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**A SEDIMENTOLOGICAL AND GLACIOLOGICAL STUDY
OF BASAL DEBRIS ENTRAINMENT AND TRANSPORT
IN POLAR GLACIERS OF SOUTHWEST BYLOT ISLAND, N.W.T.**

**by
Christian Michel Zdanowicz, B.Sc.**

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of the requirements
for the degree of Master of Science

Department of Earth Sciences

Carleton University

Ottawa, Ontario, June 1994

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submitted by
Christian M. Zdanowicz, B.Sc.
in partial fulfilment of the requirements for
the degree of Master of Science

Thesis Supervisor

for Co-supervisor F.A. Michel

Chairman, Department of Earth Sciences

Carleton University

July 1994

Abstract

An investigation of subglacial debris entrainment processes has been conducted on polar mountain glaciers of southwestern Bylot Island, N.W.T.. Basal debris and bedrock samples were analyzed by x-ray diffraction to distinguish between clay-size debris derived from contrasting Archean-Aphebian and Cretaceous-Tertiary substrate lithologies. Results indicate that only the lowermost basal ice layers (stratified facies) in glaciers reflect incorporation and mixing of debris from the underlying bedrock, while the bulk of the basal zone (dispersed facies) carries sediment derived mostly from up-ice regions of their flowpath. Co-isotopic analysis ($\delta^{18}\text{O}$ - δD) of dirty ice in the dispersed facies indicates a process of basal accretion by regelation. To account for these observations, a model of debris entrainment and transport in glaciers of the study area is presented, which suggests the existence of polythermal conditions at the glacier bed and the existence of several distinct zones of basal debris accretion.

Acknowledgements

My involvement in this project was initially suggested by Dr. Michel A. Bouchard of Université de Montréal. I am very grateful to him for introducing me to a fascinating field of research. The thesis was co-supervised by Dr. William Shilts of the Geological Survey of Canada and Dr. Frederick Michel of the Dept. of Earth Sciences. I am thankful to Bill Shilts for sharing some of his experience with me, and to Fred Michel for providing insights on isotopic glaciology and its applications. Critical review by Peter G. Johnson (University of Ottawa) was highly appreciated. Ron DiLabio (GSC) gave me access to unpublished data from Bylot Island. Isotopic data analysis was made following helpful suggestions by Dr. Roland A. Souchez (Université Libre de Bruxelles) and Dr. Chris Burn (Carleton University).

The project was mainly supported through a grant from the National Science and Engineering Research Council held by Dr. Shilts. Additional field expenses were covered by the Northern Science Training Program. Helicopter support on Bylot Island was provided by the Polar Continental Shelf Project (PCSP), and field equipment was loaned by EMR-Technical Field Support Services. NSERC also covered conference expenses for the Geological Society of America 1993 meeting (Boston, MA) and at the 1994 INSTAAR Arctic Workshop (Boulder, CO). Additional travel grants from the GSA and the National Science Foundation (held through INSTAAR) are gratefully acknowledged.

Field work on Bylot Island was conducted under the guidance of Bill Shilts and Fred Michel, and with the appreciated participation of Rein Vaikmae (Estonian Academy of Sciences) and of Thomas W. Edwards (University of Waterloo). I am very grateful to the PCSP staff at Resolute Bay and to the helicopter crew - Walter Sizinski, Dan Ireland and John McEmpherson - for their support and patience. Field assistance was ably provided by Shirley McCuaig, Mark Elver, Anne-Marie Bourgeois and Kate Ault.

Several people have contributed to the analytical work. My gratitude first goes to Dr. Ken Torrance (Dept. of Geography), who gave me access to facilities at the Geotechnical Science Laboratories and showed endless patience during my frequent consultations. Thanks also to Ron Conlon at the X-Ray Diffraction Laboratory (Dept. of Earth Sciences) for swiftly running all of my analyses (without breaking a single slide). Jim Howes gave me many useful tips on XRD work. Laser scanning particle-size data were helpfully provided by Patti Lindsay and Preston Wilson at the Sedimentology Laboratory of the GSC. Kellie Eamon kindly took charge of the dry weighing procedures, while thin sections were prepared by John Everest. The stable isotope analyses were conducted by Nathalie Morisset at the University of Ottawa.

Last but not least, I wish to address special thanks to all of my fellow students in room 246 for providing much-needed therapeutic silliness in moments of stress over these last three years.

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Introduction

Since the early 1950s, an increasing number of researchers have investigated processes active in the basal zone of modern glaciers, both in temperate alpine environments and at polar latitudes. The initial concern was mostly related to rates of glacial erosion on bedrock, the dynamics of glacier sliding, and the genesis of till and associated landforms. With the start of deep ice coring projects in the 1970s, the scope of these studies has expanded to ice sheets, as knowledge of subglacial processes is needed to allow reconstruction of the paleoclimatic record in deep ice layers. Recently, the question of subglacial sediment transport has seen a revival of interest related to the hypothesis of deformable till beds and their control on large-scale ice sheet dynamics.

Weertman (1961), and later others (Kamb and Lachapelle 1964, Lliboutry 1964, Nye 1970, Robin 1976) introduced the concept of subglacial debris accretion by the refreezing of water at the glacier sole through several processes (regelation, cavitation regelation, net freeze-on). Field studies by Kamb and Lachapelle (1964), Boulton (1970), Souchez *et al.* (1978), Lawson (1979) and others have since provided support to this concept, and stressed its importance to subglacial entrainment of basal debris in polar and sub-polar glaciers.

An important step in the understanding of subglacial processes in glaciers and ice sheets was made possible by the theoretical and experimental work of Souchez and Jouzel (1982, 1984) who elaborated a method allowing basal ice formed by melting-refreezing to be distinguished from precipitation-derived glacier ice on the basis of its composition in stable oxygen and hydrogen isotopes. The souchez-Jouzel model was subsequently refined to account for the effects of mixed water input situations (Souchez-DeGroote 1985) and of varying freezing rates (Souchez *et al.* 1987).

Since it was introduced, the method has been applied by numerous authors to the study of debris entrainment in temperate, sub-polar, and polar glaciers (Lorrain *et al.* 1981, Gordon *et al.* 1988, Sugden *et al.* 1987, Knight 1989, Souchez *et al.* 1990).

The consensus in current literature generally supports subglacial accretion as a predominant process for debris entrainment in temperate and sub-polar glaciers. The detailed work of Boulton in Svalbard (1970, 1972) has suggested that subglacial accretion may result in the formation of relatively conformable sequences of debris bands in the basal zone of glaciers. Unless major distortion of the basal zone occurs up-ice from the compressive marginal zone, this sequence may be preserved in the glacier over a considerable distance without major homogenization of the debris load occurring during transport.

To date, few efforts have been made to verify the validity of this contention. The present study will attempt to do so by applying current knowledge on processes occurring at the ice-bedrock interface to a situation in which a fairly simple compositional constraint may be used to relate the debris load of glaciers to a specific source area. The work of DiLabio and Shilts (1978, 1979) on southwestern Bylot Island, N.W.T. has demonstrated the area's potential for such a study, as it combines two major advantages:

- 1) glaciers on southwestern Bylot Island flow from an alpine Archean-Aphebian crystalline core area on to lowlands underlain by soft Cretaceous-Tertiary clastic sedimentary rocks; the strongly contrasting lithologies underlying these two terranes provide an excellent opportunity to link mineralogical characteristics of glacier debris to source outcrops.

- 2) large piedmont outlet glaciers extending onto the outer lowlands are characterized by high (10-60 m) near-vertical margins that provide excellent exposure of internal structure and stratigraphy.

The aim of this thesis is to establish a simple model of basal incorporation and transport of glacial debris in glaciers of southwestern Bylot Island. This investigation will be conducted by utilizing the contrasting nature of the bedrock sources in the study area as a constraint for mapping the spatial mineralogical variations of the transported debris in the ice mass. Isotopic characterization of the basal ice by the Souchez and Jouzel method will help define the processes responsible for debris entrainment.

Specific research objectives:

1^o To describe spatial (vertical and longitudinal) variations in the basal debris load of some of the large glaciers in the study area. To this end, the clay mineralogy of the debris will be examined by semi-quantitative X-ray diffraction analysis, based on the investigative approach outlined by DiLabio and Shilts (1979);

2^o to link the various components of the clay fraction of basal debris to their potential bedrock source by comparison with samples collected from the various terranes of the study area, and to establish the relative importance of each terrane's contribution to the debris load;

3^o to use the observed mineralogical variations in the basal debris load of glaciers and their relation to specific bedrock terranes to characterize patterns of subglacial debris incorporation and transport in the ice.

4^o to determine if significant melting-refreezing occurs in the basal zone using the co-isotopic analysis applied to debris-rich basal ice layers. This will help to establish whether subglacial accretion of debris by freezing-on or regelation processes is important in glaciers of the study area.

Interpretation of data will consider *a)* the efficiency of the sampling strategy in view of glacial features observed in the field; and *b)* the accuracy of analytical procedures, based on estimated error on the various techniques employed. The discussion will examine the significance of results in the light of previous work, and the conclusions will comprise some tentative implications regarding glacial geological processes and their relation to the thermal regime of glaciers in polar regions.

Chapter 1

Theoretical Background

In this chapter, some important concepts and theories relating to subglacial processes are reviewed in order to establish a theoretical background for data interpretation. Special attention is given to high-latitude mountain glaciers such as those in the study area.

Note: This chapter is intended to be introductory. For an exhaustive discussion of glacial thermodynamic and sedimentological processes, see Patterson (1981), Drewry (1986), Hubbard and Sharp (1989) and Souchez and Lorrain (1991).

1.1) Glacier thermodynamics and the subglacial zone

Thermal regime of a glacier

If we assume that temperature distribution inside a glacier is solely governed by thermal conduction, then the vertical temperature gradient inside the ice mass is essentially a function of the temperature at the ice surface and the bed. The surface temperature is determined by local meteorological conditions and long-term climatic variations while the bed temperature is governed by the heat flux from the bedrock substrate, itself a function of the geothermal gradient and bedrock thermal conductivity. Thus, the temperature profile through a glacier essentially reflects the balance between cold surface temperatures and warmer basal conditions.

However, various factors complicate this simple model (Fig.1.1). Strain heat associated with mechanical deformation of ice, and frictional heat resulting from basal sliding during flow contribute to internal warming of the glacier. Both terms are proportional to the ice flow velocity.

Fig. 1.1. Factors affecting the vertical temperature profile in a moving ice mass. MAST = mean annual surface temperature; PMP = pressure-melting point; A = snow accumulation; A_b = snow ablation; U = advection of heat by ice movement. (from Drewry 1986)

Another important effect is that of advection, i.e. the longitudinal transfer of heat along a glacier as a result of differential accumulation and ablation. Progressive burial of annual snowfall carries cold temperatures down into the ice body. Where snowfall is important, as in the higher regions of a glacier, low temperatures will be found to a greater depth than where accumulation is small. The transfer of "cold" ice through the glacier is a function of the net snowfall input and the ice velocity: as both terms vary considerably from the accumulation to the ablation area, a different temperature profile will be found along different points of the flow line.

Thermal classification of glaciers

Snowfall accumulation and surface temperature are geographically variable factors, governed by local meteorological conditions. The same applies to relief which determines the bed slope gradient and strongly controls ice flow velocity. Thus, the thermal regime of a glacier is linked to its geographical location, mostly in terms of latitude and altitude. A common system classifies glaciers by thermal regime as well as geographical location:

- 1) A glacier at the pressure-melting temperature throughout is designated as a *warm-based* or *temperate* glacier. Basal melting occurs at the sole of temperate glaciers, and liquid water can exist in equilibrium with ice inside its mass. This reflects a situation in which the mean annual temperature at the glacier surface is higher than that of the bed, so the vertical thermal gradient is reversed. Basal heat is in part advected through ice flow while another fraction is consumed by englacial melting. Excess heat is dissipated by melting of the ice at the glacier-bedrock interface. As a result, the ice velocity and erosional potential of temperate glaciers is generally high. Temperate glaciers are distributed over a wide range of latitudes ranging from subpolar to tropical and are mostly located in high altitude alpine environments.

2) By contrast, *cold-based* or *polar* glaciers are defined as having ice below the pressure-melting point from bed to surface. This is the case for glaciers through which most of the heat flow can be conducted out, i.e. the vertical thermal gradient is such that the surface is colder than the bed. Cold-based glaciers are so called because their sole is frozen to the substrate. Their flow velocities are generally lower than those of temperate glaciers owing to the fact that flow occurs mostly by internal deformation with limited basal sliding. Erosional rates are also comparatively lower. Polar glaciers are found in areas of low mean annual temperatures, mostly at high latitudes where they occur either in alpine settings or as outlets of large ice caps or ice sheets.

This classification is of course an oversimplification: warm-based and cold-based conditions often coexist in the same glacier, and this is probably the case for most glaciers. Therefore, the terms temperate and polar glaciers should be applied to glaciers that are perennially at or below the pressure-melting point throughout most of their mass, respectively. The terms *Sub-polar* or *polythermal glaciers* may be used to designate those glaciers in which, at least seasonally, both warm-based and cold-based zones occur.

Zonation in glaciers

In this work, the following common terminology is used to designate different parts or zones of the glacier system:

Supraglacial zone: refers to the upper layers of the glacier affected by surface temperature variations where firnification¹ and surface ablation occurs. The supraglacial zone extends beyond the firn layers to the depth where ice reaches the density of glacier ice ($\approx 830 \text{ kg/m}^3$).

¹ Firnification designates the gradual transformation of snow into glacier ice by compaction, moisture diffusion, and recrystallization. Firn is the intermediate phase between snow and ice.

Basal or subglacial zone: the lower zone of glaciers affected by processes occurring at or near the ice-bedrock interface. The basal zone contains debris-rich ice which is produced at, and interacts with the glacier bed as a result of various mechanisms involving melting-refreezing, diffusion, and plastic deformation of ice.

Englacial zone: the bulk of the glacier mass, enclosed between the supraglacial and the subglacial zones. Englacial ice ("*glacier ice*") originates from firnification of snow. It is typically coarse, clean (debris-poor), contains abundant air bubbles, and is foliated on a large scale by polycrystalline deformation associated with ice flow.

Subglacial boundary conditions

Thermal boundary conditions at the ice-bedrock interface are critical to the dynamics of glacier movement as well as to processes of subglacial erosion and entrainment of debris. The main driving force of glacier movement is gravity which causes the ice mass to expand and flow by internal creep and shearing. However a large number of glaciers (mostly temperate, but also polar) exhibit ice flow velocities which cannot be entirely accounted for by internal deformation processes.

Basal sliding occurs when temperature at the bed is at or close to the freezing point and may account for most of the momentum in some temperate glaciers. Much of the resistance to glacier flow is due to friction at the ice-bed interface. The bed roughness of glaciers can be defined in terms of amplitude and wavelength for a series of successive obstacles (or "hummocks") and varies widely with the nature of the substrate. Depending on the scale of roughness, different mechanisms may operate to enhance or reduce basal ice flow rates. Ice flowing over bedrock obstacles of an amplitude higher than about 0.5 m is subjected to an increase in the longitudinal strain rate which results in enhanced flow velocities proportional to the size of the obstacles. This process, known as *enhanced plastic flow*, is increasingly effective with both amplitude and wavelength of the bed roughness.

Concurrently, increase of the stress field around bed hummocks may induce local pressure-melting if basal ice is close to the pressure-melting point. Meltwater thus produced flows along the hydraulic pressure gradient to the low pressure area on the lee side of the obstacles, where it refreezes. The simultaneous release of latent heat of fusion may contribute to further melting on the stoss face of the obstacles, provided heat can be efficiently conducted through it.

This process of mass transfer is known as *regelation sliding* (Weertman 1961) and is regarded as being of major importance beneath temperate glaciers. However, observations by Kamb and Lachapelle (1964) indicate that regelation may only be effective around obstacles of an amplitude lower than ≈ 1 m, as heat cannot be conducted through larger hummocks. Thus the scale of bed hummocks is a controlling factor in determining which process (enhanced plastic flow or regelation sliding) operates predominantly at the ice-bedrock interface. Additionally, in warm-based glaciers a thin film of pressurized subglacial water may enhance regelation sliding velocity by reducing the effective bed roughness.

Polythermal glaciers will exhibit zones of contrasting ice flow rates reflecting the variability of thermal conditions at the bed. Basal melting and rapid flow rates will occur beneath those parts of the glacier bed where heat supplied to the sole is greater than can be conducted away from it by the temperature gradient. Near the glacier margin, thinning of the ice may steepen the temperature gradient until it allows removal of the entire basal heat flux, thus creating a low-velocity zone of basal freezing (Fig. 1.2).

Subglacial erosion occurs by a combination of processes that are highly dependant on the mechanical properties of the substrate and on the thermodynamic behavior of basal ice. Abrasional wear is caused by debris transported in the basal zone in contact with the glacier bed. This debris may be derived from rockfalls in the accumulation zone or along

Fig. 1.2. Variable thermodynamic regime in a valley glacier. Conditions at the bed and at the surface are given for various parts of the glacier. Vertical temperature profiles in A and B are shown for different values of accumulation rate ($\dot{A}$), ablation rate ($\dot{A}_{bb}$), and longitudinal ice flow velocity (U). (from Drewry 1986)

the valley walls of the glacier, or may be directly removed from the underlying substrate by cryostatic load failure and plucking.

A high flow velocity near the ice-bedrock interface is conducive to higher erosion rates as it allows fractured debris to be evacuated from its resting site and dragged away in contact with the bed. Initial removal of subglacial debris may also be facilitated by the presence of pressurized meltwater at the sole acting as a hydraulic jack to displace fractured bedrock. These conditions are generally associated with temperate or sub-polar glaciers (hence their high erosive potential) although in cold-based glaciers, the intense shear created by differential flow rates in the lowermost ice layers may generate enough horizontal momentum in imbedded clasts to result in limited abrasion of the bed. Erosional effects associated with subglacial water flow are also constrained by thermal bed conditions, as the presence of a film of water at the glacier sole is limited to those zones where ice is at or near the pressure-melting point.

1.2) Subglacial debris entrainment

Shear planes

In an early study, Goldthwait (1951) suggested that marginal moraines of the Barnes Ice Cap (Baffin Island) may be derived by entrainment of debris from the glacier bed along thrust planes. This process would occur as active ice, flowing from the interior of the ice sheet, overrides a zone of stagnant marginal ice by thrusting along a system of active shear planes extending to the glacier bed (Fig. 1.3). Weertman (1961) criticized the hypothesis, arguing that shearing-in is unlikely to occur beneath a cold-based ice sheet, as the limited basal sliding would not allow efficient debris entrainment in the ice.

However, recent observations by Echelmeyer and Zhongxiang (1987) have shown that rapid motion of cold-based glaciers may result from internal shearing of a subglacial

Fig. 1.3. Glaciotectonic structural deformation of basal ice resulting from a change of thermal conditions at the glacier bed. A: Velocity profiles in a warm-based zone of basal sliding (right) and a cold-based zone of differential shearing (left). Dotted lines represent the position of debris layers. Deceleration of ice flow and increase of compressive strain at the bed thermal boundary may result in asymmetric folding (B) or thrusting (C) of basal ice layers. (From Drewry 1986)

bed of unconsolidated sediment. These conditions would allow limited basal debris incorporation by the propagation of shear planes into overlying glacier ice. Alternatively, impacts between particles transported along parallel pathways in a basal shear zone may cause them to gradually diffuse into overlying ice, but dispersal distances associated with this process were estimated to be no more than about one order of magnitude above the size of the particles themselves (Weertman 1968).

Thus, generally speaking, the agreement in the literature remains that while thrusting and shearing may contribute to debris entrainment and thickening of basal debris sequences, neither mechanism can properly account for the thick sequences of closely stacked debris layers commonly observed under glaciers.

Net freeze-on

A more accepted hypothesis initially proposed by Weertman (1961) involves large-scale debris accretion by direct freezing of water onto the glacier sole. This process may operate when the thermal gradient of the temperate or sub-polar glacier is locally enhanced in response to sustained changes of surface temperature or overlying ice thickness. To readjust to the new gradient, the vertical temperature distribution inside the glacier changes, causing the 0°C isotherm to migrate downward into the substrate. If the latter consists of a soft debris bed, the result will be the accretion of relatively undeformed layers of sediment onto the glacier sole that can subsequently be entrained forward by ice flow (Fig. 1.4).

If a steady supply of subglacial meltwater exists, repeated freeze-on events may lead to the formation of a relatively conformable basal debris sequence. Basal accretion by freezing-on is probably restricted to a relatively narrow marginal zone in glaciers, as the thermal conditions that favour its occurrence are associated with limited ice thickness.

Fig. 1.4. Weertman's (1961) hypothesis for debris accretion by net freeze-on at the bottom of an ice sheet experiencing non-steady state conditions. Migration of the 0°C isotherm from (a) to (b) causes subglacial water from an up-ice warm-based zone to freeze in the unconsolidated debris bed (c) and incorporate some of the debris into the overlying ice. Repetition of the cycle results in the accretion of an interlayered basal ice and debris sequence (d). (from Weertman 1961)

Regelation

Direct observations of basal sliding in temperate and sub-polar glaciers have led several authors (Weertman 1961, Kamb and Lachapelle 1964, and others) to emphasize the role of regelation sliding as a mechanism for incorporating fine subglacial debris into an ice layer. During regelation sliding, loose fine debris present at the ice-rock interface may become entrapped in new ice layers formed on the lee-side of bed hummocks; if the process is repeated over a series of successive obstacles, a basal sequence consisting of alternating clean and dirty ice laminae will result (Fig. 1.5).

Thickness of regelation layers is controlled by ice viscosity, roughness of the bed, and physical properties of the debris. Weertman (1961) estimated that regelation may only operate around small (0.5 to 1m high) obstacles, and that it is unlikely to form thick basal debris sequences, as individual regelation layers produced on the lee-side of an obstacle are likely to be destroyed on the stoss-side of the following one. Recent experimental modelling by Hubbard and Sharp (1993) has confirmed this.

Entrainment of debris by regelation may occur preferentially in subglacial cavities opening on the down-ice side of bed obstacles as a result of high flow velocities, a process referred to as *cavitation* (Lliboutry 1968). If such cavities remain open for a sufficient length of time, unconsolidated debris (mostly silt- to sand-size) may become entrapped within regelation layers a few mm to cm thick. Although widespread subglacial cavitation remains hypothetical, several observations have been reported that support its occurrence underneath temperate glaciers (Souchez *et al.* 1973, Tison and Lorrain 1987).

Detailed observations in Spitsbergen by Boulton (1970, 1972) led him to conclude that basal freezing processes, in particular regelation, account for the accretion of thick basal debris sequences in polar glaciers. Boulton also stressed that while thrusting may result in some subglacial debris entrainment (e.g. where ice flow over large bed obstacles raises the basal shear stress), it is chiefly responsible, with recumbent folding, for the

Fig. 1.5. Subglacial debris entrainment by regelation. Pressure-melting and subsequent refreezing of basal ice around a bed obstacle results in the incorporation of fine debris in newly accreted ice laminae. A decrease of amplitude in successive bed hummocks favours a net gain of regelation ice (graph), as the uppermost laminae are preserved from disruption at the glacier bed. (from Drewry 1986)

considerable thickening of the basal sequence that occurs in the marginal compressive ice flow zones. Based on these premises, Boulton suggested a controversial generalization² following which polar glaciers may generally contain larger amounts of subglacially derived debris than temperate glaciers, as cold-based conditions are favourable to basal accretion.

This contention was supported by Boulton's observations at Aavatsmarkbreen (Boulton 1970) where the upward sequence of basal debris layers reflects, in reverse order, the composition of the successive lithologies over which the glacier flows. Such conformable stacking may only be attained if the dominant process of subglacial debris entrainment is one of accretion by freezing-on at the sole. This also suggests that spatial compositional variations in till deposits derived from cold-based ice sheets may conceivably be linked to their source, even over areas of geologically contrasted substrate.

Further refinements to the basal freezing accretion concept were introduced by Robin (1976) who hypothesized the existence of cold-based zones below glaciers that are predominantly at the pressure-melting point as a result of rapid changes of pressure occurring when moving ice encounters a bedrock irregularity. The Robin effect occurs as ice flowing on the stoss side of the obstacle undergoes pressure-melting due to the local increase of the stress field. If some of the expelled meltwater is lost in intergranular veins, refreezing on the lee-side of the hummock cannot restore enough latent heat to allow surrounding ice to adjust to the new pressure-melting point. As a result, a cold-based zone extending laterally for several metres may form where any other meltwater locally available can freeze onto the glacier sole. This effect may also be associated with rapid changes in the basal stress distribution in response to rapid subglacial water pressure variations.

In his classic study of Matanuska Glacier (Alaska), Lawson (1979) proposed a facies system for identifying distinct stratigraphic zones encountered in glaciers (Fig.1.6).

²Cf correspondence between G.S. Boulton and J.T. Andrews, *J.Glacio.* 10, 1971.

Fig. 1.6. The ice facies system of Lawson (1979). Relative thickness shown for basal ice facies are those of Matanuska Glacier (Alaska).

Lawson's system includes two major subdivisions of the basal zone: the lower *stratified* facies, characterized by a high debris to ice ratio and complex layering structures, which grades into the overlying massive *dispersed* facies (often referred to as *clotted ice* facies) where debris content is comparatively lower and uniformly distributed. Based on morphological and isotopic criteria, Lawson suggested that the stratified facies may result mostly from a net freezing-on process occurring in the marginal part of the glacier with possible input of supraglacially-derived meltwater at the sole. The upper dispersed facies is more problematic, as it presents characteristics compatible with both subglacial or supraglacial processes. Despite these uncertainties and the highly variable aspect of the basal zone between different glaciers, Lawson's system has been widely adopted by researchers as it provides a useful common reference.

Overriding of ice and debris

Incorporation of a frontal ice and debris apron may account for some basal debris entrainment in the marginal regions of glaciers. Hooke³ suggested that marginal ice-cored moraines may develop from the combination of compressive marginal ice flow coupled with ice movement over snow and superimposed ice accumulated at the glacier margin. Based on observations in Antarctica, Shaw (1977) argued that frontal apron incorporation may be an important mechanism for subglacial debris entrainment in polar glaciers where low meltwater availability and the presence of deeply frozen substrate are unfavourable to subglacial accretion processes. Souchez *et al.* (1973) and Tison and Lorrain (1987) have reported the incorporation of snow and ice coatings in subglacial cavities as folded layers on to the base of Switzerland glaciers. Boulton (1975, 1983) stressed the importance of plastic deformation of basal ice during overriding of bed obstacles and debris in the genesis of subglacial bedforms and the translocation of basal debris to higher positions in the basal

³Hooke, 1973: cited in Lawson (1979).

ice sequence. Overriding and incorporation of subglacial meltwater channel deposits may also account for debris incorporation in localised parts of the glacier bed.

Glaciotectonic thrusting

Permafrost beneath the ice margin of a glacier underlain by soft sediments is conducive to glaciotectonic thrusting and shearing. When meltwater produced under the warm-based inland regions of a polythermal glacier migrates to the outer frozen fringe, the resulting hydrostatic pressure build-up in the substrate may reduce its shear strength and allow the moving ice to incorporate large unfrozen rafts of material. Glaciotectonic structures usually disrupt the conformable sequence of previously accreted englacial debris and, when coupled with compressive marginal flow, may result in the formation of imbricate thrust plates (Eyles 1977, Klassen 1982a). Overriding of pre-existing ice and glaciotectonic thrusting are likely to be more effective underneath the net ablation zone, where thermal balance and the occurrence of cold-based ice favors thrusting or basal freezing (*re* Fig. 1.3).

1.3) Stable isotope studies of basal ice

The use of stable oxygen and hydrogen isotopes in glaciology is based on kinetic effects that result in fractionation of heavy (^{18}O and ^2H) and light (^{16}O and ^1H) isotopes during phase transformations of water. In snow, firn and glacier ice, the ^{18}O - ^2H relation is essentially dictated by the equilibrium condensation of water vapour, during which ^{18}O and ^2H become relatively enriched in the liquid phase as a result of their lower vapour pressure.

Craig (1961) empirically established a linear equation which correlates the isotopic composition of precipitation (given in δ values⁴) over the range of variation in natural fresh waters:

$$\delta D = 8\delta^{18}O + 10 \text{ ‰} \quad (\text{Craig 1961})$$

This relation is known as the *precipitation effect* and the line obtained from the previous equation is the *Global Meteoric Water Line*. In a given geographic region, the local *Precipitation Line (PpL)* is expressed by the generalized relation $\delta D = b\delta^{18}O + a$, where b and a are controlled by evaporation-condensation effects and the isotopic composition of the local moisture source, respectively. Divergence from the precipitation line may characterize waters affected by intense evaporation relative to atmospheric input. This effect also applies to snow if intense sublimation occurs.

While the $\delta^{18}O$ - δD relation in snow essentially conforms to the precipitation effect, the actual ^{18}O and D concentrations in snow are dependent on the temperature at which condensation occurs: the greater the fall in temperature, the lower will the heavy isotope concentrations be relative to the moisture source (i.e increasingly negative δ values). This relationship is linear and its parameters are geographically variable (Picciotto *et al.* 1960). Heavy isotope depletion in precipitation can also be attributed to the *altitude effect*, which results from adiabatic cooling of the precipitating air mass as it moves to higher elevations. Altitude-related ^{18}O -depletion has been reported to range from -0.2 ‰ per 100 m in the

⁴Concentration of the ^{18}O isotope relative to ^{16}O for a given sample is commonly given by its $\delta^{18}O$ value, expressed in per mille (‰), representing departure from the universal water standard SMOW (Standard Mean Ocean Water) following the relation:

$$\delta^{18}O = \frac{(^{18}O/^{16}O)_{\text{sample}} - (^{18}O/^{16}O)_{\text{SMOW}}}{(^{18}O/^{16}O)_{\text{SMOW}}} \times 10^3 \text{ ‰}$$

The heavy hydrogen isotope 2H (= deuterium D) is quantified in a similar way with respect to 1H , and its relative concentration is expressed as δD relative to SMOW.

western Alps (Ambach *et al.* 1968) to -0.9 ‰ per 100 m in Queen Maud Land, Antarctica (Gonfiantini *et al.* 1962).

$\delta^{18}\text{O}$ variations in basal ice

Stable isotopic methods were applied by Gow *et al.* (1979) to characterize basal layers of the Byrd Station deep ice core from West Antarctica. Results showed that the mean $\delta^{18}\text{O}$ for basal ice is comparable to that of clean, air-rich glacier ice, while debris-bearing ice layers display typically lower $\delta^{18}\text{O}$ values with variations of a much higher magnitude. Subglacial meltwater extracted from the hole was also found to be ^{18}O -depleted by about 3 ‰ with respect to the average value of the basal ice. Gow *et al.* interpreted these results as indicative of basal melting-refreezing, the ^{18}O depletion of the meltwater being similar to that reported by O'Neil (1968) for oxygen isotope fractionation between ice and fresh water during equilibrium freezing.

The diffusion coefficients of HDO and H_2^{18}O molecules in ice being very low (in the order of $10^{-11} \text{ cm}^2\text{s}^{-1}$), melting of compact ice is likely to occur without isotopic change. However, isotopically homogeneous liquid water is characterized by comparatively higher diffusion coefficients (in the order of $10^{-5} \text{ cm}^2\text{s}^{-1}$). As a result, freezing of water involves isotopic fractionation as a consequence of different water molecules freezing at slightly different temperatures.

If the freezing rate is sufficiently slow to allow for a Rayleigh-type equilibrium process to apply, ice layers formed by the progressive freezing of water will experience a relative enrichment of ^{18}O and D, while a parallel depletion in heavy isotopes will occur in the remaining meltwater fraction. Therefore, the isotopically light basal layers observed by Gow *et al.* in Antarctica may result from the refreezing of meltwater previously depleted in ^{18}O as a result of successive previous freezing events.

Lawson and Kulla (1978) reported contrasting high $\delta^{18}\text{O}$ values (relative to glacier ice) for presumed regelation ice in the basal zone of Matanuska Glacier (Alaska) which they interpreted as resulting from the basal freezing-on of isotopically heavier meltwater originating from the supraglacial zone near the equilibrium line of the glacier. On theoretical grounds, Boulton (1983) suggested that ^{18}O enrichment in regelation ice may also reflect the varying proximity of the parent meltwater source. Depending on the distance traveled by the water, ice formed from partial refreezing will either show a net enrichment in ^{18}O as a result of fractionation during freeze-up (close to water source), or a net depletion because the remaining water available has been gradually impoverished in its heavy component during transport (away from water source).

Co-isotopic method: the Souchez-Jouzel model

Using theoretical and empirical data, Souchez and Jouzel (1982, 1984) have introduced a new isotopic method for genetic characterization of basal ice whereby accreted ice (originating from regelation or net freeze-on) can be distinguished from englacial ice by its characteristic co-isotopic ($\delta^{18}\text{O}$ - δD) composition. The method applies the fact that ice having undergone melting-refreezing is characterized by a different $\delta^{18}\text{O}$: δD relationship than ice derived by firnification in which the stable isotope composition is dictated by the precipitation effect.

When englacial ice and refrozen ice samples are plotted on a $\delta^{18}\text{O}$ - δD diagram, the former will lie along a straight line of slope ≈ 8 (corresponding to the local precipitation line), while the latter will lie on a *freezing line* of slope less than 8, its precise value depending on the initial isotopic composition of the parent water admitted to freeze. The initial composition of the water source from which glacier basal ice layers are derived may be determined by calculating the intercept of the freezing line with the line giving the $\delta^{18}\text{O}$: δD relation of englacial ice.

In a closed system situation where no mixing of input sources occur, successive ice laminae formed by the progressive freezing of a water reservoir will initially display heavy isotope (^{18}O -D) enrichment relative to the initial water composition. This trend will be reversed after about 66% of the water volume has undergone freezing, as a result of the remaining water supply being progressively impoverished in its heavy components by fractionation (Fig. 1.7). Therefore, ice laminae formed in the later stages of the freezing process will display lighter $\delta^{18}\text{O}$ and δD values relative to that of the initial water source.

However, isotopic fractionation effects in basal regelation ice may not be detected if the scale of sampling is in the same order as that of the total ice layer thickness, as the bulk isotopic composition of the layer will be similar to that of the water source. Hubbard and Sharp (1993) have recently demonstrated by computer simulation that isotopic composition of regelation ice laminae evolves down-glacier in such a way that both the basal ice layer and the net basal meltwater film display large-scale fractionation effects with the former being consistently less depleted in heavy isotopes than the latter. The magnitude of this fractionation has been estimated in the order of 3‰ for ^{18}O , which is similar to that previously proposed by O'Neil (1968). Hubbard and Sharp's conclusions stress the fact that both morphology and isotopic composition of regelation ice layers formed underneath temperate glaciers can be extremely variable between adjacent laminae as a result of multiple refreezing events at the bed. This contributes to considerable scatter effects in isotopic characterization of the ice, as samples recovered by traditional methods may be composed of several such laminae.

Other developments have been proposed by Boulton and Spring (1986) and Souchez and DeGroote (1985) to account for the effects of a mixed water input situation (such as seeping of lake water into the basal zone of a glacier), and by Souchez *et al.* (1987) and Hubbard (1991) regarding the effect of freezing rate variations. These will be considered later.

Fig. 1.7. Isotopic evolution of ice formed by the progressive equilibrium freezing of a water source of initial composition equal to SMOW. The $\delta^{18}\text{O}$ - δD composition of ice is shown for ten successive frozen fractions of the water reservoir. Only ice layers formed after 66% of the initial water volume has frozen are ^{18}O - and D-depleted relative to SMOW. (from Jouzel and Souchez 1982)

1.4) Summary

1) Field observations suggest that processes involving basal ice accretion by freezing of water at the glacier sole may be responsible for the initial entrainment of most of the debris load in the basal zone of polythermal glaciers. Other processes such as overriding of stagnant ice and debris, glaciotectionic thrusting, and upward shearing-in are generally limited to the marginal compressive flow zone and cannot account for the total volume of debris in transport within the basal ice layers.

2) Basal ice accretion can occur by a wide range of mechanisms. *Regelation* is generally recognized as an important process for basal ice flow as well as subglacial debris entrainment in temperate and sub-polar glaciers. Robin's *heat pump effect* can allow for regelation to occur underneath glaciers that are predominantly warm-based. In polythermal glaciers, subglacial water derived from basal melting in warm-bed zones may lead to large-scale accretion of debris by *net freeze-on*. Supraglacial or marginal water infiltrations to the glacier sole may contribute strongly to basal ice formation.

3) Theoretical and experimental considerations both indicate that regelation is unlikely to produce basal ice layer sequences of a thickness comparable to those observed in the field. In fact, the thickening of the basal ice zone results largely from structural deformation such as plastic deformation and thrusting which causes overstacking of basal ice layers after initial entrainment of the debris by subglacial accretion processes.

Upward dispersion of clasts within the ice matrix as a result of shear deformation also contributes to this effect, although to a much lesser extent.

4) Most observers recognize different ice types within the basal zone of glaciers. The descriptive facies system of Lawson subdivides the basal zone between a lower, highly deformed and debris-rich *stratified* facies, and an overlying thicker and massive *dispersed* facies with an overall lower debris content. Whereas researchers generally agree on a basal

accretion origin for ice in the stratified facies, the origin of the dispersed facies remains problematic, as its characteristics are indicative of both subglacial and supraglacial processes.

5) In some instances, subglacial entrainment of debris by basal accretion leads to the formation of an ice/debris sequence which may compositionally reflect in a conformable manner the successive lithologies over which the glacier flows, the uppermost debris bands being related to substrate units located further up-ice along the flow lines, and vice-versa. This sequence may not be preserved over long distances, as the basal ice zone is submitted to intense shear stress and undergoes structural deformation by folding, shearing and thrusting, particularly near the compressive margin. Moderate ice flow rates and smooth bed topography are more likely to favor the preservation of a conformable basal sequence in glaciers.

6) The co-isotopic $\delta^{18}\text{O}$ - δD method developed by Souchez and Jouzel is employed to distinguish between accreted basal ice layers and glacier ice derived by firnification. When plotted together on a $\delta^{18}\text{O}$: δD plot, glacier ice samples lie close to the local meteoric water line (slope ≈ 8), whereas refrozen ice samples are aligned on a regression freezing line with a slope < 8 reflecting isotopic fractionation effects of freezing. The method has proven applicable for both open and closed system situations, but its results are highly dependant on the scale of ice sampling, as isotopic fractionation due to melting-refreezing effects may not be detectable in ice layers several cm thick.

Chapter 2

Presentation of the Study Area

Bylot Island (area: 11 067 km²) is located on the eastern margin of the Canadian Arctic Archipelago between latitudes 72° and 74 N°, and longitudes 75° and 82° W. The island is nestled on the northeast corner of Baffin Island at the entrance to Lancaster Sound (Fig. 2.1). To the south, Bylot Island faces Baffin Island and Borden Peninsula across Eclipse Sound. The island is easily accessible via scheduled flights to the settlement of Pond Inlet on northern Baffin Island.

2.1) Physiography and geomorphology

The central part of Bylot Island is occupied by the northwest-trending Byam Martin Mountains with peaks averaging an altitude of 1400 m a.s.l. This range is part of the Davis Highland, a mountain belt associated with the Hudsonian orogeny (~ 1.8 Ga ago) that stretches over 1500 km from southeastern Baffin Island to central Ellesmere Island. Along the western and southwestern coasts of Bylot Island, smooth, gently rolling hills extend from the piedmont to the shores of Eclipse Sound and Navy Board Inlet. The Bylot Lowlands are an extension of the Lancaster Plateau, itself a part of the Arctic Lowlands (Klassen and Shilts 1987).

The Davis Highland presents a typical alpine setting with high summits and plateaus dissected by major valleys and inlets. In the Byam Martin Mountains, sharp peaks and horns extending to 1800 m a.s.l. are separated by deep cirque basins with narrow walls and saw-tooth crests. The central part of the mountains is occupied by an icefield enclosing several major drainage basins from which ice flows outward to coalesce into large

(Figure provided by B.Moorman)

composite outlet glaciers, two of which calve in Lancaster Sound. Facing southwest, several major glaciers extend in piedmont lobes over the Bylot Lowlands.

Summits of the piedmont area are broad, rounded, with commonly flat hilltops that may correspond to the ancient erosional surface of the Baffin Upland to the south. An extensive cover of frost-shattered rubble drapes this landscape. The lowlands along Eclipse Sound present a contrasting smooth hummocky topography with numerous small kettle lakes. Periglacial features associated with permafrost conditions such as patterned ground, solifluction lobes and mudboils are widespread.

The area selected for this investigation (Fig. 2.2) encloses three major glaciers flowing to the southwest and extending in piedmont lobes across the foothills and onto the Bylot Lowlands. From east to west, those glaciers are Sermilik (B-37), Aktineq (B-17) and Ujarasukjuit (B-7)¹: all three have their source in the same major ice drainage basin. Numerous smaller tributary and outlet glaciers such as B-26 and B-28 flow between them.

2.2) Bedrock geology

The bedrock geology of Bylot Island and adjacent northern Baffin Island has been mapped at the 1: 250 000 scale by Jackson and Davidson (1975) and Jackson *et al.* (1975). Chronology and setting of Precambrian metamorphism on Baffin Island and Bylot Island was studied by Jackson and Morgan (1978). Miall *et al.* (1980) established the stratigraphy of Cretaceous-Tertiary sedimentary strata from Eclipse Trough. Recently, Jackson and Sangster (1987) presented an assessment of the natural resource potential of Bylot Island and northwest Baffin Island.

¹ The glaciers are identified in accordance with official toponymic designations (NTS map sheets) and with the Glacier Atlas of Canada, Bylot Island Area 46201 (Inland Waters Branch, 1969). An exception is made for glacier B-7 for which the local name *Ujarasukjuit* although unofficial, is employed.

Regional setting

Bylot Island is located at the junction between the Churchill province of the Canadian Shield and the eastern reaches of the Arctic Platform. The Canadian Shield terranes mostly include igneous and high-grade metamorphic crystalline rocks of Archean-Aphebian age that outcrop in most of the Byam Martin Mountains. Along parts of western and northern Bylot Island, this complex is overlain by a succession of unmetamorphosed, relatively undeformed Proterozoic (Neohelikian) sedimentary and volcanic rocks. The Bylot Lowlands are essentially underlain by Arctic Platform terranes, with the exception of the westernmost area where Proterozoic Shield rocks are exposed. On southern Bylot Island, the Arctic Platform bedrock is composed of flat-lying clastic, poorly-consolidated sedimentary rocks of Cretaceous-Tertiary age.

Neohelikian, Paleozoic, and Cretaceous-Tertiary strata on northern Baffin Island and Bylot Island are preserved in a series of parallel, northwest-trending grabens separated by horsts of crystalline rocks. The bounding faults dip steeply toward the grabens and probably originated from an extension phase during crustal attenuation and sea-floor spreading associated with the mid-Cretaceous to mid-Eocene Eurekan rifting episode (Miall *et al.* 1980). The Cretaceous-Tertiary sequence of Eclipse Trough that underlies most of the Bylot Lowlands was deposited in one such fault-controlled basin. The trough is bounded to the north and south by the Navy Board and Byam Martin Precambrian highlands. On Bylot Island, the faulted contact between Archean-Aphebian rocks in the core of the island and the Cretaceous-Tertiary sedimentary strata coincides approximately with the topographic transition between the Byam Martin Mountains and the Bylot Lowlands (Fig. 2.3).

Note: To simplify terminology, the Archean-Aphebian and Cretaceous-Tertiary terranes of Bylot Island will be referred hereafter by the abbreviations *AA* and *KT* respectively.

Fig. 2.3. Simplified geological map of southwestern Bylot Island, NWT.

Compiled from Jackson and Davidson (1975) and Miall, Balkwill and Hopkins (1980).

Archean-Aphebian (AA) crystalline basement

Shield lithologies in the study area consist mostly of banded granitic to granodioritic migmatites. These were derived from granitic rocks initially emplaced in Archean time (about 2.8 Ga ago) and subsequently metamorphosed to upper granulite or amphibolite facies during tectonic activity associated with the Hudsonian orogeny 1900 to 1600 Ma ago (Jackson and Morgan 1978). On southern Bylot Island the hypersthene isograd which marks the upper amphibolite-granulite facies transition runs approximately parallel to the southern fringe of the uplands.

On the western side of the Byam Martin Mountains, Jackson and Davidson (1975) reported large areas of metasedimentary and metavolcanic rocks within the banded migmatitic complex. These units occur as dispersed narrow southeast-striking belts up to several tens of km long and striking southeast. The volcano-sedimentary units are most likely related to the Archean supracrustal sequence of the Mary River Group which hosts rich iron ore deposits on north-central Baffin Island (Jackson and Sangster 1987). Several such belts have been mapped in the study area and underlie part of Sermilik, Aktineq and Ujarasukjuik Glaciers. Whether these belts host iron formations is not known precisely.

The northernmost sector of the study area is underlain by part of the Bylot Batholith, a 2500 km² pluton of relatively uniform monzocharnockitic (hypersthene-quartz monzonite) composition. Jackson and Morgan (1978) established that the batholith was metamorphosed to the granulite facies in late Aphebian time, but its emplacement age remains unknown.

Cretaceous-Tertiary (KT) sedimentary beds

The *KT* sedimentary strata exposed in the Bylot Lowlands were interpreted by Jackson *et al.* (1975) as a molasse sequence deposited in a shallow lagoonal or littoral environment in response to uplift during fault reactivation in the North Baffin Rift Zone. Miall *et al.* (1980) have subsequently divided the sedimentary sequence of Eclipse Trough in three formations, two of which, the Kanguk and Eureka Formations, are exposed in the study area.

The Kanguk Fm and the lower member of the Eureka Sound Fm are exposed along a narrow (maximum width: 5 km) stretch of land trending southeast and bounded to the north by the faulted-contact with the AA basement rocks. The Kanguk Formation consists of lower grey mudstone member overlain by an upper white to orange-brown, medium- to coarse-grained sandstone member. The upper member of the Kanguk Fm and the lower overlying lower mudstone member of the Eureka Sound Fm are exposed in a small area between Aktineq Glacier and the south coast near Sermilik Glacier. Their absence northwest of Aktineq Glacier are probably the result of a lateral facies change or an unconformity.

The Eureka Sound Fm lower member consists of dark grey, variably silty mudstones containing siderite concretions and thin lenses of soft sand and carbonate-cemented sandstone. The upper sandstone member is exposed over large areas of the southern Bylot Lowlands, except where it is overlain by glacial drift or outwash deposits. Lithologically, the sandstones are typically white, orange or brown, massive and structureless, although occasional silty and argillaceous interbeds occur which provide some evidence of bedding.

The abundance of fresh and highly angular granitic fragments in the sandstones indicates a first-cycle origin, while their petrography points at a protolith of quartz monzonitic, gneissic or granitic composition. A paleogeographical reconstruction proposed by Miall *et al.* (1980) suggests that most of the detrital material found in Eclipse Trough lithologies was derived from the surrounding Byam Martin, Navy Board and Baffin Uplands. Dark grey to black carbonaceous beds, plant remains, and chips of dark shale and coal occur throughout the *KT* sedimentary sequence, but the only significant thickness is found in the lowermost Cretaceous Hassel Formation which is hardly exposed in the study area (not shown on Fig 2.3).

2.3) Surficial geology and glacial history.

Hodgson and Haselton (1974) provided a description of marginal and recessional moraines on Bylot Island as part of a reconnaissance study of the surficial geology of northeastern Baffin Island. This formed the background of a detailed investigation of the Quaternary geology of Bylot Island by Klassen (1982b, 1985, 1993). Aspects of the surficial geology and glacial history of Bylot Island presented in this section are mostly based on Klassen's interpretation.

Surficial deposits

The modern surficial geology of Bylot Island is a product of previous glacial episodes and of the severe arctic climate that has prevailed on the island in recent time. The Quaternary glacial record reveals several periods of glaciation by regional ("foreign") ice sheets that flowed onto Bylot Island across the deep marine trenches which surround it, and by local ("native") mountain glaciers or ice caps. Frost-shattering and mass-wasting

associated with periglacial conditions have produced abundant scree and rubble which is sometimes difficult to differentiate from Quaternary glacial deposits.

Klassen (1993) mapped the surficial geology of Bylot Island at the 1:250 000 scale by air photo interpretation and ground truth evidence. On a large part of the ice-free coastal areas of Bylot Island, surficial deposits consist of glacial drift associated with either native or foreign glaciations. The distinction between native and foreign drift units was chiefly established on the basis of drift provenance as revealed by the dispersal of coarse erratics. Carbonate rocks of Proterozoic (nonfossiliferous) and Paleozoic (fossiliferous) age derived from the uplands of Borden Peninsula and Brodeur Peninsula are associated with northward foreign ice advances onto Bylot Island. Erratics of this type are useful indicators of drift provenance as they are easily detectable in areas of crystalline terrane where they stand out in sharp colour contrast as a result of their alkalinity inhibiting vegetation growth. Poorly-consolidated *KT* sediments of the Bylot Lowlands could not be used to this end because they do not form cobble- or boulder-size erratics.

In the study area (Fig. 2.4), foreign drift is mostly associated with the mid-Wisconsinan *Eclipse* glaciation and consists of sediment transported by foreign ice sheets and deposited as a supraglacial debris veneer or as a grey to brown sandy diamicton with a variable (1 to >80 weight %) content of foreign debris. The foreign drift cover is thin, often discontinuous, and varies texturally from a thick muddy diamicton to a boulder gravel. On the southeastern coast of Bylot Island, Eclipse drift is found up to 300 and 600 m a.s.l. across coastal mountain slopes and up to 50 m to 250 m a.s.l. within valleys. Further inland and at higher elevations, some scattered foreign erratics occur as well. Two of the major glaciers within the study area (Sermilik and Aktineq) have their modern ice margin

overriding the inland limit of Eclipse drift. Near Aktineq Glacier, the extent of foreign drift is marked by a morainic bench across a nearby hillside to the east of the glacier.

Native drift on the Bylot Lowlands forms a discontinuous cover extending from the foothills of the Byam Martin Mountains to approximately 3 km south of the known inland limit of Eclipse drift. In the piedmont area of southwestern Bylot Island, native drift deposits are chiefly associated with Neoglaciation², or with an older readvance episode that occurred during retreat of Eclipse Ice. Native drift chiefly consists of a veneer of crystalline erratics overlying disaggregated KT bedrock regolith. Thick hummocky accumulations of ice-proximal native drift are found only within two broad valleys, one of which is occupied by Ujarasukjuit Glacier. The hummocks are the product of stagnant melting of local valley glaciers. Crystalline erratics associated with pre-Neoglacial native drift are lichen-covered and show signs of weathering due to prolonged exposure while Neoglacial and modern deposits are angular, unvegetated and unweathered.

Chronology of glacial events

Because of the critical position it occupied beyond the northeastern margin of the Late Wisconsinan Laurentide Ice Sheet, Bylot Island has undergone a complex glacial history (Fig. 2.4). Andrews and Barry (1978)³ suggested that inception of the Laurentide Ice Sheet growth may have been centered on the Davis Highlands. North of Bylot Island, the Lancaster Sound channel was a major pathway for ice flowing eastwards into Baffin Bay.

² *Neoglaciation* is employed to designate a worldwide readvance of mountain glaciers which occurred in the late Holocene. In western Europe and North America, Neoglaciation is often considered synchronous with the Little Ice Age (LIA) interval between the 15th and 19th centuries (Bradley 1985).

³ Cited in Klassen (1993).

Surficial sediment cover is discontinuous over most of Bylot Island and useful stratigraphic sections are restricted to coastal areas where they mostly consist of glacial marine sediments. Erosional bedforms are also rare, suggesting that during major glaciation episodes on Bylot Island, ice was cold-based and non-erosive. Using lithological contrasts in the regional bedrock to distinguish between foreign and native drift units, Klassen established a stratigraphy of glacial deposits on Bylot Island. Stratigraphic sections were subdivided with reference to organic horizons assumed to represent interglacial conditions.

Foreign glaciations

Klassen recognized at least three distinct foreign glaciations that directly affected Bylot Island. The oldest and most extensive is designated as *Baffin* glaciation, and may represent one or more pre-Eclipse regional events. Scattered foreign erratics found at higher elevations than that of the maximum inland extent of continuous foreign drift cover are believed to be associated with Baffin glaciation. Their general distribution suggests that foreign ice during Baffin glaciation may have been thick and extensive enough to flow across the topographic divide of the Byam Martin Mountains.

An extensive moraine system adjacent to Lancaster Sound and Pond Inlet marks the upper limit of foreign ice against the coastal mountains during Eclipse glaciation. Eclipse Moraine is best defined along the northeastern and southeastern sides of Bylot Island where it lies along coastal cliffs and crosses inland valley floors at low elevations. The moraine is composed almost entirely of foreign debris including most bedrock lithologies represented in the northern Baffin region, with carbonate sedimentary rock being a major component. Based on radiocarbon evidence, Eclipse glaciation is believed to have occurred over 43 000 years ago, in pre-Late Wisconsinan times (Klassen 1985).

Subsequent to Eclipse glaciation, *Button* glaciation appears to have been much less extensive. Stratigraphic evidence for Button glaciation is very limited and may represent one or more glacial events subsequent to Eclipse glaciation. Because of the rarity of Button drift, the extent of ice during Button glaciation is mostly speculative, and was probably restricted to the outer margins of the island below 140 m a.s.l.

Native glaciations

Three distinct periods of native glaciation have been identified on Bylot Island based on the geological relations of native drift units with deposits and landforms of Eclipse glaciation. *Bylot* glaciation is the oldest and most extensive and may include several distinct glacial events of pre-Eclipse age. Widespread distribution of crystalline erratics on the outer Byam Martin Mountains and the lowlands, as well as deeply incised meltwater channels occurring at high elevations suggest that ice thickness during Bylot glaciation was considerable and probably covered most or all of the island.

In front of Aktineq Glacier, landforms define a major advance of native ice postdating Eclipse glaciation. However, evidence for this native glacial advance is restricted to a few glaciers. Klassen (1993) hypothesized that this asynchronous advance may have resulted from a deglacial readjustment of outward ice flow in native glaciers that were dammed on the island during Eclipse glaciation. This episode is designated as the *Aktineq phase* and is interpreted as a late glacial (deglacial) phase of Eclipse glaciation.

McCuaig (1994) recently dated native erratics exposed near the margins of large glaciers on southwestern Bylot Island using cosmogenic ^{10}Be . Consistent exposure ages around 60 Ka were obtained, suggesting a rapid, almost simultaneous retreat of Eclipse Ice

and native ice which most likely corresponds to the end of the Aktineq deglacial phase as originally hypothesized by Klassen.

Neoglaciation

Several glaciers on Bylot Island are surrounded by marginal moraines that lie within 100 to 1000 m of their modern termini. These moraines are considered to outline the maximum extent of ice during a *Neoglacial* advance of native glaciers. Neoglacial moraines are often ice-cored and are typically composed of angular crystalline boulders, unweathered and unvegetated in contrast to older (underlying) native and foreign deposits. However, some glaciers terminate directly on older (pre-Neoglacial) sediments, indicating they are now at or beyond to their Neoglacial maximum positions.

An approximate timing of Neoglacial advance on Bylot Island is given by radiocarbon dating of organic sediments in debris underlying the marginal moraines of some major glaciers. DiLabio and Shilts (1978) reported ages of 450 ± 70 ^{14}C years B.P. (GSC-2597) and 7860 ± 100 ^{14}C years B.P. (GSC-2541) for two peat layers underlying lichen-free morainic material on the southeast margin of Aktineq Glacier. The first date was obtained from the uppermost part of the section; the second from a position 2 m below the first. These dates suggest that Aktineq Glacier advanced to its present position within the last 400-500 years and is as far advanced as it has been in the last 7000 to 8000 years (DiLabio and Shilts 1978).

A sample of organic-rich material thought to represent incorporated lake sediment was collected from the supraglacial surface of Aktineq Glacier and analyzed by Bouchard and Shilts (1991). The sample gave an essentially modern date of 130 ± 50 ^{14}C years B.P. (uncorrected); indicating that the source of the sediment, an ice-marginal lake located 6.5

km upstream, may have been overridden by ice very recently. An unpublished older date of 460 ± 120 ^{14}C years B.P. (GSC-5189, uncorrected) was also obtained by Bouchard and Shilts from in-ice organic debris collected about 1 km up-ice from the snout of Sermilik Glacier.

Frozen peaty sediment collected by this author from a subglacial section exposed on the western margin of Ujarasukjuit (B-7) Glacier was dated at 130 ± 50 ^{14}C years B.P. (GSC-5354, uncorrected). This is consistent with a similar date of 120 ± 80 ^{14}C years B.P. (GSC-3227) obtained by Klassen (1982a) for twigs collected from the uppermost bed of recent glaciotectionic thrust plates exposed in front of Glacier C-79 (west of glacier B-7). A comparable date of 120 ± 120 ^{14}C years B.P. (GSC-2529, uncorrected) was also reported by DiLabio and Shilts (1978) for peat-covered clasts of outwash that melted out of the ice 30 m above the bed of glacier B-7.

Despite the uncertainty associated with modern or uncorrected radiocarbon dates, the consistency of ages cited above suggests that the Neoglacial advance of native glaciers on the southwestern coast of Bylot Island has taken place within the last 500 years, probably more recently.

2.4) Modern glacial environment

Modern ice on Bylot Island covers about 4900 km² (nearly half of the island area) and forms a continuous icefield in the Byam Martin Mountains. The northwest-southeast trending topographic divide separates ice drainage into northwards and southwards flow components. Interconnected cirques define five major basins from which modern ice flows to coalesce into large composite outlet glaciers. On northern Bylot Island, two glaciers calve directly in Lancaster Sound.

Regional glaciation limits

Modern glaciation limits (*GL*) and lower equilibrium line altitudes (*ELA*)⁴ were estimated by Andrews and Miller (1972) for the Baffin Island region. Their work indicates that while *GL* and *ELA* respectively average 700 m and 400 m along most of the northern Baffin Island coast, an enclosed high *GL* contour reaching 1100 m exists over the Bylot Island icefield and adjacent Pond Inlet area (Baffin Is.) with associated *ELA* contours of up to 800 m. These high limits may result from local precipitation deficiencies (i.e. low accumulation rates) associated with the relatively high altitude and continentality of the icefields compared with glaciers of the same latitude occurring at lower elevations.

Recent glacial fluctuations

Most glaciers on Bylot Island have their modern margins either stationary at, or retreating from, positions attained during Neoglacial maximum. Using historical documents Falconer (1962) established that some ice marginal recession took place in the late 19th century. Comparison of 1948, 1958, and 1961 air photographs also showed that while glacier margins on the northern coast of Bylot Island had been almost stationary for the previous few decades, some large glaciers on the southeastern coast had recently undergone slight recessions of up to 90 m. Falconer could not extend his conclusions to the southwestern coast glaciers, due to the lack of records.

Klassen and Shilts (1987) reassessed the situation based on field observations and concluded that marginal retreat of several hundreds of metres had occurred for numerous

⁴ *Glaciation limits* of a given locality designate the lowest elevation at which glaciers can accumulate. The *equilibrium line* is the imaginary boundary on the glacier surface which demarcates the net accumulation zone (up-ice) from the net ablation zone (down-ice).

glaciers since Falconer's work. As well, several small ice caps in the area recorded on topographic maps were found to have completely melted since the photographic base was prepared in the early 1960's. A similar situation exists on north-central Baffin Island where small ice caps (e.g. Tiger Ice Cap) are presently retreating from Neoglacial maximum positions.

Comparison of 1982 air photographs with those of 1958-1961 shows that ice recession seems more accentuated on the southern coast of Bylot Island, although a few large glaciers (e.g. glaciers D-91 and E-131) on the north coast have also retreated from their 1958 positions. Recession on the southern coast is not uniform either. Some glaciers which formerly extended to Eclipse Sound indicate recent frontal retreat of up to several hundred meters (e.g. Sermilik Glacier) while other glaciers seem unaffected (e.g. Ujarasukjuik Glacier).

Climate

The arctic climate that prevails over Bylot Island and northern Baffin Island is characterized by cold temperatures and low annual precipitation. Klassen and Shilts (1987)

give -27°C and -35°C as maximum and minimum values for daily average temperatures during the winter months (November to April) in the Bylot Island area. The Canadian Climate Normals (1982) give similar figures for Pond Inlet⁵ (northern Baffin Island), with a daily minimum winter temperature of -31.8°C and a maximum of -24.0°C . Summer daily temperature averages are above freezing for June to August only and usually range between $+0^{\circ}\text{C}$ and $+5^{\circ}\text{C}$ (although temperatures of $+10^{\circ}\text{C}$ are not uncommon).

⁵ Average values for two stations in Pond Inlet for the period 1951-1980.

Despite overall frigid winter temperatures, the northern Baffin region sometimes experiences events of warm weather during the winter months with air temperature rising above 0°C as a result of winter cyclones migrating up the Davis Strait-Baffin Bay low pressure trough (Gilbert and McKenna-Neuman 1978).

Moisture is mostly supplied to Bylot Island glaciers from storms moving northwards from Davis Strait into Baffin Bay and accompanying warm air advection from the south. Most precipitation falls during early autumn and late spring before Baffin Bay is ice-covered (Kiassen and Shilts 1987). About half of the annual precipitation falls as snow and total accumulation over the winter months averages 4 to 9 cm (Canadian Climate Normals 1982). Precipitation likely to survive the spring-summer melt season and thus accumulate in glaciers mostly falls in late autumn and early spring. Some moist air is also supplied perennially to Bylot Island from the North Water polynia (*re* Fig. 2.1) at the head of Baffin Bay, which feeds the Devon Ice Cap.

Koerner (1977) established that the combined effect of perennially open water in Baffin Bay and the persistence of a low pressure cell there throughout the year brings high accumulation to land slopes facing the bay. Thus over the winter months, that part of Bylot Island facing Lancaster Sound and Baffin Bay may total more accumulation compared to glaciers to the south of the topographic divide, which are chiefly fed by northward moving moist air. This may explain the relative stability of ice margins on the north coast of the island, compared to the general recession observed on glaciers of the southern coast.

2.5) Previous work

Only recently has the extensive field work conducted by the Geological Survey of Canada demonstrated the research potential of Bylot Island, chiefly through the work of

Hodgson and Haselton (1974) and Klassen (see previous section). Prior to this, regional surveys of the Eastern Arctic glacial environment were published by various authors (Falconer 1962, Andrews and Miller 1972, Field 1975). Of particular importance to this thesis are two recent series of investigations conducted on southwestern Bylot Island:

DiLabio and Shilts (1978, 1979)

A study of compositional variations in glacial debris was carried out by DiLabio and Shilts (1978, 1979) on several large glaciers of southern Bylot Island flowing across the AA / KT geological boundary. The chief objective of the study was to ascertain the extent of mixing and the spatial distribution of the various lithological components of glacial debris being transported in glaciers. To this end, samples of debris collected from vertical glacier ice profiles and adjacent lateral moraines were analyzed for mineralogy and trace element geochemistry.

It was found that over 99 % of the sand and coarser debris in the ice and moraines is derived from the Precambrian crystalline terrane. The heavy mineral suite of sand-size glacial debris did not allow clear distinction between material derived from Shield lithologies and those of the KT sedimentary beds, because of the mineralogical immaturity of the latter. This was confirmed by Bernier (1985), who examined the heavy mineral suite of lateral moraine samples collected by DiLabio and Shilts and found that only the occasional presence of pseudorutile in debris derived from the sedimentary terrane suggested a different source than the crystalline basement.

Examination of lateral moraine and vertical ice profiles on Aktineq and Ujarasukjuit Glaciers established that an inverse correlation exists between the concentration of Shield-derived trace metals relative to the ratio of expandable clays and chlorite-kaolinite in the clay

fraction of glacial debris. Montmorillonite-group minerals in the clay fraction probably originated from the *KT* substrate, while the well-crystallized micas and chlorites were most likely derived by glacial comminution of phyllosilicates eroded from the crystalline Precambrian rocks in the upper reaches of the glaciers. From this contention, DiLabio and Shilts inferred that down-ice decrease of metal concentrations in the clay fraction observed in lateral moraines may reflect the progressive mixing and dilution of AA-derived debris by metal-poor clays from the sedimentary beds during glacial transport.

The use of clay mineralogy as a provenance tracer and the general investigative approach described above supports part of this work. Consequently, analytical results presented in Chapter 5 will be discussed in the light of those obtained by DiLabio and Shilts.

Lorrain et al. (1981)

Lorrain *et al.* (1981) conducted a study on glaciers of the Canadian Arctic to determine the chemical and isotopic composition of basal ice and its relation to glacier dynamics. As part of this work, they sampled ice from Aktineq, Ujarasukjuit, and Sermilik Glaciers on southwestern Bylot Island.

Using the co-isotopic analytical approach outlined in Chapter 1, Lorrain *et al.* concluded that some of the basal ice in these glaciers originated from a meltwater refreezing process at the sole, which could be partly responsible for debris entrainment. Their results, developed in subsequent publications (Souchez and Jouzel 1982, Souchez *et al.* 1988) will be discussed in greater detail in Chapter 6.

Chapter 3

Field Work and Observations

Field work was carried out on southwestern Bylot Island in July and August 1991-1992 from a series of helicopter-supported field camps established in the vicinity of Aktineq and Ujarasukjuit Glaciers (*re* Fig. 2.2). All camps were set along the foothills of the Byam Martin Mountains, in close proximity to the mapped location of the faulted AA-KT geological contact, and within walking distance to major glaciers.

3.1) Principal glacier features

Numerous reconnaissance traverses were made onto and around the main glaciers of the study area in order to ascertain active glaciological and glacial sedimentological processes. Observations summarized hereafter comprise descriptions of the various basal ice facies that were sampled. Plates are annexed in Appendix I. Sediment texture descriptions made are based on the standard Udden-Wentworth granulometric scale.

Large-scale features

The piedmont glaciers of southwestern Bylot Island display structural features similar to those described by Shaw (1977), Lawson (1979) and others on polar and sub-polar glaciers of the northern and southern hemispheres. The glaciers are shaped as elongated tongues with smooth longitudinal and transversal profiles. The surface of glaciers is scattered with highly angular rubble derived from valley walls or nunataks as medial moraines and rockfall debris (Plate 1). Various supraglacial features such as

crevasse scars and debris trains define transverse arc-shaped folds with sub-vertical hinges reflecting large-scale deformation of the ice due to differential flow rates.

The englacial zone which includes the bulk of glacier ice is nearly free of debris and is pervasively foliated on a large scale as a result of differential flow rates. Ice foliation is generally horizontal to sub-horizontal and is defined by textural variations in ice crystals, air bubble content, and the occasional presence of dirty ice layers. Observations at glacier margins suggest that most of the sediment load is transported in the basal zone as a relatively thin (5 m to 10 m thick) layered sequence of ice and debris. Near the glacier terminus, the basal zone thickens as a result of compressive flow, recumbent folding and complex thrusting. Some debris from the basal zone appears to be brought up along shear planes and is extruded on the stagnant glacier surface in the form of wedges, elongated ridges, or mounds (Plate 2). The accumulation of a debris veneer on the supraglacial surface insulates ice against rapid ablation and, combined with differential melting rates, results in the formation of large extents of hummocky ice-cored moraine surrounding the termini of glaciers.

Morphology of the ice margins

Several glaciers flowing over the southwestern Bylot Lowlands have undulating margins that display contrasting zones of steep, near-vertical ice cliffs and smooth, low-angle slopes. The large cliffs result from the erosive action of meltwater streams flowing in direct contact with the ice which are actively undercutting the margin during the melt season. Steep ice walls along Aktineq Glacier may reach 30 m in height, but are typically lower, in the order of 15 to 20 m high, whereas cliffs on Ujarasukjuit Glacier are often up to 60 m high or more (Plate 3).

The near-vertical margins displayed by glaciers on southwestern Bylot Island provide excellent exposure of internal structure and stratigraphy. Most direct deposition and reworking of glacial sediments occurs near the ice margin as it advances and retreats during glacial episodes. Thus, ice-marginal processes play a key role in glacial sedimentation, even though they may not represent conditions prevailing underneath the central parts of the glacier.

Basal ice stratigraphy

The basal zone of Bylot Island glaciers presents a variety of aspects that makes it somewhat difficult to establish a consistent description encompassing all features. Consequently it is more convenient to consider different glaciers separately:

Aktineq Glacier (B-17)

Along the eastern margin of Aktineq Glacier the exposed part of the basal zone ranges in height from 1.5-8 m near the Byam Martin Mountains foothills to 10-15 m down-ice over the lowlands. Such estimates may not reflect accurately the true thickness of the basal zone, as they apply only to that part extending from the surface of adjacent moraine up to the contact with the overlying englacial zone. Thus, the thinner aspect of the basal zone in the upper reaches of Aktineq Glacier may be partly accounted for by the presence of thick accumulations of lateral ice-cored moraine that obscure part of the section (Plate 4).

Stratigraphy of the basal zone consists of a lower, crudely stratified to massive debris-laden ice facies a few m thick, overlain by a considerably thicker (up to 12 m) clear ice facies with occasional thin debris-rich layers. These units, also observed in other glaciers of the field area, may be equivalent, respectively, to the *stratified* and *dispersed* facies of Lawson (1979). In general, both debris layers and ice foliation dip into the glacier

at angles varying from 25° to 35° over several tens of metres. In the piedmont area, the attitude of ice foliation becomes increasingly steeper to reach a sub-vertical orientation near the point where the glacier flows out of the mountains.

The lower stratified facies is rarely observed along the upper 3 km of the eastern ice margin: best exposures are located on the western margin, at the foot of ice sections eroded by cascading meltwater. The exposed part of the stratified facies is only a few metres thick (max. 2.3 m) and contains highly variable amounts of debris¹, ranging in size from fine sand to boulders, all of which is seemingly derived from Precambrian crystalline basement rocks. Larger clasts appear randomly oriented, but their considerable degree of rounding and sphericity precludes effective fabric measurements. There is a crude stratification between thick (5-12 cm) frozen debris bands with minor amounts of interstitial dirty ice, and layers of dark, bubble-free, macrocrystalline ice bearing thin (mm-thick) laminations of silt-size to sand-size debris. The stratification is laterally discontinuous and commonly distorted into near-horizontal folds.

Contact with the upper dispersed facies is commonly sharp although irregular, and is marked by the disappearance of the thick frozen debris bands and an overall decrease in the debris content of the ice (Plate 5). The bulk of the clotted ice facies is clear or slightly dirty, mostly coarse-grained and bubble-free, but shows occasional gaseous inclusions stretched in the plane of foliation along grain contacts. Aggregates of silty debris mixed with some sand and fine gravel are dispersed throughout the entire facies, sometimes forming sinuous, poorly-defined planes rarely more than a few cm long.

¹As previously observed by Gow *et al.* (1979) and Lawson (1979), only small amounts of dispersed debris are required to give ice layers a deceptively dirty appearance. Visual estimates of debris volume content in different ice facies are therefore likely to be exaggerated.

Several dirty ice layers occur in the dispersed facies. They are considerably thinner than those of the stratified facies and appear to contain a higher ice-debris ratio. Ice associated with such layers is generally laminated, coarse to macrocrystalline, while debris is mostly sandy but ranges in size from silt to fine gravel. Angular to sub-angular cobbles and boulders are present in the facies as well. Cobbles seem mostly associated with the thin debris bands, while boulders are distributed randomly without preferential orientation, except when they intersect one of the debris bands. In the latter case, they are often oriented in the plane of ice foliation and ice laminations within the debris layer appear to "stream" around the obstacle, possibly as a result of stress-induced recrystallisation (Plate 6). Except for occasional large clasts, the overall size range of sediment in clotted ice ranges from sand- to clay-size. The debris load in the upper part of the basal zone thus appears considerably finer in average than that of the stratified facies where material ranges from clay-size particles to boulders. This is consistent with observations made by Lawson (1979) and others on alpine glaciers.

Features reported above are in general accordance with those described by Lorrain *et al.* (1981) on Aktineq Glacier. Their published observations, although detailed, were restricted to a 14 m high section located about 2 km up-ice from the snout of the glacier along the eastern margin. In addition to the overall presence of coarse and fine bubble-free ice and the occurrence of sandy layers, Lorrain *et al.* reported isolated layers of bubbly ice associated with thick bands of frozen morainic debris layers. A thin (8 cm) band of frozen sediment containing foliated clear ice lenses was also observed at the base of the section (R. Souchez, personal communication). No such layers were identified by this author, although occasional coarse debris layers were observed in the upper part of the dispersed facies. It is not known if these layers correspond to the ones described by Lorrain *et al.*,

but the associated ice appeared darker than would be expected from bubbly ice (commonly white), which suggests a different origin.

The contact between the basal and englacial zones of Aktineq Glacier is not always well-defined. In some locations, meltwater runoff on smooth slopes deposits a widespread film of algae² and sandy supraglacial debris over the glacier surface, which obscures considerably the aspect of the basal sequence. On other sites, the transition is abrupt and coincides with a coarse debris layer. More often, the passage into the englacial zone is gradual over a short vertical distance and is marked by the rapid decrease in the number and thickness of dirty ice layers.

The internal layering of the englacial zone was examined on a section exposed by the collapse of a large portion of ice on the eastern margin of Aktineq Glacier. Englacial ice is for the most part composed of alternating bands of white, coarse bubbly ice and greyish blue bubble-poor ice. The difference in colour reflects the difference in grain size and/or gas content, and defines a textural layering observable both on a large-scale and in hand-size samples. Large angular clasts (size up to over 1 m), probably of supraglacial origin, are scattered throughout the whole facies. As well, aggregates of silt-size debris with occasional granules, possibly of eolian origin, are suspended in the bulk of the ice, and sometimes define irregular layers associated with thin zones of finer-grained bubble-poor ice. Measurements on *in situ* englacial ice showed the internal foliation to be sub-horizontal. A few rare but well-defined debris layers also occur randomly within the englacial zone: their debris content appears considerable, with coarse, angular clasts ranging in size from sand to cobbles (Plate 7). These layers, which were probably derived by burial of supraglacial debris in the accumulation area, sometimes appear to terminate

²*Ancylonema nordenskiöldii* is a type of dark-coloured algae which preferentially grows on the glacier's surface (Bouchard 1991).

with a large imbedded clast. Similar observations were made in Antarctica by Shaw (1977), who suggested such clasts were incorporated by falling in a supraglacial crevasse that subsequently closed. Features of the englacial clean ice and its associated debris layers at Aktineq Glacier are comparable to those described by Lawson (1979) as that of the englacial *diffused* and *banded* facies, respectively.

Ujarasukjuit Glacier (B-7)

In almost all sites examined on Ujarasukjuit Glacier, the englacial zone forms a distinctive overhang above the basal zone (Plate 8). This contrast results from preferential melting of the low-albedo debris-laden dirty ice in the basal zone. Interestingly, the overhang does not always coincide with a sharp change in the aspect of the ice, but rather seems to extend over a narrow zone where the basal dispersed facies grades into the englacial zone. Along the eastern margin, the basal zone typically consists of a very thin (< 1 m thick) massive to poorly stratified debris-rich facies, often unexposed, overlain by a much clearer ice zone up to 12 m high (Plate 9).

Where exposed, the lower facies was found to be essentially similar in aspect to that exposed along Aktineq Glacier. However, the overlying dispersed facies differs from that of Aktineq Glacier by its overall lower content in debris and a greater textural variability of the ice matrix. Layers of clean coarse-grained to macrocrystalline ice up to 8 cm thick alternate with zones of fine-grained ice bearing discontinuous laminae of silty to sandy material. Thin, discrete debris-rich ice layers and dispersed clasts of varying sizes are found throughout the basal zone, dipping at sub-horizontal angles (2° to 3°) into the glacier. Limbs of long inclined folds, and oblique shear planes are sometimes recognizable, but are generally overprinted by the sub-horizontal planar ice foliation. Contact with the

englacial zone is sometimes marked by a coarse debris layer (Plate 10), but is more often gradual. Englacial ice in Ujarasukjuit Glacier appears to contain fewer discrete debris layers or scattered boulders than on Aktineq Glacier.

On the western margin, a 7 m high basal ice section exposed by meltwater erosion near the glacier terminus displays features that offer a remarkable contrast with those of the eastern margin (Plate 11). The lowermost 3 m of the ice section consists of frozen morainic sub-angular debris crudely interbedded with lenses of coarse dirty ice. Sigmoidal foliated lenses of clear white bubbly ice up to 1 m long are interlayered with the dirty ice and stand out in positive relief. The absence of debris within those lenses and their bubble-rich content suggests an origin by basal incorporation of wind-blown snow from an overridden marginal cavity, as was suggested by Tison *et al.* (1987) for similar features at Glacier de Tsanfleuron (Swiss Alps). The whole of the debris-rich sequence is strongly distorted into complex folds, and the ice foliation is contorted around large unstratified aggregates of frozen debris.

A discontinuous horizon of frozen peaty sediment capping the lower facies was sampled for radiocarbon dating and gave an age of 130 ± 50 ^{14}C years B.P. (GSC-5354, uncorrected). Klassen and Shilts (1987) had previously reported the occurrence of pre-existing sediments incorporated within the glacier as intact rafts. DiLabio and Shilts (1979) described such a clast of slightly deformed waterlain sediment with preserved internal bedding melting out of the ice some 30 m above the bed of Ujarasukjuit Glacier. In situ peat capping the sediments was dated at 120 ± 120 ^{14}C years B.P. (uncorrected). This is suggestive that at least part of the debris load within the lower stratified facies may result from recent overriding and incorporation of marginal frozen debris.

Over 4 m of coarse to macrocrystalline dirty ice interbedded with thick debris-laden layers overlie the lower part of the section and extend up to the englacial zone. Lens-shaped layers of white bubbly ice up to 80 cm thick occur as well, mostly in the lower part of the dirty ice zone. The upper contact with englacial ice is foliated, sheared, and marked with large augen structures cored by aggregates of coarse debris or large isolated clasts.

The occurrence of a thicker basal sequence containing large amounts of debris, some of which may have been sheared-in, generally agrees with the notion of a marginal cold-based zone in polythermal glaciers where subglacial freezing-on and glaciotectionic thrusting are favoured.

Other glaciers

As most of the field work was conducted on Aktineq and Ujarasukjuit Glaciers, descriptions of features relating to other glaciers in the study area will only be discussed briefly. B-26 and B-28, located east of Aktineq Glacier, are considerably smaller glaciers draining the same accumulation area. These glaciers do not extend onto the Bylot Lowlands and are assumed to be entirely underlain by crystalline rock of the AA complex.

A large (1 x 1.5 km) meltwater-fed icing extends from the snout of Glacier B-26 and obscures the basal zone at the frontal margin. A single exposure of basal ice was examined in a narrow meltwater-cut gully on the northern side of the glacier. The 5 m high section offers a very uniform appearance, with basal ice grading upwards into the overlying englacial ice. Debris-bearing ice essentially consists of layers of clear macrocrystalline ice interbedded with thin (mm-thick) laminae of silt- to fine sand-size sediment. Sub-angular pebbles, cobbles and boulders are scattered throughout the sequence without preferential orientation or sorting. Debris content decreases upwards and the ice lamination grades into

alternating clean white and blue ice layers similar to those found associated with englacial ice of Aktineq and Ujarasukjuit Glaciers.

A spectacular distorted debris layer was observed on the southern margin of B-26 Glacier (Plate 12). The base of the layer coincides with a descending break-in-slope in the glacier bed. Debris (mostly coarse) appears to be sheared upwards into the ice, and the layer stretches down-glacier over a distance of nearly 200 m, cross-cutting the ice foliation at about 20° and extending to the top of the glacier. Sediment is released on the supraglacial surface by ablation and drapes the ice with a train of debris. A few metres below the surface, the debris layer is distorted into a S-shaped fold probably resulting from differential ice movement rates along the principal flow line.

B-28 Glacier differs from B-26 by a smoother profile and by the extensive recessional moraines surrounding its margins. Near the glacier snout, wedge-shaped aggregates of frozen debris up to 1m long are protruding upwards from the bed into the overlying ice along shear planes dipping up-glacier. Debris in the wedges is mostly coarse, and ranges in size from fine sand to cobbles with interstitial laminated ice and silt layers. It is difficult to assess whether the wedges extend to the glacier sole, as meltwater has undercut the base of the features, leaving only the uppermost part visible.

Since these wedges occur in the compressive marginal zone, their origin is probably linked to a shearing mechanism. Basal ice on B-28 Glacier was also examined in subglacial cavities formed by meltwater erosion beneath stagnant ice. The ice is macrocrystalline to very coarse and interlaminated at tight intervals with sandy and silty material. The presence of a veneer of well-sorted sandy sediment on the floor of the subglacial cavities suggests that the coarser component of the debris lamination may be derived from glaciofluvial sediment incorporated in overlying moving ice by regelation.

3.2) Sampling methods for ice and debris

During the 1991 and 1992 field seasons, over 140 samples of glacial debris and ice were collected from the basal zone of Aktineq, Ujarasukjuit, B-26 and B-28 Glaciers. The field work strategy aimed at providing a database for mapping two-dimensional (longitudinal and vertical) mineralogical variations in the basal debris load. Most of the sampling was performed in the vicinity of the Byam Martin Mountains foothills to make use of the geological constraint provided by the contrasting *AA* and *KT* terranes. Systematic traverses were conducted along the lateral margins of Aktineq and Ujarasukjuit Glaciers, which provide the best exposures of the basal ice sequence. Additional samples of ice and debris were collected from B-26 and B-28 Glaciers for comparative purposes.

The adopted investigative approach was to make use of the clay fraction ($< 2\mu\text{m}$) mineralogy as a discriminant factor between *AA*- and *KT*-derived debris following the suggestions of DiLabio and Shilts (1979). Accordingly, care had to be taken to collect samples with sufficient clay-sized material to satisfy analytical requirements.

Aktineq Glacier. (Fig. 3.1)

Basal debris sampling on Aktineq Glacier was carried out along a longitudinal traverse on the eastern margin. 40 samples were collected from basal ice at 75 m intervals over an total distance of nearly 3 km stretching down-ice from the approximate mapped location of the *AA-KT* contact. Samples were collected within the lowermost accessible debris-rich layers of the basal zone, which was generally located in the upper part of the stratified facies, the lowermost stratified facies being only exposed in 12 locations out of 40 along that part of the eastern margin covered in the traverse.

Fig. 3.1. Sample location map
Aktineq Glacier - Ice and debris samples

Each sample consisted of about 500 ml of sediment, ranging in size from fines to pebbles. Samples were bagged for transport and numbered AK-91-15 to AK-91-54. The debris : ice ratio in the sampled horizons was not measured, but unpublished data provided by R. DiLabio gives volume concentrations³ of debris in basal layers ranging from 2.9 to 36.5%, with a mean value of 21.6%. Debris concentration in the basal zone was found to vary greatly from station to station: an estimate can be given by the number of debris bands over a 1 m section, which varies from 3 to 22 depending on location.

The slope of the margin and the dip of ice foliation were measured at each sampling location. Slopes vary randomly from 17.5° to 61.5° (averaging 29.5°) while dip ranges from 18.5° to 40.5° and increases to sub-vertical in the Byam Martin Mountains foothills. In general, higher abundance of debris layers is associated with steeper slopes and ice foliation. This is probably related to the effects of meltwater erosion, as meltwater cascading over steep slopes erodes deeper into the ice toe, exposing the lower, debris-rich part of the basal ice sequence. In several locations along the margin of the glacier, surface meltwater runoff tends to disperse the finer components of debris released by ablation. Consequently, special care was taken during sampling to ensure the sediment being collected was not winnowed or contaminated by surface runoff effects.

In addition to the eastern traverse, six samples (AK-91-01 and AK-91-08 to AK-91-14) of basal debris were collected along part of the western margin near Bigfoot Lake. Poor exposure did not allow the different basal ice facies to be identified at this location. Sediment collected within the lower part of the section appears to be generally coarser (sandy to gravelly) than that of the eastern margin, probably as a result of winnowing by the waters of Aktineq Creek.

³ Volume percentages of debris are given relative to ice content in samples expressed as water equivalents.

Supraglacial material was obtained from various locations on Aktineq Glacier. Samples AK-91-02 to AK-91-07 were collected from a series of elongated debris ridges trending parallel to the ice margin and containing unusual amounts of fine (sand- to silt-size) material. These ridges appear to be formed by sediment brought up from an englacial or subglacial position along shear planes and extruded on the glacier surface. Samples AK-91-55 to AK-91-57 were collected in a supraglacial stream bed near the Byam Martin Mountain foothills and are essentially composed of sandy angular sediment which probably represents reworked rockfall debris.

Ujarasukjuik Glacier (Fig. 3.2)

Sampling on Ujarasukjuik Glacier was carried out on various locations over the two field seasons. In 1991, basal ice and debris samples (B7-91-01A to E and B7-91-03A to E) from the stratified facies were obtained from two 6 m high sections during a reconnaissance traverse near the glacier terminus. Easy access to well-exposed basal sections allowed a more systematic sampling campaign to be conducted on the eastern margin during the summer of 1992.

Two distinct methods were employed. First, basal debris was collected along a longitudinal traverse using the same methodology previously described for Aktineq Glacier. The lower stratified facies being rarely exposed in the basal zone of Ujarasukjuik Glacier, sediment was obtained from the lowermost exposed debris layers of the dispersed facies instead. The uniform ice margin configuration along the eastern side of Ujarasukjuik Glacier ensured a high consistency in the stratigraphic position of the samples. A total of 25 samples were collected at average intervals of 70 m. The traverse covered a segment of the

ice margin starting about 500 m up-ice of the assumed location of the *AA-KT* contact and extending roughly 1.8 km down-ice.

Additionally, ice and sediment were obtained from 10 vertical profiles through the basal sequence in various positions along the margin (Plate 13). Two profiles (VS1 and VS2) were sampled up-ice of the assumed position of the *AA-KT* contact, the eight others (VS3 to VS10) in locations ranging from 0.3 km to 7.2 km down-ice of the contact. As the ice margin on Ujarasukjuit Glacier is steep and no ice climbing equipment was available, the profiles were only sampled within an accessible range of heights which varied with the local basal zone configuration. Approximate correlation between different profiles may be established by referring to the transitional contact marking the base of the englacial zone. In most profiles, the lowermost layer sampled was at or near the elevation of the surrounding ground, close to the glacier sole.

An exception is profile VS7, located near the glacier terminus, where the basal ice sequence is only exposed 5.5 m above ground due to the presence of a large marginal moraine. Surprisingly, the ice stratigraphy at VS7 was found to be comparable to that of other profiles, despite the nearby occurrence of large folding and shearing structures disrupting the internal ice foliation (Plate 14).

From each profile, an average of seven 250 ml samples of sediment and ice were collected in the thin debris-rich layers occurring throughout the dispersed basal facies. Special care was taken during sampling, as the ice matrix was to be preserved for stable isotope analysis. To avoid contamination by isotopic exchanges with percolating meltwater, each sample was collected from a depth of 10-12 cm within the ice wall; persisting cold weather facilitated this task by greatly reducing meltwater runoff during the sampling period.

Melted ice samples were double-bagged and sealed prior to shipping from the field, with as little free air as possible remaining inside the bags to prevent isotopic fractionation due to evaporation. Upon reception of the shipment in Ottawa, meltwater was transferred into 20 ml polyethylene vials with airtight screwcaps. A single sample was lost during transport, and the amount of water which seeped from others was minimal. As mixing between the content of bags during transport is very unlikely, the meltwater samples were considered to be pristine and uncontaminated.

Other glaciers

For comparative purpose, a few specimens of basal ice and debris were obtained from various locations on nearby B-26 and B-28 glaciers. Samples B26-92-16 to B26-92-24 were collected in the laminated ice sequence exposed on the northern margin of B-26 Glacier. On the eastern margin of B-28 Glacier, a single sample (B28-92-47) was obtained from a well-defined 10 cm-thick debris layer dipping steeply into the glacier. Although the basal zone was not clearly exposed at this location, the dip of the debris layer suggests a subglacial transport position. Additionally, R. DiLabio and W. Shilts provided some unpublished mineralogical and textural data obtained from analyses of moraine and basal debris from Aktineq, Ujarasukjuit and Sermilik Glaciers.

3.3) Sampling of bedrock and erratics (Fig.3.3)

Bedrock samples were collected in 1991 and 1992 from various locations over the AA (BYR-91-01 to BYR-91-05 and BYR-92-01/02) and KT (BYR-92-03 to BYR-92-06) terranes to determine the provenance of specific mineralogical components of the debris load from glaciers in the study area.

Fig. 3.3. Sample location map
Bedrock and erratic samples

- Bedrock sampling sites
- ▲ Erratic sampling sites

Samples numbered BYR-91 and BYR-92 were collected by this author in 1991-1992. Samples numbered 77-BR were collected by H. Balkwill and analyzed by R. DiLabio and W. Shilts in 1977-1978

Archean-Aphebian rocks

The limited mobility of the field party did not allow bedrock to be sampled from outcrops located far up in the highlands. To compensate, crystalline erratics were collected from lateral moraines and supraglacial positions along the major glaciers, in addition to bedrock specimens. Because these erratics differ from the metamorphic rocks exposed near the foothills, they are presumably derived from lithologies outcropping considerably higher up-ice. For the scope of their reconnaissance mapping, Jackson and Davidson (1975) grouped crystalline bedrock units outcropping on southwestern Bylot Island in a single undifferentiated banded migmatite unit. However field observations suggest an important lithological variability within the unit. In spite of this, crystalline rock samples collected in the study area are considered representative of the overall petrographic composition of the migmatite unit.

Cretaceous-Tertiary rocks

Rock samples from the *KT* terrane were obtained from exposures of the Eureka Sound and Hassel Formations. Eureka Sound Fm specimens are from the upper sandstone member which outcrops over most of the lowlands in the study area (Miall *et al.* 1980). The samples were collected from a section exposed in an incised meltwater gully east of Ujarasukjuit Glacier, and typically consist of poorly-consolidated olive-grey fine-grained sandstones or siltstones displaying complex lamination and flow structures. Quartzose sandstones of the Hassel Fm were obtained from the north shore of Bigfoot Lake, east of Aktineq Glacier. Additionally, R.DiLabio and W. Shilts provided results of analyses performed on *KT* bedrock samples collected by H.R.Balkwill (# 77-BR-03 to 77-BR-28 on Fig. 3.3).

Chapter 4

Analytical Methodology

The analytical methodology adopted for this study was determined by the research objectives and the available laboratory facilities (Fig. 4.1). Work was essentially divided in three steps:

A) Using dry weight and scanning laser methods, the particle size distribution of glacial debris and pulverized bedrock samples was determined in order to link textural characteristics of the basal debris load with its mineralogical composition, position of transport in the ice, and possible mode of entrainment.

B) X-ray diffraction analysis of the silt and clay fractions from basal glacial debris and crushed bedrock samples aimed at identifying and defining patterns of spatial compositional variations of the basal ice load relative to transport position in the glacier.

C) Co-isotopic ($\delta^{18}\text{O}$ - δD) analysis of basal ice samples by mass spectrometry was performed to determine the genetic signature of basal layers and to define the nature of subglacial debris entrainment processes for glaciers in the study area.

Pre-analytical treatments of debris samples and XRD analysis were performed in the Geotechnical Science Laboratories (Department of Geography) and in the X-Ray Diffraction Laboratory (Department of Earth Sciences) at Carleton University, Ottawa. Stable isotopic analyses of ice samples were conducted at the Ottawa-Carleton Geoscience Center Stable Isotope Facility, Department of Geology, University of Ottawa.

Fig. 4.1. Analytical flowchart

4.1) Particle size analysis

Principles of particle size analysis are reviewed in several recent texts (Gee and Bauder 1986, McCave and Syvitski 1992). Commonly employed methodologies combine various techniques such as sieving, hydrometer, settling and centrifuge to produce an estimate (volumetric or weighted) of particle size distribution. Results are expressed with reference to standard granulometric scales such as the USDA, Atterberg, or modified Udden-Wentworth scales. For glacially derived sediments, results are often plotted over a continuous size scale to allow detection of any modal recognition that may result from glacio-sedimentary processes.

Note on grain size class designations

The grain-size separation method chosen for this study is fairly simple and aimed at efficiently isolating the clay fraction from glacial sediment. Samples were separated into three size fractions designated as *clay*, *fine*, and *coarse*. The grain-size range of each fraction was determined by splits at 0.002 mm and 0.074 mm (200 mesh - US standard) as shown below.

<i>Designation</i>	<i>Size range (μm)</i>
Clay	< 2
Fine	2 - 74
Coarse	> 74

Time did not allow a more sophisticated separation to be performed but the method was judged satisfactory in view of its aim. However, this fairly crude size separation limits the possibilities of interpretation of grain size distribution data.

Analytical methodology

Basal debris samples collected from glaciers in the study area typically consist of silty to sandy diamicton containing abundant pebbles and an estimated 10 % (Vol.) cobbles > 1cm. For all samples, grain-size separation was performed as follows:

1) A specific volume of sediment was initially dried at low temperature in an air-circulation oven and weighed on an electronic balance with a precision of ± 0.01 g. The volume of sediment available varied with the sampled medium. For analytical purposes, the following amounts were utilized: 150 ml for the AK and B7 sample series (long profiles - Aktineq and Ujarasukjuit Glaciers) and for samples obtained from Glaciers B-26, B-28, and Sermilik (78GS series) Glaciers ; and the total volume of sediment available for the vertical profiles collected on Ujarasukjuit Glacier (in average 53 g of sediment for 250 $\pm$ 10 ml of ice).

2) Standard grain size analysis techniques require that sediment be dispersed in an aqueous solution by both chemical and physical means (Gee and Bauder 1986) prior to size separation. As well, various pre-analytical chemical treatments are recommended in order to remove organic matter, oxide coatings, and soluble salts. Such treatments mostly apply to soils in which chemical weathering and biogenic effects may have altered the mineralogical composition. Glacial debris samples from Bylot Island appear to consist mostly of fresh rock fragments physically comminuted in an environment unfavourable to intense chemical weathering. Secondary minerals present in the debris are thus more likely to have a

metamorphic or diagenetic origin. Thus, no chemical treatments were applied to the debris prior to size segregation procedures. However, some samples were found to contain organic material (mostly algal remains derived from the ice surface) mixed with sediment. This material was successfully removed by thoroughly rinsing samples with pH-controlled deionized water and discarding the supernatant once the sediment had settled.

3) After initial weighing and rinsing, each sample was saturated with water and disaggregated using a Quigley-Rochester ultrasonic probe at maximum setting for 1.5 min. This operation was performed primarily to remove the coating of silt and clay from coarser fragments (Edwards and Bremner 1967). Samples were then washed through a 1mm diameter copper sieve and a 200 mesh (0.074 mm) stainless steel sieve after clasts over 1 cm in had been temporarily removed to avoid obstructing the sieves. The > 74 μ m fraction was dried and stored. The finer fraction encompassing very fine sand, silt, and clay was further subdivided as described hereafter.

4) To separate clay-size material from the > 74 μ m fraction, a settling method was used. The technique is based on Stoke's Law, which predicts the gravitational fall velocity V_g of a particle of radius r submerged in a static fluid of viscosity η , such that:

$$V_g = \left\{ \frac{4gr\Delta\rho}{18\eta} \right\} \quad (\text{Leeder, 1982})$$

where g is the gravitational constant and $\Delta\rho$ the difference between the density of the fluid and that of the particle. If the falling distance d of the particle is known, the time t required for a particle of specific radius r to cover this distance can be calculated since $V_g = d/t$.

Using this relation, removal of clay-size particles from a homogenized solution can be accomplished by siphoning the solution above level h after a time t has elapsed, where t is given by:

$$t = 18\eta h \{ g\Delta\rho(2r)^2 \} \quad (\text{Gee and Bauder 1986})$$

The method used in this work was adapted from that developed by Berry (1987). For each sample, the $< 74 \mu\text{m}$ fraction was placed into a 10 cm high cylindrical plastic container filled with distilled water. The solution was then stirred for one minute and left to settle. Using a standard value of $\Delta\rho = 1.65 \text{ g}\cdot\text{cm}^{-3}$ for clay-sized material, a minimum particle radius of $1 \mu\text{m}$ and a viscosity of $0.01 \text{ g}\cdot\text{cm}^{-1}\text{s}^{-1}$ for water at 20°C , the time required for most of the $2\text{--}74 \mu\text{m}$ material to settle was estimated at 6 hours, 10 minutes. Once settling was completed, the uppermost 8 cm of the solution were siphoned out from the cylinder into a separate container using a "J"-shaped siphon. This technique minimizes inclusion of oversized particles in the suspension being siphoned off. The cylinder was then re-filled and the procedure repeated until the remaining water was virtually clear of suspended sediment. It was found that three successive decantation and siphoning rounds were sufficient to separate clay from most samples. The silty deposit in the cylinders was transferred into vials, dried at low temperature and stored. Material within this fraction includes minor amounts of clay-size particles, but is primarily very fine sand to silt. The volume of clay lost in this procedure was estimated as negligible.

5) Separation of clay suspensions from the siphoned solution was performed without the use of flocculating agents to avoid contamination of the precipitate that may complicate x-ray diffraction interpretation. Instead, settling and centrifugation methods

were employed. A 72 hours long decantation step allowed the volume of solution to be reduced to 250 ml. The concentrate was then centrifuged for one hour at 2000 RPM according to calibration set by R.W.Berry. This step proved effective in settling most of the suspended clay from the solution. Supernatant water was then siphoned out and the remaining clay slurry was transferred into vials to be used for x-ray α -radiation procedures.

6) The coarse, fine and clay fractions obtained were separately weighed on an electronic scale with a precision of ± 0.01 g. As drying may alter clay mineralogy, the < 2 μ m fraction was kept in aqueous solution and weighed in the containing vials. Identical vials containing an equal volume of water were weighed separately and their weights were subtracted from those of the clay samples. This procedure introduced an error unaccounted for by the size separation method employed, as samples containing very low amounts of clay sometimes gave weights inferior to that of the water-filled vials. When this occurred, the weight of the clay fraction was considered equal to 0.

Quality control

By comparing the bulk sample weight's with the total weight of the separate fractions, a maximum closure error inferior to 5 % was found. Further quality control testing was done at the Sedimentology Laboratory of the Geological Survey of Canada where selected samples of the clay and fine fractions of debris were analyzed using a laser particle size analyzer. This instrument calculates the volumetric size distribution of particulate matter in aqueous suspensions by measuring the transition time needed for a 0.5 μ m wide He-Ne laser beam to cross the diameter of single particles as they pass through the focal zone of the scanning cell. The laser analyzer generally gives reliable data, despite some minor error due to the difficulty of accurately computing the volume distribution of

particles smaller than 0.5 μm (i.e. the analyzer slightly "underestimates" the relative volume of clay-size material in the sample).

Results (Appendix II) showed that the particle size range in the clay split falls mostly below 2 μm , although occasional coarser particles 2-4 μm in size occur in some samples, sometimes accounting for up to 17 % (Vol.), but more typically for about 10 % or less. Essentially, these results testify that the clay separation was performed quite accurately, although the < 2 μm fraction retained some of the silt-sized material from the sediment.

4.2) X-ray diffraction analysis

Note on terminology and classification of clay minerals

The expression *clay minerals* is used here in a broad sense to designate the various phyllosilicates commonly present in the silt and clay fractions of sediments. The adopted clay mineral classification, shown in Table 4.2, is that of Brindley and Brown (1984). Groups and subgroups in this system are defined based on mineral crystallographic structures, while species are identified mostly by chemical composition.

Expansible or "*swelling*" clays designate certain mineral groups or species, such as smectites or vermiculites, that exhibit intracrystalline swelling by incorporation of water and organic materials between silicate layers. *Interstratified clay minerals* are phyllosilicate structures in which two or more kinds of silicate layers occur in a vertical stacking sequence, commonly along the *c*-axis. The most commonly reported mineral components found in mixed-layer structures are illite¹, smectite, vermiculite, and chlorite.

¹ *Illite* is a term commonly used to designate hydrous fine-grained micas such as hydromuscovite.

**Table 4.1 - Classification of phyllosilicate clay minerals
(after Brindley and Brown 1984)**

Layer Type¹	Groups	Sub-groups²	Species²
1:1	Serpentine-Kaolin	Serpentines Kaolin	Chrysotile, antigorite, lizardite Kaolinite, halloysite, dickite, nacrite
2:1	Talc-Pyrophyllite	Talcs Pyrophyllites	Talc Pyrophyllite
	Smectites (E)	Saponites Montmorillonites	Saponite Montmorillonite, beidellite, nontronite
	Vermiculites (E)	Trioctahedral Diocahedral	Trioctahedral vermiculite Diocahedral vermiculite
	Micas	Trioctahedral Diocahedral	Phlogopite, biotite, lepidolite Muscovite, paragonite
	Brittle micas	Trioctahedral Diocahedral	Clintonite Margarite
	Chlorites	Trioctahedral Diocahedral	Clinochlore Donbassite
2:1 (Ribbon structure)	Sepiolite-Palygorskite	Sepiolites Palygorskites	Sepiolite Palygorskite
Interstratified Clay Minerals		Arrangement²	Species²
Regularly stratified		Muscovite - montmorillonite Chlorite-smectite Chlorite- vermiculite Mica-vermiculite	Rectorite (E) Corrensite (E) Corrensite (E) Hydrobiotite (E)
Randomly interstratified		Illite-smectite Mica-vermiculite Smectite-chlorite Chlorite-vermiculite Kaolinite-smectite Mica-chlorite	Illite-smectite (E) Mica-vermiculite (E) Smectite-chlorite (E) Chlorite-vermiculite (E) Kaolinite-smectite (E) Mica-chlorite
¹ Layer types are identified by the ratio of tetrahedral (silicate) to octahedral sheets in the mineral structure. ² Only the principal sub-groups and species are presented here. (E) designates expansible clays.			

Interstratified minerals nearly always contain an expansible phase, usually smectite or vermiculite, although some rare species such as mica-chlorite where neither component is expansible are known (Wilson and Nadeau 1985).

Basic principles

XRD analysis of debris from Bylot Island glaciers was performed using oriented clay mounts, a method abundantly discussed by several authors (Brindley and Brown 1984, Carroll 1970, Whittig and Allardice 1986). The process involves exposing dried clay slurry samples mounted on glass slides to Ni-filtered, Cu K α x-rays of known wavelength inside a diffractometer. For each sample, a characteristic *diffraction* is produced, which consists of an x-y plot of diffraction angle (expressed in $^{\circ}2\theta$) versus intensity (I) of diffracted radiation. Using Bragg's equation $n\lambda = 2d\sin\theta$, the diffraction angles of peaks on the diffractogram can be related to molecular spacings d in the corresponding mineral structure. Most modern diffractometers are equipped with an automated processor which computes and prints the d spacing values (in Å) and sometimes the relative intensity (in %) of each major diffraction peak.

The x-ray diffractogram of a specific sediment sample can be viewed as a signature representative of the mineral species in the sample and of their relative proportions. The ratio of diagnostic diffraction peaks associated with specific mineral species may be used as a semi-quantitative index of mineralogical composition, and can thus serve to distinguish between mineral assemblages from different facies, provenance or states of weathering (Dean 1969, Shilts 1978, DiLabio and Shilts 1979).

Preparation of clay mounts

Clay mounting methods are numerous and vary with the nature of the sediment being analyzed and the aim of the procedure. A fairly simple drop-mounting technique was adopted here, in which a layer of clay slurry of uniform thickness is deposited on a glass slide using a dropper. The slurry is then left to dry at room temperature and low humidity

conditions. Used with clay-size material, this method gives *oriented samples*: phyllosilicate particles settling within the suspension layer come to rest with their main crystallographic (*c*) axis perpendicular to the glass surface. Oriented clay samples produce diffractograms that display clearly the 00 ℓ peaks of basal reflections, thus facilitating mineral identification.

Gibbs (1965) tested and compared several existing clay mounting methods in terms of accuracy and precision, and concluded that although the precision of most methods is comparable (about ± 10 % error in every case), methods involving settling of particles in aqueous solution are less accurate than others such as the suction-on-tile and smear-on-glass techniques. This can be attributed to the terminal settling velocity of microcrystalline mineral species being several orders of magnitude lower than that of larger crystals. As a result, a surface segregation occurs during settling, with smaller smectite crystals preferentially overlying larger chlorite, kaolinite or illite crystals.

This may lead to a filter effect, as x-ray reflections from surface layers of a sample are sharper and stronger than those from underlying laminae. The objection is serious if one is concerned with accurate detection and identification of specific mineral species, but is not a major inconvenience for relative abundance analysis. Surface settling of smectites or vermiculites may lead to an overestimation of their relative abundance, but if the error is consistent it should not significantly modify variation patterns of relative abundance between different samples.

To obtain reliable XRD signatures from oriented clay slides, it is also necessary to ensure that the size range of particles in the slurry falls within that of clay minerals. Empirical evidence has demonstrated that the $< 2 \mu\text{m}$ fraction gives best results for XRD and that separation of clay from silt and larger mineral grains is theoretically sound (Carroll 1970). Obtaining reproducible XRD data is increasingly difficult with coarser sediment, in

part because of the presence of tectosilicates which become preferentially concentrated within the silt- to sand-size range due to their resistance to physical weathering or comminution. Large quartz or feldspar crystals in samples will often "mask" clay minerals or reduce the relative intensity of their XRD peaks. As well, considerable signature variations in a given sample can result from the random orientation of large particles.

Howes (personal communication) has reported differences in peak intensities of up to 15% (within two standard deviations) for duplicate analyses of silt-size loess samples. For the purpose of this work, it is assumed that size requirements have been met: laser scanning analyses results confirmed that particles in the clay suspensions are mostly below or close to 2 μm in diameter.

Some simple quality control measures suggested by Howes (1991) were applied during clay mounting procedures. To ensure that a uniform clay concentration would be applied to each glass slide, specific volumes of concentrated clay solution and distilled water (measured with a graduated dropper) were mixed and the resulting solution was drop-mounted onto the slides. A maximum of consistency was obtained by placing each slide over a piece of newsprint: once the proper volume of clay slurry had been applied, its opacity was such that characters on the underlying paper became hardly visible. After drying, each slide was examined through a binocular microscope to detect any major inhomogeneities of the clay laminae. Imperfect slides were rinsed off and re-mounted.

Pre-analytical and analytical procedures

X-ray diffractograms from oriented clay mounts give reflections corresponding to layer spacing periodicities 00ℓ along the *c*-axis of minerals. Common XRD identification procedures involve analyzing sediment samples before and after auxiliary treatments have

PM-1 3½"x4" PHOTOGRAPHIC MICROCOPY TARGET
NBS 1010a ANSI/ISO #2 EQUIVALENT

been carried out: this is particularly useful in oriented mounts to allow distinction between mineral species in which some basal reflections may overlap. Basal interlayer spacings of clay minerals may vary with humidity, dehydration, glycolation, cation exchange and acid leaching. Identification of specific clay species is based on the reaction of minerals to such treatments under controlled conditions (Carroll 1970).

For this study, a fairly simple mineralogical breakdown was required. It was previously established that the clay-sized fraction of basal debris transported in Aktineq and Ujarasukjuit Glaciers consists of a mixture of expansive, montmorillonite group clay minerals and well crystallized micas and chlorite- and/or kaolinite-type minerals (DiLabic and Shilts). These results were obtained by XRD analysis prior to and after saturation with ethylene glycol ($\text{C}_2\text{H}_4(\text{OH})_2$), as described below. A few additional tests were also conducted on selected samples to distinguish between some clay mineral species, and to test the variability and reproducibility of results.

Glycolation treatment

Ethylene glycol treatment was used to identify expansive clays or interlayered complexes with expansive phases for which the (001) basal reflection ($d \approx 15 \text{ \AA}$) may overlap that of chlorite ($d \approx 14 \text{ \AA}$). When clay minerals such as smectites are saturated with ethylene glycol vapour, glycol molecules invade the interlayer positions of crystals and link with cations (such as Ca^{2+}) to form large polyhedral organic complexes that cause the interlayer spacing to expand, the degree of swelling varying with mineral composition and layer charge. Expansion of smectites by glycolation is marked by a strong glycol-(001) peak at approximately $5^\circ 2\theta$, corresponding to d spacings in the $16.8 \text{ \AA}$ - $17.8 \text{ \AA}$ range, coupled with a proportional reduction in intensity of the $14 \text{ \AA}$ chlorite peak relative to that of

untreated samples. This effect also applies to some interstratified clay minerals such as chlorite-smectite.

Glycolation is often performed at about 60°C to increase vapour pressure. For this work, the procedure was conducted at room temperature: mounts were placed in an sealed glass chamber containing about 470 ml ethylene glycol and were left for 36 hours to allow optimal saturation, following recommendations by Carroll (1970) and Berry (1987). XRD analyses were performed immediately after slides were removed from the container.

Most mineralogical data used in this study was obtained from XRD analysis of untreated and glycolated clay mounts. Over 160 mounts were analyzed, including all samples collected from traverses along Aktineq and Ujarasukjuik Glaciers (AK-91 and B7-92 series) and from vertical profiles on Ujarasukjuik Glacier (VS series). All slides were scanned at room conditions prior to and after glycolation on a Phillips Norelco automated diffractometer at $2^\circ 2\theta$ /minute over a range of $3^\circ 2\theta$ to $34^\circ 2\theta$ using monochromatic Cu K α radiation.

550 °C heating test

Kaolinite is hard to distinguish from chlorite when they occur together, as the 002 and 004 reflections of chlorite ($\approx 7.13 \text{ \AA}$ and $3.55 \text{ \AA}$) overlap the 001 and 002 reflections of kaolinite ($\approx 7.15 \text{ \AA}$ and $3.58 \text{ \AA}$). Furthermore, the basal reflection of chlorite varies considerably as a function of its content in iron and magnesium. If the chlorite is magnesium-rich, distinction can be made by heating samples to 550°C heating test, which causes the intensity of the $14 \text{ \AA}$ peak of chlorite to sharpen while the intensity of the $7 \text{ \AA}$ reflection diminishes considerably, sometimes to zero (Berry 1987). Any remaining $7 \text{ \AA}$ reflection may then be attributed to kaolinite, although this is not fully conclusive.

This test was conducted on representative clay specimens from Aktineq (# AK-91-30/40/54) and Ujarasukuit (# B7-92-10/20) Glaciers to detect the presence of kaolinite. Selected clay mounts were heated for an hour at 550°C in a high temperature convection oven, then placed in a dessicator for cooling prior to x-ray diffraction to avoid rehydration by atmospheric humidity. Diffraction was performed with the same instrumental settings as specified for untreated and glycolated samples.

Reproducibility tests

The semi-quantitative clay mineralogy analysis performed in this study is based on the contention that XRD signature variations reflect compositional changes in the sample population. This assumption was tested by estimating the variance of results obtained from duplicate analyses of chosen samples. Three successive XRD runs were performed on representative clay samples selected from traverses along Aktineq and Ujuarasukjuit Glaciers. Each selected sample was analyzed prior to and after glycolation with the diffractometer settings specified earlier. Untreated and glycolated XRD signature variations were also compared for the whole sample set.

Additionally, three test slides were prepared from a given sample (B7-92-18), each with a different clay concentration. These were analyzed separately to estimate error resulting from possible inconsistencies in the clay mounting procedure. Estimates of analytical precision derived from this data are discussed in Chapter 5.

4.3) Analysis of bedrock samples

Various analytical tests were performed on bedrock samples collected in the study area to obtain petrographic and mineralogical compositional data. Analyses were conducted in three steps:

1) Thin sections were prepared from 9 samples (# BYR-91-01 to -03 and BYR-92-01 to -06). These were used to provide basic petrographic descriptions of the bedrock samples as well as establishing their relative degree of alteration and weathering.

2) Fragments of each of the 9 samples were mechanically crushed in a percussion rock mill. A mixture of sand-to granule-sized debris was obtained, which was then passed three consecutive times through a grinding mill in which particles are forced by centrifugal spin between two revolving plates of resistant steel alloy separated by a 0.5 mm space. An assessment of the resulting grain-size distribution was obtained by scanning laser analysis of three representative samples (BYR-91-01, BYR-91-03 and BYR-92-04).

3) The clay-size fraction of pulverized rock samples was separated using the method previously described for glacial debris (combined settling and centrifugation techniques). Clay slides for XRD analysis could only be obtained from 4 of the 9 samples, as some specimens did not yield sufficient amounts of clay-sized material. Alternatively, silt mounts were prepared for these samples using the bulk pulverized powder after passing through a 200 mesh sieve. All slides were analyzed at room temperature prior to and after glycolation, using equipment settings specified earlier.

4.4) Stable co-isotopic ($\delta^{18}\text{O}$ - δD) analysis

Stable isotope analyses of basal ice samples were performed at the University of Ottawa Stable Isotope Facility. A total of 15 specimens were selected from 8 of the 10 vertical profiles sampled along the eastern margin of Ujarasukjuit Glacier (VS1, VS3, VS5 and VS6 to VS9). The samples were obtained from debris layers occurring within the dispersed ice facies 0.34 to 10.61 m above ground level. Based on visual estimates, dissolved and suspended matter concentrations of the samples were assumed low enough not to interfere with analytical procedures. Each specimen was analyzed for co-isotopic ($\delta^{18}\text{O}$ - δD) composition in order to detect possible evidence of freezing effects. Analytical specifications provided by the Stable Isotope Facility are given below.

The deuterium-hydrogen ratio in water is determined by the analysis of H_2 gas generated via zinc reduction at 430°C . H_2 is analyzed on an automated double collector VG 602D mass spectrometer. Results are normalized to SMOW. The routine analytical precision, expressed by two standard deviations (2σ), is ± 1.3 ‰. The oxygen isotopic composition of water is determined by CO_2 -water equilibration at 25°C using an automated system. CO_2 is analyzed on a triple collector VG SIRA 12 mass spectrometer. The CO_2 -water fractionation factor used at 25°C is 1.0412. The routine precision (2σ) of the analyses is ± 0.10 ‰.

Chapter 5

Results and Interpretation

In this chapter, analytical data obtained from Bylot Island glacial debris and ice samples are presented and examined. Section 5.1 considers the results of particle-size and XRD analyses. A description of the interpretative method for semi-quantitative XRD analysis is provided, with an estimate of precision based on reproducibility of results and other sources of error. Textural and compositional variations in the basal debris load of glaciers are examined in detail. Section 5.2 focuses on the isotopic composition of basal ice from Ujarasukjuit Glacier. An interpretation is proposed and its significance statistically tested.

5.1) Composition of glacial debris

The mineralogy of clay-size debris in glaciers in the study area was examined by XRD analysis of over 160 samples collected from various locations. Because of the considerable volume they represent, original x-ray diffractograms are not included in the manuscript. Two copies are available for examination: one was deposited to the care of Dr. W.W. Shilts at the Geological Survey of Canada (Terrain Sciences Division) in Ottawa; the other is at the Department of Earth Sciences at Carleton University. Data obtained by XRD analyses are summarized in a series of tables included in this chapter and in Appendix III.

Clay mineralogy of the basal debris load

Diffraction patterns of clay-size glacial debris from Bylot Island display a consistent pattern with major well-defined reflections in the 4-14° 2 θ and 24-30° 2 θ angle regions (Fig. 5.1a). The diffractometer provides basal spacing measurements for all peaks showing a relative intensity significantly higher than that of the background noise. This facilitates the identification of mineral species, which is based on diagnostic XRD peaks in accordance with the equivalences given in Table 5.1.

Results of XRD analyses showed that although some considerable quantitative variations exist between samples collected at different localities, the bulk mineralogy of clay-size debris from southwestern Bylot Island glaciers essentially contains the following components:

<i>Mineral class</i>	<i>Species</i>
Phyllosilicates	Mg-rich chlorite, kaolinite, halloysite, mica (biotite) or/and illite, smectite (montmorillonite), chlorite-smectite
Tectosilicates	Quartz, plagioclase feldspars
Inosilicates	Amphiboles (probably hornblende)

Smectite and chlorite-smectite were identified by a 16.8-17.1 Å glycol-(001) peak (Fig. 5.1b). A second (8.9 Å) order smectite basal reflection was sometimes observed as well, although it usually overlapped the 8.5 Å amphibole peak. Vermiculite also reacts to glycolation, but its basal spacing rarely expands beyond 14.5 Å (16.3 Å in some Mg-rich vermiculites), whereas the 17 Å spacings computed by the diffractometer are consistently in the 16.8-17.1 Å range, which is more typical of smectites. Therefore it may be assumed

Fig. 5.1. Identification of major x-ray diffraction peaks.
Sample AK-91-20 - < 2 μm fraction
Monochromatic $\text{CuK}\alpha$ radiation; $2^\circ 2\theta/\text{min}$.

Table 5.1. Diagnostic XRD peaks selected for mineralogical analysis

Adapted from Carroll (1970) and Brindley and Brown (1984)

Basal spacings (rounded average, in Å)	Possible mineral species (00 l) = order of reflection, (G) = glycolated
18 - 22 (diffuse)	mica-smectite (G) (rare)
16.8 - 17.1	smectite (G)
14.8 - 15.5	smectite, vermiculite (G), chl.-smectite (G), chl.-vermiculite (G)
14.3	chlorite (001), chlorite-smectite, vermiculite
12.4	smectite
10.1	micas (001), illite, halloysite
9.20 - 9.70	illite, talc, pyrophyllite
8.97	smectite (G)
8.51	amphiboles
7.13	chlorite (002), kaolinite (001), halloysite, Al-serpentine
4.98 - 5.04	micas (002) or (004)
4.74	chlorite (003), vermiculite (006), illite, magnetite
4.26	quartz
3.66 - 4.04	plagioclase feldspar (several low-order reflections)
3.54 - 3.56	chlorite (004), kaolinite (002)
3.35	mica (003), quartz (low-order reflection), sillimanite
3.18 / 3.24	plagioclase feldspar (double peak)
2.84	quartz, smectite (g), vermiculite

that if vermiculite is present, it is likely to be a minor component relative to smectite. An occasional 15.5 Å glycol peak was attributed to interstratified chlorite-smectite, which is also likely to occur in view of the abundance of chlorite and smectite in the clay mineral assemblage.

Some additional difficulty was met in discriminating chlorite from kaolinite. The limited decrease of the 7 Å peak relative to the 14 Å peak after glycolation is suggestive of the presence of kaolinite. Contradictory evidence was obtained from samples heated to 550°C, which showed a marked increase of intensity of the 14 Å peak while the 7 Å peak disappeared entirely (Fig.5.1c). This reaction is associated with Mg-rich chlorite, but the absence of a remaining 7 Å reflection may also imply that kaolinite is not part of the assemblage. However, this may also result from the transformation of microcrystalline kaolinite to nearly amorphous metakaolin at high temperatures (Berry 1987).

A slow x-ray scan of the 24.5-25.5 °2θ region would be required to separate the 3.54 Å chlorite (004) and the 3.58 Å kaolinite (002) peaks and help determine if the latter is present. This method could not be applied here due to time constraints. Thus, in the absence of discriminating evidence, the presence of kaolinite as well as chlorite remains possible. This also applies to halloysite, which can not easily be distinguished from chlorite and kaolinite in its 7 Å hydrated form.

The 10 Å peak may be attributed to both mica or illite, as the similarity of their basal spacings does not allow easy distinction. In general, the strength and sharpness of the reflection is more suggestive of well-crystallized mica than of illite. Yet, the strong sharpening of the 10 Å peak after heating of samples at 550°C may indicate dehydration of illite or 10 Å halloysite, although this effect also applies somewhat to micas. Therefore, the possibility that both mica and illite may be present in the glacial debris cannot be ruled out.

Comparison with bedrock mineralogy

Clay-size mineral species identified in pulverized rock samples from the study area are listed in Table 5.2. Interpretation was chiefly based on XRD analysis using the criteria described above. Glass mounts prepared with 1-300 μm size sediment did not allow a very accurate compositional interpretation because strong quartz-feldspar reflections masked those of finer-grain components. To compensate for this, XRD results were verified by petrographic examination of thin sections prepared from the rock samples. This procedure allowed positive identification of chlorite and of some mica species, and provided information on the state of alteration of the mineral assemblages. Included in Table 5.2 for comparison are unpublished results of XRD analyses conducted by DiLabio and Shilts on *KT* rock samples (numbered 77-BR) collected in the study area by H.R. Balkwill.

Archean-Aphebian terrane. Jackson and Davidson (1975) described the banded migmatites outcropping on the Precambrian crystalline terrane as dominated by granitic to intermediate gneisses. The most common mineralogical assemblage includes potassium feldspar and plagioclase (frequently combined), and biotite, hornblende, garnet and clinopyroxene as major mafic components. This bulk composition is found in samples BYR-91-01 to BYR-91-03 and in sample BYR-92-01, which may be regarded as representative of the banded migmatites. XRD analysis established that biotite is by far the most abundant phyllosilicate mineral in the $< 2 \mu\text{m}$ size range. Muscovite was identified in thin sections and could either result from weathering of primary biotite, or be associated with the regional retrogressive metasomatism to greenschist facies reported by Jackson and Morgan (1978).

Table 5.2. X-ray diffraction mineralogical analysis of bedrock samples			
Sample Number¹	Rock type (Unit²)	Size fraction analyzed	Principal mineral species Phyllosilicates: major, minor /Others
BYR-91-01 BYR-91-02 BYR-91-03	Biotite-feldspar gneiss (mg)	[< 2 μ m] and [1-350 μ m]	Biotite, kaolinite, muscovite, smectite, illite (?), mixed-layer clays (?), quartz, plagioclase, minor k-feldspar
BYR-92-01	Bio-Gt-Opx gneiss (mg)	[1-350 μ m]	Biotite, chlorite-kaolinite, quartz, plagioclase
BYR-92-02	Epidotized gneiss (mg)	[1-350 μ m]	Chlorite, muscovite, quartz, plagioclase, k-feldspar, spinel (magnetite)
BYR-92-03	Quartz arenite (Hassel Fm)	[1-350 μ m]	Mica, quartz, minor feldspars, spinels (magnetite)
77-BR-27 77-BR-03	Quartz arenite (Hassel Fm)	[< 2 μ m]	Kaolinite, quartz
77-BR-28	Fine-grained sst (Kanguk Fm)	[< 2 μ m]	Vermiculite and/or chlorite-smectite, kaolinite, quartz, orthoclase
BYR-92-04 BYR-92-05 BYR-92-06	Laminated siltstone (Eureka Sd Fm)	[< 2 μ m] and [1-350 μ m]	Smectite, biotite, glauconite (?), chlorite-kaolinite, quartz, feldspars, magnetite
77-BR-23	Sandstone (Eureka Sd Fm)	[< 2 μ m]	Smectite, quartz, feldspars
¹ Location map: Fig. 3.3. ² Units designation: Fig. 2.3			

This is supported by the widespread occurrence of epidote and chlorite as pervasive alteration or fracture fillings of potassium feldspar-rich granitic gneisses (DiLabio and Shilts 1979). Evidence of this alteration is displayed in sample BYR-92-02 which contains abundant well-crystallized to microcrystalline chlorite accompanied by minor muscovite (Plates 15 and 16).

More uncommon is the presence of kaolinite and minor expansible clays in the mineralogical suite of Archean gneisses. These unstable, low-temperature minerals are very unlikely to occur as primary constituents of high-grade crystalline rocks. Alternatively, they may be derived by authigenic alteration of alkali feldspars and aluminosilicates such as primary biotite and metasomatic muscovite. Examination of thin sections from samples BYR-91-01 and BYR-91-03 reveals some degree of alteration affecting feldspars and to a lesser degree other silicates (Plate 17). The alteration is more accentuated in mylonitized zones, where it forms aggregates of small micaceous crystals, unidentifiable with a petrographic microscope (Plate 18). Alteration minerals may consist of illite (sericite) or nontronite, which are produced from acid rocks under moderate weathering conditions. In more alkaline and magnesium-rich lithologies, kaolinite, beidellite or montmorillonite would form preferentially (Keller 1962, Evans 1992).

Cretaceous-Tertiary terrane. Sedimentary strata exposed in the study area include siliciclastic rocks of varied texture and composition. Sample BYR-92-03 is a very pure, quartzose sandstone that was collected from a narrow exposure of the Cretaceous Hassel Formation underlying the Kanguk Formation mudstones. The high degree of sorting and roundness of quartz grains (Plate 19) reflects the fluvial-deltaic environment in which the unit was deposited (Miall *et al.* 1980). XRD analysis of the pulverized sample showed virtually no other components than quartz and some minor feldspar and mica. The latter could not be identified in thin section, but is most likely biotite, although illite could be present also.

The Eureka Sound Formation is represented by samples BYR-92-04 to BYR-92-06, which consist of finely laminated siltstones and fine-grain sandstones showing a highly immature texture and composition. Angular quartz and feldspar grains are largely

dominant, with garnet, magnetite, and amphiboles as accessory constituents. Glycolation revealed the presence of abundant smectite and possibly other expansible clays. These minerals are most likely syngenetic or diagenetic alteration products and may represent a major constituent of the argillaceous matrix of the siltstones (Plate 20). Mica occurs as well, and was identified in thin sections as fine-grained biotite. Glauconite, a common clay mineral in marine sediments, has been reported in siltstones and mudstones of the Eureka Sound Formation and may contribute to the basal 10 Å reflection attributed to mica and illite (Miall 1986).

A small 14 Å peak indicates that some chlorite is present, although in considerably lesser amounts than in the crystalline rock samples. Chlorite is probably mostly derived from the crystalline parent rock, but some may have been formed diagenetically during sedimentation: chlorite has been reported as a pore cement in quartzitic sandstones where it is neoformed by recrystallisation of Mg-Al-rich gel in pore spaces (Velde 1985). In Alaska, soils of the dry tundra zone contain primary chlorite and kaolinite derived from underlying sandstones of Mesozoic to Tertiary age, as well as mica and slightly weathered illite showing some interstratification (Campbell and Claridge 1992). Kodama and Foscolos (1981) have also reported the presence of relict chlorite in young skeletal soils developed from quartz-mica-schist on a recently deglaciated region of maritime Antarctica.

In essence, results of the analyses of Bylot Island glacial debris samples appear to support the original contention of DiLabio and Shilts (1979) that:

"The probable source for the expansible materials is the Cretaceous-Tertiary bedrock that underlies the sample sites at 'Camp' (Ujarasukjuit) and Aktineq Glaciers. The well-crystallized chlorites and micas are typical for tills derived from crystalline basement

rocks of the Canadian Shield and are most likely derived by glacial comminution of phyllosilicates eroded from the crystalline Precambrian substrate in the upper reaches of 'Camp' (Ujarasukjuik) and Aktineq Glaciers'.

(DiLabio and Shilts 1979, p. 148)

However, the presence of expansible clays in gneissic samples of the Precambrian basement indicates that this statement should be regarded with caution, especially in view of the small number of bedrock samples examined, and of the limitations of XRD analysis. The occurrence of secondary minerals in crystalline basement rocks of Bylot Island will be discussed in greater detail in Chapter 6.

5.2) Textural characteristics of debris

The considerations that follow are based on granulometric data given in Appendix II. Results of the dry weight grain size distribution analyses of debris from Aktineq and Ujarasukjuik Glaciers are plotted graphically on Fig. 5.2. Table 5.3 summarizes these results. Precision of figures is about ± 0.01 %.

In general, grain-size distribution in basal debris from Aktineq and Ujarasukjuik Glaciers is comparable. The debris is essentially coarse and consists mostly of sand-size or coarser grains ($> 74 \mu\text{m}$), as is shown by the cluster of points about the summit of the ternary diagram on Fig. 5.2. Basal debris from Ujarasukjuik Glacier appears to offer a wider range of textures with larger amounts of silty material (< 200 mesh) than does Aktineq Glacier.

This is particularly obvious in samples obtained from vertical sections, in which the silt-size fraction accounts for up to 97 % of the weight of debris. However, because the

weight of debris mostly reflects that of the coarse fraction, these results may not be representative of the true grain-size distribution in the basal debris load of glaciers.

Table 5.3. General textural characteristics of basal debris in Aktineq and Ujarasukjuik Glaciers					
Weight % of size fraction	Number of samples	Minimum	Maximum	Mean	Standard deviation
<i>Aktineq Glacier - Eastern margin - Long profile*</i>					
Wt % [$<2\ \mu\text{m}$]	40	0.08	5.53	3.32	1.68
Wt % [$2\text{--}74\ \mu\text{m}$]	40	1.56	11.19	4.49	2.23
Wt % [$>74\ \mu\text{m}$]	40	86.34	97.59	92.19	2.57
<i>Ujarasukjuik Glacier - Eastern margin - Long profile*</i>					
Wt % [$<2\ \mu\text{m}$]	25	0.00	27.55	7.77	6.60
Wt % [$2\text{--}74\ \mu\text{m}$]	25	1.01	56.46	14.24	13.03
Wt % [$>74\ \mu\text{m}$]	25	15.99	98.81	77.99	18.87
<i>Ujarasukjuik Glacier - Eastern margin - Vertical sections**</i>					
Wt % [$<2\ \mu\text{m}$]	55	0.00	27.49	3.41	4.88
Wt % [$2\text{--}74\ \mu\text{m}$]	55	3.56	97.10	21.72	15.12
Wt % [$>74\ \mu\text{m}$]	55	0.00	94.89	75.02	17.45
*Lowermost debris layers exposed on ice margin; **Dispersed facies only					

To provide further insight on the textural nature of basal debris, the $< 74\ \mu\text{m}$ fraction of selected samples was compared with that of pulverized bedrock specimen using laser scanning analysis. Figures 5.3a to 5.3c show compared results for basal debris samples of Aktineq (AK-91-21 and -45) and Ujarasukjuik Glaciers (B7-92-22, VS1-B,

Fig. 5.2. Textural characteristics of basal debris from Aktineq and Ujarasukju'it Glaciers: relative weight % of three size fractions.

Key to symbols

- Aktineq Glacier - E margin
Longitudinal profile (AK series)
- Ujarasukju'it Glacier - E margin
Longitudinal profile (B7 series)
- △ Ujarasukju'it Glacier - E margin
Vertical profiles (VS series)

VS1J and VS6-E), and for rock samples (BYR-91-01, BYR-91-03 and BYR-92-04). Particle-size distributions in the 2-300 μm range are displayed as probability volume density curves (local and cumulative). Finer particles are underrepresented on this type of graph because of their small volume; therefore their estimated relative abundance should be regarded as minimal.

Results clearly indicate that basal debris (Fig. 5.3a and b) contains significantly higher amounts of fines than the pulverized rock samples (Fig. 5.3c). The texture of basal debris is remarkably uniform over the 2-300 μm range and approaches a normal grain-size distribution with a mode around 20 to 30 μm (very fine sand to silt). The asymmetry of the size distribution curves can be accounted for by the cut at 74 μm resulting from sieving. Small shoulders to the right of the curves are due to the presence of particles larger than 74 μm that passed through the sieve. The textural predominance of fine sand and silt in this size range is widely recognized in sediment formed by glacial comminution, and represents the terminal grain size mode of minerals such as quartz, feldspars, garnet and amphiboles derived from the eroded substrate (Dreimanis and Wagner 1971).

The overall higher content of fines and the modal grain size distribution in basal debris relative to pulverized rock samples indicate that the former have undergone a much higher degree of wear than can be simulated by the grinding mill. This agrees well with an origin for debris in the basal traction zone of glaciers where abrasion and attrition are the dominant processes of comminution (Haldorsen 1981).

However, there is no major textural difference between samples such as AK-91-15, collected in the lowermost part of the basal zone (Fig. 5.3a), and samples VS1-J and VS6-E (Fig. 5.3b) which were obtained several metres above the glacier bed. Thus, it appears that the latter have been entrained from an initial subglacial position up to their actual

Fig. 5.3a
Particle size distribution in basal debris
from Aktineq and Ujarasukjuit Glaciers
Range 2 - 300 μm

Fig. 5.3b
Particle size distribution in basal debris
Vertical profiles -Ujarasukjuit Glacier
Range 2 - 300 μm

Fig. 5.3c
Particle size distribution
in pulverized bedrock samples
Range 2 - 300 μm

location within the basal zone. Despite the small number of samples analyzed, a similar origin may be inferred for the silty sediment found in discrete layers of the dispersed facies on Ujarasukjuut Glacier.

5.3) Semi-quantitative XRD analysis

Various methods exist for semi-quantitative analysis of XRD data of clay-size sediment. When high analytical precision is available, it is customary to use peak areas or peak heights corrected for x-ray absorption and mineral crystallinity in order to calculate relative weight percentages of those mineral species (Griffin 1971, Carroll 1970, Brindley and Brown 1984). In view of instrumental limitations, a simple method of peak intensity measurements was adopted for this study, based on previous work by DiLabio and Shilts (1979). Five intensity ratios were calculated for major diagnostic peaks represented in most samples after glycolation:

<i>XRD peak intensity ratio</i>	<i>Relative abundance ratio of mineral species</i>
17 Å : 14 Å	Expansible clays vs Chlorite
17 Å : 10 Å	Expansible clays vs Micas-Illite
17 Å : 7 Å	Expansible clays vs Chlorite-Kaolinite
14 Å : 10 Å	Chlorite vs Micas-Illite
10 Å : 7 Å	Micas-Illite vs Chlorite-Kaolinite

The Carleton University XRD facility does not provide diffractograms with a flat reference base line, as some instruments do. Instead, the printout displays a sharp rise in the low angle region and a broad swell centered below the 3.35 Å quartz peak (*re* Fig. 5.1a). To measure relative reflection intensities, a smooth, curving base line was drawn on

each diffractogram along the base of the peaks. As each sample gives a different diffraction pattern, the exact shape of the base line is somewhat variable and may introduce a nonsystematic source of error in peak height measurements. In order to minimize this error, base lines on all diffractograms were drawn so that they consistently correlated at their maxima, minima, and inflexion points (Fig.5.4). Peak heights were measured relative to base line with a precision of ± 0.05 cm, which translates to a ± 0.01 unit precision for calculated intensity ratios. Appendix III lists values of all ratios computed for clay-size debris samples collected in the study area, with specifications regarding pre-analytical treatments and diffractometer settings.

Estimated precision and significance of results

Segall *et al.* (1987) have identified various sources of error involved in semi-quantitative XRD analysis, some of which apply to this work: *i)* instrumental error; *ii)* peak variations resulting from changes in humidity and temperature during analytical procedures; *iii)* insufficient testing of the consistency of results; and *iv)* human error in estimation of peak intensities. In the given context, error related to instrumentation limitations could not be controlled, but was assumed to be minimal. Reproducibility of results was tested by performing three successive XRD runs on some representative clay samples, as described earlier. An overview of results is given in Table 5.4.

The maximum variance of peak intensities (and peak ratios) for glycolated samples is about twice that of untreated samples; this can be attributed to the difficulty of duplicating the same degree of glycol saturation in samples in spite of the consistency of the procedure.

Fig. 5.4. Correlation of x-ray diffractograms
for semi-quantitative variation analysis.

Glycolated samples ($< 2 \mu\text{m}$ fraction)
Monochromatic $\text{CuK}\alpha$ radiation; $2^\circ 2\theta / \text{min}$.

*Base Lines are correlated at their minima,
maxima, and inflexion points (dashed lines).*

Table 5.4. Estimated error of semi-quantitative XRD analysis				
	<i>Untreated samples</i>		<i>Glycolated samples</i>	
	Max. standard deviation (σ)	Maximum error (%)	Max. standard deviation (σ)	Maximum error (%)
AK-91-17 (3 runs)				
Peak heights	0.13	3.72	0.27	6.51
Peak height ratios	0.02	3.53	0.08	8.64
B7-92-09 (3 runs)	Max. standard deviation (σ)	Maximum error (%)	Max. standard deviation (σ)	Maximum error (%)
Peak heights	0.05	3.98	0.66	7.40
Peak height ratios	0.02	5.27	0.11	13.36
B7-92-18 (3 runs**)	Max. standard deviation (σ)	Maximum error (%)	Max. standard deviation (σ)	Maximum error (%)
Peak heights	4.00	53.37	9.93	43.07
* Error percentages are based on standard deviation relative to average values. ** Three mounts with a different thickness of the same clay slurry.				

For sample B7-92-18, analyses were performed on three mounts: A, B and C, prepared with different concentrations of clay slurry such that $[B] = 2[A]$, and $[C] = 3[A]$. The wide deviation in results indicates that XRD peak intensities are very sensitive to the thickness of clay applied on the glass slide. In view of the careful control exerted during preparation of the clay mounts, it is unlikely that variations of the magnitude shown in sample B7-92-18 might occur for most specimens analyzed. If error due to inconsistency in the preparation steps can be assumed to be minimal, the range of variation on peak ratios for three analytical runs of a given glycolated sample is less than 15 %.

Semi-quantitative estimates of the amounts of clay minerals have been stated by various investigators to be probably accurate to about 5-10 % (Carroll 1970). Dalrymple and Maas (1986) have reported an analytical precision of ± 5 %, while Segall *et al.* (1987)

have obtained errors of up $\pm 40\%$ by weight, although on average considerably less. The estimated $< 15\%$ error on results thus seems to compare well with those obtained by other methods. However this may not be conclusive, as duplicate analyses were only performed on a small set of samples.

As a means of further testing the significance of data, variations of selected XRD peak intensity ratios in two series of samples (AK-91 and B7-92 series) were compared prior to and after glycolation (Fig.5.5a and 5.5b). Results indicate that glycolation does not significantly change the general pattern of XRD peak ratio variations between samples. The most noticeable effect is the overall reduction of amplitude of the 14:10 Å ratio, which can be attributed to the reduced intensity of the 14 Å peak due to expansion of smectite. On these grounds, and based on estimates of precision shown in Table 5.4, it is assumed that the reproducibility of XRD semi-quantitative analysis is sufficient to allow interpretation of results with a reasonable level of significance.

5.4) Compositional and textural variations in the basal zone

On Figures 5.6a and 5.6b, textural variations in basal debris along the eastern margins of Ujarasukjuit and Aktineq Glaciers are plotted with respect to down-ice distance from stations AK-91-54 (Aktineq) and B7-92-01 (Ujarasukjuit), respectively. Grain-size distribution curves were smoothed with a three-point moving average. Correlation with bed topography and bedrock geology along the ice margin is shown with reference to the assumed position of the faulted AA/KT geological contact. Unit symbols are defined as on Fig. 2.3, and attitude of the contacts is approximated.

The most obvious pattern of textural variations is a gradual down-ice enrichment in fines ($< 74\ \mu\text{m}$) observed on Ujarasukjuit Glacier (Fig.5.6b). This enrichment appears to

Fig. 5.5a.
Compared XRD peak ratio variations:
untreated vs glycolated samples
Aktineq Glacier - Eastern margin profile

Fig. 5.5b.
Compared XRD peak ratio variations:
untreated vs glycolated samples
Ujarasukjuit Glacier - Eastern margin profile

coincide with the passage of the glacier onto the Cretaceous sandstones of the Kanguk Fm, although the textural nature of basal debris over the crystalline terrain is not well-documented. No clear correlation with bed topography can be established over that part of the margin covered by the transect.

Aktineq Glacier (Fig.5.6a) shows a comparatively more complex pattern. No definite trend is observed in the coarser ($>74\ \mu\text{m}$) fraction of the debris load, which varies widely and unsystematically. However, the clay ($< 2\ \mu\text{m}$) and fine ($2-74\ \mu\text{m}$) fractions display variations that may be related to the nature of the underlying substrate. A noticeable increase in the proportion of clay-size debris occurs southwest of the AA-KT geological contact in the upper part of the profile. This trend is interrupted about 2 km down-ice from station AK-91-54, where the relative weight of clay-size debris drops abruptly by about 3 %. The initial weight increase in clay may be related, as for Ujarasukjuit Glacier, to the passage of the glacier from the crystalline basement rocks to the soft sedimentary strata of the Cretaceous Kanguk Fm. Similarly, the subsequent decline can be correlated to the transition between the lower mudstone and upper sandstone member of the Kanguk Fm. Interestingly, this transition also coincides with a peak in the relative weight of the $2-74\ \mu\text{m}$ fraction. The Kanguk Fm sandstones are lithologically similar to those of the Eureka Sound Fm which are for the most part medium- to coarse-grained (Miall *et al.* 1980). Therefore it is likely that glacial comminution of these rocks would yield lesser amounts of clay-size material than would mudstones. The relatively high proportion of clay in basal debris over the upper part of the profile might reflect the presence of abundant phyllosilicates derived from the volcano-sedimentary units that outcrop on the eastern margin of Aktineq Glacier.

Figures 5.7a and 5.7b show down-ice variations in the clay mineralogy of basal debris along Aktineq and Ujarasukjuit Glaciers, as reflected by the calculated XRD peak

intensity ratios. As before, the smoothed curves are correlated with bedrock topography and geology with reference to the AA-KT faulted contact. It can be observed that the 17:14 Å, 17:10 Å and 17:7 Å peak ratios are comparatively greater in basal debris from Aktineq Glacier (Fig. 5.7a) than in that of Ujarasukjuit Glacier (Fig. 5.7b). This indicates that while the relative proportions of chlorite-kaolinite and mica-illite are essentially similar in both glaciers, the amount of expansible clays is significantly greater in Aktineq Glacier than in Ujarasukjuit Glacier. This is in accordance with results reported by DiLabio and Shilts (1979) for debris layers in four ice profiles studied on those glaciers.

Surprisingly, however, ice marginal profiles on both Aktineq and Ujarasukjuit Glacier indicate a down-ice decrease of the proportion of expansible minerals in the basal debris load away from the AA-KT contact. This trend is more noticeable on Ujarasukjuit Glacier than on Aktineq Glacier, which suggests a local lithological control on the clay mineralogy of debris in the glaciers. On Aktineq Glacier, the passage over the KT terrane is initially marked by a sharp increase in the proportion of expansible clays (and to a lesser degree of mica-illite) in the basal load. A second peak, of lower amplitude, occurs 1 km down-ice of the contact, followed by a marked decline which brings the proportion of 17 Å clays below its initial level.

The first increase is conceivably related to the erosion and entrainment of smectite- and illite-rich clays from the underlying soft Kanguk Fm mudstones. The second peak may result from a local increment in subglacial debris accretion in response to compressive basal ice flow over the steeper bed gradient. This compositional shift does not find a corresponding expression in textural variations of the basal debris, but the two parameters do not necessarily correlate. The down-ice decline of abundance in 17 Å clays over the lower part of the profile may be linked to the passage of the glacier over the coarser

Fig. 5.6a
Textural variations in basal debris load
vs bed geology and relief
Aktineq Glacier - Eastern margin profile

Fig. 5.6b
Textural variations in basal debris load
vs bed geology and relief
Ujarasukjuit Glacier - Eastern margin profile

AA KT (Geological symbols defined as on Fig. 2.3.)

Fig. 5.7a
Mineralogical variations in basal debris load
vs bed geology and relief
Aktineq Glacier - Eastern margin profile

Fig. 5.7b
Mineralogical variations in basal debris load
vs bed geology and relief
Ujarasukjuit Glacier - Eastern margin profile

sandstones of the Kanguk and Eureka Sound Formations. This could also account for the configuration of the 17:14 Å profile on Ujarasukjuik Glacier, which indicates a relative impoverishment in expansible clays over the Eureka Sound Fm sandstone beds. The abundance of 17Å clays in basal debris up-ice of the *AA-KT* contact on both Aktineq and Ujarasukjuik Glaciers is suggestive of the presence of a subglacial debris source for these minerals over the Precambrian terrane. Expansible clays were also identified in supraglacial debris on Aktineq Glacier, and in lateral moraine debris from B-28 and Sermilik Glaciers¹, which are both supposedly underlain by Precambrian crystalline bedrock only. This will be discussed in greater length in Chapter 6.

The seven vertical ice profiles studied on the eastern margin of Ujarasukjuik Glacier are displayed in Appendix IV. Each is represented by a schematic stratigraphic column (scaled in metres) showing the different basal facies observed and the positions of the debris-rich layers sampled (bold capitals). A key to symbols used in the stratigraphic sections is provided. Variations in the clay mineralogy and texture of debris-rich layers in each profile are indicated by a series of scaled vertical graphs.

Comparison of the different profiles is made on Figures 5.8a to 5.8e. Vertical sections are represented with respect to their position along the eastern margin of Ujarasukjuik Glacier and the assumed location of the *AA-KT* geological contact (bold dotted line). The lower margin scale gives down-ice distance from station VS1. Correlation of the profiles is made using the basal zone-englacial zone contact as a reference datum. Compositional variations of debris in the profiles (XRD peak ratios - top scale) are plotted against the weight percentage of the <2 µm fraction (bottom scale).

¹ See original diffractograms for glycolated samples AK-91-56, B28-92-47, and 78GS0066, respectively.

The range of values for any given XRD peak ratio varies considerably in different profiles, the largest amplitudes being associated as before with the expansible clays to chlorite (17Å:14Å) and mica-illite to chlorite-kaolinite (10Å:7Å) ratios. There does not appear to be any consistent vertical pattern of increase or decrease in clay mineral abundances throughout the basal zone. This observation also applies to textural variations within the sections.

However, if profiles are compared with respect to their relative positions along the ice margin, some degree of correlation between clay mineralogy and texture in debris layers can be observed in those profiles located down-ice of the *AA-KT* contact. In profiles VS5, VS8 and VS9, variations in the relative amount of expansible clays seem to match that of the clay-size fraction of the debris. This tendency is reflected in the 17Å:14 Å ratio, but does not seem to hold for the 17Å:10Å ratio. There also appears to be a slight increase in the overall relative amount of expansible minerals from profile VS3 to VS9, but this increase is within the range of variations of individual profiles and thus may not be significant. Positive correlation is also apparent between the 14Å:10Å ratio and the proportion of clay-size debris in profile VS7.

These observations may possibly reflect a gradual down-ice enrichment of 17Å clays in debris layers of the dispersed facies resulting from progressive smectite-rich clay enrichment of the debris load over the *KT* substrate. The fact that profile VS7 does not conform to this trend could be attributed to the presence of far-travelled sediment largely derived from the crystalline basement rocks in the upper layers of the basal zone. This could also account for the positive correlation between the proportion of chlorite (14Å:10Å ratio) and that of fines displayed in the profile.

However, in view of the important variability of results in other profiles, this interpretation remains questionable. In general, it rather seems that composition of debris layers in the dispersed facies does not noticeably reflect that of the underlying substrate, and the range of variations from profile to profile is less considerable than that observed in the lowermost basal debris load. Implications of these observations for subglacial debris entrainment and transport will be considered in Chapter 6.

5.4) Co-isotopic signature of basal ice.

Results of the co-isotopic analysis of basal ice from Ujarasukjuit Glacier are given in Appendix V. Analytical error is ± 0.10 ‰ for $\delta^{18}\text{O}$ and ± 1.3 ‰ for δD ; values are given with according numerical precision.

Estimation of the $\delta^{18}\text{O}$ - δD signature for englacial ice

Detection of freezing effects by co-isotopic analysis requires that the $\delta^{18}\text{O}$ - δD signature of basal ice be compared to that of englacial ice. Because it is mostly controlled by the precipitation effect, the $\delta^{18}\text{O}$ - δD relation in englacial ice conforms to the linear equation $\delta\text{D} = b\delta^{18}\text{O} + a$ and is usually characterized by a slope close to 8. An approximation of the linear δD - $\delta^{18}\text{O}$ signature of englacial ice for the site was calculated using data from a 5.5 m firn core extracted with an auger drill from the higher parts of the Bylot Island icefield during the 1992 summer field season (Drill site location shown on Fig. 2.2). The top 40 cm of the core were sampled at 5 cm intervals and analyzed for $\delta^{18}\text{O}$ - δD by M.Elver (Elver 1994) with the same precision as that obtained for basal ice samples.

Results (Appendix V) range from -22.40 ‰ to -25.80 ‰ for $\delta^{18}\text{O}$ and from -173.3 ‰ to -189.6 ‰ for δD , with an average composition of $\delta^{18}\text{O} = -24.24$ ‰ and $\delta\text{D} = -180.2$ ‰. These figures may not be representative of the entire firn zone, as they reflect the isotopic signature of firn which probably formed in the late spring when surface temperatures are relatively high. Depleted $\delta^{18}\text{O}$ values in the -30 ‰ to -36 ‰ range associated with winter precipitation are found in deeper core layers (M. Elver, personal communication). However, unless intense evaporation or sublimation took place in the accumulation area prior to compaction, the isotopic signature of the upper firn layers should essentially reflect the precipitation effect. This was verified by applying regression analysis to the firn core data. Using the least-squares method, the following linear relation was determined:

$$\delta\text{D} = 7.8(\delta^{18}\text{O}) + 10.6 \text{ ‰} ; r^2 = 0.94 \quad (5.1)$$

To test the validity of this relation, an estimate of the $\delta^{18}\text{O}$ - δD signature of precipitation in the accumulation zone of nearby glaciers was computed using data obtained by F.A. Michel from the Pond Inlet weather station located about 50 km southeast of the sampling site across Eclipse Sound. The data covers a 20 month period (July 1990-Feb.1992) and provides the $\delta^{18}\text{O}$ - δD composition of cumulated precipitation samples for each month in record (Appendix V). Least-square regression analysis of the data gave the following equations:

$$\delta\text{D} = 8.0(\delta^{18}\text{O}) + 12.0 \text{ ‰} ; r^2 = 0.99 \quad (5.2)$$

Equation (5.2) gives the $\delta^{18}\text{O}$ - δD relationship for precipitation in Pond Inlet for the 20 months of record and corresponds to the local meteoric water line. Comparison of equations (5.1) and (5.2) is made assuming that: *a*) southern Bylot Island experiences the same general weather conditions that Pond Inlet does; and *b*) the isotopic altitude effect² on precipitation falling in the central Byam Martin Mountains (elevation ≈ 1550 m a.s.l.) does not affect the $\delta^{18}\text{O}$ - δD relationship in firn collected at the drill site (although δ values may be expected to be more negative in average).

Fig. 5.9 shows the linear relations obtained from equations (5.1) and (5.2) plotted on a graph of $\delta^{18}\text{O}$ vs δD for comparison: a strong agreement between both lines is obvious. This high covariance of equation (5.1) ($r^2 = 0.96$) in view of the small sample population ($n = 8$) appears to validate it as an estimate of the $\delta^{18}\text{O}$ - δD relation in englacial ice for the study area.

Comparison of $\delta^{18}\text{O}$ - δD signatures from basal and englacial ice

On Fig. 5.10, the 15 basal ice samples collected on Ujarasukjuit Glacier are plotted (with appropriate analytical error bars) on a scatter graph of $\delta^{18}\text{O}$ vs δD with the linear relation for englacial ice (*EL*) shown for comparison. All samples analyzed were collected from the dispersed basal facies with the exception of VS9-A (circled) which was obtained from a small exposed section of the stratified facies.

Basic statistical parameters of $\delta^{18}\text{O}$ and δD values for the sample population are given in Table 5.5. These figures indicate that the average $\delta^{18}\text{O}$ - δD composition of basal ice samples from Ujarasukjuit Glacier (-21.41 ‰; -170.6 ‰) is isotopically enriched in ^{18}O and D by 3 ‰ and 10 ‰ respectively, relative to the mean isotopic composition of

² Estimated at 0.15 ‰ per 100 m for $\delta^{18}\text{O}$ and 1.12 ‰ per 100 m for δD by Elver (1994).

Fig. 5.9. Estimated $\delta^{18}\text{O}$ - δD signature of englacial ice (EL) in Ujarasukjuit Glacier compared to the precipitation line (PpL) in Pond Inlet, northern Baffin Island.

Fig. 5.10. : isotopic (^{18}O -D) composition
of basal ice in Ujarasukjuit Glacier.
Analytical error (2σ): $\delta^{18}\text{O} = \pm 0.10 \text{ ‰}$; $\delta\text{D} = \pm 1.3 \text{ ‰}$.

shallow firn (-24.24‰ ; -180.2‰) in the accumulation area. In fact, only the basal ice sample with the lightest isotopic content (VS10-A: -24.02‰ ; -184.5‰) approaches the composition of the upper firn layers. Since the latter can be viewed as a seasonal high value, it ensues that most basal ice samples collected on Ujarasukjuit Glacier are isotopically heavier than firn in the accumulation area and possibly heavier than the bulk of englacial ice, although glacier ice with heavier isotopic signatures may occur closer to the equilibrium line.

Table 5.5. Statistical parameters for ^{18}O - D signatures of basal ice samples from Ujarasukjuit (B-7) Glacier.		
Element	Oxygen 18 ($\delta^{18}\text{O}$)	Deuterium (δD)
Count	15	15
Minimum	-24.02‰	-184.5‰
Maximum	-19.31‰	-160.7‰
Range	4.71‰	23.8‰
Math. mean	-21.41‰	-170.6‰
Median	-21.54‰	-169.9‰
Standard deviation σ	1.18‰	6.14‰
Variance σ^2	$1.40 (\text{‰})^2$	$37.7 (\text{‰})^2$

The possibility that such an enrichment may represent an altitude effect can be ruled out for simple reasons: englacial ice in the deeper parts of the glacier most likely originates from a point located near the accumulation zone and should therefore display an isotopic signature comparable to or lighter than that of the upper firn zone. This is obviously not the case in basal ice samples of Ujarasukjuit Glacier. Isotopic exchange between ice and free

water during snow metamorphism in the accumulation zone may lead to a post-depositional ^{18}O -D isotopic enrichment in firn layers of the same magnitude as that observed (Arnason 1969). However this applies mostly to temperate glaciers for which meltwater production and rainfall are intense during the melt season. Such is not the case for sub-polar glaciers of Bylot Island.

Another effect attributed to isotopic exchanges with snow and firn involves the smoothing of seasonal δ variations in ice below the uppermost winter layer. This effect was not observed in the $\delta^{18}\text{O}$ variations of the 5.5 m firn core extracted from the Bylot Island icefield. Smoothing of the isotopic record might occur at greater depths in the firn zone, but it is very unlikely: although the core was extracted in late July at the height of the melt season, temperature in the borehole was so low it caused the drill core barrel to freeze onto the ice.

A third possibility is that heavy isotope enrichment in dirty ice layers of the basal zone results from a process involving melting and refreezing at the glacier sole, as proposed by Souchez and Jouzel (1982). To test this hypothesis, least-squares regression analysis was applied to the basal ice data, and the following linear relation was found:

$$\delta\text{D} = 4.4(\delta^{18}\text{O}) - 76.3 \text{ ‰} \quad r^2 = 0.72 \quad (5.4)$$

The 4.4 slope of the relation falls within the range of reported values for basal ice in glaciers of the northern and southern hemispheres (see Chapter 6). This suggests that the relation may indeed represent a freezing effect. However, in view of the large scatter of data ($r^2 < 0.8$), further statistical testing of the relationship is required.

Statistical validation of the freezing line

On the suggestion by R.Souchez (personal communication), the validity of the freezing line relation in basal ice was assessed by calculating the *theoretical freezing slope* of the data in accordance with the Jouzel and Souchez model for a closed system situation, and testing its value against a 90%-level interval of confidence. The theoretical freezing slope B represents the predicted linear relation between the $\delta^{18}\text{O}$ and δD values of ice derived by progressive freezing of an initial water source of composition ($\delta^{18}\text{O}_i$; δD_i). The value of B may be calculated from:

$$B = \left(\frac{\alpha - 1}{\beta - 1} \right) \left(\frac{1 + \delta\text{D}_i}{1 + \delta^{18}\text{O}_i} \right) \quad (5.5)$$

where α and β are the isotopic diffusion coefficients between ice and water for deuterium (0.0208) and ^{18}O (0.003), respectively (Souchez and Jouzel 1984). The initial water composition ($\delta^{18}\text{O}_i$; δD_i) is obtained by resolving the equation system given by the $\delta^{18}\text{O}$ - δD linear relation in englacial ice (equation 5.1) and that of basal ice (equation 5.4).

However, the latter may not accurately represent the true $\delta^{18}\text{O}$ - δD relation in basal ice, as the least-squares regression algorithm does not account for the covariance of $\delta^{18}\text{O}$ and δD . Therefore, for the purpose of testing the validity of the relation, a least-squares cubic *functional fit* was used in addition to the least-squares fit, as recommended by Burn and Maxwell (1993). Parameters for the functional fit (slope and δD intercept) were derived from those of the least-squares approximation using the algorithm of Church and Mark (1980) which accounts for the analytical error on $\delta^{18}\text{O}$ (± 0.10 ‰) as well as δD (± 1.3 ‰).

As most research relating to glaciological co-isotopic problems applies the least-squares regression method, the theoretical freezing slope β was calculated from both the functional and least-squares fits to allow comparison of results with those of others. Upper and lower confidence limits on β were calculated at the 90 % level for the least-squares and functional fits using algorithms given by Younger (1979) and Church and Mark (1980). Results are shown below.

Table 5.6. $\delta^{18}\text{O}$-δD relationships in basal ice from Ujarasukjuik Glacier			
Linear fit (Method)	Equation $\delta\text{D} = b\delta^{18}\text{O} + a$	90 %-level confidence limits for slope $[b_L, b_U]$ *	Theoretical freezing slope B_f *
Least-squares fit	$\delta\text{D} = 4.4\delta^{18}\text{O} - 76.3$	$[b_L = 3.9 ; b_U = 4.9]$	$B_f = 5.8$
Functional fit*	$\delta\text{D} = 4.6\delta^{18}\text{O} - 72.1$	$[b_L = 3.9 ; b_U = 6.5]$	$B_f = 5.8$
* Plotted through centroid $[-21.41 \text{‰}; -170.6 \text{‰}]$ on Fig. 5.11.			

Figures above indicate that the least-squares and functional fits diverge only by a minor difference (0.2) of their freezing slope. The 90%-level interval of confidence for the slope of the relation is wider on the functional fit because it accounts for variance on both $\delta^{18}\text{O}$ and δD , as opposed to the least-squares method which attributes total variance of data to δD . The divergence between the theoretical freezing slope (5.8) and that of the linear approximations (4.4 - 4.6) can be partly attributed to the large scatter of data, but probably also reflects a complex history of freezing effects in the basal ice layers.

Confidence testing of the functional fit slope is illustrated graphically on Fig. 5.11. All lines are plotted through the central mean value of the data cluster (the *centroid*) to facilitate comparison. It can be seen that the calculated theoretical slope B_f falls within the confidence limits established at the 90 % confidence level ($3.9 < 5.8 < 6.5$).

Fig.5.11.
 Ujarasukjuik Glacier basal ice.
 Statistical validation of the $\delta^{18}\text{O}$ - δD freezing slope.
 b_f : slope of least-square cubic functional fit (FF)
 B_f : calculated theoretical freezing slope
 $[b_U, b_L]$: 90%-level confidence limits on B_f

Thus, the $\delta^{18}\text{O}$ - δD signature of debris-rich basal ice at Ujarasukjuik Glacier may indeed represent the effects of a freezing process. This interpretation applies essentially to ice collected within the dispersed facies. No discernable change in the freezing slope or the confidence limit values are obtained if sample VS9-A (stratified facies) is not accounted for in the calculations.

Interpretation of the freezing line

Fig. 5.12 shows the $\delta^{18}\text{O}$ - δD signature of basal ice at Ujarasukjuik Glacier in relation to the least-squares linear fit and the Englacial ice Line (*EL*). On the graph, all basal ice samples plot to the right of the *EL* and above its intersection with the linear fit. The intercept of these two lines represents the average composition of the water source from which ice in the debris-rich layers was derived. Resolving the equation system of the *EL* and linear fit for $\delta^{18}\text{O}$ and δD at the intercept gives the following solution:

$$\delta_i^{18}\text{O} = -25.56 \text{ ‰} ; \delta_i\text{D} = -188.8 \text{ ‰} \quad (5.6)$$

This composition is remarkably close to the average for the upper firm layers in the accumulation area (-24.24 ‰; -180.2 ‰), and probably approaches the mean isotopic composition of englacial ice for Ujarasukjuik Glacier. This approximation does not account for the possible occurrence of isotopically lighter or heavier ice reflecting colder or warmer climatic episodes deeper in the glacier. As no information is available on the potential preservation of such a signal in glacier ice on Bylot Island, the possibility can neither be confirmed nor denied.

Fig.5.12 indicates that all basal ice samples collected on Ujarasukjuik Glacier are to various degrees isotopically enriched in ^{18}O and D relative to their hypothetical source.

Fig. 5.12. Compared $\delta^{18}\text{O}$ - δD signatures of basal ice (*LSR*) and englacial ice (*EL*) from Ujarasukjuut Glacier.

This enrichment ranges from 1.54 ‰ to 6.25 ‰ for ^{18}O and from 4.3 ‰ to 28.1 ‰ for D. Sample VS9-A, which was collected in the stratified facies, is enriched by 3.06 ‰ in ^{18}O and 16.4 ‰ in D relative to $[\delta_i^{18}\text{O}; \delta_i\text{D}]$. The overall heavy isotopic composition of basal ice layers may be interpreted as representing the early steps in an equilibrium freezing process occurring in a closed system, i.e. without significant mixing of isotopically contrasted sources over the course of freezing. At the scale to which sampling was performed, the isotopic composition of samples most likely represents an average for multiple ice laminae, each with different $\delta^{18}\text{O}$ - δD signatures corresponding to successive steps of a freezing event. The Souchez and Jouzel model predicts that only ice laminae formed in the latter stages of freezing (after 66 % of the initial water volume has frozen) will display heavy isotope depletion relative to the initial source. It follows that the probability of obtaining an ice sample with a heavier $\delta^{18}\text{O}$ - δD signature than the source is considerably higher than that of obtaining an isotopically lighter sample. This is consistent with the high $\delta^{18}\text{O}$ - δD values observed for basal ice at Ujarasukjuit Glacier. Thus, the co-isotopic characteristics of basal ice layers on Ujarasukjuit Glacier generally appear compatible with an origin linked to freezing accretion processes at the glacier sole.

Chapter 6 Discussion

In this chapter, sedimentological and glaciological data are jointly examined and discussed to provide the empirical basis of a model of incorporation, transport, and spatial distribution of debris in glaciers of southwestern Bylot Island. The model is discussed with regards to contrasting features observed in different glaciers of the study area.

6.1) Composition and provenance of basal debris

As observed by DiLabio and Shilts (1978), coarse debris transported in large glaciers of southwestern Bylot Island appears to be composed of more than 99 % crystalline rock detritus derived from the Precambrian terrane, regardless of the distance of glacial flow over the younger lithologies of the lowlands. This probably results from the combination of two factors:

1^o Because of the marked physiographic contrast between the Bylot Lowlands and the Byam Martin Mountains, those parts of glaciers flowing over the KT substrate are probably eroding and incorporating lesser amounts of debris than those flowing over the Shield bedrock. In the upland alpine setting, colder temperatures, steeper slope gradients and greater ice velocities result in high glacial erosion rates and abundant mass wasting that deliver debris to glaciers in the form of rockfalls, medial moraines, and subglacial debris. The piedmont ice lobes flowing on the lowlands are not constricted between narrow valley walls and thus are deprived of a considerable input of debris. As well, lower ice velocities coupled with lower bed gradient and relief near the ice margin probably reduces considerably rates of subglacial erosion.

2° Poorly-consolidated sedimentary rocks of the lowlands tend to be rapidly disaggregated by glacial action into sand-size or finer components that cannot be distinguished megascopically from those of the crystalline detritus. With the exception of the quartzose sandstones of the Hassel Formation, most of the KT units exposed in the study area are texturally and lithologically immature, and contain a suite of minerals or polymineralic grains comparable to that of the Precambrian parent rocks (DiLabio and Shilts 1979, Miall *et al.* 1981, Bernier 1985). Results of the grinding mill experiment and petrographic observations presented in Chapter 5 confirm this.

These factors do not preclude the possibility of a contribution of debris derived from the KT sedimentary beds. As the smooth topography of the lowlands limits any supraglacial input, KT-derived debris would have to be chiefly incorporated in the basal zone by subglacial accretion processes such as those described in Chapter 1. DiLabio and Shilts (1979) have argued that expansible clays such as smectites found in the clay fraction of basal debris and lateral moraines may represent a contribution from the KT lithologies. This contention was based in part on the apparent negative correlation between trace metal concentrations and the relative abundance ratio of expansible clays to chlorite and kaolinite in the clay fraction of basal debris (Fig. 6.1).

XRD results discussed in Chapter 5 have shown that expansible clays are indeed major constituents of the $< 2 \mu\text{m}$ fraction of basal debris in both Aktineq and Ujarasukjuit Glaciers, as well as in the clay mineral suite of pulverized rock samples of the Kanguk and Eureka Sound Formations. However, if expansible clays were exclusively derived from the sedimentary rocks of the lowlands, they should not be found in glacial debris derived and transported from the highlands without mixing. Yet, XRD analysis has identified smectites in basal debris collected up-ice of the AA-KT contact on Aktineq and Ujarasukjuit Glaciers, and in greater relative abundance than in debris collected several km down-ice from the contact. As well, expansible clays were identified in supraglacial debris on Aktineq Glacier,

Fig. 6.1. Relation between Cr and Zn and amount of expandible minerals ($17:14 \text{ \AA}$) in clay fraction of debris sampled from four vertical ice profiles on Aktineq and Ujarasukjuit Glaciers. See Fig. 3.1 and 3.2 for location of profiles. (from DiLabio and Shilts 1979)

and in lateral moraine debris from B-28 and Sermilik Glaciers, which are both supposedly underlain by Precambrian crystalline bedrock only.

The possible occurrence of secondary minerals such as smectites or mixed-layer clays in high-grade crystalline rocks of Bylot Island is somewhat unlikely, as the extensive ice cover and the severe, cold environment which prevail today on the island are unfavourable to chemical weathering of the substrate. Yet, analyses of gneissic erratics and bedrock samples have indicated the presence of expansible minerals, although in lesser abundance than in debris samples (if the height and sharpness of the 17 Å XRD peak may be considered a valid index of relative abundance).

These minerals may indicate that the surface of the Precambrian substrate has retained weathering alteration from previous exposure to different surface conditions. Some recent paleoclimatic studies have suggested that a climate warmer and moister than present, and thus more conducive to surficial alteration, may have existed in the eastern Arctic until at least mid-Quaternary time. Morgan *et al.* (1993) inferred mean annual temperatures up to 4-5 °C higher than today based on microfaunal analysis of interglacial peat deposits in north-central Baffin Island. Buried soils and organic debris of interglacial character are also found on northeastern Baffin Island and southern Bylot Island, suggesting a period of warmer climatic conditions and a native ice cover less extensive than at present (Klassen and Shilts 1987, Klassen 1993). Thus, it is conceivable that some degree of chemical weathering might have affected rocks of the crystalline basement in the past, which could explain the presence of kaolinite and expansible clays in their mineralogical suite. This would also imply that modern glacial erosion has not yet entirely removed the surface layer of crystalline substrate affected by surficial weathering.

Expansible clays may also occur as hydrothermal or secondary alteration products along joints, fractures and in shear zones within crystalline rocks. The occurrence of montmorillonite and other swelling clays in fracture fillings has been recognized before in

Archean rocks (Brekke and Selmer-Olsen 1965). Expansible clays may be present as the end product of hydrothermal alteration of an old fault zone, and montmorillonite is often found within mylonitized zones up to several metres wide. Smectites may also occur as a result of the alteration of a diabase dyke to amphibolite and finally to clay (Barton 1974).

Alternatively, some of the alteration observed in erratics may have been inherited during erosion and subsequent transport in the ice. This may apply somewhat to debris transported supraglacially, but is unlikely to affect debris derived from the basal zone. Diagenetic pressure-solution effects related to ice load pressure have been documented (e.g. Fairchild *et al.* 1993) but only for soluble carbonate rocks in the basal zone of temperate glaciers.

The high abundance of 17Å clays in basal debris immediately up-ice of the AA-KT contact on both Aktineq and Ujarasukjuit Glaciers is suggestive of the presence of a local subglacial debris source for these minerals over the Precambrian terrane. The narrow volcano-sedimentary units that outcrop on the upper margins of Aktineq and Ujarasukjuit Glaciers (*re* Fig. 2.3) could conceivably represent such a source, as weathering of Mg-rich silicates in basalts favours the formation of chlorite-smectite, and subsequently of montmorillonite (Keller 1962, Wilson and nadeau 1985). This may explain the apparent contradiction between the low abundance of expansible clays in the mineralogical suite of the Precambrian migmatites, and their occurrence in glacial debris derived from the uplands.

Another, more potential source of smectites may be foreign drift derived from sedimentary strata of northeastern Baffin Island, Eclipse Trough or the Bylot Lowlands, and transported onto the highlands during a previous regional glacial event such as the Wisconsinan Eclipse glaciation (*re* Fig. 2.4). Klassen (1993) has reported silica-cemented quartz sandstone erratics (probably derived from the Hassel Fm) occurring at 1136 m a.s.l.

beside Sermilik Glacier, 10 km inland of the coast. Furthermore, a recent study of erratic dispersal patterns by McCuaig (1994) has shown the inland extent of foreign drift to be more considerable than previously assessed by Klassen. This may account for the presence of expansible clays in basal debris from B-26 and Sermilik Glaciers, both of which flow several km down-ice of the mapped inland extent of foreign drift.

The occurrence of Cenozoic (or older) outliers underneath the modern ice cover of Bylot Island may also be considered. Isolated outcrops of the Hassel Formation are found on the lowlands surrounding Pond Inlet on Baffin Island (Miall *et al.* 1980). In the central Baffin uplands, Andrews *et al.* (1972) have reported a Paleogene limestone outlier about 26 km north of the Barnes Ice Cap, at an altitude of 730 m a.s.l. The nearest deposits of similar age only occur 650 km to the southeast near Cape Dyer. On Bylot Island, the mapped inland extent of the *KT* terrane on the island remains somewhat approximate as it is obscured by the modern ice cover and the extensive layer of erratics and frost-shattered regolith. Consequently the possible occurrence of sedimentary outliers within the crystalline terrane of the Byam Martin Mountains cannot be ruled out.

These considerations indicate that the relative abundance of expansible clays and other mineral constituents in glacial debris may not constitute a reliable indicator of sediment provenance or source mixing for glaciers of southwestern Bylot Island. This is due to the fact that clay abundance index are mostly based on the ratio of secondary clays vs chlorite and micas. Whereas chlorite, and to a lesser degree mica and illite, are ubiquitous products of erosion and comminution in glacial deposits derived from the Canadian Shield (Dean 1969, Dreimanis and Wagner 1969, Shilts 1978), secondary clays such as smectites may reflect either a specific source provenance and/or the result of authigenetic or diagenetic processes, which renders any interpretation based on their presence (or absence) problematic. Despite these uncertainties the relatively high

concentrations of smectites in *KT* rocks can be used to ascertain the nature of basal debris entrainment processes in the piedmont glacier lobes of southwestern Bylot Island.

6.2) Patterns of basal debris entrainment

The observed textural and compositional variations of debris in the basal zone of Aktineq and Ujarasukjuik Glaciers may be summarized as follows:

A) Down-ice variations in grain-size distribution of basal debris appear to correlate with changes in the nature and texture of underlying lithological units. The passage of ice over the geological *AA-KT* contact is marked in both glaciers by a gradual increase in the proportion of fines ($< 74 \mu\text{m}$) of the basal debris, although this increase is considerably greater in Ujarasukjuik Glacier than in Aktineq Glacier. In the latter, the transition from the mudstone member to the sandstone member of the Kanguk Fm is reflected by a marked decline in the content of clay-size ($< 2 \mu\text{m}$) basal debris.

B) The transition from the *AA* to the *KT* terrane is also reflected by compositional changes in the clay mineralogy of basal debris, although in contrasting manners. In Aktineq Glacier, a sharp increase of the relative proportion of expansible clays occurs immediately down-ice of the contact, and is maintained over some distance before reducing to its initial level. This trend is somewhat similar to the head-and-tail pattern (Shilts 1992) of a glacial dispersal train and may represent the incorporation and subsequent down-ice dilution of 17 Å clays from the Kanguk Fm mudstones in the basal debris load. In Ujarasukjuik Glacier, the transition from the Shield onto the lowlands is marked by a decline in abundance of 17 Å clays which may reflect an impoverishment in expansible clays of the underlying Eureka Sound Fm sandstones relative to an unmapped up-ice source.

C) By contrast, mineralogical and textural variations in the dispersed facies profiles on Ujarasukjuik Glacier display no clear correlations with the underlying substrate, even

over considerable distances. There is, in certain profiles, an apparent agreement in the proportion of expansible clays and that of clay-size debris, but this relation is within the range of variability of other profiles and is probably not significant.

These different variation patterns may be related to features observed in the basal zone of glaciers in the study area. Despite the considerable contrast in their overall debris load, Aktineq and Ujarasukjuik Glaciers are similar in many regards. In both glaciers, the bulk of the basal zone offers an aspect that generally corresponds to that of the dispersed facies as described by Lawson (1979), but considerable differences are also noticeable, such as the occurrence of closely spaced and well-defined debris layers near the sole of Aktineq Glacier. On Ujarasukjuik Glacier, the basal zone is almost entirely included within the dispersed facies over considerable parts of the margins, and only close to the snout is the lower debris-laden stratified facies observed.

The same applies somewhat to Aktineq Glacier, although a relatively thin (< 2 m) stratified facies is exposed on several locations in the vicinity of the piedmont. The contact between the basal zone and the overlying englacial zone also differs between glaciers: relatively sharp in Ujarasukjuik Glacier, more gradual in Aktineq Glacier. The lowermost exposed dirty ice layers on lateral ice sections systematically appear to be relatively debris-rich compared to overlying ones, but this is deceptive, as some of this debris content may be derived by ablation of dirty ice layers located higher in the basal zone. Laminated ice assumed to represent regelation ice is frequently observed near the base of ice sections and decreases in frequency upwards.

If considered in the light of XRD and particle-size data, these features suggest that under the piedmont lobes of Aktineq and Ujarasukjuik Glaciers, subglacial debris entrainment is restricted to the lowermost layers of the basal zone (within the first 1-2 m above the bed). Debris transported in the upper dispersed facies (which represents most of the basal zone in Ujarasukjuik Glacier) is probably derived from regions of the glacier bed

located higher up-ice over the highlands and transported without considerable mixing over the length of the glacier to the marginal zone.

Four basal ice profiles examined by DiLabio and Shilts (1979) on Aktineq and Ujarasukjuit Glaciers are shown on Fig. 6.2. Apart from the overall greater concentrations of 17 Å clays in the debris load of Aktineq Glacier (previously noted in Chapter 5), no strong correlation between the height of debris layers and the amount of expansible clays is found. Only profile 2, sampled on the western side of terminal margin of Aktineq Glacier, shows a relative enrichment of 17 Å clays in the lowermost layers of the basal zone. Unpublished data provided by W. Shilts and R. DiLabio indicate that debris volume in the layers displaying greater concentrations of expansible clays is about 10 times that of overlying ones. These debris-rich layers probably correspond to the lower facies of the basal zone, which is more likely to reflect subglacial accretion of underlying sediment, as indicated by their high 17: 14 Å ratios.

However, this relation was not found in any of the other profiles. On Ujarasukjuit Glacier (profiles 4 and 6), the absence of correlation between the 17:14 Å ratio and the height of layers (already noted in vertical sections; see Fig. 5.7a to 5.7g) may be attributed to the very limited exposure of the lower stratified facies over most of the margin of Ujarasukjuit Glacier. Profile 5, located on the upper western margin of Aktineq Glacier, was found to contain very uniform debris concentrations in the 35 to 50 % (Vol.) range (DiLabio and Shilts, unpublished data), and yet shows no downward enrichment of expansible clays in debris, as in profile 1. If any such pattern was initially present in the basal sequence, it appears to have been disrupted since, possibly by glaciotectionic deformation of the basal zone.

Observations cited above tend to indicate that only the lowermost debris-rich layers of the basal zone in piedmont lobes reflect the composition of the underlying substrate in a noticeable manner. This reinforces the contention that the bulk of the basal ice might be

Fig. 6.2. Relative amounts of expansive clays (17:10 Å) in debris sampled from four vertical ice profiles on Aktineq and Ujarasukjuit Glaciers. See Fig. 3.1 and 3.2 for location of profiles. (from DiLabio and Shilts 1979)

affected by subglacial processes active under the upper reaches of the glaciers, and that debris carried within it is primarily derived from the Precambrian substrate. However, since most of the basal debris load volume appears to be concentrated in the lower stratified facies, it can be expected that the bulk ablation product of the basal ice will reflect a locally entrained compositional signature, at least within the finer size fraction.

The occurrence of a frozen sediment raft 30 m up in the ice margin of Ujarasukjuik Glacier (Klassen and Shilts 1987) also indicates that locally-derived debris may be entrained high into the basal zone over relatively short transport distances, probably by a shearing-in or glaciotectionic thrusting mechanism.

On the eastern margin of the same glacier, vertical profiles VS1 and VS9 (*re* Appendix III) show high concentrations of expansible minerals in debris layers elevated 3 to 5 m above the glacier sole. The characteristic signature of these layers may reflect the occurrence of smectite-rich material being transported and comminuted along distinct shear planes in the ice. DiLabio and Shilts (1978) have reported similar findings in chlorite-rich green debris layers occurring 20 m above the bed of Ujarasukjuik Glacier, which were found to represent *in situ* crushing of epidotized gneissic rocks from the Precambrian substrate.

If it may be assumed that the high 17 Å signatures in profiles VS1 and VS9 indicate the presence of *KT*-derived debris, it follows that this debris was either entrained into its position of transport from the glacier bed, or incorporated at its present height from an incoming tributary glacier such as the one flowing into Ujarasukjuik Glacier from the Byam Martin Mountain foothills. Since it is not known whether this tributary is underlain by *KT* beds, the latter possibility cannot be verified.

However generally speaking, compositional analysis of debris layers in the basal zone of Bylot Island glaciers indicates that little homogenization in the debris load occurs after initial entrainment, and that each debris layer retains a characteristic mineralogical

signature reflecting its debris content. This was also implied by the work of Bernier (1985) who examined basal debris samples from Aktineq Glacier provided by J.L. Tison (*re* Lorrain *et al.* 1981). Bernier suggested that the variability of the heavy mineral suite from different samples was indicative that virtually no mixing occurred between adjacent debris bands after initial entrainment.

The presence of regelation ice in the lowermost basal zone of Aktineq and Ujarasukjuit Glaciers suggests that even though subglacial conditions are probably predominantly cold-based, some parts of the bed may be subjected to pressure-melting. However, it is difficult to determine if regelation ice is mostly being accreted under the upper parts of the glacier, or if it is also produced underneath the piedmont lobes. The situation is different closer to the glacier termini, where thick sequences of debris-laden ice comparable to the stratified facies are observed, as for instance on the western terminal zone of Ujarasukjuit Glacier. Features of this lower facies suggest a combination of various processes, including basal debris accretion by net freeze-on (thick frozen debris bands), regelation (laminated ice sequences), overriding of subglacial cavities (sigmoidal lenses of clean bubbly ice), and intense plastic deformation (complex folding, augen structures).

The presence of a peaty layer capping frozen sediments in the basal ice zone exposed near the snout of Ujarasukjuit Glacier also indicates that overriding of older stagnant basal ice has taken place in recent time. These processes are not incompatible with subglacial thermal conditions that characterize polar glaciers: large-scale accretion of debris by basal adfreeze, overriding or glaciotectionic thrusting have been reported before in cold-based glaciers by Boulton (1975), Shaw (1977) and Klassen (1982a). However, the fact that features such as compressive folds and thrust planes are restricted in extent to near the glacier termini indicate that other mechanisms must operate under the up-ice parts of the bed.

Thus, it is suggested that a major part of the basal zone in the Bylot Lowland piedmont glaciers probably represents the product of subglacial processes active underneath the up-ice parts of those glaciers, over the Byam Martin Mountains. Some support to this contention is provided by the results of co-isotopic analysis of debris-rich ice from the clotted ice zone of Ujarasukjuit Glacier.

6.3) Interpretation of co-isotopic data

The slopes of 4.4 and 7.8 obtained for the $\delta^{18}\text{O}$ - δD relation in basal and englacial ice on Ujarasukjuit Glacier are in close accordance with the 4.9 and 7.8 slopes reported by Lorrain *et al.* (1981) for Aktineq Glacier. These figures are also well within the range of slope values published by others (Table 6.1) for subpolar and polar glacier ice in northern and southern hemisphere locations.

A basal melting-refreezing mechanism at the ice-bedrock interface may thus be considered to explain some of the isotopic and morphological characteristics of discrete debris layers occurring throughout the basal dispersed facies of Ujarasukjuit Glacier. Subglacial debris accretion by regelation generally results in sequences of laminated macrocrystalline ice and fine layers of sand- to silt-size sediment (Hubbard and Sharp 1994). This type of ice has been widely reported before and is usually associated with the lower debris-rich facies of the basal zone where it occurs in association with thick frozen debris bands (Lawson 1979).

In the study area, regelation ice has been identified in the basal zone of Aktineq, Ujarasukjuit, and B-26 glaciers. On all sites, laminated ice was found to be interlayered with coarser debris-laden bands of ice, and was mostly restricted to the lowermost part of the basal zone with the stratified facies (where observed).

Table 6.1. $\delta^{18}\text{O}$ - δD slopes of englacial ice and basal ice of temperate, subpolar and polar glaciers in the northern and southern hemispheres.

Location	Englacial Ice	Basal Ice	Author(s)
Aktineq Glacier, <i>Bylot Is.</i>	7.8	4.9	Souchez and Jouzel 1982
Gl. de Tsanfleuron, <i>Switz.</i>	8.2	6.4	
Gannakhui Glacier, <i>India</i>	8.2	---	Grabczak <i>et al.</i> 1983
Broggi Glacier, <i>Peru</i>	7.3	---	
Jacobshavn Isbrae, <i>W.Greenland</i>	7.6	6.35	Sugden , Knight <i>et al.</i> 1987
Nordenskjold Glacier, <i>South Georgia Island</i>	8	6.4	Sugden <i>et al.</i> 1987
Balgesvarri Ice Cap, <i>Lyngen Pen., Norway</i>	7.72	5.45	Gordon <i>et al.</i> 1988
Glaciers B-7, B-17, & B-37 <i>Bylot Island</i>	7.63	---	Souchez <i>et al.</i> 1988
Søndre Strømfjord, <i>W. Greenland</i>	7.95	6.5	Knight 1989
Russel Glacier	8.22	4.83	Souchez <i>et al.</i> 1990
Qigssertâq Glacier <i>W.Greenland</i>		3.71	

The bulk of the ice in the dispersed facies of Ujarasukjuik Glacier differs from the stratified facies by its massive aspect, its coarser crystallinity and its overall low debris content. However, laminated macrocrystalline ice is also found in thin zones (relative to the total thickness of the facies) sometimes associated with the debris layers sampled.

The basal stratified facies in glaciers is generally considered to result from net freezing-on combined with small-scale cavitation regelation (Hubbard and Sharp 1989). However, the origin of the dispersed (or "clotted ice") facies remains speculative. Lawson (1979) has observed that ice with dispersed debris in the Matanuska Glacier (Alaska)

presents characteristics that may be attributed to both supraglacial and subglacial origins and may represent a transitional or boundary zone between regions of the glacier where either process dominate.

Souchez *et al.* (1988) have conducted co-isotopic analysis on ice from the dispersed facies of Aktineq Glacier and have found it to be isotopically homogeneous within a range of $\delta^{18}\text{O}$ - δD values similar to that of englacial ice. Therefore, the two ice types may not be distinguished in terms of $\delta^{18}\text{O}$ - δD signature alone. A progressive melting experiment conducted on clotted ice samples from Russel Glacier (West Greenland) has led to the same conclusions.

DiLabio and Shilts (unpublished data) sampled debris from 24 dirty ice layers in vertical sections on Ujarasukjuit Glacier (*re* Fig.6.2). Debris volume content in the layers varied from 0.9 % to 37.7 %, with an average of 8.19 %. Comparatively, sediment content in samples from profiles VS1 to VS10 was visually estimated to vary from 0.1 to 35 % around an average of 10% (Vol.). Most of the volume is that of the coarser fraction (> 200 mesh), which also accounts for 75 % to 95 % of the weight in most layers sampled (*re* Fig. 5.7a-5.7g). Pebbles up to 7 cm in diameter are frequent, and decimetric boulders have been observed embedded in layers at several locations. Larger clasts are sub-angular in shape, occasionally sub-rounded, and sometimes striated.

Debris concentration and coarseness in layers of the dispersed facies are in general above the range associated with regelation ice and are considerably higher than that of aggregates occurring throughout clotted ice. Occasional pebbles and boulders are also found scattered in the dispersed facies. In the dirty ice layers, the degree of pebble roundness, the presence of abundant fines (up to 20 % weight), and the particle-size distribution of fines (*re* Fig. 5.2b) point to a debris source at the ice-bedrock interface.

Ice crystallinity is generally coarse and does not show signs of mylonitization associated with shearing-in of debris at sub-freezing temperatures (Echelmeyer and

Zhongxiang 1987, Tison *et al.* 1993), but this observation may not be conclusive, as ice recrystallisation after accretion can overprint an older fabric.

To account for the characteristics of the sediment load and the co-isotopic signature of the ice matrix, a mechanism of subglacial debris accretion by *regelation coupled with shearing-in* is proposed for debris-rich layers in the dispersed facies on Ujarasukjuik Glacier. This type of process has been previously inferred for basal debris incorporation in sub-polar glaciers by several authors (Boulton 1975, Sugden *et al.* 1987).

The presence of water under a glacier experiencing sub-zero surface temperatures throughout most of the year can be explained by pressure-melting in up-ice zones where steep bed gradients may generate intense basal shear stress conditions at the sole. This would support the suggestion of Souchez *et al.* (1988) that pressure-melting and regelation could occur along steep valley heads under the upper reaches of Aktineq Glacier. In the marginal supraglacial zone of Ujarasukjuik Glacier, murky, silt-rich water was observed spouting out from crevasses. This water was strikingly different from crystal-clear meltwater outpouring from lateral ice fractures in several other locations, and its suspended sediment load was probably derived from the inner debris-rich basal zone of the glacier. However, circulation of water in subglacial conduits is a common feature in glaciers and is not necessarily indicative of internal melting or basal conditions allowing for regelation to occur (Hooke 1989, Souchez and Lorrain 1991).

Alternatively, water may be supplied to the glacier sole by percolation from the supraglacial zone or by infiltration from a source outside the glacier itself. Lawson and Kulla (1978) have suggested that an input of supraglacial meltwater from the central part of the glacier into the basal zone could possibly account for the relatively ^{18}O -enriched composition of ice layers produced by basal refreezing in the Matanuska Glacier, Alaska. This hypothesis has recently received support by the discovery of high tritium levels in basal ice of the Matanuska Glacier (Larson *et al.* 1994).

Several large ice-dammed periglacial lakes fed by nival and glacier meltwater exist along the upper margins of both Aktineq and Ujarasukjuik Glaciers, and could conceivably provide water to the inner regions of these glaciers. However, in isotopic terms ice formed by the freezing of subglacial water mixing with an isotopically depleted input is likely to display a $\delta^{18}\text{O}$ - δD signature close to the meteoric water line as a result of the re-homogenization of the reservoir through the course of freezing (Boulton and Spring 1986, Souchez and DeGroote 1985). This does not appear to be the case for dirty ice layers in the dispersed facies at Ujarasukjuik Glacier.

Given the high latitude and altitude, it is likely that subglacial thermal conditions for large parts of the glaciers are cold-based. Yet, in up-ice regions, the steep bed gradients and relief (inferred from elevation contours along the ice margins) are probably conducive to intense shear stress and frictional heating at the sole, which may lead to localized pressure-melting and regelation. Parts of the glacier bed affected by these processes may thus be regarded as polythermal to some extent. Regelation events in the upper cold-based parts of the glacier bed are probably linked to local increases of the basal stress field occurring as ice flows over bedrock obstacles. These events are likely to occur episodically rather than in a continuous manner, as would be the case if the ice-bedrock interface was close to the pressure-melting point.

The amplitude of ^{18}O fractionation in a single freezing event is limited by the fractionation coefficients in ice and water to about 3 ‰ (O'Neil 1968). In Ujarasukjuik Glacier, the most isotopically enriched sample (VS1-F: $\delta^{18}\text{O} = -19.31$ ‰) differs from the hypothetical water source composition by more than twice this value ($\Delta\delta^{18}\text{O} = 6.25$ ‰). This suggests that ice within some of the debris-rich layers may have originated by multiple freezing events. If such layers have undergone successive episodes of melting-refreezing without significant net loss of water, the evolution of their isotopic composition through

time may be reflected by gradual variations within the glacier, with increasingly depleted δ values occurring further down along the flow lines.

This contention was tested on Ujarasukjuit Glacier by comparing the isotopic composition of the basal ice samples in terms of their position in the basal zone. To do so, $\delta^{18}\text{O}$ and δD values for each sample were plotted on an X-Y scatter graph (Fig.6.3) against down-ice lateral distance (measured from profile VS-1) and stratigraphic position (as expressed by the ratio of height of sample above ground vs thickness of the basal sequence). No clear correlation was found between the isotopic signature of the ice samples and their position in the basal zone. To obtain more conclusive results would have required a single layer of debris-rich ice to be sampled at various intervals along its length, but the discontinuous character of most of these layers precluded such a procedure.

Despite the scarcity of samples for the areal extent considered, this result suggests that if multiple freezing events are involved in the accretion of the dirty ice layers, they occur episodically and not in a continuous pattern related to ice flow in the basal zone. Regelation may lead to the formation of continuous sequences of laminated debris-rich ice layers if the glacier bed is at or close to the pressure-melting point. However, if cold-based thermal conditions prevail at the glacier sole, entrainment of debris by regelation may be considerably limited and occur only where sufficient heat is produced by friction and intense basal shear stress to bring the ice-bedrock interface to the pressure-melting point.

Thus, it is suggested that discrete dirty ice layers found in the basal ice sequence of Ujarasukjuit glacier may result from the episodic entrainment of basal debris by regelation in zones of high basal stress. The discontinuous aspect and irregular spacing of the layers in the clotted ice facies may be explained by such a process. Preservation of the layers over considerable transport distances would be made possible by their transposition to a higher

Fig.6.3. Basal ice - Ujarasukjuit Glacier
Spatial $\delta^{18}\text{O}$ - δD variations in the basal zone

stratigraphic position in the basal zone along shear planes that are subsequently overprinted by recrystallization in the dominant stress field associated with ice flow.

The scattered aggregates of debris in the "clotted ice" matrix of the dispersed facies could arguably represent fine subglacial debris entrained mostly by differential shearing above the glacier bed where a strong vertical velocity gradient exists, in a process somewhat similar to that described by Tison *et al.* (1993) for "amber ice" in the bottom layers of a deep ice core in coastal Antarctica.

This process might explain the absence of a $\delta^{18}\text{O}$ - δD regelation signature in clotted ice, as only limited isotopic fractionation would occur at crystal boundaries. Freezing effects may also be hindered if the freezing rate in the water film is fast enough to prevent efficient isotopic diffusion in the liquid phase away from the freezing front during regelation (Souchez *et al.* 1987, 1988; Hubbard 1991). This is likely to be the case in ice subjected to cold temperatures and high strain rates.

In the down-ice segment of the polythermal zone, there may be a net accretion of laminated regelation ice at the glacier sole. Net regelation in this zone could be favoured by the presence of a supercooled meltwater film flowing through subglacial sediment, provided the latter is permeable enough to accommodate it (Alley *et al.* 1993). Alternatively, local changes in pore water pressure within the subglacial sediment may allow penetration of ice into it by regelation and lead to a net gain of debris-laminated ice (R.LeB.Hooke, personal communication, and Iverson 1993). Basal features in the piedmont margins of Aktineq and Ujarasukjuit Glacier indicate that regardless of the process involved, regelation is accreted at the sole, although not in considerable thickness. It is probable that the thicker laminated regelation sequences observed closer to the snout are the result of overstacking and folding by plastic deformation in the basal zone.

Although the interpretation presented above accounts for several features observed in the basal zone of Aktineq and Ujarasukjuit Glaciers, it should be regarded as tentative.

Knight (1989) has shown that a process of net freezing-on alone could not account for basal features observed at the margin of Søndre Strømfjord, West Greenland. Co-isotopic evidence from this location indicates that debris-rich and debris-poor elements of the basal sequence have different origins: while the debris bands display isotopic fractionation consistent with a freezing-on origin, the intercalated clean ice layers do not and must be accounted for by some other genetic mechanism.

More recently, Souchez *et al.* (1990) have reported isotopic data from Qigssertâq (Greenland) suggesting that ^{18}O enrichment in basal ice may result from isotope exchange with hydroxyl-bearing minerals. Souchez *et al.* postulate that in the presence of small amounts of Pleistocene age ^{18}O -depleted meltwater, exchange of oxygen isotopes can occur with hydroxyl-bearing clay minerals from basal ice layers with sufficient magnitude to markedly affect the initial composition of the meltwater. Although the possibility of such exchanges has been documented (James and Baker 1976), the suggestion of Souchez *et al.* remains questionable: Gow *et al.* (1979) did not report any systematic correlation between fine debris concentrations and δ values in ice from deep Antarctic cores, despite the presence of subglacially-derived meltwater.

6.4) Entrainment and transport of debris in glaciers of SW Bylot Island

Fig. 6.4 presents a schematic model of incorporation and transport of debris in glaciers of southwestern Bylot Island, based on features observed in Aktineq and Ujarasukjuit Glaciers. The model mostly emphasizes basal debris entrainment processes occurring in different parts of the glacier bed along the central flowline. The main elements of the model are summarized as follows:

1^o - At the head of glaciers, debris incorporation is mostly restricted to an input of coarse angular debris on the glacier surface and in crevasses. Extensional crevasses

observed at the edges of the accumulation area of Ujarasukjuit Glacier indicate that active ice flow occurs only tens of metres from the cirque walls. Debris brought onto the glacier surface by mass wasting is in part transported down-ice in a supraglacial position, but mostly buried under new snow accumulation or in crevasses, and eventually transferred by internal flow to englacial or subglacial transport positions. This is facilitated if a net loss of ice occurs in the basal zone through basal melting (Boulton 1978). Subglacial erosion in valley heads essentially occurs by plucking and loading fatigue of the underlying bedrock. Abrasion and debris entrainment are limited to the effects of differential laminated flow in the high basal shear stress zone above the glacier bed.

2^o - Over a long segment of the transport distance along its principal valley, the glacier bed experiences polythermal conditions, with localized zones of basal melting associated with intense shear stress, possibly coupled with some basal sliding by frictional heating at the ice-bedrock interface. It is difficult to assess if the same occurs under tributary valley glaciers, but their smaller ice volumes may not be sufficient to generate basal stress regimes that would favor polythermal conditions. As cumulative amounts of debris are derived from the valley walls and the bed, subglacial erosion becomes increasingly abrasive and the texture of subglacial debris gradually evolves toward a polymodal distribution reflecting the terminal comminution sizes of the various lithological fractions.

Debris entrainment at the glacier sole occurs by the combination of episodic regelation associated with shearing-in of debris where ice overrides steep bed obstacles. Discontinuous debris layers produced in this manner are preserved over considerable transport distances and are found throughout the basal zone in irregularly spaced positions. The clotted ice matrix which forms the bulk of the dispersed facies represents a zone of basal ice in which fine aggregates of debris were entrained and subsequently scattered by

differential shearing near the glacier bed, possibly coupled with limited regelation. The extent of the subglacial polythermal zone cannot be determined easily, but its lower limit may be located down-ice of the geological contact with the sedimentary terrane of the Bylot Lowlands. This is suggested by the increasing amounts of fines in basal debris beyond the AA-KT contact which indicate that some subglacial accretion is occurring.

3^o - Observations of ice marginal features suggest that only limited basal accretion occurs underneath the piedmont lobes of glaciers. The surface of glaciers is littered with supraglacial debris (including medial moraines) accumulated in the highlands. Occasional coarse debris bands in the englacial zone are probably derived by burial of supraglacial material in the upper reaches of the glaciers (Boulton 1978). The dispersed facies encloses most of the exposed part of the basal zone on Ujarasukjuik Glacier, and is also extensive on Aktineq Glacier. The stratified facies is only exposed in a fairly narrow zone of the ice margin, and is often obscured by the presence of ice-cored lateral moraines. The stratigraphy of the basal zone appears generally conformable over considerable distances, except where large scale features disrupt the ice flow pattern. This is the case on the eastern margin of Ujarasukjuik Glacier where a tributary glacier encounters the main ice stream, and on the western flank of Aktineq Glacier where ice calves into Bigfoot Lake.

4^o - In the vicinity of the glacier terminii, a considerable thickening of the basal zone occurs, coupled with a relative increase in height of the stratified facies relative to the overlying dispersed facies. Features of the stratified facies indicate a combination of processes are responsible for its formation, including regelation, glaciotectionic thrusting, net freezing-on, and overriding of stagnant ice. It is not possible to quantify the relative contribution of these processes, as the entire facies is intensely deformed by folding and thrusting associated with marginal compressive flow. Some basal sediment transfer to the

ice surface is believed to occur along upward-oriented meltwater conduits emerging near the snout of glaciers (W.W. Shilts, personal communication). Debris mounds and ridges protruding from the supragacial zone near the glacier termini may represent channel deposits that froze into the ice mass after meltwater was diverted to other drainage outlets.

6.5) Contrasting aspects in glaciers

The tentative model presented above accounts for some of the main physical features observed on Aktineq and Ujarasukjuik Glaciers, and proposes an interpretation for the mineralogical and isotopic variations of debris and ice in the basal zone of those glaciers.

Yet, field evidence indicates that a wide range of debris entrainment processes occur in different glaciers. The basal zone of glacier B-26 displays several metres of laminated ice and fine debris extending to the sole, the bulk of which offers the aspect of regelation ice. No distinct transition between different basal facies is observed and the debris-rich ice grades upward into englacial ice. On the frontal margin of glacier B-28, debris "wedges" are protruding into basal ice from the sole, and sandy glaciofluvial sediment is being incorporated into ice in subglacial cavities. The variability of basal features observed in different glaciers clearly demonstrates that any inferences made regarding processes of debris entrainment and transport are largely site-specific, and reflect the general characteristics of the glaciers being considered.

One of the most striking contrasts observed in the study area is the overall greater debris volume being carried in Aktineq Glacier relative to Ujarasukjuik Glacier. The textural characteristics of debris are also significantly different in the two glaciers: grain-size distribution estimates based on dry weight measurements suggest that the basal debris load of Ujarasukjuik Glacier offers a wider range of textural variability than Aktineq Glacier (*re*

Fig. 5.2). However, figures based on relative weight strongly reflect that of the coarser fraction. Since Aktineq Glacier carries larger amounts of debris than Ujarasukjuik Glacier, the actual proportions of the sand-size and finer fractions in its basal load may be significantly different than Fig. 5.2 would suggest.

Fig. 6.5 displays the compared textural characteristics of basal debris collected by R.DiLabio and W. Shilts (unpublished data) from the eastern and western margins of Aktineq and Ujarasukjuik Glaciers. Grain-size distributions are given as relative weights of the sand, silt and clay fractions, and are thus restricted to a size range of < 2 mm. Within those limitations, basal debris on the eastern margin of Aktineq Glacier appears texturally finer than that of the western margin or that of Ujarasukjuik Glacier, although the latter may be misrepresented by the comparatively smaller number of data shown. The greatest amounts of clay-size material are found in debris from the western margin of Ujarasukjuik Glacier, but comparable concentrations occur on the western margin as well.

The silty component of debris from the eastern margin of Aktineq Glacier may be derived from fine-grained metapelites associated with the volcano-sedimentary belt outcropping in the piedmont area, or, alternatively, may originate from the *KT* mudstone or sandstone beds. However, since the *KT* lithologies outcrop on both sides of the glacier, it is difficult to explain why basal debris from the western margin does not offer a comparable texture to that of the eastern margin.

It appears more probable, and more in accordance with the model presented in section 6.4, that spatial textural differences in basal debris from different glacier margins are linked to lithological contrasts in the crystalline bedrock units underlying the upper reaches of those glaciers. DiLabio and Shilts (1979) had previously proposed a similar interpretation to account for the wide variations of trace element concentrations in debris from lateral moraines of Aktineq and C-93 Glaciers. Similar findings have been reported by

Fig. 6.5. Textural characteristics of basal debris from Aktineq and Ujarasukjuit Glaciers:
relative volume % of sand, silt and clay fractions.

Figure provided by R.DiLabio and W.Shilts.

Stephens *et al.* (1983), who emphasized their application for bedrock mapping and drift prospecting in alpine regions.

It is suggested that the important difference of debris volume transported in Aktineq Glacier relative to Ujarasukjuit Glacier is simply a reflection of the much larger drainage basin of the former. Along its main ice stream, Aktineq Glacier flows over a distance of 35 km and an altitude difference of 1500 m, while Ujarasukjuit Glacier covers about 20 km and drops in elevation by 1090 m from the head of its main cirque to the snout. Additionally, Aktineq Glacier is fed over its course by about 15 tributary glaciers, some of which are as wide as the main ice stream. Comparatively, only 6 major tributaries flow into Ujarasukjuit Glacier. It follows that a large part of the coarse crystalline debris transported in Aktineq Glacier might have been acquired at the confluence of tributaries in the alpine uplands. Boulton (1978) has suggested that in compressive flow zones such as those at the confluence of tributary glaciers or to nunataks, or on the lee-side of a spur in a glacier headwall, basal flowlines can be expected to transport material from the basal transport zone up into high level transport.

In a glacier such as Aktineq, the large number of tributaries flowing into the main ice stream may conceivably contribute a large quantity of material to the upper parts of the basal zone, probably in the form of scattered cobbles and pebbles and aggregates of finer debris. The relative roundness of larger striated clasts found dispersed throughout the clotted ice facies would appear to confirm this hypothesis.

Thus, the important basal debris load of Aktineq Glacier probably reflects in part an input from tributary glaciers, and a basal accretion component derived from subglacial material, as suggested earlier. In this regard, the contrast offered by Aktineq and Ujarasukjuit Glaciers illustrates well the control exerted by the general configuration of the glacier on processes of debris entrainment and transport.

Summary and Conclusions

An investigation of debris entrainment and transport processes was conducted on the piedmont outlets of large composite glaciers flowing out of the Byam Martin Mountains onto the lowlands of southwest Bylot Island, N.W.T. These glaciers were selected because *A)* the morphology of their ice margin offers excellent exposure of the internal stratigraphy of the basal and englacial zones; and *B)* the glaciers are flowing from an area underlain by Precambrian high-grade crystalline rocks of the Canadian Shield onto poorly-consolidated sedimentary beds of Cretaceous to Tertiary age.

Following an investigative approach suggested by DiLabio and Shilts (1978, 1979), an attempt was made to use this lithological contrast as a constraint to map the relative distribution of debris derived from the two terranes in the basal zone of Aktineq and Ujarasukjuit Glaciers, and to estimate the pattern of entrainment and transport of these various debris components.

The relative abundance of specific mineral species in the clay fraction of basal debris was utilized as a tool to discriminate between sediment derived from the Cretaceous-Tertiary (*KT*) and Archean-Aphebian (*AA*) terranes. To do so, the clay mineralogy of glacial debris and of pulverized bedrock and erratic samples from both terranes was first determined by x-ray diffraction techniques.

Semi-quantitative indexes of relative abundance for expansible clays, chlorite-kaolinite, and mica-illite were calculated and plotted against distance of transport and height for a series of sampling traverses and profiles conducted on the ice margins of Aktineq and Ujarasukjuit Glaciers. Additionally, an assessment of the particle-size distribution in glacial debris and pulverized rocks was made using dry weighing and laser scanning methods.

Results have established that:

1o Clay-size debris in Aktineq and Ujarasukjuit Glaciers is essentially composed of a mixture of well-crystallized biotite and chlorite-kaolinite, and of expansible minerals, the latter being most probably montmorillonite-group smectites and interlayered chlorite-smectite. This is in accordance with previous findings by DiLabio and Shilts (1979).

2o Siltstones and sandstones of the *KT* terrane appear to contain a high proportion of expansible clays relative to *AA* migmatitic gneisses, the latter yielding larger amounts of chlorite and/or biotite. However, the presence of expansible clays in some of the gneissic samples and the wide lithological variability of both terranes suggests that these observations may not apply to the all bedrock units exposed in the study area.

3o Basal debris collected on the eastern margin of Aktineq Glacier immediately down-ice of the *AA-KT* faulted contact display an important enrichment in expansible clays relative to chlorite, kaolinite and mica with respect to samples obtained further up-ice. However, the opposite trend is observed on a profile sampled along the eastern margin of Ujarasukjuit Glacier, with the highest concentrations of expansible clays being found in basal debris collected over the *AA* terrane. Thus, it appears that if any mixing of clay-size debris derived from the two terranes occurs in the basal zone of Aktineq and Ujarasukjuit Glaciers, it does not result in a simple down-ice linear increase of the *KT* component relative to the *AA* one.

4o This observation also infers a possible source of secondary minerals in the mineralogical assemblage of some of the *AA* lithologies. These minerals may be derived from preserved weathering surfaces reflecting past milder climatic conditions, or originate from foreign drift or Paleozoic outliers occurring under the modern ice cover.

5o The basal zone on Aktineq and Ujarasukjuit Glaciers includes two distinct parts: a lower debris-rich facies, and an upper massive facies with dispersed debris and occasional dirty ice layers.

In essence, those facies have features similar to those assigned by Lawson (1979) to the *stratified* and *dispersed* facies, respectively. Observations show that the stratified facies becomes gradually thicker with down-ice distance, and attains a considerable height near the frontal margin of glaciers, partly as a result of compressive flow and plastic deformation of the basal zone. Features of the stratified facies suggest a subglacial accretion origin by combined processes of regelation, net freeze-on, glaciotectionic thrusting and folding. The dispersed facies is more conformable over the length of the ice margin and appears less affected by marginal disruptions.

On a local scale, mineralogical and textural variations in the lowermost layers of the basal zone appear to reflect changes in the nature of the underlying substrate. This relation does not hold for the bulk of the basal zone, which is enclosed in the dispersed facies. It is thus suggested that the lowermost part of the basal zone, which corresponds approximately to the stratified facies, is formed in part by accretion processes which incorporate debris from the basal traction zone over short transport distances, while the overlying dispersed facies originates from other subglacial entrainment processes in the higher regions of the glacier bed.

To elucidate the origin of debris-rich layers occurring throughout the dispersed facies, the co-isotopic $\delta^{18}\text{O}$ - δD relation of the ice matrix was determined using the Souchez and Jouzel (1984) method applied for a closed system composition. By linear regression method, a freezing slope of 4.9 was obtained, relative to a slope of 7.8 in firn collected from the central Bylot Icefield. This isotopic evidence and the general texture of debris transported in the layers suggest an origin by regelation in the basal traction zone of the glacier. The discontinuous and irregular aspect of the dirty ice layers in the dispersed facies may result from episodic regelation occurring in zones of high basal shear stress over steeper parts of the glacier bed.

The different conclusions presented above were integrated in a composite model of debris entrainment and transport, the features of which are summarized in Chapter 6. Generally speaking, results of this work indicate that the spatial distribution of debris in the basal zone of polar mountain glaciers does not form a continuous, two-dimensional zonation. The lowermost layers of the basal zone reflect to a certain degree the gradual mixing of components from the various lithologies it flows over. However, the nature of the debris load remains largely dominated by the lithological material entrained in the highland areas, where most of the erosion occurs.

Furthermore, the distribution of debris within the basal zone is strongly constrained by the general configuration of the glacier and by the thermodynamic conditions which prevail at its bed. Basal debris entrained in the glacier from lithologies outcropping underneath the piedmont lobes is transported close to the glacier bed within the lower stratified facies, while the upper dispersed facies includes material derived from rocks underlying the upper reaches of the glacier. There is no strong evidence for a vertical compositional gradation between the two facies, which are usually in sharp contact. It also seems very likely that an important fraction of the scattered debris in the dispersed facies is acquired in the confluence zone of tributary glaciers.

Till is deposited in part as lodgement till underneath the active glacier itself, but is mostly released by ablation in the marginal zone. Extensive till deposits are formed by the time-transgressive migration of the ice margin in response to changes in the glacier mass balance or thermal regime. Ablation processes in marginal areas result in an important homogenization of debris as a consequence of the reworking of sediment by meltwater during deposition. Typically, only debris released by the ablation of the lowermost parts of the basal zone is likely to preserve some of the stratification inherited during transport (Lawson 1979, 1981). Furthermore, this stratification is generally disrupted near the ice margin as a result of compressive glaciotectionic deformation.

Therefore, it appears unlikely that under the present climatic conditions, any vertical compositional zonation in the basal zone would be preserved after deposition, especially in view of the fact that most of the basal zone is enclosed within the dispersed facies which is characterized by a high ice to debris ratio. This does not preclude the possibility that longitudinal variations in basal till may be preserved, but results from the XRD analyses indicate that such variations may only be properly recognized if a better assessment of the lithological variability of the substrate is made.

The mechanisms which govern the distribution of debris inside glaciers are still poorly understood, as most observations reported to this day were made in marginal positions or, in a few cases, in tunnels excavated underneath the thinner edge of temperate glaciers. Attempts have been made at mapping the spatial variations of thermal conditions at the glacier bed by borehole temperature measurements, but these methods are often inaccurate as they introduce a change in the temperature distribution at the ice-bedrock interface.

However, recent developments in isotopic and physical glaciology now allow important inferences to be made on debris entrainment processes by the examination of basal ice retrieved from cores. It is the opinion of this author that a well-coordinated program of coring, coupled with the use of remote-sensing techniques such as ground-penetrating radar, would constitute the best approach to ascertain the internal structure and stratification of a glacier. An interesting focus for such a study could be the compressive stress zones where confluence of tributaries or overriding of nunataks occur, as very little is known of the erosional and debris entrainment processes active in these parts of glaciers. Remote-sensing of the internal structure of polar glaciers on Bylot Island may also provide information on their potential for the preservation of recent paleoclimatic records.

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APPENDIX I

Plates

(See list of plates on page xii)

PLATE 1. Coarse angular debris on the supraglacial surface of Aktineq Glacier.
Rockfall talus from adjacent valley wall in background. Ice pick is 1 m long for scale.

PLATE 2. Debris ridge extruded along a thrust plane on the surface of Aktineq Glacier,
near the frontal margin. Ice flow from left to right. The coarser debris represents the
remnants of a large frost-shattered supraglacial boulder.

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PLATE 3. Vertical ice cliff on the eastern margin of Ujarasukjuit Glacier. Height of the ice margin is about 70 m. The glacier section on this picture is entirely comprised within the englacial zone. Ice flow from left to right.

PLATE 4. The basal zone of Aktineq Glacier, exposed along the eastern margin 1 km down-ice of the Byam Martin Mountain foothills. Debris in the foreground is mostly of supraglacial origin. Ice flow from left to right.

PLATE 5. Sharp contact between the basal and englacial zones of Aktineq Glacier. Note well-defined sub-horizontal debris-rich ice layers. Ice flow from right to left. Hammer is $\approx$ 35 cm long for scale.

PLATE 6. "Ice streaming" around a large boulder embedded in the basal zone of Aktineq Glacier. Ice flow from left to right. Note the subangular aspect of the boulder, and the smaller one overriding it. Ice pick 1 m long for scale.

PLATE 7. Sandy debris layers exposed in a collapsed segment of englacial ice on the eastern margin of Aktineq Glacier. Ice pick 1 m long for scale.

PLATE 8. A section of the dispersed facies in the basal zone of Ujarasukjuik Glacier. The ice pick rests in the plane of ice foliation. The overhang at the top of the picture marks the transition with the englacial zone.

PLATE 9. Sharp contact between the stratified facies (bottom) and the dispersed facies (top) in the basal zone of Ujarasukjuik Glacier. The ice margin is undercut by a meltwater stream. The cavity in the top right corner is the former transport position of a boulder. Ice flow from left to right. Ice pick 1 m long for scale.

PLATE 10. Dirty ice in the basal dispersed facies of Ujarasukjuik Glacier. Silty sediment released by ablation gives the ice its overall dark colour. The thick debris layer in the middle of the picture marks the contact with the englacial zone. Ice pick 1 m long for scale.

PLATE 11. Basal ice section exposed by meltwater erosion on the western frontal margin of Ujarasukjuik Glacier. The lower part consists of overridden, debris-rich stagnant ice (A) capped by a frozen peaty layer (B). The upper part comprises thick debris bands and lenses of white bubbly ice (C). The strong overhang (D) marks the contact with the overlying englacial zone.

PLATE 12. Distorted debris layer entrained along a shear plane on the ice margin of glacier B-26. Ice flow from right to left. The total length of the layer is about 200 m. In its steepest part, the shear plane cuts the dominant ice foliation at about 20°.

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PLATE 13. Sampling of ice and sediment in debris-rich layers of a basal ice section exposed near the snout of Ujarasukjuit Glacier. Ice and debris are collected from about 10 to 15 cm deep into the ice to avoid isotopic contamination by surface processes.

PLATE 14. Glaciotectonic deformation features near the marginal compressive flow zone of Ujarasukjuit Glacier. (A): Complex set of overturned folds about a thrust plane. Figure in foreground for scale. Ice flow from left to right. (B): Crenulation folding at the base of an ice section. Ice flow from left to right. Ice pick 1 m long for scale.

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PLATE 15. Metasomatic retrograde alteration of biotite-feldspar gneiss: epidote-chlorite intergrowth in a fracture. Sample BYR-92-01. *Cross-polarized light. Magnification $\times 2.5$.*

PLATE 16. Pervasive metasomatic alteration of biotite-feldspar gneiss. Well-crystallized chlorite (center) and epidote (smaller angular crystals) in a matrix of calcite (pale beige). Sample BYR-92-01. *Cross-polarized light. Magnification $\times 10$.*

PLATE 17. Surface alteration of plagioclase feldspar in a biotite-feldspar gneiss. The alteration may produce microcrystalline sericite and hydromicas, kaolinite, smectite, and possibly interlayered clay minerals. Sample BYR-91-03. *Cross-polarized light.*
Magnification $\times 10$.

PLATE 18. Alteration of feldspars in a mylonitized biotite-feldspar gneiss. The fine-grained micaceous minerals may be in part recrystallized biotite. Sample BYR-91-03. *Cross-polarized light. Magnification $\times 2.5$.*

PLATE 19. Texturally mature quartz arenite of the Hassel Formation. The high level of sorting and roundness of grains reflects the fluvio-deltaic environment of deposition. Sample BYR-92-03. *Cross-polarized light. Magnification x 2.5.*

PLATE 20. Argillaceous matrix in a siltstone from the Eureka Sound Formation. Note the angularity and poor sorting of grains denoting the immature character of the sediment. Opaque minerals are probably spinels. Sample BYR-92-06. *Cross-polarized light. Magnification x 10.*

APPENDIX II

Particle-size data

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Table II-A: Grain size distribution in basal debris	
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Table II-A
 Grain size distribution in basal debris
 Aktineq Glacier: Eastern margin transect
 Dry weight measurements

Sample No	Distance * (km)	Total weight of sample (g) before split	Weight (g) Size fractions in microns			Weight % Size fractions in microns			Total weight of sample (g) after split	% Error on total weight
			Clay**	[< 2] Fine	[2-74] Coarse	Clay**	[< 2] Fine	[2-74] Coarse		
AK-91-15	2.925	182.65	0.90	3.37	172.56	0.51	1.91	97.59	176.83	3.29
AK-91-16	2.850	171.67	0.13	15.44	148.76	0.08	9.40	90.53	164.33	4.47
AK-91-17	2.775	169.4	3.57	9.87	151.59	2.16	5.98	91.86	165.03	2.65
AK-91-18	2.700	186.8	2.59	11.04	165.74	1.44	6.15	92.40	179.37	4.14
AK-91-19	2.625	172.39	1.90	7.07	156.52	1.15	4.27	94.58	165.49	4.17
AK-91-20	2.550	140.76	0.60	7.94	125.83	0.45	5.91	93.64	134.37	4.76
AK-91-21	2.475	174.38	1.54	5.26	159.82	0.92	3.16	95.92	166.62	4.66
AK-91-22	2.400	161.58	1.36	5.02	152.36	0.86	3.16	95.98	158.74	1.79
AK-91-23	2.325	172.25	0.99	11.60	154.48	0.59	6.94	92.46	167.07	3.10
AK-91-24	2.250	55.97	1.28	5.20	47.53	2.37	9.63	88.00	54.01	3.63
AK-91-25	2.175	119.56	2.83	12.83	98.98	2.47	11.19	86.34	114.64	4.29
AK-91-26	2.100	193.48	1.29	5.38	182.35	0.68	2.85	96.47	189.02	2.36
AK-91-27	2.025	170.49	7.66	8.82	148.84	4.63	5.34	90.03	165.32	3.13
AK-91-28	1.950	175.48	8.29	7.56	154.50	4.87	4.44	90.70	170.35	3.02
AK-91-29	1.875	168.01	7.81	9.26	147.51	4.75	5.63	89.63	164.58	2.08
AK-91-30	1.800	169.39	8.19	4.61	151.42	4.99	2.81	92.21	164.22	3.15
AK-91-31	1.725	177.07	7.73	7.21	157.73	4.48	4.18	91.35	172.67	2.55
AK-91-32	1.650	174.36	7.74	8.30	154.19	4.55	4.88	90.58	170.23	2.43
AK-91-33	1.575	156.69	8.19	12.06	131.85	5.38	7.93	86.69	152.10	3.02
AK-91-34	1.500	188.38	7.73	5.59	167.54	4.27	3.09	92.64	180.86	4.16
AK-91-35	1.425	164.47	7.74	5.38	146.62	4.85	3.37	91.79	159.74	2.96
AK-91-36	1.350	179.02	7.31	4.41	159.96	4.26	2.57	93.17	171.68	4.28

Table II-A
(Continued)

Sample No	Distance * (km)	Total weight of sample (g) before split	Weight (g)			Weight %			Total weight of sample (g) after split	% Error on total weight
			Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse	Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse		
AK-91-37	1.275	172.42	7.43	3.34	154.91	4.48	2.02	93.50	165.68	4.07
AK-91-38	1.200	157.4	6.05	4.20	140.94	4.00	2.78	93.22	151.19	4.11
AK-91-39	1.125	179.38	7.69	4.86	163.90	4.36	2.75	92.89	176.45	1.66
AK-91-40	1.050	148.34	6.16	8.21	131.07	4.24	5.64	90.12	145.44	1.99
AK-91-41	0.975	163.59	7.51	6.93	146.21	4.67	4.31	91.01	160.65	1.83
AK-91-42	0.900	180.33	6.36	5.52	163.21	3.63	3.15	93.21	175.09	2.99
AK-91-43	0.825	168.77	6.06	7.45	151.79	3.67	4.51	91.83	165.30	2.10
AK-91-44	0.750	156.44	8.51	7.70	137.61	5.53	5.01	89.46	153.82	1.70
AK-91-45	0.675	132.37	6.58	8.59	114.23	5.09	6.64	88.28	129.40	2.30
AK-91-46	0.600	162.58	0.85	4.78	153.02	0.54	3.01	96.45	158.65	2.48
AK-91-47	0.525	184.59	7.73	5.15	167.36	4.29	2.86	92.85	180.24	2.41
AK-91-48	0.450	174.47	6.09	2.84	161.19	3.58	1.67	94.75	170.12	2.56
AK-91-49	0.375	186.58	6.74	6.37	169.57	3.69	3.49	92.82	182.68	2.13
AK-91-50	0.300	177.84	6.09	2.73	165.65	3.49	1.56	94.94	174.47	1.93
AK-91-51	0.225	166.38	7.57	6.44	149.26	4.64	3.94	91.42	163.27	1.90
AK-91-52	0.150	152.47	6.27	5.30	136.60	4.23	3.58	92.19	148.17	2.90
AK-91-53	0.075	181.67	7.26	4.39	165.62	4.10	2.48	93.43	177.27	2.48
AK-91-54	0.000	185.45	7.27	9.86	164.37	4.01	5.43	90.56	181.50	2.18

(* distance measured from station AK-91-54)

(** Aqueous clay suspensions weighed in vials)

Table II-B
 Grain size distribution in basal debris
 Uluksuk Glacier: Eastern margin transect
 Dry weight measurements

Sample No	Distance * (km)	Total weight of sample (g) before split	Weight (g) Size fractions in microns			Weight % Size fractions in microns			Total weight of sample (g) after split	% Error on total weight
			Clay**	Fine	Coarse	Clay**	Fine	Coarse		
B7-92-01	0.05	180.24	5.66	1.78	169.27	3.20	1.01	95.79	176.71	2.00
B7-92-02	0.10	195.52	6.22	3.83	180.65	3.26	2.01	94.73	190.70	2.53
B7-92-03	0.15	161.14	0.00	1.89	156.28	0.00	1.19	98.81	158.17	1.88
B7-92-04	0.20	185.16	7.64	5.47	164.99	4.29	3.07	92.64	178.10	3.96
B7-92-05	0.30	196.23	5.83	4.93	178.54	3.08	2.60	94.32	189.30	3.66
B7-92-06	0.40	151.00	0.47	12.54	131.61	0.32	8.67	91.00	144.62	4.41
B7-92-07	0.50	80.19	7.93	16.07	53.05	10.29	20.86	68.85	77.05	4.08
B7-92-08	0.60	158.14	6.34	13.26	135.74	4.08	8.54	87.38	155.34	1.80
B7-92-09	0.65	158.03	6.05	5.72	141.51	3.95	3.73	92.32	153.28	3.10
B7-92-10	0.70	147.93	8.04	17.15	119.04	5.57	11.89	82.53	144.23	2.57
B7-92-11	0.75	125.15	5.71	20.55	94.02	4.75	17.09	78.17	120.28	4.05
B7-92-12	0.85	129.90	7.88	4.55	113.22	6.27	3.62	90.11	125.65	3.38
B7-92-13	0.90	71.78	7.16	7.27	54.27	10.42	10.58	79.00	68.70	4.48
B7-92-14	0.95	178.34	9.13	13.09	151.38	5.26	7.54	87.20	173.60	2.73
B7-92-15	1.00	108.75	7.32	27.46	69.64	7.01	26.30	66.69	104.42	4.15
B7-92-16	1.05	179.14	6.67	32.39	135.01	3.83	18.61	77.56	174.07	2.91
B7-92-17	1.10	136.23	8.01	10.78	113.80	6.04	8.13	85.83	132.59	2.75
B7-92-18	1.15	64.08	7.00	5.83	49.20	11.28	9.40	79.32	62.03	3.30
B7-92-19	1.20	103.56	8.79	21.13	69.93	8.80	21.16	70.04	99.85	3.72
B7-92-20	1.25	65.40	8.28	6.42	47.72	13.26	10.29	76.45	62.42	4.77
B7-92-21	1.30	37.17	9.87	20.23	5.73	27.55	56.46	15.99	35.83	3.74
B7-92-22	1.35	106.90	8.76	26.45	66.85	8.58	25.92	65.50	102.06	4.74

Table II-B
(Continued)

Sample No	Distance * (km)	Total weight of sample (g) before split	Weight (g)		Weight %			Total weight of sample (g) after split	% Error on total weight
			Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse	Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse	
B7-92-23	1.40	39.10	9.82	14.92	13.00	26.02	39.53	34.45	3.60
B7-92-24	1.45	145.33	11.59	31.76	97.99	8.20	22.47	69.33	2.82
B7-92-25	1.50	102.20	8.87	15.04	74.36	9.03	15.30	75.67	4.00

(* distance measured from station B7-92-01)

(** Aqueous clay suspensions weighed in vials)

Table II-C
 Grain size distribution in basal debris
 Ujarasutjuit Glacier: Vertical profiles
 Dry weight measurements

Sample No	Height * (m)	Total weight of sample (g) before split	Weight (g) Size fractions in microns			Weight % Size fractions in microns			Total weight of sample (g) after split	% Error on total weight
			[< 2] Clay**	[2-74] Fine	[> 74] Coarse	[< 2] Clay**	[2-74] Fine	[> 74] Coarse		
VS1-A	0.00	39.40	7.38	4.89	25.69	19.44	12.88	67.68	37.96	3.65
VS1-B	0.12	48.29	6.78	14.45	25.66	14.46	30.82	54.72	46.89	2.90
VS1-C	0.34	81.03	8.36	7.29	63.08	10.62	9.26	80.12	78.73	2.84
VS1-D	0.85	102.17	7.03	16.38	75.58	7.10	16.55	76.35	98.99	3.11
VS1-E	1.13	15.95	0.35	3.88	11.47	2.23	24.71	73.06	15.70	1.57
VS1-F	2.24	81.03	0.68	12.51	64.89	0.87	16.02	83.11	78.08	3.64
VS1-G	3.30	21.38	0.07	1.64	18.73	0.00	8.05	91.95	20.44	4.40
VS1-H	3.71	39.88	0.57	3.45	34.24	1.49	9.02	89.49	38.26	4.06
VS1-I	4.24	89.19	7.24	11.60	67.19	8.42	13.48	78.10	86.03	3.54
VS1-J	5.44	59.58	1.01	5.20	51.34	1.75	9.04	89.21	57.55	3.41
VS3-A	2.54	102.57	0.42	15.89	83.22	0.42	15.97	83.61	99.53	2.96
VS3-B	3.51	60.34	1.53	12.35	44.30	2.63	21.23	76.14	58.18	3.58
VS3-C	3.87	53.66	0.82	10.91	39.61	1.60	21.25	77.15	51.34	4.32
VS3-D	4.18	28.92	0.52	5.68	21.53	1.88	20.48	77.64	27.73	4.11
VS3-E	5.61	21.98	0.00	4.49	16.44	0.00	21.45	78.55	20.93	4.78
VS3-F	6.14	39.36	0.83	10.18	26.56	2.21	27.10	70.69	37.57	4.55
VS5-A	0.05	79.84	4.90	29.81	41.43	6.44	39.15	54.41	76.14	4.63
VS5-B	0.54	16.37	0.70	5.65	9.41	4.44	35.85	59.71	15.76	3.73
VS5-C	0.86	16.27	0.65	5.32	9.58	4.18	34.21	61.61	15.55	4.43
VS5-D	1.51	60.14	0.00	4.36	53.45	0.00	7.54	92.46	57.81	3.87
VS5-E	1.80	99.92	8.09	3.54	85.09	8.36	3.66	87.98	96.72	3.20
VS5-F	2.41	86.26	2.15	16.21	65.41	2.57	19.35	78.08	83.77	2.89

Table II-C
(Continued)

Sample No	Height * (m)	Total weight of sample (g) before split	Weight (g)			Weight %			Total weight of sample (g) after split	% Error on total weight
			Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse	Size fractions in microns [< 2] Clay**	[2-74] Fine	[> 74] Coarse		
VS5-G	4.85	21.89	0.61	20.41	0.00	2.90	97.10	0.00	21.02	3.97
VS6-A	1.08	15.60	4.11	10.84	0.00	27.49	72.51	0.00	14.95	4.17
VS6-B	1.42	39.87	1.34	9.02	27.96	3.50	23.54	72.96	38.32	3.89
VS6-C	1.87	43.70	1.77	9.47	30.61	4.23	22.63	73.14	41.85	4.23
VS6-D	2.35	20.92	0.00	4.82	15.51	0.00	23.71	76.29	20.33	2.82
VS6-E	4.33	34.30	0.66	14.24	18.65	1.97	42.44	55.59	33.55	2.19
VS6-F	4.56	18.45	0.51	4.03	13.23	2.87	22.68	74.45	17.77	3.69
VS7-A	5.91	23.88	0.59	3.63	18.98	2.54	15.65	81.82	23.20	2.85
VS7-B	6.46	61.48	1.56	11.10	46.40	2.64	18.79	78.56	59.06	3.94
VS7-C	6.86	19.64	0.21	2.08	16.62	1.11	11.00	87.89	18.91	3.72
VS7-D	7.77	39.54	0.55	7.98	29.34	1.45	21.07	77.48	37.87	4.22
VS7-E	8.15	61.01	0.00	10.18	48.41	0.00	17.58	83.58	58.59	3.97
VS7-F	8.51	118.68	2.66	16.40	97.26	2.29	14.10	83.61	116.32	1.99
VS7-G	9.00	32.62	1.35	6.41	23.64	4.30	20.41	75.29	31.40	3.74
VS7-H	9.77	115.16	3.94	21.05	87.48	3.50	18.72	77.78	112.47	2.34
VS7-I	10.61	142.94	1.48	19.64	117.48	1.07	14.17	84.76	138.60	3.04
VS8-A	1.60	28.22	0.60	4.19	22.40	2.21	15.41	82.38	27.19	3.65
VS8-B	2.04	21.01	0.00	4.49	15.90	0.00	23.12	81.87	20.39	2.95
VS8-C	2.36	32.51	1.37	10.97	18.91	4.38	35.10	60.51	31.25	3.88
VS8-D	3.01	54.79	0.42	5.79	46.35	0.80	11.02	88.18	52.56	4.07
VS8-E	3.30	17.57	0.41	4.66	11.96	2.41	27.36	70.23	17.03	3.07
VS8-F	3.91	150.93	1.23	25.64	119.25	0.84	17.55	81.61	146.12	3.19
VS8-G	6.35	26.14	0.61	5.53	19.10	2.42	21.91	75.67	25.24	3.44
VS8-H	7.31	105.56	0.82	13.49	87.26	0.81	13.28	85.91	101.57	3.78
VS8-I	8.04	129.85	0.44	13.99	111.92	0.35	11.07	88.58	126.35	2.70
VS9-A	0.45	45.36	0.82	11.64	31.03	1.89	26.76	71.35	43.49	4.12

Table B-C
(Continued)

Sample No	Height * (m)	Total weight of sample (g) before split	Weight (g) Size fractions in microns		Weight % Size fractions in microns		Total weight of sample (g) after split	% Error on total weight
			[< 2] Clay**	[2-74] Fine Coarse	[< 2] Clay**	[2-74] Fine Coarse		
VS9-B	0.94	68.68	1.99	14.26	3.01	21.59	66.05	3.83
VS9-C	1.26	49.57	0.00	9.85	0.00	20.72	47.86	3.45
VS9-D	1.91	33.79	0.51	9.14	1.57	28.17	32.45	3.97
VS9-E	2.20	24.40	0.43	3.92	1.84	16.75	23.40	4.10
VS9-F	2.81	6.71	0.10	0.23	1.55	3.56	6.46	3.73
VS9-G	5.25	6.17	0.04	0.71	0.66	11.68	6.08	1.46

(* Height measured from ground level)

Table II-D
Scanning laser analysis results
Quality control test for grain size separation procedure

Sampling location	Sample No	% Volume		
		Scanning range:		
		Particle diameter in microns		
		0-2	0-4	0-60
Aktineq Gl. E transect	AK-91-20	100.00	0.00	0.00
	AK-91-25	100.00	0.00	0.00
	AK-91-45	83.43	16.57	16.57
Ujarasukjuit Gl. E transect	B7-92-07	89.57	10.43	10.43
	B7-92-17	100.00	0.00	0.00
	B7-92-22	89.51	10.49	10.49
Ujarasukjuit Gl. Vert. profiles	VS1-B	100.00	0.00	0.00
	VS1-J	84.31	15.69	15.69
	VS6-A	100.00	0.00	0.00
	VS6-E	100.00	0.00	0.00

APPENDIX III

X-ray diffraction data

All analyses performed on Phillips Norelco automated diffractometer with monochromatic Cu K α radiation ($K\alpha_1 = 1.54056 \text{ \AA}$; $K\alpha_2 = 1.54439 \text{ \AA}$) at $2\theta/minute$. Unless specified otherwise, all XRD peak and peak ratios are given for samples after saturation with ethylene glycol to define the $17\text{\AA}$ reflection of expansible clays.

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Table III-A
X-ray diffraction data
Aktineq Glacier: Eastern margin transect
XRD peak heights (cm)

Sample No	Distance* (km)	Untreated samples			Glycolated samples			
		14 Å	10 Å	7 Å	17 Å	14 Å	10 Å	7 Å
AK-91-15	2.925	3.85	6.90	7.15	5.80	2.60	6.50	6.80
AK-91-16	2.850	3.20	7.35	5.80	3.20	2.15	6.70	5.70
AK-91-17	2.775	2.80	8.10	6.45	2.95	1.90	8.20	6.20
AK-91-18	2.700	3.30	10.25	7.50	2.80	1.90	8.75	6.25
AK-91-19	2.625	4.70	8.35	6.35	3.00	1.90	6.35	5.90
AK-91-20	2.550	4.10	10.35	8.80	3.65	2.50	10.00	8.40
AK-91-21	2.475	3.10	8.25	6.75	3.20	2.10	7.65	6.45
AK-91-22	2.400	3.00	8.80	6.35	3.05	2.00	8.05	5.60
AK-91-23	2.325	2.05	6.60	5.65	2.30	1.75	5.90	5.15
AK-91-24	2.250	2.60	9.65	6.95	2.75	2.00	7.95	6.05
AK-91-25	2.175	3.60	9.55	7.75	3.30	2.70	9.50	7.75
AK-91-26	2.100	2.90	9.10	6.55	2.60	2.30	8.00	5.75
AK-91-27	2.025	2.80	8.50	6.50	2.90	1.90	8.10	6.20
AK-91-28	1.950	3.00	9.05	7.05	3.55	2.10	8.05	5.85
AK-91-29	1.875	2.95	12.30	7.85	2.95	2.20	10.35	6.90
AK-91-30	1.800	2.95	11.20	7.50	2.55	2.40	9.30	6.70
AK-91-31	1.725	2.70	9.80	6.65	3.80	2.35	9.60	6.80
AK-91-32	1.650	3.00	9.60	7.65	2.60	2.00	8.55	7.05
AK-91-33	1.575	2.95	12.15	8.35	4.65	3.75	13.50	7.45
AK-91-34	1.500	3.40	10.15	8.60	3.60	2.10	8.45	6.70

Table III-A (continued)
XRD peak heights (cm)

Sample No	Distance* (km)	Untreated samples			Glycolated samples			
		14 Å	10 Å	7 Å	17 Å	14 Å	10 Å	7 Å
AK-91-35	1.425	3.40	9.20	8.40	3.80	2.10	6.80	7.10
AK-91-36	1.350	5.25	12.55	9.55	6.85	3.15	11.95	9.95
AK-91-37	1.275	3.75	10.45	8.00	3.15	2.50	8.65	7.50
AK-91-38	1.200	3.45	9.85	7.90	4.60	2.20	8.45	7.15
AK-91-39	1.125	3.40	9.85	6.90	4.60	2.25	8.35	6.75
AK-91-40	1.050	4.45	11.60	7.80	4.85	3.15	9.50	7.65
AK-91-41	0.975	2.95	7.80	5.50	2.85	2.05	6.45	5.40
AK-91-42	0.900	4.00	10.00	7.80	5.00	2.45	9.45	6.90
AK-91-43	0.825	2.85	8.50	6.50	3.10	1.75	7.20	6.45
AK-91-44	0.750	3.55	8.80	7.30	3.70	2.00	9.00	7.00
AK-91-45	0.675	2.95	7.85	6.20	3.45	1.85	7.95	5.55
AK-91-46	0.600	3.95	9.60	7.45	6.30	2.45	8.60	6.25
AK-91-47	0.525	3.70	11.80	7.40	4.85	2.10	11.00	6.25
AK-91-48	0.450	3.60	9.80	7.05	4.40	2.35	9.40	6.60
AK-91-49	0.375	3.30	8.85	7.40	3.50	1.70	8.05	6.35
AK-91-50	0.300	3.60	9.60	6.25	4.40	2.20	7.90	5.60
AK-91-51	0.225	2.95	13.95	7.80	3.05	2.10	11.95	7.45
AK-91-52	0.150	2.65	6.75	6.35	3.05	1.70	6.40	5.75
AK-91-53	0.075	4.00	8.15	8.10	6.25	2.80	6.80	6.85
AK-91-54	0.000	3.50	10.50	9.20	3.30	2.20	8.75	8.60

(* distance measured from station AK-91-54)

Table III-B
X-ray diffraction data
Aktineq Glacier: Eastern margin transect
XRD peak ratios

Sample No	Distance* (km)	Untreated samples			Glycolated samples				
		14 Å:10 Å	10 Å:7 Å	14 Å:7 Å	17 Å:14 Å	17 Å:10 Å	17 Å:7 Å	14 Å:10 Å	10 Å:7 Å
AK-91-15	2.925	0.56	0.97	0.54	2.23	0.89	0.85	0.40	0.96
AK-91-16	2.850	0.44	1.27	0.55	1.49	0.48	0.56	0.32	1.18
AK-91-17	2.775	0.35	1.26	0.43	1.55	0.36	0.48	0.23	1.32
AK-91-18	2.700	0.32	1.37	0.44	1.47	0.32	0.45	0.22	1.40
AK-91-19	2.625	0.56	1.31	0.74	1.58	0.47	0.51	0.30	1.08
AK-91-20	2.550	0.40	1.18	0.47	1.46	0.37	0.43	0.25	1.19
AK-91-21	2.475	0.38	1.22	0.46	1.52	0.42	0.50	0.27	1.19
AK-91-22	2.400	0.34	1.39	0.47	1.52	0.38	0.54	0.25	1.44
AK-91-23	2.325	0.31	1.17	0.36	1.31	0.39	0.45	0.30	1.15
AK-91-24	2.250	0.27	1.39	0.37	1.38	0.35	0.45	0.25	1.31
AK-91-25	2.175	0.38	1.23	0.46	1.22	0.35	0.43	0.28	1.23
AK-91-26	2.100	0.32	1.39	0.44	1.13	0.32	0.45	0.29	1.39
AK-91-27	2.025	0.33	1.31	0.43	1.53	0.36	0.47	0.23	1.31
AK-91-28	1.950	0.33	1.28	0.43	1.89	0.44	0.61	0.26	1.38
AK-91-29	1.875	0.24	1.57	0.38	1.34	0.29	0.43	0.21	1.50
AK-91-30	1.800	0.26	1.49	0.39	1.06	0.27	0.38	0.26	1.39
AK-91-31	1.725	0.28	1.47	0.41	1.62	0.40	0.56	0.24	1.41
AK-91-32	1.650	0.31	1.25	0.39	1.30	0.30	0.37	0.23	1.21
AK-91-33	1.575	0.24	1.46	0.35	1.24	0.34	0.62	0.28	1.81
AK-91-34	1.500	0.33	1.18	0.40	1.71	0.43	0.54	0.25	1.26
AK-91-35	1.425	0.37	1.10	0.40	1.81	0.56	0.54	0.31	0.96
AK-91-36	1.350	0.42	1.31	0.55	2.17	0.57	0.69	0.26	1.20
AK-91-37	1.275	0.36	1.31	0.47	1.26	0.36	0.42	0.29	1.15
AK-91-38	1.200	0.35	1.25	0.44	2.09	0.54	0.64	0.26	1.18
AK-91-39	1.125	0.35	1.43	0.49	2.04	0.55	0.68	0.27	1.24
AK-91-40	1.050	0.38	1.49	0.57	1.54	0.51	0.63	0.33	1.24

Table III-B (continued)
XRD peak ratios

Sample No	Distance* (km)	Untreated samples			Glycolated samples				
		14 Å:10 Å	10 Å:7 Å	14 Å:7 Å	17 Å:14 Å	17 Å:10 Å	17 Å:7 Å	14 Å:10 Å	10 Å:7 Å
AK-91-41	0.975	0.38	1.42	0.54	1.39	0.44	0.53	0.32	1.19
AK-91-42	0.900	0.40	1.28	0.51	2.04	0.53	0.72	0.26	1.37
AK-91-43	0.825	0.34	1.31	0.44	1.77	0.43	0.48	0.24	1.12
AK-91-44	0.750	0.40	1.21	0.49	1.85	0.41	0.53	0.22	1.29
AK-91-45	0.675	0.38	1.27	0.48	1.86	0.43	0.62	0.23	1.43
AK-91-46	0.600	0.41	1.29	0.53	2.57	0.73	1.01	0.28	1.38
AK-91-47	0.525	0.31	1.59	0.50	2.31	0.44	0.78	0.19	1.76
AK-91-48	0.450	0.37	1.39	0.51	1.87	0.47	0.67	0.25	1.42
AK-91-49	0.375	0.37	1.20	0.45	2.06	0.43	0.55	0.21	1.27
AK-91-50	0.300	0.37	1.54	0.58	2.00	0.56	0.79	0.28	1.41
AK-91-51	0.225	0.21	1.79	0.38	1.45	0.26	0.41	0.18	1.60
AK-91-52	0.150	0.39	1.06	0.42	1.79	0.48	0.53	0.27	1.11
AK-91-53	0.075	0.49	1.01	0.49	2.23	0.92	0.91	0.41	0.99
AK-91-54	0.000	0.33	1.14	0.38	1.50	0.38	0.38	0.25	1.02

(* Distance measured from station AK-91-54)

Table III-C
X-ray diffraction data
Ujaraukuk Glacier: Eastern margin transect
XRD peak heights (cm)

Sample No	Distance* (km)	Untreated samples			Glycolated samples			
		14 Å	10 Å	7 Å	17 Å	14 Å	10 Å	7 Å
B7-92-01	0.05	5.05	6.65	11.20	3.00	3.45	6.40	10.05
B7-92-02	0.10	3.90	12.45	9.85	2.40	2.90	12.55	9.70
B7-92-03	0.15	3.75	13.35	10.20	2.60	2.60	11.95	8.80
B7-92-04	0.20	3.95	11.55	9.90	2.65	3.10	11.20	9.20
B7-92-05	0.30	3.65	8.10	8.50	2.45	2.90	8.05	7.65
B7-92-06	0.40	5.40	10.75	11.40	3.85	3.20	9.20	9.45
B7-92-07	0.50	3.35	10.15	9.20	1.80	2.40	7.85	8.50
B7-92-08	0.60	3.00	14.45	10.50	1.00	2.45	13.55	9.80
B7-92-09	0.65	2.90	10.75	9.15	1.40	1.90	9.80	8.70
B7-92-10	0.70	2.80	16.60	11.70	1.01	2.85	15.10	11.40
B7-92-11	0.75	4.30	15.60	11.55	1.80	3.20	14.70	11.70
B7-92-12	0.85	4.50	9.15	10.90	2.05	3.40	10.90	9.85
B7-92-13	0.90	3.00	8.55	7.75	1.50	2.60	7.85	7.50
B7-92-14	0.95	2.90	14.95	9.55	1.75	2.20	13.10	8.60
B7-92-15	1.00	2.60	9.05	6.40	1.50	2.00	8.40	6.60
B7-92-16	1.05	2.50	9.80	7.70	1.50	2.40	7.75	7.30
B7-92-17	1.10	3.20	12.30	8.60	1.20	2.35	11.60	8.65
B7-92-18	1.15	3.95	11.75	13.95	1.20	3.60	11.65	11.50
B7-92-19	1.20	2.00	8.45	6.30	1.50	1.90	7.25	5.20
B7-92-20	1.25	3.15	13.70	10.40	1.20	3.00	13.75	10.05
B7-92-21	1.30	2.15	8.20	7.00	1.70	2.00	7.40	7.35
B7-92-22	1.35	3.00	9.75	8.70	1.55	2.50	9.20	7.50
B7-92-23	1.40	2.30	5.60	5.10	1.30	2.00	5.55	5.70
B7-92-24	1.45	2.30	6.70	6.15	1.50	2.15	7.00	5.90
B7-92-25	1.50	2.40	6.35	5.90	1.90	1.90	9.60	7.60

(* Distance measured from station B7-92-01)

Table III-D
X-ray diffraction data
Ujarsaukquit Glacier: Eastern margin transect
XRD peak ratios

Sample No	Distance* (km)	Untreated samples			Glycolated samples				
		14 Å:10 Å	10 Å:7 Å	14 Å:7 Å	17 Å:14 Å	17 Å:10 Å	17 Å:7 Å	14 Å:10 Å	10 Å:7 Å
B7-92-01	0.05	0.76	0.59	0.45	0.87	0.47	0.30	0.54	0.64
B7-92-02	0.10	0.31	1.26	0.40	0.83	0.19	0.25	0.23	1.29
B7-92-03	0.15	0.28	1.31	0.37	1.00	0.22	0.30	0.22	1.36
B7-92-04	0.20	0.34	1.17	0.40	0.85	0.24	0.29	0.28	1.22
B7-92-05	0.30	0.45	0.95	0.43	0.84	0.30	0.32	0.36	1.05
B7-92-06	0.40	0.50	0.94	0.47	1.20	0.42	0.41	0.35	0.97
B7-92-07	0.50	0.33	1.10	0.36	0.75	0.23	0.21	0.31	0.92
B7-92-08	0.60	0.21	1.38	0.29	0.41	0.07	0.10	0.18	1.38
B7-92-09	0.65	0.27	1.17	0.32	0.74	0.14	0.16	0.19	1.13
B7-92-10	0.70	0.17	1.42	0.24	0.35	0.07	0.09	0.19	1.32
B7-92-11	0.75	0.28	1.35	0.37	0.56	0.12	0.15	0.22	1.26
B7-92-12	0.85	0.49	0.84	0.41	0.60	0.19	0.21	0.31	1.11
B7-92-13	0.90	0.35	1.10	0.39	0.58	0.19	0.20	0.33	1.05
B7-92-14	0.95	0.19	1.57	0.30	0.80	0.13	0.20	0.17	1.52
B7-92-15	1.00	0.29	1.41	0.41	0.75	0.18	0.23	0.24	1.27
B7-92-16	1.05	0.26	1.27	0.32	0.62	0.19	0.21	0.31	1.06
B7-92-17	1.10	0.26	1.43	0.37	0.51	0.10	0.14	0.20	1.34
B7-92-18	1.15	0.34	0.84	0.28	0.33	0.10	0.10	0.31	1.01
B7-92-19	1.20	0.24	1.34	0.32	0.79	0.21	0.29	0.26	1.39
B7-92-20	1.25	0.23	1.32	0.30	0.40	0.09	0.12	0.22	1.37
B7-92-21	1.30	0.26	1.17	0.31	0.85	0.23	0.23	0.27	1.01
B7-92-22	1.35	0.31	1.12	0.34	0.62	0.17	0.21	0.27	1.23
B7-92-23	1.40	0.41	1.10	0.45	0.65	0.23	0.23	0.36	0.97
B7-92-24	1.45	0.34	1.09	0.37	0.70	0.21	0.25	0.31	1.19
B7-92-25	1.50	0.38	1.08	0.41	1.00	0.20	0.25	0.20	1.26

(* Distance measured from station B7-92-01)

Table III-E
X-ray diffraction data
Ujarasukjuik Glacier: E margin vertical profiles

Sample No	Height* (m)	XRD peak heights (cm)				XRD peak height ratios					
		17 Å	14 Å	10 Å	7 Å	17 Å:14 Å	17 Å:10 Å	17 Å:7 Å	14 Å:10 Å	10 Å:7 Å	
VS1-A	0.00	16.00	34.50	132.00	113.00	0.46	0.12	0.14	0.26	1.17	
VS1-B	0.18	16.00	29.50	141.00	101.00	0.54	0.11	0.16	0.21	1.40	
VS1-C	0.64	21.50	35.00	77.00	132.50	0.61	0.28	0.16	0.45	0.58	
VS1-D	0.85	14.50	25.50	171.50	116.50	0.57	0.08	0.12	0.15	1.47	
VS1-E	1.13	14.00	21.50	107.50	99.50	0.65	0.13	0.14	0.20	1.08	
VS1-F	2.24	32.00	33.00	150.00	104.00	0.97	0.21	0.31	0.22	1.44	
VS1-G	3.30	25.00	40.00	89.00	149.00	0.62	0.28	0.17	0.45	0.60	
VS1-H	3.71	181.00	109.00	38.50	250.00	1.66	4.70	0.72	2.83	0.15	
VS1-I	4.24	24.00	36.00	161.00	111.00	0.67	0.15	0.22	0.22	1.45	
VS1-J	5.44	35.00	39.00	137.50	153.00	0.90	0.25	0.23	0.28	0.90	
VS3-A	2.54	10.00	21.50	90.50	82.00	0.47	0.11	0.15	0.24	1.10	
VS3-B	3.51	9.50	22.50	137.00	97.50	0.42	0.07	0.12	0.16	1.41	
VS3-C	3.87	17.00	35.00	169.00	129.00	0.49	0.10	0.18	0.21	1.31	
VS3-D	4.18	10.50	27.00	119.00	103.50	0.39	0.09	0.25	0.23	1.15	
VS3-E	5.61	11.00	24.00	134.50	107.50	0.46	0.08	0.13	0.18	1.25	
VS3-F	6.14	16.00	29.00	167.50	127.00	0.55	0.10	0.16	0.17	1.32	
VS5-A	0.05	13.00	19.00	90.00	63.00	0.68	0.14	0.2	0.21	1.43	
VS5-B	0.54	10.00	15.00	100.00	67.00	0.67	0.10	0.11	0.15	1.49	
VS5-C	0.86	8.50	16.50	76.00	53.50	0.52	0.11	0.18	0.22	1.42	
VS5-D	1.51	17.00	30.50	35.50	80.00	0.56	0.48	0.2	0.86	0.44	
VS5-E	1.80	17.00	18.00	240.00	104.00	0.94	0.07	0.26	0.08	2.31	
VS5-F	2.41	16.50	20.00	112.00	58.50	0.82	0.15	0.22	0.18	1.91	
VS5-G	4.85	12.00	25.50	167.50	103.00	0.47	0.07	0.15	0.15	1.63	
VS6-A	1.08	10.00	19.00	149.00	67.00	0.53	0.07	0.21	0.13	2.22	
VS6-B	1.42	8.50	18.00	122.00	71.00	0.47	0.07	0.15	0.15	1.72	
VS6-C	1.87	13.00	23.00	127.00	74.00	0.57	0.10	0.19	0.18	1.72	
VS6-D	2.35	13.00	17.00	123.00	53.00	0.76	0.11	0.19	0.14	2.32	
VS6-E	4.33	9.50	19.00	108.00	75.00	0.50	0.09	0.18	0.18	1.44	
VS6-F	4.56	9.00	17.00	70.00	58.00	0.53	0.13	0.14	0.24	1.21	

Table III-E
(continued)

Sample No	Height* (m)	XRD peak heights (cm)				XRD peak height ratios					
		17 Å	14 Å	10 Å	7 Å	17 Å:14 Å	17 Å:10 Å	17 Å:7 Å	14 Å:10 Å	14 Å:7 Å	10 Å:7 Å
VS7-A	5.91	10.00	18.00	77.00	51.00	0.56	0.13	0.22	0.23	0.23	1.51
VS7-B	6.46	7.50	20.00	83.00	66.50	0.38	0.09	0.16	0.24	0.24	1.25
VS7-C	6.86	10.00	15.00	77.00	55.00	0.67	0.13	0.15	0.19	0.19	1.40
VS7-D	7.77	12.00	20.00	100.50	61.50	0.60	0.12	0.22	0.20	0.20	1.63
VS7-E	8.15	13.00	16.50	77.00	50.00	0.79	0.17	0.12	0.21	0.21	1.54
VS7-F	8.51	13.50	17.50	75.00	62.00	0.77	0.18	0.12	0.23	0.23	1.21
VS7-G	9.00	12.00	22.00	102.00	77.50	0.55	0.12	0.1	0.22	0.22	1.32
VS7-H	9.77	13.00	25.00	98.00	79.00	0.52	0.13	0.13	0.26	0.26	1.24
VS7-I	10.61	12.50	18.00	91.00	59.00	0.69	0.14	0.1	0.20	0.20	1.54
VS8-A	1.60	9.50	13.50	83.00	51.00	0.70	0.11	0.13	0.16	0.16	1.63
VS8-B	2.04	9.50	15.00	87.50	50.50	0.63	0.11	0.17	0.17	0.17	1.73
VS8-C	2.36	15.50	26.00	129.00	87.50	0.60	0.12	0.21	0.20	0.20	1.47
VS8-D	3.01	9.00	19.00	84.00	62.50	0.47	0.11	0.15	0.23	0.23	1.34
VS8-E	3.30	10.50	17.00	75.00	48.00	0.62	0.14	0.16	0.23	0.23	1.56
VS8-F	3.91	15.50	28.50	124.00	97.50	0.54	0.12	0.21	0.23	0.23	1.27
VS8-G	6.35	11.00	21.50	114.50	71.00	0.51	0.10	0.16	0.19	0.19	1.61
VS8-H	7.31	16.50	21.00	188.50	73.50	0.79	0.09	0.28	0.11	0.11	2.56
VS8-I	8.04	11.00	22.00	130.00	91.50	0.50	0.08	0.12	0.17	0.17	1.42
VS9-A	0.5	10.00	16.00	86.00	60.00	0.62	0.12	0.17	0.19	0.19	1.43
VS9-B	0.94	13.50	21.50	130.00	73.00	0.63	0.10	0.18	0.17	0.17	1.78
VS9-C	1.26	12.00	26.00	126.50	83.00	0.46	0.09	0.14	0.21	0.21	1.52
VS9-D	1.91	10.00	17.50	82.50	66.50	0.57	0.12	0.15	0.21	0.21	1.24
VS9-E	2.20	10.50	18.00	89.50	72.00	0.58	0.12	0.15	0.20	0.20	1.24
VS9-F	2.81	16.50	25.50	121.00	89.50	0.65	0.14	0.18	0.21	0.21	1.35
VS9-G	5.25	15.00	29.00	23.50	103.00	0.52	0.64	0.15	1.23	1.23	0.23

(* Height measured from ground level)

Table III-F.
XRD reproducibility test results
 (U) = Untreated samples; (G) = Glycolated samples

XRD Peak (d-spacing)	Peak height (cm)			Mean μ (cm)	Std. Deviation σ (cm)	% Error (= $(\sigma/\mu) \times 100$)
AK-91-17	RUN A	RUN B	RUN C			
14 Å (U)	2.80	3.05	2.85	2.90	0.11	3.72
10 Å (U)	8.10	8.25	8.35	8.23	0.10	1.25
7 Å (U)	6.45	6.75	6.50	6.57	0.13	2.00
17 Å (G)	2.95	3.15	2.90	3.00	0.11	3.60
14 Å (G)	1.95	2.20	1.90	2.02	0.13	6.51
10 Å (G)	8.15	7.80	8.10	8.02	0.15	1.93
7 Å (G)	5.80	6.35	6.40	6.18	0.27	4.40
B7-92-09	RUN A	RUN B	RUN C			
14 Å (U)	3.30	3.00	3.10	3.13	0.12	3.98
10 Å (U)	10.05	10.35	9.85	10.08	0.21	2.04
7 Å (U)	9.25	9.15	8.75	9.05	0.22	2.39
17 Å (G)	1.75	1.50	1.65	1.63	0.10	6.29
14 Å (G)	2.35	2.20	2.40	2.32	0.08	3.67
10 Å (G)	7.95	9.45	9.20	8.87	0.66	7.40
7 Å (G)	8.50	8.65	7.55	8.23	0.49	5.92
B7-92-18	RUN A	RUN B	RUN C			
14 Å (U)	3.95	2.00	1.90	2.62	0.94	36.06
10 Å (U)	11.75	3.95	4.40	6.70	3.58	53.37
7 Å (U)	13.95	5.10	5.90	8.32	4.00	48.06
17 Å (G)	1.20	1.20	0.50	0.97	0.33	34.14
14 Å (G)	3.60	2.00	1.75	2.45	0.82	33.45
10 Å (G)	11.65	6.05	4.25	7.32	3.15	43.07
7 Å (G)	11.50	7.75	5.70	8.32	2.40	28.88

Table III-F (continued)

XRD Peaks (d-spacing)	Peak height ratios			Mean μ (cm)	Std. Deviation σ (cm)	% Error (= $(\sigma/\mu) \cdot 100$)
AK-91-17	RUN A	RUN B	RUN C			
14 Å:10 Å (U)	0.35	0.37	0.34	0.35	0.01	3.53
14 Å:7 Å (U)	0.43	0.45	0.44	0.44	0.01	1.86
10 Å:7 Å (U)	1.26	1.22	1.28	1.25	0.02	1.99
17 Å:14 Å (G)	1.51	1.43	1.53	1.49	0.04	2.90
17 Å:10 Å (G)	0.36	0.40	0.36	0.37	0.02	5.05
17 Å:7 Å (G)	0.51	0.50	0.45	0.49	0.03	5.39
14 Å:10 Å (G)	0.24	0.28	0.23	0.25	0.02	8.64
14 Å:7 Å (G)	0.34	0.35	0.30	0.33	0.02	6.55
10 Å:7 Å (G)	1.41	1.23	1.27	1.30	0.08	5.92
B7-92-09	RUN A	RUN B	RUN C			
14 Å:10 Å (U)	0.33	0.29	0.31	0.31	0.02	5.27
14 Å:7 Å (U)	0.36	0.33	0.35	0.35	0.01	3.60
10 Å:7 Å (U)	1.09	1.13	1.13	1.12	0.02	1.69
17 Å:14 Å (G)	0.74	0.68	0.69	0.70	0.03	3.73
17 Å:10 Å (G)	0.22	0.16	0.18	0.19	0.02	13.36
17 Å:7 Å (G)	0.21	0.17	0.22	0.20	0.02	10.80
14 Å:10 Å (G)	0.30	0.23	0.26	0.26	0.03	10.89
14 Å:7 Å (G)	0.28	0.25	0.32	0.28	0.03	10.12
10 Å:7 Å (G)	0.94	1.09	1.22	1.08	0.11	10.56

APPENDIX IV

Vertical profile logs Ujarasukjuit Glacier - Basal Zone

*Ice stratigraphy / Mineralogical
and textural variations of basal debris
(Key to symbols inserted next page)*

**Basal ice sections
Eastern margin - Ujarasukjuit Glacier
Key to symbols**

* Relative debris content is based on visual estimates and expressed as volume percentage of debris in ice

APPENDIX V

Stable isotopic data

(All figures given with a precision of ± 0.10 ‰ for $\delta^{18}\text{O}$ and ± 1.3 ‰ for δD .)

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Table V-A
 Monthly average isotopic composition
 of precipitation in Pond Inlet, NWT
 (July 1990 to February 1992)

Month	Year	Del values (per mil)	
		Deuterium	Oxygen 18
Jul.	1990	-153.0	-22.90
Aug.	1990	-153.0	-19.20
Sept.	1990	-157.0	-20.75
Oct.	1990	-191.0	-25.60
Nov.	1990	-287.0	-37.40
Dec.	1990	-281.0	-37.00
Jan.	1991	-338.0	-43.60
Feb.	1991	-278.0	-35.60
Mar.	1991	-210.1	-27.57
Apr.	1991	-233.4	-32.12
May	1991	-163.6	-20.03
Jun.	1991	-158.4	-21.25
Jul.	1991	-126.9	-17.62
Aug.	1991	-168.7	-23.04
Sept.	1991	-173.8	-22.97
Oct.	1991	-176.2	-23.93
Nov.	1991	-225.8	-29.42
Dec.	1991	-300.5	-39.47
Jan.	1992	-321.1	-40.97
Feb.	1992	-325.0	-40.87

(Data provided by F.A. Michel)

Table V-B
Isotopic composition of firn in a core
extracted from the Bylot Island icefield
in August 1992

Depth to (cm)	Del values (per mil)	
	Deuterium	Oxygen 18
5	-173.3	-23.00
10	-188.8	-25.40
15	-173.7	-24.00
20	-170.8	-23.20
25	-165.6	-22.40
30	-186.5	-24.60
35	-193.1	-25.80
40	-189.6	-25.50

(Data provided by M.S. Elver)

Table V-C
Isotopic composition of basal ice samples
from Ujarasukjuit Glacier
(Dispersed facies, except for VS9-A: stratified facies)

Sample No*	Del values (per mil)	
	Deuterium	Oxygen 18
VS1-F	-160.7	-19.31
VS1-I	-168.1	-20.96
VS3-A	-168.1	-20.88
VS3-F	-167.3	-21.75
VS4-A	-174.4	-21.20
VS4-F	-181.5	-22.40
VS5-F	-168.7	-21.17
VS5-G	-162.5	-19.62
VS6-D	-170.9	-22.49
VS6-E	-171.8	-21.54
VS7-E	-169.9	-20.41
VS7-I	-169.5	-21.21
VS8-G	-168.4	-21.63
VS9-A	-172.4	-22.50
VS10-A	-184.5	-24.02

(* Samples are identified with reference
to the vertical sections shown in Appendix IV)

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