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GLACIAL EROSION BY THE LAURENTIDE ICE SHEET AND ITS RELATIONSHIP TO ICE, TOPOGRAPHIC AND BEDROCK CONDITIONS

by

D. E. SUGDEN

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The aim of this project was to map and analyse landscapes of glacial erosion associated with the Laurentide ice sheet and to relate them to the main variables affecting glacial erosion: basal thermal regime of the ice sheet, the topography and geology of the bed. A reconstruction of the basal thermal regime of the Laurentide ice sheet was carried out using and adapting a model developed for existing ice sheets. Using Landsat imagery, maps of landscapes of glacial erosion were compiled for the whole of the Laurentide

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ice sheet area with more detailed maps for the eastern Canadian Arctic. A conceptual model is developed.

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MAP

Distribution of landscapes of glacial erosion in the eastern Canadian Arctic, compiled from LANDSAT imagery and conventional air photographs. Scale 1:50,000,000.	Inside back cover

SUMMARY

The aim of the research project was to map and analyse landscapes of glacial erosion associated with the Laurentide ice sheet and to relate them to the main variables affecting glacial erosion:- basal thermal regime of the ice sheet, the topography and geology of the bed. A reconstruction of the basal thermal regime of the Laurentide ice sheet was carried out using and adapting a model developed for existing ice sheets. Using LANDSAT imagery, maps of landscapes of glacial erosion were compiled for the whole of the Laurentide ice sheet area with more detailed maps for the eastern Canadian Arctic. A conceptual model is developed and hangs round the following postulates:-

(1) Landscapes of glacial erosion are related primarily to the basal thermal regime of the ice sheet.

(2) Intense areal erosion is favoured above all by the availability of efficient mechanisms of debris evacuation, rather than mechanisms of abrasion or fracture.

(3) Landscapes of glacial erosion are equilibrium forms related primarily to maximum glacial conditions. This latter conclusion implies that at some stage in the Pleistocene the Laurentide ice sheet was in a stable maximum condition for a long period of time.

STATEMENT OF PROBLEM

The aim of the project was to test and refine theories of glacial erosion by field testing. In particular it was hoped to test the hypothesis that different landscapes of glacial erosion could be related to varying basal thermal regimes beneath the former Laurentide ice sheet. It was proposed to use mathematical models to predict the former basal thermal regime and then to compare the result with maps of erosional landscapes of the same area. Such a comparison was intended to throw light on the relative importance of glacial, topographic and geological variables in influencing the nature of glacial erosion. A subsidiary result of the work was to be the production of two maps of landscapes of glacial erosion, one of Baffin Island at a scale of 1:1,000,000 and one of northern Canada at a scale of 1:20,000,000.

BACKGROUND

The main reasons for the research project were outlined in the initial grant proposal as follows:-

"(1) To help understand the role of glaciers as erosive agents on the earth's surface. In particular the study will clarify the question of whether the Laurentide ice sheet accomplished deep erosion (White, 1972) or relatively minor erosion (Bird, 1967). Clearly the answer will have important implications concerning such problems as the rates and total amount of erosion achieved by ice sheets, the reasons for the widespread variation in the effects of ice sheets on landscape, the origin of till and the nature of the processes of glacial erosion. At present there are widely divergent views on all these problems.

(2) To develop a geomorphological line of enquiry which has not contributed its potential to the study of glaciers. For a variety of reasons it can be argued that glacial geomorphology is characterised by a surfeit of qualitative description at the expense of quantitative description, a surfeit of description of form at the expense of an understanding of the processes involved, and a failure to apply glaciological theory. As a result the field of study has not progressed effectively in recent decades and is unable to contribute much useful expertise to glaciology. It is significant that at the Cambridge meeting of the International Glaciological Society in the

Autumn of 1974 there were several pleas from glaciologists for information on glacier beds and the processes operating at the ice/rock interface. Lack of information is limiting the development of more sophisticated models of glacier behaviour. This Baffin Island proposal is in effect a study of the morphological characteristics of the glacier bed of part of the Laurentide ice sheet.

(3) To attempt a project on a scale which is suitable for the application of LANDSAT imagery. In the past most work in glacial geomorphology has concentrated on small scale problems. Now with imagery available at a large scale the time seems appropriate for an attempt to tackle problems at a larger scale."

The project was carried out much as planned with nine months of the study period based at the Institute of Arctic and Alpine Research, University of Colorado, and a summer field season in Baffin Island. Also as planned, I spent over two weeks in Ottawa where I was able to examine LANDSAT images and air photographs and discuss my work with Officers of the Geological Survey of Canada.

During the course of the work it was felt necessary to change the emphasis and scope to some extent. It proved difficult to reconstruct the dynamic characteristics of the former Laurentide ice sheet without approaching it as a whole. Accordingly the scope of the project was broadened to encompass the whole Laurentide area and the detailed studies involved the whole of the eastern Canadian Arctic (rather than just Baffin Island as planned). As a result of this broadening of the scope of the project the original scales of the two maps of erosional landscapes were inappropriate and the maps are now combined to include one of the eastern Canadian Arctic at a scale of 1:5,000,000.

APPROACH

Three main and complementary approaches to the study of erosional landforms seem appropriate. Firstly, the problem can be tackled in the laboratory. Secondly, the problem can be tackled by direct observation beneath glaciers, as for example demonstrated by Kamb and LaChapelle (1964) and Boulton (1972, 1974). Thirdly, the problem can be tackled by testing theoretical predictions in the field. This holds particular promise in formerly glaciated areas once covered by mid-latitude ice sheets where the former glacier bed is accessible for study. The present research project adopted the last approach.

There were three main strands to the project:-

(1) Review and critical assessment of alternative views on the amount of glacial erosion by ice sheets. In particular it was necessary to consider the hypothesis of W.A. White (1972) which was in complete disagreement with the work carried out here.

(2) Reconstruction of the morphology, dynamics and thermal characteristics of the Laurentide ice sheet at its maximum.

(3) Compilation of maps of landscapes of glacial erosion and the relationship of the pattern to ice, topographic and geological variables.

Each theme is the subject of a separate paper and these form the body of this report. Theme one is covered in the paper entitled "A Case Against Deep Erosion of Shields by Ice Sheets", and has been published in <u>Geology</u> in September 1976. Theme two is the subject of a paper entitled "Reconstruction of the Morphology, Dynamics and Thermal Characteristics of the Laurentide Ice Sheet at its Maximum", and has been accepted for publication in <u>Arctic and Alpine Research</u> in January 1977. Theme three is covered in a paper entitled "Glacial Erosion by the Laurentide Ice Sheet". It is planned to submit the paper to the <u>Journal of Glaciology</u> in due course. In this report the Figure numbers run consecutively and there is one consolidated bibliography at the end.

THEME 1

A CASE AGAINST DEEP EROSION OF SHIELDS BY ICE SHEETS

ABSTRACT

The hypothesis of deep erosion of shields by continental ice sheets put forward by W.A. White (1972) is being used by neighbouring disciplines. Although valuable as a speculative hypothesis, it is felt that there is field evidence to contradict the hypothesis. This evidence is discussed and it is concluded that Pleistocene glacial erosion has removed no more than a few tens of metres of material from the shields of the northern hemisphere.

INTRODUCTION

In 1972 W.A. White presented a case for widespread and deep erosion of shield areas by ice sheets. The view was in sharp contrast to a long-held view epitomized by Flint (1971) that shield areas had experienced relatively little lowering by glacial erosion during the Pleistocene. The presentation of such a bold hypothesis by White is valuable because it draws attention to the nature of the available evidence, challenges assumptions and raises new questions. Nonetheless, there may come a point where a speculative hypothesis in one discipline becomes accepted and uncritically incorporated into the beliefs of neighbouring disciplines. This is especially likely to occur if a hypothesis seems also to be a synthesis of past observations.

There are indications that this pattern of events applies to White's hypothesis. Dent (1973) used the contrasts in depth of glacial erosion postulated by White to explain the asymmetric age distribution of impact craters on the Canadian shield. Boulton (1974) used some of White's conclusions in support of his broad theory of glacial erosion. Moreover, at the time of writing, White's views are being used as a basic input parameter in the reconstruction of the characteristics of the former northern hemisphere ice sheets being carried out as part of the CLIMAP project (T.E. Hughes, Personal Communication, 1976). If the hypothesis of White is to be widely used, then it is important that the evidence on which it is based is weighed against any contradictory evidence.

Only in this way can the full worth of the hypothesis be assessed. It is the purpose of this paper to attempt such a review by testing key postulates against the available field evidence.

The crux of White's hypothesis is that the spatial correlation of shields with the sites of former ice sheets implies a causal correlation whereby the ice sheets actually exposed the shields by deep erosion. For example, White argued that prior to Pleistocene glaciation central Canada resembled the central United States and was completely covered by Paleozoic sediments. In addition, White postulated two zones of erosion. The first included the central inward sloping parts of the Canadian and the Fennoscandian shields which were argued to have been transformed to basins by massive glacial erosion of overlying Paleozoic and shield rocks. No estimate of the depth of erosion was given, but if the conceptual model (White's Figure 5) is applied to Hudson Bay, a figure of over 1,000 m. of erosion is implied. Hudson Bay and the Gulf of Bothnia were regarded as the largest in a hierarchy of ellipsoidal lakes created essentially by glacial erosion. A second, outward-sloping, peripheral shield zone was also argued to have been exposed by the removal of the overlying Paleozoic rocks by glacial erosion, although in this latter zone, the shield itself was not thought to have been materially lowered.

White recognised that the hypothesis of deep erosion was contrary to the long-held view of minimal glacial erosion in shield areas. He considered that these past views were based on two types of evidence and that they could be explained away. One line of evidence is that the volume of glacial drift created by the Fennoscandian and Laurentide ice sheets reflects only limited lowering of the central areas by ice erosion. White argued that such a conclusion could not be justified, since no one knows how much sediment has been washed into the ocean. The other line of evidence is that the existence of weak sediments, for example in the area around Hudson Bay (Prest, 1970), argues against deep erosion by ice sheets. White argued that this simply reflects that ".... ice sheets can be able or feckless erosive agents in the same place at different times." (White, 1972, p. 1037). A reader of White's paper might reasonably assume that there are no other grounds for suggesting that ice sheets accomplished relatively little lowering of shield areas. There exists, however, a wide and varied body of evidence in support of a view of limited glacial erosion.

The two fundamental pillars of White's hypothesis are the postulates that: (1) Hudson Bay and the Gulf of Bothnia occupy massive basins eroded by Pleistocene ice sheets; and (2) the shield rocks have been exposed by deep glacial erosion. It is felt that the validity of both postulates can be tested by reference to the available evidence in the northern hemisphere shield areas.

DEEP GLACIAL EROSION OF THE CENTRAL SHIELD AREA

A TEST IN HUDSON BAY

The Hudson Bay basin and the Gulf of Bothnia were argued to be depressions because they have been eroded by ice. Moreover, ".... these largest basins seem to owe little to structural or lithologic weakness or to pre-glacial topography ...." (White, 1972, p. 1053). Taking the case of Hudson Bay, there is a considerable body of evidence which does not support such a view.

Thick Paleozoic rocks

Much of the Hudson Bay depression is underlain not by crystalline basement rocks but by Paleozoic rocks (Grant, 1969). Moreover, the Paleozoic rocks display a broad basin structure and are estimated to be at least 1,190 m. thick (Grant, 1969) and probably more than 2,440 m. thick (Hodgkinson, 1969; Hood et al., 1969). The presence of such thicknesses of Paleozoic rocks agrees with the long-held view that the Hudson Bay depression is an old structural feature (see for example King, 1965; Sanford and Norris, 1975).

Pre-glacial valley pattern

Another line of evidence contradicting deep erosion over Hudson Bay is the presence of a dendritic river valley pattern visible on bathymetric maps of Hudson Bay. The river pattern and its reconstructed form are discussed by Bird (1967), Pelletier (1969) and Cumming (1969). In places, parts of the former valleys are filled with Pleistocene sediments (Grant, 1969). The argument for a pre-glacial age for the river valleys hinges mainly on the scale of the pattern:— namely the way it forms part of a subcontinental drainage system and the way the valleys are associated with broad upland surfaces which also form a consistent pattern on a subcontinental scale (Bird, 1967). It is difficult to imagine that such a large drainage pattern could have evolved in anything other than pre-glacial times. The only alternatives - post-glacial or inter-glacial time - are simply too short. The c. 7,500 years of post-glacial time has seen minimal modification of even glaciated rock surfaces, while it is difficult to imagine that even the combined length of inter-glacial intervals could have provided enough time for anything other than the initiation of small scale river patterns.

It is important to stress that the presence of a pre-glacial valley pattern on Paleozoic rocks on the floor of Hudson Bay carries the implication that ice has been unable to modify either the Paleozoic rock cover or the crystalline shield rock beneath the centre of the Laurentide ice sheet.

EXPOSURE OF SHIELD ROCKS BY GLACIAL EROSION

The other fundamental line of argument used by White provides another opportunity to test the hypothesis in the field. White concluded that the spatial correlation of shield rocks with former Pleistocene ice sheets implies that the ice sheets accomplished the exhumation of the shield rocks. If it can be shown that the main features of the glaciated shield areas were in existence in Tertiary times, then this would apparently contradict this line of argument. There would seem to be two main ways of approaching this problem - firstly, by analysis of the deposits around the present limits of the shield rocks and secondly, by the search for preglacial topography and associated deposits within the area underlain by shield rocks.

Deposits around the shield

Gravenor (1975) has used the first approach and has discussed the rock fragment composition and the heavy mineral assemblages in the oldest tills in Illinois and adjacent areas around the Great Lakes. Gravenor was able to conclude from the nature of the heavy mineral assemblages in the oldest tills that the shield was exposed before the first glaciation for which there is evidence. Moreover, from analysis of the way in which shield-derived material varied from place to place, he deduced that the pre-glacial shield boundary must have been in approximately the same position as today.

Mathews (1975) discussed the evidence now available on the Cenozoic sediments off the eastern seaboard of North America. The volume of sediment per unit length of continental margin is lowest in the vicinity of the Canadian shield and this is in a zone where the source area is most extensive. This suggests low Cenozoic denudation rates in the shield area, a conclusion which Mathews notes does not agree with a view of deep glacial erosion.

Pre-glacial topography within the shield

The second approach involves examination of the topography of glaciated shield areas. Many such areas in the northern hemisphere retain Tertiary fluvial features, providing evidence contrary to the view of deep erosion by ice of either a former Paleozoic cover or of shield rocks. When Tertiary deposits exist, then there can be little doubt concerning the dating. When no deposits are known, then the evidence is in the form of former river valleys and associated upland surfaces, and dating is only relative to the subsequent Pleistocene glaciation.

An example of this latter morphological type of argument is illustrated by the writer's work in the glaciated shield area of part of west Greenland, where there are indications that the ice has only etched a pre-existing surface. Over thousands of square kilometres in the area east of the Sukkertoppen ice cap the main rivers maintain integrated drainage patterns and regular valley-floor long profiles reminiscent of non-glacial humid environments. "These patterns occur regardless of the abundant signs of scouring and indeed the only exceptions are in the vicinity of distinct troughs. Since integrated patterns on such a scale are unlikely to have formed on a newly-created glacial surface without greater modification of the glacial forms, one is led to conclude that the ice was unable to change the land surface sufficiently to derange a pre-existing pattern. This view is confirmed by evidence from the summits between the rivers. Adjacent summits tend to be conformable in altitude and a generalised contour map of their elevations produces a surface which is conformable with the direction of river flow. Both lines of evidence imply that a pre-existing surface has been modified by ice without major transformation." (Sugden, 1974, p. 179). It could be suggested that in the above example the ice sheet eroded a uniform depth of rock from the area while maintaining the pre-existing morphology and that the evidence does not preclude deep glacial erosion. However, it is unlikely that an ice sheet, which is notoriously selective in its erosion (Embleton and King, 1975), could erode topography of varying altitude and valleys of varying size and orientation without greater signs of differential erosion.

Arguments similar to those above have been used elsehwere. Perhaps it will suffice to take a few well documented examples from different shield areas. In northwest Scotland an erosion surface at an altitude of 60-180 m. is cut across Precambrian and other rocks. It forms one of several surfaces which, by their association with dated volcanic rocks and relict soils, are thought to have formed during the Tertiary epoch (Godard, 1965; Sissons, 1967, 1976). This surface has been scoured by ice yet its former pediment, associated river valleys and prominent inselbergs are still preserved. It is apparent that modification by Pleistocene ice has been relatively limited. In Sweden similar arguments are developed in detail by Rudberg (1954). Here, a detailed regional study revealed a series of ancient erosion surfaces and an integrated pre-glacial river valley system. The surfaces and river valleys are often largely intact and, indeed, Rudberg has used their reconstruction as a means of assessing the degree of glacial modification. In Norway similar arguments were used by Strøm (1948) when discussing the geomorphology of the uplands of the country. In Finland, Niini (1968) suggests overall glacial erosion of less than 10 m. This estimate is based on several previous research studies on the reconstruction of the morphology of river valley profiles. In Canada there is an impressive array of data about the pre-glacial shape of the Canadian shield. Bird (1967) brought together the results of a series of individual Rand Memoranda describing in detail a large part of Arctic Canada between the Arctic Archipelago and Hudson Bay. The surveys suggested the existence of a series of upland surfaces which are clearly unrelated to present processes and yet which bear signs of glaciation. Some surfaces, for example in Somerset Island, are cut across Paleozoic and Precambrian rocks and therefore cannot have been exhumed from beneath a Paleozoic rock cover (Bird, 1967). Many other surfaces are associated with river valleys and, since there is no known way in which they could have been created by glacial erosion, it seems reasonable to attribute them to pre-glacial subaerial erosion. A recent case of such an argument is illustrated in the Barrow Strait area by Bornhold et al. (1976). The preservation of such surfaces throughout the Pleistocene epoch points to relatively minor modification by glacial erosion. Bird summarised the implications of this type of evidence in the following words: "By the beginning of the Pleistocene epoch the main elements in the landscape of arctic Canada were already in existence. Glacierization led to modifications in detail of the scenery, and only in the Highlands along the east coast was the landscape changed fundamentally by the Pleistocene glaciations." (Bird, 1967, p. 89).

In certain situations there are Tertiary deposits associated with these ice-scoured shield rock surfaces. Such deposits provide firm evidence that the surface on which they lie was in existence before Pleistocene glaciation. Andrews et al. (1972) found a thin layer of early Tertiary limestone on a glaciated shield surface in north-central Baffin Island. The deposit was dated on the basis of the contained micro flora which indicated a warm temperate environment. In Scandinavia deep chemical weathering on certain shield rocks was discussed by Niini (1968). The depth of the weathered rock, its association with surface rock fractures and its character all suggested that a long period of warmer climate was required for its formation. Such a climate was last experienced in the Tertiary epoch.

The various lines of evidence mentioned above all imply that the main features of the shield surfaces were exposed in preglacial times. In such a case it is difficult to entertain a view of deep erosion by ice sheets in shield areas.

CONCLUSION

It has been the purpose of this paper to test White's theory of deep erosion by ice sheets against some of the available evidence in the northern hemisphere shield areas. The two critical postulates which form the basis of White's hypothesis are found difficult to reconcile with the evidence presented. Indeed, the overwhelming conclusion is that ice sheet erosion over shield areas has been restricted to the modification of pre-existing surface characteristics. Rather than deep erosion, one might perhaps be thinking in terms of the removal of material (previously weathered regolith?) with an average thickness of the order of a few tens of metres.

THEME 2

RECONSTRUCTION OF THE MORPHOLOGY, DYNAMICS AND THERMAL CHARACTERISTICS OF THE LAURENTIDE ICE SHEET AT ITS MAXIMUM

ABSTRACT

On the assumption that the Laurentide ice sheet attained a steady-state maximum condition at some time during the Pleistocene, a reconstruction of its characteristics was made largely by analogy with existing ice sheets. The maximum surface altitude was ca. 3,400 m., maximum thickness ca. 4,200 m., and total ice volume 37 x 10^6 km.^{3}. Surface mean temperatures were below -40^{0}C over a large central and northern part of the ice sheet, while accumulation decreased from 10 to 80 cm. water equivalent near the periphery to 5 cm. water equivalent near the centre. Velocities increased from less than 10 m. yr.$^{-1}$ over much of the ice sheet centre to 50 to 800 m. yr.$^{-1}$ near the periphery. Outlet glacier velocities were 300 to 800 m. yr.$^{-1}$. Basal shear stresses increased from <0.25 bars in the central areas to 1 bar near most peripheries. An exception was the high western plains area where surface gradients and basal shear stresses were low. Using a model developed by Budd et al. (1970), the basal temperatures of the ice sheet were calculated. From centre to periphery there were basal zones of warm melting, warm freezing, cold-based ice, and finally a peripheral warm melting zone. Over the Queen Elizabeth Islands the ice was cold-based except over some straits. A sensitivity test suggested that the broad pattern was stable although the exact position of the zone boundaries was tentative. The basal temperature pattern has profound implications concerning processes of glacial erosion and deposition and the stability of the ice sheet.

INTRODUCTION

Someone viewing the course of glacial geomorphology over the last fifteen years might well be struck by the contrast between the great progress made in understanding the pattern and chronology of glacier fluctuations and the relatively meagre progress in understanding process/form relationships. Perhaps the main reason for this stress on historical rather than functional modes of explanation has been the failure to construct an effective working link between glacial geomorphologists and glaciologists. Nevertheless, there is now a powerful core of glaciological theory concerning processes of glacial erosion and deposition and it would seem a propitious time for geomorphologists to test and refine these ideas by study of glacial landforms.

The most accessible areas with information about glacial landforms are the sites of former Pleistocene glaciers where the former glacier bed is now exposed. However, these are the areas for which there is little glaciological information and thus it is difficult to relate form to glaciological processes. The aim of this paper is to reconstruct the basal thermal regime of the Laurentide ice sheet at its Pleistocene maximum. If broad variations can be recognised, then it should be possible to compare the predicted distribution of erosive and depositional processes associated with certain thermal regimes against the landform pattern on the ground. Comparison should then allow refinement of the original theory and improved understanding of process/form relationships. Although this paper is confined to the calculation of the basal thermal regime, the following paper compares the pattern with erosional landscapes (Theme 3).

The importance of basal thermal regime in influencing the processes of glacial erosion and deposition has been stressed above all by Boulton (1972, 1974, 1975) who built on ideas developed by Weertman (1961a, 1966). In particular, Boulton has emphasised the fundamental contrast between processes associated with warm-based and cold-based ice as well as the importance of the spatial distribution of zones of warm- and cold-based ice in influencing how different glaciers erode and deposit. These ideas have been explored in relationship to landscapes of glacial erosion in Greenland (Sugden, 1974) and in the more general context of glacial geomorphology by Sugden and John (1976). A study of the role of varying thermal regime beneath the Laurentide ice sheet was made by Hughes (1973a) who was concerned with its effect on the occurrence and type of permafrost.

THE MODEL

Ice temperatures at the base of ice sheets are influenced by many factors and it is useful to discuss the influence of each before going on to calculate their role in combination:

Ice thickness is a critical variable since, like most earth surface materials, the temperature tends to rise with depth. Although there are important exceptions which are discussed later, the greater the ice thickness the warmer base temperatures tend to be.

Ice surface temperature is important in that it determines one boundary condition. The warmer the surface temperature, the warmer the base tends to be.

The accumulation rate is important because it determines the rate at which accumulating firn is carried down into the ice. As demonstrated by Robin (1955), a high accumulation rate carries cold down into the ice more effectively than a low accumulation rate. Base temperatures will tend to be cooler with a high accumulation rate and warmer with a low accumulation rate.

The rate of surface warming as ice flows from the central cold zone towards lower warmer zones is an important variable which is affected by ice velocity and the vertical temperature gradient on the ice sheet surface. The higher the velocity and steeper the gradient, the greater the magnitude of the effect tends to be. This advection of cold ice has a cooling effect and may cause a decline in temperature with depth in the upper layers of the ice (Robin, 1955).

The geothermal heat conducted into the base of the ice is an important source of heat and its magnitude varies significantly from one geologic structure to another. Other things being equal, the higher the geothermal heat flux, the warmer base temperatures tend to be.

Frictional heat produced by internal deformation or basal sliding as the ice sheet flows is another means of providing heat at or near the ice sheet base. This heat is concentrated near the base since these are the layers subjected to most deformation. The heat supplied will vary with ice velocity and shear stress. At the stresses common in glaciers a velocity of 20 m. per year provides approximately as much heat as that provided by the average geothermal heat flux (Paterson, 1969). The higher the frictional heat produced the warmer base temperatures tend to be

Individually, the role of each variable seems clear. In combination, however, they operate in a highly complex manner. This can be illustrated by considering the full role of an increase in accumulation, which has a three-fold effect, namely to carry cold down into the ice mass (which tends to reduce base temperatures), to produce frictional heat due to higher velocities (which tends to increase base temperatures), and to cause horizontal advection of cold surface ice (which tends to cool base temperatures).

The most successful mathematical model available which treats all these variables and allows calculation of basal ice temperatures has been developed by Budd et al. (1970). In this model a column fixed in space over a given point on the ground is considered and the conduction of heat in ice moving through the column is calculated. For the simplest case with ice thickness Z, accumulation rate A, strain rate A/Z, horizontal advection and warming rate $\frac{D\Theta}{Dt}$ and thermal diffusivity of ice $\varkappa$ (all constant with depth), the differential equation for temperature θ at level z above the bedrock may be written:

$$\varkappa \frac{d^2\theta}{dz^2} + \frac{Az}{Z}\frac{d\theta}{dz} = \frac{D\theta}{Dt} \qquad (1)$$

The boundary conditions used are the basal temperature gradient γ_b and the surface temperature θ_s. For a balanced state the surface warming rate following movement is given by:

$$\frac{D\theta}{Dt} = \frac{V\delta\theta_s}{\delta x} = \alpha V\lambda \qquad (2)$$

where α is the surface slope, V is the forward velocity and λ is the vertical temperature gradient measured along the surface. This model assumes that all the frictional heat is supplied at the base and the basal gradient can therefore be calculated from the combination of the geothermal heat flux γ_g and the frictional heating by:

$$\gamma_b = \gamma_g + \frac{\tau_b V}{JK} \qquad (3)$$

where τ_b is the basal shear stress, K is the ice conductivity and J the mechanical equivalent of heat.

The calculations can be carried out so long as the following input data is available for each point: ice thickness (Z), ice surface temperature (θ_s), accumulation rate (A), velocity (V), surface warming rate ($\alpha V\lambda$) and the base gradient (γ_b).

The model has been tested against the temperature profiles obtained from the Camp Century core in Greenland and the 2,164 m. core at Byrd Station in the Antarctic. In both cases the fit is good and the error standard deviation for the Camp Century profile is only 0.3°C and for the Byrd profile 0.4°C (Budd, Jenssen and Radok, 1971).

So far the model has only been applied to existing ice sheets, where, with the exception of the geothermal heat flux, the necessary input data are available (even if they are not always known!). The problems of applying such a model to an ice sheet which has disappeared are formidable indeed, because it is necessary first to estimate values for the input data - an exercise requiring many assumptions. An assumption which is quite fundamental to this reconstruction is that the Laurentide ice sheet at its maximum was relatively stable for a long period of time (50,000-100,000 years). This assumption is made for two reasons. In the first place it allows the calculation of input data by analogy with the ice sheets of Greenland and East Antarctica, which are generally thought to be stable and in an approximate steady-state condition. In the second place it is a necessary prerequisite for the use of Budd _et al._'s model which was developed for an ice sheet which has a temperature profile dominated by approximate steady-state conditions in climate and ice sheet dynamics. Budd _et al._ (1971) have shown that in the East Antarctic ice sheet surface ice may take c. 100,000 years to reach the bottom layers; this carries the implication that a steady-state temperature profile may take this long to become fully established, although it may be approximately achieved in half this time. It may turn out that the Laurentide ice sheet was intrinsically unstable and that it never attained a maximum position for a sufficiently long period for steady-state temperature profiles to be established at any stage in the Pleistocene. In such a case the interpretations reached in this paper will have to be reassessed, a point taken up further in the conclusion.

Notwithstanding such problems, this paper attempts to reconstruct the input data for the Laurentide ice sheet by analogy with the existing ice sheets of Greenland and Antarctica. Where appropriate these input data are displayed in the form of maps. In other cases they are simply calculated for each of 117 points which were selected as representative of the whole Laurentide ice sheet. The paper goes on to calculate the basal thermal regime and finally includes a sensitivity analysis in order to gain some perspective on the reliability of the results.

INPUT DATA

Ice morphology

Any consideration of the morphology of the Laurentide ice sheet involves discussion firstly of the position of its margin and secondly of its surface profile.

The margin selected was the maximum extent of the ice sheet at any time. The land boundaries were plotted using the maximum drift extent and ice limits shown by Flint (1971). The seaward margin was assumed to lie on what is now the 200 m. submarine contour. There seems clear evidence that the ice extended across the continental shelf in the northwest (Pelletier, 1966), off Baffin Island (Løken and Hodgson, 1971) and off Labrador and Newfoundland (Fillon, 1975). Although in places, the ice may have been grounded further offshore than the 200 m. submarine contour, it seems a reasonable approximation (Flint, 1971). In the northeast the ice sheet is assumed to have been linked with the Greenland ice sheet, a conclusion based on the existence and orientation of submarine troughs between Ellesmere Island and Greenland (Pelletier, 1966).

The choice of a surface profile for an ice sheet the size of the Laurentide ice sheet is fraught with difficulty. The only similar-sized ice sheet is in East Antarctica and is partially centred on a mountain massif; the Laurentide ice sheet was centred over a lowland. Topographic contrasts at this scale can introduce variations in ice surface elevations. Nevertheless, there are theoretical and empirical grounds for believing that large ice sheets

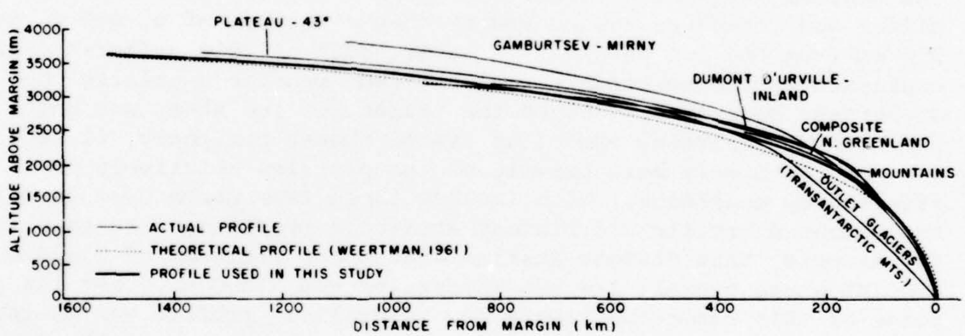


Figure 1. Surface ice profiles in East Antarctica and Northern Greenland.

have broadly similar profiles (Robin, 1964), and it is possible to draw some conclusions from existing surface profiles. Figure 1 shows a selection of profiles drawn perpendicular to the surface contours on parts of the East Antarctic and Greenland ice sheets. No profile is taken from the potentially unstable West Antarctic ice sheet (Hughes, 1975). Most Antarctic profiles are taken from the most accurate map available which covers the area between 90°E and 180° (Scott Polar Research Institute, 1974). Surface contours on the map are thought to be accurate to within 30 m. (Drewry, 1975). The Plateau Station profile is taken from the American Geographical Society map of Antarctica (1970) and is probably accurate to within 50 m. The composite north Greenland profile is taken from five profiles based on the Quaternary map of Greenland (Weidick, 1971). The outlet glacier profile is composite and based on several large outlet glaciers crossing the Trans-Antarctic Mountains.

Excluding the outlet glacier profiles, there is close agreement between the profiles representing different ice sheets in different climate regimes. For example, 100 km. from the ice margin the maximum vertical difference between the profiles is 110 m. The difference increases inland and reaches a maximum of c. 400 m. some 900 km. from the ice margin. However, much of this difference is explained by the relatively high altitude profile overlying the Gamburtsev Mountains. Since the Laurentide ice sheet was not affected by mountains except at its northeast periphery, it is reasonable to rely more heavily on the profiles relatively unaffected by mountains, which include those from north Greenland and the Dumont d'Urville and Plateau Antarctic profiles. It is notable, for example, that Plateau Station lies at an altitude of 3,624 m. at a point where overall ice thicknesses exceed 3,000 m. For the purposes of this reconstruction a new generalised profile was fitted using a summit altitude of 3,600 m. and approximating to the lower existing ice sheet profiles. It is close to the Plateau profile in the centre and the north Greenland profile near the edge (Fig. 1). The new profile has the form ($y = a + bx + cx^2$), where $a = .607$, $b = .005$, and $c = -.2 \times 10^{-5}$. For comparison the theoretical profile of Weertman (1961b) is drawn assuming the central surface ice altitude is 3,600 m.

To reconstruct the Laurentide ice sheet, the profile was extrapolated backwards at right angles to the ice margin and the ice surface contoured appropriately. Three main assumptions were made. The first was that the ice sheet formed one main dome. The second was that the ice sheet profiles were based on the 200 m. submarine contour. The latter assumption means that the full profile is only relevant along the eastern and northwestern boundaries of the ice sheet where the margin was on the 200 m. submarine contour. At higher elevation, especially on the western plains, allowance was made for the altitude of the ice margin. This means, for example, that at an ice margin altitude of 1,000 m., the lower 1,000 m. of the ice sheet profile was omitted. Thus it is implied that ice margin gradients in such areas were likely to be lower than average, a feature which is apparently confirmed by geological evidence (Mathews, 1974). Similar situations exist in many parts of Greenland today where high altitude margins have lower gradients than low

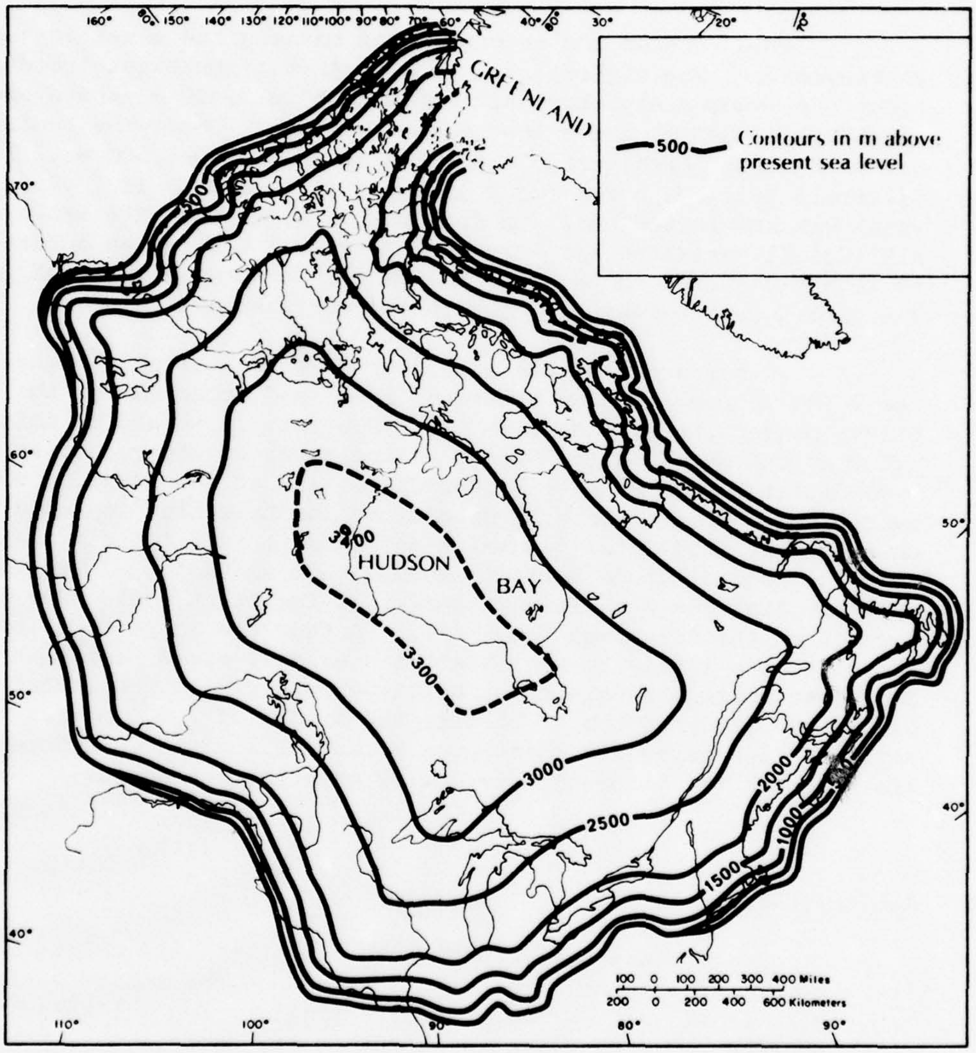


Figure 2. Morphology of the Laurentide ice sheet at its maximum.

altitude margins. At a large scale such a contrast can be seen between the high eastern margin and the low western margin of the Greenland ice sheet, as well as at the scale of individual massifs or nunataks. The third assumption was that large troughs represent the sites of former outlet glaciers, and in such situations the appropriate outlet glacier profile was used.

The form of the reconstructed maximum ice sheet is shown in Figure 2. The highest area is a northwest/southwest trending ridge over Hudson Bay where altitudes rise to 3,400 m. above present sea level (3,600 m. above the seaward margin). Over the central Canadian Arctic Archipelago altitudes attain 2,200-2,500 m. It is difficult to place error terms on these altitudes. If the ice sheet was stable and built up to form one dome, then the range of altitudinal variation illustrated in Figure 1 implies an accuracy to within at least 400 m. near the centre and probably around 200 m. The error would be smaller towards the peripheries.

A surface altitude of 3,400 m. is considerably higher than the 2,700 m. suggested by Paterson (1972) and accepted for the CLIMAP projections (CLIMAP, 1976). However, it should be pointed out that the much smaller Greenland ice sheet of today is over 3,000 m. high over a central bedrock lowland at sea level. Moreover, ice flow in a westerly direction from this high ice-shed is unimpeded by subglacial mountains and thus it is difficult to imagine that the high altitude reflects topographic control. It would seem unlikely that the much larger Laurentide ice sheet would have been lower than the Greenland ice sheet. Rather the Laurentide ice sheet is more likely to resemble the similarly sized East Antarctic ice sheet where the main crest altitude over varied subglacial topography is 3,600-4,000 m. To take the figure of 3,600 m. for the ice altitude of a large ice sheet over a central lowland is entirely consistent with the evidence provided by existing ice sheets.

Ice thickness

Ice thicknesses were calculated for the 117 points selected for the main experiment after making allowance for relief variations and glacio-isostatic movements. A generalised ice thickness map is shown in Figure 3. The following method was used:

(1) For each point an area of 2,500 km.^{2} was scrutinised and an average altitude assigned. In practice this method obliterates relief variations represented by individual mountains and valleys unless they are more than c. 30 km. across.

(2) An estimate of residual isostatic rebound was made in order to obtain the altitude of the ground surface when isostatic equilibrium is achieved following deglaciation. Opinions as to the amount remaining vary. For example, Walcott (1972a) mentioned 300 m. for central Hudson Bay while Andrews (1970) suggested 160 m. and Cathles (1975) believed recovery is virtually complete. For the purpose of this reconstruction an intermediate figure close to Andrews' estimate is used for the central region and

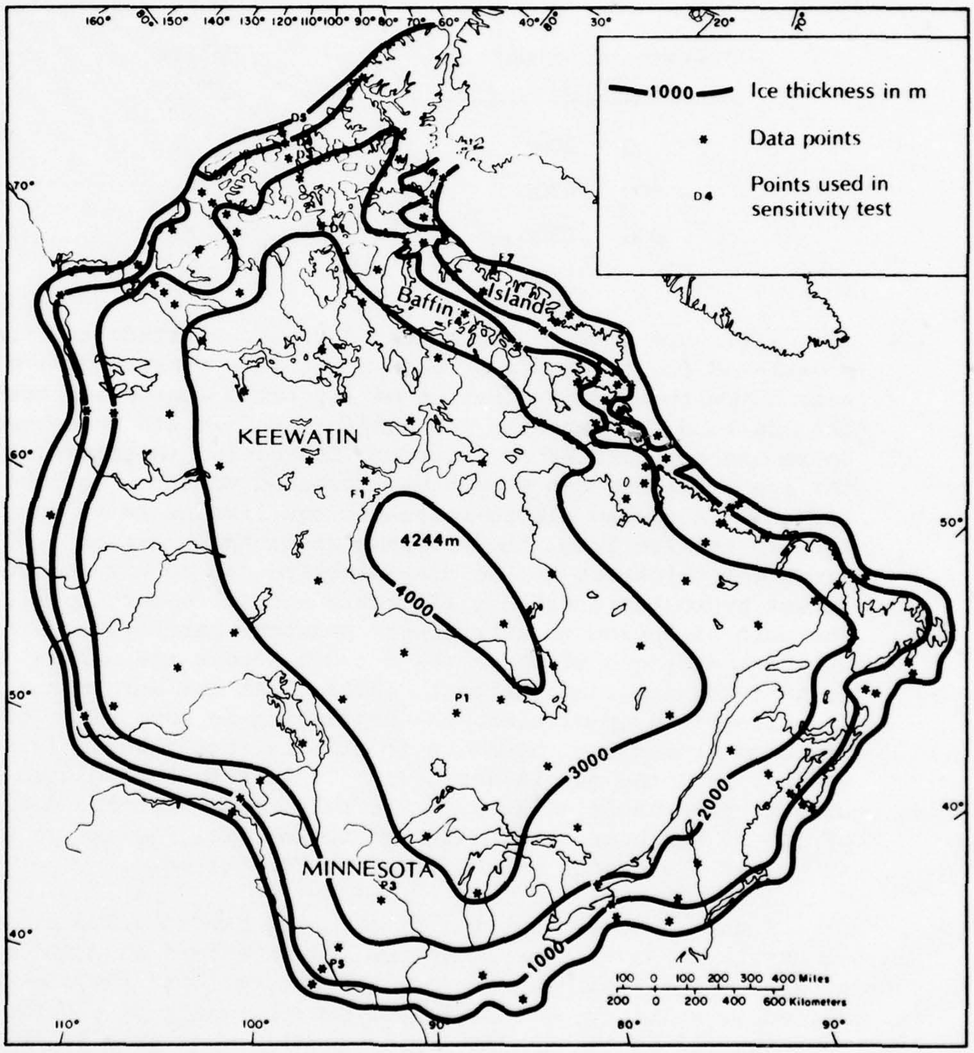


Figure 3. Generalised ice thicknesses, maximum Laurentide ice sheet. The map was compiled on the basis of calculations at the 117 data points marked. The data points used later in the sensitivity test are labelled.

the amount reduced towards the periphery where it is generally agreed to be negligible (Table 1). These values of residual rebound were added to the average altitude of each point.

TABLE 1

ALLOWANCE FOR RESIDUAL ISOSTATIC REBOUND

Distance from margin (km.)	Uplift (m.)
0 - 300	0
300 - 800	50
800 - 1300	100
1300	150

(3) The difference between the adjusted ground altitude and the postulated ice thickness at each point is a measure of ice thickness but without consideration of isostatic depression beneath the ice load. Since the ice surface profile can be assumed to remain constant after isostatic depression (Weertman, 1961b), the ice thickness can simply be increased to allow for the effect. Assuming *glacio*-isostatic equilibrium is reached beneath the ice load, then depression of the order of .267 times the thickness of the overlying ice can be assumed due to direct hydrostatic effects (Brotchie and Silvester, 1969). The main exception to this occurs near the periphery where the inherent strength of the earth's crust across the margin helps to bear the ice load (Walcott, 1970). At the margin and for a few kilometres up-glacier, the depression is more than would be expected from direct hydrostatic effects, but then as far as 400 km. from the margin depression is less than predicted by direct hydrostatic effects. For this reconstruction depression of 100 m. has been assumed at the ice margin, increasing to 420 m. at a distance of 400 km. from the ice edge.

Figure 3 shows that ice thicknesses exceed 4,000 m. in a small central zone over Hudson Bay but are in excess of 3,000 m. over a broad area. An estimate of 37×10^6 km.^{3} for the ice volume is obtained by multiplying the area of the ice sheet by .76 times the ice thickness at the centre (Paterson, 1972). This volume represents a potential sea level lowering of c. 100 m.

Ice surface temperature

Surface temperatures were obtained by estimating the full glacial temperatures ar various points around the margin of the ice sheet and then extrapolating temperature gradients inland (Figure 4).

There is doubt about full glacial temperatures around the Late Wisconsin Laurentide ice sheet, where there is pollen and other

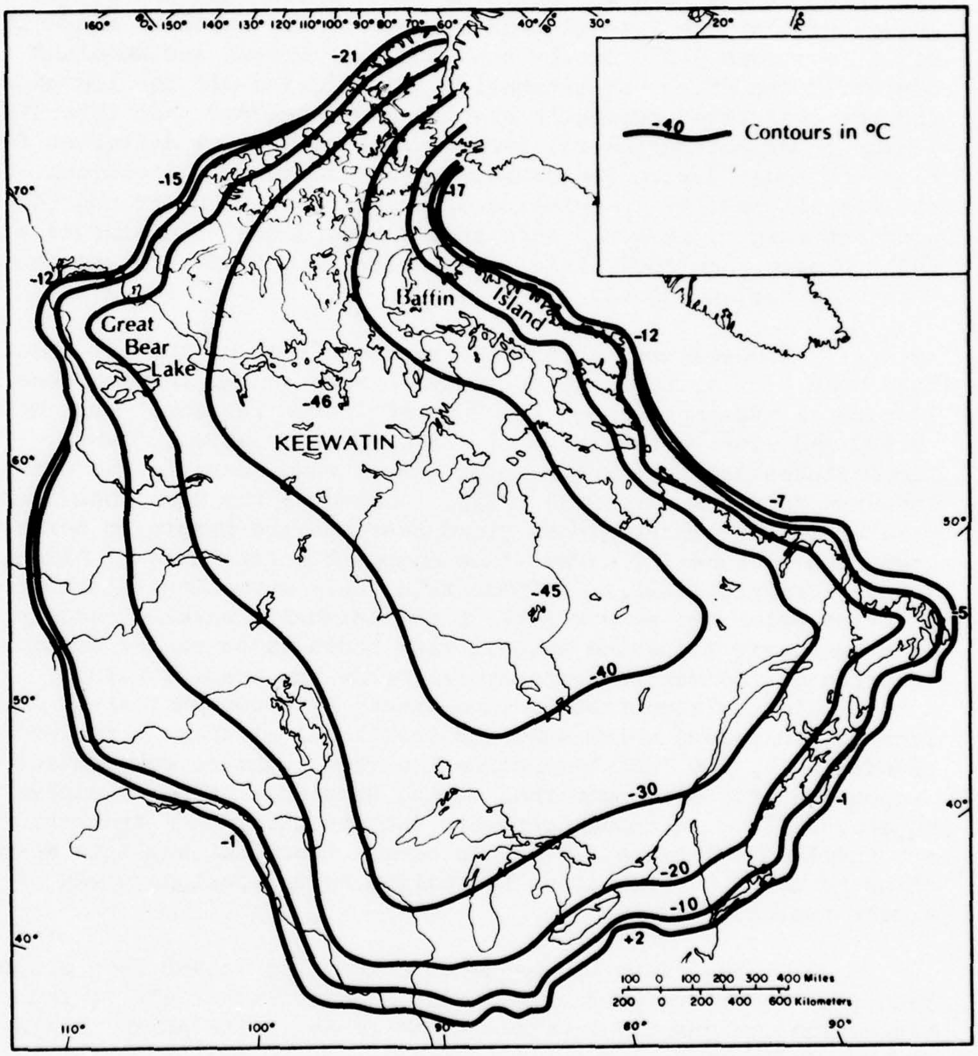


Figure 4. Suggested surface mean annual temperatures on the Laurentide ice sheet at its maximum.

evidence available, let alone temperatures during earlier glacial phases. Obviously there is no space for a full review here. Nonetheless, there are some broadly acceptable indications of the main features. Wright (1971) summarised much of the pollen evidence and concluded that there is firm evidence of boreal forest with some temperate forest trees around much of the southern margin of the ice sheet at its Late Wisconsin maximum, while Webb and Bryson (1972) suggested that the late-glacial climate of the northern mid-west U.S.A. was some 3.3°C cooler than today. Bryson and Wendland (1969) discussed the effect of katabatic winds flowing off the ice sheet and the associated adiabatic heating and concluded that there is no reason to expect peripheral temperatures to be very different from those of today during the maximum. Taking these indications and the overall paucity of evidence of tundra conditions around the southern margin, it would seem that a mean annual temperature some 4-6°C cooler than today is a reasonable estimate for the southern margin of the ice sheet.

Temperatures around the northern edge of the ice sheet may have been similar to those of today. Mean annual temperatures at 12 coastal stations around the East Antarctic ice sheet vary between -9.5°C and -18°C with a mean of -12°C (Bentley et al., 1964). These temperatures are comparable to or warmer than those in the coastal Canadian Arctic Archipelago today. Moreover the mean annual temperature at Jørgen Brønlunds Fiord near the ice margin in north Greenland is some 3°C warmer than comparable latitudes in Ellesmere Island (Fristrup, 1961). These relatively warm temperatures around existing polar ice sheets reflect the adiabatic warming associated with katabatic winds and suggest that there is no reason to expect a lowering of ice margin temperatures below the present values. Such a conclusion is supported to some extent by geomorphological evidence from Baffin Island which suggests that local glaciers which escaped inundation by the last Laurentide ice sheet were no more extensive during the Late Wisconsin than during Neoglacial times (Andrews et al., 1972). Although climatic factors other than temperature are involved (Williams, 1975), it seems clear that any Late Wisconsin climatic deterioration along the Baffin Island periphery was of slight amplitude.

On the basis of the above discussion it has been assumed that mean annual temperatures around the margins of the maximum Laurentide ice sheet were the same as today in the north (north of mid-Baffin Island and Great Bear Lake) and 5°C cooler along the southern margin. Appropriate intermediate values were interpolated for the western plains margin and the Labrador-Baffin Island margin. It is emphasised that even if the estimates are out by several degrees such errors will have only a minor effect on the ice sheet surface because of the high amplitude of the overall temperature range.

The surface temperature gradient is difficult to estimate largely because the present day analogue of the Antarctic is situated on the Pole and thus any temperature gradient includes the effects of both altitude and latitude. In the case of the Laurentide ice sheet the latitudinal component has already been included by postulating a

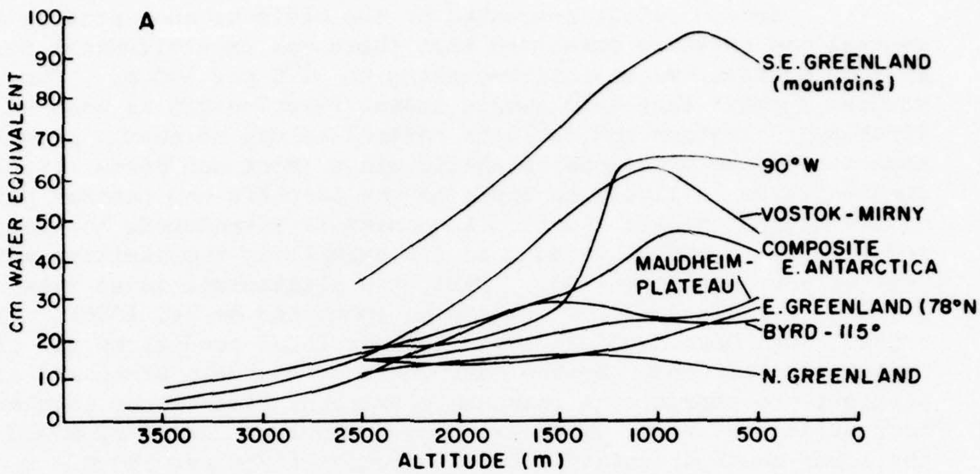
$23^{o}C$ difference between the temperatures at the southern and northern margins. Thus it is necessary to find some estimate of the altitudinal component of the temperature gradient on existing ice sheets.

Benson (1962) suggested on the basis of snow studies in central and northern Greenland that there was an altitudinal decline in snow surface temperature amounting to $-1^{o}C$ per 100 m. Subsequent studies suggest that this simple linear relationship is good as a first approximation and reflects control by dry adiabatic processes associated with downslope katabatic winds (Mock and Weeks, 1966). Such a figure is likely to apply to the northern and central parts of the Laurentide ice sheet. A problem is introduced, however, when there are upslope winds, as for example in the maritime environment of southern Greenland. Here, the altitudinal lapse rate varies between $-1.2^{o}C$ and $-2.1^{o}C$ per 100 m. (Mock and Weeks, 1966). It is argued that these high gradients reflect local conditions and are especially influenced by the narrowness of southern Greenland and the proximity to storms. A reasonable extrapolation to the southern Laurentide ice sheet, which was bigger and less maritime, would be the lower south Greenland rates of around $-1.2^{o}C$ per 100 m.

In the light of the above, a gradient of $-1^{o}C$ per 100 m. was applied to the northern half of the Laurentide ice sheet and $-1.2^{o}C$ per 100 m. to the southern half. Figure 4 shows that temperatures near the centre and north of the ice sheet may have been below $-40^{o}C$ with a coldest point of $-46^{o}C$ attained over Keewatin. A comparison with existing ice sheets suggests the results are of the right order. A central zone at $-40^{o}C$ is midway between the $-50^{o}C$ attained in central East Antarctica in higher latitudes (Bentley et al., 1964) and the $-30^{o}C$ which occurs in the centre of the smaller Greenland ice sheet at similar latitudes (Fristrup, 1966). The cooling over Keewatin is $34^{o}C$ compared with today and this figure may be compared with cooling of up to $40^{o}C$ suggested for parts of the Arctic by the NCAR atmospheric circulation model during the 'Ice Age' (Williams and Barry, 1975).

Accumulation rate

Chorlton and Lister (1970) have shown by regression analysis that there is a close correlation between ice surface altitude and accumulation rate on the Antarctic ice sheet. This is particularly true for the outermost 800-1,000 km. where altitudinal contrasts are strong. However, the relationship also applies near the centre, although in the case of Antarctica there is also an apparent latitudinal effect. Chorlton and Lister (1970) conclude that snow accumulation increases inland from the ice margin to a point about 1,600 m. in elevation. Thereafter there is a decline towards the centre. Figure 5a shows the relationship between altitude and accumulation for parts of the Antarctic and Greenland ice sheets using data provided in map form by Bull (1971) and Fristrup (1966) respectively. It is clear that there is wide variability at lower altitudes but also, with the exception of the steep mountain border of East Greenland, there is considerable con-



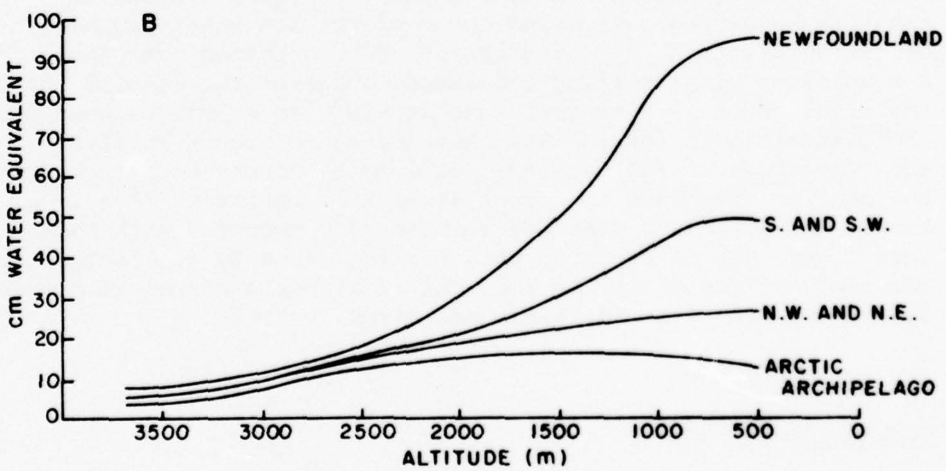


Figure 5. (a) The relationship between snow accumulation and altitude on the Greenland and Antarctic ice sheets.

(b) Four type curves used to characterise four climatic regimes.

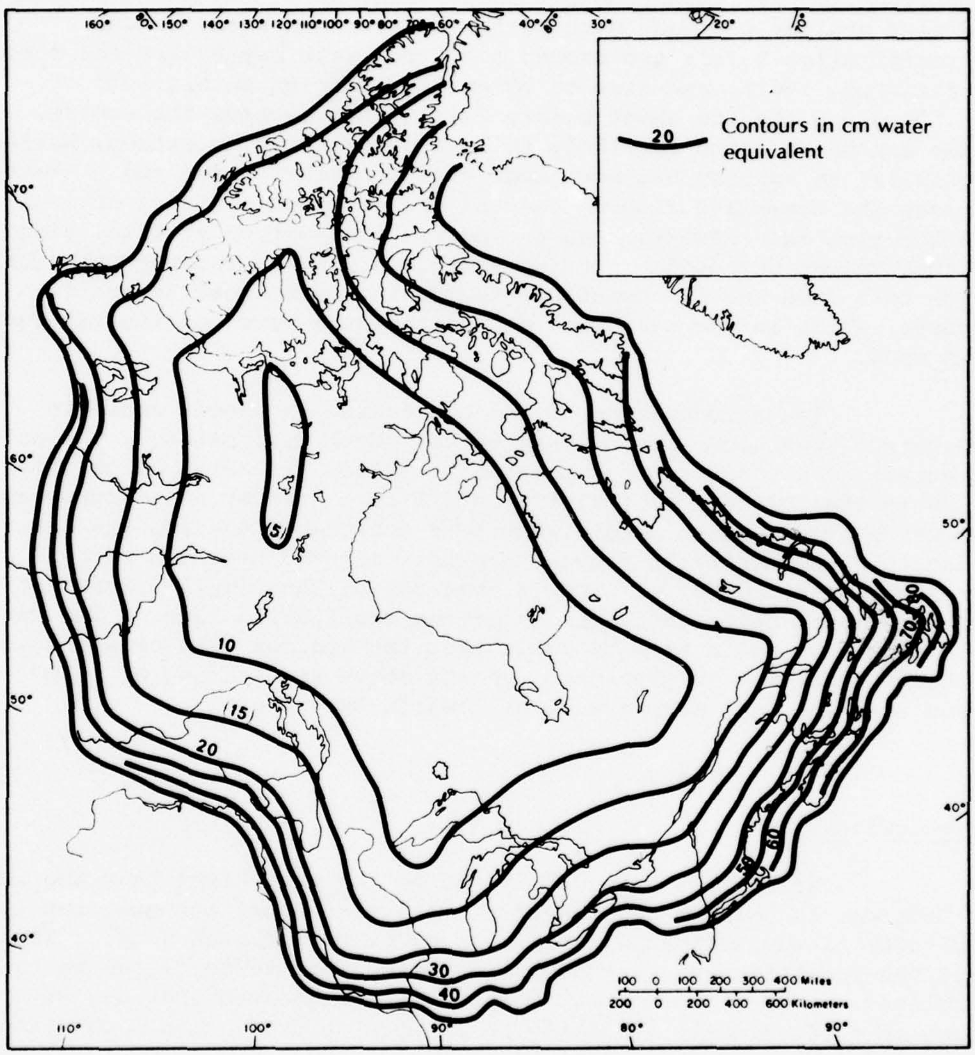


Figure 6. Accumulation rate on the Laurentide ice sheet at its maximum.

sistency above altitudes of c. 2,000 m. When one remembers that all except a narrow periphery of the Laurentide ice sheet is higher than this, then there are grounds for confidence in values obtained by analogy, at least for most of the ice sheet. The problem involves the periphery, and here four type curves were derived from the values in Figure 5a to represent four main climatic regimes relevant to the Laurentide ice sheet (Fig. 5b). In the north a direct analogy was made with northern Greenland where annual precipitation totals are around 5 cm. per year beyond the ice margin (Fristrup, 1961), and rise to 20 cm. per year at an altitude of 1,500 m. on the ice sheet before falling off towards the centre. In the southeast which is likely to have experienced an extreme maritime climate, an analogy has been made with southeast Greenland. However, since the Greenland figures reflect the orographic effect of mountains which rise more steeply than the normal ice surface profile, slightly lower values are used. In the south and west an intermediate curve has been used and is comparable to the composite East Antarctic curve, while in the northwest and northeast a more continental curve is used.

The accumulation of the Laurentide ice sheet shown in Figure 6 represents a synthesis of the above discussion. The main feature is a large central area where values are below 10 cm. with a low in the northern centre of around 5 cm. Values rise appreciably towards the southern margin, but less noticeably towards the northern margin. Accumulation figures are not included near the margin below an altitude of c. 1,000 m. because of the complications introduced in and below the vicinity of the equilibrium line. The data in Figure 5a would seem to imply that the analogy with existing ice sheets is likely to give good results above an altitude of 2,000 m. and that the main errors will be towards the margin.

Ice velocity

Velocities were calculated on the assumption that the ice sheet was in balance and therefore that the annual accumulation up-glacier of successive points had to be evacuated each year. Thus, if the ice thickness is known for a point, the velocity can be calculated. Budd, Jenssen and Radok (1971) suggested that any such estimate of velocity is likely to be more accurate than a solution based on the flow characteristics of ice.

Velocities were calculated for each of the 117 points. In order to allow for the effect of diverging and converging ice flow, the calculations were made for a 100 km. wide 'channel' positioned over each point. The area of the catchment for each 'channel' was calculated by extending boundaries up-glacier perpendicular to the surface contours. The volume of water equivalent accumulating in each catchment was calculated by multiplying the areas by appropriate accumulation values. The volume was then divided into the cross-sectional area to give the mean ice velocity through each 100 km. wide 'channel'. Calculations were not carried out for altitudes less than 1,000 m. near the margins of ice domes

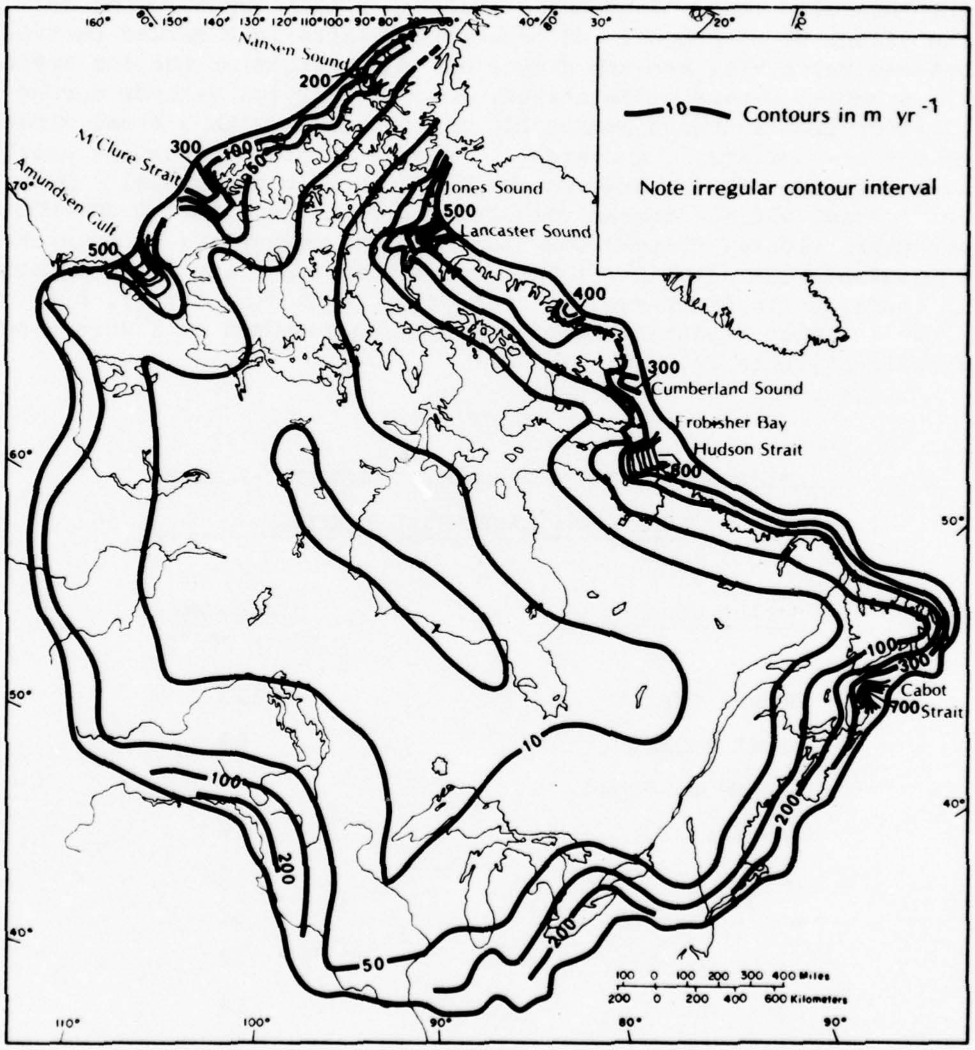


Figure 7. Balance velocities of the Laurentide ice sheet at its maximum.

where an allowance would have to be made for loss of ice volume by ablation.

Figure 7 shows the main pattern of ice velocity. There is a large central zone where velocities are less than 10 m. per year. Towards the edge of the ice sheet there is a progressive increase in velocity to values of 50-200 m. per year, reflecting the increase in accumulation as well as the thinning of the ice mass. Overall the velocities and pattern are similar to that of the Antarctic (Budd, Jenssen and Radok, 1971). By allowing for the effect of divergence and convergence there is a marked contrast between velocities beneath ridges and depressions on the ice surface. For example, a nose of relatively low velocity ice extends across Labrador from southern Hudson Bay and coincides with a broad ridge on the ice surface. Conversely, relatively high velocities exist over the site of Lake Erie which is a zone of convergence. Outlet glacier and ice stream velocities are of the order of 300-800 m. per year, figures closely comparable to the velocities of Antarctic equivalents (Swithinbank, 1964). The most important outlet glacier in terms of ice discharge was that occupying Hudson Strait, but those in Cabot Strait, Lancaster Sound and Amundsen Gulf were also important (Table 2).

TABLE 2

CALCULATED DISCHARGE OF MAIN OUTLET GLACIERS DURING THE LAURENTIDE MAXIMUM

Outlet glacier	Discharge (km.^{3} yr.$^{-1}$)
Hudson Strait	123
Cabot Strait	69
Lancaster Sound	58
Amundsen Gulf	52
Cumberland Sound	40
M'Clure Strait	35
Frobisher Bay	17
Jones Sound	12
Nansen Sound	12

In the Arctic Archipelago sector of the ice sheet, outlet glaciers account for approximately 75% of the ice discharge, in the eastern Canadian Arctic about 40%, and in the south and west virtually zero.

It is important to stress that the velocities are values averaged over 100 km. Thus in areas of high relief the average may mean little. For example, the relatively high values over northern Baffin Island are averages which take no account of any channelling effect through valleys and fiords. Such channelling is likely to

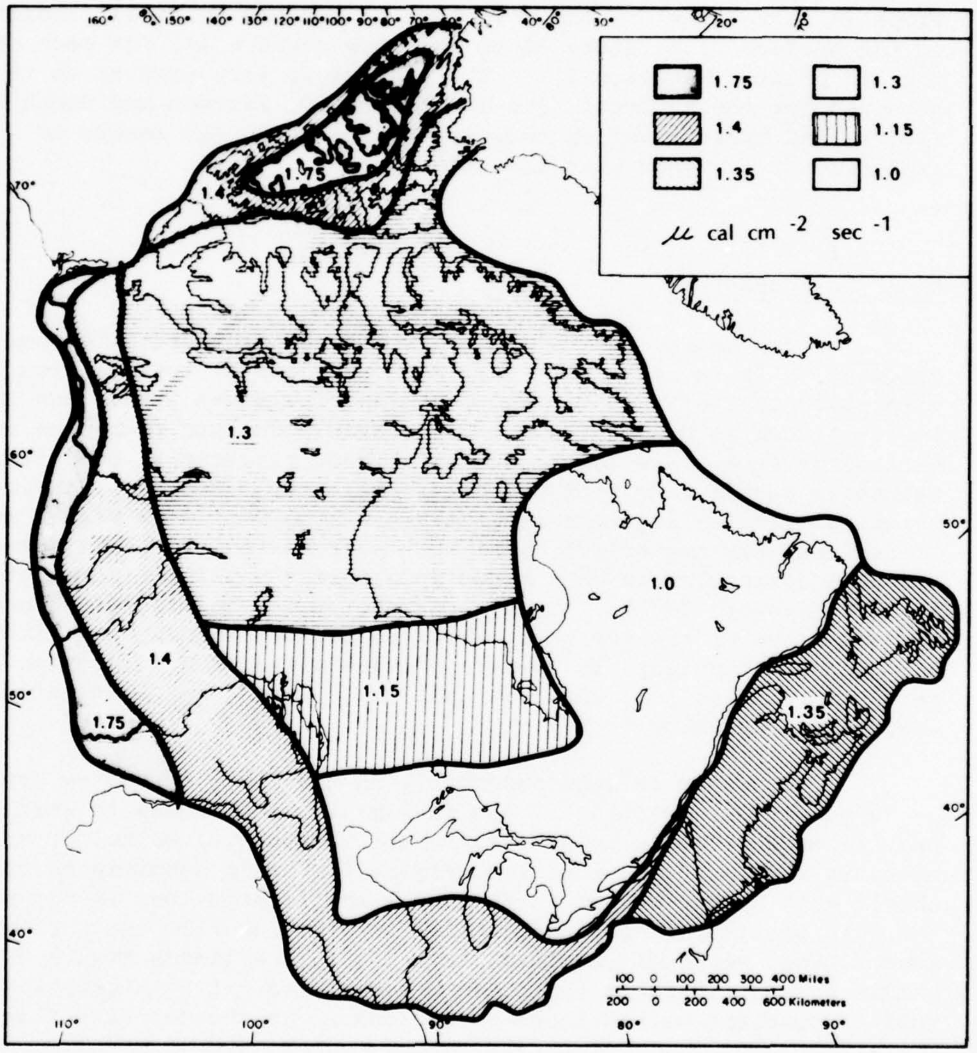


Figure 8. Geothermal heat flow in the Laurentide ice sheet area, based mainly on data collated by Judge (1973a, 1973b).

give low velocities over the plateau areas and high values in the troughs.

Surface warming

Surface warming was calculated by multiplying the surface slope (α in radians) by the velocity (V) and the vertical gradient of the surface temperature along the snow surface (λ) for each of the 117 points (Equation 2). The pattern is very similar to that obtained for the Antarctic ice sheet by Budd, Jenssen and Radok (1971), and typical values increase from zero at the centre to around .06°C per year near the edge.

Basal heat gradient

The geothermal heat flux is one contributor to the basal gradient. It is relatively well known for the Laurentide area in comparison to the Antarctic and Greenland ice sheet areas, but there is still much to discover and it is possible only to recognise rough variations from place to place. Published terrestrial heat flow values in Canada vary from 0.6 to 2.1 cal. cm.$^{-2}$ sec.$^{-1}$. An intriguing circular argument is involved: some heat flow values are 'corrected' for the effect of an ice cover during the Pleistocene and this correction is made assuming a former basal ice temperature of -1°C (Jessop, 1971). Such a correction may be misleading beneath certain zones of the ice sheet where basal temperatures are likely to have been considerably lower. Nevertheless, the correction rarely amounts to more than 0.2 cal. cm.$^{-2}$ sec.$^{-1}$ and is usually only half this value.

Figure 8 is compiled mainly on the basis of review articles by Judge (1973a, 1973b). Judge (1973b) includes a map in which 33 published records are used to attribute average values to certain tectonic regions. In addition, Figure 8 shows a subdivision of the shield with higher values in the north and lower values in the south, and this partly reflects unpublished information from the northern zone (Judge, Personal Communication, 1975). Although Figure 8 allows for considerable variation between areas it is highly likely that it conceals marked local variations in geothermal heat flow.

Frictional heat is the second component of basal heating and involves knowledge of ice velocity and shear stress (Equation 3). Following Nye (1952), shear stresses have been calculated using the equation:

$$\tau = \rho g h \sin \alpha \qquad (4)$$

where τ = basal shear stress in bars, ρ = density of ice, g = the acceleration due to gravity, h = ice thickness and α = surface slope. Figure 9 shows the general pattern of shear stress as calculated for 117 points. There is a central zone with shear stresses of less than .25 bars; from here values rise towards the periphery and in the north-

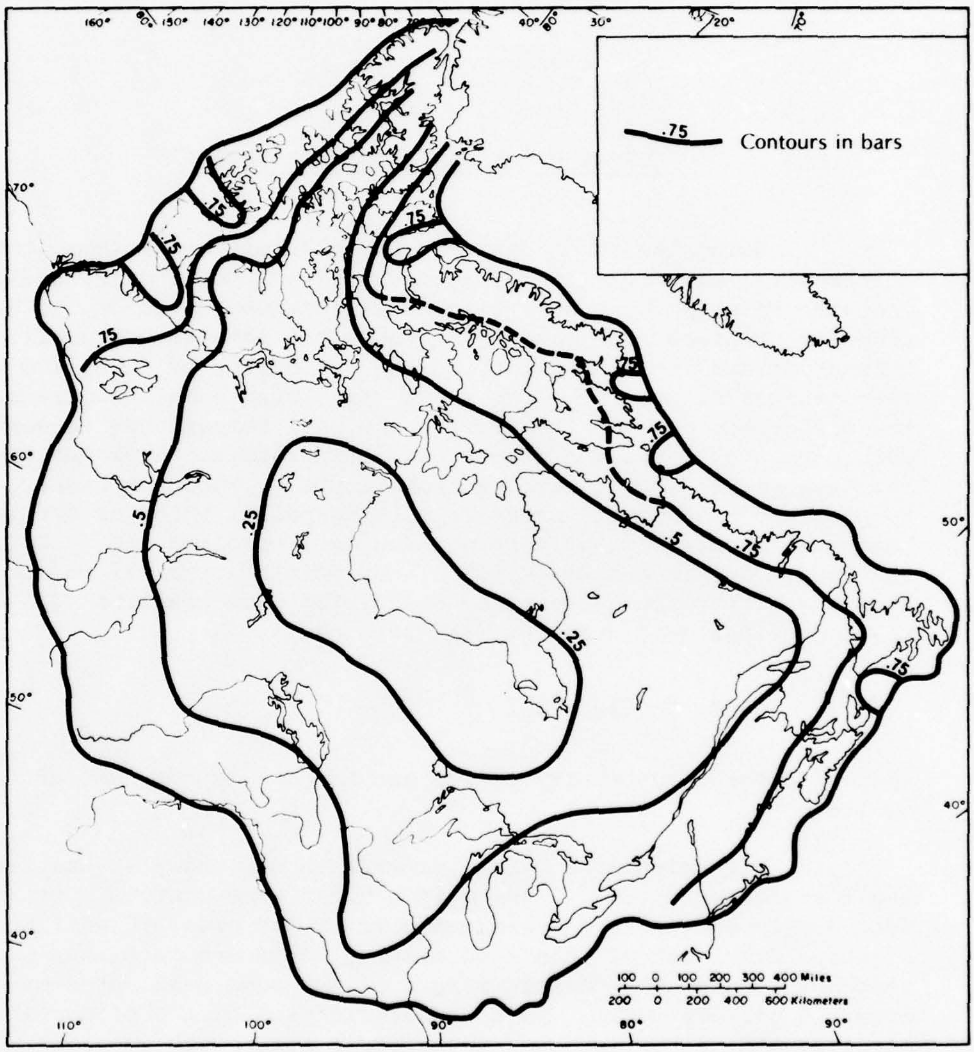


Figure 9. Calculated basal shear stresses beneath the Laurentide ice sheet at its maximum.

west, east and southeast shear stresses attain approximately 1 bar. The main exception is in the western plains area where shear stresses do not rise above .64 bars, a reflection of the relatively low surface gradients postulated for this high altitude margin. The overall pattern is very similar to that calculated for the Antarctic ice sheet (Budd, Jenssen and Radok, 1971).

OUTPUT - THE BASAL THERMAL REGIME

Using the input data for the 117 selected points, the differential equation (1) was solved for each point using a direct computer solution by numerical finite difference methods. The computer print-out for each point shows the temperature profile throughout the ice column for incremental depths of 50 or 100 m. To construct Figure 10, which shows calculated basal temperatures, the difference between the surface and base temperature in each column was taken and the actual base temperature calculated in relation to the actual surface temperature. When the basal temperature reaches the pressure melting point, then any surplus heat must be used for melting a thin layer of basal ice. Following Budd, Jenssen and Radok (1970) the melting rate (M) can be found from the difference between the calculated base gradient (γ_c) and the geothermal heat gradient (γ_g) such that:

$$M = (\gamma_c - \gamma_g) \frac{K}{L} \qquad (5)$$

where K = the conductivity of ice and L = the latent heat of fusion of ice.

The calculated basal temperature map shows several interesting features (Fig. 10). There is a broad zone centred over southwest Hudson Bay where the basal ice is at the pressure melting point. Over most of this zone melting rates are less than 1 mm. per year but towards the eastern margin of the zone melt rates may attain 4 mm. per year. This is a reflection of the higher shear stresses and velocities over the eastern part of the zone when compared to the west. Surrounding this central melting zone is a cold-based ring. In the northwest and west the calculated temperatures are usually below -5°C while in the south and east they are below -10°C and attain values as low as -21°C over central Labrador and southern Baffin Island. This cold-based zone extends northward to cover the northern Arctic Archipelago, although warm-based ice exists over the sites of deep straits. Finally, a peripheral zone at the pressure melting point exists around most of the ice sheet; exceptions are in the extreme north and the southwest. In the south the peripheral melting zone is 200-500 km. wide and it tends to become narrower towards the north. Calculation of the melting rates beneath the peripheral melting zone suggests values of 40 mm. per year in the east and around 15 mm. per year in the west.

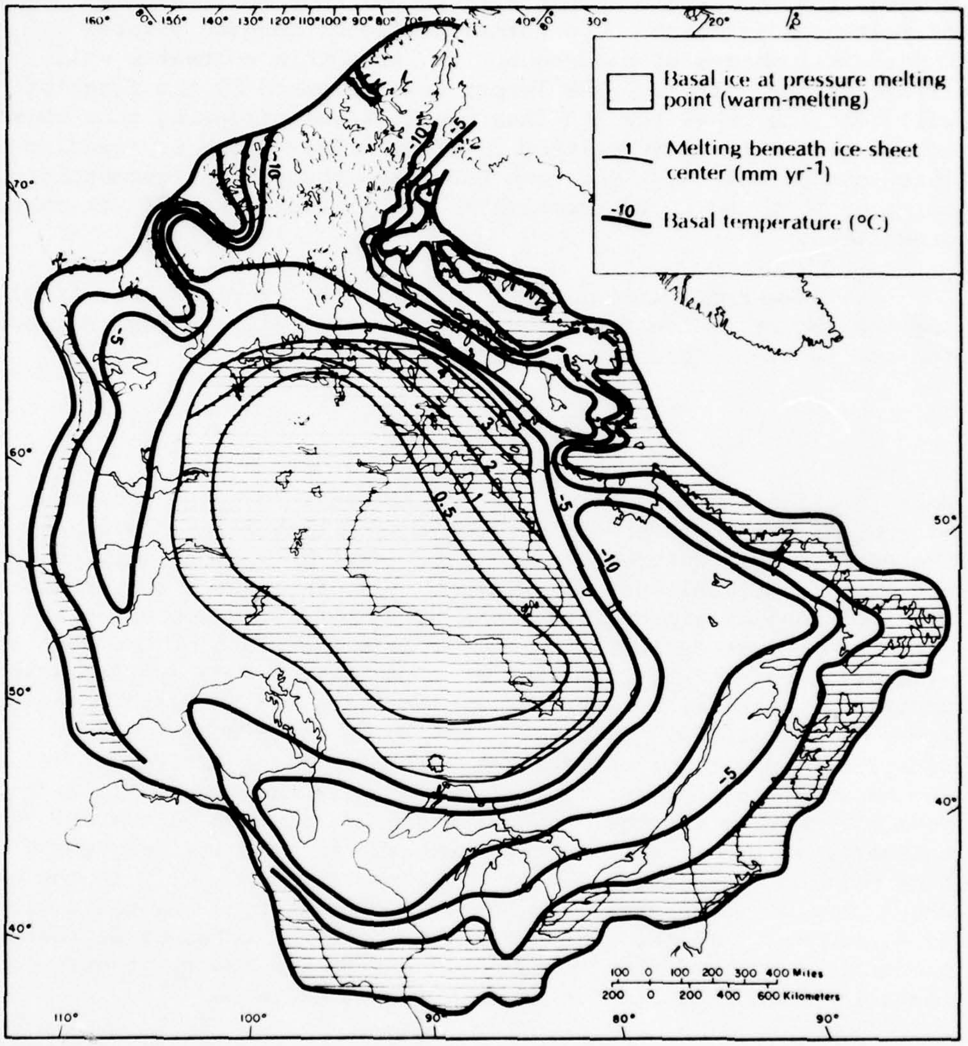


Figure 10. Calculated basal temperatures and zones of basal melting beneath the Laurentide ice sheet at its maximum.

Figure 10 fails to take into account the basal heating which might be expected to occur when the basal meltwater produced beneath the ice sheet centre freezes on to the bottom of the cold-based ring. Weertman (1972) and Shreve (1972) have shown how meltwater formed at the base of an ice sheet will tend to flow outward towards the ice margin. Since it is unlikely that the small amounts of meltwater involved could maintain streams beneath several hundred kilometres of cold-based ice, then this meltwater will freeze to the bottom. The latent heat released by the freezing will heat the basal ice and thus create a warm-freezing zone where ice is at the pressure melting point but where there is freezing on. This is a potentially important zone from the glacial geomorphological point of view and it is thus useful to gain some idea of its possible dimensions.

Weertman (1961a) has considered the problem theoretically and the amount of ice frozen on to the bottom (A_b) in regions where the ice is at the pressure melting point is given by:

$$A_b = \frac{1}{L}\left(\frac{k\Delta\tau}{h} - (Q_g + Q_s)\right) \qquad (6)$$

where L = the heat of fusion, k = coefficient of heat conductivity of ice, $\Delta\tau$ = the difference between the melting point of ice and the surface temperature, h = ice thickness, Q_g = geothermal heat and Q_s = frictional heat of sliding. If one assumes that the warm-freezing zone exists until all the meltwater is frozen on to the glacier base, it is possible to calculate the width of the zone of warm-freezing. First, the volume of meltwater made available at a particular boundary between the warm-melting and the warm-freezing zones is calculated. Then by multiplying the right-hand side of Equation 6 by various incremental distances and substituting values in the equation appropriate for that part of the ice sheet, it is possible to discover the width of the zone required before all the meltwater is consumed. In the north where there are relatively high melting rates the width of the zone is c. 130 km. In the east where melting rates are lower, several calculations suggest a width of c. 110 km. In the west and south the lower rates of meltwater production are more than compensated for by the lower intensity of the cold-based ring and the zone is c. 160 km. wide.

The positions of the various basal thermal zones is portrayed in Figure 11. The map is derived from Figure 10 but includes the additional warm-freezing zone. It is emphasised that the width of the warm-freezing zone is sensitive to the amount of meltwater involved. This in turn is sensitive to the many other variables discussed in this paper. Thus it is the broad pattern that is of interest and not the precise position of the boundaries which can be no more than a rough approximation. Nevertheless, it is important to record that the cold-based ring seems persistent everywhere except for two locations; the vicinity of Hudson Strait and Amundsen Gulf. In both cases there is a continuous zone of basal ice at the pressure melting point from centre to periphery.

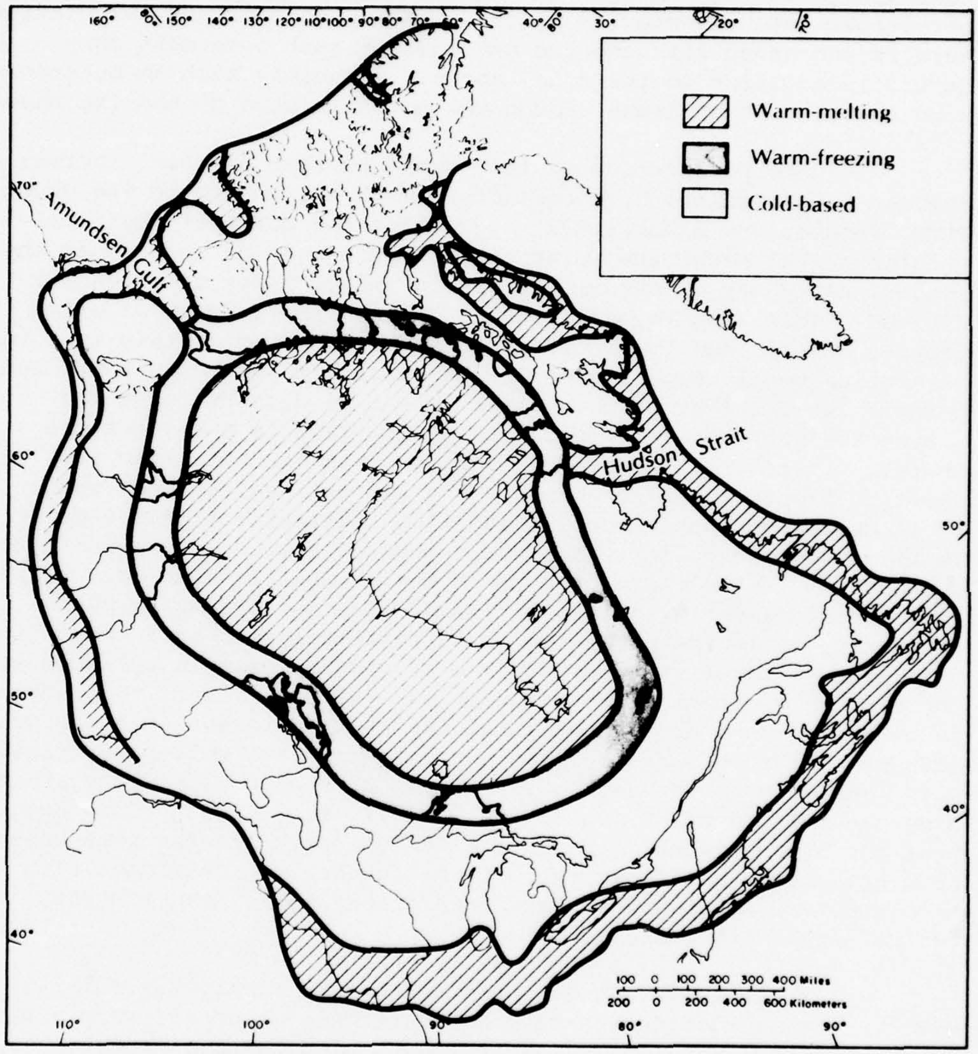


Figure 11. Zones of contrasting basal thermal regime beneath the Laurentide ice sheet at its maximum.

It is important to state that in addition to the various zones discussed above there is the possibility of an outermost cold-based zone in the ablation zone. Such a zone is discussed by Paterson (1969) and can be envisaged as an extension of the ground permafrost into the glacier margin. Since none of the calculations here is concerned with the ice margin, any such outermost zone occurs in addition to the zones above. Probably such an outermost zone was present at least around the northern edge of the ice sheet.

Some perspective on this broad pattern of basal thermal regimes can be gained by a comparison with the Antarctic ice sheet. Budd, Jenssen and Radok (1971) suggest on the basis of their calculations that basal ice is at the pressure melting point near the coastal periphery of Antarctica and in the vicinity of ice shelf basins. This zone is analagous to the peripheral zone of warm-melting beneath the Laurentide ice sheet. Inland of this zone in Antarctica basal temperatures are calculated to be below the freezing point. At first sight this might appear to disagree with the Laurentide pattern. However, it is important to note that the overall altitude of the bedrock base beneath the Antarctic ice sheet is higher than that beneath the Laurentide ice sheet when it is depressed isostatically, and also that Antarctic surface temperatures are colder; one would thus expect cooler temperatures beneath the central Antarctic ice sheet. What is much more relevant in any comparison is the similarity of the main trends. Such similarity is indicated by the way a relatively modest increase in basal temperatures in Antarctica, achieved for example by increasing the average geothermal heat flux by 0.4 cal. $cm.^{-2}$ $sec.^{-1}$, produces an extensive central zone of basal ice at the pressure melting point (except over mountains), a prominent intermediate cold-based ring with temperatures between $-10^{o}C$ and $-20^{o}C$, and an outer zone at the pressure melting point (Budd et al., 1971, Map 4/1c). This suggests that the broad trends in basal ice temperatures are the same beneath both ice sheets, and that the main difference is restricted to absolute temperatures which are predictably cooler beneath the central Antarctic ice sheet.

It is important to stress that this apparent similarity between the Laurentide and East Antarctic ice sheets is only a test of the internal consistency of the model used in this paper. Given the same model and using similar input data, one would obviously expect a similar pattern in the basal thermal regime beneath both ice sheets. However, there is independent evidence for believing that parts of the basal ice beneath the interior of the East Antarctic ice sheet are at the pressure melting point. This evidence is the occurrence of subglacial lakes which have been identified by radio echo sounding (Oswald and Robin, 1973).

SENSITIVITY TEST

The value of maps such as Figures 10 and 11 depends on the confidence that can be placed on them as models of conditions beneath the Laurentide ice sheet. This in turn depends on the sensitivity of the pattern to changing input values. For this reason a series of tests were run on three complete transects, one across the Arctic Archipelago, one from Keewatin to Baffin Island and one from near Hudson Bay through Minnesota (Fig. 3). The vital statistics of three points along each transect are given in Table 3. The points have been grouped so as to be representative of a central ice sheet zone, an intermediate zone and a peripheral zone. In addition, profile F is representative of converging flow with higher than average velocities, whereas profile P represents diverging flow with lower than average velocities.

The sensitivity of input variables was assessed by (1) altering each input variable in turn, (2) calculating anew the temperature difference between base and surface, and (3) comparing this figure with the results of the main experiment. The differences between the two sets of calculations are given in Table 4. For perspective, an overall decrease in basal temperature by 13°C (a difference of -13°) would be required to eliminate the central zone of pressure melting, though a decrease of 5°C would halve its area. An overall basal temperature rise of 21°C would be required to eliminate the intermediate cold-based ring completely. A rise of 10°C would have little effect on the extent of the ring in the east and south, but could eliminate the ring in the west. An overall temperature decrease of 24°C would be required to eliminate the outer warm-melting zone over the southern half of the ice sheet whereas a decline of 10°C would achieve this in the northwest.

Geothermal heat is perhaps the most important variable because any variations can have a profound effect on the central zone of pressure melting. A decrease of 25%, for example, reduces base temperatures in the centre by 7-12°C (Table 4), a reduction which would be almost enough to eliminate the central zone of pressure melting. An increase of 25% on the other hand would increase the area of pressure melting, the amount of basal meltwater, and thereby the width of the warm-freezing zone. In the intermediate zone and the peripheries, however, geothermal heat is relatively unimportant and any realistic variations have little effect on the overall pattern.

Basal shear stress was changed by .2 bars and found to be important only when velocities are high (over 70 m. per year). This is because it affects the amount of basal heat produced by internal deformation or sliding. In the peripheral zones a change in the value of basal shear stress produced a change of 9-55°C, with the lower values represented by P5 which is affected by diverging ice flow. This means that the existence and intensity of the peripheral warm-melting zone is very sensitive to changes in basal shear stress. However, any changes in basal shear stress are unlikely to have much effect in the central zone of melting or the intermediate cold-based ring, although the intensity of the latter could be modified.

Table 3

Vital Statistics of the Data Points Used in the Sensitivity Test Applied to the Reconstructed Basal Thermal Regime

Position on ice sheet	Centre			Intermediate			Periphery		
Site number (see Fig. 3)	D_1	F_1	P_1	D_3	F_4	P_3	D_5	F_7	P_5
Ice thickness (m.)	2939	3864	3090	1910	2320	2661	630	670	1234
Accumulation rate (m.)	.12	.06	.06	.15	.18	.16	.12	.22	.32
Velocity (m.)	2.5	5	2	14	82	11	61	464	39
Surface temperature (^oC)	-42	-41	-40	-30	-33	-28	-20	-16	-10
Surface warming rate (^oC yr.$^{-1}$)	.0001	0	0	.0013	.0030	.0003	.0142	.055	.0024
Basal shear stress (bars)	.37	.24	.25	.83	.61	.48	.66	.71	.53
Basal heat gradient (^oC m.$^{-1}$)	-.027	-.027	-.022	-.05	-.097	-.026	-.084	-.497	-.056
Temperature difference between surface and base (^oC)	+30	+54	+38	+20	+22	+18	-11	+27	+12

(+ means base is warmer than surface.)

Table 4

The Effect on Basal Temperatures of Changes in the Input Variables

Changes in Input Variables	Centre			Intermediate			Periphery		
	D_1	F_1	P_1	D_3	F_4	P_3	D_5	F_7	P_5
Reduced geothermal heat (-25%)	-7	-12	-9	-7	-6	-5	-5	-3	-2
Increased geothermal heat (+25%)	+8	+12	+8	+7	+5	+5	+1	+2	+3
Reduced basal shear stress (-.2 bars)	0	-3	-1	-6	-22	-56	-22	-56	-10
Increased basal shear stress (+.2 bars)	0	+2	0	+3	+21	+3	+22	+55	+9
Reduced surface warming rate (-20%)	0	0	0	+6	+13	+3	+9	+36	+3
Increased surface warming rate (+20%)	0	0	0	-6	-14	-3	-12	-36	0
Reduced accumulation gradient	0	-1	-1	-1	-3	-2	-1	-8	-8
Increased accumulation gradient	0	+1	+1	+1	+1	+1	+2	+4	+6
Reduced accumulation (-20%)	+1	+4	+2	+5	+1	+3	-4	-2	-1
Increased accumulation (+20%)	-2	-3	-3	-4	-7	-2	+5	+3	+2
Reduced accumulation in centre (-5 cm.)	+6	+31	+16						
Increased accumulation in centre (+5 cm.)	-4	-10	-9						
Reduced thickness (-300 m.)	-1	-2	-2	+1	+5	0	+19	+23	+1
Increased thickness (+300 m.)	+1	+2	+2	-2	-5	0	-17	-23	-2
Reduced surface temperature (-5°C)	-5	-5	-5	-5	-5	-5	-5	-5	-5
Increased surface temperature (+5°C)	+5	+5	+5	+5	+5	+5	+5	+5	+5
Calculated basal temperature (°C) Main Experiment	-12	+13	-2	-10	-11	-10	-30	+11	+2

The sites are grouped into three ice sheet zones, namely centre, intermediate and periphery. For the purposes of the sensitivity test, the calculations ignore the effect of the pressure melting point which in reality provides an upper boundary condition for basal temperatures. Because of this some basal temperatures, for example at F_7, are recorded as positive in this Table.

The surface warming rate was increased and decreased by 20%. Since the rate increases towards the periphery, this is where changes have most effect. The magnitude of the changes near the periphery, (0-36°C) shows that the existence and intensity of the outer warm-melting zone is susceptible to changes in the surface warming rate. This in turn reflects surface temperature gradients and velocities near the peripheries and would be affected by changes in either or both of these. Surface warming rate changes have a negligible effect near the centre but could affect the intensity of the intermediate cold-based ring.

Changes in the accumulation rate have a complex effect and it was decided to test their role by changing both absolute values as well as the gradient from ice sheet centre to periphery. Since the accumulation rate also affects the base gradient (through its effect on velocity and this frictional heating) and the surface warming rate, both were recalculated as appropriate. The main conclusions are that the various effects largely cancel out and that changes in accumulation rate are relatively unimportant, except in the centre. The accumulation gradient was modified by changing the peripheral and central values by 10% so as to increase or decrease the rate of decline of accumulation from periphery to centre. This had the effect of modifying temperatures only up to 3°C except near the periphery where a change of 8°C was induced. An overall increase or decrease in the accumulation rate by 20% had a similar minor effect and does not affect the overall pattern of basal zones significantly. However, a percentage variation is perhaps misleading in the centre where total accumulation is low. For this reason the accumulation in the centre was increased by 5 cm. and at P1 and F1 this almost doubles the original value. The effect of this is to reduce basal temperatures by 9-10°C, which is sufficient to severely reduce the central area of basal melting. A decrease of 5 cm. would increase the potential temperature difference between surface and base by 16-31°C and thus drastically increase the area of central melting.

Ice thickness changes have surprisingly little effect except near the peripheries. Ice thickness was varied by 300 m. while all other variables were held constant. In the centre an increase of 300 m. had the effect of increasing the base temperatures by 1-2°C and a decrease of 300 m. had the opposite effect. A surprising effect was discovered in the intermediate and peripheral zones, where an increase in thickness had the effect of cooling base temperatures and vice versa. This seems to be related to the balance and relative strength of sources of basal heating and surface cooling. In certain situations an increase in thickness allows the surface cold inversion (where temperatures fall with depth) to grow at the expense of the deeper zone where temperatures rise with depth. The magnitude of the changes induced by varying ice thickness suggest that it is a relatively unimportant variable except in the peripheral zone. However, since ice thickness is thought to be most accurate in this zone, it is reasonable to conclude that ice thickness changes are unlikely to modify significantly the overall pattern of basal thermal regimes.

Surface temperature changes do not affect the magnitude of the difference between ice temperatures at the surface and base. However, they do change one boundary condition and an overall rise or fall in surface temperature would have an equal effect on base temperatures. Whereas overall warming by $5^{o}C$ would reduce the intermediate cold zone in the west, warming of over $10^{o}C$ would be necessary to significantly change the pattern in the east.

The discussion of the role of individual variables allows some broad conclusions to be drawn about the stability of the pattern portrayed in Figures 10 and 11. The existence of the central zone of melting is threatened only by large changes in two variables. A decrease in geothermal heat by 25% or a doubling of the accumulation rate would almost suffice to remove the zone. A change in these variables in the opposite direction (which is equally likely) would increase the size of the zone and the adjacent zone of warm-freezing. The existence of the intermediate cold-based ring is not threatened by changes in individual variables except in the west. (It should be stressed that F4 is an extreme example of sites in this zone with atypically high velocities and variability and that most points in the zone resemble D3 and P3.) Within this intermediate zone basal temperatures are most sensitive to changes in basal shear stress, the surface warming rate and, to a lesser extent, geothermal heat. Basal temperatures in the peripheral zone of warm-melting are highly sensitive to changes in basal shear stress, ice thickness and the surface warming rate. Changes in any of these could remove the warm-melting zone completely, though of course it is equally likely that the variables could change in the opposite direction and increase the width of the zone.

Overall it seems reasonable to suggest that the pattern of a central warm-melting zone and surrounding zones of warm-freezing and cold-based ice is stable. The outer zone of warm-melting is less stable. The zones can be ranked in order of decreasing stability as follows: (1) cold-based ring (at least in the east and south), (2) central melting zone and thus the adjacent warm-freezing zone, and (3) peripheral melting zone. The cold-based extension over the Arctic Archipelago is particularly stable, both because of its intensity and because its extent is not influenced by changes in melting rates beneath the ice sheet centre. To claim that the pattern is relatively stable is not to suggest that the positions of the boundaries between the zones are stable. Quite clearly there are many variables which can shift boundaries by hundreds of kilometres. For this reason Figure 11 is intended to be no more than a first approximation of basal thermal zones beneath the maximum Laurentide ice sheet.

CONCLUSION AND IMPLICATIONS

As hinted at earlier, the type and distribution of thermal zones beneath an ice sheet has profound influence on the type of geomorphological process operating at the base of the ice. Some of the implications have been conceptualised in model form by Sugden and John (1976). Figure 12 is a development of these models and illustrates the variation in erosive processes that can be expected to occur beneath different parts of the Laurentide ice sheet. In the north there is either continuous cold-based ice or a transition from cold-based ice to warm-melting. Containing no debris, such cold-based ice is unlikely to be able to erode (Fig. 12a). Further south the transition is warm-melting, warm-freezing, cold-based, warm-melting (Fig. 12b). In the vicinity of Hudson strait the sequence is warm-melting, warm-freezing, warm-melting (Fig. 12c). Finally, Figure 12d shows a situation where there is in addition an outermost zone of cold-based ice in the ablation area. Here the sequence may be warm-melting, warm-freezing, cold-based, warm-melting, warm-freezing, cold-based. In all the latter examples there are different relationships between erosion per se and entrainment. Probably when cold-based ice carries debris there can be some erosion. However simple such models appear to be, they do offer a framework from which to investigate the landform evidence on the ground. This is a theme which will be taken up in relation to glacial erosion in northern North America in a separate paper. In addition, one can imagine similar models being used for the study of depositional processes. Most deposition is likely to occur in the warm-melting zone where debris in the ice is brought into contact with the glacier bed by basal melting. The peripheral warm-melting zone might be expected to experience more deposition than the central warm-melting zone because (1) basal melting rates are higher and (2) there is a greater amount of debris available in the lower layers of the ice sheet, inherited from the upstream warm-freezing zone.

A further implication concerns the role of basal thermal regime in influencing ice sheet stability and wastage. It can be suggested, for example, that the existence of a cold-based ring makes the Laurentide ice sheet intrinsically stable. However, one can also point to the existence of two sites where this cold-based ring may be breached by basal ice at the pressure melting point, namely inland of Hudson Strait and Amundsen Gulf. As suggested on theoretical grounds by Hughes (1973b), these two areas may be points of potential instability, something already hinted at by geological evidence of almost catastrophic rates of retreat in the Hudson Bay/Hudson Strait area during the last phase of deglaciation (Andrews and Falconer, 1969; Andrews and Peltier, 1976).

It is important to conclude by stressing once more the assumption that is central to this reconstruction of the Laurentide ice sheet. It has been assumed that at least once, and possibly several times, the Laurentide ice sheet attained a maximum position for a sufficiently long time for steady-state temperature profiles to be approximately established. As mentioned earlier this period is probably of the order of 50,000-100,000 years. If all the

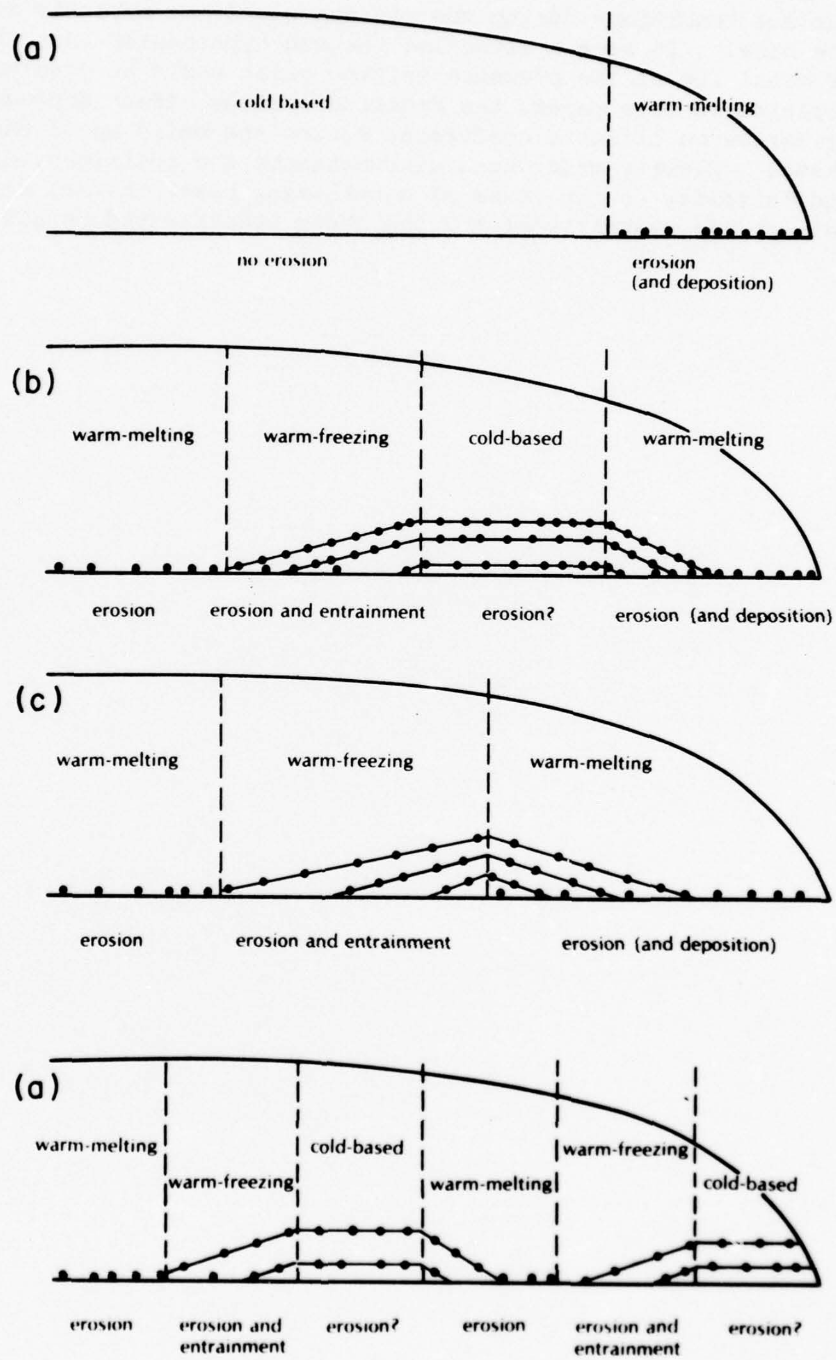


Figure 12. Various sequences of basal thermal regime and associated erosional processes encountered beneath the Laurentide ice sheet at its maximum.

Laurentide ice sheet maxima prove to be significantly shorter than this, then one can expect some modification to the pattern displayed in Figures 10 and 11. The main result of a short-lived period in a maximum steady-state condition would be that the effect of cold surface conditions during the maximum would not be fully felt at the base. In such a situation one can hypothesise that the area of basal ice at the pressure melting point would be greater than depicted in this paper, the magnitude of the effect depending primarily on climatic conditions during the build up of the ice sheet. Clearly under such circumstances the positions, dimensions and intensity of the zones of contrasting basal thermal regime as well as the stability of the ice sheet itself would be affected.

THEME 3

GLACIAL EROSION BY THE LAURENTIDE ICE SHEET

ABSTRACT

The aim of the paper is to map and analyse landscapes of glacial erosion associated with the Laurentide ice sheet and to relate them to the main variables affecting glacial erosion, namely, the former basal thermal regime of the ice sheet, the topography of the bed and the geology of the bed.

Areal scouring is associated with zones of basal melting beneath much of the former ice sheet centre and in places where the topography favoured converging ice flow. The landscape type may also form beneath cold-based ice when it is carrying debris inherited from an upstream zone of regelation. Areas with little or no sign of glacial erosion occur primarily in the north in the Queen Elizabeth Islands, but also further south on uplands associated with diverging ice flow. They coincide with areas calculated to have been covered by cold-based ice devoid of debris. Landscapes of selective linear erosion are common on uplands near the eastern periphery of the ice sheet. In these situations preexisting valleys channelled ice flow and led to a contrast between warm-based ice over the valleys and cold-based protective ice over the intervening plateaux. Variations in the permeability of the bedrock base can modify the landscape pattern on a local scale, mainly in those areas where there was a change from one basal thermal regime to another. In general permeable rocks tend to have experienced less erosion than impermeable rocks. Using lake density as an indication of the intensity of glacial erosion, a zone of maximum erosion is identified and forms a ring between the centre of the former ice sheet and its periphery. This ring coincides with a zone where meltwater from the ice sheet centre froze on to the bottom of the ice sheet. This regelation incorporated basal debris into the ice forming a basal layer 20-50 m. thick and afforded an efficient means of debris evacuation.

A conceptual model is developed and hangs round the following postulates: (1) landscapes of glacial erosion are related primarily to the basal thermal regime of the ice sheet; (2) intense areal erosion is favoured above all by the availability of efficient mechanisms of debris evacuation rather than mechanisms of abrasion or fracture; (3) landscapes of glacial erosion are equilibrium forms related above all to maximum glacial conditions. This latter conclusion implies that at some stage in the Pleistocene the Laurentide ice sheet was in a stable maximum condition for a long period of time.

Figure 13. Vertical air photograph of a landscape of areal scouring near Nettilling Lake, Baffin Island.

INTRODUCTION

The most abundant evidence of glacial erosion by ice sheets exists in those areas formerly covered by the Laurentide and Scandinavian ice sheets. Although modified by post-glacial weathering, the beds of these former ice sheets are accessible and offer a potentially rewarding field of study whereby the glacial landforms on the ground may be examined and related to glaciological processes occurring beneath the former ice sheets. At present too little is known about the landform patterns to be able to relate them to glaciological process and as a result there is something of a gulf between the progress of theoretical studies of processes occurring at the bottom of ice sheets and the field evidence which is required to constrain and further develop such theory. It is the purpose of this paper to map and analyse landscapes of glacial erosion in Canada and to offer some perspectives on the relative importance of the main variables believed to affect glacial erosion, namely (1) basal ice characteristics, (2) topography of the glacier bed, and (3) geology of the bed.

THE LANDSCAPE EVIDENCE

Classification

The mapping of the landscapes of glacial erosion employs a simple morphological classification used by the writer in Greenland and elsehwere (Sugden, 1973; Sugden and John, 1976). The main characteristics in so far as they apply to Arctic Canada are described below.

A landscape of areal scouring comprises an irregular rock surface which has everywhere been shaped by the action of ice (Fig. 13). This landscape type has been described by Linton (1963) as knock-and-lochan topography. A good example of the landscape type occurs around Frobisher Bay settlement (Fig. 14). Here the most obvious feature is the widespread occurrence of glacially abraded rock surfaces, occasionally with striations and crescentic gouges still preserved. At a scale of hundreds or thousands of metres these rock surfaces form part of smooth and elongated hills streamlined in the direction of ice flow. In between the hills are structurally controlled depressions which often contain irregular lakes. Typical measurements reveal a 1,000 m. long hill to be commonly 60 m. high, and 200 m. long hill to be 10-15 m. high. At this scale the profile of the hills is not broken up by irregularities induced by plucked or other lee-side effects. At a scale of 10-100 m. the streamlined hills can be seen to comprise a series of individual rock knobs and depressions. Typically the knobs are roches moutonnées with ice-moulded upstream flanks and plucked and craggy lee sides. Boulders bounded by clear joint

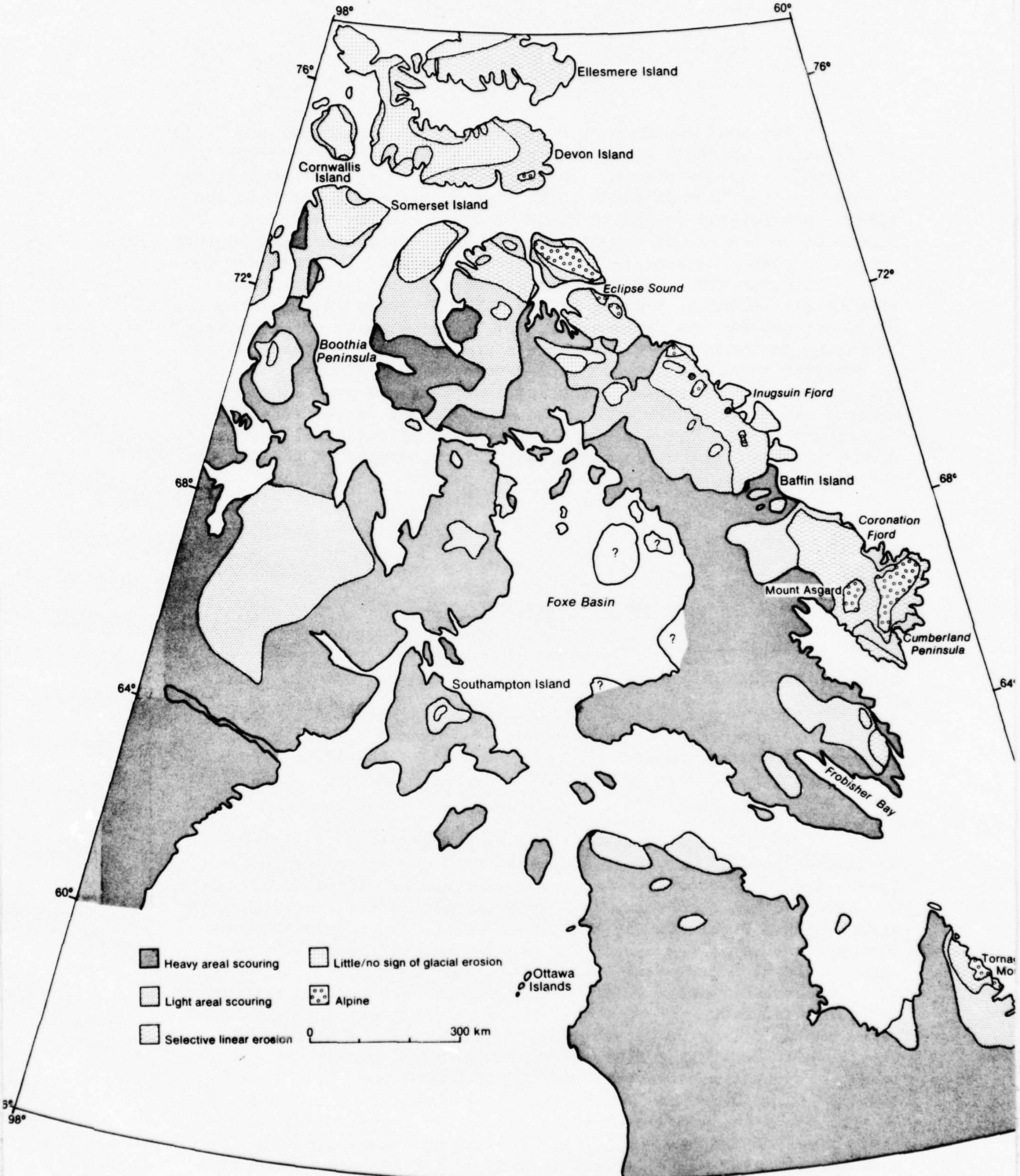


Figure 14. Distribution of landscapes of glacial erosion in the eastern Canadian Arctic, compiled from LANDSAT-1 imagery and conventional air photographs.

Figure 15. A landscape of selective linear erosion adjacent to Coronation Fiord, Baffin Island. Tors and felsenmeer occur on the upland plateau.

faces are common and include some which are still in place as well as those which have been moved horizontally some tens of metres from their original starting place.

A landscape of selective linear erosion consists of glacially excavated troughs separated by upland plateaux with little or no sign of glacial erosion. An example is provided by the Coronation Fiord area in the Cumberland Peninsula area of Baffin Island (Figs. 14 and 15). The fiords are glacial troughs with steep cliffs rising precipitously above the fiord water to altitudes in excess of 950 m. The upland areas on either side of the trough consist of a gently sloping area marked by broad and open river valleys. The surface supports regolith derived in the main by weathering from the underlying bedrock but also from till containing distinctive glacial erratics. For example, at an altitude of 910 m. on the plateau just north of Coronation Fiord glacially faceted boulders of grey fine-grained gneiss and red coarse-grained pigmatite occur on bedrock consisting of coarse-grained garnetiferous biotite gneiss. Bedrock tors are characteristic of the upper surfaces and may be surrounded by erratics. One of the most striking features of this landscape type is the clarity of the break between the cliff top of the glacial trough which truncates the gentle upper slopes. Other examples of this landscape type with seemingly identical characteristics have been described in the Torngat Mountains of Labrador and the Inugsuin Fiord area of eastern Baffin Island by Ives (1957, 1975).

A landscape of little or no sign of glacial erosion consists of an extensive area where the dominant landforms are river valleys and where long gentle slopes are swathed in regolith derived in the main from the underlying bedrock. Often tors occur on interfluves. Such landscapes are common in the northern Canadian Arctic and are easily distinguished on aerial photographs. The landscape type is characteristic of much of Somerset Island and is described in detail by Dyke (1976) (Fig. 16). Although there is no obvious sign of glacial erosion, a former ice cover is frequently indicated by the occurrence of glacial erratics, for example on Somerset Island (Dyke, 1976) and elsewhere in the Arctic Archipelago (Blake, 1964, 1975).

Alpine landscapes comprise massifs shaped essentially by the action of local valley glaciers constrained in their flow by the topography of each individual massif. A coarsely dendritic trough pattern, horns and arêtes are characteristic. A spectacular example of the landscape type, much in demand for the making of Canadian whisky advertisements and Secret Service movies, occurs in the Mount Asgard area of Cumberland Peninsula, Baffin Island.

Distribution of the landscape types

The mapping of erosional landscapes relies heavily on the interpretation of LANDSAT-1 images, backed up by the scrutiny of several thousand oblique and vertical air photographs as well as fieldwork in sample locations on Baffin Island. In practice it was

Figure 16. Photograph from approximately 50 m. above the ground of a landscape of little/no sign of glacial erosion on Somerset Island. Erratics occur around the tors. (Photograph by A.S. Dyke, GSC-203014-D)

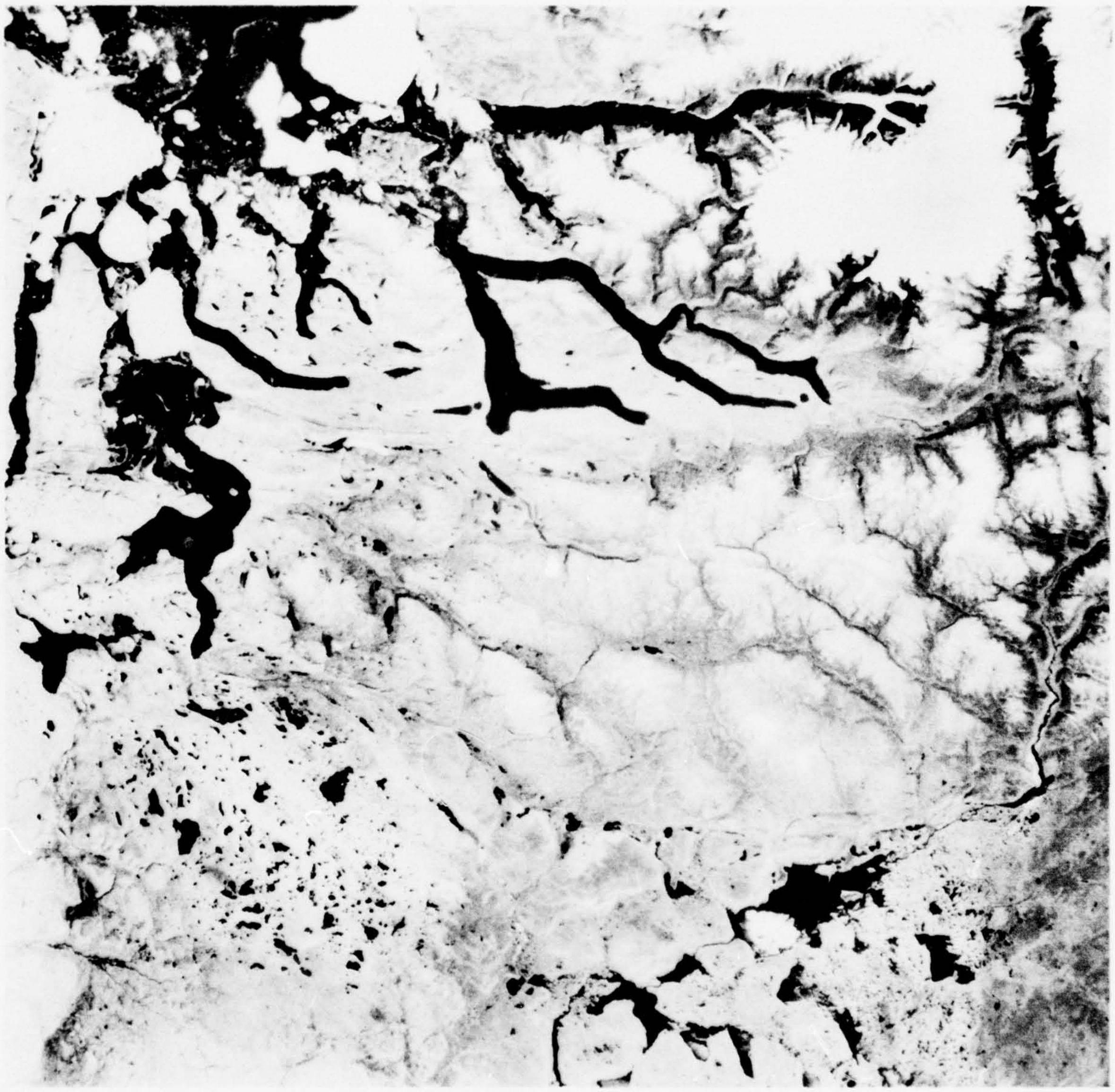


Figure 17. The appearance of different landscape types on LANDSAT-1 imagery. The image is of an area south of Eclipse Sound in Baffin Island and shows areal scouring in the south, selective linear landscapes in the northeast and little/no sign of glacial erosion in the intermediate area in the east. North is at the top of the image. (E-1746-16434-6)

found that LANDSAT-1 images (especially bands 5 and 6) were easily interpreted in terms of the classification given above. Landscapes of areal scouring can be picked out by the irregular lake pattern, presence of clear structural lineations revealing the outcropping of bedrock at the surface and the absence of dendritic river patterns (Fig. 17). Areas of little or no sign of erosion are characterised by regular dendritic river patterns, the absence of structurally controlled lakes and a texture indicating the presence of regolith. An intermediate category (light areal scouring) was introduced to cover those areas where the river pattern is clearly preserved but where there are also a few lakes and some structural lineations. Landscapes of selective linear erosion are distinguished by troughs with clearly marked edges and intervening plateau areas covered in regolith and devoid of lakes. In those cases where the intervening upland is ice scoured, then the whole landscape type is classified as one of areal scouring. Alpine landscapes are marked by a complex coarsely-dendritic trough pattern and the presence of pointed peaks.

Figure 18 depicts the broad distribution of main landscape types for the whole of the area covered by the Laurentide ice sheet at its maximum. In addition to the erosional categories discussed above a depositional zone is included and describes those areas where bedrock is largely obscured by drift. The map, covering as it does such a large area, is obviously highly tentative and is intended to be no more than a guide to the broad pattern of landscape variation. The main feature is a broad area of areal scouring extending over the central area of the former ice sheet. Along the southeastern maritime margin the zone extends at least as far as the coast, while along the southern and western land margins the zone gives way to a peripheral zone of deposition. In the north the areal scouring gives way to landscapes with little or no sign of glacial erosion. This transition takes place mainly beneath the waters of Viscount Melville Sound but in several places the transition is abrupt and takes place on land. For example, on Somerset Island Netterville et al. (1976) have described the change between the characteristic ice-scoured depressions and rock knobs south of the boundary and the fluvial topography, regolith and tors of the area immediately to the north. Landscapes with little or no sign of glacial erosion include most lowlying islands in the Queen Elizabeth Islands. Although there is some discussion about the status of these islands during the Late Wisconsin glaciation (Blake, 1975; England, 1976), there is general agreement that they were covered by Laurentide ice at some stage during the Pleistocene, a conclusion apparently confirmed by the existence of Laurentide shield erratics on the Paleozoic rocks of the islands. The topography beneath the straits is obviously not well known. However, there are indications that it includes both areal scouring and areas with little or no sign of erosion. For example, Bornhold et al. (1976) describe well preserved river patterns as well as ice-scoured topography from the floor of Barrow Strait immediately north of Somerset Island. Landscapes of selective linear erosion are, with one exception on Melville Island, restricted to the eastern upland periphery of the former Laurentide ice sheet. The landscape type is most common in the north and most restricted in the south-east. Probably the extent of selective linear erosion is under-

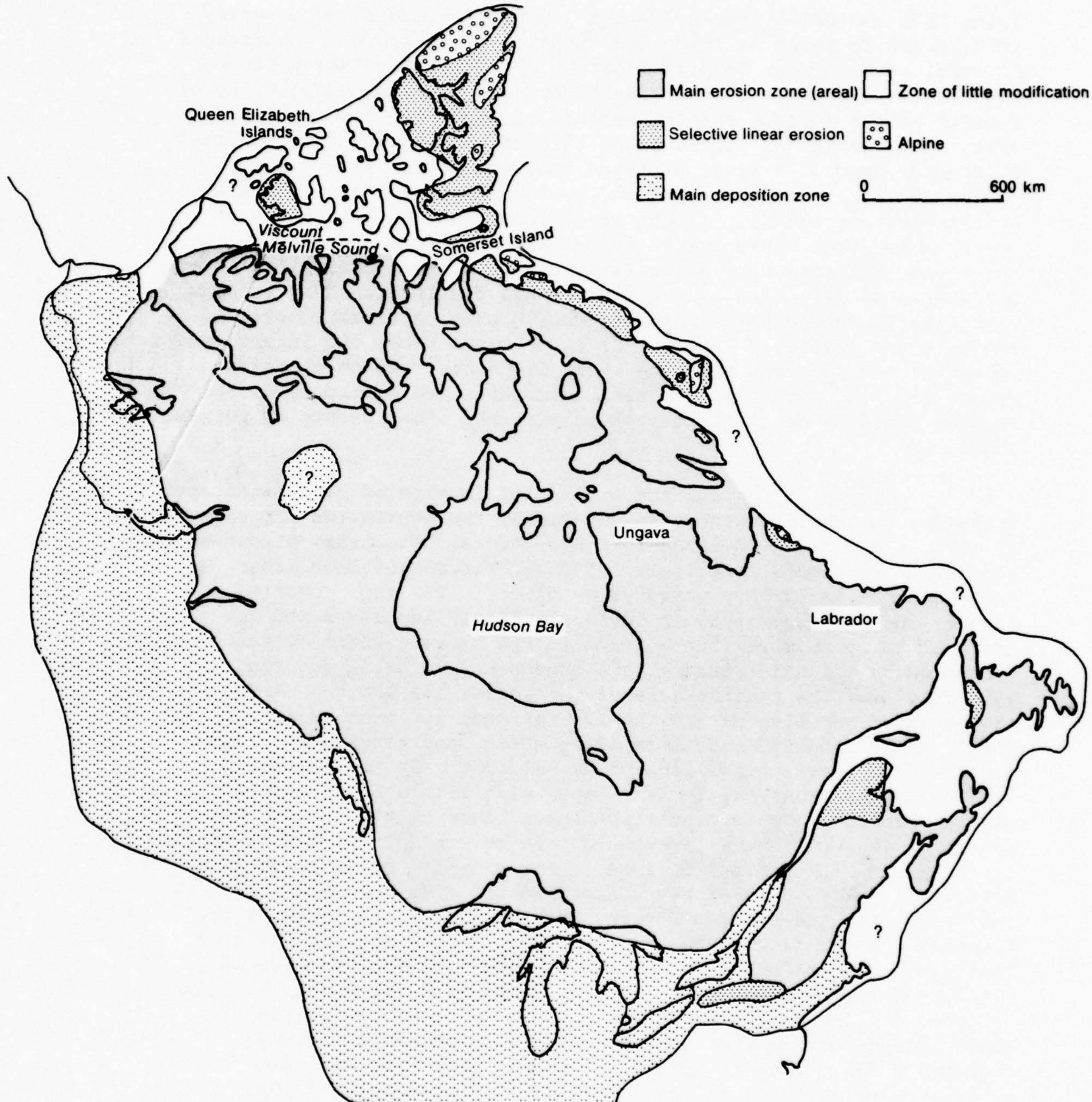


Figure 18. Main landscape zones, Laurentide ice sheet, compiled from examination of LANDSAT-1 imagery.

estimated in the northwest. Here glacial troughs appear to occur in some straits (Pelletier, 1966) and these could justify classification of more of the Queen Elizabeth Islands as selective linear landscapes, albeit partly submerged. Alpine scenery within the confines of the former Laurentide ice sheet occurs patchily along the northeastern periphery and in one part of Labrador. In all cases Alpine areas are adjacent to or surrounded by selective linear landscapes.

Figure 14 shows the erosional landscapes of the eastern Canadian Arctic in more detail, and relies mainly on the examination of several thousand oblique air photographs. Clearly in any such map there will be areas of doubt, but the map is offered as a representation of the more detailed pattern and provides a basis for discussion. Landscapes of areal scouring cover most of the area other than the northern and eastern periphery. Both flanks of Hudson Bay and Foxe Basin, islands rising out of these basins and southern Baffin Island are all heavily ice scoured. One clear pattern is that uplands are often less intensely ice-scoured than surrounding adjacent lowlands. Many good examples of this relationship can be seen along the axes of peninsulas in southern Baffin Island, Boothia Peninsula and on Southampton Island. Landscapes with little or no sign of erosion form isolated pockets over the northern part of the area. It is clear that one category of this latter landscape type is associated with uplands. In the north the bulk of the central high parts of Devon and Cornwallis Islands fall into this category. Further south the occurrences are less extensive but the same relationship is apparent. In several places on Baffin Island, Southampton Island and Boothia Peninsula, areas of little or no sign of erosion include the culminating points of uplands which are increasingly modified by areal scouring down their flanks. Another type of landscape with little or no sign of glacial erosion includes the lowlying eastern forelands of central and northern Baffin Island. Landscapes of selective linear erosion are extensive in the north in eastern Devon Island and Ellesmere Island and in northern and central Baffin Island, but are restricted to small upland areas in southern Baffin Island and the Torngat Mountains of Labrador.

Intensity of glacial erosion

Although Figures 14 and 18 are helpful in pinpointing the location of different types of glacial erosion, they are not able to provide quantitative information about the intensity of erosion from place to place. The information is potentially important and several attempts were made to derive it. Simple tests using such indicators as degree of stream derangement, streamlining as reflected by lakes, and preferred orientation of depressions proved either unwieldy or unhelpful. However, density of lake basins proved more helpful. Lake basins have long been regarded as diagnostic of glacial erosion and indeed Hobbs (1945) used their distribution as a means of determining the former extent of the Laurentide ice sheet. For the purpose of this article lake basin

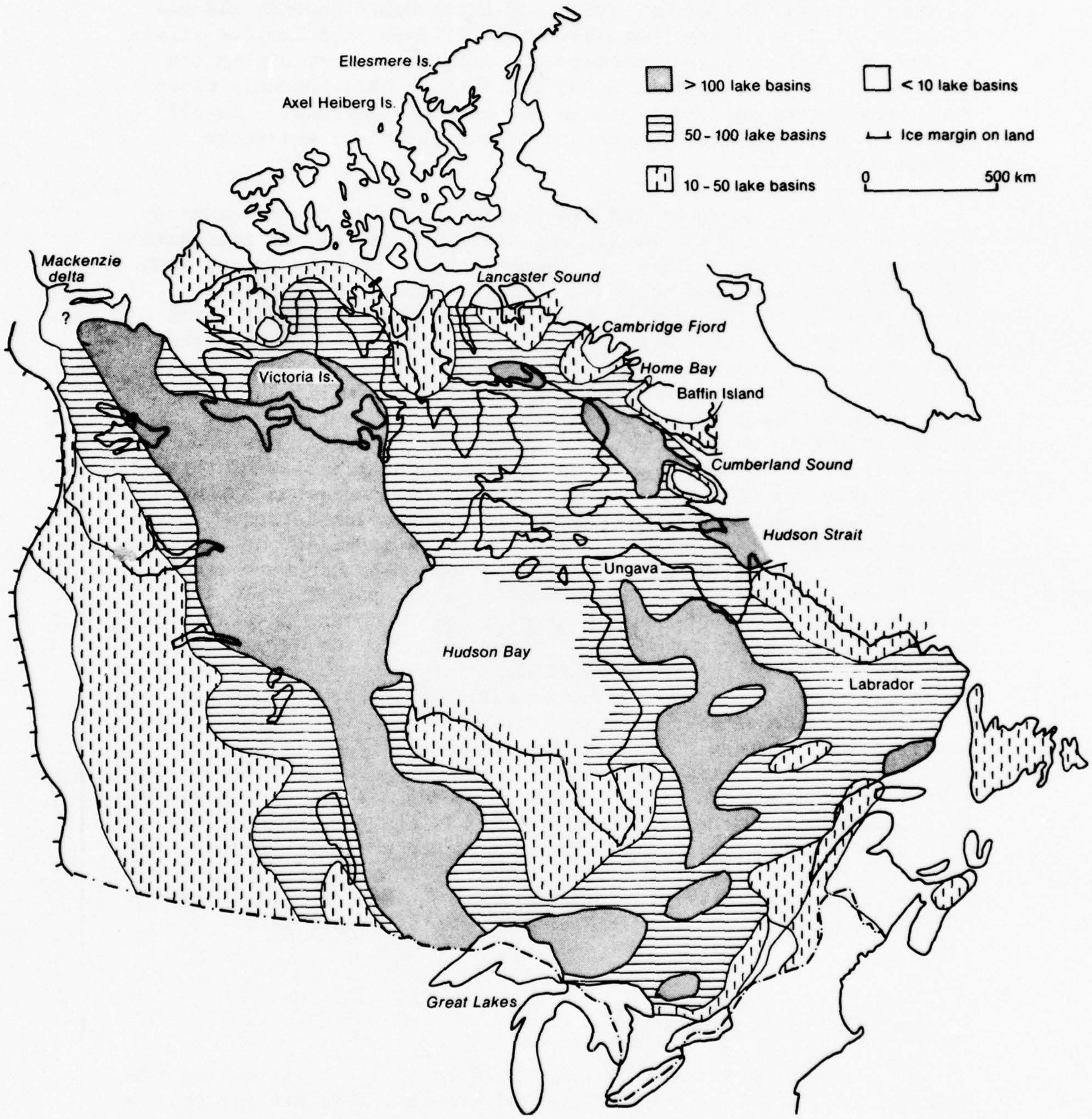


Figure 19. The density of lake basins (0.5 to 2.0 km. in diameter) in unit areas of 400 km.^{2} in Canada. The data was obtained by sampling from flat areas on topographic maps at a scale of 1:500,000. Lake basin density at this scale is thought to be a measure of the intensity of glacial erosion.

density was measured for that part of Canada formerly covered by the Laurentide ice sheet. Each Canadian topographic map at a scale of 1:500,000 (latest edition available in 1975) was divided into four equal parts. Within each sector a quadrat representing an area 20 x 20 km. in size was located by random methods and lake basins of 0.5 to 2.0 km. in diameter were counted. Areas of sloping topography were avoided where possible as well as areas of large lakes. It is impossible to know how many of the lakes sampled were actually excavated in bedrock. However, structural control of the lake borders was obvious in the vast majority of cases and it is unlikely that the inclusion of some non-bedrock lakes will seriously affect the pattern. One obvious area of difficulty was in the Mackenzie River delta where there are many 'periglacial' lakes and as a result this area was left blank.

The resulting map (Fig. 19) reveals a clear pattern which is in full agreement with that already mentioned. The boundary between little or no sign of erosion in the Queen Elizabeth Islands and areal scouring to the south is born out by the contrast in lake density. To the north there are less than 10 lake basins per 400 km.^{2}, whereas immediately to the south there are between 10 and 50 lake basins per 400 km.^{2}. This latter density soon increases southward to densities of 50-100 lake basins per 400 km.^{2} or more. Also of interest is a zone with a low density of lakes along the eastern board between Labrador and Lancaster Sound. Within this zone locally high densities occur in the vicinity of Hudson Strait, Cumberland Sound, Home Bay and the Cambridge Fiord area.

Perhaps the most intriguing pattern is the zone of high density basins (>100 per 400 km.^{2}), situated approximately midway between the shores of Hudson Bay and the former ice sheet margin. In the west the zone is continuous between the high western plains and the Arctic Archipelago. In the east it is less continuous, but a recognisable zone swings round from the Great Lakes through the Ungava Peninsula and is continued in two places in Baffin Island. If lake basin density is accepted as one measure of the intensity of glacial erosion, then one can suggest that there is a zone of maximum erosion situated between the former ice sheet centre and the periphery. The reality of this pattern is borne out by two independent maps which are reproduced in the National Atlas of Canada (1975). One, showing the <u>Area of Fresh Water as Percentages of Grid Units 10,000 square kilometres in Area</u> brings out the same zone as one with generally 10-20% of its surface area comprising water. The other, showing <u>Surface Materials</u> portrays areas where bedrock comprises most or half the surface materials. In addition to steep mountain areas along the east coast, there is an arcuate zone running from the Great Lakes to Victoria Island which is coincident with the postulated zone of maximum erosion. On either side most of the surface material consists of unconsolidated drift. Such a pattern would be expected if there were a zone of maximum glacial erosion approximately midway between Hudson Bay and the western ice sheet margin.

At present there seems no evidence to contradict this view of a zone of maximum glacial erosion midway between the ice sheet centre and periphery. For long there has been a view that an arc of exhumation between shield and Paleozoic rocks may be the result of glacial erosion, but as Gravenor (1975) has pointed out, there is often good local evidence that this existed as a depression in preglacial times and has merely been deepened by erosion in places. It is notable that the greatest deepening (in the vicinity of the Great Lakes) coincides with the intermediate zone of maximum erosion postulated in this paper. In 1972 White suggested that the zone of maximum glacial erosion occurred beneath the centre of the former Laurentide ice sheet, namely over Hudson Bay. This view was put forward as a speculative hypothesis, but it has subsequently been suggested that it is contradicted by a great deal of field evidence (Gravenor, 1975; Theme 1).

RELATIONSHIP OF LANDSCAPE TYPE TO PROCESS

The maps in the preceding section provide information about the variation in glacial erosion beneath the former Laurentide ice sheet both on a sub-continental scale and at a more detailed scale. It is the purpose of this section to relate the pattern to the role of the three main variables influencing glacial erosion - namely basal ice conditions, bedrock topography and bedrock geology.

Of all glaciological variables there are grounds for believing that the basal thermal regime is the most critical affecting the nature of erosion beneath ice sheets. Following from the work of Weertman (1961, 1966), Boulton (1972) has shown how the basal thermal regime controls the occurrence or otherwise of basal slip between ice and rock, the amount of meltwater at the base, the importance of the mechanisms of abrasion and plucking and the processes of entrainment of rock particles. Some of the possible relationships have been summarised by Sugden and John (1976) and are illustrated in Figure 20 which shows an idealised sequence of zones and suggested primary erosional processes associated with each. From this it can be suggested that cold-based ice is essentially protective unless it contains basal debris, inherited perhaps from a zone of basal regelation up-glacier. In this latter situation slight convexities on the bed may cause ice to diverge round the obstacle and bring debris into contact with the bed, as suggested by Röthlisberger (1968). Also some entrainment of debris may take place by plastic flow of ice round already loosened debris. Zones of warm melting (ice at the pressure melting point and subjected to melting) are associated with erosion by abrasion, fracture and meltwater, while zones of warm freezing (ice at the pressure melting point but with successive freezing on of meltwater) are associated in addition with entrainment by regelation and plucking.

The different types of erosional landscape may be related to differing basal thermal regimes. Areal scouring reflects extensive abrasion and plucking and would be most favoured by basal ice at the pressure melting point. Possibly areal scouring can

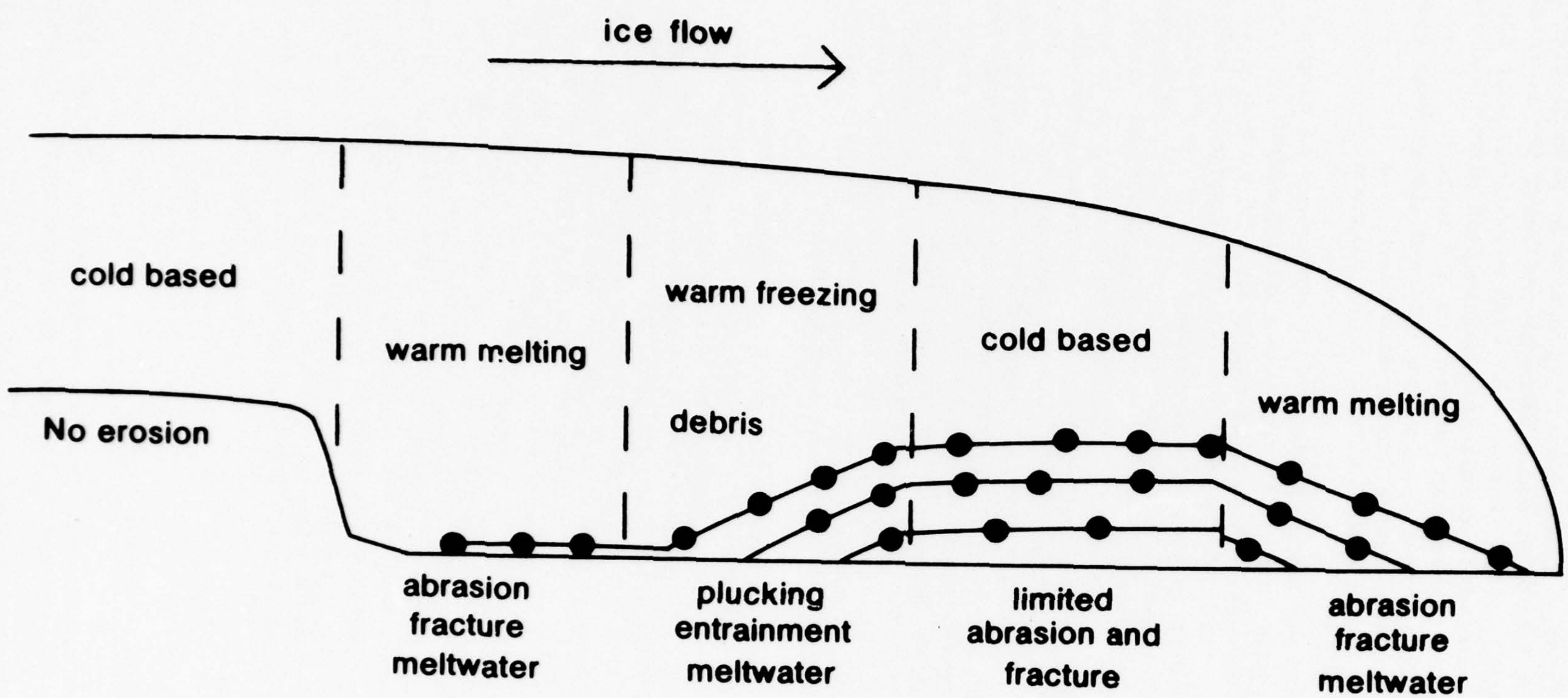


Figure 20. Idealised model of the relationship between processes of glacial erosion and the basal thermal regime of an ice sheet.

also result from the passage of cold-based ice when it is carrying a basal debris load. Landscapes of little or no sign of glacial erosion would be likely to occur whenever the basal ice is cold-based and carrying little debris. In such situations most forward movement of the ice would result from deformation within the lower layers of the ice, rather than between ice and bedrock. Landscapes of selective linear erosion are intermediate in that they may reflect a situation where ice is warm-based over the sites of troughs and cold-based over the intervening plateaux.

Topography has an influence on landscapes of glacial erosion in two main ways. It can have a direct effect by determining whether or not a particular massif will be submerged beneath the ice sheet. Thus it can be suggested that topography will be the dominant factor in determining whether or not a particular massif is eroded by ice sheet ice or local valley glaciers. Topography can also play an indirect role by influencing the basal thermal regime beneath the ice sheet. This may be through its influence on the thickness of the overlying ice. Temperatures beneath thick ice tend to be higher than those beneath thin ice because, with the exception of the surface layers, temperatures tend to rise with depth. Topography also affects basal temperatures through its influence on converging or diverging flow. In an area where the bedrock topography favours convergent ice flow then ice velocities will be increased and this will produce more heat by internal deformation and/or basal sliding. On the other hand topography which causes ice divergence will tend to reduce overall velocities and thus the amount of frictional heat generated.

The geology of the ice sheet bed can be expected to have important implications on the pattern of glacial erosion at a broad and local scale. The influence of joint patterns, fracture characteristics and hardness are all variables which have yet to be considered from a detailed theoretical standpoint and little can be said about them at present. However, permeability is known to be important in so far as it affects (1) the amount of water present beneath warm-based ice and thus the rate of basal slip (Weertman, 1966), and (2) the effective normal pressure applied to the glacier bed (Boulton, 1974, 1975). Beneath the thick ice of the Laurentide ice sheet it is reasonable to suppose that the main effect of permeable beds, especially if they are extensive, would be to allow basal meltwater to flow through the beds rather than at the ice rock interface. This in turn is likely to reduce the rate of erosion. In part this is because the rate of basal slip is reduced and thus less glacier sole and abrasive passes a given point on the bed, and in part because there would be a tendency for deposition to occur as particles lodge beneath the high overburden pressures (Boulton, 1975).

The role of basal thermal regime

The relationship between landscapes of glacial erosion and basal thermal regime can be examined by comparing the landscape distribution and character with the distribution of the differing basal thermal regimes beneath the former Laurentide ice sheet. Reconstructing the basal thermal regime of a former ice sheet is obviously an exercise fraught with difficulty and, in order to give adequate treatment of its limitations, I have felt obliged to complete the reconstruction in a separate paper (Theme 2). Figure 11 shows the result of this reconstruction of the basal thermal regime beneath the Laurentide ice sheet at its Pleistocene maximum.

The reconstruction of the ice sheet was made for maximum steady state conditions on the assumption that most erosion was achieved during maximum ice conditions. There are suggestive pieces of evidence which seem to confirm this assumption as realistic. One is simply that such ice sheet landforms as the larger fiords of eastern Baffin Island occur at the peripheries of the ice sheet and can hardly have been occupied by ice unless the ice sheet was very close to its maximum. Another tentative line of evidence is that erosional forms seem to relate to maximum ice flow conditions and are relatively unaffected by subsequent changes, as for example occur during deglaciation. An example of this situation is provided by the Ottawa Islands in eastern Hudson Bay. Here Andrews and Falconer (1969) noted that the streamlined outline of the islands reflected maximum ice flow conditions out of Hudson Bay, and yet overlapping striations of completely different orientation tell of the phase of deglaciation during which the ice centre migrated to Labrador-Ungava. The deglaciation phases were capable only of carving some striations into the previously ice moulded surfaces. Even if the view that most erosion was achieved during maximum conditions must remain an assumption at this stage, there is evidence that the northern hemisphere Late Wisconsin ice sheets accomplished little erosion and that therefore any reconstruction should be of earlier ice sheets. This view of limited Late Wisconsin erosion is based on the juxtaposition of erosional forms, for example in Europe (Ahlmann, 1919; Sugden, 1969) and on the location of inter-glacial deposits on glaciated surfaces, for example in West Greenland (Sugden and Miller, 1977) in eastern Baffin Island (Feyling Hanssen, 1976) and around southern Hudson Bay (Prest, 1970; Macdonald, 1969).

The main feature of Figure 11 is a zone of warm-melting beneath the central area of the ice sheet. Surrounding this central zone is a ring of warm-freezing followed by a ring of cold-based ice and finally in most places a peripheral zone of warm-melting. The pattern was subjected to a sensitivity test by changing the input data and it was concluded that the sequence of zones was relatively stable but that small changes in the input data could shift zone boundaries and widths by hundreds of kilometres. This was especially true of the western plains where the pattern was least stable. The cold-based ice over the Queen Elizabeth Islands was found to be especially stable and persisted in spite of all reasonable changes in input variables.

Comparison of the map of basal thermal regime with those of landscape variation poses several suggestive correlations. One is that the area of little or no erosion over the Queen Elizabeth Islands coincides with an area where the ice sheet was cold-based.

In the absence of any obvious correlation with either topography or geology, it is reasonable to suggest that there may be a causal link between the former occurrence cold-based ice and the areas of little or no erosion. Devoid of basal debris it is to be expected that such cold-based ice would be essentially protective. The only exceptions occur in the area of the deeper straits where calculation suggested that the basal ice could have been at the pressure melting point. It is notable that the deeper straits often contain glacial troughs (Pelletier, 1966) and also that zones of light areal scouring are sometimes associated with the sides of straits, for example between Cornwallis and Devon Islands (Fig. 14).

The central area of areal scouring around Hudson Bay coincides with an area where the basal ice was at the pressure melting points. This relationship would be expected. Especially interesting however is the apparent coincidence of the zone of most intensive glacial erosion with the zone of warm-freezing. The warm-freezing zone provides an efficient means of debris evacuation because successive regelation of meltwater and associated debris allows the build-up of a considerable thickness of debris bearing ice. Sample calculations for the Laurentide ice sheet suggest that the layer would be of the order of 20-50 m. thick. It is tempting to suggest that the zone of most intense erosion is associated with this zone of efficient debris evacuation. There is not a perfect spatial correlation between the zone of intensive erosion and the warm-freezing zone, but neither should this be expected. In the first place the width and position of the zone as portrayed in Figure 11 is highly sensitive to the amount of meltwater produced beneath the centre of the ice sheet. It may be that climatic conditions were different to those used in the reconstruction with the result that the position and width of the zone would have been different. In the second place, it is highly probable that the position of the zone fluctuated during the course of a glaciation as maximum equilibrium conditions took time to be established. The remarkable feature is that the correlation is as good as it is. One might point out that the correlation agrees over such details as the relative width of the zone which is wider in the west than in the east.

It might be suggested that geologic variations affect the position of the zone of intense erosion. For example, the western edge of the zone on the western plains coincides with the edge of the shield. Whereas such geologic conditions may partly influence the position of the zone in this area it is sufficient at this stage to note that the pattern also occurs in Labrador-Ungava where there is a uniform spread of shield rocks both within and on either side of the zone of intense erosion.

The areal scouring that occurs between the zone of intense erosion and the former ice sheet periphery coincides with both cold-based and warm-melting zones. There are several possible explanations and a full answer must await more detailed field study than has been attempted in this paper. One possibility is that the zone has been eroded beneath cold-based ice which was carrying a considerable thickness of debris inherited from the warm-freezing zone immediately upstream. Another possibility is that some areas, especially nearer the margin ice sheet periphery, have been subjected to erosion beneath the outer zone of warm-based ice. This zone was found to be the least stable of all and relatively modest changes in the climatic input variables could dramatically increase the width of the zone. Yet another possibility is that the areal scouring in this intermediate zone relates to conditions before or after the steady state pattern of Figure 11 was established.

There is no obvious and direct relationship between landscapes of selective linear erosion and basal thermal regime. In Greenland there is a tendency for the landscape type to occur in environments where the ice sheet was nourished under a continental climatic regime rather than a maritime climate (Sugden, 1974). Such a tendency may also be reflected in eastern North America where it can be seen that the landscape type occurs more frequently in the north than in the south.

The role of topography

At the scale of the Laurentide ice sheet as a whole the role of topography seems subdued and relatively unimportant. Much of the bed of the former ice sheet is relatively flat and gently sloping. The main exception is in the east where the land rises to form an uplifted, westward sloping plateau. It is generally accepted that important topographic features such as the straits of the Arctic Archipelago are pre-glacial, at least in their main outlines (Bird, 1967; Fortier and Morley, 1965). It is in these eastern and northern areas that topography has played an important role in glacial erosion.

Alpine landscapes occur along the eastern periphery of the Laurentide ice sheet and as might be expected appear to coincide with those massifs which rose above the maximum ice sheet surface profile (Fig. 21). The ice sheet profile in Figure 21 is based on the assumption that the ice was grounded as far as the 200 m. submarine contour. Inland of this margin the surface profile was obtained by analogy with existing ice sheets and has the form $y = a + bx + cx^2$ (where $a = .607$, $b = .005$ and $c = -.2 \times 10^{-5}$) (Fig. 1). The mountain altitudes were obtained by taking the mean of the three highest peaks in each Alpine landscape area and these were then plotted at the appropriate distance inland from the ice margin. In addition the closest adjacent massif unaffected by Alpine glaciation was also plotted on the diagram. Following Walcott (1970) an allowance was made for isostatic depression of the ground following inundation by the ice sheet and amounts to c. 100 m. at the ice margin and c. 300 m. at a distance of 200 km. from the ice margin. These figures give differential isostatic depression of 200 m. between the ice margin and a point 200 km. inland and this has been built in to Figure 21. Bearing in mind the assumptions involved in constructing such a diagram, the apparent correlation in Figure 21 between Alpine scenery and those massifs which rose above the ice sheet profile seems too good to be for-

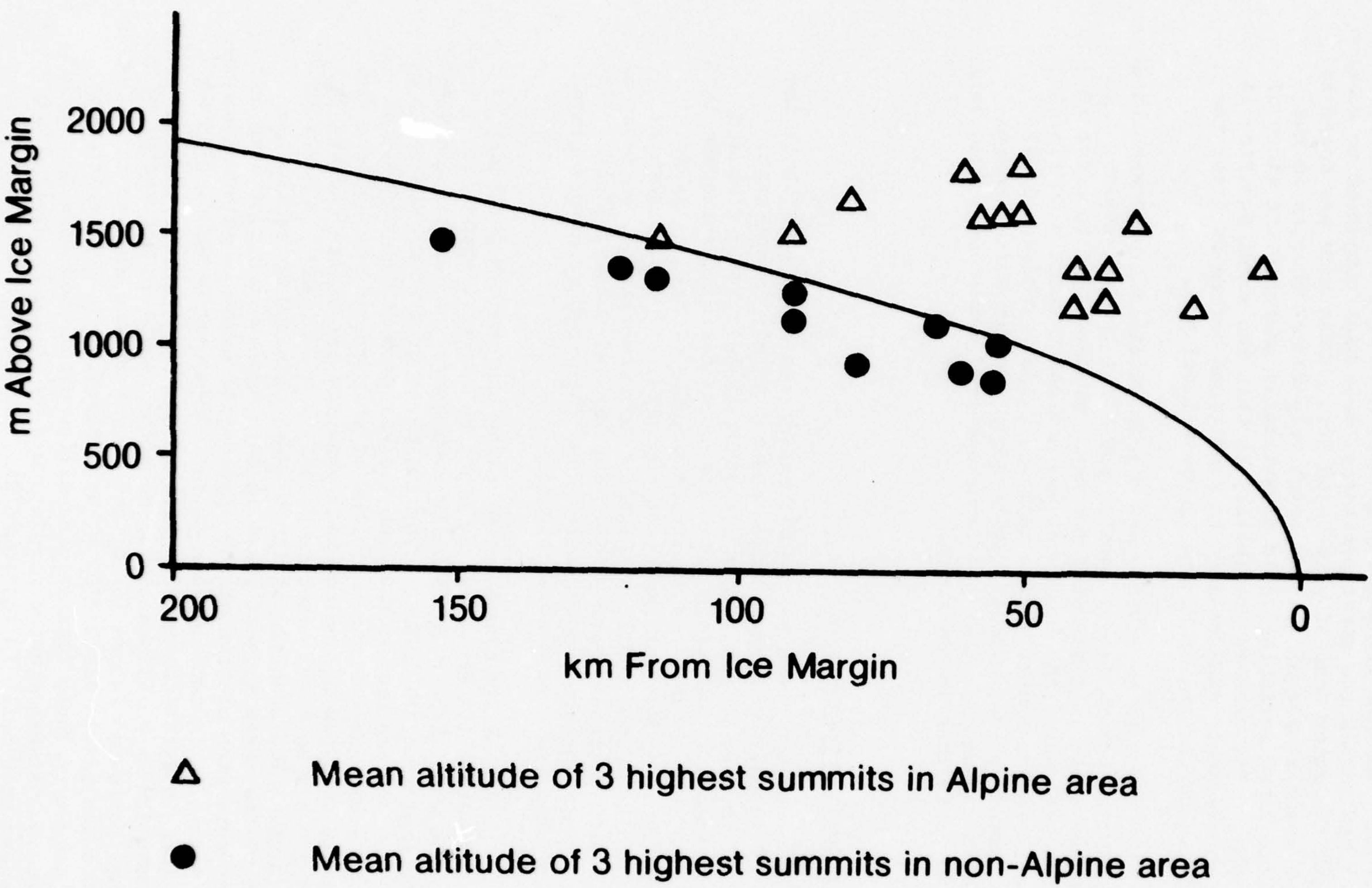


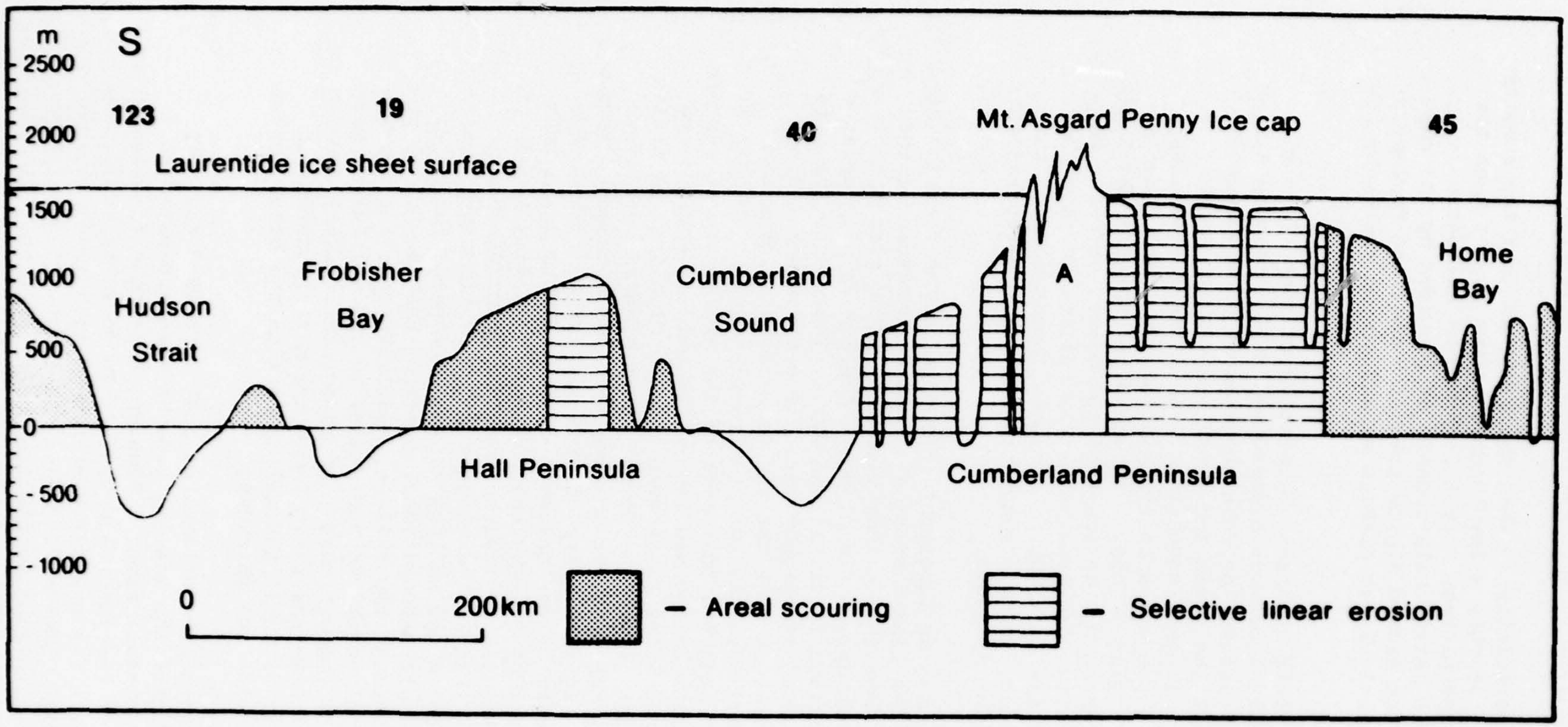
Figure 21. The relationship of Alpine landscapes to the former maximum Laurentide ice sheet profile. Alpine landscapes are confined to massifs sufficiently high to have escaped inundation by the ice sheet.

tuitious. This is especially so when one realises that the summits of an Alpine area may be only a few hundred metres higher than the summits of a non-Alpine plateau only a few tens of kilometres away. A good example of this latter relationship is provided by the contrast between the Mount Asgard Alpine area of Cumberland Peninsula and the adjacent slightly lower plateau upland around the Penny Ice Cap (Fig. 22).

In view of this apparent correlation it seems fair to suggest that the contrast between plateau and well developed Alpine scenery reflects the contrast between those areas once submerged beneath the Laurentide ice sheet and those areas which were sufficiently lofty to escape inundation. As a result these upstanding massifs were subjected to periglacial and local glacier action throughout the glacial age. As would be expected the local glaciers were constrained by the shape of the massif on which they formed and thus exploited pre-existing river valleys to form the dendritic trough pattern of today. At lower levels some of these massifs are traversed by troughs whose orientation clearly reflects the action of ice sheet drainage.

Topography has an important influence on the location of landscapes of selective linear erosion. All occurrences of the landscape type are associated either with the upland plateaux which form the eastern upland rim of North America or with uplands in the Arctic Archipelago (Figs. 14 and 22). The reason for this relationship may be that topography influences the basal thermal regime of the ice sheet so that the ice is warm-based over the troughs and cold-based over the intervening plateaux. The map of basal thermal regime suggests that most of the upland rim of eastern Canada was covered with cold-based ice. The main exceptions are parts of Baffin Island where there were locally high ice velocities (owing to the thinness of the ice cover) which created sufficient frictional heat near the base for the ice to be warm-based. However, the calculations in Figure 11 were carried out for an average ice thickness over a 100 km. wide 'channel' of ice. This means that any channelling effect of valleys was ignored. In such a case it is probably more realistic to imagine most ice being discharged down the valleys and that ice velocities and thus frictional heat was also concentrated in the valleys at the expense of the intervening uplands. Sample calculations, assuming that 50-75% of the ice was discharged through the valleys, confirmed that the ice was cold-based over the plateaux and warm-based in the valleys. Moreover the calculations suggest that this topographic effect causing convergence or divergence was about ten times more important in creating this pattern of basal thermal zones than the other contributing factor of varying ice thickness.

The topographic relationships just described may help explain the pattern once the glacial troughs were excavated to approximately their present size. However, they do not necessarily explain the origin of the landscape type. To do so it would seem necessary to postulate the existence of pre-glacial river valleys incised into the upland edge of Canada. Any such irregularities would chnnel ice flow as soon as the ice sheet built up. It is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– Areal scouring
– Selective linear erosion

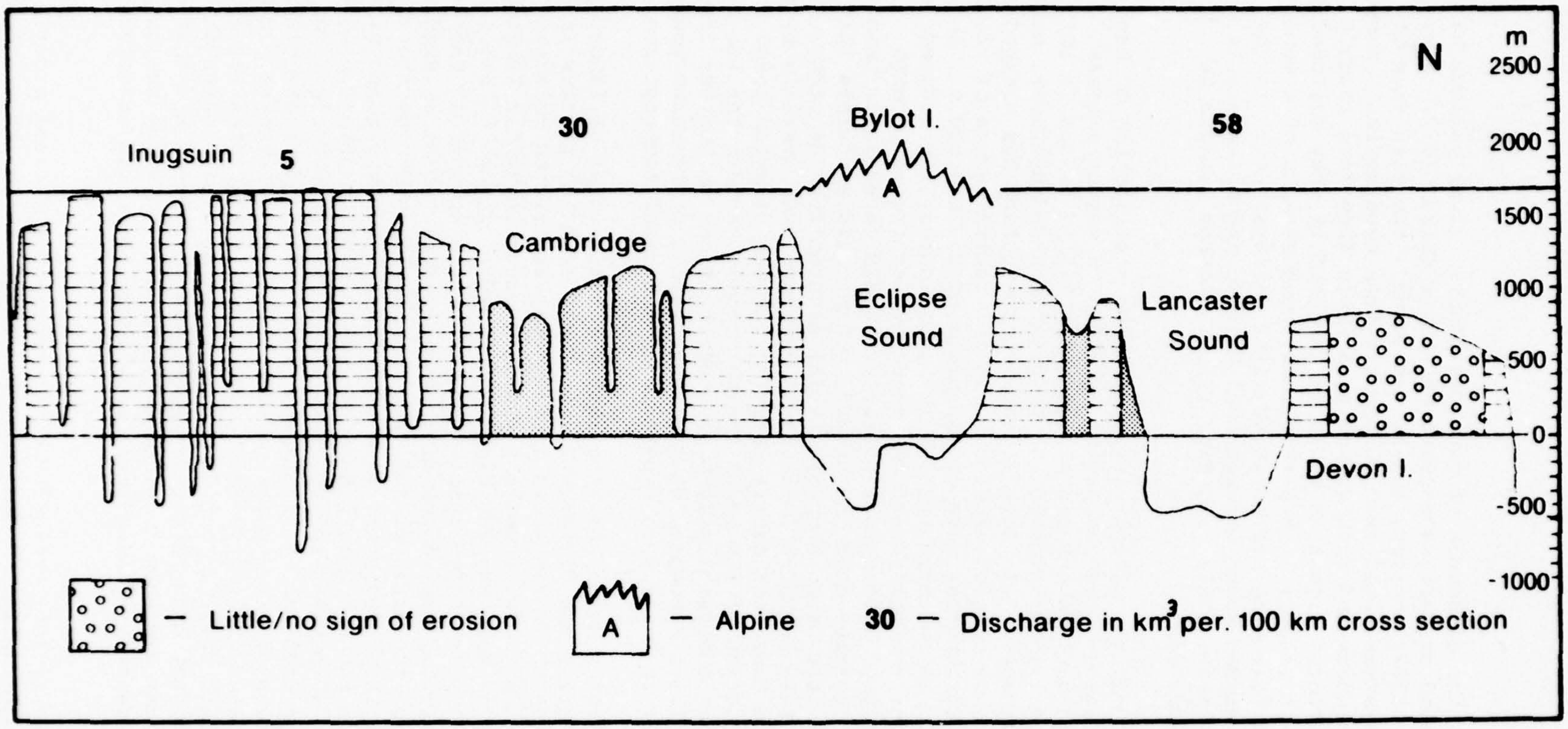

Figure 22. Profile parallel to the eastern margin of the Laurentide ice sheet and 130 km. in from the margin, showing relationships between landscape type, topography and ice sheet thickness. Differential isostatic depression between the margin and the line of the profile has been allowed for.

notable that the existence of such pre-glacial river valleys has been postulated on various grounds by others (Bird, 1967; Sim, 1964; Ives, 1957; Mercer, 1956). Moreover, the plan view of many fiord systems would seem to support this hypothesis. Many fiords are sinuous and when viewed as a whole they form roughly dendritic patterns - just as would be expected if they followed pre-existing river valleys. A particularly good example of a dendritic fiord system following pre-existing river valleys occurs in the vicinity of Nansen Sound between Axel Heiberg and Ellesmere Islands (Hattersley-Smith, 1961; and Fig. 19). Another example is given in Figure 23.

Topography clearly influences the distribution of landscapes of areal scouring at a local scale. A common pattern around Hudson Bay, Foxe Basin and southern Baffin Island is for uplands to be lightly scoured and to be surrounded by lower, heavily scoured landscapes. In some situations the upland has escaped all visible signs of glacial erosion. In the central areas of the ice sheet where thicknesses were of the order of 2,000-4,000 m., there would seem to be two main reasons for this, namely the reduced ice thickness over the upland and the tendency for ice to diverge round the upland. Both characteristics would cause lower than average temperatures over the upland. One could speculate that areas with little or no sign of glacial erosion such as parts of Southampton Island and Hall Peninsula may have been overlain by cold-based ice and that most debris carrying ice was diverted round them. The reconstruction of the basal thermal regime beneath the ice sheet confirmed that the relatively minor modifications to the ice temperature profiles accompanying such topographic differences would be sufficient to cause such a pattern in the basal temperatures.

Further north the topographic relationships of landscapes of areal scouring are different. As areal scouring becomes less common towards the north it is increasingly restricted to areas of ice convergence. Two excellent examples occur in Baffin Island in the vicinity of Home Bay and Cambridge Fiord. Figure 2? shows how these two areas are topographic lows across the uplands of Baffin Island. Calculations of ice discharge along the coast of Baffin Island made on the basis of the form and flowlines of the ice sheet suggest that these two zones were evacuating between six and nine times as much ice per unit length of coast than the intervening area. Clearly such a contrast can be expected to lead to important contrasts in the creation of frictional heat. Other fine examples of the association of areal scouring with zones of convergence can be seen at the head of Eclipse Sound, on either side of Somerset Island and on southern Devon Island (Fig. 14). In all these cases it can be suggested that much of the ice sheet was cold-based and that frictional heat was able to produce enough heat to raise the basal ice to the pressure melting point only where there was convergent ice flow.

These relationships between topography and areal scouring may be conceived as different ends of a continuum in that they are local variations superimposed on an overall tendency for basal ice temperatures to fall from the ice sheet centre towards the north. Areal scouring becomes progressively less common towards the north where it is increasingly confined to especially favourable topographic

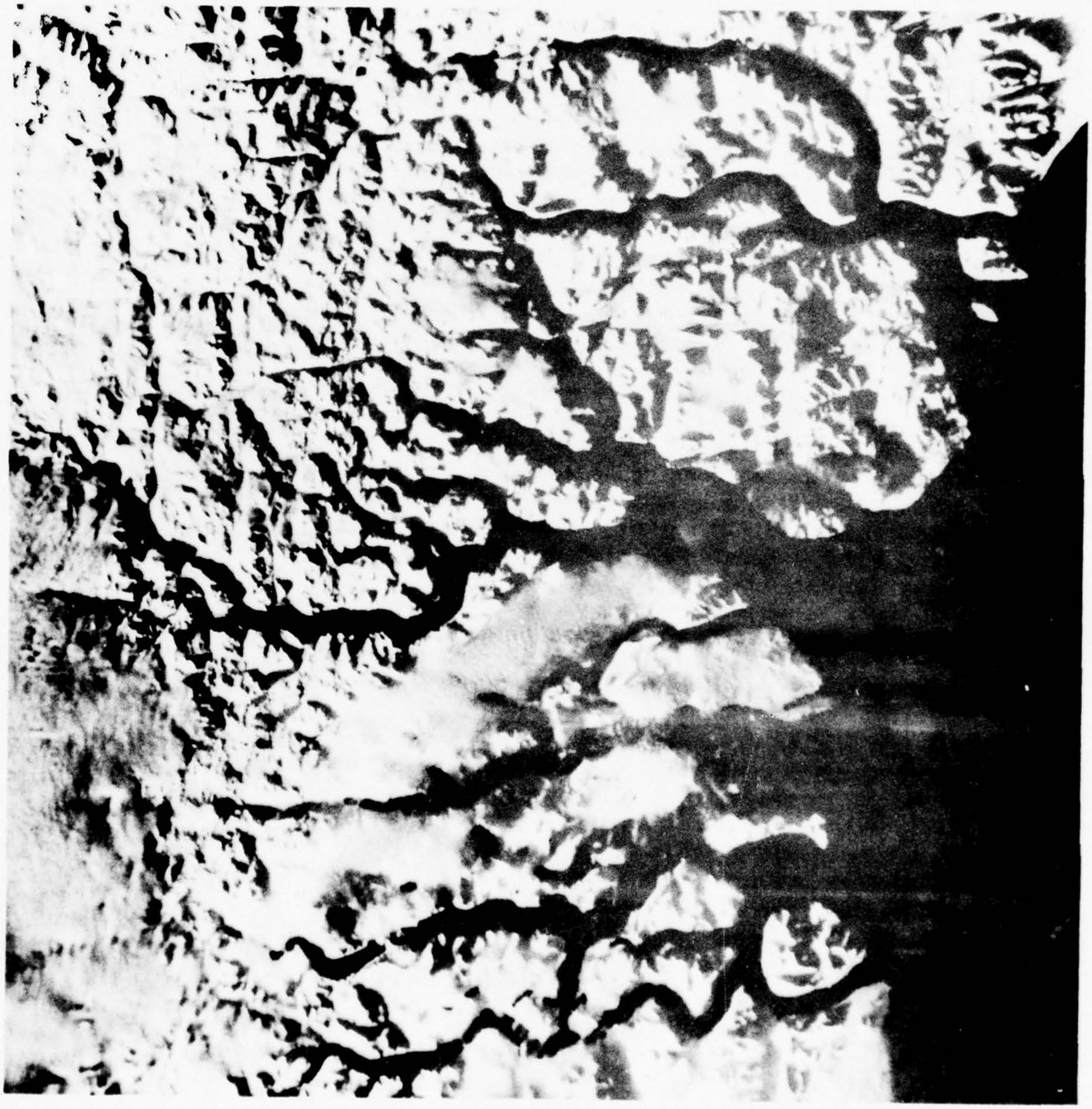

Figure 23. Dendritic fiord pattern in the Cambridge Fiord area, northeast Baffin Island as seen on a LANDSAT-1 image.

locations. The converse argument also applies to landscapes with little or no sign of glacial erosion.

The role of geology

Geological variations seem most important at a local scale, though their importance at a macro-scale cannot be overlooked in the case of North America. It can be suggested, for example, that there is a tendency for areal scouring to be associated with shield rocks and minimal erosion or deposition to be associated with younger Paleozoic sediments. However, examination of the shield boundaries reveals many anomalies which do not fit this simple generalisation. A particular example is Victoria Island which is clearly affected by areal scouring and yet which is largely underlain by Paleozoic rocks. At a more local scale the role of geological variation is clearer. Examples may be taken from northern Baffin Island and Somerset Island where there are extensive outcrops of Paleozoic rocks, mainly limestones. These rocks might be expected to be more permeable than the adjacent shield and one might therefore expect less erosion. Such appears to be the case. On Somerset Island the western boundary between areal scouring and areas with little or no sign of glacial erosion closely coincides with the boundary between Paleozoic limestones and shield rocks. Moreover an area of little or no sign of glacial erosion on Brodeur Peninsula coincides almost exactly with an exposure of Silurian limestone. To the west and south there is no such relationship between permeable rock and landscapes with little or no sign of erosion. Large areas of Melville Island, Southampton Island and southern Baffin Island are underlain by Paleozoic limestones and yet are heavily ice scoured.

A partial answer to this apparent paradox may concern the relationship of geological variations to other broad trends within the Laurentide area. The north Baffin and Somerset Island areas lie close to a boundary between areal scouring to the south and little or no sign of erosion to the north. In such a sensitive position variation in rock type may be able to play a role. Further south any local variation introduced by varied rock type is insufficient to overcome the overall tendency for areal scouring. As with topography, geological variations seem to be able to produce only limited local variations to a dominant overall pattern.

A MODEL OF LAURENTIDE ICE SHEET EROSION

Figure 24 is an attempt to conceptualise the possible relationships between erosional landscapes and the main variables affecting glacial erosion for the Laurentide ice sheet. The left-hand part of the ice sheet may be envisaged as an approximately north-south profile from Ellesmere Island to Hudson Bay, while the right-hand side is a composite profile from the east coast to Hudson Bay. Areal scouring affects most of the central ice sheet zone where basal ice conditions change from warm-melting at the centre, through warm-

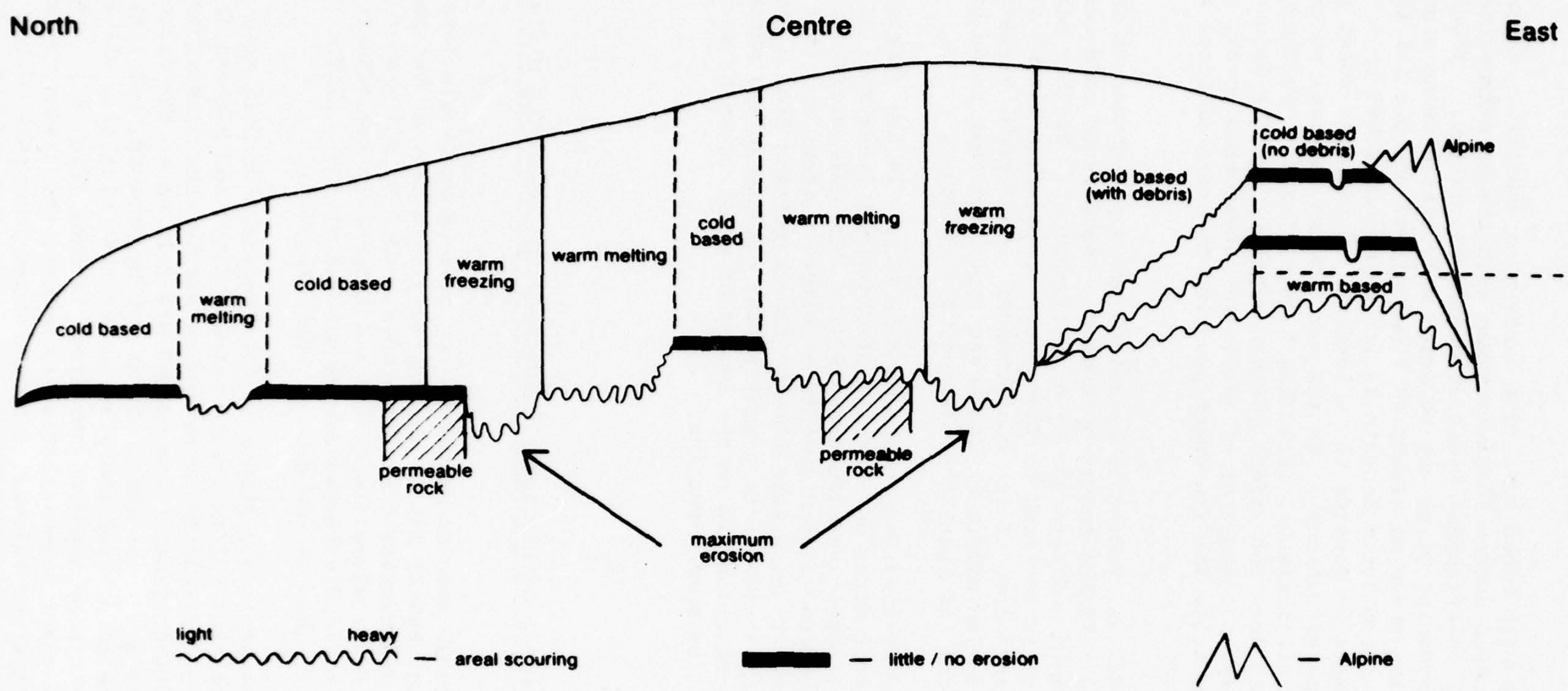

Figure 24. A model of the possible relationships between erosional landscapes and the main variables thought to affect glacial erosion, namely basal thermal regime, topography and bedrock geology.

freezing to a cold-based zone (but containing debris). A maximum amount of erosion occurs in association with the warm-freezing zone which affords an efficient means of debris evacuation. Towards the north areal scouring gives way to areas with little or no sign of glacial erosion where an essentially unmodified pre-glacial landscape has been protected beneath an ice sheet nourished in a continental climate. Towards the upland rim of the east coast there are three type situations. In the maritime south areal scouring extends over all surfaces, including high plateaux. Further north areal scouring does not affect plateau surfaces and the landscape type gives way to landscapes of selective linear erosion or, when the land rose above the ice sheet profile, to areas of local Alpine relief.

Local and regional variations are superimposed on these broad trends. In the centre uplands are overlain by cold-based ice and have escaped modification by areal scouring. Further north areal scouring is confined to favourable topographic sites, such as depressions and zones of ice convergence. Geological variations in bedrock have an effect on landscape type only near zones of transition from one basal thermal regime to another.

The conclusions represented by Figure 24 are very similar to those reached after an analysis of erosional landscapes around the peripheries of Greenland (Sugden, 1974). The main difference is that the topography around Greenland is more uniform and that one can therefore recognise more clearly the transition from a maritime climate in the southwest to a continental climate in the north and east and its relationship to the amount of areal scouring which falls off in the same direction.

RELATIONSHIP OF EROSIONAL LANDSCAPES TO LAURENTIDE MAXIMA

Though long suspected in the case of small glaciers and ice caps (e.g. Penck, 1905; Haynes, 1972), it has not yet proved possible to demonstrate that landscapes eroded by ice sheets are equilibrium forms whose dimensions and morphology are adjusted to the amount of ice discharged during glacial maxima. There are indications that this is the case in North America.

The near coincidence of the ice sheet profile with the plateau of central Baffin Island between Clark and Macbeth fiords offers a rare opportunity to determine whether the dimensions of the fiords which transect the upland are related to the discharge of Laurentide ice. In this case it can be assumed that little ice sheet ice moved over the plateau and that virtually all crossed the mountain rim by means of the fiord troughs (Fig. 22). Using the same ice sheet reconstruction as before, the discharge passing down each fiord was calculated by measuring the width of the ice mass tapped by each fiord. Maximum depths of the fiords occur at the point where the fiords pass the mountain crest (Løken and

Hodgson, 1971) and consequently cross-sectional areas were measured at this point, using 1:250,000 scale maps. Where known the true depth of the fiord was used in the measurements. In the remaining cases where the true depth was not known a median depth of 600 m. was taken (median between the shallowest and deepest fiords described by Løken and Hodgson, 1971). Bearing in mind the many possible sources of error it was a surprise to discover a strong and significant relationship between fiord size and ice discharge (Fig. 25). There are signs that this relationship continues offshore where troughs are cut into the continental shelf. The deepest offshore troughs are associated with the convergence of several fiords. Moreover the two deepest offshore troughs lie in the vicinity of Home Bay and Cambridge Fiord - the two main low-lying discharge routes across Baffin Island (Løken and Hodgson, 1971). This tentative evidence of a relationship between trough size and ice discharge at the maximum implies that the troughs have achieved some sort of equilibrium with ice flow conditions of the maximum.

A similar conclusion is also suggested by the relationship of the Alpine scenery to the projected ice profiles. It can be suggested that during the main periods of erosion the profile must have been lain in approximately the projected position. If it was higher more Alpine landscapes would have been submerged. If it was lower then more of the high plateau areas would have been modified by local valley glaciers. The consistent and clear distinction between the two sets of summit heights strongly suggests long periods of maximum ice conditions.

A final piece of evidence concerns the broad landscape patterns. It has been suggested in this paper that there are consistent and predictable relationships between landscape patterns on the ground and the basal thermal regime reconstructed for full Laurentide maximum conditions. This coincidence implies that the erosional landscapes may reflect conditions of the full maximum.

The conclusion that ice sheet erosional landscapes reflect maximum glacial conditions has several important implications:

(1) It affects the type of questions which may be asked about ice sheet erosion. In particular it suggests that the concept of an equilibrium landscape is potentially helpful in the analysis of links between process and form.

(2) It suggests that erosional evidence can be used to gain some indication of past ice sheet behaviour. It is sometimes argued that the Laurentide ice sheet was intrinsically unstable and that it never attained maximum steady state conditions for a long period of time. However, a period of the order of 50,000 to 100,000 years is probably necessary before steady state thermal conditions as represented in Figure 11 become fully established. If this is the case and if the apparent relationship between basal thermal regime and erosional landscapes is justified, then a long period (or periods) in a maximum steady state condition are implied. Certainly the events of the

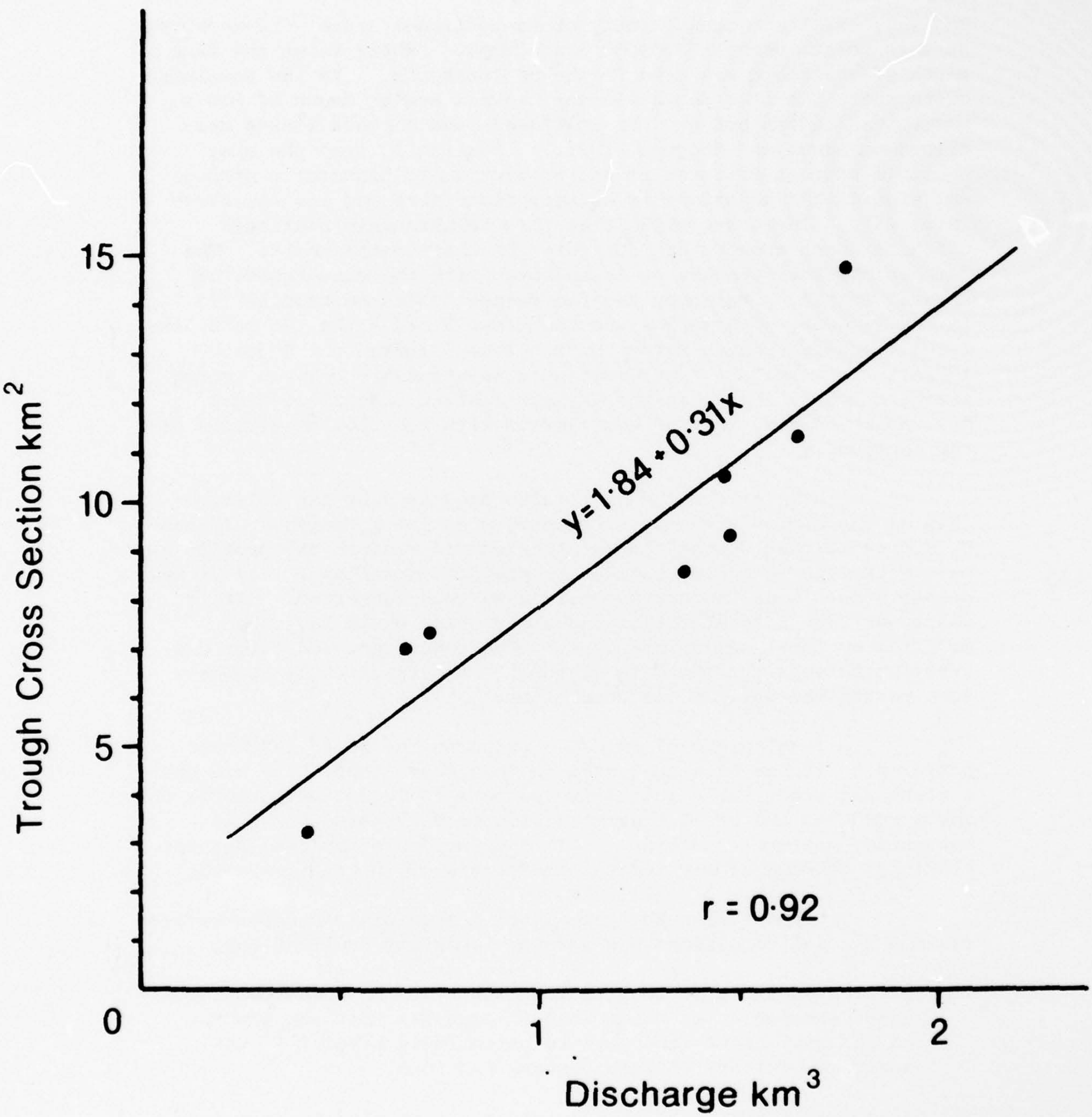

Figure 25. The relationship between fiord size and ice discharge from the Laurentide ice sheet, central Baffin Island.

Late Wisconsin glacial age are not typical of earlier maximum phases responsible for erosion of the main landforms.

(3) Another implication concerns the long-term possibility of improving the understanding of basal temperature variations in ice sheets through study of the landforms on the ground. This offers the prospect of working back from the field evidence <u>via</u> basal temperature regime to the refinement of the original input data such as ice thickness, surface climate and geothermal heat flow. This in turn provides another way of tackling the problem of reconstructing past Quaternary environments.

CONCLUSION

It is perhaps useful to conclude on a cautionary note. The purpose of this paper has been to present some information about landscapes of glacial erosion in the area formerly covered by the Laurentide ice sheet. Relying on spatial correlation it was suggested that the landscapes can be interpreted in terms of variations in the basal thermal regime of the ice sheet at its maximum, with modifications introduced by topography and bedrock geology. The spatial correlations are of course no proof of the reality of the postulated relationships. Together they merely comprise a hypothesis which is available for further testing and falsification. The hypothesis can hang around the following statements:

(1) Landscapes of glacial erosion are related primarily to the basal thermal regime of the ice sheet.

(2) Intense areal erosion is favoured above all by the availability of efficient mechanisms of debris evacuation rather than mechanisms of abrasion or fracture.

(3) Landscapes of glacial erosion are equilibrium forms related primarily to maximum glacial conditions.

GENERAL CONCLUSION

Although the main thrust of the research project is summarised in the conclusions of the individual papers provided above, it is helpful to provide a general conclusion by reference to the three reasons for the research (pp. 1-2).

The overall aim was to help understand the role of glaciers as erosive agents on the earth's surface. The main conclusion here is that the Laurentide ice sheet did not accomplish deep erosion in North America but merely etched a pre-existing surface.

An important specific aim was to develop a geomorphological approach which can contribute to the study of glaciology. It is hoped that the project has made a contribution by allowing the following postulates to be put forward as hypotheses available for further testing:

(1) that erosion by ice sheets is influenced above all by the basal thermal regime existing at the glacial maximum;

(2) that mechanisms allowing evacuation of debris rather than those of abrasion or fracture may be the most important in influencing the amount of glacial erosion achieved by an ice sheet;

(3) that cold-based ice may accomplish erosion if it has a debris load inherited, for example, from an upstream zone of regelation;

(4) that the morphology of the Laurentide ice sheet bed was in equilibrium with maximum glacial conditions;

(5) that the Laurentide ice sheet seems to have attained a steady state maximum condition at some stage in the Pleistocene; and

(6) that landscapes of glacial erosion may be used as a means of reconstructing former glacial climates and glacier morphology.

A further aim of the research project was to assess the value of satellite imagery as an aid to the study of macro-scale geomorphological problems. It is hoped that the study has shown that the use of such imagery allows the geomorphological mapping of vast areas with relatively small resources. This means that imagery allows geomorphologists to tackle macro-scale projects.

RECOMMENDATIONS

(1) A similar project relating landscapes of glacial erosion in Scandinavia and Britain to former basal thermal regime would further develop and refine ideas of glacial erosion.

A European study would have the particular advantage that detailed geological information would be available. This would allow more detailed understanding of the role of geological variables in influencing glacial erosion.

(2) The approach relating landscape type to basal thermal regime could be profitably applied to glacial deposition. In general more is known about glacial deposits than glacial erosion and this would allow relatively sophisticated treatment of the causes of deposition and the prediction of the character and distribution of glacial deposits. Such an exercise would be of possible practical value beneath the waters of the North Sea and would also have implications for the Sand and Gravel industry in formerly glaciated areas.

(3) One implication of this research project is that the distribution of different landscape types can be used to reconstruct the basal thermal regime and thence the morphology and climatic environment of former ice sheets. This would seem to offer a powerful new tool for the study of Quaternary climatic fluctuations in Europe and North America.

(4) A further implication of the research project is that the landforms of the bedrock base of existing ice sheets can be predicted. Such an exercise carried out for Antarctica and Greenland would be of future academic and economic importance.

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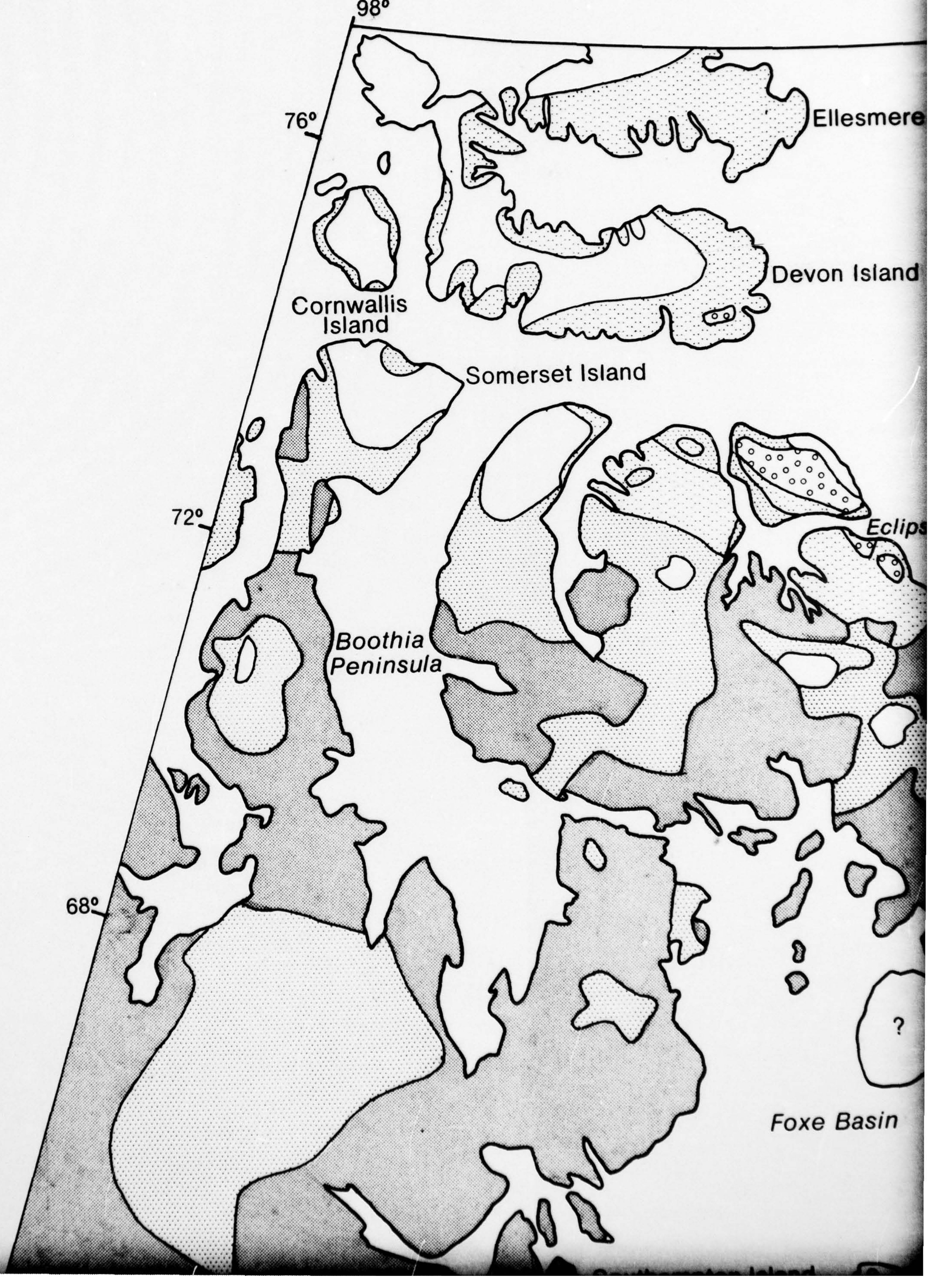
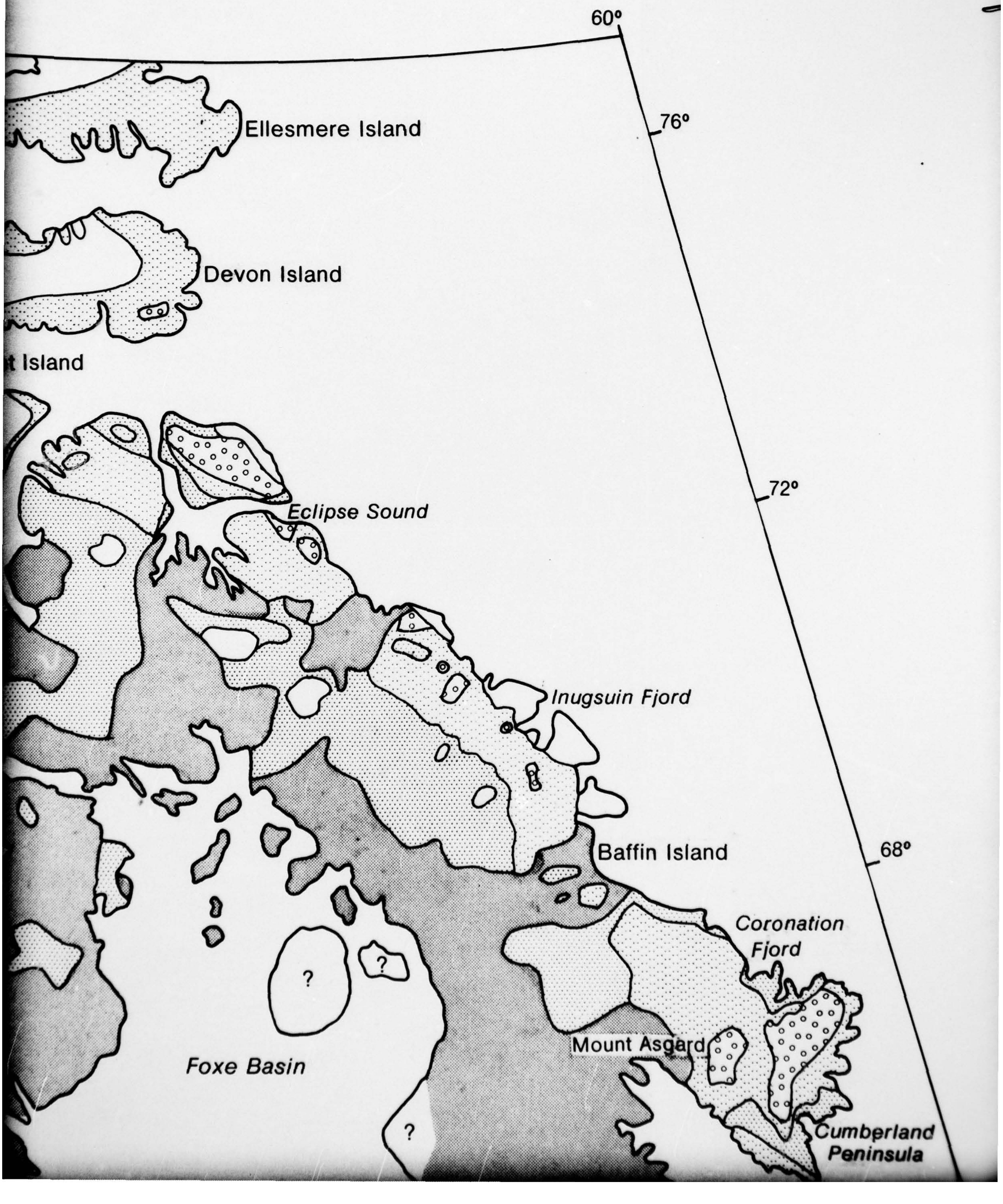


98°
76°
72°
68°
Ellesmere
Devon Island
Cornwallis
Island
Somerset Island
Eclips
Boothia
Peninsula
?
Foxe Basin


60°
76°
72°
68°
Ellesmere Island
Devon Island
t Island
Eclipse Sound
Inugsuin Fjord
Baffin Island
Coronation
Fjord
Mount Asgard
Foxe Basin
Cumberland
Peninsula
?
?
?

AD-A036 246 ABERDEEN UNIV (SCOTLAND) DEPT OF GEOGRAPHY F/G 8/7
GLACIAL EROSION BY THE LAURENTIDE ICE SHEET AND ITS RELATIONSHI--ETC(U)
SEP 76 D E SUGDEN DA-ERO-76-G-001

UNCLASSIFIED NL

2 OF 2
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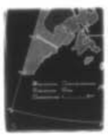
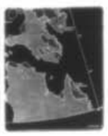
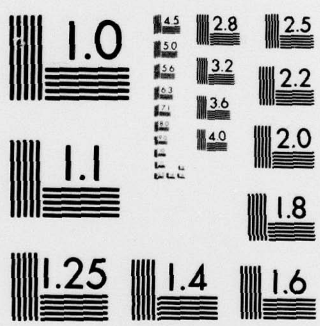
A036246

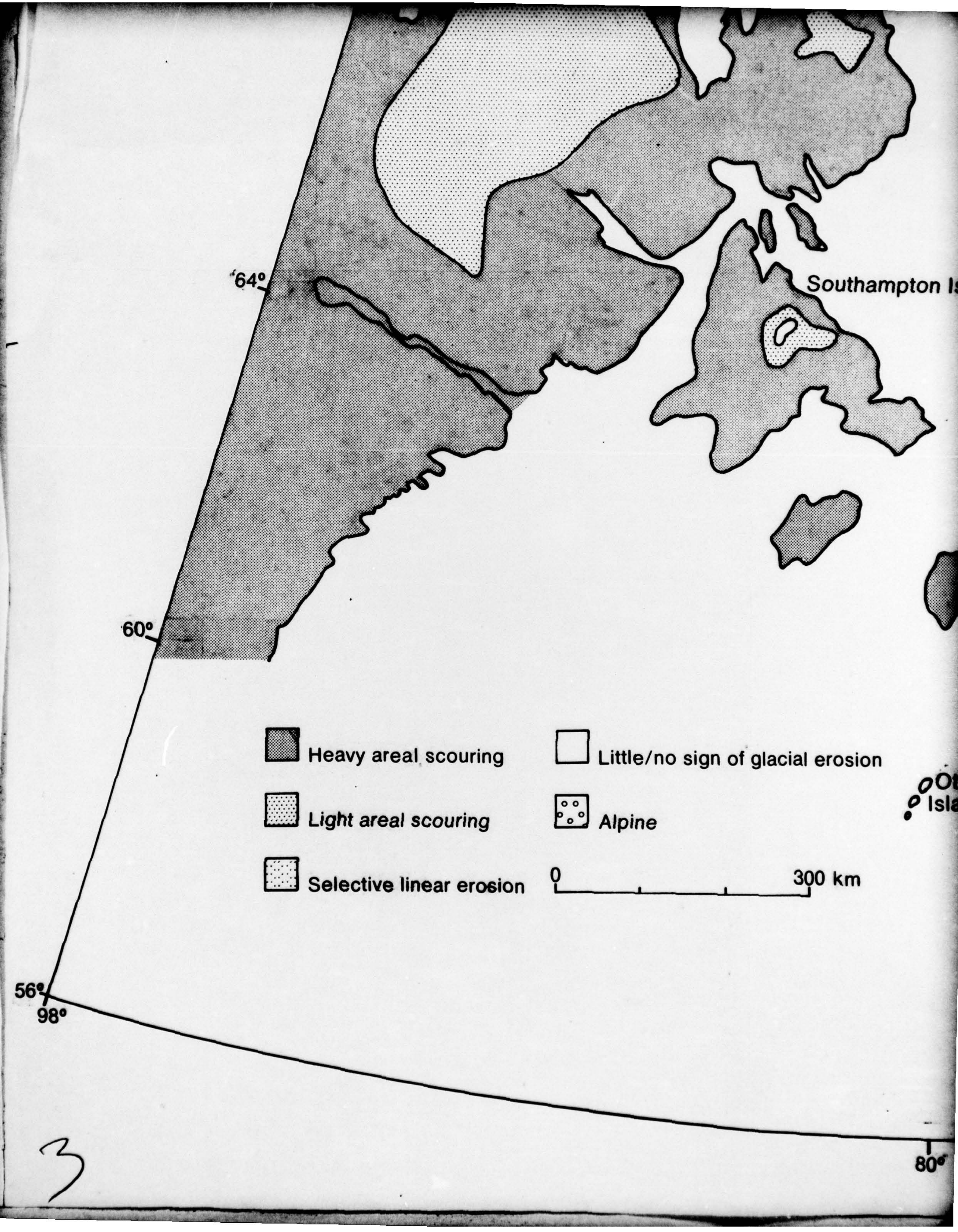




END
DATE
FILMED
3-77

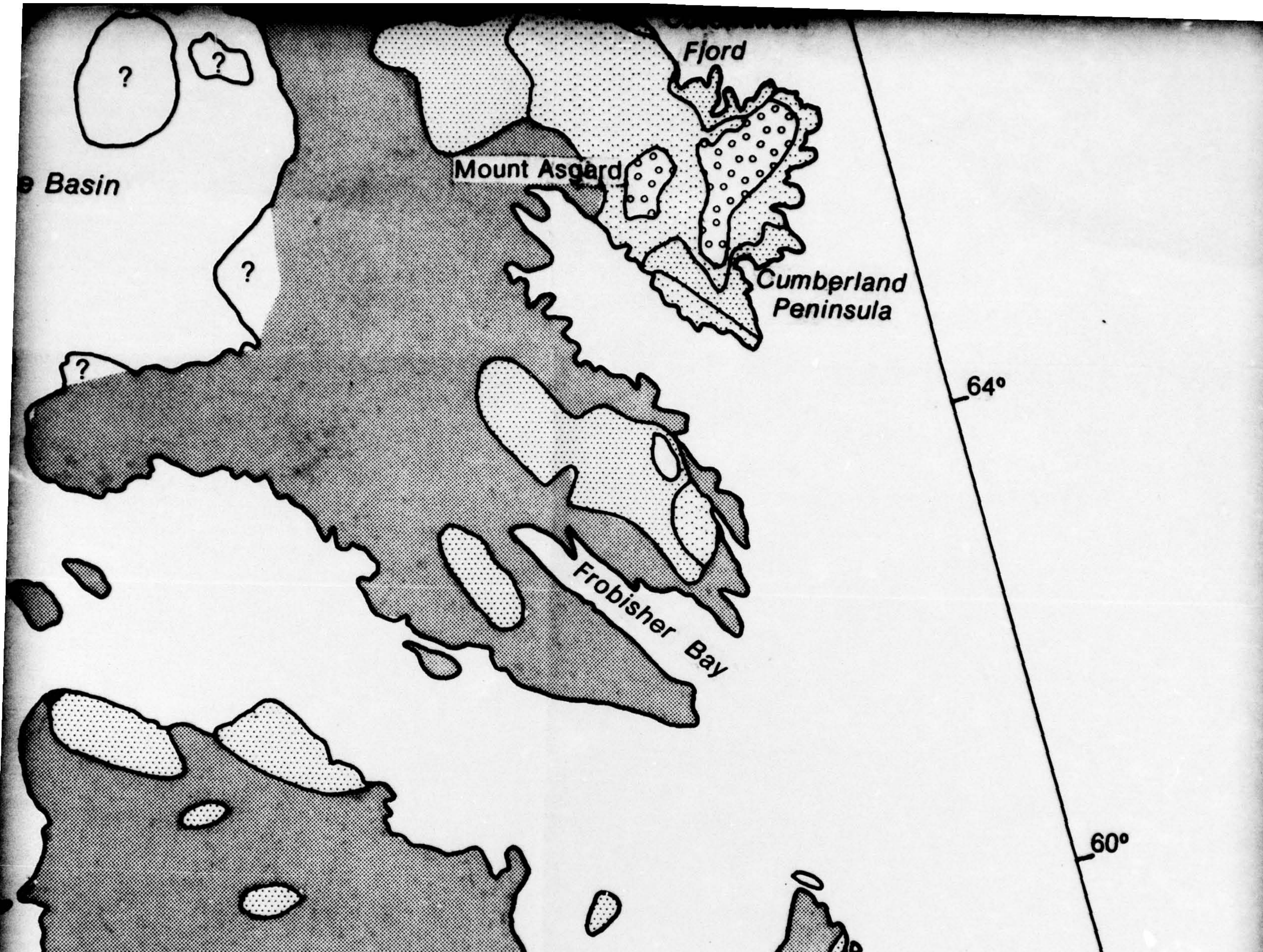


MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS-1963-A



Southampton Is
64°
60°
56°
98°
80°
Heavy areal scouring
Light areal scouring
Selective linear erosion
Little/no sign of glacial erosion
Alpine
0
300 km

3



Fjord
Mount Asgard
Basin
?
?
?
?
Cumberland
Peninsula
64°
Frobisher Bay
60°
Torngat
Mountains
56°
60°

1:5,000,000 4