2 km east-north-east of Ellis Rapids where two north-westerly trending dolerite dykes (dipping 70° to the north-east) were sheared. Patterns similar to those formed by these features can be identified in other places on aerial photographs. For example, another fault line scarp of the same character, running from the east end of Lake Druzhby to the margin of the ice sheet, forms Trajer Ridge on one side and a valley on the other. Those two fault zones define an *en echelon* fault system on the eastward elongation of the Ellis Fjord fracture zone. Other east-north-east trending scarps, e.g. from between the west end of Lake Cowan and Stalker Hill to the northern side of Stinear Lake, coincide with the strike of the foliation and lithological boundaries.

2.3 LINEAR VALLEYS

Linear valleys are more common and morphologically more pervasive than the linear scarps. Apart from the east-north-east trending fjords and linear basins there are two other groups of linear valleys. These are situated in fault zones. In one group the valleys trend north-east, in the other north-west.

Two north-easterly trending linear valley systems are apparent. Firstly there is one occupied by Collerson Lake, Club Lake and Oval Lake and extending across Long Fjord to Pioneer Crossing to total about 14 km in length. The second one features Tassie Lake, Triple Lake and Ekho Lake, before crossing Long Fjord to Tryne Crossing, where, connecting with the Tryne Fjord fault zone, it totals 15 km or more in length.

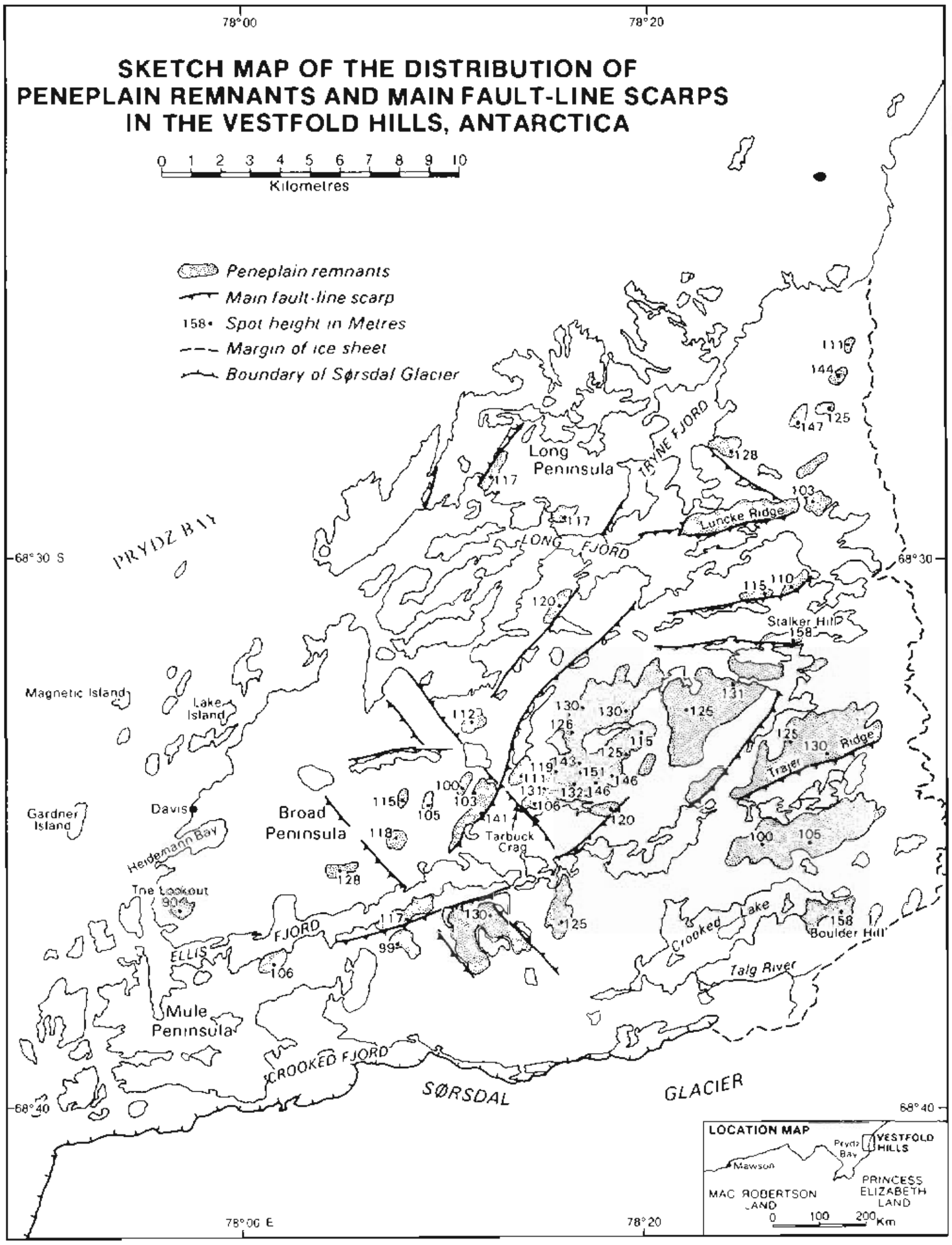


Figure 5. Sketch map showing the distribution of fault line scarps and erosion surface remnants, Vestfold Hills.

Within those two zones, fault line scarps of more than 100 m (e.g. Tarbuck Crag and Lied Bluff) and linear basins of several kilometres in length are found with many such features of smaller size over the rest of the area, especially on Long Peninsula and Broad Peninsula.

Evidence indicates that many linear valleys used to be arms of the sea during the Middle Holocene (Part B).

Further structural control is apparent in the north-west trending linear valleys of comparatively smaller size and in relatively lower relief. The most obvious and representative of this type of feature are from the mouth of Weddell Arm via Deep Lake to the east of Tarbuck Crag, and Lichen Valley (about 6 km and 3.5 km long respectively).

2.4 ASYMMETRIC VALLEYS

Many structural (fault line or foliation) valleys, especially those trending east-north-east, have asymmetrical cross sections. The northern side of the valley bounded by Trajer Ridge, for example, is formed by a very steep cliff (relief more than 100 m) while the slopes along the southern side are without cliffs and are much lower and gentler (Plate 2). Such asymmetry may be found in both north-east and north-west trending valleys.

Interpretation of valley asymmetry should be made with great caution (Washburn 1973 p. 214) because many factors may be involved in a complex landscape history as in the Vestfold Hills. Several denudation cycles may have preceded the Quaternary glaciation which would have exhumed structural features so that linear scarps and valleys appeared again. The character of relief mainly depends on the essential features of structure itself (dip, of fault or foliation, and the lithology either side of the faults). Valley asymmetry can arise from these influences alone. Thus relief is rugged, although not universally because, for example, the southern and central portions of Mule Peninsula have gentle hills and shallow pools.

After ice retreat very active periglaciation modified the landscape, as it does at present, and valley asymmetry was probably emphasised. This would seem to apply especially to the valleys trending east-north-east.

However, the post-glacier period has not been sufficiently long in the Vestfold Hills for valley asymmetry to be entirely attributable to freeze/thaw processes. Severity varies with aspect as it does in some arctic areas where periglacial conditions obtained during both glacial and interglacial times.

Therefore the asymmetric valleys are not produced by periglacial or structural influences alone but by the combined effects of structure, glaciation and periglaciation. In essence they reflect the pervading influence of geological structures.

2.5 SUMMARY

The landforms of the Vestfold Hills reflect the importance of geological structure over long periods and the influence of cold climates during the Quaternary. The planation surfaces probably formed well before the Cainozoic ice sheet covered the continent (including the present ice-free area) so that structurally determined features were exhumed from beneath continental-scale weathering mantles.

Present structural landforms were mainly sculptured by glacial erosion along zones of weakness (e.g. fault line and foliation trends) and reworked by post-glacier periglaciation. Three sets of fault lines trending east-north-east, north-east, and north-west respectively are now revealed as linear scarps and valleys. Relicts of the peneplains are still preserved on the summits of hills. The structurally controlled features are not derived from neotectonic movement but date back to the formation of the Archaean rocks which received the final imprint in a long series of petrogenetic and tectonic events as long ago as the late Proterozoic (Collerson et al. 1983).

3. Glacial geomorphology

Products of glacial erosion and glacial deposition occur in the Vestfold Hills. Former ice inundation is demonstrated by the often recognised presence of rock pavement (bearing glacial striae), numerous erratics, hillocks in the form of roches moutonnées, and extensive deposits of glacial drift and moraine (Law 1959; Blandford 1975; Johnstone et al. 1973a; Zhang et al. 1983). During glaciation the minimum ice thickness was 158 m over the Vestfold Hills because the whole area was covered. Thus most evidence of previous Quaternary ice-free landscapes has been removed so that glacio-erosional forms date mainly from the period of the last major ice advance. Zhang et al.(1983) and Pickard and Adamson (1983) date this event as lasting from about 30 000 BP until about 10 000 BP (see Part B). Glacial deposits include till dating from these times as well as younger ice-cored moraines near the present ice edge dating from about 3000-2000 BP (Adamson and Pickard 1983; Zhang et al. 1983).

3.1 LANDFORMS DUE TO GLACIAL EROSION

The Vestfold Hills are characterised by a high percentage of bedrock outcrop and this is attributed to the removal of weathered mantles by successive ice advances during the Quaternary ice age. Post-glacier periglacial activity has resulted in the initiation of frost shattered regolith, among other things. Even so, small scale ice erosion features (e.g. parallel groovings, and mammilated surfaces including roches moutonnées) are well preserved in many places as are the larger scale evidences of glacial erosion referred to in the discussion of structural geomorphology. In some places streamline features (e.g. striae, roches moutonnées and crag and tail, Price 1973 p. 58) indicate the direction of ice movements.

3.1.1 Parallel grooves

Well-developed parallel grooves stretching north-west, tens of metres wide, and from between 100 to 1000m long, occur in the south-east part of the Vestfold Hills adjacent to the ice sheet. An example is located between Lake Chelnok and Crooked Lake (Plate 4) where north-westerly trending parallel grooves cut across the structure of the country rock. In cross profile they are U-shaped with rounded and polished crests and range from 20 m to 50 m in relief.

3.1.2 Depressions and irregular knobs

In some areas erosion appears to have been more selective and an irregular surface featuring depressions and rock knobs has developed.

There are more than 700 lakes in the Vestfold Hills. Most of them are small (less than 0.01 km^2) and shallow (less than 3 m in depth). Some are floored by ablation till with erratics, boulders and other debris while others have a bare rocky bottom. The many smaller lakes are also glacially eroded as indicated by the abraded bedrock around them. Their orientation is not always coincident with that of local striations (mostly north-westerly) but reflects structural control with their elongation in east-west, north-east and north-west directions. This illustrates again that in the case of erosion by ice caps structurally determined zones of weakness will find expression whatever their orientation in relation to ice movement direction (e.g. Jennings and Ahmad 1957).

Lithology and structure has also controlled the form of the hills and rocky knobs. Some larger eminences, Stalker Hill (158 m asl), Lied Bluff (125 m asl), Tarbuck Crag (141 m asl) and The Lookout (90 m asl) for example, lack the gentle stoss and steep lee slopes exhibited by less structurally dominated glacially eroded landforms. On Tarbuck Crag the striations stretch north-west across the axis (north-east) of the hill itself. Thus the direction of ice movement is not immediately obvious from the landscape lineations (Embleton and King 1968) and has to be deduced from the study of striations and erratic trains.

3.1.3 Roches moutonnées

Roches moutonnées are most prominent in the Talg River and in the vicinity of Crooked Lake and Zvezda Lake. They appear in groups in the Talg River and range from 5 m to 20 m high and from tens to hundreds of metres in length (Plates 5 and 6). Generally, they are smoothed on the stoss side and are broken and shattered into irregular surfaces on the lee side. Their east-west axial orientations roughly correspond to the flow direction of the Søvrsdal Glacier (Zhang, in press) during Middle to Late Holocene advance that covered the Talg River valley. However, many roches moutonnées in this valley have asymmetric forms. Long cross profiles with steep southern sides and gentle northern sides are controlled by the foliation. However, where the bedrock dips away from the direction of ice advance, 'reversed' roches moutonnées with steep stoss and gentle lee slopes have formed in the vicinity of Crooked Lake and Zvezda Lake (Plate 7). Demorest (1939) regarded such features as 'pre-glacial structures which have successfully resisted the last abrasive action of the ice while the area immediately surrounding them has been cut lower.'

3.2 LANDFORMS DUE TO GLACIAL DEPOSITION

In the Vestfold Hills glacial deposits can be subdivided into two morphological groups. The first group, characterised by very little local relief, is best described as till-plain or till-sheet (Price 1973 p. 67). An obvious example is Marine Plain (4.5 km^2) bordering the northern side and 20 m above the end of Crooked Fjord (Plate 8). Field evidence indicates that up to 2 m of glacial drift overlies marine deposits of an unknown thickness and dating from at least 31 000 years BP (Zhang et al. 1983). The glacial drift post-dates this marine phase and probably was laid down in the last glacial phase as an ablation till during ice cap retreat. Thin fossiliferous marine deposits partly masking the drift are dated about 7000 BP. They testify to marine transgression in the early post-glacier period.

The second group of glacial deposits displays very irregular surface forms including simple linear ridges, or complexes of ridges, hollows and mounds (as typified by Price

1973 p. 67). Actively forming ice-cored moraines near the land-based ice edge testify to glacial thinning, retreat and freezing of glacial ice to the bedrock beneath. One simple linear ridge can be clearly seen broadly paralleling the ice edge and traces of similar ridges also occur. The ice margin immediately east of the Vestfold Hills is associated with ice-cored moraine in a zone 5-20 m high, hundreds of metres wide, and 20 km long.

The moraine is only a few hundred metres from the ice edge and, together with those along the edge of the Sørsdal Glacier, must be among the youngest local glacial landforms described here. These easternmost ridges are exposed in summer but always at least partly buried in snow during winter (Plate 9).

The evolution of surface drift forms is well illustrated in a zone parallel to and including the land-based margins of the Sørsdal Glacier. Here a series of three ice-cored shear moraine complexes in various stages of modification by core disintegration testify to retreat from Middle to Late Holocene advances that may be equated with the Late Holocene advances recognised in glaciated regions outside Antarctica. The innermost and youngest ice-cored shear moraine is actively forming and consists of a low ridge just inside the Sørsdal Glacier. It is at Lake Chelnok that the ice edge is closest to the oldest remnants of shear moraines. They are part of a line of isolated hummocks 10-25 m high and tens of metres wide and 10-100 m long that can be traced along the Talg River valley as far east as 1 km north of a prominent ice-cored debris pile of irregular surface form known locally as the 'Garbage Dump' (Plate 10). This debris represents a still-stand younger than that from which the hummocks developed and older than those currently forming, and may represent a separate glacier margin change. The morphological contrast between the 'Garbage Dump' and the legacy of still-stands either side is quite obvious. It is not a ridge but a large (2.5 km x 1 km) elliptical hummock with its long axis aligned east-west parallel to the ice front. The surface form is complex due to non-uniform disintegration of the ice core. Nevertheless sufficient remains of a series of parallel ridges to indicate that, like the other two ridges, this ice margin deposit originated as ice-cored shear moraine formed by the transport of debris to the glacier surface along shear zones. These shear zones were formed by active ice behind a glacier margin that, being too thin to promote basal pressure melting, froze to the bed (e.g. see Goldthwait 1951). Thawing and thermokarst formation has reworked the original surface. Many pools, lakes and valleys are formed (Plate 11). The process continues and active thawing can be seen during the summer.

The moraine ridge, on the northern side of Sørsdal Glacier from Lake Chelnok westward to the end of Crooked Fjord, ranging from 10 to 50 m high and tens of metres to hundreds of metres wide, is continuous for some 10 km. It appears as either a simple or complex ice-cored moraine ridge (Plate 12). Three lagoons have been dammed by this ridge at the end of the Crooked Fjord but the morainic dams have been broken down after ice retreat during the last 1000-2000 years (Plate 13).

In some places, the south-east corner of the present ice-free area for instance, many arch-like terminal ice-cored moraine ridges 40-50 m high and 100-200 m wide have dammed valleys (Plate 14). Behind them small glacial lakes or pools have appeared nourished by melt-water from dead glacial ice and drifting snow. It is supposed that these terminal ice-cored moraine ridges were formed during a period of neoglaciation during the period of ice margin fluctuations during the Middle to Late Holocene when the ice sheet expanded into the valleys. The subsequent ice retreat left debris marking the former ice front which is now seen as terminal ice-cored moraine ridges about 0.5 to 1.5 km from the present ice edge. The youngest ice-cored shear moraine is, as described previously, closest to the ice edge and represents the most recent ice margin advance.

3.3 SUMMARY

In the Vestfold Hills a variety of erosional and depositional glacial landforms provide evidence not only of the power of glaciers to modify landscapes but also of the sequence of events that marks the interaction between glacial and other influences during the late Quaternary.

Apart from Marine Plain, depositional glacial landforms of the late Pleistocene were almost completely destroyed or reworked, mainly by periglaciation, during the Holocene. Erratics and boulders still remain on the tops of hills and fringe islands in the ice-free area. Despite the important influence of geological structure on the pattern of ice erosion striations, roches moutonnées and parallel grooves coincidentally indicate that ice flow was north-westerly from the ice plateau to the coast during the last (Pleistocene) glaciation. The nature and position of the glacial drift overlying marine deposits that pre-date the latest glacial stage suggest that during the retreat of the last Pleistocene ice cover ablation tills accumulated on frozen marine deposits.

Some isolated moraine ridges and ice-cored moraine ridges regularly occur 0.1 to 3.0 km in the front of the present ice sheet and the Sørsdal Glacier. The youngest ice-cored moraines are actively forming on the margin of the present ice sheet. It is suggested that the marginal ice froze to the bed so that active ice formed shear zones. This probably occurred during advances coincident with the well known advances during the second half of the Holocene, sometimes referred to as 'Neoglaciation' (Denton and Porter 1970).

4. Periglacial landforms

Many factors determine the nature of particular periglacial environments but the overriding controls are regional climate and topography. Local factors may modify the regional climate and strongly influence cryogenic processes (Washburn 1979). Washburn recognised that the zonation of periglacial processes is demonstrably dependent on the presence or absence of permafrost. From climatic data collected over the last thirty years or so it is possible to classify the Vestfold Hills' climate in Tricart's category A — 'cold dry climate with severe winters' (Tricart 1969). Characteristics include very low temperatures, short summers, permafrost, low precipitation, and violent winds. Consequently there is intense freezing, reduced or even negligible activity of running water, and important wind action (Washburn 1979 p. 6).

Periglacial landforms in the Vestfold Hills are less developed than in similarly cold areas that have been ice-free during both glacial and interglacial stages of the present ice age. Explanations for this are offered in Section 4.6. Twelve types of periglacial landforms due to frost wedging, frost heaving, wind action and thawing have been found (Zhang 1983) (Figure 6).

4.1 LANDFORMS DUE TO FROST WEDGING

4.1.1 Block slopes

Block slopes (e.g. Washburn 1979, his Figure 4.7) are confined to steeper (25°-40°) slopes on fractured gneiss (Plate 15). In some cases the block slope includes erratics from up-slope. The blocky mantles are without interstitial fines or ice, and harbour neither long lying snow drifts nor summer melt-water. They show no sign of contemporary mass movement and probably accumulated by rock fall from up-slope and possibly aided by minor frost creep.

4.1.2 Summits

Frost shattered bedrock summits carry accumulations of blocks, frost wedged *in situ* to resemble single cycle tors. Some stand 1 m or so above the general surface, near Mossy Lake for instance, and between Crooked Lake and Lake Druzhby (Plate 16).

4.2 LANDFORMS DUE TO FROST HEAVING

4.2.1 Patterned ground

Six varieties of patterned ground have been found. They are widely distributed on the eastern part of Mule Peninsula close to the ice plateau and in some areas adjacent to fresh water lakes on Broad Peninsula.

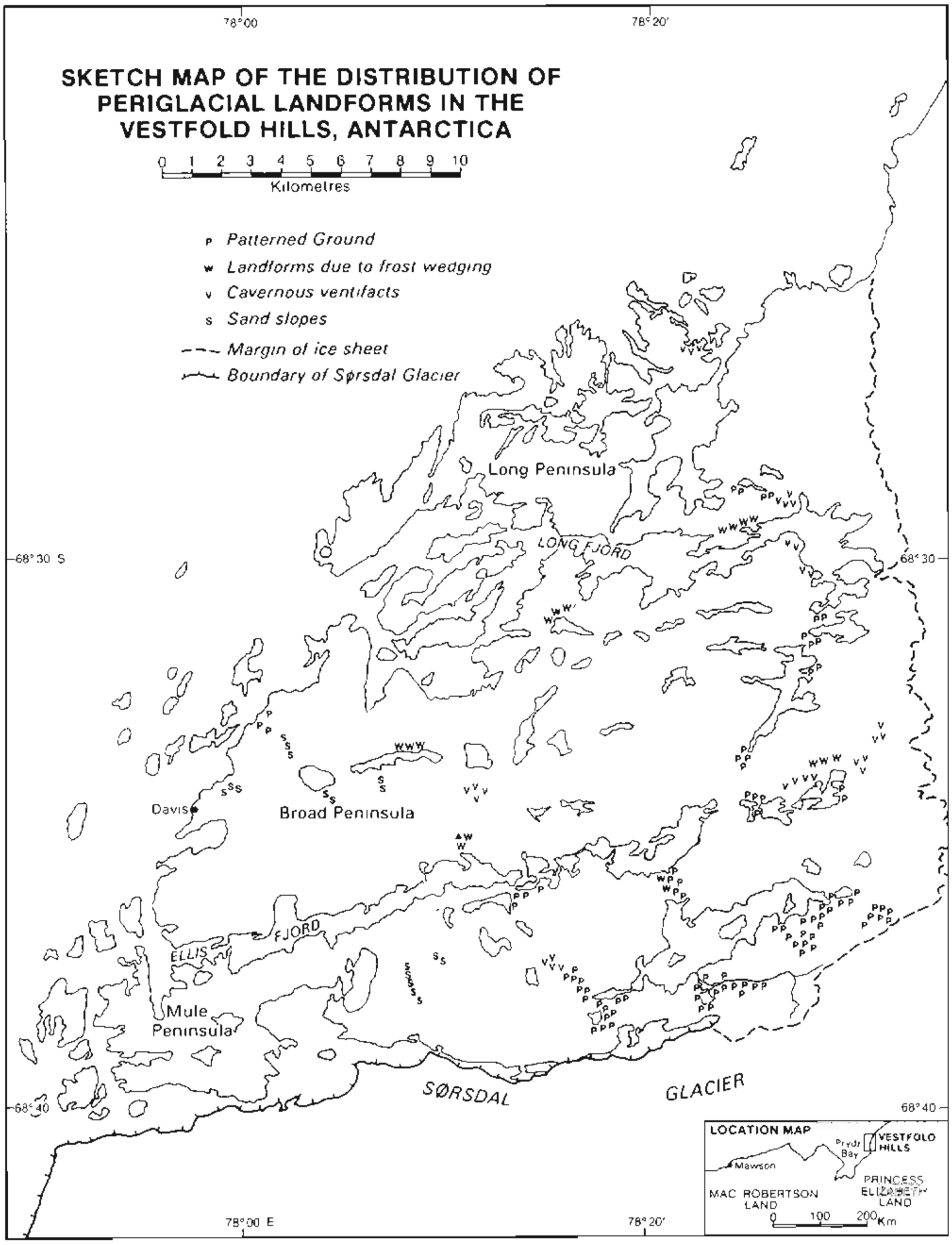


Figure 6. Sketch map of the distribution of periglacial landforms in the Vestfold Hills.

4.2.2 Large sorted circles

These are 2-6 m in diameter and are found where surface materials are seasonally saturated by summer melt-water such as on the floors of broad valleys or on beaches of gentle slope (2-3 °) around fresh water lakes. Sand has been blown off the surface of these features but comprises 5-35% by weight of the material beneath the surface within the stony borders. Grain size generally decreases with circle diameter. Well-developed large sorted circles contain secondary polygons 5-15 cm in diameter (Plate 17). These features are actively forming.

4.2.3 Small sorted circles

These are 0.3-2.0 m in diameter and occur on the young moraine deposits in the southern Vestfold Hills and therefore are probably younger than the large sorted circles. They are composed of coarser materials which probably inhibit the rate of development due to high permeability and low porosity.

4.2.4 Sorted polygons

These are rare and occur in low lying areas on the edge of the Talg River bed where there is an abundance of silt and sand. They have irregular stony borders and are usually saturated, often flooded, during summer. Surfaces may carry desiccation cracks forming small non-sorted polygons.

4.2.5 Large sorted stripes

These occur on gentle slopes (5 °-10 °) on the lee side of hills below winter snow drifts (Plate 18). Stony stripes are up to several metres apart and tens of metres long. Their distribution appears to be controlled by availability of melt-water from up-slope snow drifts and a plentiful supply of frost shattered stones mantling slopes of suitable gradient.

4.2.6 Debris islands

These are known from the Talg River bed (Plate 19). They have irregular rocky borders surrounding non-sorted muddy islands often with a block in the centre. They differ from the type of examples quoted elsewhere (e.g. Washburn 1979) and may represent the development of these features on flatter ground where moisture for ground ice is supplied from a river bed rather than melt-water run off on valley sides. During the summer the fines are subject to diurnal freezing and thawing. These features are actively forming and probably comparatively young because the Talg River bed is among the most recently deglaciated parts of the Vestfold Hills (Zhang et al. 1983).

4.2.7 Stone pavements

These occur in locations similar to those in which sorted stripes are found but the two are not found together. Pavements were noted at 'Airport Beach', Heidemann

Valley and on some of the terraces surrounding Crooked Lake. The pavements appear to favour locations where stones are flat sided or facetted and there is a deeper active layer formed in regolith with abundant matrix. Salt in the matrix might favour greater and longer mobility in the active layer thus maximising the opportunity for flat sided stones to distribute themselves on the surface.

4.3 LANDFORMS DUE TO WIND ACTION IN THE PERIGLACIAL ENVIRONMENT

4.3.1 Cavernous ventifacts

Outcrops facing east and close to the ice cap north of the Søfrsdal Glacier are in the path of the diurnal katabatic winds which may sustain high velocities for 6 to 8 hours. Such outcrops exhibit well-developed ventifacts which, in the coarse-grained intercalations of the layered paragneiss (see Oliver et al. 1982) at the head of Long Fjord, take on a form reminiscent of cavernous or even honeycomb weathering (Plate 20). The aspect and distribution of outcrops like this suggest that sand, including coarse grades, travels with drifting snow derived from the snow patches that cover dead ice now lying asymmetrically in narrow valleys in the ice-cored moraine at the ice cap edge. Firnification in the snow patch is partly by infiltration congelation and the resultant ice layers are periodically dismembered along with the overlying snow and firn, especially during blizzards. Thus ice particles, as well as sand and snow, contribute to ventifact formation.

Further downwind (i.e. to the west) similar weathering can be found on suitably placed outcrops of orthogneiss where grain size is coarse. Cavernous weathering forms here include hollows up to 75 cm across and 25 cm deep. Dolerite dikes stand out in high relief in both areas and do not exhibit such weathering.

4.3.2 Sand slopes

These are lee side deposits which have accumulated from residual sand left behind after the melting of snow patches that form during the colder months (Plate 21). Despite their south-westerly aspect they receive sufficient solar radiation during the months of higher sun angles for some melting to occur. The orientation and height of the ridges above the sand slopes makes them less efficient 'snow fences' than those associated with the permanent snow patches and also facilitates the receipt of summer solar radiation. The surface of the sand slopes is quite dry but some of the melt-water is refrozen in the lower layers thus accounting for the general stability of these features. The top several centimetres are mobile during summer and may form miniature barchans (less than 0.5 m high) migrating down-slope. Particle size distribution is dominated by the coarse sand fraction with some fine gravel derived from weathering of nearby bedrock. It is estimated that the thickness of these deposits is in all cases less than 10 m.

4.4 LANDFORMS DUE TO THERMOKARST

These occur because of melting in ice-cored moraines and of ground ice in permafrost areas.

4.4.1 Thaw slumps

Thaw slumping is only seen on the 'Garbage Dump'. They can form semicircular hollows opening down-slope, are initiated by exposure and thawing of ice-cored moraine and are excavated by mudflow or gelifluction of the resulting saturated debris during summer (Plate 22).

The head of the slumping slope retreats rapidly. No measurements were taken but an estimated 7-10 m of retreat each summer is reported in the western Canadian Arctic (French and Eggington 1973). Topographical features probably change sufficiently quickly in this area for the difference eventually to be apparent from the study of time-lapse photographs.

4.4.2 Thaw lakes

Lakes are found in the 'Garbage Dump' and on Marine Plain. They are in basins that have formed or enlarged by thawing of frozen ground (Plate 11) in the manner reported from other polar areas (Black 1969; Ferrians et al. 1969; Hussey and Michelson 1966; Sellmann 1975). They are thus a kind of thermokarst feature and are sometimes referred to as thermokarst lakes (Klassen 1979). Washburn (1979) has suggested that lakes caused by thawing alone rarely exceed 3 m in depth. The lakes both in the 'Garbage Dump' area and on Marine Plain are very shallow (estimated depth 2-3 m). The diameter of the lakes varies from tens to hundreds of metres. Washburn (1979) has noted the difference, both in origin and environmental implications, between lakes in kettleholes formed by melting of buried glacial ice (glacial thermokarst) and lakes in thaw depressions caused by melting of other ground ice. The thermokarst lakes in the 'Garbage Dump' are glacial thermokarst and those on Marine Plain are due to melting of permafrost. Although there are fundamental similarities in formative process the inherently more unstable environment of the 'Garbage Dump' area will probably ensure that shorter histories and more rapid changes will characterise the lakes there compared with those on the Marine Plain.

4.5 POSSIBLE PINGO FORMATION

Pickard (1983) reported dome-like features in the Pelite Lake area near the 'Garbage Dump' and describes them as pingos. No examination has been made to confirm that ice cores occur under the apparently dilational cracks, and because the distribution of shear planes in the ice marginal zone there have not been discussed, the present authors cannot confirm the presence of pingos.

4.6 MEASUREMENT OF ACTIVITY OF SORTED CIRCLES

Measurement of seasonal changes in patterned ground was used to establish the degree of activity. Eight sets of steel rods (120 rods each 2 cm diameter, 45 cm long) were driven vertically into the active layer, in the centres and on the borders of large sorted circles formed in low lying flat ground near Mossy Lake. Horizontal displacements were monitored by Zhang between February and December 1981 and measurements have since been continued by ANARE personnel.

Distinct seasonal activity was evident in the behaviour of the large sorted circles under observation. Expansion took place in February and March with the heaving on the permafrost, and contraction occured during the annual active layer development in November and January. Over the winter the circles were stable. Over two seasons mean extension varied between 3 mm and 9 mm (Table 2).

Variation in annual rates of expansion of the large sorted circles is indicated by distinct differences between respective radii. The limited data suggest that these differences correlate with the size of sorted circles. Expansion rates fall exponentially with increasing size. The most important factor influencing the annual growth rate is the net contraction of the width of the borders (wedges), the wider of which appear to be subject to lower rates of narrowing (Table 3). Field evidence indicates that smaller sorted circles normally have wider borders. They therefore appear to offer greater scope for narrowing in the process of circle expansion than do the narrow border wedges, tightly filled with pebbles, that define the larger sorted circles.

4.7 DISCUSSION

4.7.1 The uniformity of periglacial types

Compared with the range of active periglacial processes and resultant forms to be found in some other periglacial areas, the suite of features found in the Vestfold Hills lacks diversity.

For example, there are more than 50 types of periglacial landforms engendered by 7 periglacial agents on the Tibetan Plateau of China (Cui 1982). In the Vestfold Hills only 12 types engendered by 4 agents are found and the majority of features are not well-developed. This is a result of limitations imposed by the Vestfold Hills environment. Relatively low mean temperatures coupled with comparatively short periods of freeze/thaw activity and widespread dryness in the regolith conspire to limit the scope for the development of periglacial forms. In addition, saturation of the active layer is in many places due to saline water which inhibits and, close to hypersaline lakes and in deposits that include evaporites, prevents freezing even at temperatures as low as − 19 °C thus eliminating the possibility of periglacial activity.

4.7.2 Age of the periglacial landforms

The maximum age for some of the periglacial features (e.g. patterned ground) is constrained by the age of the deposits in which they have developed. For example,

Site	Diam. (m)	Sand content of interior circle (1%)	Movement* (mm)								
			Feb 81/Mar 81		Mar 81/Nov 81	Nov 81/Jan 82		Jan 82/Feb 82	Feb 82/Mar 82		Mar 82/Aug 82
			Mean	Max.	Mean	Mean	Max.	Mean	Mean	Max.	Mean
Group G-1	3.4	—	+2.5	+ 8	0	−6.4	−11	−2	+ 6.0	+11	0
Group B	2.9	45	+5.0	+16	0	−9.4	−21	0	+ 7.6	+12	0
Group G-2	2.4	40	+9.5	+19	0	−3.0	− 7	+1	+14.0	+21	0

Site	Net Movement (mm) Feb 81-Mar 82	
	Mean	Max.
Group G-1	1.2	6
Group B	3.0	14
Group G-2	10.0	23

*Positive movement is expansion; negative is contraction.

Distance is taken from the centre to the margin of large sorted circles (nominal accuracy 1 mm).

Table 2. Seasonal expansion and contraction and annual growth rate of large sorted circles.

Black and Berg (1963) and Berg and Black (1966) have dated periglacial landforms in Victoria Land as younger than 10 000 years. Large sorted circles and some types of patterned ground may be found on raised beaches and terraces that have been dated as 7000-10 000 years old by Pickard and Adamson (1983) and 5000-6000 years old by Zhang et al. (1983). The ice-cored moraines probably represent Middle to Late Holocene advances (Zhang et al. 1983) so that small sorted circles on their surfaces would be dated no more than 2000-3000 years.

On those moraines still characterised by substantial ice cores much younger ages are to be expected. The debris islands in the Talg River must have been formed during this interval.

Site	Width (mm)	Net Contraction (mm)*		Duration
		Mean	Maximum	
Within Group A	4.5	−26	−29	February 1981-March 1982
Surrounding Group G-1	1.2	−14	−19	February 1981-March 1982

*Negative movement is contraction. Distance is between pegs set on the borders of wedges measured monthly (nominal accuracy 1 mm).

Table 3. Net contraction of wedges between circles.

5. Coastal beaches and lake terraces

5.1 CONTINENTAL SHELF

The sea bed morphology of Princess Elizabeth Land is little known. However, a large continental shelf about 400 m deep is depicted on the IOC GEBCO chart of 1960 (Figure 7). The zone within the 200 m isobath is only tens of kilometres wide, being a platform-like part of the continental shelf on which islands are found in a belt within 5km of the present coast. Further north of the 'platform', the continental shelf descends gently to about 67 °S. The continental slope begins at about this latitude plunging steeply northward to a depth of some 3000 m.

Although the scale of the GEBCO chart does not allow depiction of detailed sea bed morphology the pattern of isobaths suggests that the structural trends (east-north-east, north-east, and north-west trending lineaments) that dominate the geomorphology of the Vestfold Hills are also important determinants of the gross morphology immediately offshore (Figure 7). Geophysical data from the Søgrsdal Glacier area interpreted as showing that Crooked Fjord extends to 750 m below sea level (Wellman and Williams 1982) confirms the importance of the east-north-east

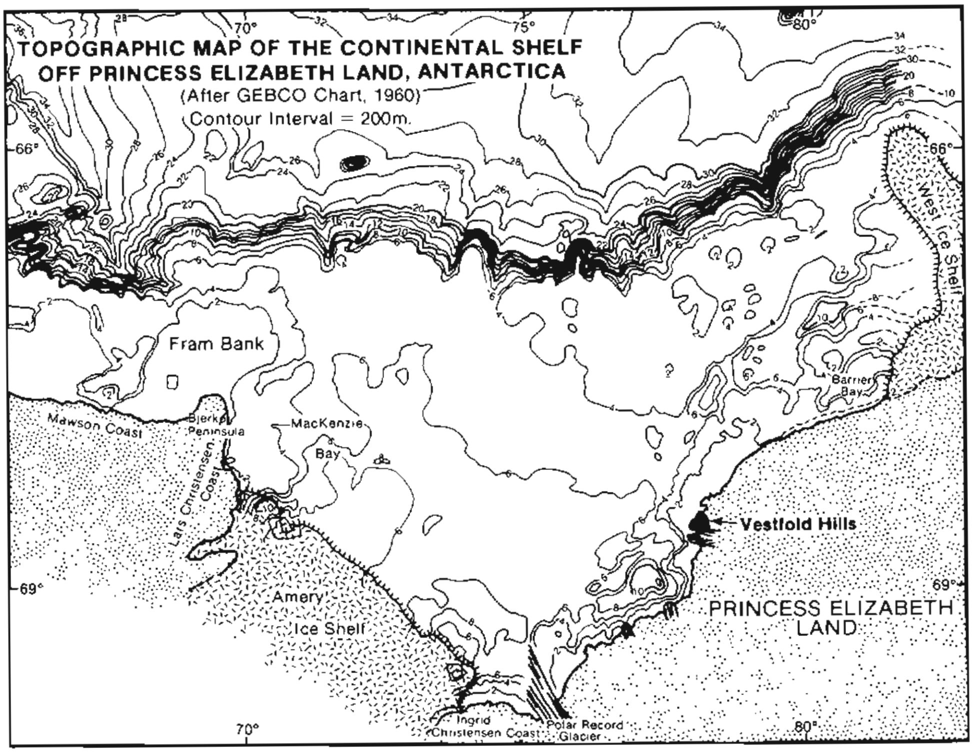


Figure 7. Bathymetry of the continental shelf in the Prydz Bay area.

trending lineaments. Thus Crooked Fjord (and probably other fjords) marks a trough that lies across the innermost continental shelf and reaches well into the Vestfold Hills area. Pleistocene glaciation has affected the whole of the present ice-free area and an unknown proportion of the continental shelf. The legacy of these glaciations lies not only in the modification of topography but also in the effects of deglaciation such as eustatic and glacio-isostatic changes in sea level.

5.2 COASTAL BEACHES

With a few exceptions, bedrock outcrops form the open sea coast on the Vestfold Hills' coast and offshore islands. The orientation of many segments of the coastline, and especially of the coastal inlets, reflects the pervading influence of geological structure in the predominance of north-westerly, east-north-easterly, and north-easterly trends. The latter is, in addition, the overall trend of the Vestfold Hills coastline.

Bedrock slopes on the modern coast are commonly steeply inclined (more than 20°) and lack regolith. On the more gentle slopes of the coastal plain (e.g. at Heidemann Bay, 'Airport Beach', and the south-east end of Long Fjord) beaches have formed but are always less than 100 m wide (Plate 23). This is expected in low energy environments where ice-free conditions obtain for only 12 to 14 weeks each year. Off Davis fast ice forms in late March and does not break up until early December. Even during the summer pack ice offshore reduces the effect of waves generated by high winds and mitigates the effect of swell. This part of the littoral is therefore classified as low energy coast or 'protected sea' (Davies 1980).

Wave energy reaching the coast is expended across a restricted intertidal zone (the tidal range is about 1.2 m, diurnal regime) that is 'microtidal' like most of the Antarctic coastlines (Davies 1980, John and Sugden 1975).

5.3 RAISED BEACHES AND TERRACES

Raised beaches and terraces, many of which are depositional, fossiliferous and datable, are widely distributed around present and former arms of the sea. The highest terraces (Plates 24 and 25), between 10 and 15 m asl, are rare and show signs of modification by progressive collapse and/or burial under the influence of subaerial processes. Terraces at 3 m and 6 m asl (Plates 26 and 27) are well preserved and clearly indicate a stage of inundation that prevailed before the sea retreated from the major valleys to its present position. Radiocarbon dating (Tables 4 and 5) of marine shells in the 6 m terrace yields ages of about 6000 BP and similar material in the 3 m terrace is about 3500 radiocarbon years old (Zhang et al. 1983). The raised shorelines around the hypersaline lakes of the central part of Broad Peninsula exemplify the sequence and morphology of terraces. Raised beaches 6 m in elevation on Heidemann Bay and 'Airport Beach' extend east-north-easterly and south-easterly to connect with the 6 m terraces of Lake Dingle for which a radiocarbon date of 5600 ± 77 years BP (ZDL79) has been obtained. From this point the terrace continues around Stinear lake (5740 ± 105 years BP SUA1239, Pickard and Adamson 1983), Deep Lake (6632 ± 119 years BP ZDL80) and Club Lake (5300 ± 90 years BP SUA1237, Pickard and

Position	Altitude of lake surface (m)	Height of terraces (m) (measured from lake surface)		Radiocarbon date (years BP) (uncorrected for ^{14}C age of present-day samples)***
		above lake	above sea	
Dingle Lake	−10*	16	6	5600 ± 77 (ZDL79)
Deep Lake	−56*	60	4	6632 ± 118 (ZDL80)
Triple Lake	−16*	19	3	6141 ± 90 (ZDL78)
Watts Lake	−4**	10**	6	6100 ± 108 (ZDL70)
Mud Lake	−2	5	3	3500 ± 86 (ZDL69)
Burton Lake	0	15	15	about 6000
Tryne Fjord	—	—	15	?
Pioneer Crossing	—	—	10	?

*According to the map 'Vestfold Hills' (1:100 000) Division of National Mapping, Australia, 1972.
**According to measurements in January 1983 made by Mr C. Heath.
***e.g. see date ZDL84, Table 5.

Table 4. The height of terraces in the Vestfold Hills.

Laboratory Number	Sampling Site	Material	Radiocarbon date (years BP)*
ZDL66	Mud Lake (68°39'S, 77°57'E) upper part of low terrace (3 m asl)	Fossil shells	3325 ± 103
ZDL69	Mud Lake, lower part of the terrace	Fossil shells	3500 ± 86
ZDL68	Marine Plain (68°38'S, 78°08'E) upper layer of marine sediments	Fossil shells	31 000 ± 474
ZDL70	Watts Lake (68°36'S, 78°12'E) upper part of a terrace (3 m asl)	Fossil shells	6100 ± 108
ZDL71	Watts Lake, top part of the terrace	Fossil algae	3600 ± 95
ZDL78	Triple Lake (68°32'S, 78°12'E) terrace surface (6 m asl)	Fossil shells	6141 ± 90
ZDL79	Dingle Lake (68°34'S, 78°03'E) terrace surface (6 m asl)	Fossil shells	5600 ± 77
ZDL80	Deep Lake (68°33'S, 78°11'E) terrace surface (6 m asl)	Fossil shells	6632 ± 118
ZDL81	A dry basin west of Platcha (68°31'S, 78°28'E) 6 m asl	Fossil algae	5677 ± 94
ZDL84	'Airport Beach' (68°31'S, 78°01'E) surface of sea ice	Modern *Laternula* shell	1312 ± 65
ZDL85	Watts Lake (68°36'S, 78°12'E) high terraces (15 m asl)	Calcareous tufa	7616 ± 104

*Uncorrected for ^{14}C age of present-day samples.

Table 5. Radiocarbon dates from the Vestfold Hills.

Adamson 1983), then through the basin of Oval Lake to the coast of Long Fjord. The +6 m sea also occupied an adjacent valley that links Long Fjord to the east across an area that is now floored by several unnamed lakes and dry basins above which terraces at about 6 m asl (Figure 8) contain shells that yield a radiocarbon date of 5677 ± 94 years BP (ZDL81) (Table 5).

Prominent among the raised terraces on the Mule Peninsula are those around Mud Lake, Laternula Lake and Watts Lake. Shells from a section through the 3 m asl terrace around Mud Lake and Laternula Lake yield radiocarbon dates of 3500 ± 86 years BP (lower horizon ZDL69) and 3325 ± 103 years BP (upper horizon ZDL66). This terrace is continuous from Laternula Inlet through the two lakes to Crooked Fjord and provides the most southwesterly point for detailed surveying to establish the nature of the differential in the sea level curves for various Vestfold Hills locations. Notwithstandingd the lack of a detailed survey such as is available in many arctic localities (e.g. Andrews 1970) the analogy between the glacio-isostatically uplifted terraces of arctic and subarctic shores and the uplifted terraces of the Vestfold Hills is strong.

The fjords and their former arms occupy the bottoms of glacially eroded troughs along which enclosed basins have been formed. Upon regression the basins remain filled with sea water. In those basins where evaporation exceeds inflow of precipitation and melt-water, hypersalinity develops as the lake level falls. The extreme example of this is Deep Lake which has a surface level 56 m below sea level and a salinity of 273 gl^{-1} (Burton and Campbell 1980; Barker 1981; Burton 1981; Burton, personal communication). The highest terrace here is more than 60 m above the present lake surface. Between this level and the present lake surface many terraces testify to the progressive fall in water level due to evaporation since regression (Plate 27).

Investigations elsewhere in Antarctica have indicated that raised beaches form as the sea invades the continental shelf during deglaciation. Available chronology implies that ice retreat was in sympathy with ice cap retreat in the northern hemisphere (Cameron 1964; Nichols 1968; Sugden and John 1973). For example, the grounded ice sheet in the Ross Sea disappeared at about the same time (Denton et al. 1970; Calkies and Nichols 1972) as the Laurentide and Fennoscandian ice sheets retreated to their core areas. As pointed out by John and Sugden (1975) withdrawal of ice cap margins across high latitude continental shelves during the last 13000 years or so has exposed long coastal sections to marine action. The withdrawal occurred at least partly in response to the retreat of the grounding line following eustatically rising seas. In itself the shrinkage of the Antarctic ice cap has been reflected in eustatic sea level rise and in glacio-isostatic uplift.

As a result the beaches formed at the coast following exposure of bedrock during deglaciation are now the highest terraces formed during glacio-isostatically induced regression. Correlation of raised shorelines from one Antarctic locality to another depends on dating because the chronology and spacing of each suite of terraces is dependent on local deglaciation dates and rate of isostatic recovery.

For the Vestfold Hills it is difficult to define the date of ice retreat from the present ice-free area although a minimum date is provided by the age of the 6 m raised beaches. Unfortunately the 15 m shorelines have not yet yielded datable material. The Vestfold Hills coasts are still subject to isostatic rise as indicated by the apparent progressive isolation of Burton Lake from Crooked Fjord, the former lake becoming a lagoon. Rookery Lake on the south-east end of Long Peninsula and the lake on Lake Island

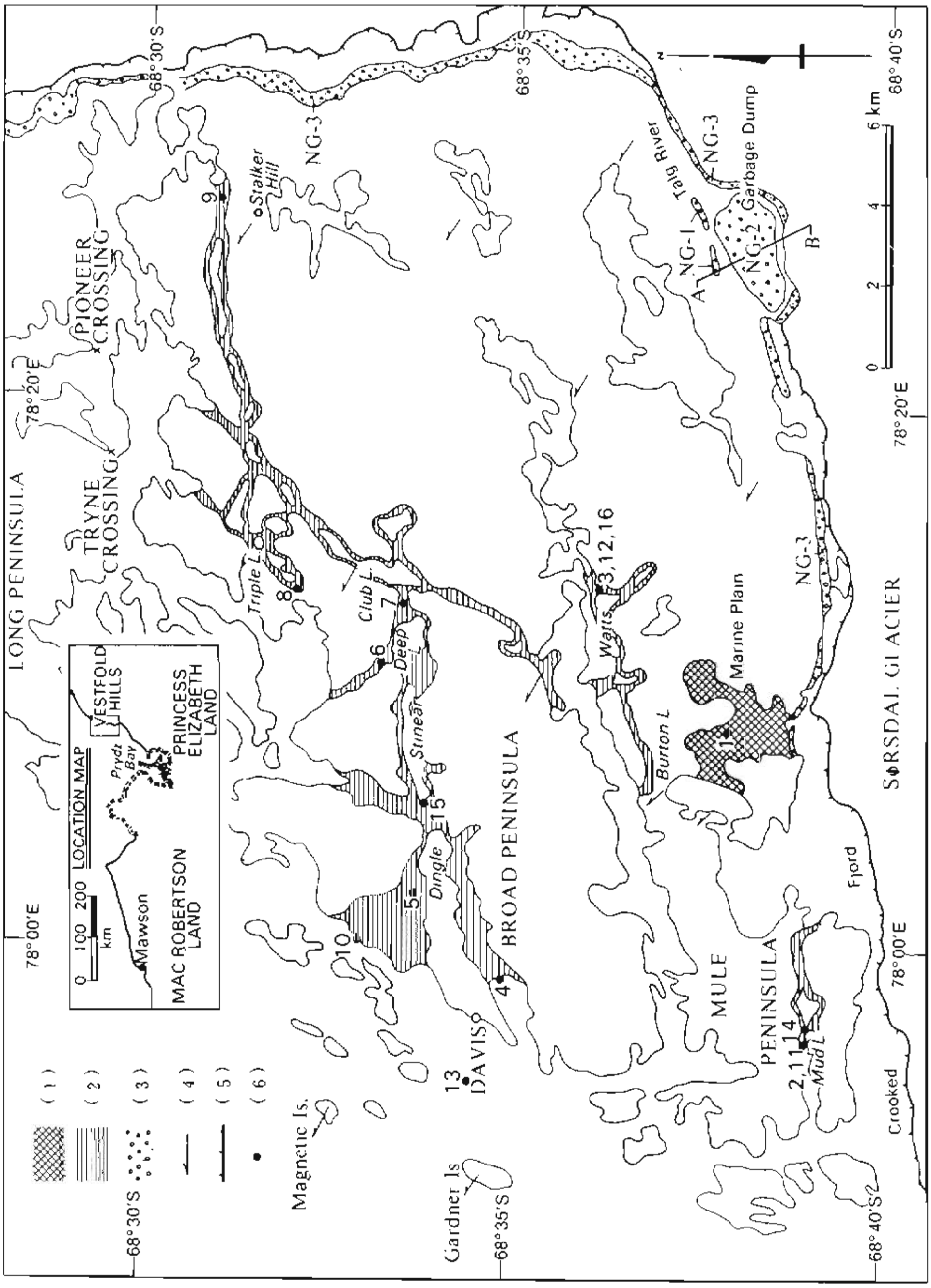

(1)
(2)
(3)
(4)
(5)
(6)
Magnetic Is.
Gardner Is
68°30'S
68°35'S
68°40'S
78°00'E
78°20'E
LOCATION MAP
0 100 200 km
Mawson
MAC ROBERTSON LAND
Prydz Bay
VESTFOLD HILLS
PRINCESS ELIZABETH LAND
LONG PENINSULA
TRYNE CROSSING
PIONEER CROSSING
Triple L.
Club L.
Deep
Stinear
Dingle
Stalker Hill
NG-3
BROAD PENINSULA
Watts
Burton L
Marine Plain
Talg River
NG-1
NG-2
Garbage Dump
A
B
0 2 4 6 km
N
SØRSDAL GLACIER
Fjord
Crooked
MULE PENINSULA
Mud L
DAVIS
1
2,11,14
3,12,16
4
5
6
7
8
9
10
13
15

Figure 8. Distribution of glacial and interstadial deposits in the Vestfold Hills. (1) Marine Plain; (2) Terraces and raised beaches; (3) Neo-glacial deposits; (4) Orientations of striae; (5) Margin of the ice sheet and Søsdal Glacier; (6) Sites of ^{14}*C dates see key to numbers:*

1. *Marine Plain, Shell, ZDL68, 31 000 ± 474 yrs BP.*
2. *Terrace Mud Lake,* **Laternula** *Shell, ZDL69, 3500 ± 86 yrs BP.*
3. *Terrace, Watts Lake, Shell, ZDL70, 6100 ± 108 yrs BP.*
4. *Davis shore, modern* **Laternula** *Shell, SUA1235, 1295 ± 105 yrs BP (Pickard and Adamson, 1983b).*
5. *Terrace, Dingle Lake.* **Laternula** *Shell, ZDL79, 5600 ± 77 yrs BP.*
6. *Terrace, Deep Lake,* **Laternula** *Shell, ZDL80, 6632 ± 118 yrs BP.*
7. *Terrace, Deep Lake and Club Lake, Worm tubes, SUA1237, 5340 ± 90 yrs BP (Pickard and Adamson, 1983b).*
8. *Terrace, Triple Lake,* **Laternula** *Shell, ZDL78, 6141 ± 90 yrs BP.*
9. *Dry basin near Platcha, algae, ZDL81, 5677 ± 94 yrs BP.*
10. *'Airport Beach', modern* **Laternula** *Shell, ZDL84, 1312 ± 65 yrs BP.*
11. *Terrace, Mud Lake,* **Laternula** *Shell ZDL66, 3325 ± 103 yrs BP.*
12. *Terrace, Watts Lake. Calcareous tufa, ZDL85, 7616 ± 104 yrs BP.*
13. *Modern marine sediment, 17m water depth, algal mud, SUA1236, 1310 ± 125 yrs BP (Pickard and Adamson, 1983b).*
14. *Terrace, Laternula — Mud Lake,* **Laternula** *Shell, SUA1411, 2880 ± 85 yrs BP (Pickard and Adamson, 1983b).*
15. *Terrace, Stinear Lake,* **Laternula** *Shell, SUA1239, 5740 ± 105 yrs BP (Pickard and Adamson, 1983b).*
16. *Terrace, Watts Lake, marine sediment after acid treatment, SUA1410, 8700 ± 100 yrs BP (Pickard and Adamson, 1983b).*

All radiocarbon dates here are uncorrected. The correction probably will be – 1300 years (see dates for sites 4, 10 and 13) (see Zhang et al. 1983).

(1 km east of 'Airport Beach' have only recently been raised beyond the reach of the sea as their surrounding beaches are less than 2 m asl. A glacio-isostatic uplift rate can be inferred for the Vestfold Hills from the data presently available. A rate of 1.0-2.5 mm y^{-1} for the last 6000 years (Zhang et al. 1983) is similar to that calculated for other Antarctic coasts (Table 6). The inevitable result of continued isostatic uplift will be further emergence which in the case of Taynaya Bay and the shallow fjords may lead to the evolution of more 'Death Valley Systems' (see Adamson and Pickard 1983).

There is scope for detailed work once a survey datum and network of benchmarks has been established.

Placc	Height (m asl)	Date of ^{14}C dating (years BP)	Rate of Uplift (mm y^{-1})	Reference
Windmill Islands (66°20'S, 110°20'E)	23	6040	3.9	Cameron and Goldthwait 1961
Marble Point (77°30'S, 163°50'E)	13	4450	2.9	Nichols 1968
East Ongul Island (69°01'S, 39°35'E)	3-4	3840	1.0	Meguro et al. 1984
Vestfold Hills (78°20'S, 68°35'E)	3-15	3500-6000	1.0-2.5	this report

Table 6. A comparison of raised beaches in Antarctica.

Part B. Late Quaternary geology

While the physical geography of the Vestfold Hills has been described by several workers (Crohn 1959; Law 1959; McLeod 1963; Johnstone et al. 1973a; Burton and Campbell 1980) the Quaternary deposits have only recently attracted particular attention (Adamson and Pickard 1983; Pickard and Adamson 1983a and 1983b; Zhang et al. 1983; Zhang, in press).

The environmental evolution of the area can be further elucidated by detailed analysis of deposits on raised beaches and terraces, especially if the deposits are fossiliferous. Deposits refer to both glacial and interglacial regimes.

Evidence of at least two late Quaternary glacial and interglacial episodes (Figure 8) may be inferred.

6. Interstadial marine sediments and glacial deposits of the last glacial stage

6.1 INTERSTADIAL MARINE SEDIMENTS OF LATE PLEISTOCENE AGE

Late Pleistocene fossiliferous marine sediments underlie ground moraine and form a plain 4.5 km^2 at 20 m asl opening on to the south shore near the head of Crooked Fjord. Sections through these deposits were formed during the Holocene after deglaciated terrain was invaded by the sea and before any glacio-isostatic uplift. It appears that when uplift did occur it was episodic as some of the sections are in a series of terraces above the eastern shore of Burton Lake. Typical sections show moraine thicknesses of 0.5-2.0 m, and up to 7 m of marine sediments are exposed in some instances. Shells from among fossils collected near the top of the marine section yield a radiocarbon date of 31 000 ± 474 (ZDL68) (Figure 8 and Table 5).

These deposits are mentioned by Pickard and Adamson (1983a) and described by Zhang (in press). The stratigraphy is summarised in Figure 9. Three characteristics should be mentioned.

First, layers 1 to 4 (Figure 9 and Plate 28) are very rich in fossil diatoms (31 genera and 60 species).

The principal species of diatoms are as follows (Li Jiaying, in press):

Diploneis sejuncta (A.S.) Jore
Diploneis splendida (Greg.) Cl.
Coscinodiscus symbolophorus Grun.
Coscinodiscus symbolophorus var. *oamoruensis* A. Schm.
Stephanopyxis sp.
Thalassiosira gravida Cl.
Melosira sulcata (Ehr.) Cl.
Hyalodiscus radiatus (O'Meora) Grun.

Coscinodiscus apiculatus Ehr.
Coscinodiscus biradiatus Grev.
Coscinodiscus moelleri A.S.
Coscinodiscus oculus iridis
Thalassiosira undulata (Mann) Sheshuk
Stephanopyxis turris (Grev. et Ar.)
Licmophora communis (Heib.) Grun.
Grammatophora arcuata Ehr.
Fragilariopsis curta
Cocconeis extravagans Jansch.
Cocconeis imperatrix A.S.
Cocconeis pinnata Greg.
Diploneis smithii (Breb.) Cl.
Pinnularia quadraterea var. *baltica* Grun.
Pinnularia quadraterea var. *costricta*
Trachyneis aspera var. *intermedia* Grun.
Melosira omma Cl.
Cocconeis antiqua Temp. et Br.

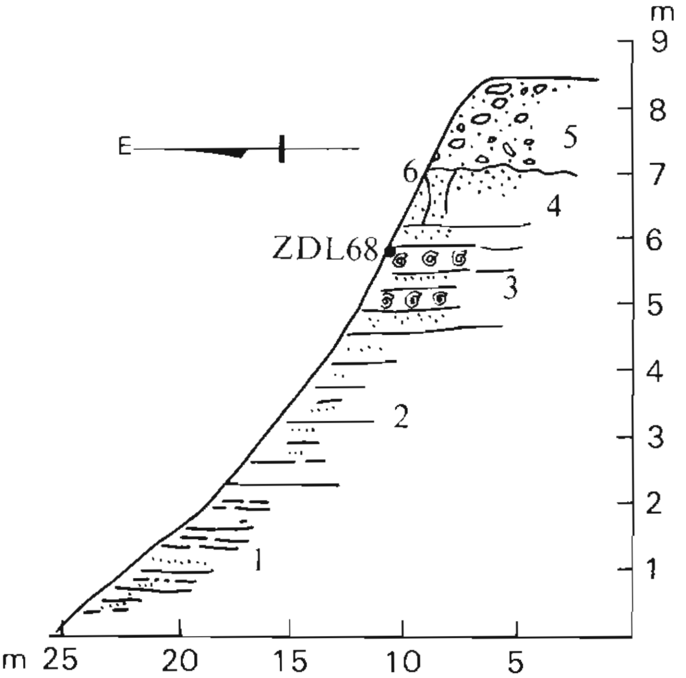


Figure 9. Section of marine deposits at Marine Plain. (1) Blueish grey silt and fine sand; (2) Greyish yellow sand cross bedding with fossiliferous coarse sand; (3) Grey coarse sand containing large number of cemented molluscs; (4) Weathering zone consisting of loose greyish yellow sand with abundant diatoms; (5) Till; (6) Fossil sand wedges.

A large number of fossil lamellibranchs (3 genera and 4 species) including *Hiatella aff. arctica* Linné, *Nucula oblique* (Lamarek) and *Chlamys (Athlopecten)* sp. are found in layers where they are cemented together with $CaCO_3$ (see Lan Xiou, in press, his Plates V and VI). A marine environment is indicated for the deposition of layers 1 to 4.

In addition to the fossil content the marine origin of the deposits is apparent from aspects of the geochemistry. In layers 1 to 4 the mean contents of oxides is 66.10% (Si), 7.63% (Al) and 8.69% (Fe). This compares with values of 51.23%, 16.9% and 17.9% respectively in layer 5 (Zhang, in press). High silica contents in sediments dredged from offshore along the coast of Princess Elizabeth Land have been attributed by Lisitzin (1972) to generally high concentrations of amorphous silica.

Secondly, the exposures indicate, from their texture grading (upward coarsening) and fossil content (all organisms in layer 3 are benthic), that marine sediments were deposited in progressively shallowing conditions. Specifically, thin horizontal bedding of dark grey and grey clay and silt occurs in the lower part of the exposure (layer 1 and 2) with a clay and silt content more than 80% (total weight). In the upper part of the exposure (layers 3 and 4) sandy cross bedding with coarse sand and pebbles appears in which medium and fine sand increase rapidly up to 30%-59% (total weight).

Finally, the moraine deposit (layer 5) has buried a weathered zone (average thickness 60 cm) containing fossil sand wedges (layer 4, containing marine diatoms) indicating a significant interval of subaerial periglacial conditions before a glacial advance.

6.2 COMPARISON WITH THE McMURDO SOUND REGION

Similar marine deposits have been recorded from between Cape Royds and Cape Barne, Ross Island (David and Priestley 1914; Debenham 1920). Handy et al. (1969) describe this deposit of frozen marine sediment as having three layers between the altitudes of 28 and 31 m. They obtained a radiocarbon date of 36 000 ± 2300 ^{14}C years BP on molluscs and bryozoa ascribed to the lowest layer. Fauna from the deposit comprised a bryozoan and molluscan community which is common today in the Ross Sea at depths of between 63 m and more than 180 m.

The southern exposure of the marine deposits between Cape Barne and Cape Royds consists of a layer of sponge spicules and shells about 30 cm thick resting on an ice core and overlain by glacial drift. The radiocarbon date from shell is more than 49 000 ^{14}C BP (Y-2642). The northern exposure near Cape Royds, consisting largely of Serpulae mixed with drift yielded ages (separate determinations) of 36 000 + 1200 − 1100, and 39 000 + 2100 − 1700 BP (QL-83) (Stuiver et al. 1981). In addition, foraminifera are abundant and display considerable diversity. There were 26 genera and 86 species of foraminifera determined (Ward and Webb 1979). This comprised 3 agglutinated, 3 planktonic, and 80 calcareous benthic species. A random census count of VNA-8 (=PNW-3) suggests relative significance as follows:

Ehranbergina glabra	51.33%
Cibicides labatulus	14%
Islandiella islandica	11.67%
Cassidulinoides porrectus	4.67%
Rosalina globularis	3.67%

Epistominella vitrea	3.3%
Globocassidulina crassa	2.33%
Planispirinoides bucculentus	1.67%
Globigerina megastoma	1.33%
Pseudobulimina chapmani	1%

These results and the excellent preservation suggest that the population from VNA-8 (= PNW-3) is an almost undisturbed assemblage. Consideration of the tidal foraminiferal fauna and comparison with post-glacial faunas from the Ross Sea suggest an original bathymetry in excess of 100 m and possibly as deep as 400 m.

Investigators agree that the fossiliferous marine sediments on Ross Island were deposited during a Late Pleistocene interstadial (Handy et al. 1969; Denton et al. 1970; Ward and Webb 1979; Stuiver et al. 1981). Although there is still argument about the post-depositional history of the Ross Island marine sediments they may be compared with the Middle Wisconsin stage of the North American sequence (Stuiver et al. 1981). The importance of the Ross Island interstadial marine sediments for Antarctic stratigraphy is greatly strengthened by the dated sequence from Marine Plain in the Vestfold Hills because here an interstadial between 50 000 BP and about 25 000 BP is evident.

6.3 THE LATEST GLACIAL DEPOSITS OF THE PLEISTOCENE

The glacial deposits overlying the marine sediments are less than 31 000 years old and probably correlate with deposits of the last glacial stage as documented from many parts of the world (Denton et al. 1970; Stuiver et al. 1981; Anderson 1981; Mayewski et al. 1981; Shi and Wang 1981). In the Vestfold Hills ice advanced to cover the highest summits (e.g. Stalker Hill 158 m asl). This was a time of lower sea level and hill tops including those currently occurring as offshore islands (e.g. Gardner Island and Magnetic Island) were ice covered. Both erosional (striae) and depositional (till) evidence testify to this advance which can be extended to the glacial low sea level grounding line. In some parts of Antarctica this was situated at the edge of the continental shelf (e.g. as far north as about 67 °S, see Hughes et al. 1981). Robin (1977) has estimated former ice thicknessess at least several hundred metres thicker than at present for most currently ice-free coastal areas in eastern Antarctica.

During this glacial phase ice abrasion fashioned abundant roches moutonnées and striae. The orientations (Plate 29, also see Plates 5 and 6) of these indicate that ice flowed from south-east to north-west* except during later stages of thinning and retreat when some ice flow patterns followed the line of valleys draining west, as detailed in the following section.

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*The ice advance that inundated the presently ice-free Vestfold Hills during the last glacial stage has been dubbed the 'Vestfold Event' by Adamson and Pickard (1983) (see also Pickard 1982).

7. The post-glacial deposits of the Middle and Late Holocene

7.1 THE POST-GLACIAL DEPOSITS.

Post-glacial deposits are widely distributed in the Vestfold Hills and include depositional terraces around many of the saline lakes and a very few of the freshwater lakes (e.g. Watts Lake, see also Figure 8). From them a rich assemblage of molluscs, foraminifera, diatoms and sponge remains have been identified and described (McLeod 1963; Zhang, in press). In some localities (e.g. Watts Lake) a large number of worm tubes were found. Two similar natural sections through terraces, one at Watts Lake and the other at Mud Lake, serve to provide information on the age and environment of deposition (Figures 10 and 11, Plates 30, 31 and 32). The exposed sediments are poorly sorted with very little silt. Coarse, medium and fine sands mixed with pebbles or fragments occupy more than 95% in total weight. The bedding (inclined) and lithology indicate deposition into shallow water from immediately up-slope.

Fossils are abundant and all are benthic or coastal planktonic marine species. The deposits are especially rich in benthic animals and plants. For example, at Watts Lake (Figure 10 and Plate 30) calcareous worm tubes of *Hydroides* sp. and *Sperobis* sp. occupy more than 50% of total sediments (Plate 31). Other fossils such as the 6 species from 4 genera of lamellibranchs *(Laternula elliptica* King et Broderip, *Laternula recta* Reeve, *Adamussim colbecki* (Smith), *Limatula hodysoni* Smith, *Thracia meridonalis* Smith, *Axinopsida* sp.) and 76 species from 44 genera of foraminifera are found (Lan, in press). The principal species of foraminifera are illustrated in Plates 34 and 35. Diatoms are illustrated in Plates 36 and 37 and lamellibranch in Plates 38 and 39.

The principal species of foraminifera, identified by Li Yuanfang (in press) are as follows:

Epistominella patagonica (d'Orbigny)
Cibicides lobatulus (Walker and Jacob)
Cibicides refulgens Montfort var.
Triloculina lamellidens Parr.
Patellina corrugata (Williamson)
Cassidulina crassa (d'Orbigny)
Trochammina squamata Parker and Jones
Trochammina antarctica Parr
Discorbis globularis (d'Orbigny)
Oolina squamosa-sulcata (Heron-Allen and Earland)
Nonionella iridea (Heron-Allen and Earland)
Lingulina sp.
Fissurina annectens (Burrows and Holland)
Quinqueloculina serra Crespin
Quinqueloculina seminulum (Linné)
Bolivina earlandi Parr.
Virgulina schreibersiana Czjzek
Laryngosigma hyalascidia Loeblich and Tappan
Cassidulina subglobosa Brady

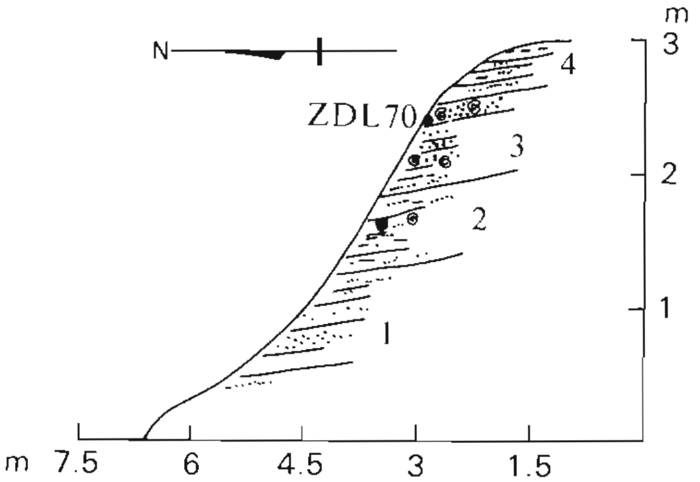


Figure 10. Post-glacial deposits on the terrace of Watts Lake. (1) Grey medium and coarse sand containing worm tubes, ***Laternula*** *and sponge etc.; (2) Greyish yellow and black sand with pebbles containing worm tubes and* ***Limatula hodysoni*** *(Smith); (3) Sand and small pebbles with large quantity of worm tubes,* ***Thracia meridonalis*** *Smith and* ***Axinopsida*** *sp.; (4) Fossil algae.*

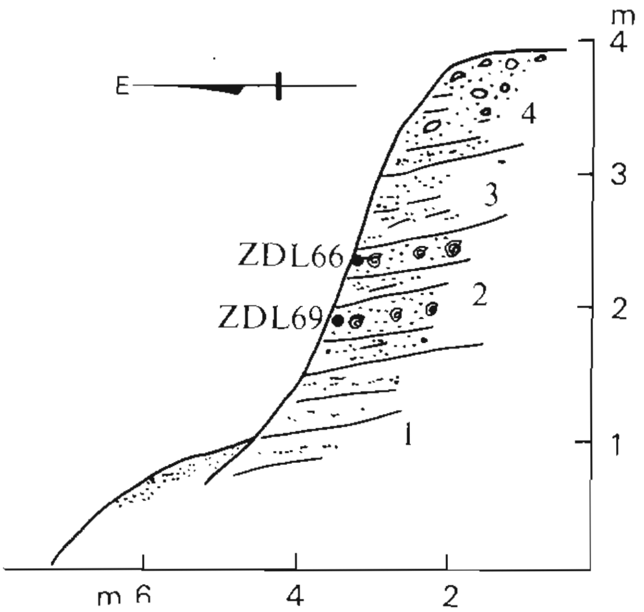


Figure 11. Post-glacial deposits on the terrace of Mud Lake. (1) Grey coarse sand; (2) Cross bedding of grey medium and fine sand, containing many ***Laternula elliptica;*** (3) Greenish grey sand and small pebbles and fragments; (4) Debris.

Parafissurina uncifera (Buchner)
Textularia earlandi Parker
Pyrgo patagonica (d'Orbigny)

7.1.1 Foraminifera assemblage

Two similar groups of foraminiferal assemblages are distinguished between lower and upper parts in the exposure at Watts Lake (Li Yuanfang, in press). The lower part (more than 6100 ^{14}C years BP, see Table 5) is very rich in foraminifera (76 species from 44 genera). This assemblage is named as the *Epistominella patagonica — Cibicides lobatulus — Triloculina lamellidens* group. The small *Epistominella patagonica* shells occupy the bulk of the fossil content. Other components include *Patellina corrugata* (Williamson), *Parafissurina uncifera* (Buchner), *Bolivina earlandi* (Parr.), *Oolina squamosa-sulcata* (Heron-Allen and Earland), *Virgulina schreibersiana* (Czjzek), *Nonionella iridea* (Heron-Allen and Earland), *Fissurina annectens* (Burrows and Holland) and *Lingulina* sp. etc. In the upper part (dated from shells at 6000 years BP (ZDL70) (see Table 5), foraminifera apparently decrease both in numbers and in genera and species, especially for the groups with small shells. Thus a group of large shells, identified as belonging to the *Cassidulina crassa* group, occupies about 77.5% of the total count of the foraminifera assemblage in the upper part of the exposure. Other components in this assemblage are
Cassidulina subglobosa (Brady)
Epistominella patagonica (d'Orbigny)
Patellina corrugata (Williamson)
Pyrgo patagonica (d'Orbigny)

The Mud Lake exposure (Figure 11) yielded ^{14}C dates of 3500 ± 86 years BP (ZDL69) in the lower part and 3325 ± 103 years BP (ZDL66) from the upper part. The foraminiferal assemblage is quite different from that of Watts Lake in that it contains only 18 species from 15 genera. Furthermore, the large shell group (e.g. *Cassidulina crassa* d'Orbigny) occupies 80-97% of the whole foraminiferal assemblage in the lower part of the Mud Lake exposure. Numbers of other components, such as *Cassidulina subglobosa, Discorbis globularis, Cibicides refulgens* and *Epistominella patagonica* are few.

In the upper part of this exposure species diversity and total numbers in the foraminiferal assemblage were rapidly decreased. Only 8 species from 4 genera such as *Discorbis globularis* (d'Orbigny), *Trochammina squamata* (Parker and Jones), *T. antarctic* (Parr.), *Discorbis margaritaceus,* and *Epistominella patagonica* (d'Orbigny) were found, indicating that a residual population of the original assemblage was surviving under difficult conditions.

The diatom assemblage identified by Li Jiaying (in press) indicates that the 20 genera and 57 species in the exposure of Watts Lake were marine species belonging to benthic and coastal planktonic types. Thus a similar environment to that indicated by the diatom assemblage from the pleniglacial deposit exposed on Marine Plain is indicated. It is a shallow water assemblage. The main species identified are as follows (see also Plates 36 and 37):
Melosira omma Cl.
Melosira sulcata (Erh.) Ktz.

Coscinodiscus biradiatus Grev.
Charcotia actinochilus (Ehr.) Hust.
Actinocyclus sp.
Asteromphalus cf. *parvulus* Karst.
Actinocyclus oculatus Jouse
Eucampia balaustium Cast.
Pinnularia quadraterea var. *constricta* Oestr.
Plagiogramma fenestra Brun.
Grammtophora arcuata Ehr.
Cocconeis pinnata Greg.
Cocconeis costata Greg.
Cocconeis fasciolata (Ehr.) Brown
Achnanthes brevipes var. *intermedia* Cl.
Cocconeis antiqua Temp. *et* Br.
Trachyneis aspera (Ehr.) Cl.
Diploneis graeffii Grun.
Navicula grevillei var. *comoidis* (Ag.) Mill.
Navicula grevillei Ag.
Pinnularia quadraterea var. *baltica* Grun.
Achnanthes brevipes var. *intermedia* Cl.
Navicula directa W. Sm.
Navicula grevillei (Ag.) Heib.

A comparison between the diatom assemblages at Marine Plain and at Watts Lake yields information about depositional environments and illustrates some of the changes that may take place during a progressive decrease in water depth. Between the two exposures the greater species diversity in the benthic diatom assemblage at Watts Lake, together with the poorly sorted texture of the sediments, establishes it as more typical of the shallowest marine deposits. Among the benthic diatom assemblage from the Marine Plain samples the greatest diversity is found in the uppermost part of the exposure, testifying to decreasing water depth there before deposition of the last Pleistocene marine deposits found. Conversely, the diversity of planktonic diatoms is much greater in the assemblage from Marine Plain, and more so in the lower than the upper part of the exposure. Overall, the data suggest that the marine deposits of Marine Plain were laid down in shallow water but significantly deeper than the environment of deposition indicated by the evidence at Watts Lake. Here large numbers of worm tubes *(Hydroides* sp. and *Sperolis* sp.) make up 50% or more of the bulk of sediments. Deposition probably took place just below the low tide level.

7.2 EVIDENCE FOR LATE HOLOCENE ACTIVITY

7.2.1 Deposits

The youngest tills are found in front of and up to 3 km from the present ice edge (Figures 8 and 12). Beyond this zone some bedrock surfaces harbour well-developed lichen cover in marked contrast to the boulders on the young moraines. Here black crustose lichens of small diameter may be found on boulders of the debris cover that settled after substantial melting of the ice in the recessional debris mantled ice-cored shear moraines that parallel the comparatively clean ice front.

In some places along the front are horizontal bands of debris concentrated in shear zones that dip towards the glacier base and strike parallel to the front. Thus the most modern moraine deposits in the Vestfold Hills are like those described from arctic locations (e.g. Goldthwait 1951; Boulton 1970) and would be expected to become more obvious along the shear zones most loaded with englacial debris as ablation proceeds until disintegration of the ice core. In some places the englacial debris appears to have been more abundant, for example, in the moraine area known as the 'Garbage Dump' which stands out prominently from the rest.

Evidence for retreat is most marked on the edge of the Sϕrsdal Glacier, east of the head of Crooked Fjord and three till types can be distinguished here (Figure 12).

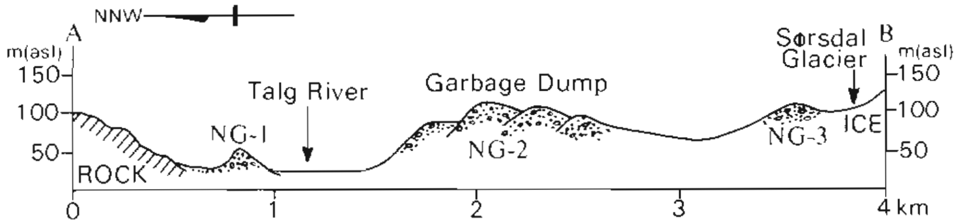


Figure 12. Late Holocene deposits in the 'Garbage Dump' area. (1) NG-1, early terminae; (2) NG-2, later terminae; (3) NG-3, latest terminae.

First, about 3 km from the ice front the outermost moraine crests protrude as a discontinuous line above the niveo-aeolian, colluvial and channel deposits of the Talg River (Plate 10).

Second, there is the blue/grey clay, fine sands and boulders occupying about 3 km^2 and reaching 70-120 m asl in the 'Garbage Dump' area (Plate 11).

The third type consists of the grey bouldery till of the ice-cored moraines that occupy a 200-500 m zone along the ice front, marking the edge of the ice-free area of the Vestfold Hills. These moraines have 10-50 m relief (Plate 9).

7.2.2 Striae

Two generations of striae have been recorded on the bedrock surfaces that border the Sϕrsdal Glacier within the zone where the youngest tills are found. An older set, orientated between 250° and 270° (true), is cut across by the other set which is orientated between 330° and 350° (true) (Plate 33). The older set therefore parallels the axis of the Sϕrsdal Glacier while the other parallels the dip direction of the shear planes in the ice-cored shear moraines. This suggests that the inland ice re-activated its outlet down the main trough of Crooked Fjord (now beneath the Sϕrsdal Glacier) after the retreat that followed the Flandrian Transgression. The Sϕrsdal Glacier, after establishment of maximum Holocene positions, to which the striations between 250° and 270° refer, underwent thinning so that its northern half sloped north and flow lines developed in that direction.

7.2.3 Age

As mentioned, an ice-free period between 6600 BP and 3500 BP is evident from documentation of marine deposits around many lakes (e.g. Mud Lake and Watts Lake). The advance of the Sørsdal Glacier therefore occurred after 3500 BP. It is interesting to note that late Holocene advances are recorded elsewhere (Clark and Lingle 1979; Shi and Wang 1981) but detailed correlation is not possible without more data. In the case of the outlet of a major ice sheet, ice margin fluctuations are greatly influenced by the dynamics of the inland ice and this may not be entirely a function of climatic change.

Summary

Available data on geomorphology, glacial geology, palaeontology, sedimentology and geochronology from the Vestfold Hills have combined to yield some insight into the history of glaciation, geomorphic and environmental evolution of eastern Antarctica since the late Pleistocene.

Inland from the trailing edges of the continental lithosphere large areas have been subjected to extended periods of subaerial denudation. The resulting surfaces sometimes date from times before the trailing edges of Gondwanaland separated each side of the spreading centres of the developing Southern Ocean. Around the Indian ocean these surfaces, termed by King (1967) the 'Mesozoic Oldlands', are the remnants of Gondwanaland landscapes. In the Vestfold Hills two planation surfaces ('erosion surfaces' of Crohn 1959) are recognised but their ages are unknown. It is supposed that planation may have ended when the Antarctic ice sheet covered the continent during the early to middle Tertiary (Denton et al. 1971; Le Masurier and Rex 1982; Tingey 1982). Therefore the ages of the Vestfold Hills' planation surfaces may be the equivalent of the 'Mesozoic Oldlands'. Alternatively one or both may date from a 'post-Gondwana' time.

The geomorphology of the Vestfold Hills is strongly influenced by geological structure. Fault line scarps and linear valleys were glacially excavated with the result that prominent lineaments, occupied by fjords, valleys and scarps, and with preferred orientations (east-north-east, north-east and north-west) dominate the physiography. Periglacial processes have modified the glaciated landscape in post-glacial times.

Marine sediments yield abundant marine fossils (diatoms, foraminifera and lamellibranchs) and a minimum radiocarbon date of 31 000 years BP from Marine Plain in the Vestfold Hills indicates an interstadial period between about 50 000 to 25 000 years BP. Such a period of high sea levels has also been documented from McMurdo Sound and many other areas of the world (Denton et al. 1970; Ward and Webb 1979; Stuiver et al. 1981; Shi and Wang 1981; Anderson 1981; and Mayewski et al. 1981). During that stage sea level was higher than at present and part of the Vestfold Hills area (e.g. Marine Plain) was submerged. Palaeontological evidence indicates submergence of between tens to a hundred metres (Li Jialing, in press). The ice-free area at this time was of similar size to the present one. Inshore water temperatures were also not unlike those of today (perhaps a little warmer) suggesting that climatic conditions during the pleniglacial here were similar to those of the present.

During the last glacial stage, which has been widely recognised in Antarctica, Europe, Asia, and North America (Denton et al. 1970; Flint 1971; Stuiver et al. 1981; Shi and Wang 1981; Anderson 1981; and Mayewski et al. 1981), the Vestfold Hills and the offshore islands were completely covered by part of the Antarctic ice sheet which grounded on the continental shelf edge possibly as far north as 67°S (Hughes et al. 1981). Evidence confirms that the ice sheet covering the present ice-free area was at least 158 m thick but it may have been much thicker (Robin 1977).

The orientation of the striae, between 290° and 310° (true), shows that the ice flow was from the south-east to north-west.

During this stage the terrain, including the remnants of former planation surfaces, was strongly modified and many glacial landforms such as roches moutonnées, polished pavements, and mamillated topography were formed. Complementary depositional glacial landforms are few and these have been modified by postglacial frost climates. The most extensive glacial deposit is the ablation-moraine mantle on Marine Plain.

Widely distributed and well preserved marine deposits identified from abundant marine fossils including foraminifera, diatoms, lamellibranchs and sponge spicules testify to relative sea levels higher during Middle Holocene times than at present. Shells from these deposits yield ages between 7500 and 5000 BP (Figure 8 and Table 5). An early Middle Holocene period as warm as the present may be inferred for eastern Antarctic regions in sympathy with similarly warm climates at this time in other parts of the world (Anderson 1981; Grove 1979; Mayewski et al. 1981; Shi and Wang 1981; and Stuiver et al. 1981). At this time the eustatic response to the world-wide deglacial hemicycle (known as the Flandrian Transgression or the Cangzhou Transgression and spanning the period 18 000 to 6000 BP) was near to maximum, especially in areas like the Vestfold Hills where glacio-isostatic response was far from complete. Thus between 7000 and 5000 years BP arms of the sea in the Vestfold Hills were more extensive than now. Glacio-isostatic recovery there is probably still incomplete. During the last 7000 years or so it has caused the innermost arms of the sea to be drained, leaving behind raised beaches up to 15 m asl. A mean rate of glacial-isostatic uplift during the last 6000 years of up to 2.5 mm y^{-1} can be inferred.

The location and age of the oldest emerged Holocene marine deposits indicates that the ice-free area in the Vestfold Hills some 7500 years ago was about 80% (possibly more) of the area exposed today. This means that over much of the area above 15 m asl and beyond the zone of the late Middle Holocene advances post-glacier landscape modification by periglacial processes has been in progress for most of the Holocene. The possibly milder climates of the early Middle Holocene, coupled with the more restricted extent of saline groundwater at that time, may have been more favourable to the development of periglacial landforms in the Vestfold Hills than is the present environment.

The distribution of patterned ground correlates with the distribution of those surficial deposits that develop a moist summer active layer lacking or very low in salt content.

Because many of the surficial deposits are in low-lying areas that were former arms of the sea in earlier Holocene times, patterned ground is of restricted occurrence.

Well-developed patterned ground, such as large sorted circles and debris islands, always appears in broad valleys close to the ice sheet and on low lying beaches of freshwater lakes into which glacial and nival melt-waters drain. Sorted circles are stable during winter when the ground is frozen but have been shown to expand in February and March after contracting in November and January. Mean net movement between the centres and borders of the circles monitored over two summers and two winters varies between 1.2 and 10.0 mm for different sizes of large sorted circles.

Compared with the range of periglacial processes and resulting landforms to be found in other cold climates such as the arctic areas (Washburn 1979) and the Tibetan Plateau (Cui 1981) the suite of features found in the Vestfold Hills lacks diversity.

The youngest tills and striae found in front of and up to 3 km from the present ice edge on the southern edge of the ice-free area is probably further evidence of the widespread nature of Late Holocene activities which have been documented in many regions of the world (Denton and Porter 1970; Shi and Wang 1981). Approximate contemporaniety with such events can be inferred from radiocarbon dates of 3325 ± 103 years BP (ZDL66) and 3500 ± 86 years BP (ZDL69) from fossiliferous marine deposits at Mud Lake, 2 km from the Sφrsdal Glacier. Late Holocene glacial fluctuations must post-date that time.

Three distinct stillstands are represented by the ice-cored shear moraines, or their remains, and separate advances and retreats may be indicated. Currently the northern edge of the Sφrsdal Glacier is thinning and retreating.

A preliminary chronology for glacial fluctuation in the Vestfold Hills during the late Quaternary and the comparison with other regions of the world is given in Table 7. It is suggested that the Vestfold Hills sequence exhibits some similarities to the sequence from McMurdo Sound as well as some from Europe, East China and North America. Global scale climatic change is indicated.

The Vestfold Hills represents one of the few relatively extensive ice-free areas in eastern Antarctica. As such it offers a link between glaciological and marine science studies that should widen the context of offshore and onshore studies of glaciers, geomorphology and geology towards elucidation of the late Cainozoic and possibly older history.

Specifically, there is scope for the following studies:

(a) *Monitoring glacio-isostatic recovery, particularly its distribution:*

This requires the installation and maintenance of a tilt meter. An array of tilt meters would allow ideas about differential isostatic recovery to be tested. The instruments should be placed in an array, with a minimum of three to cover as large an area as possible. Use of offshore islands for this purpose should be considered.

(b) *The nature and distribution of Holocene glacio-isostatic uplift:*

This project requires the detailed survey of Holocene marine terraces, and some further dating of shells found in them. The installation and maintenance of a reliable tide gauge at Davis station, and perhaps at a few other places, notably on the shores of some of the fjords, would allow use of sea level as datum for the levelling traverses necessary to establish the elevation of each terrace.

(c) *Monitoring of current geomorphic processes:*

(i) Shoreline development on the coast and around the salt lakes: some of the latter rarely freeze and so offer the rare chance to study wave activity in a polar

Area Sequence Age	Vestfold Hills (this report)	McMurdo Sound (Denton et al. 1970 Ward and Webb 1979 Stuiver et al. 1981)	East China (Shi and Wang 1981)	Eurasia and Greenland (Anderson 1981)	North America (Mayewski et al. 1981)
HOLOCENE	Glacier fluctuations since 3000 years BP	Present	Present Neoglaciation 3000-2000 years BP	Present little ice age 2500-2000 years BP	Present little ice age 2500-2000 years BP
	Post-glaciation since 7600 years BP to 3500 years BP. Optimum stage between 6500 and 5500 years BP.	Post-glaciation since 9490 years BP to about 4450 years BP. Optimum stage ranges 6600 to 5650 years BP.	Optimum climatic stage 6000-5000 years BP (Post glacial period since 10 000 years BP).	Post-glaciation period since 8000-4500 years BP. Hypsithermal between 7000 to 6000 years BP.	Altithermal maximum 7500-6000 years BP.
LATE PLEISTOCENE	Vestfold (last) glaciation since less than 30 000 to about 10 000 years BP.	Ross glaciation to about 25 000 years BP to 10 000 years BP maximum between 21 200 and 17 000 years BP.	Late Dali glaciation 25 000-10 000 years BP maximum 14 780 ^{14}C years BP.	Late Weichselian 25 000-10 000 years BP.	Late Wisconsin 24 000-10 000 years BP.
	interstadial 25 000 or 30 000 years BP ago estimate up to 50 000 years BP.	interstadial 25 000-50 000 years BP.	interstadial About 40 000-10 000 years BP.	Middle Weichselian 50 000-25 000 years BP.	Middle Weichselian 50 000-24 000 years BP.
		Ross glaciation 1.	Early Dali glaciation 70 000-40 000 years BP.	Early Weichselian more than 50 000 years BP.	Early Wisconsin more than 50 000 years BP.

Most dates quoted here refer to radiocarbon years before present.
See sources cited for methods of calibration and correction.

Table 7. Comparison of Late Pleistocene sequences.

climate without the interruption of freeze-up. Explanations for the horizontality of some southern high latitude shorelines in other places invoke the presence of shore ice and so will not be entirely appropriate for the terraces of the saline lakes of the Vestfold Hills area, and

(ii) Periglacial activity: relationships between environment, form and process. The distribution of saline groundwater has resulted in curious juxtapositions of forms that offer unusual scope for testing hypotheses about the efficiency of frost heaving and other cryogenic processes.

(d) *Quaternary stratigraphy and the identification of till types:*

Some of the more extensive deposits (e.g. those on Marine Plain) might be more fully documented after drilling. The discovery of a stratigraphic sequence to compare, even only in part, with some of the Dry Valley stratigraphies would be especially valuable. The depth of the Marine Plain deposit is unknown, as is the thickness of deposits on the bottom of any of the lakes.

(e) *The ice-cored shear moraines:*

These mark the outcrop of shear zones that might be further documented with the aid of ice sounding equipment. Of special interest would be the results of any attempt to map the distribution of the frozen bed zone of the Antarctic ice cap in the Vestfold Hills area.

(f) *Offshore deposits:*

Sampling and sub-bottom profiling should reveal some details of the glaciated area now under water offshore. Both erosional and depositional features are of interest.

Most of these projects refer to the Quaternary history of the Vestfold Hills, and more generally, to some other eastern Antarctic coastal areas. As such, Quaternary studies in the Vestfold area assume greater importance than many areas of similar size elsewhere on earth. The geomorphologic evolution of the Vestfold Hills has important implications for related work on the Quaternary history of Antarctica.

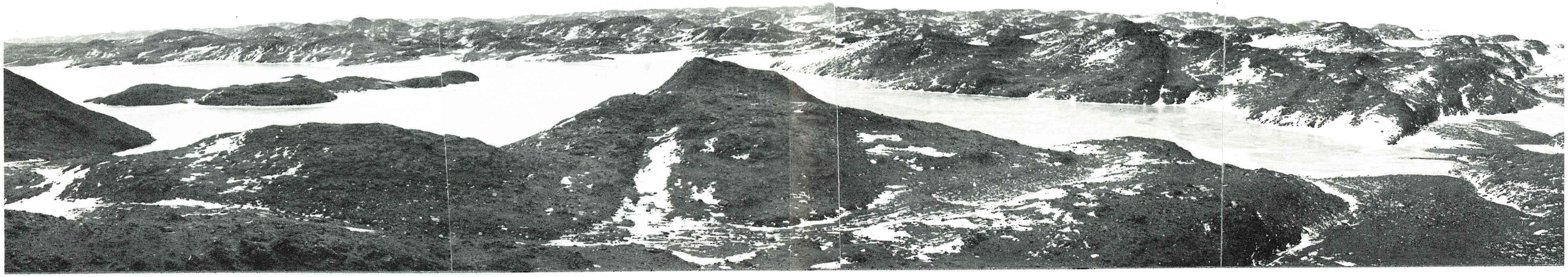


Plate 1. Panorama (120°) looking north from a hill south of the east end of Crooked Lake. Skylines like the one seen here are common and add credence to physiographic interpretations that infer the presence of old planation surfaces in the accordance of summits. 10 August 1981.

Plate 2. A typical fault line scarp trends east-north-easterly from Lake Druzhby to the edge of the present ice cap. Looking west from the edge of the ice cap; Trajer Ridge on the right rises up to 120 m above one of the valleys with asymmetrical cross profiles. 24 September 1981.

Plate 3. Silicified breccia forming the face of a steep southerly dipping fault plane trending north-north-east to form part of Luncke Ridge (north side of Long Fjord). 26 April 1981.

Plate 4. North-westerly trending parallel grooves (left to right, middle-ground) tens of metres wide and hundreds to thousands of metres long, are found between Lake Chelnok and Crooked Lake. View north-west from a hill north of Lake Chelnok. 9 August 1981.

Plate 5. A group of roches moutonnées 10-20 cm high and from tens up to hundreds of metres in length, occur on the southern side of Talg River. Dead ice left behind by retreat of the Sϕrsdal Glacier covers a roche moutonnée in the valley in the foreground. View is from a south facing slope south of Crooked Lake, looking west. 20 February 1981.

Plate 6. A roche moutonnée in the Talg River Valley. Blocks on its surface are glacial erratics. Striae on the surface are orientated west-north-west. The Talg River bed in background is filled with drift snow and dead ice. View from north side of river bed looking east. Late February 1981.

Plate 7. A 'reversed' roche moutonnée immediately north of the Talg River and close to Crooked Lake. The surface is striated and carries a few erratics. The steepness of the stoss side and the comparatively gentle slope on the lee side is attributed to the very strong influence of geological structure here. Direction of ice movement across the feature was east-south-east. View to north-west, late February 1981.

Plate 8. A till plain occupying about 4.5 km^2 at 20 m asl borders the northern side of Crooked Fjord near its head. The till, up to 2 m thick, mantles marine deposits containing shells which yielded a radiocarbon date of 31 000 ^{14}C yrs. View facing south towards the Sϕrsdal Glacier (background). 20 March 1981.

Plate 9. The youngest ice-cored moraine ridge, 5-20 m high, hundreds of metres wide, forms a continuous zone more than 20 km long parallel to the edge of the ice sheet. It is exposed in summer and partially buried by drifting snow in winter. View is to the north-east from a hill in the front of the ice sheet on the Long Peninsula. 10 October 1981.

Plate 10. An isolated hummock, 20 m high, 50 m wide and 150 m long, in the Talg River valley about 1 km north of the 'Garbage Dump'. Dead ice left behind by the retreat of the Søsdal Glacier is in the foreground. Looking to the north, 10 February 1981.

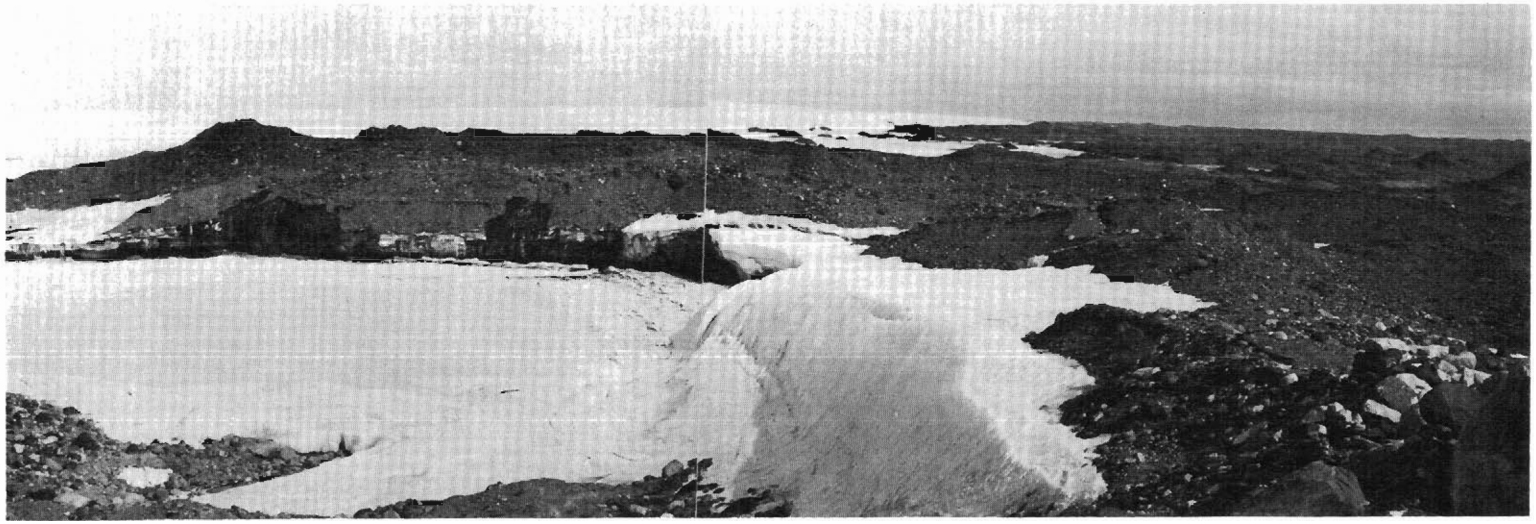


Plate 11. General view of the 'Garbage Dump', a large elliptical hummock (2.5 km^2) composed of ice-cored moraines upon which a very irregular surface relief has developed due to melting of core ice and diurnal freeze and thaw cycles in the active layer during summer. A thermokarst lake is in the left foreground, Sφrsdal Glacier left background. View facing west, 10 February 1981.

Plate 12. Parallel ice-cored moraine ridges 10-50 m high, tens to hundreds of metres wide, formed during Late Holocene changes in ice margins 3000-200 years BP. Ice-cored shear moraines formed during the glacier fluctuation, and exemplified here, can be found parallel to the northern side of Lake Chelnok, westward along the northern side of Crooked Fjord, between 200 m and 1500 m from north of Søɸrsdal Glacier margin. Facing to the west, with Sɸrsdal Glacier in left background and Chelnok Lake left middle ground, 22 September 1981.

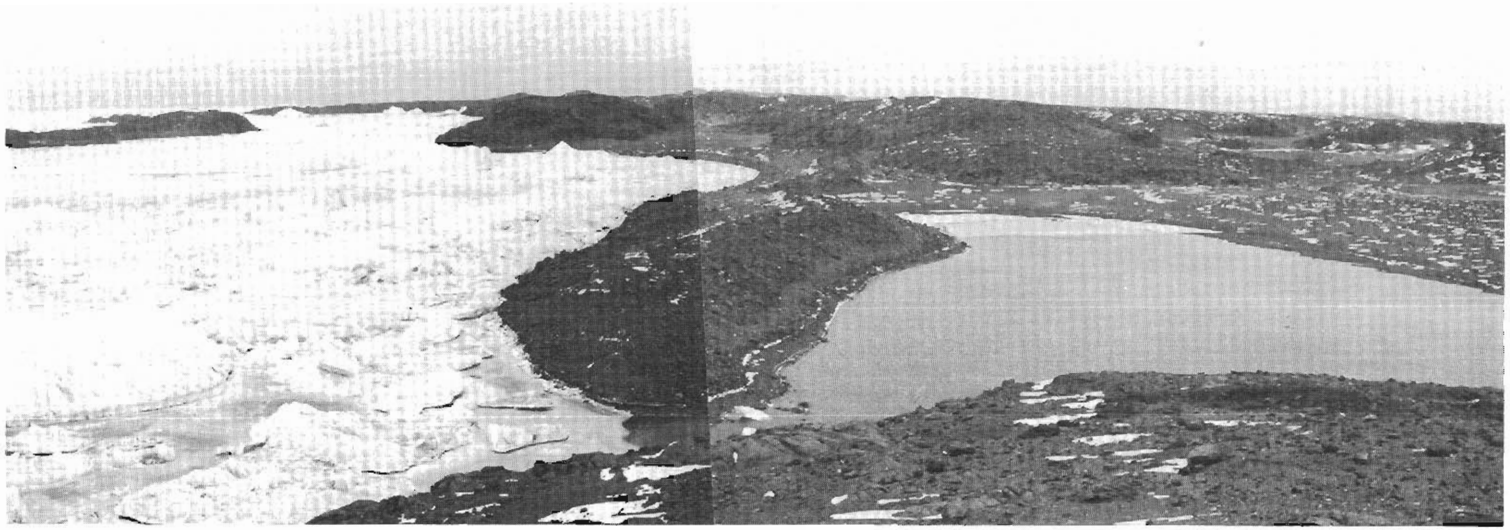


Plate 13. Ice-cored moraine ridge (centre, middle-ground), formed by Late Holocene glacier fluctuations to separate a lagoon (right) from Crooked Fjord (left). Sϕrsdal Glacier margin at the point shown here (far left background) is on dry land. View to the west, 20 February 1981.

Plate 14. Ice-cored shear moraine between bedrock (right and left foreground) and the Sørsdal Glacier (background). The moraine lies across a small valley ponding a small pro-glacial lake, fed by snow-melt and the thaw of dead ice. The moraine marks a later still stand during ice margin fluctuations of Late Holocene times. The youngest ice-cored moraine outcrops along the present ice sheet margin in the background. View is looking south from uppermost part of the Talg River valley, 22 August 1981.

Plate 15. Block slope on the east side of Jackson Hill, looking towards south-west, 20 October 1981.

Plate 16. Tors of ice-wedged country rock between Lake Druzhby and Crooked Lake. Facing west, 18 October 1981.

Plate 17. Large sorted circles on the beach of Mossy Lake. Scale given by a note book, late February 1981.

Plate 18. Large sorted stripe on a hillside above Mossy Lake. Mid-September 1981.

Plate 19. Debris islands in the bed of the Talg River. Scale given by hammer. Mid-February 1981.

Plate 20. Cavernous ventifacts developed on the east facing side of an outcrop of the coarse grained layer of the paragneiss, 0.5 km north of the head of Long Fjord. 12 October 1981.

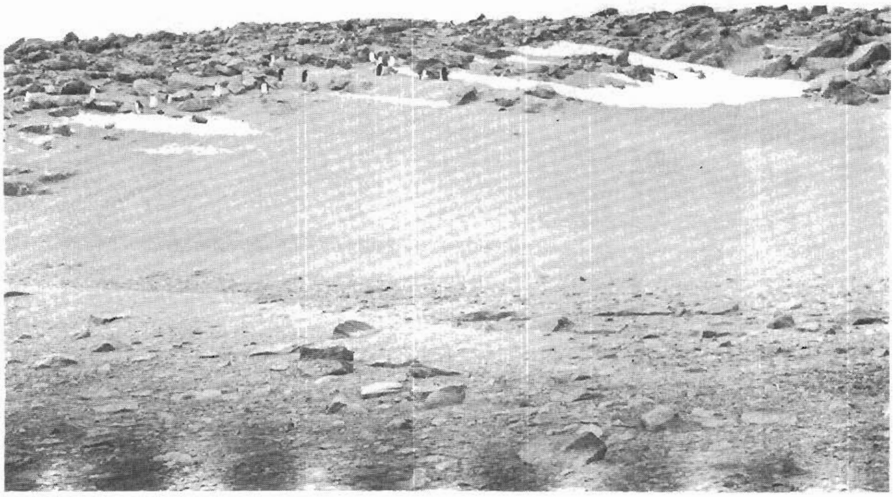


Plate 21. Sand slope and remnant of annual snow patch in the lee of a low north-westerly trending ridge (30 m high) west of Dingle Lake, looking east. Early March 1981.

Plate 22. Collapse of moraine debris mantle blanketing ice core of the disintegrating ice-cored moraine complex known as the 'Garbage Dump'. Such mass movement is most marked where the slopes face the higher sun angles during the summer. Here a saturated active layer yields mudflows and promotes gelifluction. View to the south-west, mid-February 1981.

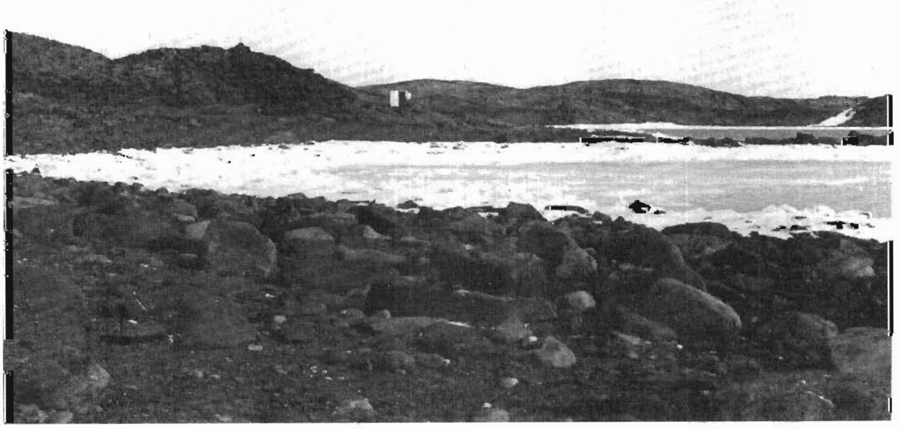


Plate 23. A modern coastal beach on the south shore of Long Fjord near Brookes Hut (middle of photograph). The intertidal zone here is between 50 and 100 m wide. The littorally reworked glacial drift in the foreground is above the present zone of wave action. View west when the new seasonal sea-ice cover was beginning to form, early March 1981.

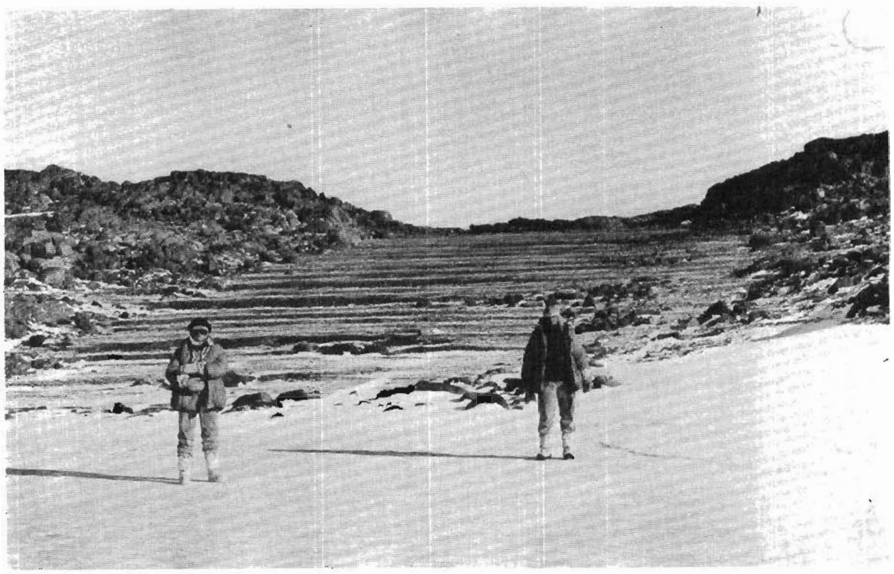


Plate 24. A well preserved high level terrace (top at 15 m asl) above the south-western tip of Zvezda Lake testifies to marine regression in this former arm of the sea. Glacio-isostatic rebound in this area has resulted in the formation of many steps on the terrace slope. 25 September 1981.

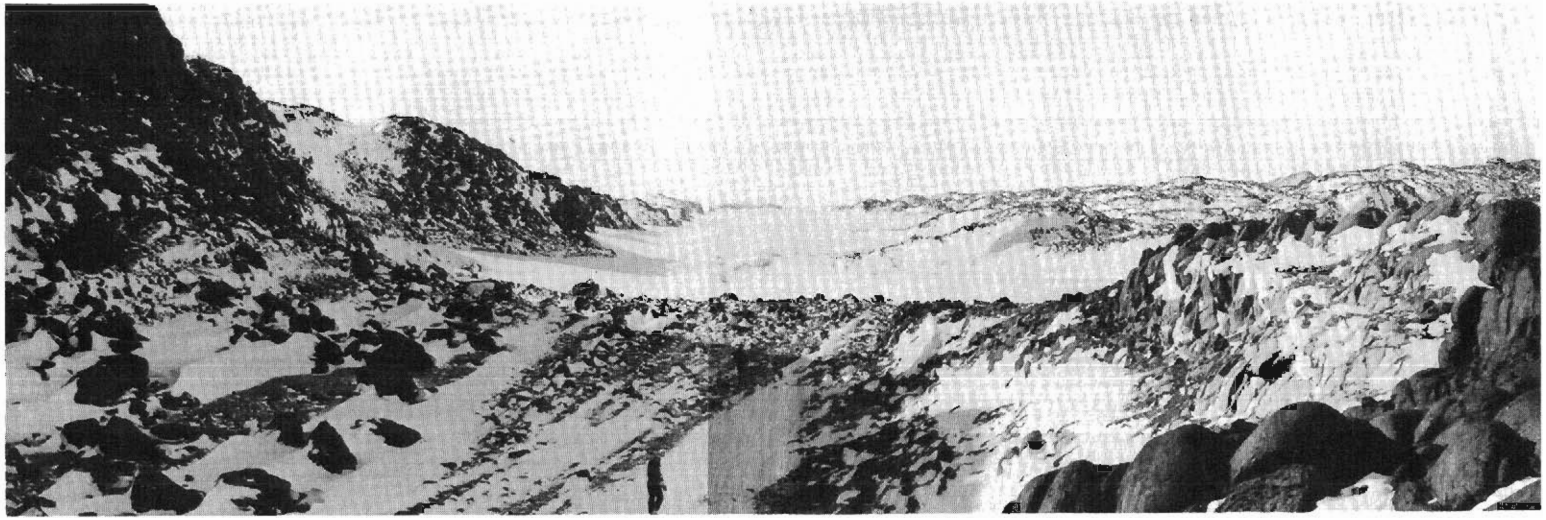


Plate 25. Raised marine terrace (10 m asl) on the interfluve at Tryne Crossing between Tyrne Fjord (background) and Long Fjord. The terrace is most probably about the same age as dated 10 m terraces at Stinear Lake and Watts Lake (i.e. 6 to 7000 years old). View facing north, 10 October 1981.

Plate 26. Well developed low level terraces (3 m and 6 m asl) appear on both sides of Ellis Fjord. View to the south-west from a hill north of Jackson Hill, 20 November 1981.

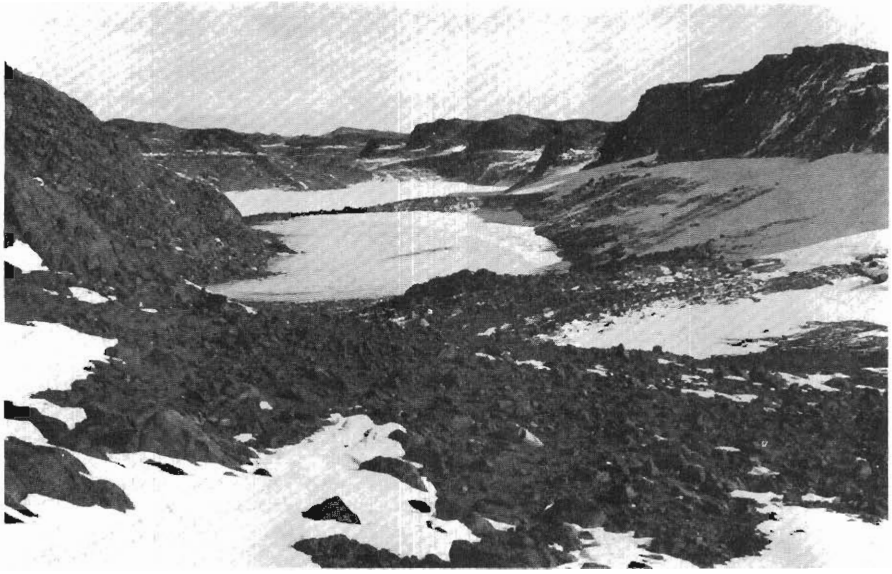


Plate 27. Sections of the 6 m asl terraces delineating the former extent of an arm of Long Fjord now occupied by the major salt lakes of Broad Peninsula. View north-east across unnamed lakes east of Oval Lake, early February 1981.

Plate 28. Section of marine deposits at Marine Plain as documented in Figure 9, looking south, mid-February 1981.

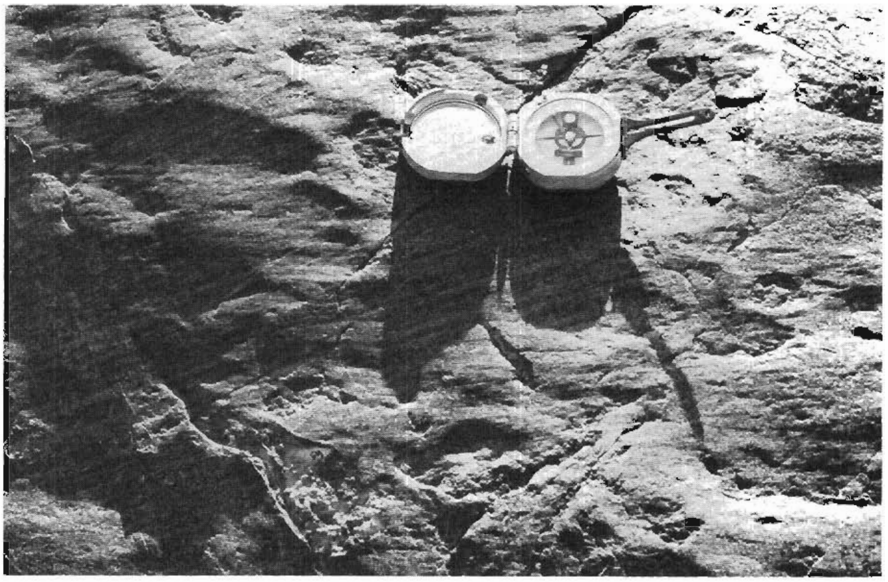


Plate 29. Striae orientated 290°-310° (true), 10°-30° (magnetic), are common on the surfaces of dolerite dykes in the Vestfold Hills. These striae indicate ice flow from south-east to north-west. Photograph taken north east of Burton Lake, 20 August 1981.

Plate 30. Section in Holocene terrace of Watts Lake. Fossil shells 30 cm from the top of the section are about 6000 radiocarbon years old. Natural exposure looking south-west. Watts Lake is in middleground right. Early February 1981.

*Plate 31. Worm tubes (**Hydroides** sp. and **Sperobis** sp.) and molluscs are preserved in sediments of a terrace of Watts Lake. Late January 1981.*

Plate 32. Section of post-glacial deposits on the terrace of Mud Lake as documented in Figure 11. Looking south-west, Mud Lake is in middleground left. Early February 1981.

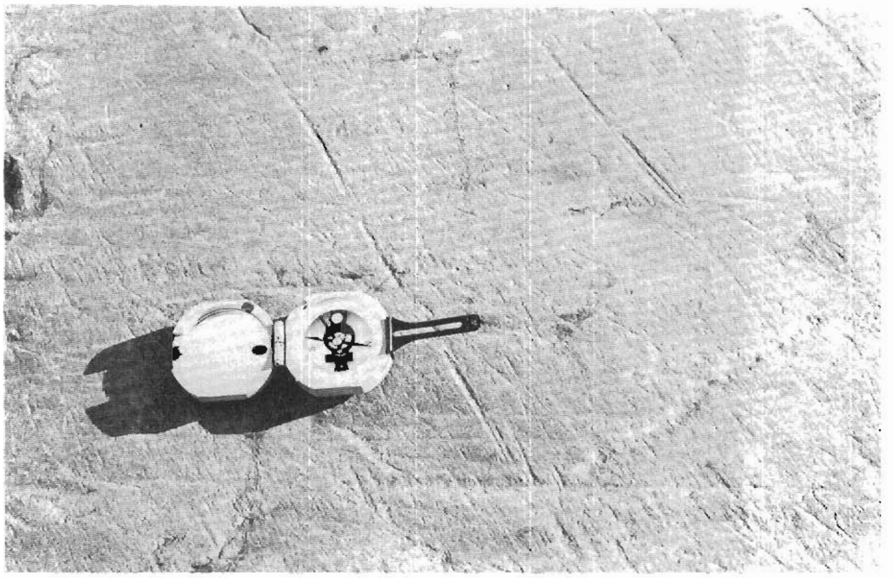
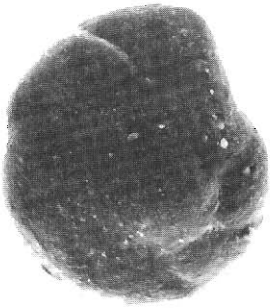


Plate 33. Two generations of youngest striae are found on the bedrock surfaces between the border of the Sφrsdal Glacier and Late Holocene ice-cored moraine ridge. An older set orientated between 250° and 270° (true), 330° and 350° (magnetic), is cut across by the other set which is orientated between 330° and 350° (true). 20 November 1981.

Plate 34. Foraminifera from Watts Lake (identified by Li Yuanfang).
1,2 **Epistominella patagonica** *(d'Orbigny), 1 x 222, 2. x 270*
3,4 **Cibicides lobatulus** *(Walker and Jacob), 3. x 72, 4. x 90*
5,6 **Triloculina lamellidens** *Parr. x 138*
7,8 **Cassidulina crassa** *d'Orbigny, x 60*
9,10 **Patellina corrugata** *Williamson, 9. x 126, 10. x 138*
11 **Trochammina antarctica** *Parr., x 120*
12 **Trochammina squamata** *Parker and Jones, x 180*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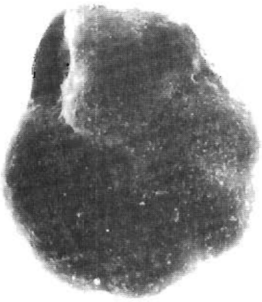
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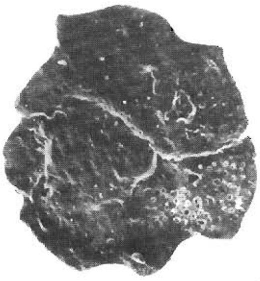
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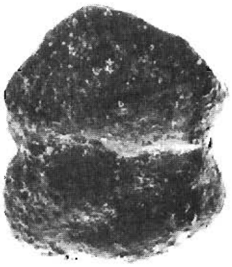
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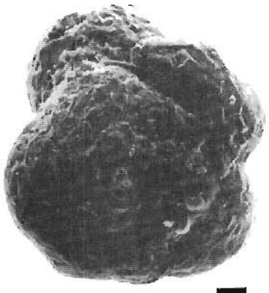
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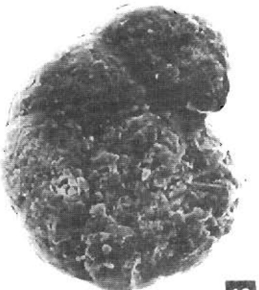
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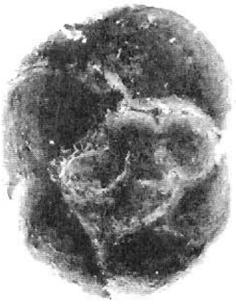
Plate 35. Foraminifera from Watts Lake (identified by Li Yuanfang).

1,2 **Discorbis globularis** *(d'Orbigny), x 60*
3 **Oolina squamosa-sulcata** *(Heron-Allen and Earland), x 120*
4,5 **Nonionella iridea** *(Heron-Allen and Earland), 4. x 150, 5. x 132*
6 **Lingulina** *sp. x 312*
7 **Fissurina annectens** *(Burrows and Holland), x 252*
8 **Quinqueloculina serra** *Crespin, x 96*
9 **Quinqueloculina seminulum** *(Linné) x 150*
10 **Bolivina earlandi** *Parr, x 138*
11 **Virgulina schreibersiana** *Czjzek, x 102*
12 **Laryngosigma hyalascidia** *Loeblich and Tappan, x 138*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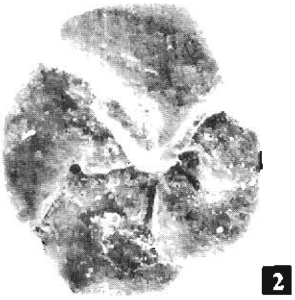
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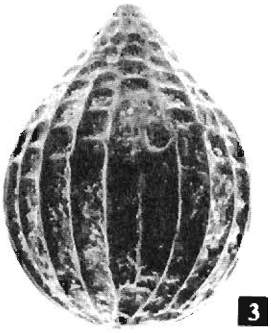
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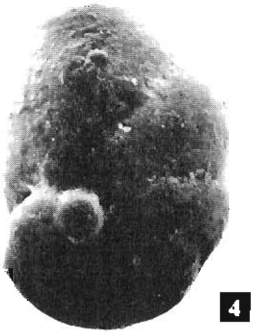
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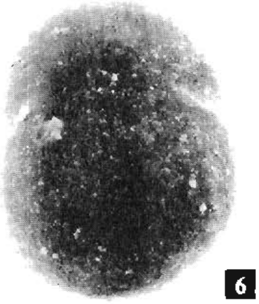
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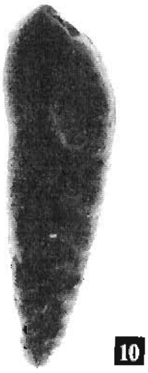
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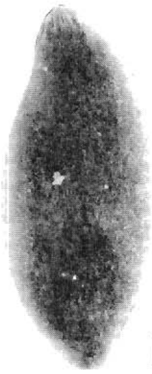
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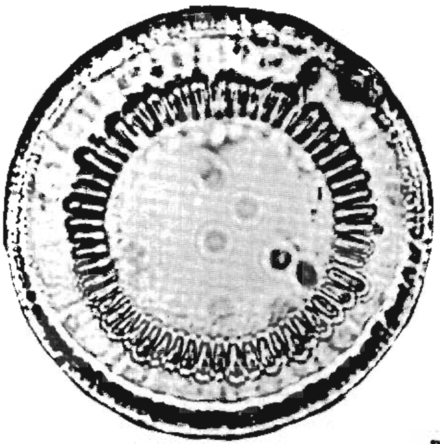
Plate 36. Diatoms from Watts Lake and Marine Plain (identified by Li JiaYing).

1 **Melosira omma** *Cl., x 1000, Lake Watts*
2-4 **Melosira sulcata** *Ktz., x 1000, Lake Watts*
5,6 **Stephanopyxis** *sp., x 1000, Marine Plain*
7 **Coscinodiscus biradiatus** *Grev., x 1000 L. Watts*
8 **Charcotia actinochilus** *(Ehr.) Hust., x 1000, L. Watts*
9 **Actinocyclus** *sp., x 1000, L. Watts*
10 **Coscinodiscus symbolophorus** *Grun., x 1000, Marine Plain*
11,12 **Coscinodiscus symbolophorus** *Var.* **oamaruensis** *A. Schm., x 1000*
13 **Asteromphalus** *cf.* **parvulus** *Karst., x 1000, L. Watts*
14 **Actinocyclus oculatus** *Jouse, x 1000, L. Watts*
15,16 **Eucampia balaustium** *Cast., x 1000, L. Watts*
17 **Thalassiosira gravida** *Cl., x 1000, Marine Plain*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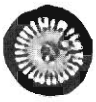
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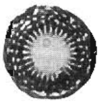
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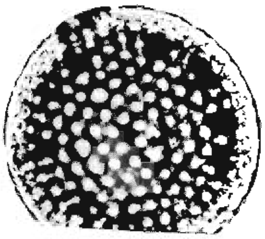
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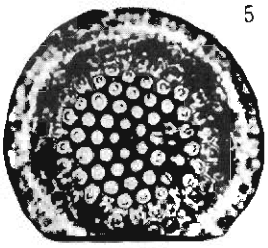
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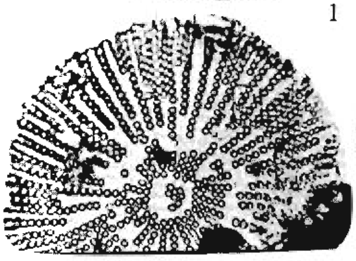
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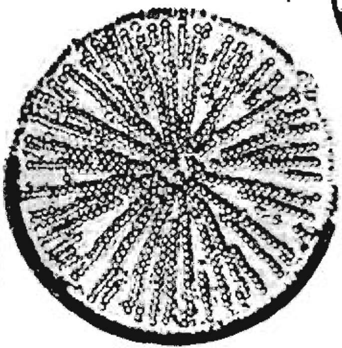
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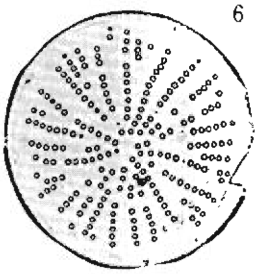
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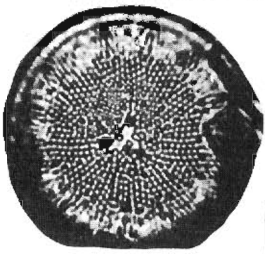
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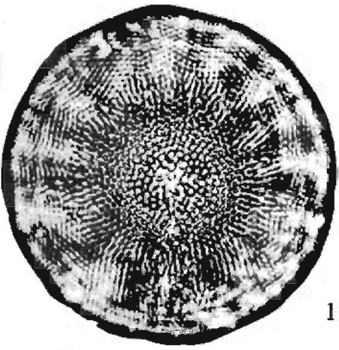
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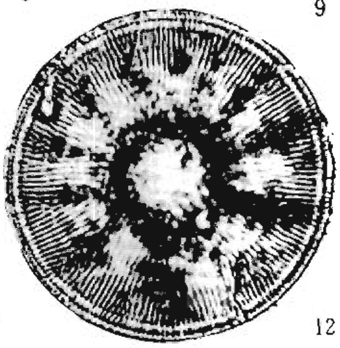
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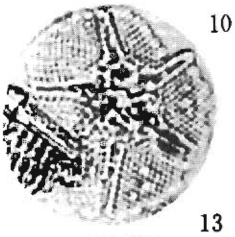
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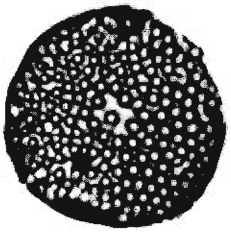
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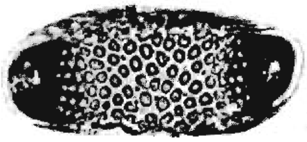
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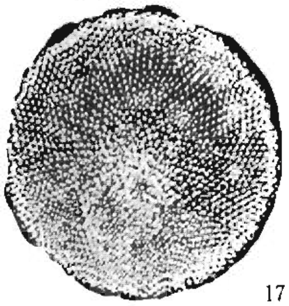
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Plate 37. Diatoms from Watts Lake and Marine Plain (identified by Li JiaYing).

1	**Plagiogramma fenestra** *Brun. x 1000, Lake Watts*
2,3	**Grammatophora arcuata** *Ehr., x 1000, L. Watts*
4	**Cocconeis pinnata** *Greg., x 1000, L. Watts*
5,6	**Cocconeis costata** *Greg., x 1000, L. Watts*
7	**Cocconeis fasciolata** *(Ehr.) Brown., x 1000, L. Watts*
8	**Achnanthes brevipea** *var.* **intermedia** *Cl., x 1000, L. Watts*
9	**Mastogloia** *sp., x 1000, L. Watts*
10,11	**Cocconeis antiqua** *Temp.* **et** *Br., x 1000, L. Watts*
12,13	**Achnanthes** *sp., x 1000, L. Watts*
14	**Trachyneis aspera** *(Ehr.) Cl., x 1000, L. Watts*
15	**Diploneis splondida** *(Greg.) Cl., x 1000, Marine Plain*
16	**Diploneis graeffii** *Grun., x 1000, L. Watts*
17	**Diploneis sejuncta** *(A.S.) Jore., x 1000, Marine Plain*
18	**Navicula grevillei** *var.* **comoidis** *(Ag.) Mill, x 1000, L. Watts*
19	**Navicula grevillei** *Ag., x 1000, L. Watts*
20	**Pinnularia quadraterea** *var.* **baitica** *Grun., x 1000, L. Watts*
21	**Pinnularia quadraterea** *var.* **constricta** *Destr. x 1000, L. Watts*

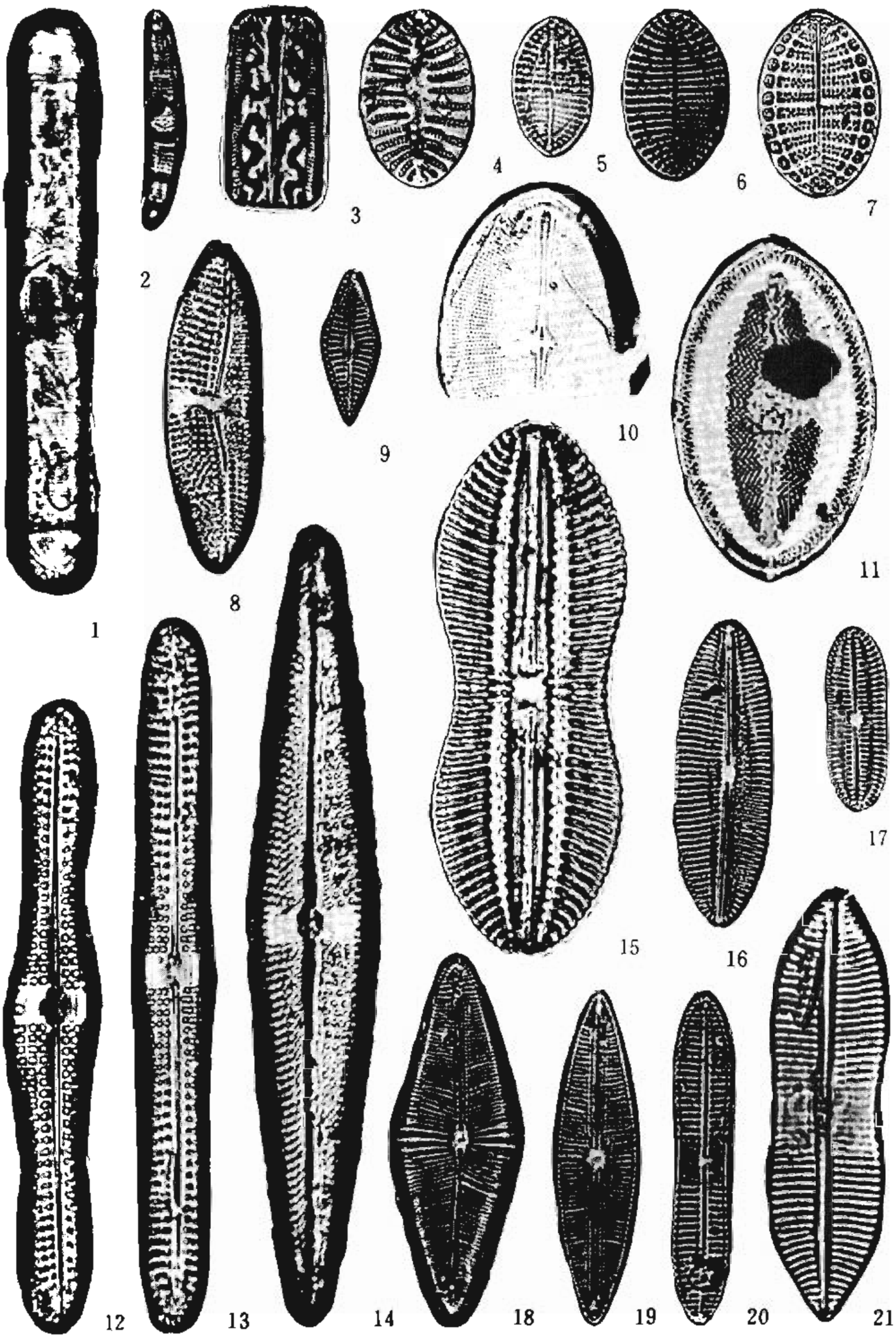
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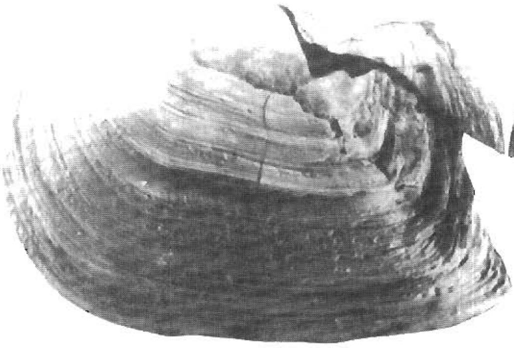
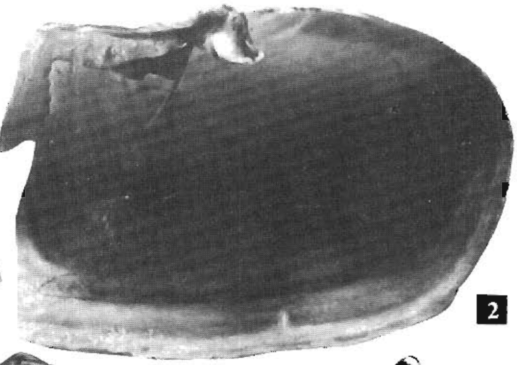
Plate 38. Lamellibranchs from Watts Lake (identified by Lan Xiou).
1,2 **Laternula** *sp. x 1*
3-6 **Axinopsida** *sp. x 4*
7-10 **Thracia meridonalis** *Smith, x 1*
11,12 **Limatula hodysoni** *(Smith), x 1*
13,16 **Laternula recta** *(Reeve), x 1*
14,15 **Laternula elliptica** *King et Broderip, x 1*





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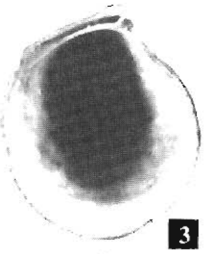

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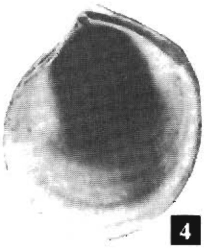

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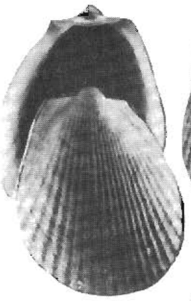

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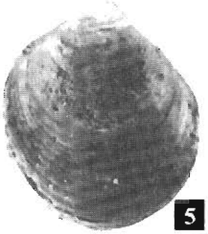

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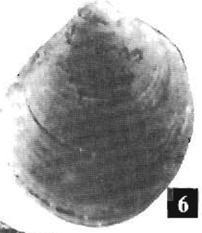

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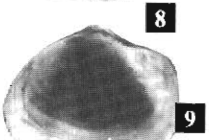

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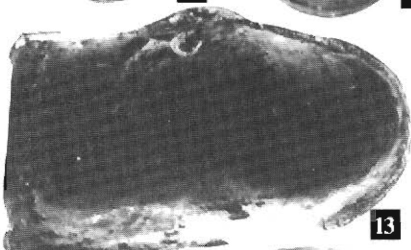

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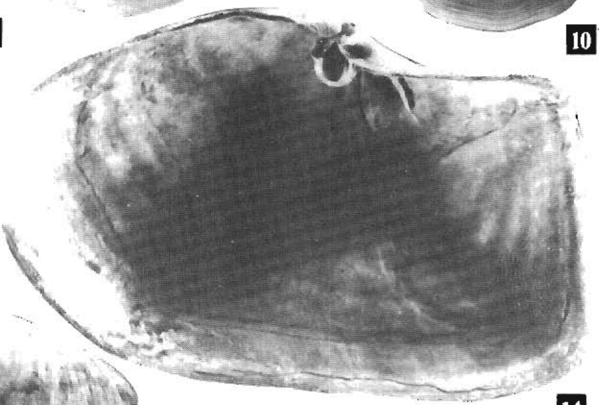

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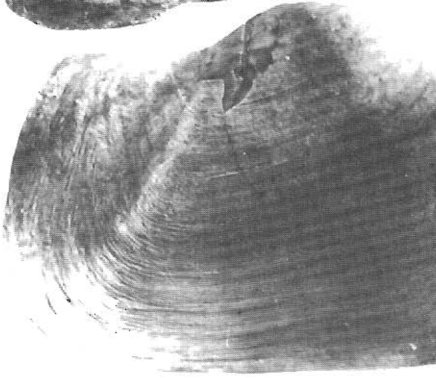

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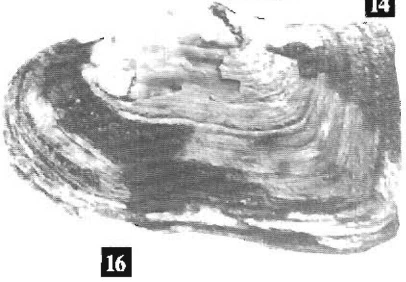

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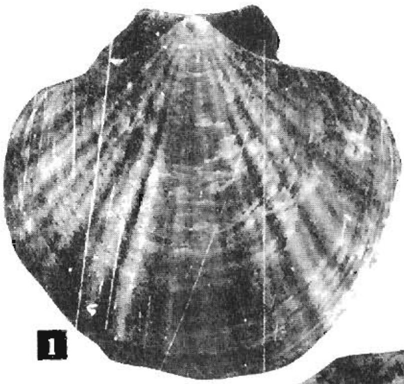
Plate 39. Lamellibranchs from Marine Plain and Deep Lake (identified by Lan Xiou) (original size).

1,2,13 **Adamussium colbecki** *(Smith), Deep Lake*
3-5,10 **Hiatella** *aff.* **arctica** *Linee, Marine Plain*
6-8,12 **Nucula obliqua** *Lamarck, Marine Plain*
9 **Chlamys** *(Athlopecten) sp., Marine Plain*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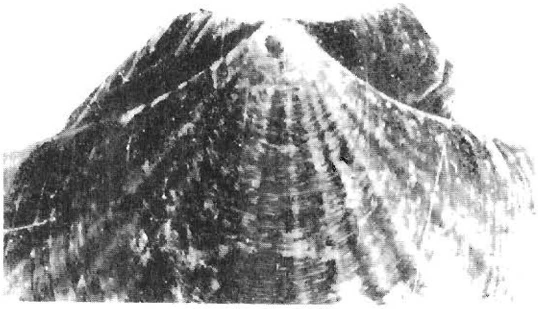
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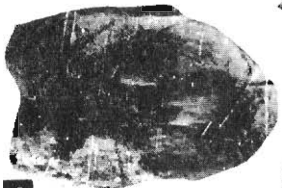
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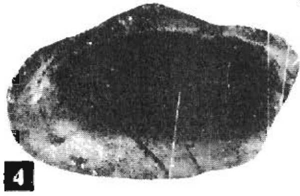
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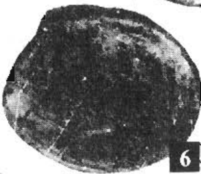
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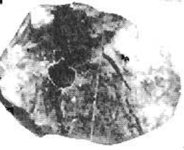
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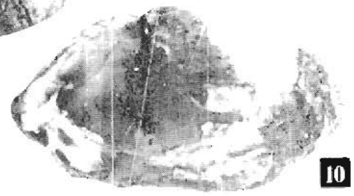
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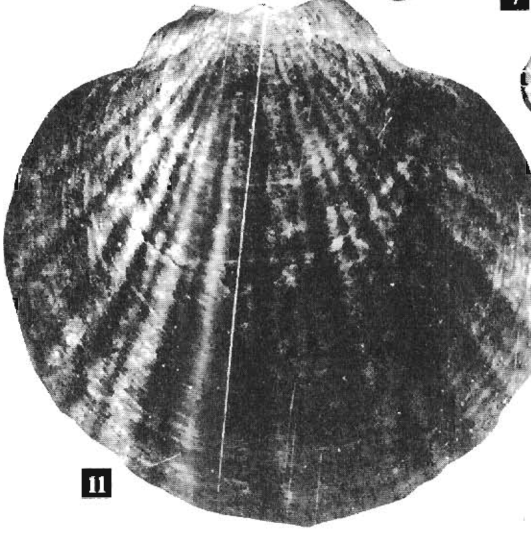
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