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MORPHOLOGY OF THE ICE-SHEET MARGIN NEAR THULE, GREENLAND

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ABSTRACT. Three types of glacier margin are found along the edge of the Greenland ice sheet near Thule: ice cliffs, ramps and ice-cored moraines. Where the glacier margin is perpendicular to prevailing katabatic winds, drifting snow accumulates along it in stagnant wind-drift ice wedges. Upward flow of active ice behind these wedges causes ice originally near the base of the glacier to rise to the surface. Where this basal ice is free of debris, a gently sloping ramp develops. However, where the basal ice contains sufficient debris, a layer of till accumulates on the glacier surface. Ice beneath the till is insulated and a debris-capped ice ridge or ice-cored moraine forms. Ice cliffs occur where the ice-sheet margin is parallel to prevailing winds and is thus swept clear of drifting snow. Although the ice sheet in the Thule area appears to have had a negative mass balance for many years, all three types of glacier margin are believed to be equilibrium forms that can develop and persist on a glacier with a balanced mass budget.

Foliation in wind-drift ice wedges generally dips down-glacier but foliation in active ice dips up-glacier. It is inferred that foliation in the wedges was once sedimentary stratification that has been tipped upward and locally overturned.

RÉSUMÉ. *Morphologie de la bordure de l'indlandsis près de Thule au Groenland.* On rencontre trois types de limite de glacier le long de la bordure de l'indlandsis du Groenland près de Thule: des falaises de glace, des talus, des moraines au centre de glace. Là où la limite du glacier est perpendiculaire aux vents katabatiques dominants, de la poudrerie au sol, poussée par le vent, s'y accumule, y formant descoins de glace stagnante. Un courant de glace active se dresse derrière ces coins et oblige la glace qui était près de la base du glacier à s'élever vers la surface. A l'endroit où cette glace de la base ne contient pas de débris, un talus en pente douce se développe. Cependant, là où la glace de la base contient suffisamment de débris, une couche de dépôt glaciaire (till) s'accumule sur la surface du glacier. La glace en dessous du till est isolée, et une crête de glace couverte de débris ou une moraine au centre de glace se forme. Les falaises de glace apparaissent lorsque la bordure de l'indlandsis est parallèle aux vents dominants qui, de ce fait, la nettoient de poudrerie au sol. Bien que le bilan de masse de l'indlandsis dans la région de Thule semble avoir été négatif depuis plusieurs années, on croit que les trois types de limite de glacier sont des formes d'équilibre, qui peuvent se développer et persister sur un glacier avec un bilan de masse équilibré.

La foliation dans le cas des coins de glace accumulés par le vent s'incline en générale vers l'amont, mais la foliation dans le cas de la glace active s'incline vers l'aval. Il est impliqué que la foliation des coins des glace fut une stratification sédimentaire qui a été poussée vers le haut et en parties retournée.

ZUSSAMMENFASSUNG. *Die Morphologie des Inlandeisrandes bei Thule, Grönland.* Drei Gletscherrandtypen sind an der Grenze des grönländischen Inlandeises bei Thule zu finden: Eiskliffe, Rampen und Moränen mit Eiskernen. Wo der Gletscherrand senkrecht zu den vorherrschenden katabatischen Winden verläuft, häuft Treibschnee in der stagnierenden Winddrift Eiskeile an. Das Aufsteigen aktiven Eises hinter diesen Keilen bringt Eis aus seiner ursprünglichen Lage nahe am Gletschergrund an die Oberfläche. Wo dieses Grundeis schüttfrei ist, entwickelt sich eine sanftgeneigte Rampe. Wo jedoch das Grundeis ausreichend Schütt enthält, häuft sich eine Moränenschicht auf der Gletscheroberfläche an. Das Eis unter der Ablagerung ist isoliert und bildet einen schüttbedeckten Rücken oder eine Moräne mit Eiskern. Eiskliffe kommen dort vor, wo der Eisrand parallel zur vorherrschenden Windrichtung verläuft und daher von Treibschnee freigefegt wird. Obgleich die Massenbilanz des Inlandeises im Thulegebiet seit vielen Jahren negativ gewesen zu sein scheint, wird angenommen, dass alle drei Gletscherrandtypen Gleichgewichtsformen sind, die sich auf einem Gletscher mit ausgeglichenem Massenhaushalt bilden und fortbestehen können.

Die Bänderung in Drifteiskeilen fällt generell gletscherabwärts ein, in aktivem Eis dagegen gletscheraufwärts. Es wird vermutet, dass die Bänderung der Keile einmal eine sedimentäre Schichtung war, die aufwärts gebogen und stellenweise überkippt wurde.

INTRODUCTION

The margin of the continental ice sheet near Thule in north-west Greenland assumes three distinct forms, each of which reflects a different combination of processes. These forms are: (1) a gently sloping ice margin leading up to an ice-cored ridge of morainal debris; (2) a gently sloping ice margin with little or no morainal material on the ice surface; and (3) ice cliffs below which morainal ridges often occur. The main objective of this study was to investigate reasons for the differences among these forms.

The first form includes the shear moraines, herein called "ice-cored moraines", previously described from this area by Bishop (1957) and Swinzow (1962). Within the area there is a narrow ice-cored moraine belt, the Siorqap moraine, extending north from Camp TUTO for a distance of about 4 km, and a much broader belt, the Pitugfiup moraine, north of the west ice cliff (Fig. 1). (These moraine names, herein used for the first time, are the Greenlandic names of the rivers draining the respective parts of the glacier margin.) A moraine near the south end of the Siorqap belt differs from those above in being roughly perpendicular to the margin. A second objective was to investigate reasons for these differences in moraine form and orientation.

A total of 3 months was spent in the field during the summers of 1965, 1966 and 1968, and 1 month was spent in the laboratory in 1967. In the laboratory, ice cores from the glacier margin were studied. Field work included mapping, trenching in moraines and ice-movement measurements.

The term "shear moraine" was used by Bishop (1957) but Weertman (1961, p. 3–4) and Hooke (1968) have argued that the shear hypothesis is mechanically unsound. Weertman used the term Thule–Baffin moraine, thus identifying the type areas for these features. I prefer the name "ice-cored moraine" (Østrem, 1963) because it is descriptive but non-genetic.

Description of moraines

Siorqap and Pitugfiup moraines

The Siorqap moraine ranges in width from 30 to 150 m and the debris layer is probably less than 0.6 m thick on the average. In contrast, the Pitugfiup moraine is two to ten times as wide and the debris layer is substantially thicker; a hole dug at the end of flag line 4 (Fig. 1) bottomed in permafrost at a depth of 1.43 m and similar debris thicknesses were estimated elsewhere in the south end of the moraine from the absence of ice in vertical exposures. At present, even during the warmest summers, debris probably is not added to this part of the moraine from the underlying ice. Farther north, ice was seen at depths of 0.3 to 0.7 m in melt streams and in local ice cliffs, suggesting that debris thicknesses may decrease northward. The larger volume of debris in the Pitugfiup moraines and the existence of permafrost within this debris layer suggest that this moraine is much older than the Siorqap moraine.

In the relatively recent past the ice margin in the Siorqap area has retreated 100 to 200 m from a lichen trim line (Fig. 2). No similar trim line is found in the Pitugfiup area, though Bishop (1957, p. 43–45) reported one farther north. However, about 100 m up-glacier from the Pitugfiup moraine a new ice-cored moraine is developing. In its central part this moraine is thinner, has less relief and is more hummocky than either the Siorqap or the main Pitugfiup moraines (Fig. 3F). Southward it peters out, but a dirt band a few centimeters wide and

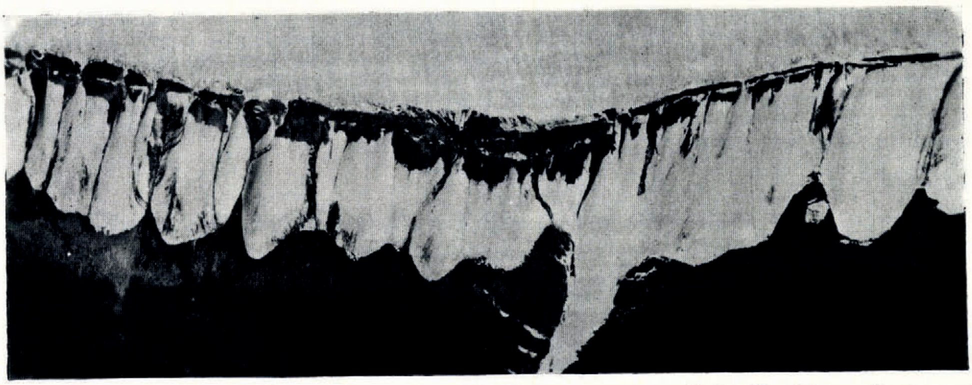

Fig. 2. July 1954 aerial photograph of Siorqap moraines. Flag line 1 was on the up-glacier side of the moraine near center of the picture. Note the relationship of wind-drift ice wedges to water gaps in the moraine, and note trim line or lichen-free zone in front of wedges. By August 1968 distal ends of many wedges had retreated (Fig. 1). Total width of photograph is about 3 km.

Fig. 1. Generalized map of moraines in Thule-TUTO area, north-west Greenland.

dipping 30° to 50° up-glacier can be traced for about 1 500 m into the west ice cliff (Fig. 1), where it marks the boundary between active glacial ice and less rapidly deforming superimposed ice (Fig. 9). Northward this new moraine becomes wider and thicker, and it merges with the main Pitugfiup moraine. Evidence presented later suggests that ice retreated from the south end of the Pitugfiup moraine in the past few hundred years, that the gap between the glacier margin and the still ice-cored Pitugfiup moraine became filled with superimposed ice, and that this new moraine resulted from a subsequent re-advance of the glacier.

These observations support the sequence of deglaciation proposed by Davies and others (1963, p. 61). In their interpretation the ice-sheet margin in the Pitugfiup area has been in its present location since the earliest phases of deglaciation (perhaps since 32 000 B.P.), but in the Siorqap area the margin has been in its present location for perhaps only one-quarter of this time.

Skyline moraine

Immediately east of Camp TUTO the Siorqap moraine swings eastward and points towards an ice-cored moraine about 4 km out on the ice sheet, herein called the Skyline moraine. Movement of the Skyline moraine is slow relative to the nearby TUTO ramp (0.7 m year^{-1} versus 2.9 m year^{-1}; Fig. 1). These observations suggest that the Skyline moraine and the east-trending end of the Siorqap moraine are surface expressions of a medial moraine derived from the subglacial bedrock ridge indicated by Barnes' gravity survey (Bishop, 1957, p. 24).

Flow measurements on flag line 1 (Fig. 1) do not support Bishop's (1957, p. 28) hypothesis that flow rates north of the east-trending moraine are significantly higher than those on the ramp, and that the east-trending moraine owes its orientation to drag of northern ice against the ramp ice. Treating these moraines as part of a medial moraine seems more logical.

Other ice-cored moraines

Just west of flag line 3 an ice-cored moraine has been built along a short north–south segment of the glacier margin. This moraine resembles the Siorqap moraines in that it is narrow and the debris layer is relatively thin. The moraine consists of a single arcuate ridge with an imposing ice-cored bulwark at its northern end. It appears that the bulwark formed first and that a 50 to 100 m retreat of the ice sheet was responsible for development of the ridge up-glacier from the bulwark.

An ice-cored moraine south-east of the east ice cliff is now completely separated from active glacial ice. Here again a recent retreat of 100 to 200 m is indicated.

Active ice-cored moraines south of TUTO near Petowik Glacier also indicate a recent retreat of 100 to 300 m.

Main ice-cliff moraine

Till melting out of the main ice cliff has formed a moraine at the base of the cliff. The base of the cliff is now 50 to 100 m back from the crest of the moraine, providing further evidence for recent retreat. This is the only major moraine in the study area that did not form on top of the ice from which the till was derived.

Bishop (1957, p. 16) stated, without citing evidence, that this moraine is ice-cored. I found no evidence for a core. Probing the bases of 1 to 4 m high stream banks and slump scars indicated that if a core exists, it is probably below the level of annual thaw. Foliated ice at the base of the cliff extends out to the moraine, but field exposures were not adequate to determine whether or not this ice was connected to an ice core.

Westward the main ice cliff dies out and wind-drift ice occupies the area between the glacial ice and the morainal ridge. The moraine continues past and beneath this wind-drift ice and reappears in front of the west ice cliff where its crest is again 50 to 100 m from the base of the cliff.

Characteristics of debris in the ice

Dirt bands in the TUTO ice tunnel were divided into four characteristic facies: *solid*, *banded*, *sandy-amber* and *amber*. Solid bands are composed of rock particles which touch each other and which are cemented by interstitial ice. The banded facies consists of 1–5 mm thick dirt bands which separate and rejoin. The dirt layers themselves are not solid but consist of dispersed silt and sand grains. Rare pebbles and cobbles cross-cut several layers. Amber ice is ice with sufficient dirt to give it an amber color but with few or no visible mineral grains. Sandy-amber ice is amber ice in which sand grains are readily visible. Ice of a particular facies may range in thickness from a few millimeters to several meters.

Concentrations of dirt in the ice were determined by melting a known weight of ice, decanting and evaporating the water, and weighing the dried residue (Table I). The larger samples had to be analyzed in a period of a few hours and some fines were lost during decanting. However, the material lost was probably a small fraction of the total present. The larger size of these samples makes them more representative of the average debris content of the ice.

Table I. Debris concentrations in ice from the TUTO ice tunnel

Ice type	*Sample size, debris and ice* kg	*Weight of debris* kg	*% debris by weight*	*% debris by volume**
Banded	35.20	2.12	6.0	2.4
Banded	1.781	0.053	3.0	1.2
Banded	0.992	0.029	2.9	1.1
Sandy-amber	34.85	0.166	0.5	0.2
Amber	35.65	0.0035	0.01	0.004

* Assume density of 2.3 Mg m^{-3} for uncompacted sediment and of 0.9 Mg m^{-3} for ice.

Moraine forms

Sharp-crested and broad-crested ridges

Ridges in ice-cored moraines can be separated into two types: sharp-crested and broad-crested (Fig. 3). Broad-crested ridges appear to be older and more stable as their debris cover is generally thicker, patterned ground is commonly developed on their flat tops, and less frequently lichens have become established. In contrast, the till cover on sharp-crested ridges commonly has tension cracks reflecting active slumping.

Because sharp-crested ridges were commonly found above solid dirt bands (Fig. 3C), it is inferred that such dirt bands are required for their formation (see also Boulton, 1967, p. 723–30). When the supply of dirt in the band is exhausted, the ridge crest becomes rounded (Fig. 3A). Subsequently other solid dirt bands may be exposed on its flanks; these produce secondary ridges (right-hand side of Fig. 3A). Broad-crested ridges are thought to form from several such solid bands melting out over a period of years, each band offset slightly from preceding ones.

Leveling of tops of broad-crested ridges occurs through flow of saturated till during the summer melt season. Such flow was observed in trenches (Fig. 3) wherever lateral movement of melt water was inhibited by a relatively level ice–debris interface. After surface topography is smoothed, local high points in the ice–till interface melt due to the lesser thickness of the debris layer over them.

This hypothesis for the origin of sharp- and broad-crested ridges is a one-cycle hypothesis which seems appropriate in the relatively young Siorqap moraine system. Over a longer time interval, flow of debris into troughs and subsequent topographic inversion by differential ablation could result in similar forms.

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Flats

Several areas in the Siorqap moraine were relatively flat and did not stand significantly above the surrounding clean ice (Fig. 3C). These flats are apparently stable as patterned ground is commonly developed on them. They are of interest because they do not stand appreciably above adjacent glacier ice, implying that little or no differential ablation has occurred, despite the till cover. They are generally found on the up-glacier side of the moraine (Fig. 3C), in a position where the contact between the flat and adjacent glacier ice may be covered by snowdrifts during much of the melt season (Fig. 12). It is inferred that this cover of superimposed snow is sufficiently persistent that net ablation beneath the flats nearly equals net ablation just up-glacier from them. This would result in the flats not being raised appreciably above the clean ice on their up-glacier sides.

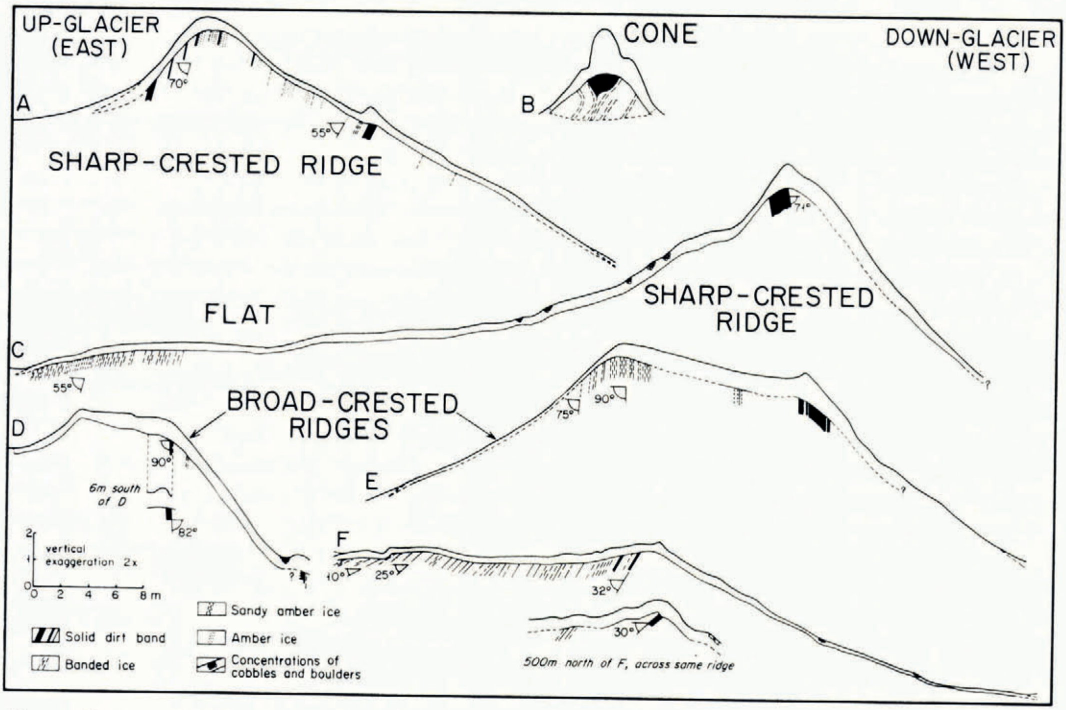

Fig. 3. Thickness of debris cover and debris facies in underlying ice as exposed in trenches through moraines. Ice–debris contact dashed where trenches did not reach contact, either due to time limitation or to presence of frozen till. Trenches A–E were on the Siorqap moraine. Approximate distances north of the ice-tunnel entrance are: A 300 m, B 3 700 m, C 4 000 m, D 3 400 m, E 650 m. Trench F was in the new moraine east of the main Pitugfiup belt and 2 800 m north-east of flag line 4.

Debris cones

Circular or elliptical cones of debris are common in the middle and along the up-glacier side of the Siorqap moraine. These cones vary in height from 0.5 to 5 or 6 m, and in diameter from 1 to 25 m or more. In some cases cones are aligned parallel to the general trend of the moraines (Bishop, 1957, p. 5). The smaller cones form when debris pods (in contrast to ribbons and sheets) melt out at the ice surface (Fig. 3B). Large cones could form either from larger pods or from debris ribbons aligned parallel to flow lines in the ice.

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Sources of debris

Till in ice-cored moraines is sorted in a manner suggesting different source areas for adjacent debris masses. Size sorting is perhaps most common; debris in one ridge may be dominantly pebble- and cobble-sized, while that in an adjacent ridge has a prominent matrix of sand. Somewhat less common are solid bands of sand or clay which retain fluvial bedding or varves, respectively (see also Bishop, 1957, p. 42). Three such bands were found in the Siorqap moraine and in a moraine south of the TUTO ramp. Goldthwait (1960, p. 59) has reported similar masses in which cross-bedding was preserved.

Another type of sorting is lithologic. Near the east end of the main ice cliff, two black debris cones about 0.75 m high contained abundant black shale chips. Such chips were lacking in adjacent brown cones of about the same size. Similarly, the south part of the new ice-cored moraine east of the main Pitugfiup moraine contains much more shale than the Pitugfiup moraine itself. These differences almost certainly reflect a difference in provenance, and suggest that masses of till that are juxtaposed in the moraines may have come from widely separated areas of the glacier bed.

Weertman (1961) suggested a mechanism of repeated freeze and thaw at the base of the glacier to incorporate material into the ice. Some such mechanism is required to account for undisturbed solid bands of fluvially sorted debris for which Bishop's (1957, p. 17) shear mechanism appears untenable. Weertman's process also seems to account for the finely dispersed sand in banded ice better than Bishop's. However, boulders with glacial striations are readily found in the moraines, indicating that differential movement across discrete shear planes does occur on the glacier bed or within the ice. The striations may mean that ice on the bed is locally at the pressure melting-point, a condition required by the Weertman freeze–thaw mechanism.

Ice-movement measurements

Procedure

Five lines of flags were established to measure horizontal movement of the ice (Figs. 1, 2 and 4). Each line originally consisted of five (flag line 5) or 20 bamboo poles set at 50 m intervals in 2 m deep holes drilled in the ice. Flag line 1 was established in August 1965 and was re-surveyed 1 year later. Flag lines 2–5 were established in August 1966, and were

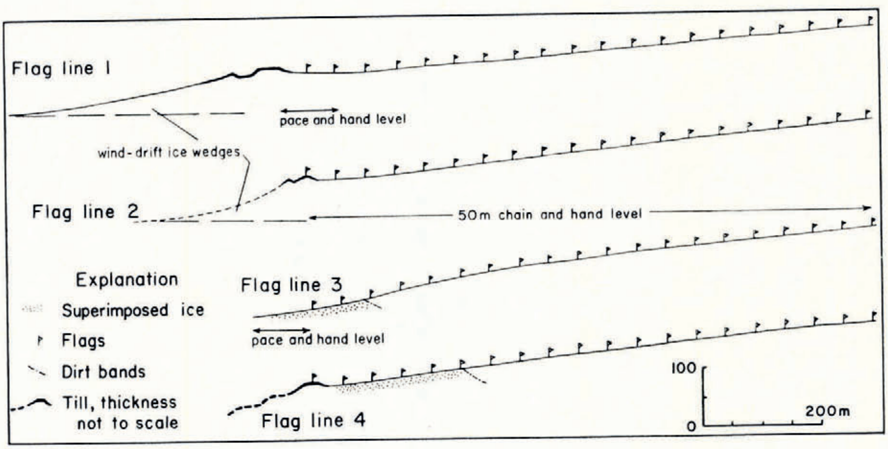

Fig. 4. Profiles along flag lines, August 1966. Dashed parts schematic. Distances determined by chaining and elevations by traverse with surveyor's level and rod unless otherwise indicated.

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re-surveyed 2 years later. Ablation rates were higher than expected in the northern part of the study area, and by July 1968 only three and six flags, respectively, remained in place in flag lines 3 and 4.

Survey points were established in front of the ice sheet (Fig. 1) and theodolite triangulation was used to locate a reference flag in each line (except flag line 1; see Table II). Each angle of each triangle was turned (direct and reverse) at least four times. In most cases closure errors were less than 5″ of arc. Flow rates of reference flags in lines 3 and 4 were measured independently by surveys from two base lines; rates differed by 0.10 m year^{-1} (flag line 3) and 0.00 m year^{-1} (flag line 4). From the reference flag, the azimuth and distance to the remaining flags in each line were determined by chaining distances between flags, and by occupying the reference flag and turning angles from one of the survey points to each flag in the line. The standard error in movement measurements is probably less than $\pm$0.10 m year^{-1} for flags within 500 m of the reference flag and less than $\pm$0.20 m year^{-1} for more distant flags.

TABLE IIa. BASE LINES USED FOR TRIANGULATION TO FLAG LINES

Flag line	*Base lines*	*Remarks*
1	A–F, B–C	A–F used to locate triangulation points B and C each year. Flags in line were located from base line B–C
2	A–X	Checked from A–B, X–Y and X–Z (1966), and Y–Z (1968). A–X considered most reliable as angles of triangle are most nearly equal. Movement determined from other base lines was in reasonable agreement considering probable error of measurement
3	X–Y and X–Z	Movement determined from both base lines and values averaged
4	X–Y and Y–Z	Movement determined from both base lines and values averaged
5	NN–SS	

TABLE IIb. DETERMINATION OF BASE-LINE LENGTHS

Base line	*Length determined by*
NN–SS	U.S. Army (Davis, 1967, p. 2)
A–F	Triangulation from temporary base line F–F′. F–F′ was chained
X–Z	Chained
X–Y	Triangulation from X–Z
Y–Z	Triangulation from X–Z
A–X	Triangulation from A–F using B and the first flag of flag line 2 as intermediate points

On flag line 4, vertical movement relative to the reference flag was measured by surveying level lines in 1966 and 1968. Closure errors suggest that movement measurements should be accurate to $\pm$0.02 m year^{-1} for flags within 250 m of the reference flag and $\pm$0.04 m year^{-1} for more distant flags. Because the reference flag has a horizontal velocity of 0.10 m year^{-1} and because the strain-rate between it and the third flag is only $-$0.00026 year^{-1}, it has been assumed that the reference flag has a negligible vertical velocity, and that measured vertical movement of the other flags equals absolute vertical movement within limits of measurement error.

Results

Selected horizontal movement measurements have been plotted in Figure 1 and more complete data are given in Figure 5. Without exception, flow is compressive and is nearly perpendicular to the local glacier margin. Maximum flow rates were 4.0–4.5 m year^{-1} on the up-glacier ends of flag lines 1 and 2, 1 000 m from the Siorqap moraine, or 1 500 m from the margin. On the Siorqap moraine, movement increased northward from 0.9 (flag C) to 1.2 (flag B) to 1.7 m year^{-1} (down-glacier end of flag line 2); on flag lines 3 and 4, in comparable locations with respect to the ice-sheet margin, the flow rate was 1.1 m year^{-1}. On the TUTO ramp (Davis, 1967), flow rates were about 3.5–4.1 m year^{-1} 2 000 m from the margin, about 1.0 m year^{-1} 750 m from the margin, and less than 0.1 m year^{-1} 500 m from the margin.

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Davis's measurements indicate that the down-glacier end of the TUTO ramp is essentially stagnant. Such nearly stagnant wedge-shaped masses of ice are herein called "wind-drift ice wedges" (Fig. 4) because, as later discussed, I believe they are maintained by, though not necessarily composed of wind-drift snow. Extrapolation of the present measurements suggests that similar nearly stagnant zones are present down-glacier from flag lines 1 and 3 (Fig. 5). In both cases the boundary between active and stagnant ice would be 50 to 100 m down-glacier from the moraine-forming zones of dirt-bearing ice which came from the base of the glacier (Fig. 6A, sketch 1). On flag line 2, the extrapolation suggests a much greater distance between the zone of dirty ice and the stagnant ice, presumably reflecting the fact that ice just north of flag line 2 continues flowing north-westward around the end of the moraine and into the topographically low area near flag lines 3 and 4 (Fig. 1).

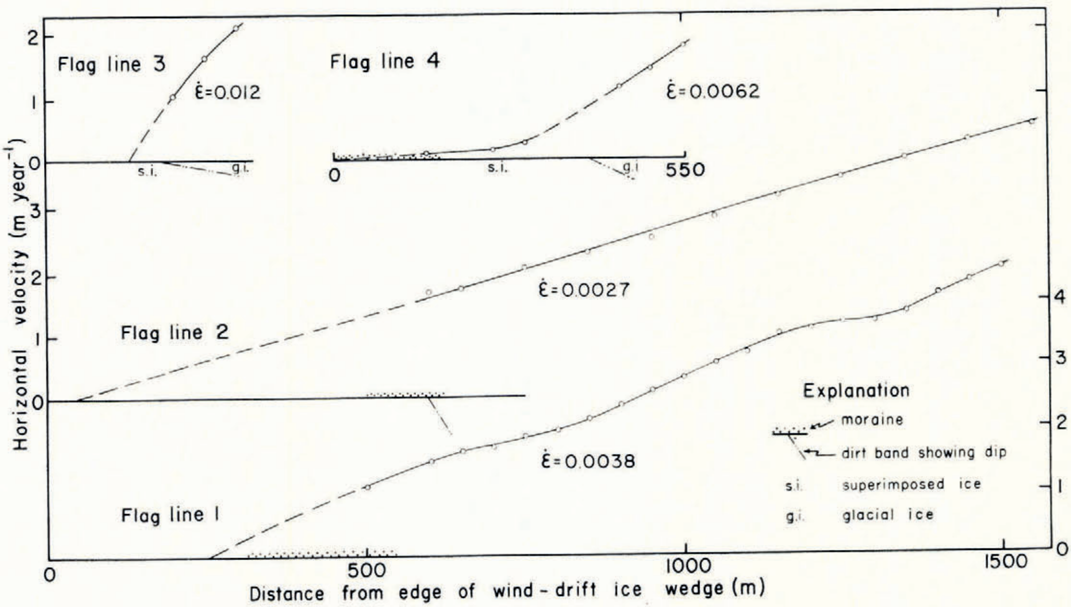

Fig. 5. Relationship between total ice movement and distance from ice-cap margin. Movement measurements extrapolated (dashed lines) to determine approximate position of boundary between active ice and wind-drift ice wedge. Mean strain-rate for flag lines 1 to 3 and the up-glacier part of flag line 4 is shown. Moraine and dirt-band symbols show where such features occur relative to flags used for movement measurements.

Data from flag line 4 suggest that an asymptotic approach to a zero flow rate may be more realistic than the straight-line extrapolations shown for the other flag lines (Fig. 5). This is supported by laboratory measurements of the flow law of ice (Glen, 1955; Nye, 1957, p. 129) which suggest that

$$\dot{\epsilon} \approx 0.15\tau^{4.2} \tag{1}$$

where ϵ is the effective strain-rate in year^{-1}, and τ is the effective shear stress in bar. The high exponent indicates that $\dot{\epsilon}$ decreases rapidly as τ decreases below 1 bar. But regardless of the extrapolation used, ice in the wind-drift wedges and in the down-glacier 250 m of flag line 4 has a relatively low velocity and lower than average strain-rate.

Strain-rate measurements

During the summer of 1968 three surface strain nets (Nye, 1959) were established along flag line 3 and one net above the main ice cliff. Each net consisted of four flags at the corners of a square with sides approximately 35.3 m long. Lengths of the sides and diagonals of each square were measured every 7–10 days, but the total length of reliable record on individual nets was short, ranging from 8 to 17 days. Study of the internal consistency of the measurements suggests that measured strain-rates should be accurate to ± 0.001 year^{-1}. However, strain-rates varied from week to week, and as estimates of the annual strain-rate these measurements are only accurate to ± 0.005 year^{-1}. The error in orientation of the principal strain axes is $\pm 25°$.

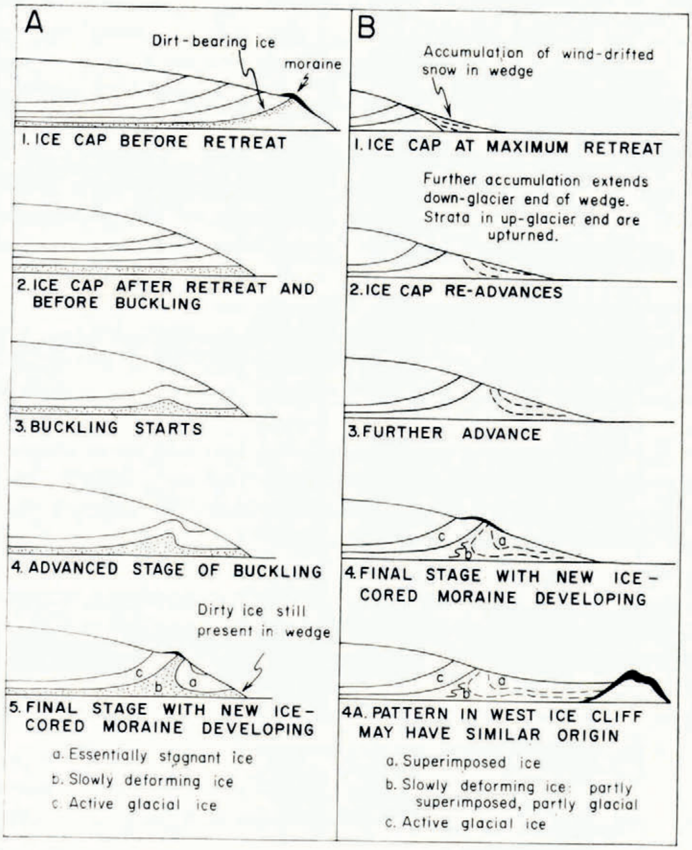

Fig. 6. Hypothetical sketches showing successive stages in development of down-glacier-dipping foliation.
A. Down-glacier-dipping foliation developed in glacial ice.
B. Down-glacier-dipping foliation developed in superimposed ice.

Strain-rates shown in Figure 1 are consistent with other evidence of the local strain distribution. The larger of the principal compressive strain-rates (-0.020 year^{-1}) measured at the down-glacier end of flag line 3 is nearly parallel to the observed flow and, if the possibility of higher summer strain-rates is recognized (Davis, 1967, p. 4), is consistent with strain-rates between nearby flags on flag line 3 (-0.011 and -0.013) measured over a 2 year period. The tensile strain-rate above the main ice cliff is consistent with the orientation of 0.1 m wide crevasses near the south-east end of the cliff. Finally, the increase in strain-rate westward along flag line 3 is logical considering the increase in ice-surface slope along this line (Fig. 4).

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Foliation and the flow field near the margin

Origin of foliation

When any plastic material is compressed in one direction and allowed to flow in a direction perpendicular to the compressive stress, inhomogeneities in the material will be stretched out and will form foliation. Laboratory studies suggest that foliation in glaciers may even develop in initially homogeneous ice (Kamb, 1964, p. 363). Three main types of ice contribute to foliation in the Thule area: dirty ice, bubbly clean ice and clear clean ice.

Local tensile or shear strain parallel to foliation is necessary for its development, but foliation is not *necessarily* parallel to stream lines within the glacier. For instance vertical foliation is commonly developed in compressive zones below ice falls (Allen and others, 1960, p. 617–18; Sharp, 1960, p. 57), whereas the dominant flow direction in such zones is longitudinal.

Foliation in the Thule area

There are three basic foliation patterns in the study area. First, foliation beneath and on the up-glacier side of the ice-cored moraines strikes parallel to the moraines and dips up-glacier. Dips of 70° to 90° are common near the moraines but dips tend to decrease progressively up-glacier (Fig. 1). Secondly, in ice cliffs in the north-west to south-east trending part of the glacier margin foliation dips are 5–10° up-glacier, an attitude typical of the termini of many valley glaciers (Sharp, 1960, p. 57).

These two foliation patterns were reproduced experimentally by compressing a vari-colored batch of modeling clay vertically and allowing it to flow horizontally. Blocks of wood were placed on three sides of the clay, thus restricting flow to one direction. When a fourth smaller block was placed in front of the clay in the direction of flow to simulate a stagnant wind-drift ice wedge, a foliation pattern was produced in the "ablation zone" (Fig. 7) similar to that observed on the ice-sheet surface up-glacier from wind-drift ice wedges. When the small block was removed, gently dipping foliation resulted. This supports the conclusion that the steeper dip of foliation up-glacier from the wind-drift ice wedges results from a higher compressive strain due to obstruction of the flow by the wedges, and perhaps also due to a higher effective viscosity of till-bearing ice.

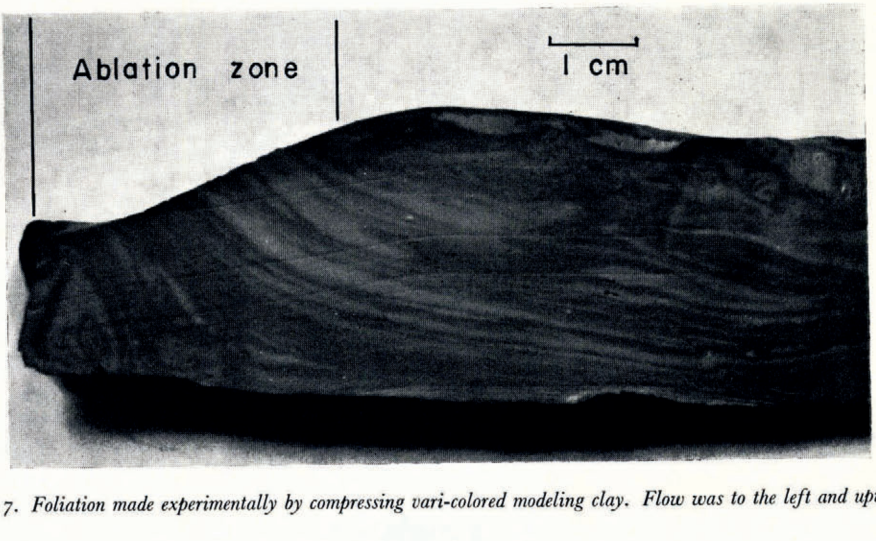

Fig. 7. Foliation made experimentally by compressing vari-colored modeling clay. Flow was to the left and upward.

Thirdly, foliation beneath and on the down-glacier side of the Siorqap moraine and on the TUTO ramp generally strikes parallel to the glacier margin and dips *down-glacier* at angles of 45–85°. This down-glacier-dipping foliation was not reproduced well in clay. The few bands with such an attitude in Figure 7 probably resulted from buckling of originally horizontal foliation during initial stages of deformation. Similar buckling could be responsible for such foliation in the wind-drift wedges. The horizontal banding at the down-glacier ends of the

Fig. 8. Ice cliff on side of wind-drift ice wedge; Lake Tuto in foreground. Foliation near moraines dips 60° up-glacier directly away from the observer. Banding on down-glacier end of wedge is believed to be sedinentary bedding that has been upturned by advancing glacial ice. This banding strikes perpendicular to the cliff face.

Fig. 9. West ice cliff. Banding on the left is probably sedimentary (Fig. 6B). This banding is wrinkled and overturned by active glacial ice advancing from the right. Boundary between active ice and less rapidly deforming superimposed ice is marked by dirt band (arrows) which can be traced into the ice-cored moraine east of the Pitugfiup moraine (Fig. 1). Note contortion of foliation beneath advancing ice.

wedges would be foliation formed near the base of actively flowing glacial ice at a time when the ice-sheet margin occupied a more advanced position. Sequential drawings in Figure 6A illustrate the hypothesized process. Alternatively, the horizontal banding could be primary sedimentary bedding (compare Figure 6B with Figures 8 and 9). Both the absence of dirt-bearing ice near the base of the wedges and the fabric studies discussed next suggest that the banding is sedimentary. If this is the case, a minor advance of the ice sheet is indicated by the steep dip of foliation in the wedges (Fig. 6B).

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Fabric studies

Data

Ice cores were collected from 15 locations along the ice-sheet margin and fabric diagrams (Fig. 1) were prepared using standard techniques (Langway, 1958). The cores were 0.075 m in diameter and averaged about 0.25 m in length. 14 of the cores were from within 1 m of the glacier surface. Evidence for recrystallization of ice near the surface (Rigsby, 1960, p. 601) was sought but not found.

Fabric diagrams are commonly contoured to show the percentage of c-axes falling within some area A. A is generally taken to be 1% of the area of the diagram (e.g. Rigsby, 1960). However, suppose M_i is the number of points falling within the ith area of size A, and E and σ are the mean and standard deviations of M_i, respectively, *in diagrams with statistically random c-axis orientations*. Then, following Kamb (1959, p. 1909), we can select the size of A such that $E = 3\sigma$. In this case, A is approximately 8.3% of the area in diagrams in which 100 c-axes are plotted. Concentrations of points exceeding $E+3\sigma$ in such an area are interpreted to represent statistically significant preferred orientations (Kamb, 1959). $E+3\sigma$ is twice the average concentration of points in a random fabric and concentrations of $E+3\sigma$ or larger have a probability of occurring of 0.0013 in random fabrics.

For each diagram in Figure 1 the maximum number of c-axes falling within 1% of the area is given as a percentage of the total number of points, and the maximum number of points falling within the appropriate area calculated by Kamb's method is given in terms of E and σ. Each point on a fabric diagram of, say, 100 points is interpreted to represent the orientation of c-axes in 1% of the volume of the core (Hooke, 1969).

Interpretation

Two cores were collected from ice that is thought to be superimposed (locations 1 and 8). This ice proved to be fine-grained (Fig. 10A) and fabrics were relatively weak, though stronger than fabrics obtained by Rigsby (1960, fig. 7) for other superimposed ice in Greenland. The fabric from location 8 is also stronger ($E+12\sigma$) than one would expect if crystal growth were controlled by randomly oriented snowflakes. The thin section from location 8 was from a bubble-free part of the core and refreezing of percolating melt water may have strengthened the fabric. The fabric from location 1 was from bubbly ice and is perhaps more typical of superimposed ice formed from compaction of snow. Ice having textures and fabrics similar to those of locations 1 and 8 is herein interpreted as superimposed.

In contrast, glacial ice is generally coarser-grained and has a stronger fabric (Fig. 11; locations 9–11). The strongest fabrics are found in ice near or beneath ice-cored moraines (location 9) and fabric strength decreases up-glacier from the moraines (locations 10 and 11). This reflects the fact that ice beneath the moraines moved near the base of the glacier where velocity gradients and hence shear strains are greatest (Fig. 6A; sketch 1) (Rigsby, 1960, p. 596, fig. 7). Ice up-glacier from the moraines moved higher above the bed of the glacier where lower shear strains would be expected. Similarly, fabrics from locations 5, 7, 9 and 13 are relatively strong, and that from location 6, up-glacier from 7, is weaker than that at location 7.

Grain-size decreases as shear strain increases (Fig. 11A–C) but even the smallest crystals in glacier ice are substantially larger than those found in superimposed ice. This decrease has been observed previously by Kamb and Shreve (1963) on a temperate valley glacier, by Rigsby (1955, p. 6; 1960, p. 605) on polar glaciers, and by Glen (1955, p. 528) in ice deformed in the laboratory.

Fabrics from locations 1–4 appear to represent successive stages in metamorphism of superimposed ice and formation of glacial ice (Fig. 10A–D). As described, ice from location 1 is believed to be superimposed. Ice from location 2, near the base of the west ice cliff (Fig. 9), has a weak fabric and a texture similar to the granulite texture in metamorphic rocks (Fig. 10B), suggesting compaction and crystal growth in the absence of shear. Location 3 is in contorted ice beneath the active glacier ice in Figure 9. This ice has clearly been deformed; crystals are larger than at location 2 (Fig. 10C), and the fabric is substantially stronger than that at location 2, though not as strong as those found in more active glacial ice (e.g. location 9). Location 4 is just up-glacier from the dirt band that apparently marks the boundary

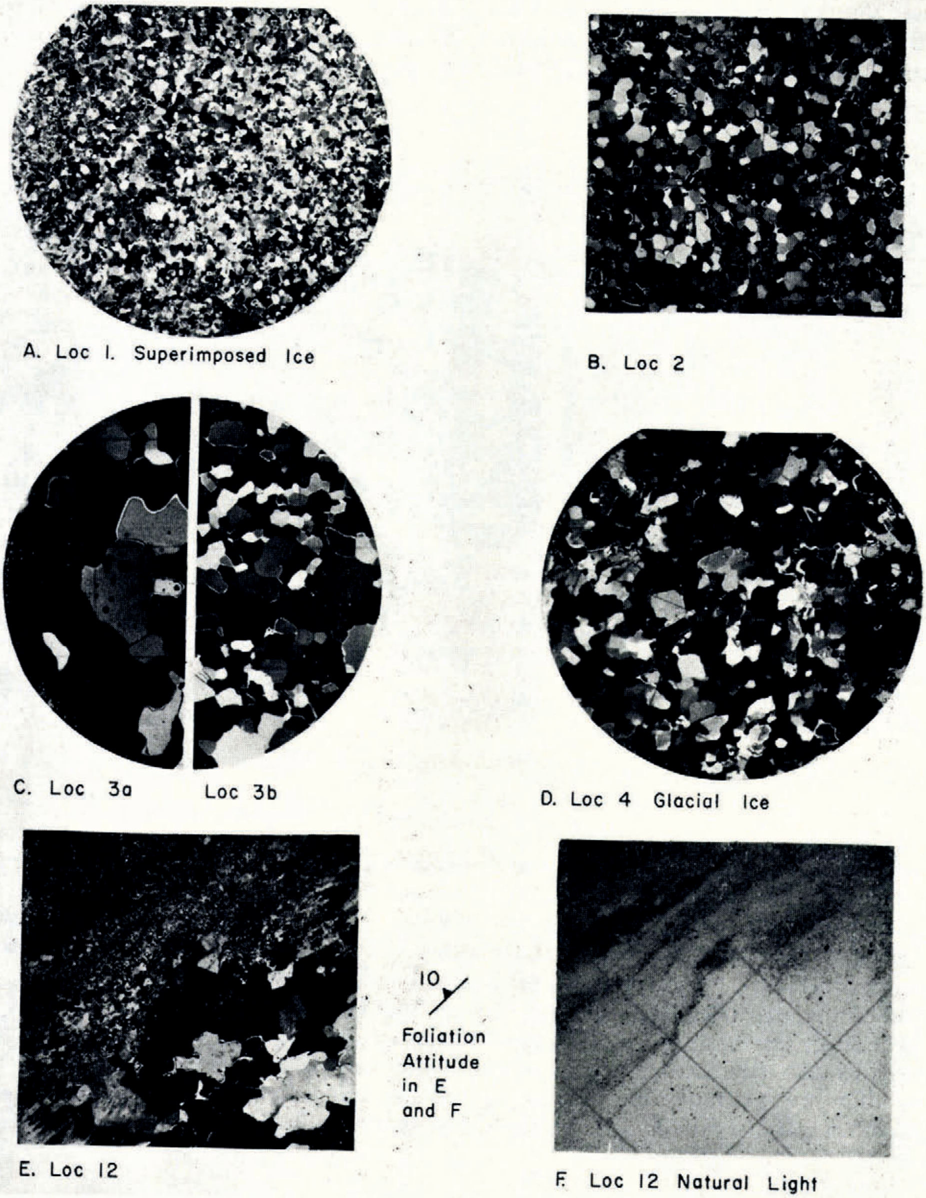

Fig. 10. Thin-section photographs. All photographs except F are taken through crossed polaroids. Grid lines are 20 mm apart. Locations 3a and 3b are from depths of 0.09 and 0.12 m, respectively. Location 3a has a stronger fabric (Fig. 1). In a longitudinal section the change in crystal size was observed to occur across a single grain boundary.

between glacial and superimposed ice. That this ice has been subjected to substantial shear strain is indicated by the strong fabric, by the texture and by the prevalence of undulatory extinction in crystals. This sequence of four cores suggests that ice between the Pitugfiup moraine and the dirt band near location 4 is superimposed, and that the structures in the west ice cliff originated as shown in Figure 6B (especially sketch 4A).

The increase in fabric strength from location 14 to location 15 on the TUTO ramp reflects a corresponding increase in ice-flow rate. Location 15 is near the line of negligible (horizontal and vertical) movement, and location 14 is closer to the line of maximum vertical flow (Fig. 1; Davis, 1967, figs. 3–5). However, in comparison with, say, locations 4, 7, 9 or 13, ice from locations 14 and 15 is coarse-grained and has a weaker fabric, suggesting lower shear strain on the TUTO ramp.

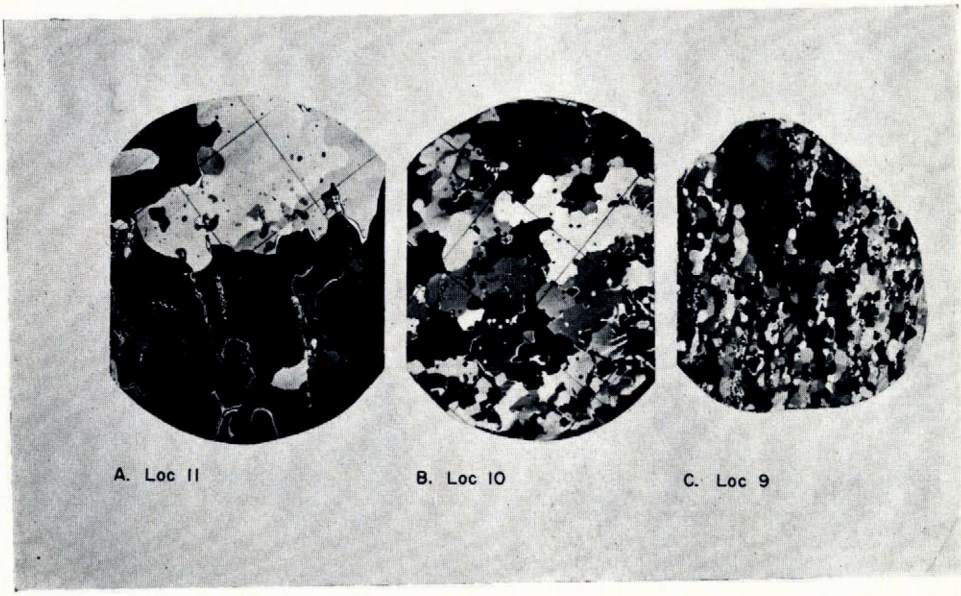

Fig. 11. Thin-section photographs through crossed polaroids. Grid lines are 20 mm apart.

Effect of debris on ice texture and fabric

Two of the cores collected for fabric analysis (locations 9 and 12) came from debris-bearing banded ice. The presence of debris did not noticeably affect the fabrics; both cores had fabrics consistent with their locations close to (location 12) or in (location 9) zones of high shear strain beneath the moraines. However, the debris did markedly affect grain-size. The small dark specks in Figure 10F are clay- to sand-sized mineral particles. Ice crystals in the neighborhood of these particles (Fig. 10E) are extremely small but ice crystals in intervening clean-ice folia occupy much of the width of the foliation band.

Several other thin sections were cut from sandy-amber or amber ice. In most of these sections there was no noticeable change in grain-size in the vicinity of the sediment particles. A slight effect was observed in a few sections in areas of locally higher sediment concentrations.

Mass budget

If an ice cap is in a steady state, mean annual upward flow in the ablation zone must equal mean annual net ablation (= excess of ablation over accumulation in mm of ice). Under

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these conditions the local profile of the ice sheet will, on the average, remain constant from year to year. Net ablation and upward flow are both measured with respect to a vertical coordinate, and upward flow is defined as shown in Figure 12A. That is

$$\text{Upward flow} = v + u \tan \alpha \tag{2}$$

where u and v are the horizontal and vertical components of the velocity vector, and α is the slope of the glacier surface.

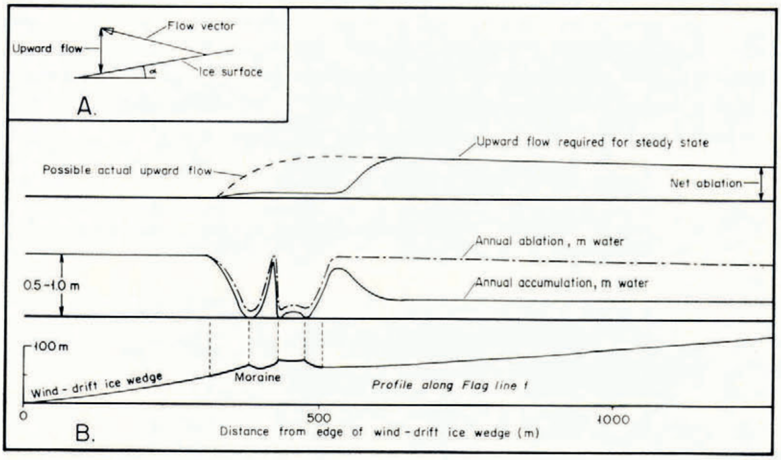

Fig. 12. A. Definition of upward flow.
B. Schematic diagram showing how excess of ablation over accumulation (net ablation) is balanced by upward flow. Formation of snow drifts results in excessive accumulation in vicinity of moraine. Thus net ablation in these areas is low and may be zero. Net ablation beneath moraines is also low because till insulates the ice.

TABLE III. MASS BUDGET ON FLAG LINE 4

Flag	*Upward flow* m year^{-1}	*Ablation* m year^{-1}
1	0.00*	0.00†
3	0.01 ± 0.02	0.07 ± 0.03
4	0.04 ± 0.02	0.25 ± 0.03
7	0.37 ± 0.04	0.68 ± 0.04
8	0.41 ± 0.04	0.64 ± 0.04

* Assumed.

† When setting flag in moraine, permafrost was encountered at a depth of 1.23 m. Base of flag was set 0.20 m into permafrost. No clean ice encountered.

Negative mass budget at present

On flag line 4, upward flow was determined from the measurements of u and v described earlier, and annual net ablation was determined by measuring the amount of each flag projecting from the ice in 1966 and 1968. During this 2 year period, net ablation exceeded upward flow by 0.2–0.3 m year^{-1} on this flag line (Table III).

Corroborative evidence for a negative mass balance at present comes from measurements made by Goldthwait* about 65 km north-east of Thule. Detailed surveys in 1955 and 1965 indicate an average lowering of the ice surface of about 5 m, or 0.5 m year^{-1}, over an area of 1 km^2.

* Paper in preparation entitled "Restudy of Red Rock ice cliff in Nunatarssuaq, Greenland".

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Although the present mass budget is negative, the observations and measurements described above suggest that ice-cored moraines, ramps and ice cliffs can all be equilibrium features on an ice sheet with a balanced budget. This conclusion, discussed below, is of interest because these forms have been previously interpreted as indicating either advance or retreat (Goldthwait, 1951, in preparation*; Bishop, 1957).

Mass budget near ice-cored moraines

The mass budget near moraines is complicated by drifting snow blown by katabatic winds. This snow accumulates in the trough on the up-glacier side of the moraines, in troughs within the moraines, and on tongue-shaped wind-drift ice wedges down-wind from water gaps through the moraines (Figs. 1 and 2). These drifts last well into the melt season and many persist for several seasons. Such drifts are easily distinguished from the dense glacier ice up-glacier from the moraines. The dense ice is only locally exposed on the wind-drift ice wedges (Fig. 1); several flags have been set 2 to 3 m into these wedges without encountering it.

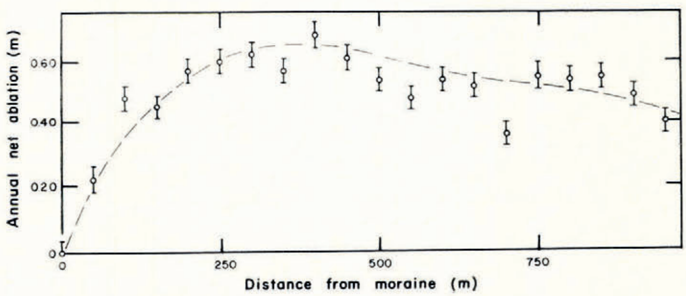

Fig. 13. Annual net ablation on flag line 2. Measurements of height of top of each flag above ice surface were made on 15 August 1966, 25 July 1968 and 22 August 1968. Data from the last two measurements were interpolated to give ablation over 2 year time span (15 August 1966 to 15 August 1968) and result divided by 2. Bar indicates estimated measuring error. Net ablation assumed to be zero at flag 1 on moraine where till cover is 0.55 m thick.

In Figure 12B the annual accumulation line is high over the wind-drift ice wedge, over the trough in the moraine, and over the trough on the up-glacier side of the moraine, reflecting the higher than average accumulation in these areas due to drifting. The annual ablation line in Figure 12B decreases sharply over those parts of the moraine that are not covered by snowdrifts. This represents insulation of ice by the till layer and the resulting low ablation beneath the moraine. Where the moraine is covered by drifts, melting of the drifts is included in annual ablation. The excess of annual ablation over annual accumulation, or net ablation, is the difference between the annual ablation and annual accumulation lines, and is plotted at the top of Figure 12B. The decrease in net ablation up-glacier from the moraines is based on net ablation measurements on flag lines 1 (not shown) and 2 (Fig. 13), and is attributed to a decrease in ablation up-glacier (Goldthwait, 1951, p. 574) rather than to an increase in accumulation.

Figure 12B implies that for a steady state to exist with net ablation equal to upward flow, there need be little or no upward flow in wind-drift wedges below moraines, as Bishop (1957, p. 17) suggested. This agrees with Davis's (1967) measurements indicating negligible flow in the wedge associated with the TUTO ramp, and is also in accord with extrapolations in Figure 5 if plane strain is assumed for, if the average horizontal strain-rate, $\dot{\epsilon}$, is low in the relatively thin wedges, annual upward flow ($= t(e^{-\dot{\epsilon}}-1)$ where t is wedge thickness) will be negligible. Alternatively, considering gravitational forces alone, a shear stress of 1 bar should

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be attained at a depth of 72 m on the wedge below flag line 1 (surface slope 9.1°), so from Equation (1) there should be little flow where the wedge is less than 72 m thick. Assuming a horizontal bed, a thickness of 72 m is attained at the crest of the outer ridge of the moraine (Fig. 12B).

For these reasons, I conclude that the wind-drift wedges are maintained primarily by wind-drift snow and only secondarily by inflow of ice from up-glacier.

Upward flow beneath moraines

Figure 12B suggests further that there need be little upward flow beneath ice-cored moraines. However, till in the moraines has undoubtedly come from the glacier bed as a result of upward flow. Furthermore, the horizontal movement data (Fig. 5) indicate a compressive strain-rate of about -0.0054 year^{-1} beneath the moraines and perpendicular to them. The mean strain-rate parallel to the moraines between triangulation points B and C is about $+0.0003$ year^{-1}. Thus a vertical strain-rate on the order of $+0.0051$ year^{-1} is implied. If the ice is 75 m thick, this would result in an upward flow of about 0.3 m year^{-1}, allowing for somewhat lower strain-rates on the bed of the glacier than at the surface.

An attempt was made to measure this flow over a 32 d period in the summer of 1968 by surveying a level line to markers set in ice beneath the moraine at the down-glacier end of flag line 1 (Table IV). The results were inconclusive in that points first seemed to move down and then up again. The movement is greater than the probable limit of random error shown in the table but undetected systematic errors could be responsible for the erratic results. On the other hand, such erratic flow has been detected previously on the Greenland ice sheet (Goldthwait, 1960, p. 33–44; Davis, 1967, p. 4). The measured movements are of the right magnitude to give the predicted upward flow rate of 0.3 m year^{-1} but they are not consistently in the right direction.

TABLE IV. UPWARD FLOW BENEATH MORAINES AT THE END OF FLAG LINE 1

	Movement cm	
Location (see Fig. 4)	*22 July–10 August 1968*	*10–23 August 1968*
Crest of distal ridge, 380 m from toe	-4.4 ± 1.0	$+2.1 \pm 1.0$
Distal side of main ridge, 430 m from toe	-4.7 ± 1.1	$+2.0 \pm 1.1$
Proximal side of main ridge, 470 m from toe	-4.8 ± 1.1	$+2.7 \pm 1.1$

Suppose there is a substantial upward flow of ice beneath the moraines, as suggested by the dashed line at the top of Figure 12B, and that, due to insulation by the debris cover, ablation beneath the moraines is substantially less than upward flow. Under these conditions the moraines will move upward as well as outward but nearby debris-free areas will not change in elevation as long as ablation equals upward flow in them. Thus the moraines will be pushed up to form ridges above the general level of the ice surface. It is important to note that this upward "pushing" is the result of differential ablation superimposed on a general upward flow of ice; it is not the result of differential flow within the ice.

As topographic relief increases, debris begins to slump off the higher steeper ridges, thus exposing the underlying ice. Some of the slumped debris accumulates on the nearby glacier surface, some rolls or slides across the wind-drift ice wedges to be deposited in front of the ice sheet, and some eventually finds it way into melt streams and is washed off the glacier.

Ice exposed by slumping generally melts rapidly because it is covered by a thin layer of dirt which absorbs heat, and because winter accumulation on the higher ridges is negligible. Ablation tends to reduce the slope of the slump scar until debris released by the melting ice can again accumulate on it and insulate it. Clearly this process will result in high ablation rates during some melt seasons and lower ablation rates as the debris cover reforms and the ridge redevelops. Thus, when averaged over a sufficiently long time, it appears that ablation can equal upward flow beneath the moraines. Furthermore, if debris-carrying ice becomes exposed on a debris-free ramp which is otherwise in equilibrium, the above discussion suggests that ridge-shaped moraines *can develop and persist without departure from equilibrium.* Bishop (1957, p. 18) and Goldthwait (1951, p. 570, 574) apparently felt that for the ridges to form a retreat, or negative mass balance, was required. Swinzow (1962, p. 13–14) realized that such ridges could be built by a glacier with a balanced budget.

Mass budget on ramps and ice cliffs

The mass budget on ramps is inferred to be similar to that shown in Figure 12B. Moraines are not present, but wind-drift ice wedges are, and the convex surface of the ice sheet (Fig. 4; flag line 3) results in a comparable wind shadow in which drifting snow accumulates. Thus net ablation on these wedges probably averages close to zero, as does upward flow (Davis, 1967). For example, net ablation on the third, fourth and fifth flags on flag line 3 (Fig. 4) averaged 0.13, 0.62 and 0.88 m year^{-1}, respectively, between 1966 and 1968. Flags 6 to 21 on this line had melted out by mid-July 1968, implying a minimum net ablation rate of about 1.1 m year^{-1}. Flags 1 and 2, on the wind-drift ice wedge, were not recovered in 1968 and were presumed to be buried.

The mass balance in ice cliffs differs from that in moraine or ramp areas. Ice cliffs occur along the east-south-east-trending part of the glacier margin. The orientation of wind-drift ice wedges (Figs. 1 and 2) indicates that prevailing winds are from the east-south-east and are thus roughly parallel to this part of the margin. The absence of persistent snowdrifts burying the ice cliffs suggests that these winds keep the cliffs swept clear of drifting snow.

Alternating with the ice cliffs are ramp areas which are down-wind from some local wind obstruction. The ramp at the down-glacier end of flag line 3 is sheltered by a slight northward bend in the glacier margin just up-wind of the flag line. Similarly, the ice-cored moraine west of flag line 3 shelters the area east of the main ice cliff.

The absence of drifts below the west ice cliff cannot be attributed entirely to wind, as part of the cliff is perpendicular to the prevailing wind direction. However, a sizeable melt stream is formed at this point by drainage from a large area of the ice sheet. This stream flows along the base of the west-facing part of the cliff and, either alone or in conjunction with wind eddies, is inferred to be responsible for clearing the cliff of drifting snow.

Because there appears to be negligible flow of ice into wind-drift ice wedges, these wedges could not exist if drifting snow did not accumulate on them. Thus in the absence of such drifting, the equilibrium profile would be established at the point where outward flow of the glacier was balanced by ablation (including dry calving). Because ice at the surface generally moves faster than that near the bed, there will be a tendency for a cliff to form. The steepness of the cliff will be limited by dry calving (Goldthwait, in preparation*) which occurs more frequently on overhanging cliffs and tends to eliminate such overhangs. Furthermore, as the slope of the ice front increases, the velocity gradient near the surface decreases; Goldthwait (1960, p. 39) found nearly uniform flow rates in the upper 60 to 75% of the Nunatarssuaq ice cliff. This also results in a tendency for a vertical cliff to be the equilibrium form.

Because ice in cliffs is unconfined, stresses sufficiently high to produce flow can occur at much shallower depths. Thus the free faces of cliffs are normally less than 30 m high in the study area.

Clearly, drifting and ice-flow patterns will vary in detail from place to place. Thus all gradations from gently sloping ramps to vertical ice cliffs are possible, and many parts of the margin in the study area have intermediate slopes.

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Ramps versus ice-cored moraines

There remains the question of why, in some places, there is enough debris in the active ice to form ice-cored moraines, whereas elsewhere there is little or no debris. Available information on bed topography (Bishop, 1957, p. 24; Röthlisberger, 1959, p. 5; Davis, 1967, p. 11) suggests that ice-flow patterns near the bed may differ substantially from those at the surface and that ice reaching the surface on ramps has never been on the bed, or has moved only short (<1 km) distances over the bed. For instance, there is a south-east-draining valley in the bed beneath the TUTO ramp, and basal ice entering this valley from the north-east and east may be diverted to the south. Similarly, the fact that the ice-cored moraine north of flag line 3 is perpendicular to the ramp at the distal end of flag line 3 suggests that basal ice in this area is moving north-westward. Considering the local geometry of the ice sheet, it seems likely that such north-westward flow may be predominant farther out on the glacier and that the south-west flow at the distal end of flag line 3 is a local effect.

Table V. Time required to form moraines in Figure 3

(1) *Profile (Fig. 3)*	(2) *Estimated volume* of debris in block of ice 1 m wide and 1 m deep beneath trench* m³ *Solid*	*Banded*	*Sandy-amber*	*Amber*	*Total*	(3) *Total volume of debris in strip of moraine 1 m wide along profile* m³
A	1.26	0.0006	0.0008	0.0001	1.26	11.7
C	1.30	0.107	0.006	—	1.41	24.8
D	1.05	0.001	—	—	1.05	10.1
F	0.50	0.063	0.00006	—	0.56	10.1

	(4) *Thickness of ice that must melt to yield volume of debris in moraine* m	(5) *Height of moraine crest above smooth profile of ice sheet* m	(6) *Time required to melt ice, assuming ablation rate of 0.3 m year⁻¹* years	(7) *Time required to produce observed relief with upward flow of 0.3 m year⁻¹* years	(8) *Total time to form moraine* years
A	9.3	4	31	13	44
C	17.6	9	59	27	86
D	9.6	>3	32	>10	>42
F	18.1	3	60	10	70

* Per cent debris by volume taken from Table IV except for solid bands. Solid bands assumed to be 100% debris by volume.

Rates of moraine formation

The steady-state model illustrated in Figure 12 can be used to estimate the length of time required to form a given ridge. Assume that the present distribution of debris in ice beneath the moraines in Figure 3 is representative of the distribution during the life time of the moraine. Assume further that the debris concentrations determined in the ice tunnel (Table I) are applicable to these dirt bands. Then the total volume of dirt in a block of ice 1 m wide and 1 m deep beneath each trench, V_i, can be estimated from detailed logs of the type of ice exposed in the trenches (Table V, column 2) and the volume of debris in the till cover, V_m, can be determined from till-thickness measurements (Table V, column 3). Dividing the second of these volumes by the first gives the thickness of ice that would have to melt to yield the observed till cover (column 4). The steady-state model requires that mean annual net ablation beneath the moraines approximately equals upward flow. Using the upward-flow rate of 0.3 m year^{-1} calculated earlier, an estimate of the time required to melt out the volume of debris in these moraines is obtained (column 6). Because no ridge would exist on the ice surface prior to deposition of the first debris, the time necessary for the observed relief to

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develop should be added to this figure. This time is equal to the height of the ridge, h, divided by the mean upward flow.

This can be expressed formally as

$$h = \sum_{i=1}^{n} (U_i - \text{abl}_i) = \sum_i U_i - \sum_i \text{abl}_i.$$

where U_i and abl_i are the upward velocity and net ablation beneath the moraine in the ith year and n is the age. We assumed that $U_i = 0.3$ m year^{-1} = constant and that for V_m to accumulate $\sum_i \text{abl}_i = V_\text{m}/V_\text{i}$. Thus

$$n = \frac{1}{0.3}\left(h + \frac{V_\text{m}}{V_\text{i}}\right).$$

Net ablation, abl_i, may vary from year to year.

The estimates of total time required to form the moraines in Figure 3 range from 42 to 86 years (column 8). The major sources of error in this calculation are the estimate of upward flow and the assumption that dirt bands presently exposed at the ice–till contact are representative of those that have been exposed there in the past. With regard to the second error, the concentration of cobbles and boulders appears to be much higher in the till than in the underlying ice, suggesting that sand and silt may have been washed or blown away, and that the time estimates are too low. However, Bishop (1957, p. 37) found that dirt mounds 0.20–0.50 m high can form in one ablation season, and that an existing mound 1.20 m × 1.85 m in plan and 0.30 m high increased in size to 1.25 m × 2.15 m × 0.35 m in a period of less than 1 month. These figures are comparable to those calculated above.

Two conclusions can be drawn from these calculations. First, it appears that significant amounts of debris are added to moraines only when a solid dirt band is exposed at the ice–till contact (Table V, column 2). Concentrations of dispersed debris in the ice are so much smaller as to be nearly negligible. Secondly, moraine building may be a very sporadic process. Ridges may develop in a few years when a solid dirt band exists at the ice–till interface, and may persist for a long time with little or no further addition of debris.

Concluding statement

The mass-balance requirements of ramps, ice-cored moraines and ice cliffs have been analyzed and it was concluded that all three of these forms could form and persist simultaneously on a glacier with a balanced mass budget. Furthermore, there is no intuitive reason why all three forms could not persist during a slow advance or retreat of the margin, as long as such advance or retreat did not alter the drifting pattern of wind-blown snow. This contrasts with Goldthwait's* conclusion that "ice cliffs on land are a product of an advancing ice margin", and that during retreat ice cliffs turn into ramps. I submit that a change in the mass balance is neither a sufficient nor a necessary condition for a change in the margin. To change from a ramp or ice-cored moraine to a cliff, or vice versa, requires primarily a change in the drifting pattern of snow. This need not result from a change in mass balance, and could occur without a change in mass balance. A change in mass balance alone, unless extreme, is probably not sufficient to cause a change in form of the margin.

The depth of wind-drift snow over and above the normal annual accumulation need not be large. Indeed any positive or negative deviation from the mean accumulation will change the amount of upward or downward flow necessary to maintain equilibrium. If such deviations occur in the same place each year, as in the case of drifting patterns controlled by permanent or quasi-permanent topographic features, the local topography of the glacier surface will tend towards a dynamic balance which reflects the effects of such drifting.

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