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Form, structure and stability of the margin of the Antarctic ice sheet, Vestfold Hills and Bunger Hills, East Antarctica

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Abstract: Studies of glacier margins have suggested that form and structure can be used to infer mass balance condition and stability. This paper examines this idea by investigating the form and structure of the Antarctic ice sheet at two coastal oases in East Antarctica. Two principal forms of the margin of the ice sheet in the Vestfold and Bunger hills that are discussed are ice cliffs and gently-sloping ice margins with an inner moraine. The form of the ice margins in both areas is primarily related to the accumulation of drift snow and superimposed ice and not to mass balance condition. It is concluded that the form and structure of ice margins are ambiguous indicators of mass balance condition and stability and that a change in mass budget is probably neither a sufficient nor a necessary condition for a change in the morphology of ice margins. Although we argue that the form and structure of the ice margins tells us little about stability, interpretation of the Holocene history and geomorphology of the oases suggests that the ice margins have been stable at least throughout the Holocene and that this condition of overall stability continues today.

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Key words: ice margin form, ice margin structure, Vestfold Hills, Bunger Hills, East Antarctica

Introduction

The form and structure of glacier and ice sheet margins reflect the dynamics and stress fields at the base of their terminal areas, and have been used as indicators of mass balance and stability (Hooke 1970, Goldthwait 1971, Chinn 1986, 1991). Although many studies have examined the physical environments of coastal oases in East Antarctica, the form, structure and stability of the ice margins have not been examined. The principal objectives of this study are to assess the stability of the ice margins at Vestfold and Bunger hills and to examine whether Chinn's (1986, 1991) model of the morphology of dry-based ice margins can be used to make inferences about mass balance and stability.

Edges of ice sheets often form cliffs, domes or gently-sloping ice ramps, each of which may have distinctive moraines (Hooke 1970, Souchez 1971). Moraines formed at the margins of ice sheets have been called "shear moraines" (Bishop 1957), "Thule-Baffin" and "inner" moraines (Weertman 1961), and to avoid genetic implications "ice-cored" moraines (Hooke 1970, Souchez 1971, Fitzsimons 1990). These moraines have been described from the Arctic in Greenland (Hooke 1970), Baffin Island (Hooke 1973a, 1973b) and Ellesmere Island (Souchez 1971). Similar moraines have been described from Antarctica near Casey Base (Hollin & Cameron 1961), southern Victoria Land (Souchez 1967) and Vestfold Hills (Fitzsimons 1990).

Formation of such moraines can be divided into three stages: 1) upwarping of debris bands so they crop out on the ice surface; 2) release of debris from the ice; and 3) dispersal of the released debris. Upwarping of debris bands in ice marginal areas has been related to longitudinal flow extension and flow compression due to deceleration caused by thermal transitions in the glacier bed (Robin 1976), to rheological transitions in the ice (Holdsworth 1969) and to bed roughness (Boulton 1970). Thermal transitions can cause flow compression when a glacier base is at pressure melting point and part is frozen to the bed. When motion changes from basal sliding to internal creep, large-scale deformation can occur. Large folds and thrust planes develop, and can result in the transportation of debris from the bed to higher levels in the ice (Boulton 1970, 1972, Robin 1976, Drewry 1986). Rheological transitions in an ice tongue were described by Holdsworth (1969) who defined four zones in an ice margin: a rigid outer zone, a semi-rigid zone; a semi-plastic zone and a plastic zone. Holdsworth (1969) and Chinn (1986, 1991) regarded the semi-rigid zone as particularly important because it formed a brittle layer about 20 m thick on the surface of glaciers. Chinn argued that, when the semi-rigid zone is grounded, it forms a stiff obstruction to flow, and determines the formation and position of the ice cliff. Depending on flow rate, the grounded zone forms a cliff, a convex, or ramped ice terminus. The effect of bed roughness is often localized and may be superimposed on the effects of thermal and rheological transitions.

Opinions are divided on whether inferences about mass balance can be made from interpretations of the form and structure of ice margins. Goldthwait (1971), suggested that ice cliffs on land are the product of advancing ice, whereas Hooke (1970) argued that a change in mass balance is neither

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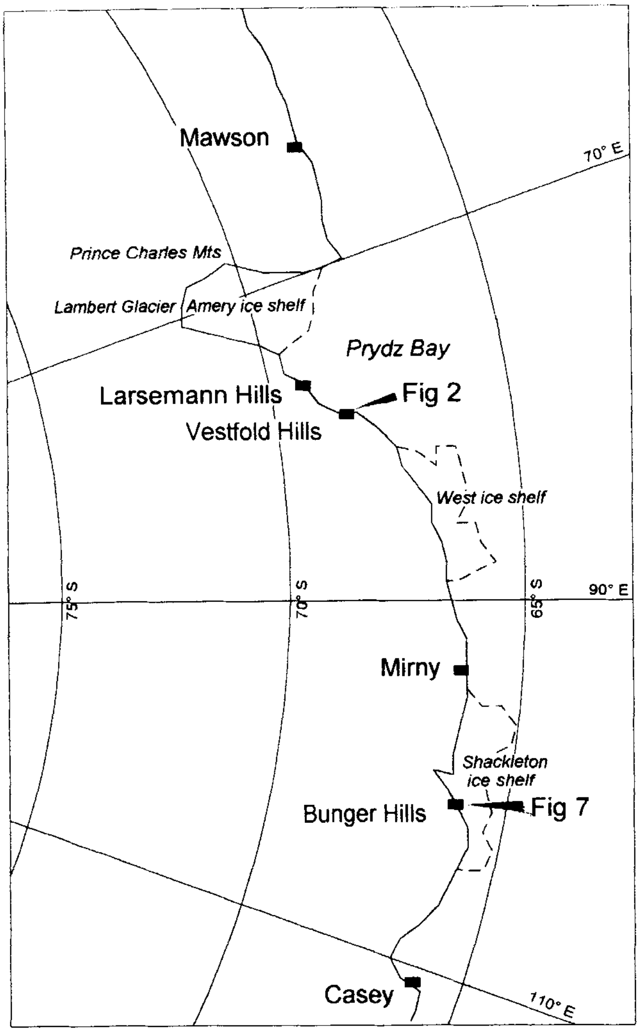

Fig. 1. Location map of the Vestfold and Bunger hills.

a sufficient nor necessary condition for a change in the form or structure of ice margins. More recently, Chinn (1986, 1991) developed a model for the dry-based glaciers of the Dry Valleys area of southern Victoria Land, Antarctica that suggests the form and structure of an ice margin is related to its stability. Chinn argued that, during advance of dry-based polar glaciers, the morphology of the ice margin goes through a number of transitional forms before developing a calving ice cliff (Chinn 1991, fig. 4). This model is partly based on Holdsworth's (1969) recognition of four different rheological zones in polar glaciers. The semi-rigid, semi-plastic and plastic zones divide gradational changes in ice behaviour with increasing depth. Holdsworth regarded the semi-rigid zone as being particularly important because, this *c.* 20 m-thick, layer contains the ice cliff (Chinn 1991, fig. 4). Where a glacier thins to less than about 20 m the more easily deformed semi-plastic and plastic layers are excluded and the semi-rigid crust becomes grounded, forming an obstacle to flow.

In Chinn's model, the pre-advance form is an ice "ramp" with an inner moraine which separates the glacier ice from the marginal snow wedge apron (Chinn 1991, fig. 4a). Ice advance leads to increased compression as the ice flows against the semi-rigid ice and the apron. Thickening of ice results and initiates steepening of the ice front to form a convex profile. As the ice thickens the basal debris zone is deformed (Chinn 1986). Large single folds developed at this stage have been associated with compression and interpreted as indicators of advance (Chinn 1989). The climax form of the advance is a calving ice cliff that overrides the thickening snow/ice apron (Chinn 1991, fig. 4c).

This paper examines the morphology and structure of the ice sheet edge at Vestfold and Bunger hills which are coastal oases in East Antarctica (Fig. 1). In both areas the continental ice sheet margin is convex and is largely obscured by ice ramps deposited as drift snow transported across the ice margin by katabatic winds. The forms and structures of the ice margins are described, compared, and contrasted with previous descriptions of ice sheet margins, and with morphological and structural models of ice advance and retreat (Hooke 1970, Goldthwait 1971, Chinn 1986, 1991).

Field observations

Vestfold Hills

The Vestfold Hills comprise a small (420 km^2) ice-free area in Princess Elizabeth Land (Fig. 2), the third largest ice free area in Antarctica after the Dry Valleys of southern Victoria Land and the Bunger Hills of Wilkes Land.

The Vestfold Hills form a complex, low-relief topography (up to 158 m a.s.l.) on gneissic rocks, and have numerous fresh and saline lakes. The southern boundary of the hills is marked by the Sørsdal Glacier, a fast-flowing outlet glacier that has a floating margin (Fig. 2). The edge of the continental ice sheet terminates for the most part in complex topography and forms the eastern boundary of the hills. The mean annual temperature of Davis Station (68°35'S, 77°58'E) is -10.2°C (Schwerdtfeger 1970), the maximum recorded temperatutre is 13°C and the minimum is -38.3°C. The climate is, on average, warmer than other Antarctic stations of similar latitude (Burton & Campbell 1980). Although there are no precipitation data, snowfall is believed to be light (<250 mm). Rainfall is rare.

The three pre-eminent characteristics of the ice margin at Vestfold Hills are its variable shape, the presence of a large sinuous ice-cored moraine (Figs 2 & 3) and the abundance of large snow drifts (Fig. 3). The ice sheet margin has a convex form and descends rapidly from 240 m within 2 km of the margin to *c.* 100 m at the margin. Where the ice margin forms the coastline it consists of 20–40 m high cliffs. On land the margin is considerably more complex, often with multiple cliffs and snow wedges (Fig. 4a). On Long Peninsula in the northern part of the hills, the ice margin is gently-sloping and merges into wide accumulations of drift snow (Fig. 3). A

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200–300 m wide ice-cored moraine inside the ice margin consists of a veneer of coarse debris and several sharp-crested ridges that are 0.5–4 m high (Fig. 4b). Some of the debris on the surface of the moraine is redistributed by meltwater streams that have formed supraglacial eskers (Fitzsimons 1991a, 1991b).

The sinuous ice-cored moraine that dominates the ice margin at Vestfold Hills is a broad, discontinuous ridge of coarse debris, 100–300 m wide and about 20 km long that occurs inside the ice margin (Figs 2 & 3). The debris is, on average, less that 0.5 m thick but accumulations up to 1.5 m thick occur on the sharp-crested ridges. The sinuous inner moraine contrasts with other moraine ridges that occur in front of ice cliffs which comprise sharp-crested ridges. Moraine ridges beyond the ice margin are much higher (up to 20 m) and much shorter than the inner moraine ridges (less than 1 km long). Most are ice-cored and highly unstable, as indicated by the occurrence of numerous sediment flows, slumps and other mass movements (Fitzsimons 1990).

On Broad Peninsula the ice margin is buried by snow drifts and terminates in a low-angle ramp. Where the margin is not buried cliffs up to 30 m high occur near the ice-cored moraines and at the heads of fjords (Fig. 5a). In these steeper sections the moraine material is concentrated below the ice cliffs (Fig. 5a & b) forming narrow, sharp-crested ridges up to 10 m high. A series of large gullies cut into the ice lie approximately 1 km inside the ice edge.

In the south-east corner of the hills, where the Sørsdal Glacier forms a distinct ice stream, the ice margin has a convex profile and has multiple ice cliffs (Fig. 4a). The slightly deformed, basal debris zone is unconformably overlain by clean white ice. This unconformity appears to record a former ablation surface that has been buried by ice which accumulated *in situ* (superimposed ice). A zone of ice-cored moraine with numerous sharp-crested ridges parallel to the ice edge occurs beyond the main ice cliffs.

The structure of the ice margin at the Vestfold Hills is revealed by exposures of the basal debris zone in ice cliffs and gullies that cross the ice margin. Deformation structures range from relatively undeformed debris bands (Fig. 4a) to intense deformation that is commonly exhibited as recumbent folds and shear structures (Fitzsimons 1990, fig. 4). Faults are relatively rare which suggests that most basal ice deformation is plastic and occurred beneath at least 20 m of ice in the 'semi plastic' or 'plastic' rheological zones (Holdsworth 1969). Deformation structures in the basal zone of the ice cap can be divided into large-scale features, which involve the entire basal debris zone, and small-scale features which occur within the basal zone. The most prominent large-scale deformation structure is the upwarping of the basal debris zone to crop-out on the surface of the glacier and form a large, sinuous ice-cored moraine (Figs 2 & 3). In an ice gully near the head of Long Fjord the upwarped basal debris zone shows a vertical contact between the debris zone and the clean superimposed ice (Fig. 6). Folds in the

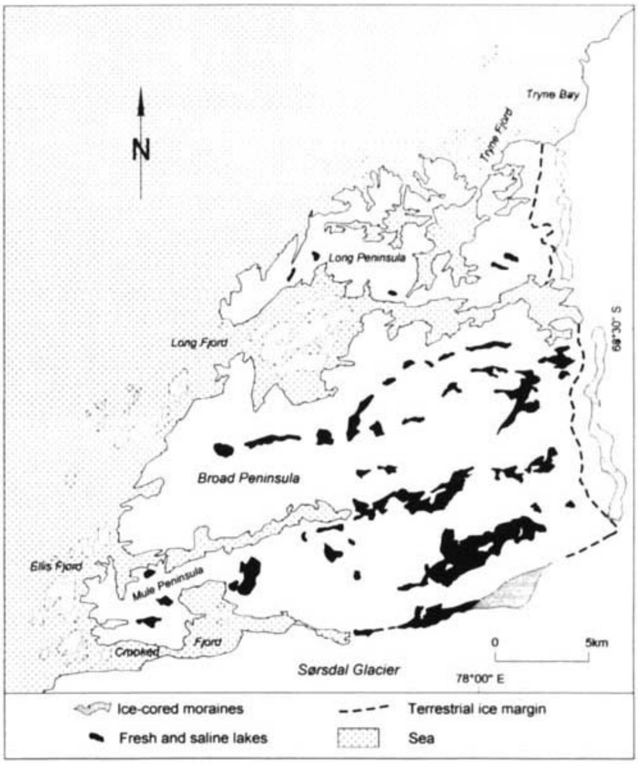

Fig. 2. Map of the Vestfold Hills showing the form of the ice margin, the position of a debris band inside the apparent ice edge, the main accumulations of ice-cored moraine, and the locations mentioned in the text.

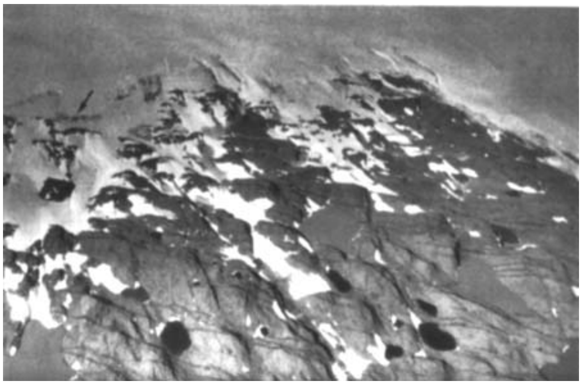

Fig. 3. Oblique aerial photograph of south-east corner of Vestfold Hills. The ice margin is partially obscured by drift snow and marked by large gullies. A partly exposed sinuous ice-cored moraine is shown on the left (arrow) and the cliffed margin of the Sørsdal Glacier is shown on the right. The photographed area is 6 km wide at the top of the photograph.

superimposed ice on the right of Fig. 6 suggest it was deformed as the ice flowed into the marginal snow wedge.

A section through an ice-cored moraine in the south-east corner of the hills shows that moraine ridges can form along the axes of a series of large recumbent folds that have

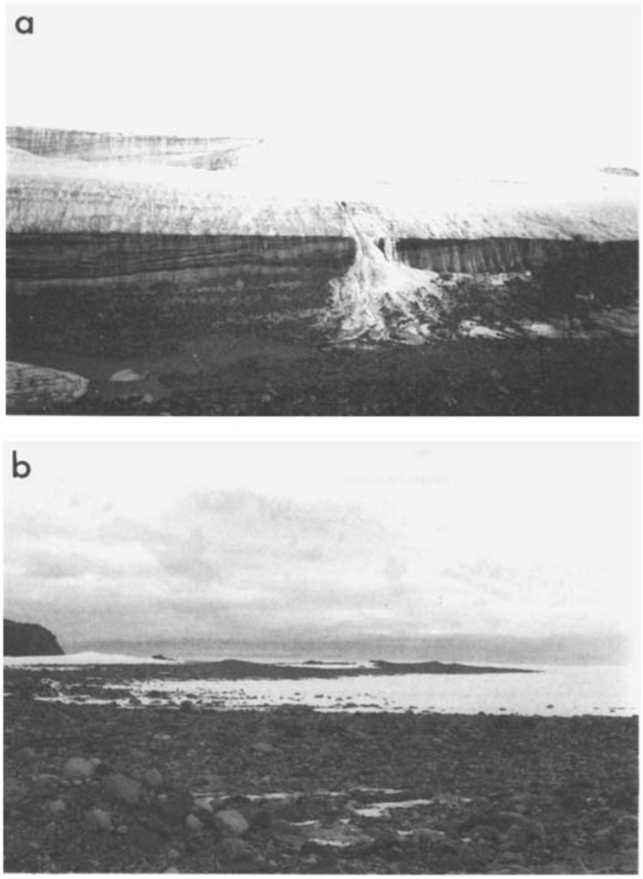

Fig. 4. Form of the Vestfold Hills ice margin. **a.** Multiple cliffs showing the snow wedge overlying foliated glacier ice. The lower cliff is about 18 m high. **b.** Sinuous ice-cored moraine on Long Peninsula. The debris cover is thin (0.5–1.5 m) and the maximum relief of the ridge is 4 m.

amplitudes over 15 m (Fitzsimons 1990, fig. 4c). Deformation structures within the basal debris zone consist of small-scale folds, shears and small parasitic faults on major folds (Fitzsimons 1990, fig. 4); the structures highlight the dominance of plastic deformation in the dynamics of the ice margin.

Measurements of debris concentrations in the basal zone ice are consistently below 10% by volume (Table I). Debris concentration of individual bands is highly variable with most of the debris concentrated close to the bed. Unusually high concentrations occur in rare debris lenses (solid subfacies of the ice facies scheme used by Lawson 1979, Table I) of sorted fluvial sediments that have been entrained by the glacier. Most of the debris consists of silt- and sand-sized particles with larger clasts either dispersed or occurring in small lenses. Gravel clasts are dominantly subrounded, rarely angular.

Bunger Hills

The Bunger Hills, also known as the Bunger oasis, have a land area of 472 km^2 (Wisniewski 1983). The major part of the land area (263 km^2) occurs in the south where the hills abut the Antarctic continental ice sheet (Fig. 7). The summits form part of a northward sloping erosion surface that was dissected during the Cainozoic to produce steeply sloping hills and many ice-scoured lakes. The largest, which include lakes Figurnoe and Dolgoe, have been eroded along major faults in the granite, gneiss and migmatite rocks; the smallest are shallow lakes formed by ice scouring.

The Bunger Hills are enclosed by ice being bounded by the Antarctic ice sheet on the south-east, the Apfels outlet glacier on the south and south-west, the Scott Glacier and Edisto Ice Tongue in the west, the Shackleton Ice Shelf in the north and the Remenchus Glacier in the east (Fig. 7).

The oasis forms a heat island in which mean temperature is approximately -9°C. Although there is no reliable record of precipitation, the area is very dry (< 200 mm water equivalent p.a.) due to the presence of the Shackleton Ice Shelf which reduces the incursions of maritime air from the east.

The Antarctic ice sheet surface decreases rapidly in height from 300 m within 1.5 km of the ice margin to approximately 120 m at the margin where it terminates against a blocking ridge of hill summits that vary from 131 m in the north through 158–132 m in the south-east. The ice flows extremely slowly against the south-eastern hills at < 1 m yr^{-1}.

A distinct boundary zone south of Lake Dalekoe separates the Antarctic ice sheet from the Apfels Glacier flowing westward along the southern margin of the hills at 113 m yr^{-1} near the centre of the glacier. However, judging by the absence of crevasses and stability of ice/water levels in ice marginal lakes, the ice flow must be much slower adjacent to the southern margin of Bunger Hills. To the west, the Apfels Glacier is deflected northwards by the large fast-flowing Scott Glacier which occupies a 700–1300 m deep trough. West of Lake Polyanskiy, Scott Glacier flows at 458 m yr^{-1} and the rates increase northward to 1280 m yr^{-1} (Dolgushin 1966).

The Antarctic ice sheet descends to sea level east of the Bunger Hills at Kapakon Inlet and Luchistaya Bay (Fig. 7). The ice edge probably terminates close to the boundary between the land and the deep marine inlets, to judge by the strongly calving ice cliffs up to 40 m high, and the islands of Geomorphologists and Skalistyi peninsulas to which the ice sheet edge is tied. Fragments of end moraine and thick perennial snowbanks occur on the windward sides of the peninsulas but owing to difficulties of access were not visited.

Farther south between Kapakon Inlet and Lake Dalekoe, where the ice edge changes trend from SSE–W, the ice sheet edge is marked by a discontinuous line of moraine that extends for 17 km with the longest sections being 1.5–2.5 km. The moraine varies from 30–125 m in width (Wisniewski 1983), and is a relatively sharp-crested sinuous ridge (Adamson & Colhoun 1992, fig. 9). It extends into the embayments between the hills and curves around the intervening hillslopes and spurs that protrude into the ice

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sheet margin.

The moraine is up to 10 m high in places and mainly consists of a thin covering of boulders of about 200–1000 mm size, with rare blocks of rock up to 2000 mm diameter resting on the ridged ice surface. The boulders are predominantly subangular in shape and are strongly polished, facetted and pitted by the wind. Some have thin crusts of hydrated iron oxide that indicate either the moraine has been in place for a long period of time or that the boulders have been derived from a pre-weathered rock surface by ice erosion.

The 130–160 m hills block mainly the northward flow of ice and only in a few places does the ice penetrate through low cols in this local watershed. An outlet lobe of ice extends almost to Schel Inlet and terminates in a gently-sloping ice margin above a shallow marginal lake. A terminal ice lobe flows through a col 3 km north-east of the head of Lake Figurnoe and reveals the structure of the ice edge (Fig. 8a). Lake Figurnoe extends to the ice edge where the lake has produced small cliffs.

West of Lake Dalekoe, a small lobe of ice extends northward and ends in a gully cut by meltwaters draining from the ice surface. Here, and farther west where the ice margin ends in two small lakes south of lakes Ptich'e and Burevestnik, the ice margin is locally cliffed. Farther west, the Apfels Glacier surface declines gently north-westward from 120–60 m and the ice edge remains below the southern rim of hills as far as tidal Lake Polyanskiy. The ice-rock margin is largely masked by drift snow but two small ice-marginal lakes are impounded between the low-gradient ice edge and the steep hill slopes. Little to no marginal moraine occurs along the boundary of the Apfels Glacier with the southern hill slopes.

South of tidal Lake Polyanskiy a 30–40 m length of ice front consists of a 10–15 m vertical cliff from which masses of ice have slumped and become stranded on the shallow floor of the marine inlet. These slumped masses have partly protected the upper ice slope which has a planar ramp to slightly convex form that is locally incised by shallow meltwater runnels.

Details of structure of the ice margin were best observed in ice gullies recessed into the ice margin by summer surface meltwater erosion. In an ice gully 3 km east of the head of Lake Figurnoe the contact between the ice sheet margin and the blocking masses of recrystallized snow and ice showed deposition and over-riding of a 5 m high and 10–15 m wide sharp-crested moraine. In addition, the ice was being deformed upwards and debris was being transported towards the ice surface by compressive ice flow.

Nearby, ice flow through the col has produced a large recumbent fold overlain by a snow wedge (Fig. 8a). The occurrence of the recumbent fold probably depends largely on the existence of the narrow col through which the ice flow is constricted. Where the ice front has degenerated to a low cliff highly encumbered with debris, the ice and debris are strongly folded on a small scale (Fig. 8b). The dominance of both large- and small-scale folds, and the scarcity of shear structures points to deformation of the ice in the marginal zone by semi-

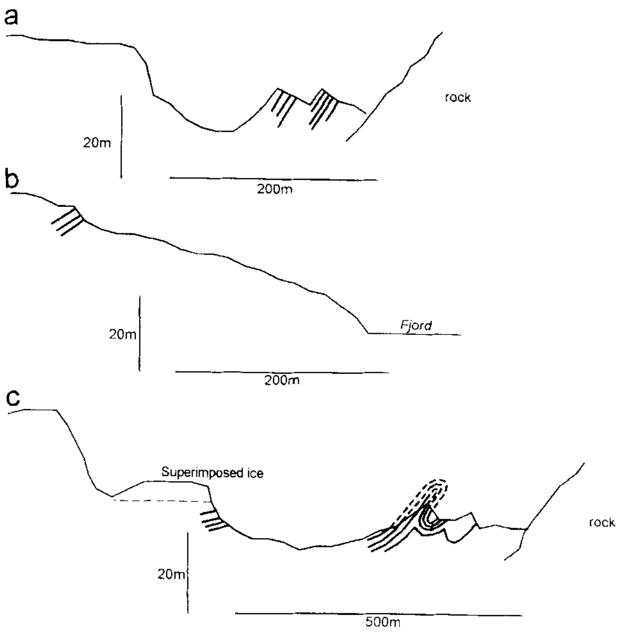

Fig. 5. Sections of the ice margin based on levelled profiles and field mapping. **a.** A cliffed ice margin and ice-cored moraines on Long Peninsula. **b.** A gently-sloping ice margin with inner moraines developed where the deformed basal debris zone crops out. **c.** Multiple ice cliffs and moraines formed on the axes of large single folds in ice the south-east corner of the Vestfold Hills at the margin of Sørsdal Glacier.

Fig. 6. Steeply-dipping, deformed debris bands near the head of Long Fjord resting against superimposed ice to the right. The cliff is 20 m high.

plastic or plastic flow.

That the ice margin has advanced recently through this col, before retreating locally due to surface meltwater erosion, is supported by the strongly hydrated iron oxide weathering of the gneiss underlying the moraine ridge that crosses the gully floor some 50 m from the ice cliff. However, given the topographic confinement of ice flow through the col it is not

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Table I. Debris concentrations in ice from cliffs at the Vestfold Hills ice margin.

Ice facies[1]	Location	% debris by volume
Stratified, discontinuous subfacies	SE corner	2.3
Stratified, discontinuous subfacies	SE corner	2.1
Stratified, discontinuous subfacies	SE corner	1.2
Stratified, discontinuous subfacies	Long Fjord	2.2
Stratified, discontinuous subfacies	Long Fjord	1.9
Stratified, discontinuous subfacies	Long Fjord	1.5
Stratified, discontinuous subfacies	SE corner	2.8
Stratified, discontinuous subfacies (deformed)	SE corner	5.8
Stratified, discontinuous subfacies (deformed)	SE corner	8.0
Stratified, discontinuous subfacies (deformed)	SE corner	6.5
Stratified, solid subfacies	SE corner	17.9
Stratified, solid subfacies	SE corner	27.8
Stratified, solid subfacies	SE corner	9.5
Stratified, solid subfacies	SE corner	7.8
Stratified, solid subfacies	SE corner	4.5
Average		**6.8**

[1] Ice facies based on the work of Lawson (1979).

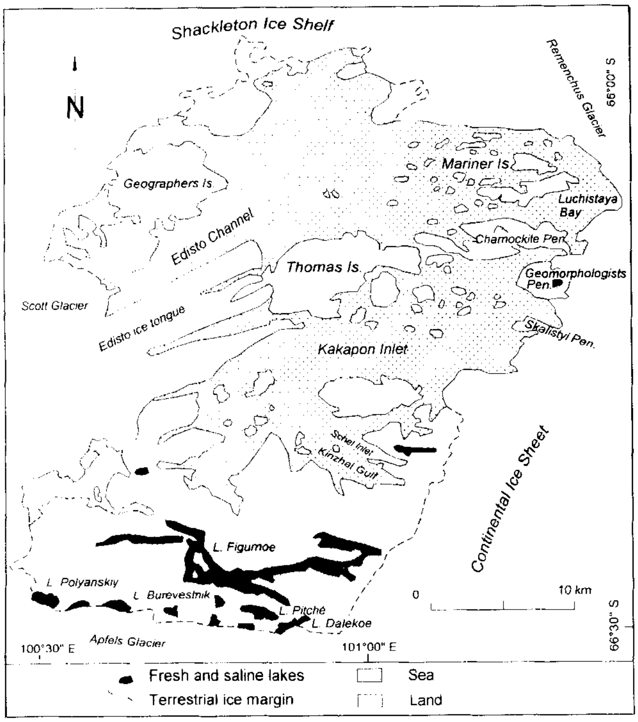

Fig. 7. Map of the Bunger Hills area showing the ice-bound nature of the oasis, the edge of the Antarctic ice sheet, the Apfels Glacier and the location of places mentioned in the text.

possible to extrapolate this interpretation to the entire ice edge.

The ice margin transports only a very small quantity of debris which is horizontally stratified and occurs mainly in the lower 2 m above bedrock. Evteev (1960) measured the quantity of debris contained in a 40 m face of banded ice in an ice gully in the Bunger Hills. He recorded 11.8% by volume debris in the basal 3 m, between 1 and 4.5% in the next 12 m and only 0.5% in the surface 25 m. Thus, nearly all the debris at this site is transported in the basal few metres of ice.

Observations from air photographs of the two ice marginal lakes east of Lake Polyanskiy show no evidence of crevassing and no evidence of recent changes in water/ice surface level. Such lakes, impounded between the ice edge and hill margins, would be very sensitive to either lowering of the ice surface or marginal retreat. If such had occurred recently, water-level benches would have been cut in the ice. There is no evidence for any change in water level.

Discussion

The margin of the continental ice sheet in the Vestfold and Bunger hills assumes two principal forms, each associated with a different combination of processes. The two forms are gently-sloping ice margins with an inner moraine, and ice cliffs.

The dominant forms in both areas are gently-sloping ice margins that are primarily related to the accumulation of drift snow and superimposed ice. Accumulation of drift snow is a consequence of relatively high coastal snowfall, downslope transportation of snow by katabatic winds, redeposition when the snow reaches the increased roughness of the ice margin and bedrock hills of the coastal oases, and the wind shadow cast by the convex ice margin and the ridges of inner moraine. Superimposed ice forms in the ice marginal area as a consequence of the accumulation of drift snow, snow metamorphism and *in situ* melting and refreezing of the snow. Extensive accumulation of superimposed ice at the ice margins highlights the complexity of determining the mass balance of continental ice sheets. Marginal areas cannot be regarded simply as ablation zones because substantial accumulation can occur in low altitude coastal locations.

Ice cliffs are uncommon in Vestfold and Bunger hills except where ice terminates at sea, in coastal inlets, and where lakes occur adjacent to the ice edge. In these circumstances cliffs are maintained by thermal erosion at the base and/or wave action combined with enhanced ablation that is a characteristic of steep ice slopes (Chinn 1990).

Gently-sloping ice margins with well-defined inner moraines inside an apparent ice edge are morphologically and structurally similar to Chinn's ramp margin. If the application of Chinn's model is accepted it could be suggested that much of the terrestrial continental ice margin at Vestfold and Bunger hills is in a stable condition. However, convex ice margins, together with evidence for the development of large single folds in the Vestfold Hills (Fitzsimons 1990, fig. 4) and Bunger Hills (Fig. 8a) could be interpreted as indicating an advancing ice margin. Although terrestrial, cliffed ice margins are rare in the Vestfold and Bunger hills they do occur in some locations at the margin of the continental ice sheet and at the margins of outlet glaciers. Using Chinn's model these cliff sections could also be interpreted as indicating advance of ice in both oases. Although it is possible to use Chinn's model as outlined above, several questions about the basis of the model need to be addressed: a) the role of drift snow and superimposed ice in the form, structure and stability of the ice margin, b) whether single folds are an unambiguous indicator of ice advance, and c) whether a particular form and/or structure can be unambiguously associated with a particular mass balance condition.

Chinn's model is largely based on studies of small dry-based alpine glaciers in the Dry Valleys of southern Victoria Land, Antarctica. This area is a polar desert with a mean annual temperature of -19.8°C (Schwerdtfeger 1984) and very light snowfall (*c.* 10 mm water equivalent per year, Chinn 1990). Most of the area has a moisture deficit, drift snow does not form significant perennial accumulations in the valley floor and superimposed ice is not significant (Souchez 1968). Thus, the model was developed in an area where accumulations of drift snow and superimposed ice do not complicate the form and structure of the dry-based polar ice margins. By contrast, in the polythermal ice margin of coastal East Antarctic oases, drift snow and superimposed ice conceal much of the ice margin and dominate the form of the terminal areas. Our field observations suggest that gently-sloping ice margins are primarily the product of drift snow accumulation and the *in situ* formation of superimposed ice.

Evidence for deformation of the basal debris zone into large single folds is seen in the core of a moraine that lies between the ice edge and rock ridges in south-east Vestfold Hills (Fitzsimons 1990, fig.4) and where the ice flows through cols in the Bunger Hills (Fig. 8a). Although these large recumbent folds could record advancing ice, our observations suggest that large-scale folding and the development of recumbent folds may be associated with variations in subglacial topography and are thus an ambiguous indicator of ice advance, particularly in regions where an ice sheet margin terminates against obstructing rock ridges.

Examination of the morphology and structure of the ice margins of the Vestfold and Bunger hills suggest that continuous drift snow accumulation will result in the maintenance of a gently-sloping ramped ice margin irrespective of whether the ice margin is retreating or advancing. Furthermore, the persistence of gently-sloping ice margins that are adjacent to and grade into ice cliffs suggest that the forms may be produced by localized flow conditions possibly related to subglacial topography and thermal erosion by sea water and ice marginal lakes.

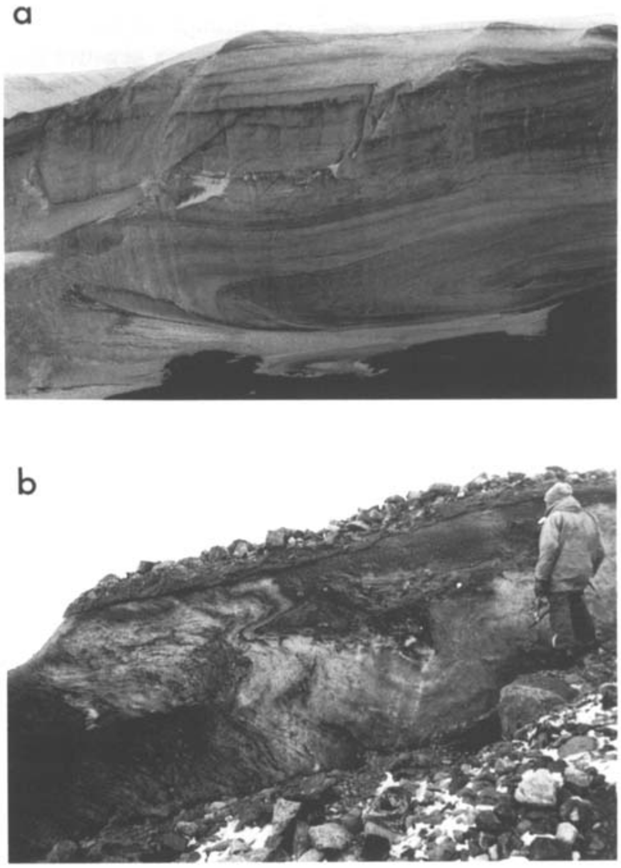

Fig. 8. *Ice margin features of the Antarctic ice sheet and the* Apfels Glacier, southern Bunger Hills. **a.** Where the ice flows through a gap in the hills east of the head of Lake Figurnoe, it is folded forward into a recumbent fold. Note the glacier ice is overlain by snow drifts. The lower drift appears only on the left (north) and has a sharp contact with the underlying glacier ice. The cliff is about 25 m high. **b.** Where the ice is thin and frozen to bedrock, retarding its advance, and where encumbered with considerable debris, it is strongly folded at the margin.

Ice cliffs

The key to understanding the relationship between form and mass balance condition is whether ice cliffs are characteristic of advancing ice. Cliffs form where basal velocities are significantly lower than surface velocities, a condition that is satisfied at most polar glacier margins. Cliffs also form where ice is removed by processes that generate or maintain a cliff form. Field observations show that such processes include thermal erosion by ice marginal lakes, thermal and wave erosion by the sea water and accelerated ablation where sun angles are low or ice is covered by a thin debris film. Regardless of the process, there is a tendency for cliffs to form at the margins of polar glaciers in a variety of situations, on land, at land/water interfaces and in water. From a ten-year study of a terrestrial ice cliff in Greenland, Goldthwait (1971)

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suggested that cliffs on land are the product of advancing, possibly surging, ice. However, throughout the ten-year study cliff forms were maintained despite a negative mass balance condition. In a study of the mass balance, morphology and structure of an ice margin near Thule, Greenland, Hooke (1970) found that drift snow and superimposed ice are important controls on the morphology of the ice margin. He suggested that a change in the principal morphology from a ramp to an ice cliff requires a change in the pattern of drifting snow. He concluded that a change in mass balance condition is neither a sufficient nor necessary condition for a change in the morphology and structure of an ice margin.

From our observations and analysis of the ice margin at Vestfold and Bunger hills, together with consideration of the work of Goldthwait (1971), Hooke (1970) and Chinn (1986, 1991), it is apparent that the morphology and structure of ice margins are ambiguous indicators of mass balance condition and stability. In particular, there is a problem with using inferences based on form and structure where drift snow and superimposed ice form a significant component of ice-margin accumulation. Study of the ice morphology of the Vestfold and Bunger hills suggests there is no specific evidence that supports interpretation of the ice cliffs as an indicator of ice advance. However, geomorphological evidence can be used to examine the stability of the ice margins.

Is the ice sheet advancing or retreating?

At Vestfold Hills the position of ridges outside the apparent ice edge (Fig. 5a) implies retreat from a former position. However, this association is rare and for the most part the continental ice sheet edge is a ramp with no ridges beyond the apparent ice edge. Boulders on the inner moraine, and rocks that protrude from the superimposed ice adjacent to the inner moraine are colonized by large lichens up to 120 mm in diameter, suggesting that the inner moraine has been stable for at least several hundreds of years and that the ice has not retreated recently although lichens could survive a cold ice cover. In the Bunger Hills, narrow zones of unweathered rock 10–50 m wide in the ice gullies indicate that local retreat has occurred. By contrast, the burial of a strongly oxidized gneiss surface beneath unweathered moraine, and the occurrence of weathered gneiss, granite and migmatite surfaces up to the ice margin indicate that slight advance is more likely than a small general retreat. In both areas, the contrast between the low debris concentrations in the ice, and the relatively large quantities in the broad, extensive end moraine formed in the ice sheet-snow wedge contact zone suggest that many millenia of nearly steady state conditions were required to form the moraine. It is difficult to envisage how either rapid advance or rapid retreat of the ice sheet margin would not destroy the moraine, especially as its very existence seems to require the interaction of a thin ice margin flowing against the rising hill slopes.

At the margin of Sørsdal Glacier, a discontinuous series of ice-cored moraine ridges record fluctuations of the ice edge. The position of the moraines imply recent retreat from the position of the outermost moraine, about 500 m from the ice edge. Radiocarbon dating of shells in glaciomarine sediment in the moraines gave ages of 9920 ± 100 yr BP (SUA 2924) from a ridge about 20 m from the ice edge, 5070 ± 80 yr BP (SUA 2923) from a ridge about 40 m from the ice edge and 2010 ± 110 yr BP (SUA 2922) from a ridge about 500 m from the ice edge (Fitzsimons 1991b). The pattern of the radiocarbon dates appears complex because the oldest date comes from material in the youngest moraine. This can be explained by episodic erosion and redeposition of the fjord-bottom sediment as the outlet glacier fluctuates (Fitzsimons & Domack 1993). The dates therefore do not reflect the age of the moraines but record glaciomarine sedimentation when marine organisms grew under floating ice or in open water. Using an Antarctic reservoir correction of 1300 years the three dates suggest that the margin of the Sørsdal Glacier was at or south of its present position by 8600 BP and that the entire set of ridges post dates 700 BP. These moraines and dates suggest that the most responsive part of the ice edge, the outlet glacier, has been relatively stable throughout much of the Holocene.

At the southern edge of the Bunger Hills the relative absence of moraine on the margin of the Apfels Glacier argues strongly for the ice margin having been in almost steady state for a long time. The presence of ice marginal lakes with no evidence of water level changes and the absence of crevasses along the margin supports the suggestion that the ice margin has been in an approximately steady state condition for a long time.

Study of raised beaches (Colhoun & Adamson 1992, Adamson & Colhoun 1992) in the Bunger Hills has shown that at Schel Inlet only 2 km from the ice sheet margin, and < 1 km from a short outlet glacier with a domed front, the ice retreated before 5600 yr BP (^{14}C age on shells of 6880 ± 160 yr BP (Beta 15831) corrected for reservoir effect by 1300 yr). The date provides a minimum age for post glacial marine transgression. Recently, radiocarbon dates of 9850 ± 600 yr BP (LU 1984) and 10070 ± 80 yr BP (LU 2932) at Lake Figurnoe show that the local Antarctic ice sheet edge must have retreated at least as far south as its present position before 10000 years ago (Bolshiyanov *et al.* 1991). The radiocarbon dating indicates that, since the ice edge has occupied approximately its present position throughout the Holocene, it is likely that the moraine, and the form and type of structural features observed at the ice margin today are characteristic of those formed under prolonged approximately steady state conditions.

Conclusions

Given the structural and morphological complexity of Vestfold and Bunger hills ice margins, the significant role played by drift snow and superimposed ice together with consideration of the literature we conclude that ice margin form and

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structure are ambiguous indicators of mass balance condition and stability. A change in mass budget is probably neither a sufficient nor necessary condition for a change in the morphology of ice margins (Hooke 1970). Nevertheless, a combination of our observations of the ice margins together with interpretations of the geomorphology and Holocene history of the oases suggests that parts of the ice edge in both the Vestfold and Bunger hills may be either retreating or advancing locally. These fluctuations appear to be of very limited extent (< 500 m). Overall, the evidence suggests that the ice margins have been stable at least throughout the Holocene and that this picture of overall stability continues today.

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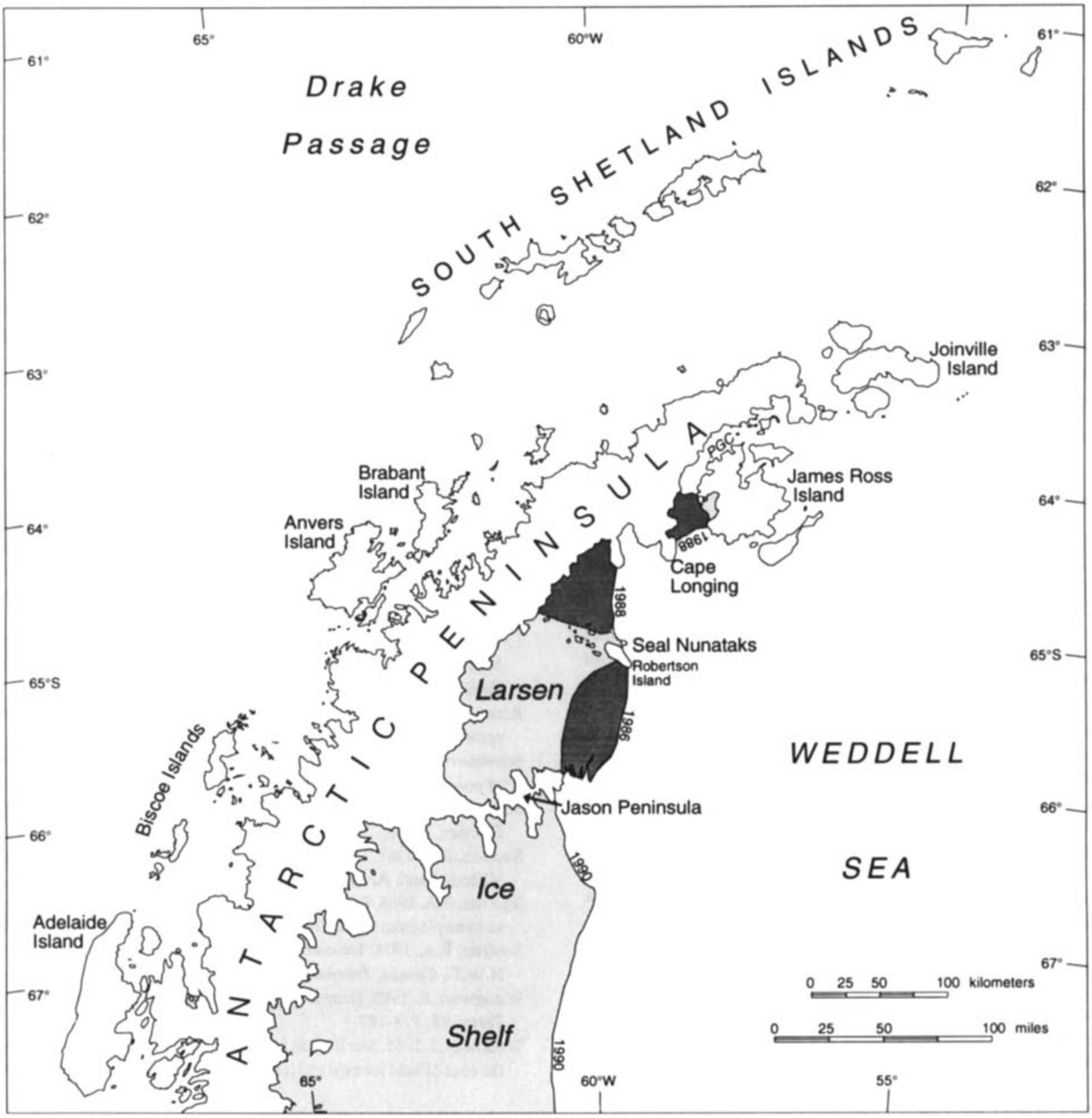
***Antarctic Science* Handy Atlas Map No. 12.**
Northern Antarctic Peninsula and Larsen Ice Shelf, derived from the 1:3 000 000 scale generalized dataset of the Antarctic Digital Database. Reproduced at 1:5 000 000 scale, with new ice front information added.

The map shows recent major reduction (dark shading) in the area of ice shelves adjacent to north-eastern Antarctic Peninsula. Analysis of NOAA satellite imagery indicates that Sjögren Glacier Tongue and its adjacent ice shelf, which once joined the peninsula to James Ross Island, disintegrated sometime before December 26 1994. The Larsen Ice Shelf, north of Seal Nunataks was in an advanced stage of breaking up by early February 1995. At about the same time, the area between Robertson Island and Jason Peninsula calved to produce a large iceberg 78 km x 37 km in extent and perhaps 200 m thick; other N-S cracks in this area of the Larsen Ice Shelf may result in the production of further icebergs in the near future. [Data: M.R.A. Thomson].



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