Studies on Ten-metre Firn Temperatures, Moraines and Blue Ice Fields in western Dronning Maud Land

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A study was made of 10 m firn temperatures to find the influence of altitude on the mean annual surface temperature and to demarcate the different facies on the ice-sheet. A lapse rate of 0,434°C/100 m was found for the part of the ice sheet above the saturation line at 320 m. From 320 m down to about 50 m a marked deviation between the 10 m firn temperatures and mean annual surface temperatures was observed. This is explained by surface melting and downward transfer of heat at these lower altitudes.

The moraines at Grunehogna, Ovenuten and Jekselen are regarded as typical shear moraines. Other possible modes of formation are briefly discussed.

Crevasse-patterns at Grunehogna were studied and are discussed in the text.

'n Studie van 10 m-sneeutemperature is gemaak ten einde die invloed van hoogte op die gemiddelde jaarlikse oppervlaktemperatuur te bepaal en die verskillende fasies op die yskop af te baken. 'n Vervaltempo van 0,434°C/100 m is gevind vir die deel van die yskop bokant die versadigingslyn by 320 m. Tussen 320 m en ongeveer 50 m is daar 'n opmerklike afwyking tussen die 10 m-sneeutemperatuur en die gemiddelde jaarlikse temperatuur. Dit word skynbaar veroorsaak deur die smelting van sneeu aan die oppervlak en 'n afwaartse verplasing van hitte by hierdie lae hoogtes bo seespieël.

Die moreenafsettings by Grunehogna, Ovenuten en Jekselen word as tipiese skuifskeurmorene beskou. Alternatiewe ontstaanwyses word kortliks bespreek.

Ysskeurpatrone by Grunehogna is ook bestudeer en word bespreek.

Firn temperature observations

Several distinct zones in the temperature regime of glaciers and ice sheets can be distinguished (Fig. 1).

On the high polar ice sheet the temperature of the firn at a depth of 10 m is approximately equal to the mean annual air temperature at the surface *(Loewe*, 1956). Therefore, in order to determine the mean annual surface temperature on the ice sheet near Sanae and the mountain base, 10 m temperatures were obtained at different altitudes between Grunehogna and Sanae (Fig. 2).

Temperatures were measured by a lagged alcohol thermometer which was lowered down a hole. The hole was covered and the temperature allowed to stabilize for a minimum of one hour. Normally stability was reached after about 30 minutes.

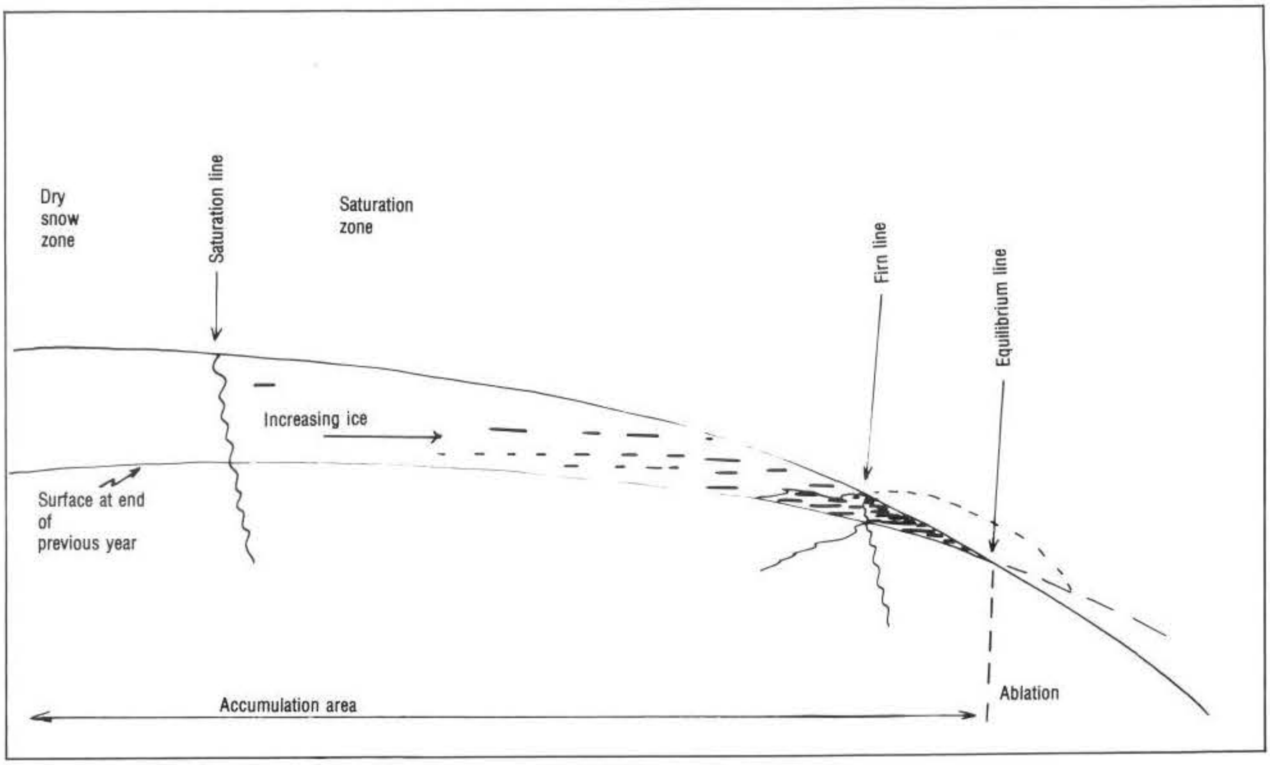

Fig. 1. Schematic zonation of the accumulation area.

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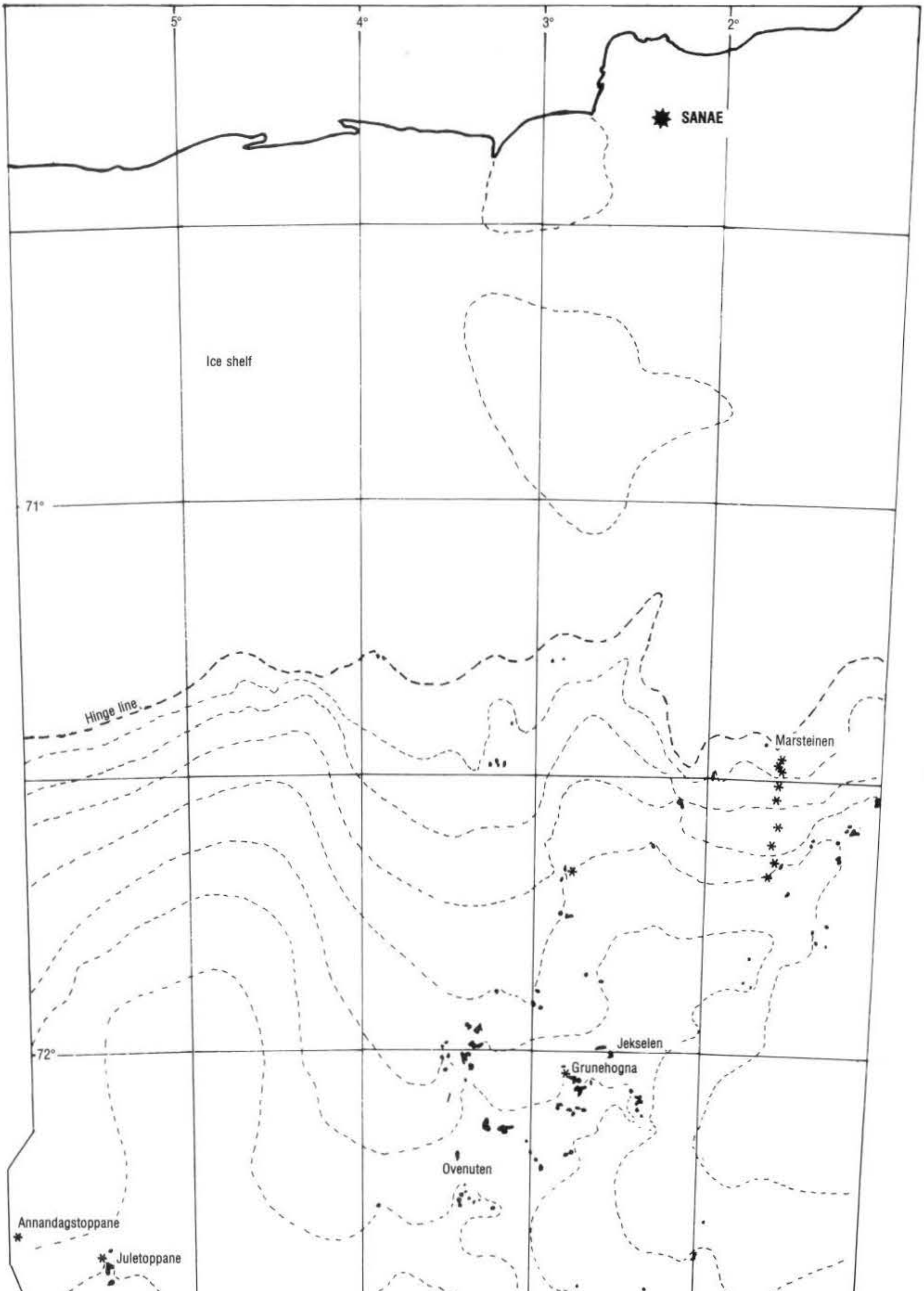

Fig. 2. Part of the Ahlmannryggen with Sipre-hole localities (asterisks).

Temperature above the saturation line

The 10 m temperatures obtained from holes made with a Sipre drill along a traverse from Juletoppane to Marsteinen are plotted in Fig. 3. On this traverse the first thick ice layers were encountered at an altitude of about 320 m. According to the definition of *Benson* (1962) this altitude represents the saturation line, marking the upper limit of complete soaking of the firn. The 10 m temperatures measured above this line represent the mean annual air temperature. When plotted against elevation, these temperatures fall on a straight line, determined by the least-squares method, for altitudes between 325 m and 1 100 m. This straight line gives a lapse rate of 0,434°C/100 m. This is somewhat lower than the values obtained from different stations along the periphery of Antarctica. *Liljequist* (1957) discussed the meteorological explanation for this. The extension of the mean annual lapse rate line (the dashed line in Fig. 3) intersects 54 m (the elevation of Sanae) at −16,3°C and sea-level at −16°C. *Burdecki* (1970) gives the mean annual surface temperature at Sanae and Norway Station over the period 1957-68 as −17,41°C and *Du Plessis* (1973) that at sea-level as −16°C. This indicates that different lapse rates apply for different altitudes from 325 m to sea-level. The *average* of these different rates is still 0,434°C/100 m as the extension of this lapse rate line intersects sea level at exactly −16°C.

The 10 m temperatures at Annandagstoppane and Juletoppane deviate from the lapse rate line of 0,434°C/100 m as a result of increasing continentality (distance from the coast) and, to a lesser extent, difference in latitude.

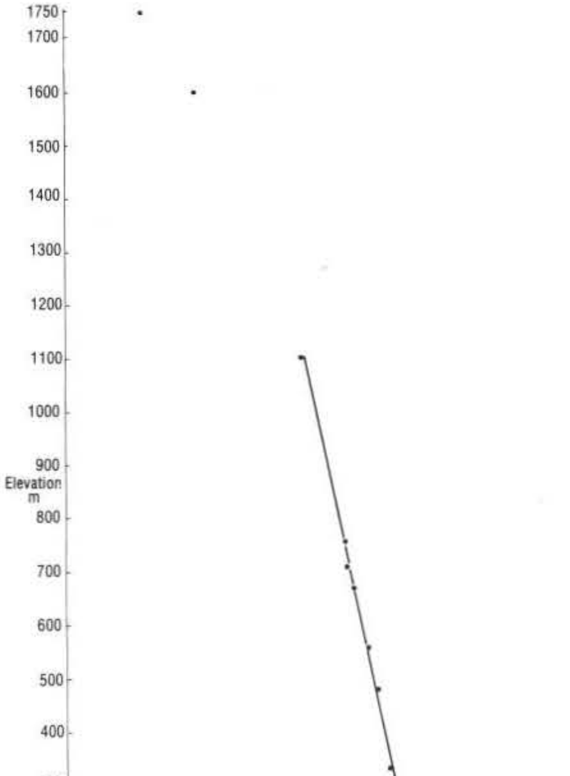

Fig. 3. 10 m firn temperature as a function of elevation.

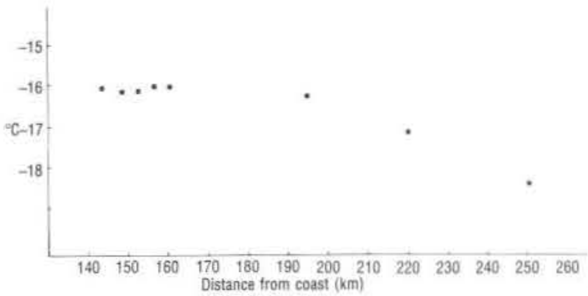

Fig. 4. Influence of distance from coast on sea-level temperatures.

In an attempt to separate the altitude effect from the other factors, all annual mean temperature values were reduced to sea-level temperatures by use of the lapse rate of 0,434°C/100 m and the resulting values were plotted as a function of distance from the coast (Fig. 4). Observations were unfortunately too widely spaced for unambiguous conclusions, but it seems that there is a definite trend towards lower sea-level temperatures at distances further than about 200 km from the coast. The latitude effect is probably insignificant as the interval between the most northerly and most southerly observations is only about 1° of latitude.

Temperatures in the saturation zone

The 10 m temperatures do not reflect the annual mean surface temperature in areas below the saturation line (approx. 320 m). The approximate surface mean temperature can be ascertained by extending the lapse rate line as determined at higher levels.

The deviations between mean annual air temperature and 10 m temperature start at 325 m, reach a maximum at about 220 m, and then decrease again.

Stratigraphic studies of drill cores have shown that the zone between about 350 m and 140 m is a soaked-firn zone where descending melt-water from the surface transports heat down to lower layers. Surface melting increases from the saturation line to the firn line because of rising summer temperatures. Thus, in the upper part of this zone, firn and ice alternate but firn predominates; in the lower part several firn layers but more ice occur and at the firn line there is predominantly ice. Although there is more melting at lower elevations the increasing amount of ice (impervious to water) decreases the downward transport of heat. Therefore there is a point between the saturation and firn lines where there is a maximum downward transport of heat due to percolation; this transport of heat decreases gradually towards the saturation line and decreases sharply near the firn line. In the area under discussion this point lies at an elevation of 200 m.

Because of shortage of time this study was not extended to the ice shelf. *Du Plessis* (1973) carried out a limited investigation of temperature conditions on the ice shelf around Sanae and reported values that are lower than those suggested for these altitudes by the extension of the lapse rate line in Fig. 3. This can be explained by the fact that the inland ice sheet is grounded, and that geothermal heat flow from the underlying rocks will result in an appreciable temperature gradient. *Cameron* (1964) found a positive temperature gradient of 4,9°C/100 m at Wilkes Station (110°E, 66° 15′S), and *Bogoslovski* (1958) found a gradient of 5,1°C/100 m at Mirny (83°E, 66° 30′S).

It is also of interest that very little evidence of downward transport of melt-water was observed in the top 10 m of the ice shelf north of Marsteinen.

Conclusions

The study of snow and ice temperatures from 150 m to 1 100 m gives a mean lapse rate of 0,434°C/100 m. This represents also the *average* lapse rate down to sea-level, but may deviate markedly from local lapse rates at these lower elevations. Above 1 100 m a higher lapse rate applies as a result of increasing continentality.

A marked change in temperature regime occurs at an elevation of about 320 m, probably due to the downward percolation of melt-water, the change from firn (high albedo) to ice (lower albedo) and subsequent increased absorption of incoming solar radiation, and also due to the increased influence of the geothermal gradient on the thinner ice found at lower elevations.

The 10 m temperatures on the ice shelf around Sanae are lower than expected from the lapse rate of 0,434°C/100 m because of the very low geothermal gradient on the floating ice shelf and also because of the lack of downward heat transfer by melt-water.

Moraines and blue ice fields

Since the entire ice sheet of western Dronning Maud Land belongs to the accumulation zone, moraine features are relatively scarce. There are a few places where ablation locally dominates accumulation, e.g. the blue ice fields north of Jekselen, south-west of Grunehogna 1285, and west of Ovenuten where moraine deposits are consequently exposed. The occurrence at Grunehogna has been studied in some detail and, as all these deposits are considered to have been formed by the same processes, the discussion will be limited to this locality.

On the south-west side of Grunehogna 1285, between the blue ice field and the foot of the mountain, there is a series (3-5) of parallel talus ridges (Fig. 5). The material is completely unsorted, ranging in size from 20 cm to 2 m, and no preferred orientation of the elongated boulders was observed. Some of the bigger boulders are fairly well rounded with well-developed striae and chatter-marks on others.

Three possible ways of formation can be postulated for these deposits:

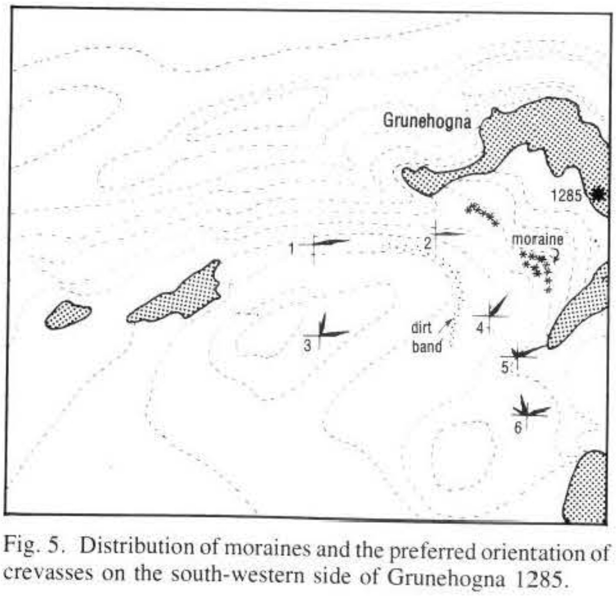

Fig. 5. Distribution of moraines and the preferred orientation of crevasses on the south-western side of Grunehogna 1285.

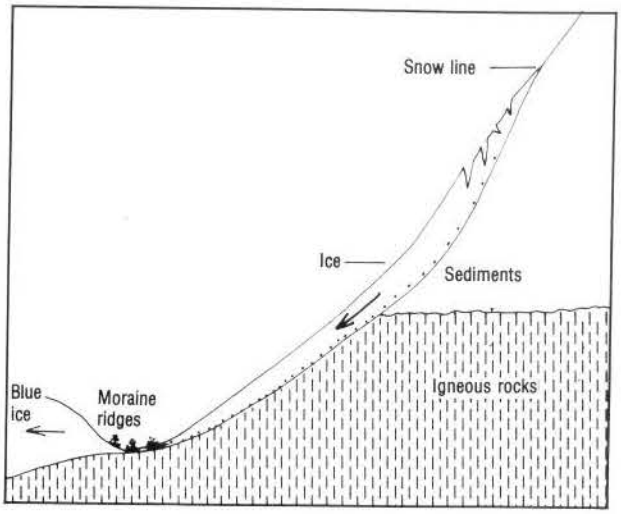

Fig. 6. Schematic representation of the ice-covered southwestern side of Grunehogna 1285.

(i) Mechanical weathering (frost-shattering) is an important erosion process in Antarctica. Rocks loosened at the top of an outcrop will roll down and, guided by the contours, come to rest in a number of favourable places. This process, however, cannot account for the rounding of some boulders and for the striae. It is also of significance that about 15-20% of the moraine material consists of igneous rock and that no igneous rocks are exposed above the snowline at Grunehogna (Fig. 6). The arrangement of the material in several parallel ridges also points to some other means of formation.

(ii) Blue ice with a thin layer of snow extends halfway up Grunehogna 1285. This ice is slowly moving downwards and is continually replenished by new snow falls, so that the snow-line will not fluctuate much. It is highly probable that this moving ice will transport loose pieces of rock. If the bedrock lies at a shallow depth under the trough at the foot of the mountain, this loose debris will again appear and be deposited there. Such a mechanism will account for the igneous rocks in the deposits, but it is doubtful if a significant part of the deposit was formed in this way.

(iii) The deposits can be of true glacial origin, i.e. deposited by a moving glacier, which would account for the fair roundness of some boulders and also for the striae.

Schytt (1961) described several moraines further to the south (near Borga) as typical shear moraines. According to him, the morainic material was brought up along different generations of shear-planes developed during a previous retreat of the ice sheet. As the moraine reduces the ablation of the underlying ice, shear-plane moraines will stand out as pronounced ridges.

Shear-plane moraines have been studied intensively in Baffin Island and in the Thule region in Greenland, and *Weertman* (1961) has proposed that they be called Thule-Baffin moraines in preference to the genetic term "shear-plane" moraine.

The Thule-moraines have been studied in detail by *Goldthwait* (1951). The moraines that he described are around the edge of the Barnes Ice-cap, which is cold and similar to the ice sheet under discussion, having a mean

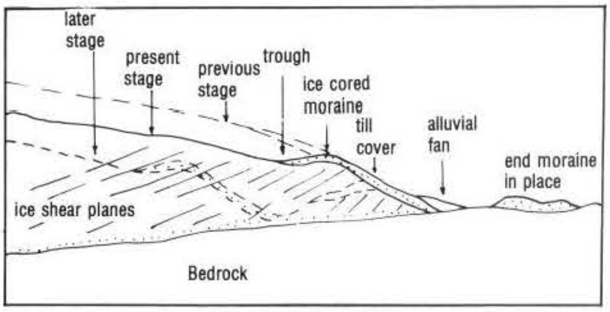

Fig. 7. Thule-Baffin cold ice moraine formation.

temperature of −10°C. The material forming the supraglacial debris is thought to have come to the surface along shear-planes from beneath the glacier, where the ice is thinning near its edge. In the Thule region the ice 30-60 m above the steep toe of the ice sheet is layered by many dirt-bands, striking almost parallel to the ice edge. This material is mostly fine dirt, with occasional pebbles and even boulders, and *Goldthwait* (1951) suggested that these layers mark shear-planes dipping down towards the bed of the ice sheet. At Grunehogna a dark band follows the contours on the blue ice field at Peak 1285 (Fig. 5). It is caused by fine debris (2-5 mm) just below the surface. In the windscoop on the north-east side similar dark bands are exposed which probably mark shear-planes in the moving ice mass adjacent to the outcrop.

According to Goldthwait, the material is brought up along the shear-planes from a dirty zone near the ice margin. The concentration of dirt only takes place where the ice thickness is less than 85 m. The dirt-covered ice melts more quickly than the uncovered ice, resulting in a steepening of the dirt-covered slope. Material then moves down the steeper slope and accumulates at the bottom to form a thick layer of debris. Thus, states Goldthwait, an accumulation of till 0,6-1,2 m thick forms on the last 30 m of slope and a depression develops on the junction of the dirty ice slope and the thicker till layer on the outer margin as a result of the more rapid melting under the thin dirt cover than under the thicker till layer. This trough separates the developing ice-covered moraine from the dirty ice beyond. In time the ice core melts and a new low terminal moraine will be left (Fig. 7).

Extensive crevassing is a feature of the blue ice field at Grunehogna and a study was made of the strike directions of these fractures in order to try and form an idea of the movement of the ice mass. In areas 1, 2, 4 and 5 (see Fig. 5) the preferred direction is parallel to the contours and it is clear that here tension down the slope (i.e. perpendicular to the contours) is the dominant crevasse-forming factor. In area 3, two intersecting sets of crevasses, the one striking E-W and the other N-S, are found. If these crevasses were formed by tangential shearing it would imply a force (movement) from the south-west to the north-east. Area 6 shows a similar pattern with an additional set of crevasses striking ESE and parallel to the contours. The sets striking N-S and ENE were formed by movement of the glacier *en masse* and the ESE set by tension along the slope.

Another conspicuous feature of the blue ice fields is a secondary layering of alternating white bubbly ice and clear blue ice. These layers are up to 1 m thick and are regarded as glacier foliation *(Embleton & King*, 1967). The shearing is developed only near the margins of the ice fields and usually runs parallel to them as well as to the dirt bands which mark shear-planes. The precise origin is still in doubt, but in general it is considered to be a structure resulting from deformation of ice during flow *(Embleton & King)* as a result of intense compression or shear.

Conclusions

The moraines at Grunehogna, Ovenuten and Jekselen appear to be true moraines of the Thule-Baffin type. The material was apparently brought to the surface along shear-planes between stagnant zones (adjacent to nunataks) and faster moving ice. The parallel ridge-like pattern of the moraines proves that the ice sheet is retreating locally. This does not, however, imply a general decrease of the snow and ice over western Dronning Maud Land. *Schytt* (1961) showed that the ice sheet in this region is in equilibrium.

The alternating blue and white ice bands observed in windscoops are glacier foliation formed by shear or compression on the margins of the moving ice mass.

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