CHANGING DEPOSITIONAL ENVIRONMENTS IN THE MARGINAL ZONE OF A HIGH LATITUDE ICE SHEET

CHANGING DEPOSITIONAL ENVIRONMENTS IN THE MARGINAL ZONE OF A HIGH LATITUDE ICE SHEET

By

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ABSTRACT

Glacial and glacially-influenced deposits examined at two sites in west-central Ellesmere Island provide insights into the nature of glaciation during the late-Wisconsinan and Holocene advances, and the modes of deposition from arid, high latitude ice bodies. Glacial lithofacies identified indicate that englacial debris content varied spatially and it is inferred that basal thermal conditions also exhibited a complex pattern.

Direct glacial deposits usually consist of unsorted diamicts with a complete size range of matrix components, indicating an absence of meltwater-sorting or winnowing during deposition. Glacially-influenced fluvial, lacustrine, and nearshore marine deposits show that most of the Quaternary sediments were deposited by low-frequency, high magnitude events during deglaciation.

A tentative reconstruction of late-glacial history in the Strathcona Fiord area proposes that an ice tongue surged down Strathcona Fiord from a previous maximum position coincident with the present day head of the fiord. This surge destabilised the margin locally, causing rapid collapse in the valleys and melting into ice-cored basins on the higher plateau areas. Periodically these ice-cored basins would drain, providing large water and sediment discharges and reworking in some sites whilst leaving other

deposits unaffected. Subsequent reworking has been minimal in the cold, arid environment where continued uplift favours fluvial incision rather than extensive sandur development.

Examination of the modern ice margin shows that the ice here is frozen to the substrate but basal debris bands indicate that at some localities basal temperatures must be above pressure melting point. Patterns of debris entrainment and deposition and debris lithologies suggest that much of the transported debris is incorporated where lobes of ice begin to flow out from the main ice cap.

ACKNOWLEDGEMENTS

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This project was made possible by the logistical support of the Polar Continental Shelf Project of Environment Canada, Department of Energy, Mines and Resources. Without the help of all their people in Ottawa and Resolute this thesis would have been very short.

I owe a great debt of gratitude to my parents for their support, albeit at a distance, for this and everything else that I do. I only wish that my father could have seen this completed.

Finally, I would like to thank Tracey for guiding me through the times when it was very late, tempers were frayed and the situation seemed hopeless. She put up with more than I could have expected, and for that I thank her most deeply.

TABLE OF CONTENTS

LIST OF FIGURES

CHAPTER ONE: INTRODUCTION

1.1: Statement of problem.

The research reported in this thesis was stimulated by current debate in the fields of glacial sedimentology and Quaternary stratigraphy. The major theme of this work is glacial sedimentology in a high arctic environment. In an area described as polar desert, outlet glaciers are usually considered to be of the polar, or cold-based, type (Lorrain et al., 1981); that is to say, the ice at the base of the glacier is below the pressure melting point (Sugden and John, 1978). It is now recognised, however, that ice-caps and glaciers cannot be neatly classified into either cold-based or warm-based; complex patterns of zonation occur beneath ice bodies in response to many external variables (Lawson, 1977; Boulton, 1972; Paterson, 1981). These in turn lead to complex patterns of deposition.

Many different deposits have been assigned to cold-based ice bodies ranging from re-sedimented drapes of diamict (Eyles and Miall, 1986) to perfectly preserved "sublimation tills" (Shaw, 1977). As Shaw recognises, till classification has evolved from investigation of modern arctic environments of a maritime nature in Spitzbergen and

Iceland where summer temperatures are commonly above freezing and rainfall is plentiful. Shaw uses deposits in the Dry Valleys of southern Victoria Land, Antarctica as models for arid polar tills; yet it has been argued that this environment is so unusual that the deposits may not be representative of all arid polar glacial deposits. With this in mind it was proposed to study glacial deposits and environments of deposition at two sites in west-central Ellesmere Island in order to identify a deposit or suite of deposits characteristic of northern, arid polar environments.

Incidental to this study is the reconstruction of a late-glacial history for the Strathcona Fiord area. Quaternary studies in the High Arctic have lead to the formulation of two widely differing hypotheses on the distribution of Laurentide ice at its northern limits; these have been termed the maximum and minimum models by Mayewski et al. (1981). Observations and descriptions at the local scale must make the tracing of larger-scale limits more precise and this is thought to be more useful than wide-ranging studies which can only present aggregate data and draw generalised conclusions.

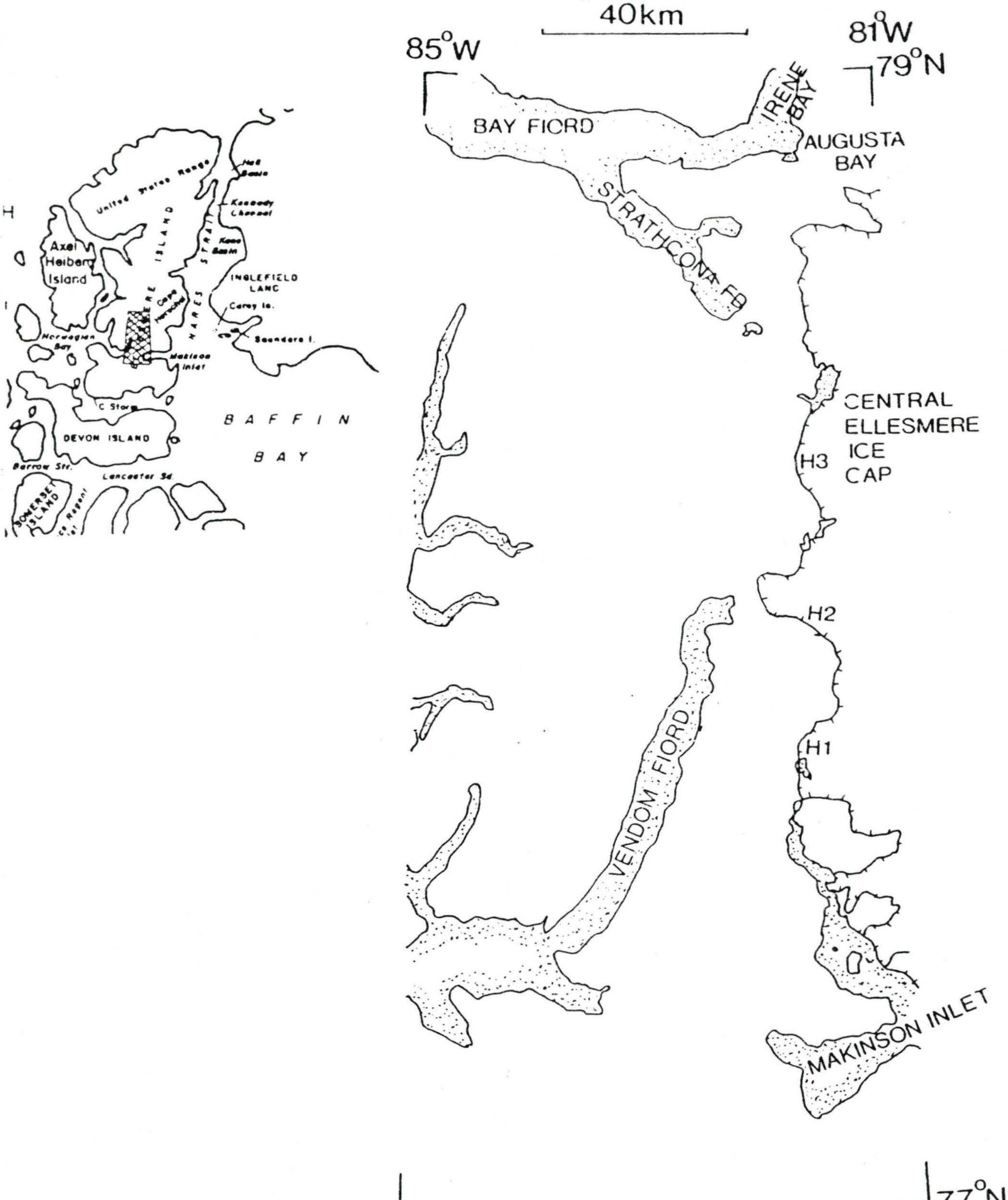

Figure 1. Location map of the field areas.

2.1: Description of the field areas.

A. Vendom Fiord.

The area under investigation is located near the head of Vendom Fiord, Ellesmere Island, Northwest Territories, Canada at latitude 78°02′ N, longitude 82°05′ W. The field area is delimited by the "Sverdrup River", the "Schei River", and the "Schei Glacier"[1] (designated glacier 78-8 by the Glacier Atlas of Canada (Hodgson, 1985)). Surficial geology was first mapped at a scale of 1:250 000 by Thorsteinsson (1972) and consists of an unconformable series of sedimentary formations of widely-differing ages overlying a Precambrian basement of gneisses, granites and migmatites. The "Hogsback Ridge" which rises to approximately 300m in the cetre of the study area is composed of rocks of the Allen Bay and Read Bay formations. These range in age from upper Ordovician to lower Devonian and are primarily dolomites and sandy and shaly limestones. Unconformably overlying the Allen Bay and Read Bay rocks and exposed on either side of the ridge are the rocks of the lower Devonian Vendom Fiord formation which are dominantly sandstones and red siltstones with some anhydrite and conglomerate. Mesozoic strata, such as those found to the west, are absent

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1 Unofficial names used in the text are those used by Ballantyne (1975) or coined by myself and are identified by their being placed in quotation marks.

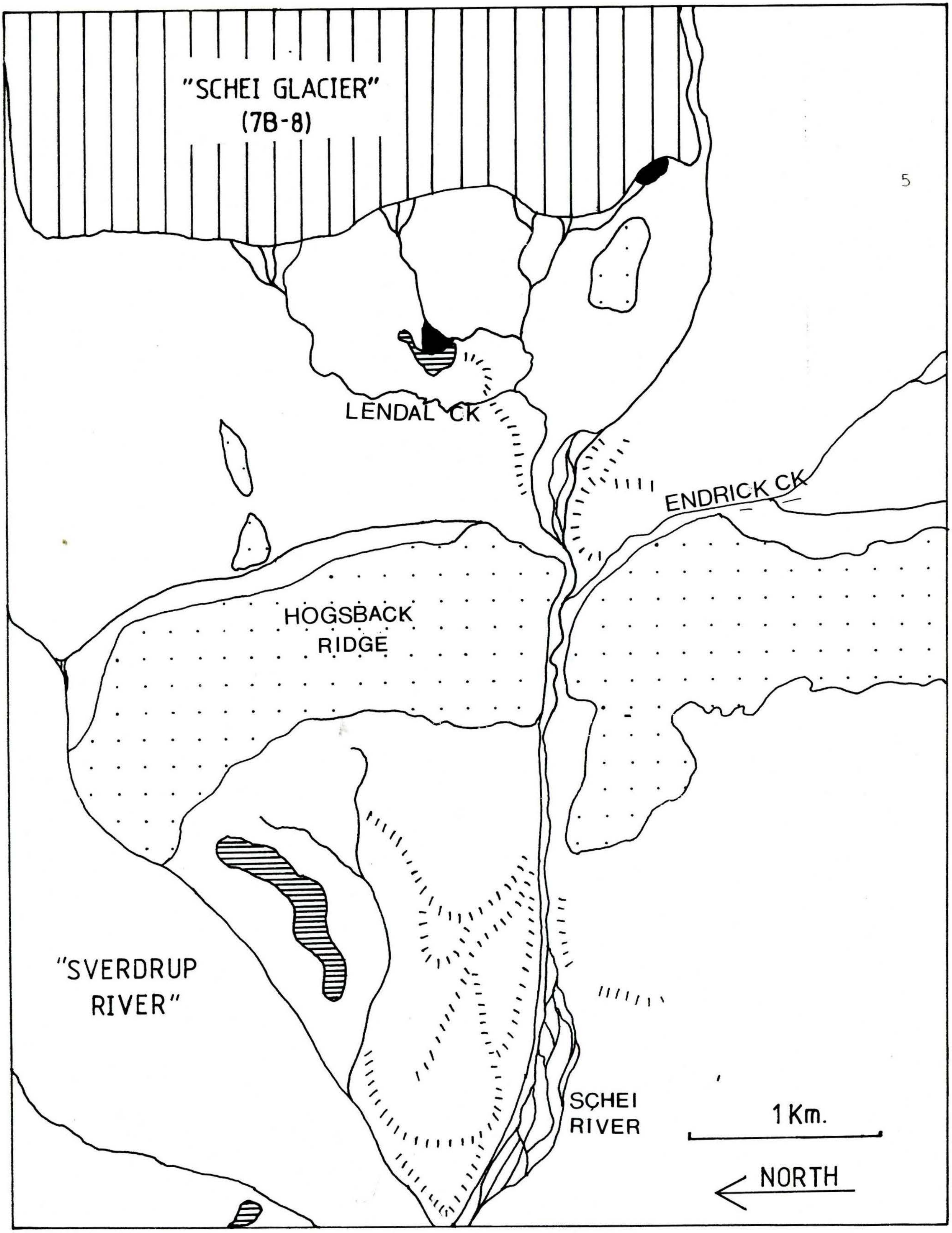
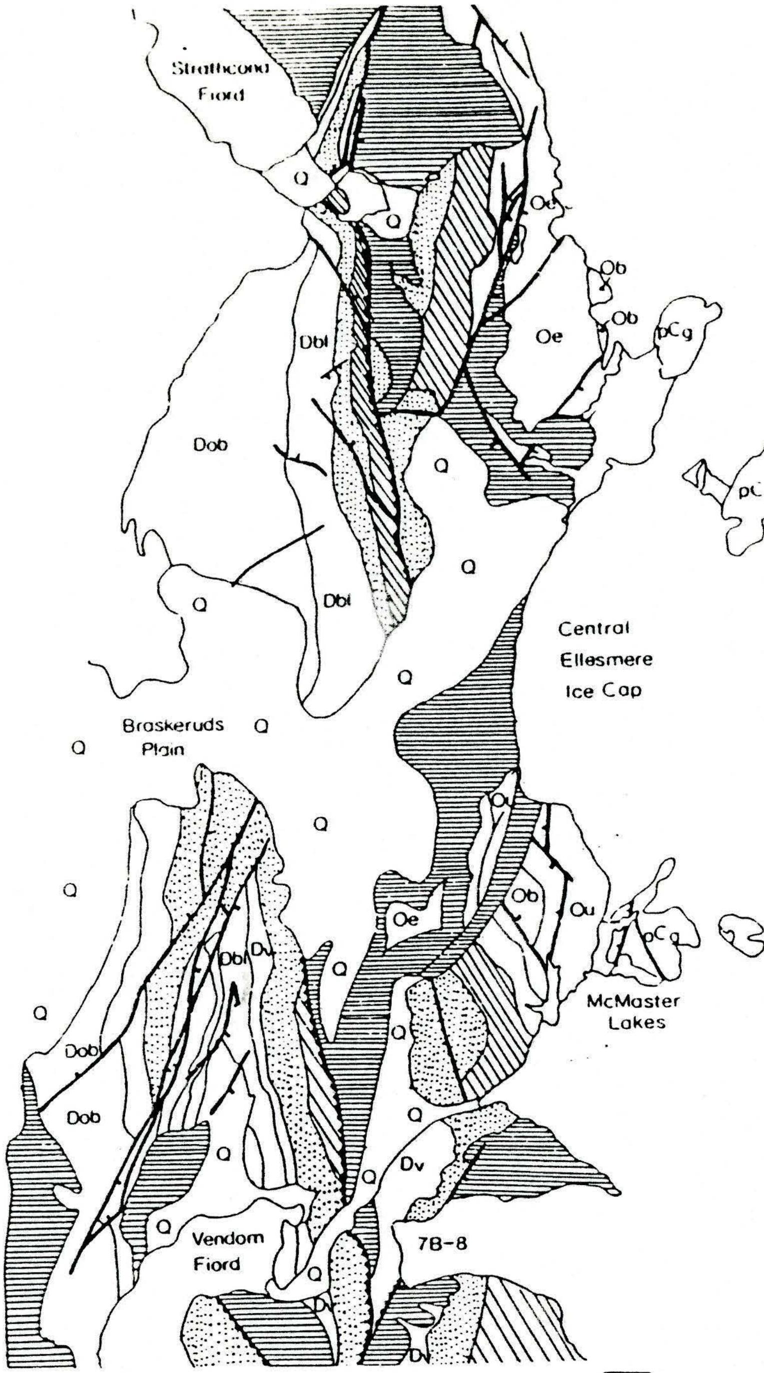


Figure 2. Vendom Fiord field area.

in the field area and the topmost member of the local sedimentary sequence is the Tertiary Eureka Sound formation, an assemblage described by Norris (1963) as heterogeneous and poorly bedded. This consists of shales, sandstones, and siltstones with some conglomerates and coal present.

Underlying the ice to the east, geology can only be inferred from nunatak outcrop in the Central Ellesmere ice-cap. Thorsteinsson (1972) identified these as being composed of Precambrian gneisses, granites and migmatites but these were not studied in detail. Frisch (1984) studied these outcrops further and assigned them Archean and/or Aphebian ages. Nunataks to the east of the "McMaster Lakes", at Rundfjeld, and to the southeast of Rundfjeld consist of hornblende-biotite gneiss and pegmatitic hornblende granite - the latter unit outcropping to the south of the former. Further to the east, approaching the watershed of the ice-cap, isolated outcrops consist of orthopyroxene granite and/or tonalite, orthopyroxene-bearing gneiss of granitic to tonalitic composition and of sedimentary and igneous origin, and other metasedimentary rocks mainly garnet-cordierite-sillimanite-biotite gneiss (Frisch, 1984).

Faulting and folding during the late Devonian Ellesmerian orogeny probably caused the uplifting of the "Hogsback Ridge" feature (Thorsteinsson and Tozer, 1960)



Q = Quaternary stream, deltaic, glacial, and marine beach sediments.

= Tertiary Eureka Sound formation - sandstone, siltstone, shale.

Dob = Devonian Okse Bay formation - sandstone, siltstone, shale.

Dbl = Devonian Blue Fiord formation - limestone and dolomite.

Dv = Devonian Vendom Fiord formation - sandstone and siltstone.

= Ordovician/Silurian/Devonian Allen Bay and Read Bay formation - dolomite and limestone.

= Ordovician Cornwallis group - limestone and dolomite.

Oe = Ordovician Eleanor River formation - limestone.

Ob = Ordovician Baumann Fiord formation - anhydrite and gypsum.

Ou = Lower Ordovician (undivided) - limestone, dolomite and shale.

pCg = Precambrian gneiss, granite and migmatite.

Figure 3. Geology of the field areas (after Thorsteinsson, 1972).

whilst the Tertiary Eurekan orogeny, marked by extensive thrust and normal faulting, resulted in local stratigraphic displacement.

Winter circulation in the Canadian Arctic Archipelago is dominated by the presence of an anticyclonic system which initially develops in the western Arctic and then expands eastwards. This anticyclone is responsible for the cold, dry weather which persists throughout the arctic winter, and does not begin to weaken until May, when eastward-moving depressions pass north of the Parry Channel (Maxwell, 1982). Only those few depressions which pass north of Parry Channel contribute to the total annual precipitation. September and October are marked by stormy weather and relatively high precipitation when cold air from the Arctic Ocean and Beaufort Sea triggers atmospheric instability (Meteorological Branch, Department of Transport, 1970).

The nearest weather station, Eureka, (80°00′ N, 85°56′ W; elevation 10m A.S.L., 450km distant) has the lowest mean annual precipitation for any station in Canada with 58.4mm. 49.6% of the mean annual total precipitation falls as rain during the months of June, July, and August; the remainder accumulates as snow or ice throughout the winter (Maxwell, 1982). An annual precipitation total between 80 and 100mm seems reasonable for both the Vendom Fiord and Strathcona Fiord field areas.

Mean annual daily temperature at Eureka is -19.4°C, in January mean monthly temperature is -36.6°C, and in July it is 5.5°C. Only June, July and August show mean monthly temperatures above freezing and there are 299 frost days per year (Maxwell, 1982). Extreme rainfall events have been recorded in which up to 54.66mm of rain was collected in an event (Cogley and McCann, 1975; Ballantyne, 1975) but these are very rare. Hydrologic and (most) geomorphic activity must be limited to the short summer season.

B. Strathcona Fiord.

The area under investigation is located at the head of Strathcona Fiord, southeast to "Foxtrap Lake" and then east to the present day ice margin. The northern limit of the field area is the "Walden Plateau". "Foxtrap Lake" is at latitude 78°34′ N, longitude 82°02′ W.

Geologically the two areas are fairly similar with all the formations present in the Vendom Fiord field area also present in this field area. In addition there are outcrops of members of the middle Ordovician Cornwallis group - the Irene Bay formation consisting of limestone with interbedded green shale, and the Thumb Mountain formation which is limestone. Other formations not seen at Vendom Fiord are the lower and middle Devonian limestones and dolomites of the Blue Fiord formation, the lower and middle Ordovician

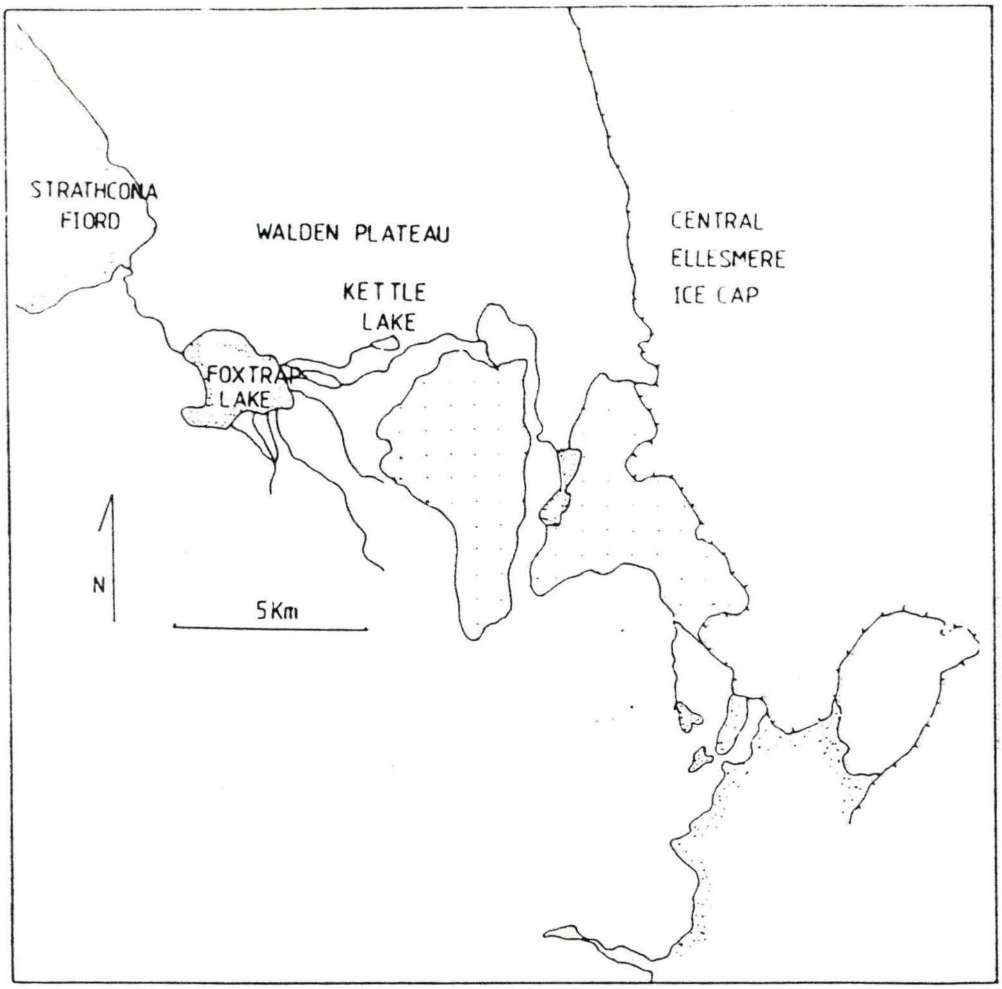


Figure 4. Strathcona Fiord field area.

limestones of the Eleanor River formation, and the anhydrites, gypsums and limestones of the lower Ordovician Baumann Fiord formation - the boundary of which approximately coincides with the ice front at the extreme east of the field area.

The area is more heavily faulted than the Vendom Fiord area as it falls within the limits of the heavily faulted belt which begins to the west of Vendom Fiord and runs north-northeast to the east of Irene Bay. The topographic manifestations of these faults are the upthrust ridge running approximately north/south between "Foxtrap Lake" and the head of Strathcona Fiord, and the deep valley running approximately north/south 3km west of the ice front. The boundary between the Palaeozoic sediments and the overlying Tertiary Eureka Sound formation appears as a topographic feature which marks the northern limit of the field area. This boundary has been eroded to form a broad, flat-bottomed valley on the northern side of which various members of the formation are well exposed. In contrast to Norris' (1963) description the strata here are well-bedded and members clearly defined as horizontal or moderately dipping beds.

Partly as a result of the area's bedrock geology, and partly because of its Quaternary history, this area exhibits a much more rugged topography than the Vendom Fiord field area and the types of Quaternary deposits found in the area reflect this. The Precambrian geology of the nunataks in the Central Ellesmere ice-cap is described in the Vendom Fiord site description as is the climate of the area.

C. Modern ice margin.

In addition to extended field work in the two areas described above, a reconnaissance flight was made by helicopter along the present day ice margin from Hook Glacier at the northern end of Makinson Inlet (77°35′ N, 82°24′ W) to Augusta Bay (78°50′ N, 81°40′ W). Photographs were taken and observations made from the helicopter and three landings were made to examine the ice front more closely. A fly camp was also set up at the extreme east of the Strathcona Fiord field area and observations of the ice were made from here.

1.3: Field methods.

Field areas were traversed on foot a number of times and interesting sections or sites noted. These were then returned to and detailed descriptions made of landforms, sediments, stratigraphy, etc. Some vertical sections required a considerable amount of digging and cleaning-off of debris before observations could be made, a few others were relatively well-exposed. Sites were plotted on aerial photographs of the area and efforts were made not to miss any regions within the field area. Unfortunately, due to late snow melt in the Vendom Fiord area, the field area was limited in its extent, and in the Strathcona Fiord area work was limited to the northern side of the "muskox" and "wolf" rivers.

In addition to traverses made on foot, a reconnaissance flight was made by helicopter along the ice margin from Makinson Inlet to Augusta Bay. Photographs were taken to record the nature of the present ice margin and landings were made at four sites for more detailed investigation.

Stratigraphic relationships in sedimentary sections are presented as sedimentary logs using the systems of Gardiner and Dackombe (1983) and Boersma (1975) and relationships between landforms are represented using geomorphological maps with symbols from Gardiner and Dackombe (1983), Cooke and Doornkamp (1974), and Blachut and Muller (1966).

CHAPTER TWO: REVIEW OF LITERATURE

2.1: High latitude glacial deposystems.

Since Ahlmann (1948) first proposed a thermal classification of glaciers, it has become increasingly apparent that the concept of a high latitude ice body being frozen to its bed over its entire area is too simplistic to be useful. Theoretical considerations by Robin (1955) showed that temperatures at the base of "cold" ice bodies may vary temporally and spatially, perhaps even causing surging behaviour. Since then a number of authors have shown through both theory and field investigations that complex patterns of thermal conditions exist at the base of supposedly "cold-based" ice bodies (Schytt,1964; Loewe, 1966; Boulton, 1972; Lawson, 1977; Andrews,1982). In many cases it appears that the margins of these ice bodies are frozen to the substrate whilst upstream, where ice thicknesses are greater, pressure melting point is exceeded at the base.

Acceptance of the existence of warm-based areas beneath a "polar" ice body is vital to explain the presence and distribution of debris within the ice. Drewry (1986) listed processes of debris entrainment without making any distinction between basal thermal states and it appears that the differences due to temperature lie in efficiency and rate of entrainment rather than process. Weertmann (1961) first proposed that ice-debris accretion may occur under relatively thick ice with changing thermal conditions at the bed. This is essentially a large-scale version of regelation which it is thought is the principal process of sub-glacial debris entrainment in warm-basd ice-bodies, yet Weertmann argues that changes in the position of the 0°C isotherm over time (in response to external stimulae) can lead to very much larger scale rafting of the refrozen zone into the ice (Drewry, 1986).

Another sub-glacial entrainment process, which assumes a priori the existence of a frozen-to-substrate outer margin is that proposed by Moran et al. (1980). These authors used the ideas of Weertmann (1961) to show how blocks of ice and frozen substrate could be sheared upwards as pressure builds up behind the frozen and largely immobile ice margin. This process would clearly disrupt pre-existing patterns of debris foliation within the ice and generate a suite of glaci-tectonic structures which may be reflected in deposits after the ice has melted. Finally, a process which is not restricted to predominantly cold-based glaciers, but appears to be most effective in that environment, is that of overriding and incorporation of frontal aprons (Shaw, 1977, 1985). Ice bodies which are cold-based at their margins often exhibit steep marginal cliffs against which fallen ice blocks, super- and englacial debris, and wind-blown deposits may accumulate as aprons or ramparts (see figure). Advancing glaciers may override this debris and incorporate it subglacially.

Super- and englacial entrainment of debris are relatively less important in "cold-based" ice bodies than they are in warm-based glaciers. This is because in a high latitude environment low precipitation will greatly slow burial and incorporation of superglacial debris. Also, the impermeability of the ice will prevent washing-in of debris through intraglacial drainage. Rockfall and avalanche inputs are less important in the large ice-domes of the Arctic and Antarctic as the relative areas affected are smaller than those in a warm-based alpine glacial setting where much of the ice is confined to rock-walled valleys. Still, locally the contribution from nunataks and valley walls may be very important in a high latitude setting. Atmospheric input of particulate matter is of little interest to sedimentologists but high latitude ice bodies, because of their large area, relatively predictable flow conditions and rates, regular accumulation, and distance from point sources, are an important store of atmospheric particulates. Much work has been done to trace changing patterns of atmospheric particulate content using ice cores from high latitude ice bodies (Windom, 1969). Volcanic ash horizons in ice sheets are also useful dating horizons as individual events can often be lithologically characterised and dated (Bradley, 198).

Having described theories of debris entrainment, the principal interest to the sedimentologist is the distribution of debris during transport and then process of deposition. A very visible difference between warm-based and predominantly cold-based ice bodies is the thickness of the basal debris layer (Boulton, 1970; Shaw, 1977; Lorrain et al., 1981; Eyles and Miall, 1986). High latitude ice bodies commonly have very thick basal debris layers because, in the absence of subglacial drainage, debris and ice will continue to accumulate subglacially until the margin is reached or the ice retreats or stagnates. In this way, thicknesses of 7m (Shaw, 1977), 10m (Lorrain et al., 1981), 30.5m (Boulton, 1971), and 11m (see chapter 5) of debris can be observed at ice margins[1]. This is in contrast to the basal debris layers seen in warm-based ice bodies which are generally only centimetres, exceptionally up to one metre, thick (Boulton, 1971; Eyles and Miall, 1986).

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[1] However, Andrews (1971, 1972) argues that these thick basal debris layers are only marginal features.

Deposition of englacial and superglacial debris takes place when the transporting medium, ice, melts and releases the sediment from entrainment. In a warm-based ice body this melting may take place subglacially during the process of lodgement, superglacially during normal ablation, or at many sites simoultaneously during stagnation. Usually the water released during melt transports all or part of the debris which is deposited downstream. In a truly cold-based ice body lodgement cannot occur. In a partly cold-based ice body it is unlikely to occur, or, more correctly, if lodgement were to occur during subglacial melting, according to Weertmann (1961) the debris would later be re-entrained during ice-debris accretion as the 0°C isotherm shifted. Thus, lodgement till would have an extremely low preservation potential in such an environment.

Release of debris at the surface during ablation is an important process in all glacial systems, yet in high latitudes the rate and amount of ablation will be considerably less than in temperate alpine areas. Boulton (1970, 1971, 1972b) presented many examples of melt-out of debris and subsequent flow of that debris from sites in Spitzbergen. There is little mention elsewhere in the literature of this process being observed on active ice margins (as opposed to stagnant) yet this was observed in the field (figure).

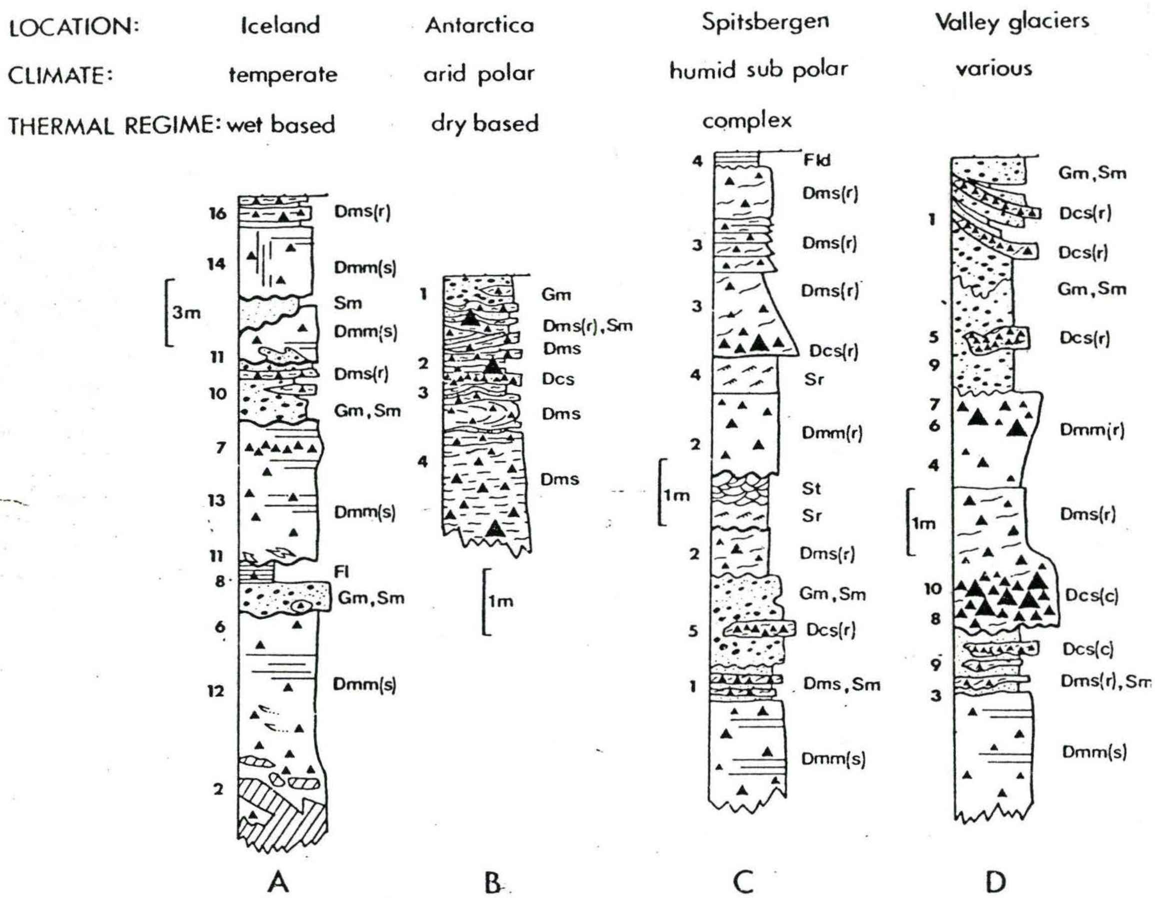


Figure 5. Vertical profile models of diamict sequences deposited by glaciers of differing basal thermal regime (after Eyles et al., 1983).

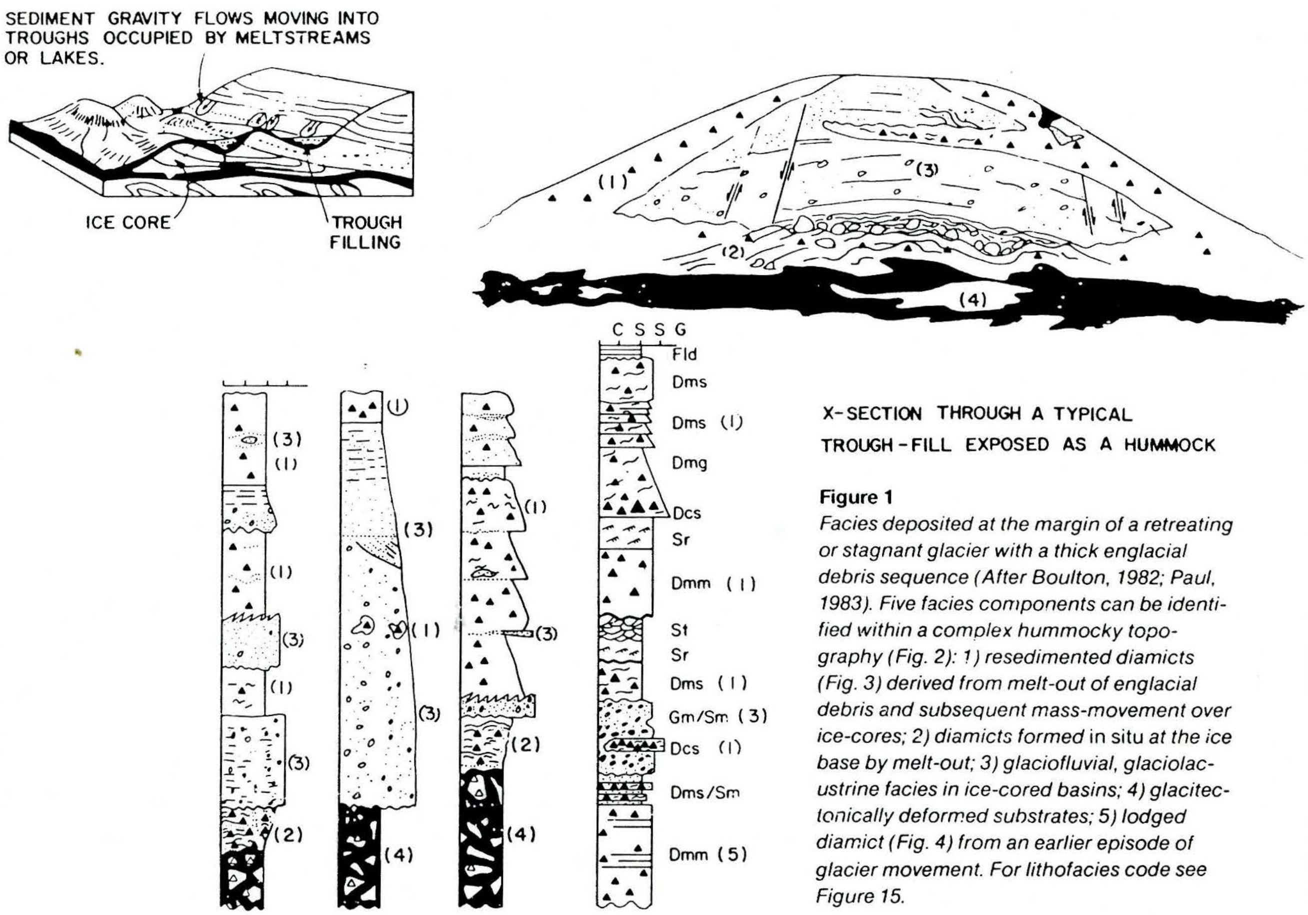


Figure 1
Facies deposited at the margin of a retreating or stagnant glacier with a thick englacial debris sequence (After Boulton, 1982; Paul, 1983). Five facies components can be identified within a complex hummocky topography (Fig. 2): 1) resedimented diamicts (Fig. 3) derived from melt-out of englacial debris and subsequent mass-movement over ice-cores; 2) diamicts formed in situ at the ice base by melt-out; 3) glaciofluvial, glaciolacustrine facies in ice-cored basins; 4) glacitectonically deformed substrates; 5) lodged diamict (Fig. 4) from an earlier episode of glacier movement. For lithofacies code see Figure 15.

Figure 6. Facies deposited at the margin of retreating or stagnant glacier with a thick englacial debris sequence (after Eyles and Miall, 1984).

The best understood depositional processes in high latitude glacial environments are those associated with stagnating ice. This is perhaps because the processes do not reflect the previous thermal regime of the ice. Shaw (1977) suggested that sublimation was a major process in sedimentation from stagnant ice in the McMurdo dry valleys of Antarctica. This process, it was hypothesized, preserved much of the fine debris foliation seen in basal layers and preserved glacitectonic structures. If this process was widespread it would result in a deposit characteristic of arid, polar glacial environments and, as such, a vertical profile containing "sublimation till" was incorporated into the model of Eyles _et al_. (1983) (figure 5). However, work reported by Robinson (1984) indicated that sublimation is not an important process of ice removal in the Dry Valleys and that most ice melts basally due to geothermal heat flux.

Eyles _et al_. (1986) subsequently suggested that the vertical profile which illustrated "sublimation till" in Eyles _et al_. (1983) (figure 5) resulted from slow basal thawing. They concluded that

> "processes that deposit debris at the base of subpolar and temperate glaciers also operate to deposit lithofacies sequences at arid polar outlet glaciers but at much reduced rates and on different scales." (Eyles _et al_., 1986, p.153.)

Processes of deposition from stagnant ice and the forms which result from these processes are well described by Embleton and King (1968), Shaw (1985), Drewry (1986) and Eyles and Miall (1986). There are only a few forms thought to be characteristic of, or more common in, high latitude glacial deposystems. Embleton and King (1968) proposed that Thule-Baffin moraines (Goldthwait, 1951; Weertmann, 1961) were characteristic of cold-based ice bodies. These are formed as debris is sheared upwards from the base and then onto the surface where differential ablation rates due to dirt coverings will cause ice-cored terminal moraines to be formed. Whilst this is not now thought to be characteristic of high latitude ice bodies, moraines which result from very large-scale glacio-tectonic thrust structures in stagnant ice may be. As high latitude ice bodies have much more debris in their basal layers, the forms resulting will tend to be considerably larger than those resulting from similar structures in stagnant temperate ice (Boulton, 1971,1972b).

Eyles and Miall (1986) claimed that the thicker basal debris sequences would cause stagnant or retreating margins to become quickly covered by a diamict drape which would then be resedimented downslope by sediment gravity flow into local basins generated by the irregular melt of buried ice (see also Lawson, 1979, 1982). Landforms which result from this include hummocks, which reflect these basin fills after topographic inversion has taken place due to ice melt, and ridge-like landforms oriented transverse to former ice-flow direction, which are remnants of englacial glacitectonic structures (Shaw,1979).

Finally, the major interest of the glacial sedimentologist is the stratigraphy or vertical profile of the deposits referred to above. Literature on this is sparse but Eyles et al. (1983) presented a vertical profile for arid polar dry-based glacial deposits and those resulting from the humid sub-polar glacial environment with a complex thermal regime (figure 5). Eyles and Miall (1986) presented a more complex model of vertical profiles characteristic of stagnation of ice with a thick basal debris layer (figure 6). In this model they identified five facies components:

1. Resedimented diamicts derived from melt-out and subsequent mass movement as described above;
2. Diamicts formed in situ at the ice base by melt-out (Shaw, 1977; Robinson, 1984; Eyles et al., 1986);
3. Glaciofluvial and glaciolacustrine facies in ice-cored basins (which are similar to those described in other environments by Jopling and McDonald (1975), Smith (1985), Smith and Ashley (1985), Miall (1983), and Quigley (1983) but are more prone to resedimentation and hence have a lower preservation potential;

4. Glaciotectonically deformed substrates; and
5. Lodged diamict from earlier glacial episodes.

The results described in the following chapters will hopefully make a useful addition to the literature in this area and will provide more data for testing and refinement of the model.

2.2: Late Quaternary history of west-central Ellesmere Island.

As mentioned in section 1.1, there has been considerable debate over the form of the last glacial advance in the Queen Elizabeth Islands. Briefly stated, there are two opposing views: one advocates the existence of a large contiguous ice-cap during late-Wisconsinan time, the Innuitian ice-cap (Blake, 1975, 1976, 1981; Mayewski *et al*., 1981; Denton and Hughes, 1981), whilst the opposing view is that pre-existing ice-caps expanded slightly yet did not amalgamate (England, 1976, 1978, 1985, 1987; England and Bradley, 1978; England *et al*., 1978; Dyke *et al*., 1982) (figure 7). There have been a number of excellent review articles outlining the evidence and drawing conclusions from it (Mayewski *et al*., 1981; Denton and Hughes, 1981; Paterson, 1977; Andrews, 1982; Fulton, 1984), so this review will concentrate on literature appropriate to the immediate area concerned - west-central Ellesmere Island.

Figure 7. "Maximum" and "minimum" models of late-Wisconsinan ice cover (after Mayewski et al., 1981).

Evidence from this area is actually fairly sparse as England has concentrated on sites in northeastern and northern Ellesmere Island and the northwestern coast of Greenland; whilst Blake has reported from sites in east-central, southeastern, and southern Ellesmere Island. Blake (personal communication, 1986) is working at a number of sites in west-central Ellesmere Island but as yet only results from sites at the head of Makinson Inlet (Blake, 1981) have been published. England (personal communication, 1986) is also moving southwards to work at sites on Greely Fiord and beyond, but as yet nothing has been published.

Hodgson (1973, 1979, 1985) has presented data from west-central Ellesmere Island which provides some clue as to the late-glacial history (figure 8). This author, however, did not place himself firmly into either camp, and at the conclusion of his most recent paper (Hodgson, 1985) presented three alternative models of the last glaciation. Two of the three alternatives presented the data in the context of the two previously outlined paradigms, whilst the third hypothesized an expansion of pre-existing ice-caps combined with formation of new ice-bodies from which outward flow originated. Very strong evidence exists for one of these ice bodies being centred on the Braskeruds Plain and, as large areas of presently unglaciated Ellesmere, Axel Heiberg, and Devon Islands lie above or close to the elevation of this plain, it is assumed that other new ice bodies also formed during the late-Wisconsinan period.

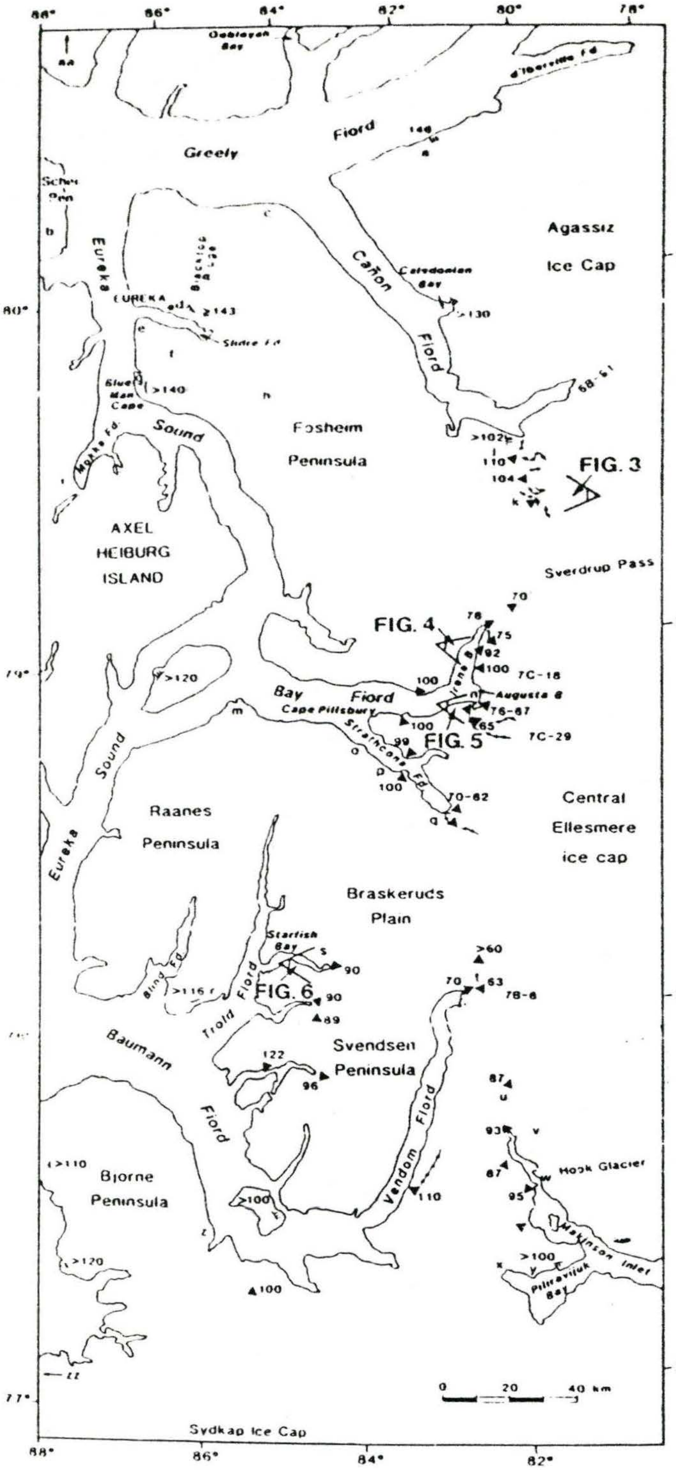
The late glacial history of the Vendom Fiord area, as reconstructed by Hodgson (1973, 1979, 1985) and Ballantyne (1975), is presented in the following chapter. There has been no previous work published on the Strathcona Fiord area and a tentative late-glacial history for the area apppears in chapters four and six of this thesis.

CHAPTER THREE: VENDOM FIORD FIELD AREA

3.1: Introduction

Over the past fifteen years the area at the head of Vendom Fiord has been subject to a number of scientific investigations. McMaster University parties concentrated primarily on the hydrology of the "Schei River" and its basin (McCann et al., 1972, 1974,1975; Woo, 1975; Ballantyne,1975; Cogley and McCann, 1975) and on a study of jokulhaups at the "McMaster Lakes" to the north (Blachut, 1975). At the same time D.A. Hodgson of the Geological Survey of Canada's Terrain Sciences Division was working on the reconnaissance mapping of surficial materials in central Ellesmere Island (Hodgson, 1972) which then developed into study of the late-glacial history of the immediate area (Hodgson, 1973) and, eventually, into ideas relevant to an even more extensive area (Hodgson, 1985).

Whilst most of the results from McMaster University research are not of any direct relevance to the present thesis, useful data was incorporated in Ballantyne's (1975) thesis. Another useful source of information, one very helpful in organising both fieldwork and results, was G.S.C.



RADIOCARBON DATES

Loc.	Age (years BP)	Lab No.	Related Sea level (m)
aa	8 280 ± 150	GSC- 593	> 83
a	8 450 ± 100	GSC-2369	>127
b	8 080 ± 160	I(GSC) 264	?
c	8 190 ± 110	GSC-1822	> 91
d	9 550 ± 250	L-6478	>110
	8 880 ± 120	GSC-1800	>125
	4 220 ± 130	GSC- 597	11
	3 420 ± 130	GSC- 589	> 3.5
	2 710 ± 130	GSC- 590	> 1.5
e	36 300 ± 2000	GSC- 111	
	19 500 ± 1100	I- 548	
f	28 700 ± 600	GSC- 51	
g	8 710 ± 140	GSC- 254	>140
h	>41 200	GSC- 268	
i	7 980 ± 150	GSC- 350	> 88
	6 010 ± 140	GSC- 351	> 37
	4 960 ± 130	GSC- 274	> 24
	4 830 ± 130	GSC- 273	> 15
	3 130 ± 140	GSC- 174	> 6
	2 900 ± 130	GSC- 267	> 3
	2 420 ± 140	GSC- 171	> 1.5
j	7 950 ± 130	GSC-3023	> 91
k	>38 000	GSC-3016-2	
	>16 000	GSC-3016	
l	9 140 ± 310	GSC-3388	> 80
	8 820 ± 90	GSC-1978	> 80
	7 340 ± 170	GSC-3397	> 75
m	1 380 ± 130	GSC- 452	> 2.5?
n	6 370 ± 100	GSC-118	> 37
o	7 230 ± 90	GSC-3728	78-85
p	7 750 ± 160	GSC- 170	> 76.5
q	7 680 ± 150	GSC- 175	
	6 780 ± 80	GSC-3765	> 56
r	8 480 ± 140	GSC- 244	>116
s	8 710 ± 120	GSC-2719	> 72
t	7 010 ± 80	GSC-1858	> 54
	6 980 ± 90	GSC-1957	> 53
u	7 730 ± 80	GSC-1972	> 37
v	>52 000	GSC-2677	
	>44 000	GSC- 140-2	
	>36 400	GSC- 140	
w	9 270 ± 110	GSC-3180	> 65
x	8 930 ± 110	GSC-2519	> 42
	8 090 ± 70	GSC-2692	> 82
	2 060 ± 50	GSC-1836	> 4.5
y	29 800 ± 220	GSC- 134	> 90?
z	30 100 ± 750	GSC-2700	>108?
zz	8 590 ± 150	GSC- 840	>107

ice cap

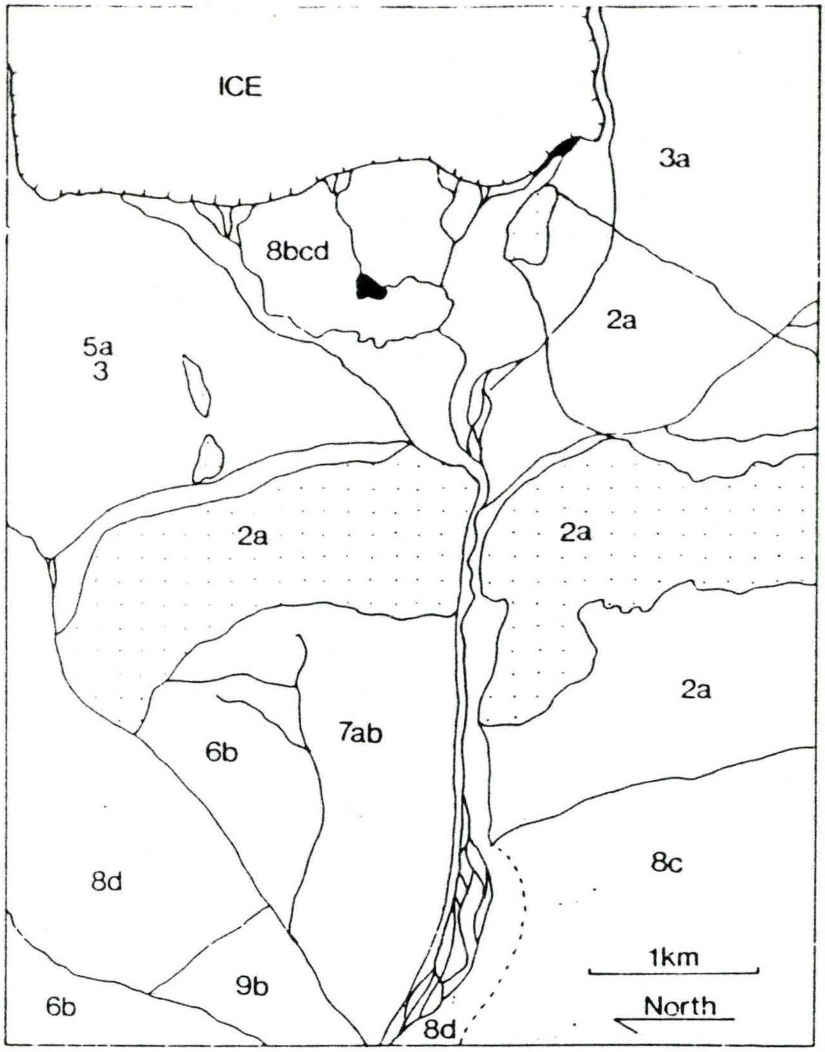
end moraine

kame delta (elevation m)

glaciomarine rhythmites

highest marine features (elevation m)

Figure 8. Map of area showing glacial features, radiocarbon sites and dates and altitudes of uplifted features (after Hodgson, 1985).



1. Precambrian rock and residual rock

2. Carbonate and Calcareous clastic rock (Palaeozoic)

   a. Generally resistant, high relief
   b. Generally recessive, moderate relief

3. Clastic, minor calcareous rock (Phanerozoic)

   c. Fine- and coarse-grained recessive rock

4. Plateau gravel

5. Morainal deposits

   a. undifferentiated till
   c. coarse-grained Holocene advance till
   c. neoglacial moraines

6. Marine sediments

   b. Fine-grained marine sediments
   c. Relict coarse-grained beaches

7. Delta sediments

   a. Coarse-grained relict proglacial delta sediments
   b. Coarse-/fine-grained inactive delta sediments
   c. Coarse-grained active delta sediments - nival regime
   d. Coarse-grained active delta sediments - glacial regime

8. Fluvial sediments - coarse-grained

   b. Fluvial terrace and inactive fan sediments
   c. Active valley flat and fan sediments - nival regime
   d. Active valley train sediments - glacial regime

9. Fluvial sediments - fine-grained

   b. Active valley flat sediments

10. Colluvium

Figure 9. Surficial materials at Vendom Fiord (after Hodgson, 1979).

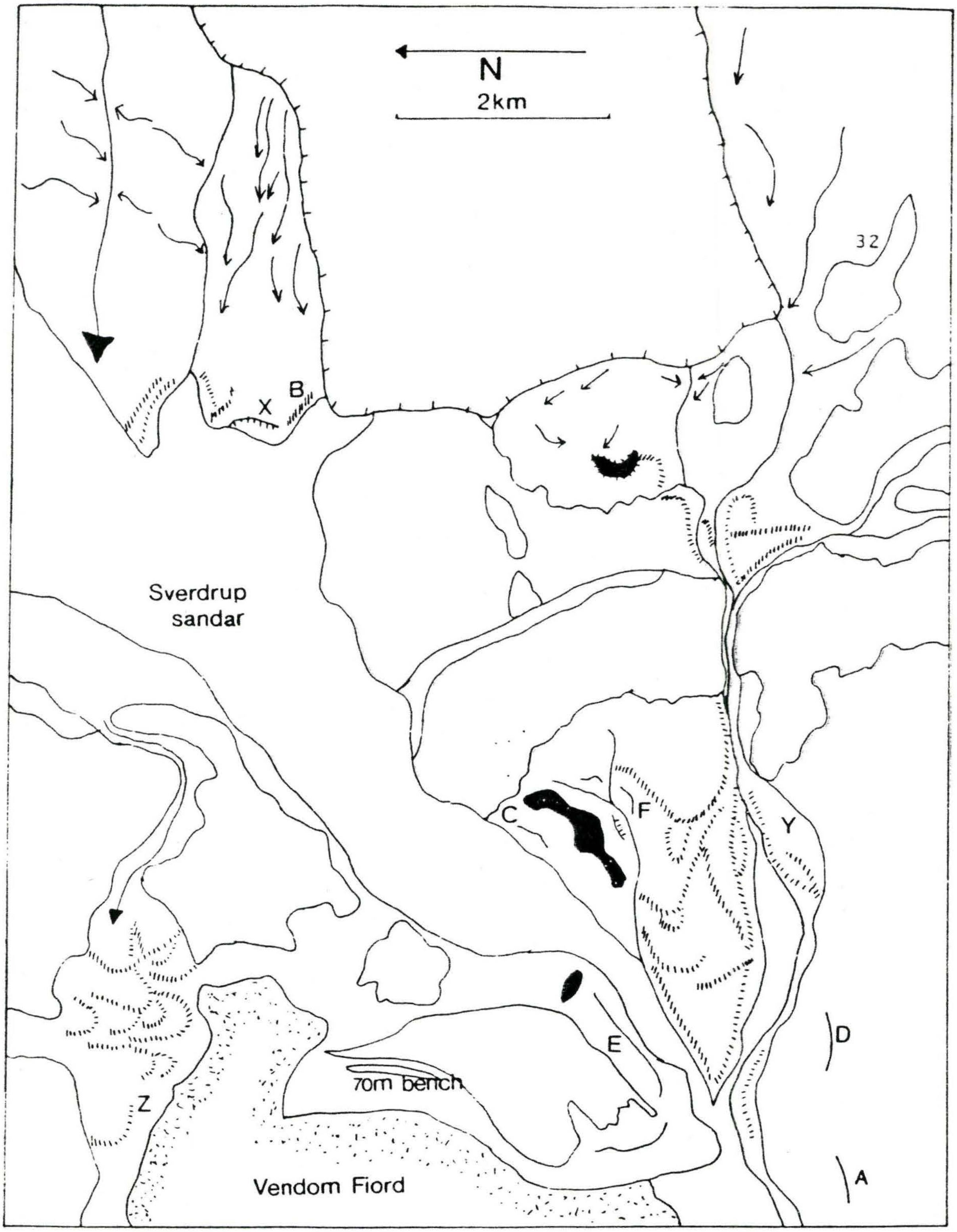


Figure 10. Landform map of Vendom Fiord area (after Ballantyne, 1975).

Open File 635 (D.A. Hodgson, personal communication, 1986; Hodgson, 1979) which is a classification of surficial materials in the area based on aerial photograph analysis and traverses in the field.

Data from previous work

Hodgson (1973) examined the question of the late-glacial history of the area by means of landform study and interpretation. Ballantyne (1975) used a similar approach. In this section their evidence is presented and in the next section new sedimentological data will be presented for part of the area.

Hodgson (1973) presented much of his data in the form of a map of geomorphological units which differs very little from that in Open File 635 which is shown as figure 9 [1]. Features Hodgson identified as being important are marginal meltwater channels and perched outwash deltas (see figure 10). Altitude measurements made on two perched deltas and a wave-cut beach yield sea levels which nowhere exceed an elevation of 70m whilst marine or estuarine silts in the area nowhere exceed an elevation of 65m. Remnants of terraces and "an extensive ice-contact outwash surface" were

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[1]This latter figure was chosen because it represents a more complete piece of work - that appearing in the 1973 publication is the result of a preliminary investigation.

also identified (Hodgson, 1973, p.134). Materials suitable for radiocarbon dating were found in one terrace remnant at location x, below delta topset beds at location y, and on a flight of beach ridges at an altitude of 5m asl. at location z.

Hodgson (1973) specifically noted that no end moraines were found in the area but Ballantyne (1975, p.46) identified "an arcuate ridge rising to 64m on either bank of the Sverdrup sandur" which he interpreted as an end moraine. This feature is composed of rounded and sub-rounded gravels of heterogeneous composition with a large component of crystalline Precambrian lithologies. Hodgson had noted (p.131) that the marine silt in this area had a discontinuous gravel cover but did not assign the observation any importance.[2]

Further evidence cited by Ballantyne refers to an area of hummocky topography, defined by a very distinct limit, which is situated on the "even surface of the highest terrace...at the distal end of the Schei Gorge" (Ballantyne, 1973, p.46). As adjacent areas of the high terrace, composed of similar material, do not have the same form, frost heave and ground ice action were rejected as causes

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[2] An important fact to note here is that pits dug by Ballantyne indicate that neighbouring silts were deposited against the flanks of the ridge, indicating that the ridge predates the silts.

for the topography seen.

Just across the river at location q an exposed scarp of a perched delta reveals horizontal or near horizontal beds of fine materials which have been truncated downstream by alternating foresets of gravel and sand (Ballantyne, 1975, p.46 fig 3.4). Both sets of beds are then truncated by a metre of sandur gravels, the elevation of which, at its distal end, is 58.9m.

Finally, Ballantyne presented data on the elevations of undisturbed marine silts. These were examined at locations a-m on figure m. The presence of paired bivalves and intact rhythmites in the silts indicated that these deposits were not actively associated with a glacial regime. Gravel veneers are all derived locally from either glacial diamict deposits or talus cones upslope. Elevations of these silts and of other features indicate a sea level of 65-70m at the time of the most recent readvance. Thus, the silts outside the glacial limit must have been deposited before withdrawal of ice from the readvance maximum as they all achieve an elevation of over 60m whilst those inside that limit must postdate that maximum as they are undisturbed.

In the area to the east of the upstanding "Hogsback ridge", Ballantyne tentatively identified three relict sandur surfaces and a ridge of reworked outwash gravels which has been deposited on top of the old sandur surface-this latter feature is interpreted as an end moraine from an advance which postdates the "Vendom Fiord stage" of Hodgson (1973). Three meltwater channels are also identified and these are thought to mark the progressive retreat of the ice from the end moraine. Ballantyne also tried to relate this advance to sea level by identifying faint strandlines from a (presumed) ice-marginal lake which emptied onto a younger relict sandur formed by the "Upper Schei River". The strandlines correspond approximately to the level of this sandur at the lake's outlet and these were related to sandur surfaces at the head of the fiord via terrace remnants through the "Schei Gorge". Although little remains of the associated terrace downstream the next highest terrace extends to the confluence of the "Schei" and "Sverdrup" rivers where it is only 2.3m. above present sea level. Therefore this second readvance must relate to a lower sea level and indicates a probable neoglacial age. Any conclusion based on such sea level data should be advanced only tentatively, however, as sea level change over the past millenia has been very slow.[3]

In the most recent published work on the area, Hodgson (1985) briefly reiterated the conclusions from his earlier

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3 A log collected at approximately 4m. asl. (thought to relate to sea level of ≥4.5m.) at Makinson Inlet yielded a ^{14}C date of 2060 ±50 B.P. (GSC-1836, Lowdon and Blake, 1978; Hodgson, 1985).

paper and then noted that

> "thick till and marginal channels on the more rugged east side of the fiord outline a former glacier with a shallow surface gradient (<12m/km). The glacier terminated near a kame delta 110m asl.; the delta is not necessarily glaciomarine, it may be glaciolacustrine." (Hodgson, 1985, p.360)

The kame delta to which Hodgson refers above is over 60km down fjord from the present head. He, unfortunately, did not provide any field evidence to back up this assertion but did note that the absence of ice-marginal landforms on the west side of the fjord may be ascribed to ice originating from the Braskeruds Plain rather than to the north east. An end moraine indicated on his figure 2 (figure 8) (Hodgson, 1985, p.349) which trends roughly northeast/southwest or subparallel to the fjord bank would tend to support this.

3.3: Original field data

A. Ice marginal zone

Disappointingly, the debris apron at the foot of the ice cliffs at the snout of the "Schei Glacier" remained snow covered and frozen throughout the field season. Some observations were made in this area, however. Figure 11 shows a small mudflow on the snow covering the debris apron. This is a small-scale, highly fluid flow which is channelised in its mid section and then forms lobes in its depositional area. This is believed to be a Lawson type II-III flow (Lawson, 1979; Drewry,1986) which acts to resediment deposits at the snout of the glacier.

It is possible such flows are initiated by the impact of, and lubrication provided by, the many slush flows observed on the snout of the glacier. These proceeded with a frequency of six to ten per hour on sunny days and appeared as "plugs" of liquid slush flowing down the snout behind a more solid flow nose, leaving a shallow channel behind them. The remaining channels varied in size from 0.1 to 0.5m in width; depth could not be estimated. Although there was no stream flow in the area usually occupied by a marginal channel there was a depression in the substrate at the foot of the debris apron which followed the ice margin. During the field season this was filled with melting snow and, latterly, thixotropic mud greater than 75cm deep. Sediment flow deposits were observed in this area and it appeared that flows ran parallel to the ice margin along the channel, although the only flow observed in action was the one described above.

At site 23 a number of solifluction lobes were identified which had flowed off the sides of the bedrock

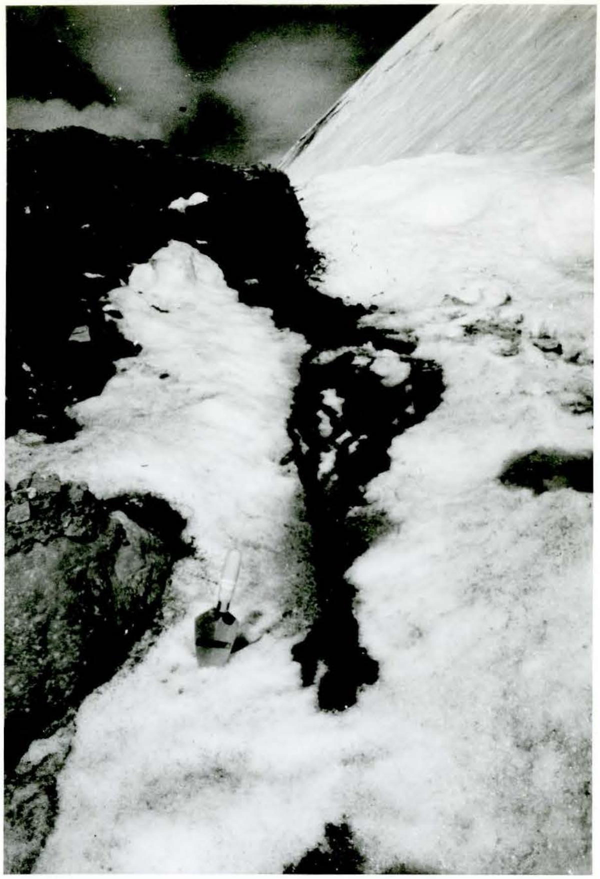


Figure 11. Small mud flow just in front of debris apron.

Figure 12. Solifluction lobe at site 23.

Figure 13. Overland rill flow on the distal side of the marginal channel.

knoll. One was surveyed and dug and a photograph of it appears in figure 12. The flow is 58m long and 15m wide defined by levees at each side for the lower two thirds of its length. At the flow nose a 69cm deep pit to the permafrost table revealed the underlying surface at 58cm depth on which preserved woody roots and stems were found. Overlying this contact is a matrix-supported unit of angular pebble and cobble-sized clasts set in silty-clay matrix. Clast frequency decreases upwards and between 38 and 24cm depth less than 10 clasts are seen. Above 24cm depth root incision and winnowing has opened up some pore space in the deposit; a few clasts are seen in the top 5cm but the majority have been heaved onto the upper surface where they lie in a matrix of fine sand, finer fractions presumably having been washed out.

Pits dug throughout the flow reflect a similar pattern with depth of the permafrost table varying from 52.5 to 74cm depth. Coarse debris is concentrated at the nose, in the levees, and on the upper surface of the flow. Very shallow sit and clay lobes extend beyond the nose and sides for 5-10m downslope.

Eyles and Paul (1983) calculate a value of 1.5° for the minimum slope on which solifluction can take place. This hillside has a mean slope of 22° and varies between 29° and 32° in the source area of the flow, so solifluction of the active layer over a frozen surface would appear to be the process in operation here. The sections indicate that this is a Lawson type II-III flow (Lawson, 1979, 1981; Drewry, 1986) if one applies the criteria used for flowtilll to solifluction deposits.

In addition to the processes of cryoturbation and solifluction which conspire to disorganise stratigraphic records, overland rill flow observed shortly after snowmelt (figure 13) also acts to winnow and resediment the surfaces of near-ice sediments. Even on a vegetated, stabilised surface this flow serves to remove fines from among the coarse clasts heaved to the surface to create the "armoured crust" seen in sections 6, 8, and 9 (figures 15, 16, 17) and over much of the proglacial zone.

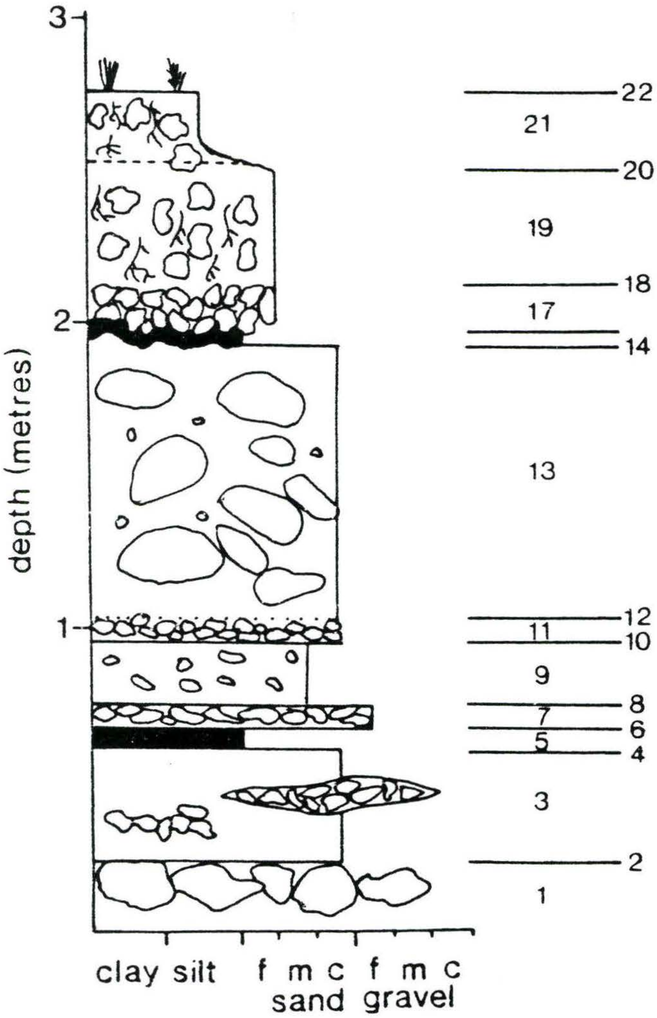
A pit dug at site 22 has this surface layer below which is a deposit consisting of an amalgam of silt and sand in which clasts are rare. The permafrost table lies at a depth of 37cm. The surface is sparsely covered with Salix arctica and isolated examples of Saxifraga. The site is at the centre of an extensive, gently sloping fan feature defined by "Lendal Creek" and the "Upper Schei River" and it appears that this whole feature is a flow feature similar to that described at site 23 but on a much larger scale. It appears that modifying periglacial processes are very active in this ice marginal zone and that sediments will tend not to preserve glacial-deposition structure but that inherited from post-depositional alteration.

B. Proximal proglacial zone.

The proximal proglacial zone exhibits landforms and sediments characteristic of fluvioglacial, paraglacial and some glaciolacustrine environments - part of what Eyles (1983) has termed the glaciated valley landsystem. Many sections were dug in this area and sedimentary logs for these sections appear as figures 14 to 18, a location map appears as figure 19.

The commonest lithofacies observed was a massive, matrix-supported cobble and gravel diamict which is seen in section 5 (unit 13), section 6 (unit 3), and section 16 (unit 3). If similar units with finer matrices are included, almost all of sections 5 to 9 is accounted for. This lithofacies is coded Dmm after Eyles et al. (1983) and is interpreted as periglacially altered glacial diamict. It has already been noted how deposits have been subject to mass movement and many authors (Lawson, 1977, 1979; Eyles and Miall, 1984) have identified the importance of resedimentation in glacial environments where ice has a thick basal debris layer.

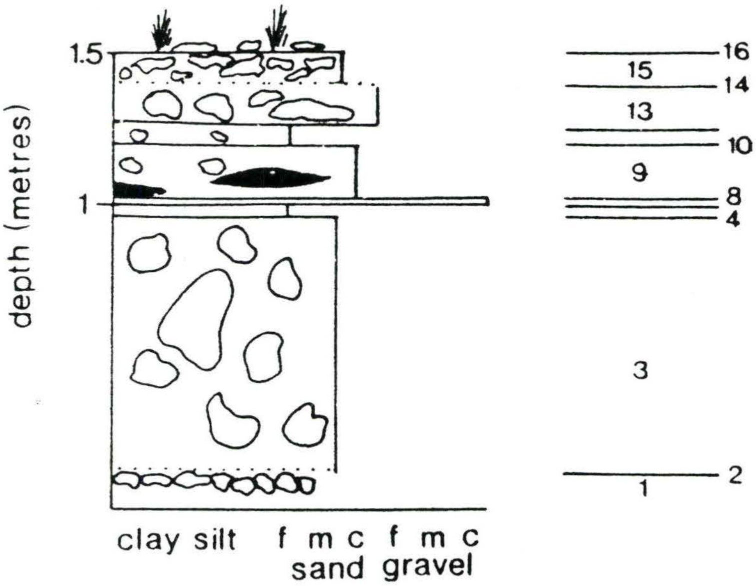
Dominant matrix size is difficult to identify in the field and varies from site to site, but what is interesting



15. Black peaty or burnt layer.

7. Clast-supported bed composed almost entirely of coal or charcoal pieces 10-20mm. long.
5. Very compacted, partly frozen bed, peaty? Contains rootlets. No clasts seen at all.
3. Massive, clean and very loose sand body - no real structure - locally lenses of coarse material.

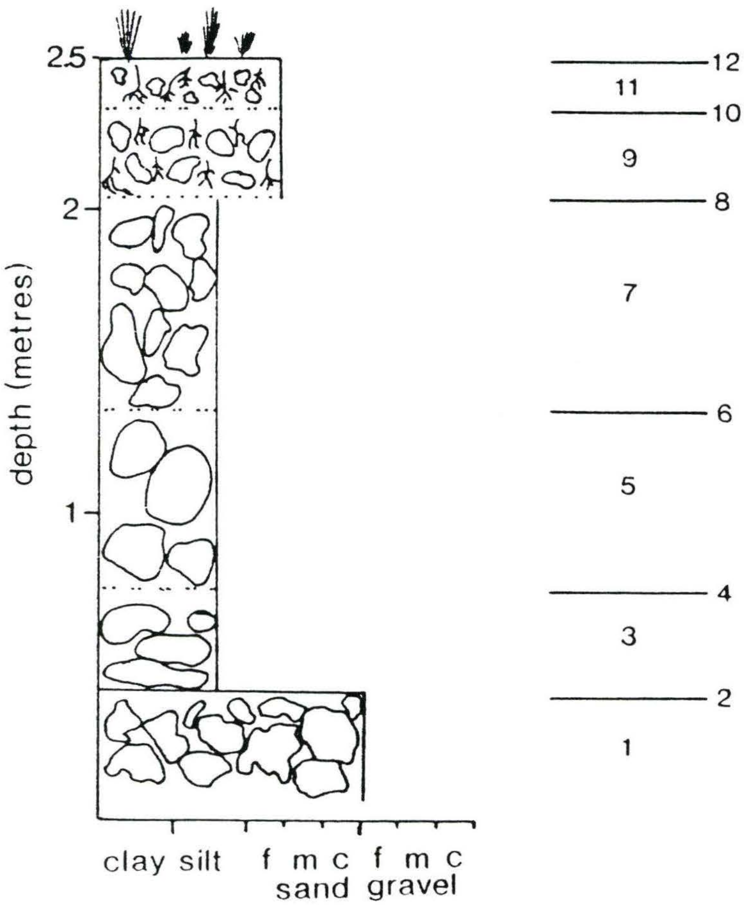
Figure 14. Section 5.



15. Some horizontal tendency in clast orientation.

13. Clasts of varied lithology.

11. Unit contains stains of darker colour - a few small clasts.
9. Unit contains some pockets of peat and/or fine coal fragments.
7. Bed of coal particles 2-3mm. in size - bed is discontinuous but laterally traceable for metres.
5. Clast free unit
3. Clasts rounded to sub-rounded, various lithologies. size from 2-500mm.
1. Clast-supported, clasts rounded and approx. 50mm. in diameter. Matrix clean white sand.



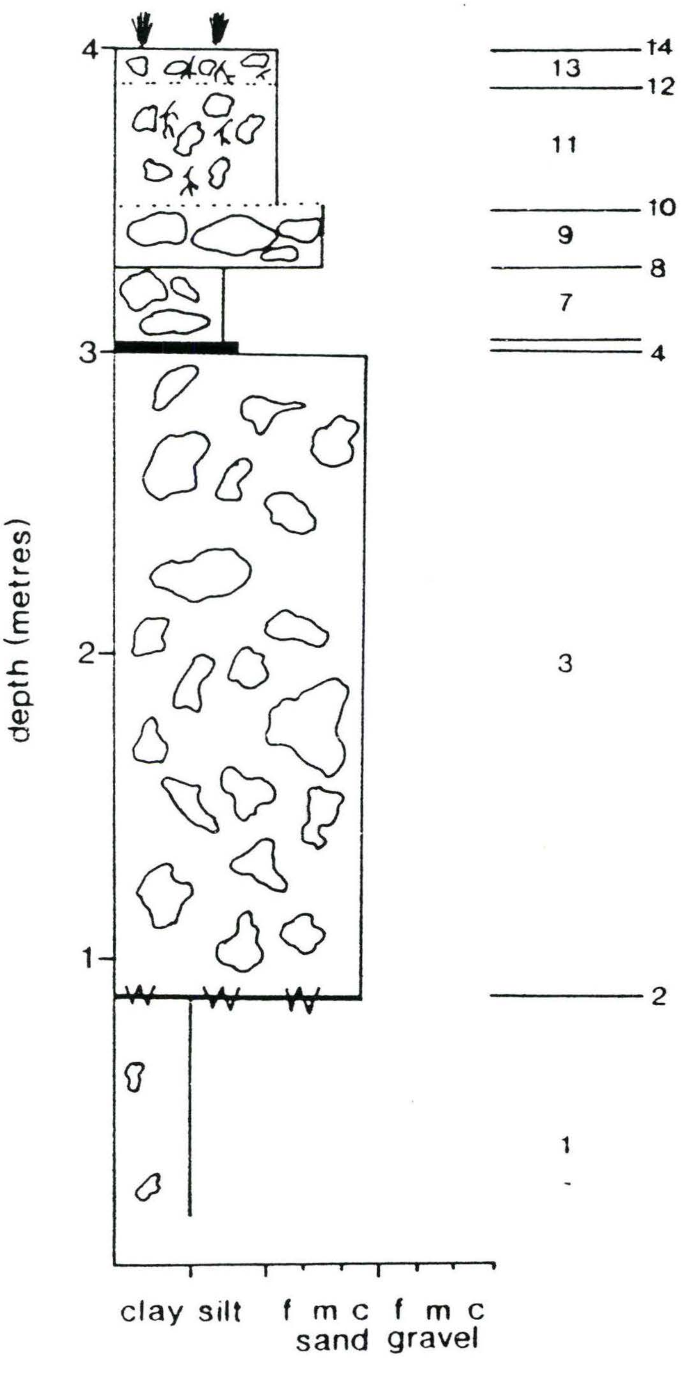
9. Matrix slightly darker brown than 11. and clasts are larger (50-150mm.) mostly rounded, no orientation, various lithologies.

7. Clasts up to 400mm., mean approx. 150mm. - more angular than above apart from largest. Almost clast-supported, no preferred orientation.

5. Clasts generally larger and more rounded than 7, more matrix-supported.

3. As for 5 except this is a unit of larger clasts 300-500mm in diameter.
1. Almost clast-supported polymict, matrix is coarse, clean white sand combined with sub angular/sub-rounded fine gravel. Roundness of clasts varies as does lithology. No orientation. Matrix discontiuously permafrozen.

Figure 15. Sections 6 and 7.



13. Clasts of various lithologies and roundness, no preferred orientation. 8-50mm. diameter.

11. Larger and more angular clasts than 13. 25-50mm in diameter.

9. More loamy matrix, clasts are larger and more rounded than those in 11. Larger component of gravel-sized clasts.

7. Mixed size range of rounded clasts, locally some areas are clast-free.

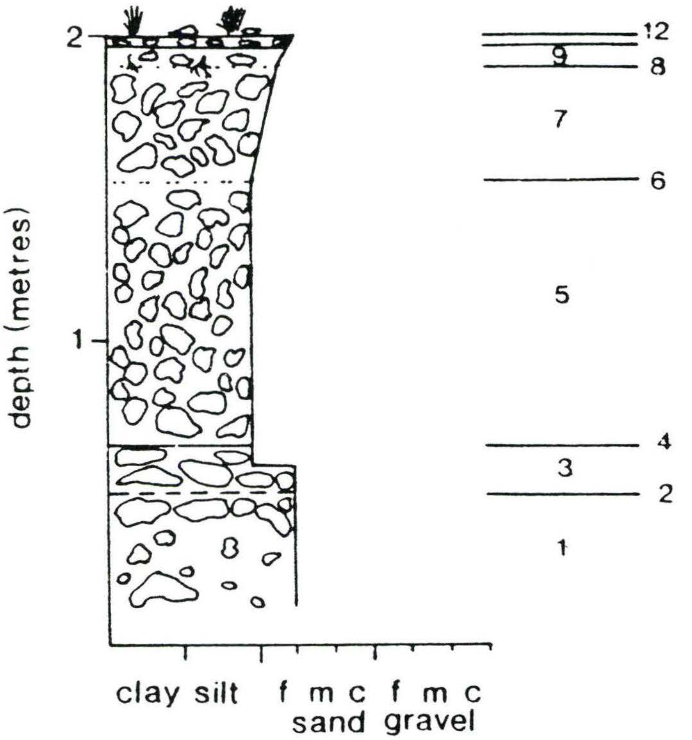
5. Silty unit of ?washed-in peat - partly frozen-some vegetal matter visible. Rare clasts up to 50mm. in diameter. Very dark brown/black.

3. Clasts range from gravel to 250mm. in size, sub-angular to sub-rounded. Various lithologies. Matrix is frozen towards bottom.

2. Sharp, erosive contact.

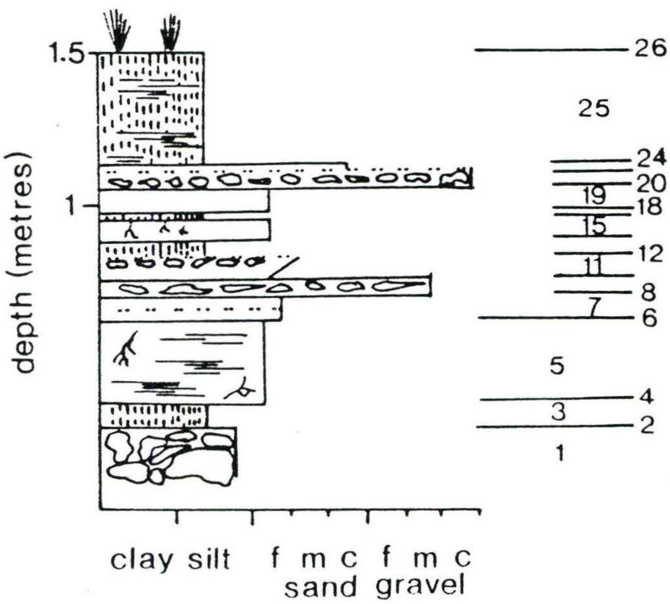
1. Permafrozen clay, structureless, rare angular clasts approx. 50mm. in diameter.

Figure 16. Section 8.



11. Surface crust of small angular clasts, horizontal planar orientation. Mixed lithologies.
9. Sub-angular to sub-rounded clasts varying from <5 to50 mm. Locally complex root mats.
7. As 9. but higher concentration of clasts, no preferred orientation, no sorting.
5. Regular clast size approximately 50-70mm., sub-angular to sub-rounded, various lithologies.

3. Clasts are large boulders up to 400mm. and cobbles approx. 100-150mm.
1. Boulders/cobbles seen in 3. in upper part, below this same clasts as in 5. but in coarser matrix with occasional cobbles - no preferred orientation.



25. Light brown dry, dusty peat formed of layers approx. 80mm. thick. Silty bands seen in peat and also fragments of coal are seen throughout unit.
21. Coarse gravel/pebbles vaguely horizontally orientated. No imbrication evident in unit as exposed. Becomes clast-supported in lower parts.
15. Fine sand/silt partly bound together by roots from overlying unit.
13. Open mossy peat, some pebbles and gravel caught in bottom layers of peat.
9. Unit of coal and shale fragments horizontally orientated.
7. Fine sand with orange staining in upper parts.
3. Peat with much void space and crystalline ice.
1. Polymict of clasts mainly 20-50mm. but up to 100mm. (large clasts are rounded) in silt matrix, no orientation or stucture.

Figure 17. Sections 9 and 21.

47. Dark to mid brown colour - bleached at the surface, some horizontal structures discernible.

45. As 47. but slightly darker and more dense, no structure evident.

41. Deformed unit of silty and sandy layers - more vegetal matter in silt. Staining from that into underlying sands.

37. Silty/peaty unit with bleached sand unit at mid-height - some identifiably vegetal remains.

35. Alternating bands of sand and silt, each 5-10mm. thick. Sand up to silt is gradual change then abrupt change back to sand.

29. Bands of silt in sand unit - heavily iron-stained.

27. Silty/clay unit with rare vegetal remains.

25. Mid brown peaty/silty unit.

23. Barely discernible plane bedding.

21. Convoluted and deformed current bedding.

19. Well preserved mossy peat band.

17. Very faint plane bedding.

15. Amalgam of clay and coarse sand in approximately equal amounts.

11. Very well-preserved mossy peat mats - mid to light brown in colour.

9. Black peat more compact than 11. well preserved.

5. Bleached sand with many loading and flame structures - no current bedding.

3. Well-preserved peat with horizontal bedding.

1. Iron stained fine sand/silt with a few rootlets.

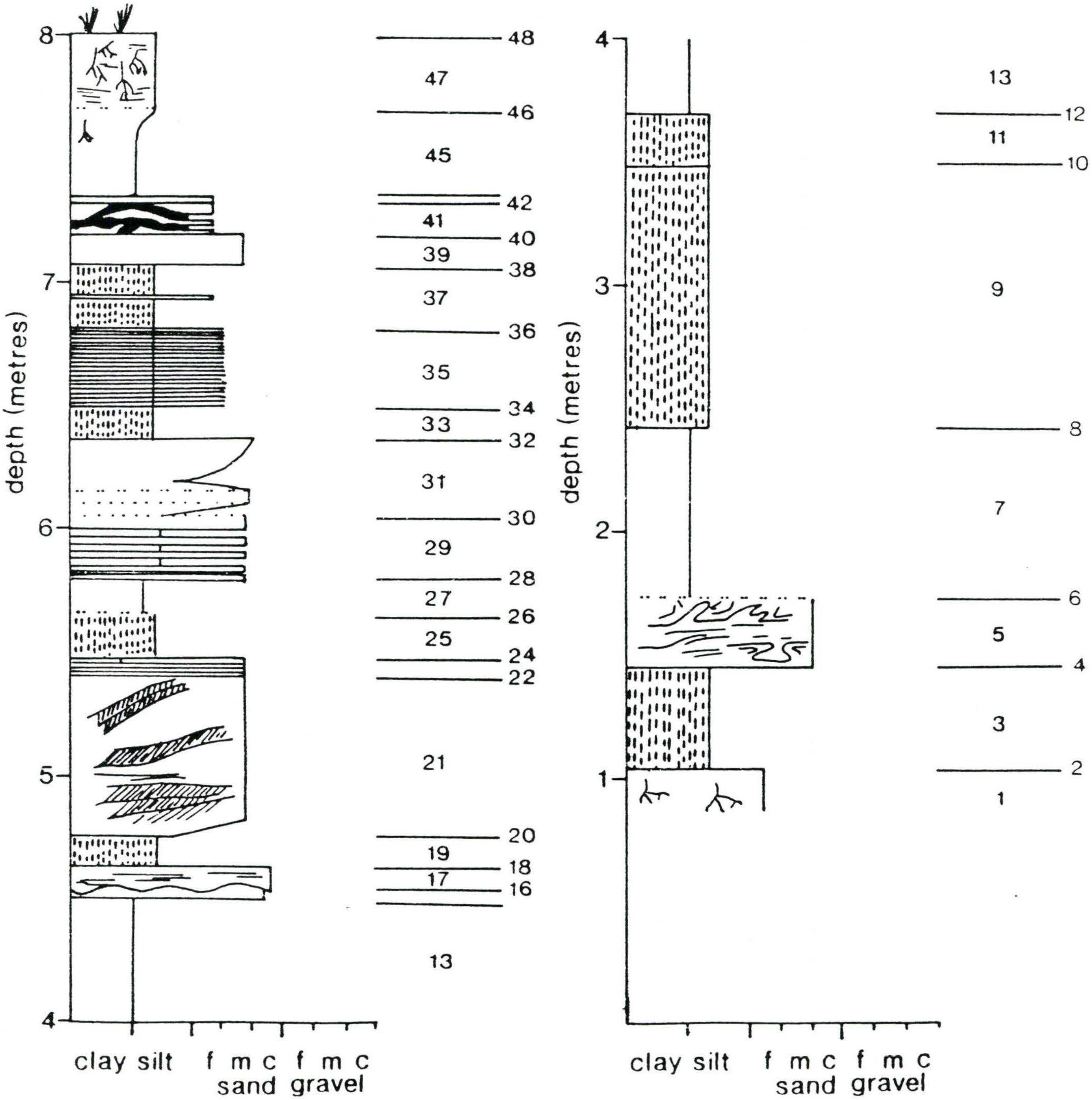


Figure 18. Section 16.

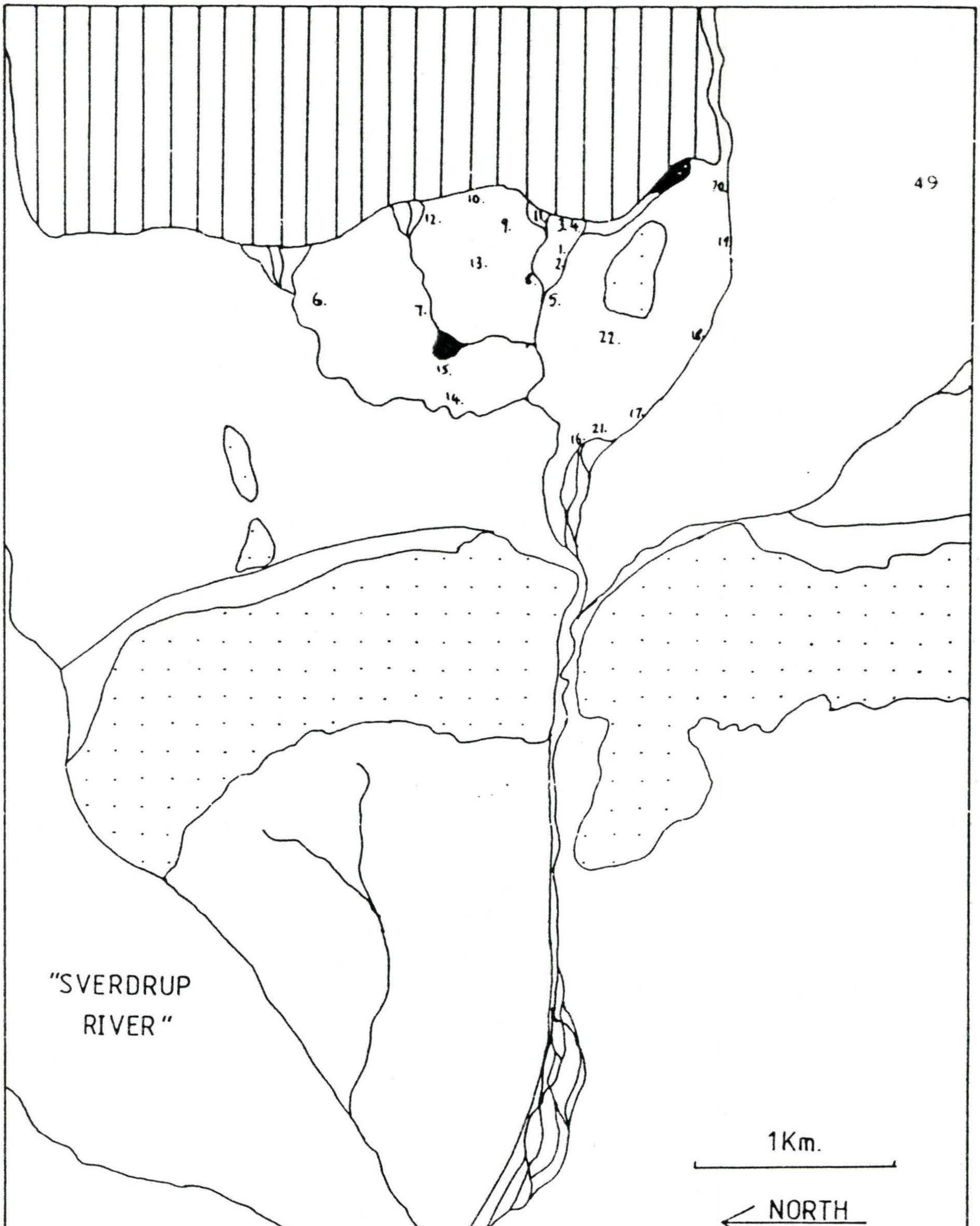


Figure 19. Location of sites and sections in Vendom Fiord area.

to note is the mixture of lithologies in both clasts and matrix. The geology of the area (figure 3) is interesting in that the boundary between Precambrian igneous and metasedimentary lithologies and Palaeozoic sedimentary lithologies approximately coincides with the present ice margin. Many of the clasts in the diamict are rounded to well-rounded gravels and cobbles and it appears likely that these are reworked fluvioglacial deposits from the local area although some may be derived from the conglomerates in both the Devonian Vendom Fiord formation and the Tertiary Eureka Sound formation. More angular clasts tend to be of all lithologies and it is hypothesized that these are debris derived from substrate or outcrop rather than reworked terrestrial sediment - these are much less common than the rounded to sub-rounded clasts. The matrix is usually an amalgam of sand sizes, silts and clays with one or more size being dominant at certain sites. In this terrestrial environment it is thought that the clays are formed by the breakdown of feldspars in the igneous rocks to form kaolinites in the sediment and this ties in with observations made later at the ice margin (chapter 6). Diamicts closer to the surface can have either a finer or a coarser matrix than underlying units but they are rarely the same. This is due to transport of finer fractions by the overland flow processes described earlier which lead to relative enrichment in fines in depositional areas and winnowing in source areas. Some units of the diamict do show some horizontal orientation but this is thought to be a relic of post-depositional mass movement as described by Lawson (1979) at the base of his type II and III flows.

A similar lithofacies to the one described above was also observed at sites 10, 12, 17, 18, 19, and 20. Sites 17 to 20 are all found along the northern bank of the "Upper Schei Gorge". Whilst safety precautions prevented close examination of an unconsolidated deposit on the banks of a 30-40m deep, vertically-sided incised gorge, it could be seen that in all places the Quaternary deposit overlies an abrupt, erosive contact with bedrock. The deposit varies in thickness from 1.5 to 2.0m at site 20 to 8.5m at site 17. At site 19, a well-sorted gravel conglomerate (a minor component of the Vendom Fiord formation; Thorsteinsson, 1972) overlies fine sandstone and, as the former is more competent than the latter, the rock is exposed and upstanding with the Quaternary deposits intermixed with talus at its margins.

Other lithofacies seen in the sections help shed more light on environments of deposition other than the reworking of glacial sediment. In section 5 there are a number of units consisting of gravels and pebbles which are clast-supported. Their occurrence below the previously described diamict units may mean that they are sieve deposits from the flows which resedimented the diamict (Miall, 1978) or they may be fluvial lag deposits or bar sediments from streams into which the diamict slumped and/or flowed. More identifiably fluvial in their origin are the deposits seen in unit 3 of section 5. These are coarse sands in which no structure has been preserved yet lenses of gravel and pebbles remain. Many analogs of this situation can be seen at present on the sandar in the region where medium sized, shallow channels move sand rapidly along the bed whilst coarser material is isolated in bars within the braided channel. The overlying sequence of a peaty bed overlain in turn by a clast-supported gravel indicates a flow bringing in vegetation and fines - probably in a very liquid state- and then the nose of a flow in which coarser clasts are relatively concentrated. If the fines and peat had been deposited during gradual abandonment of the channel there would be a gradual fining rather than the rapid transition observed. As other sections show peaty and muddy sediments were being laid down in the area during the Holocene, it is quite likely that these could be mobilised as easily as, if not more so than, the diamicts. Unit 9 in section 6 also shows lenses of muddy, peaty material which adds weight to the explanation.

Section 8 shows a very interesting contact just above the permafrost table. This sharp, erosive contact lies between the diamict previously described and an underlying clay in which there are few clasts. This clay must represent a lacustrine deposit in a proglacial environment, perhaps with a little ice-rafted debris being brought in from the margins. More detailed examples of lacustrine facies are seen in sections 16 and 21 (figures 18 and 17). In these sections peats can be seen interspersed with laminated sands and silts and chaotically distorted sand bodies. Without more detailed investigation it would be impossible to say whether the sediments are varved but many coarser units undoubtedly reflect individual events when sediment load was increased in inflow streams, by turbidity flows into the lake, and possibly the limits of terrestrial mass movements just reaching into the lake (Hakanson ; Church and Gilbert, 1975; Ashley, 1975; Evenson <u>et al</u>., 1977; Luckman, 1975; Smith and Ashley, 1985). Peat formation can be observed in the field area at present although it is doubtless very slow (Fenton,); what is more remarkable is how well-preserved the peat is, even at 7m depth - this is due to its being permafrozen fairly soon after formation as upper layers of peat act as a very good insulator to prevent lower layers from ever thawing.

The only current bedded sand encountered in the field area is seen in section 16 (figure 18). This may record an episode during which the lake dried up and a branch of the braided rivers preserved elsewhere migrated over the lacustrine deposits, or it may be a result of an overland flow into shallow waters at the edge of the lake. No consistent current direction is recorded as is normal with braided channel deposits and the sands have been deformed either shortly after deposition or with increasingly deep burial some time later.

3.4: Summary.

These, then, are lithofacies encountered in the proximal proglacial zone at Vendom Fiord. Major depositional environments appear to be mass movements of glacially deposited diamict, and glacially influenced rivers and lakes. Even in the fluvioglacial and glaciolacustrine deposits the influence of glacial and periglacial mass movement of sediments can be identified.

CHAPTER FOUR: STRATHCONA FIORD FIELD AREA

4.1: Introduction and previous work.

In contrast to the Vendom Fiord field area, the Strathcona Fiord field area has had almost no previous investigation. The sole references to the area in the literature were made by Hodgson (1979, 1985) and Dyck and Fyles (1964). These authors discuss the landforms at the head of the fiord but little mention is made of landforms or deposits inland of here.

Hodgson (1985) reported that westerly ice flow was recorded by erosional and depositional glacial and ice-marginal landforms and that patches of till flanked the fiord to at least 300m asl. Abundant marine shells and fragments in till south of Strathcona Fiord indicate that a regional westerly ice flow crossed the fiord and excavated marine sediments from it. Dyck and Fyles (1964) claimed that it was ice from the last glaciation that inundated the fiord and cited a dated peat horizon (7680 ±150 years B.P., GSC-175; Dyck and Fyles, 1964) overlying till at 400m asl to prove this. Hodgson (1985) disagreed with this and argued that the "till" was in fact colluvium. To the northwest, in

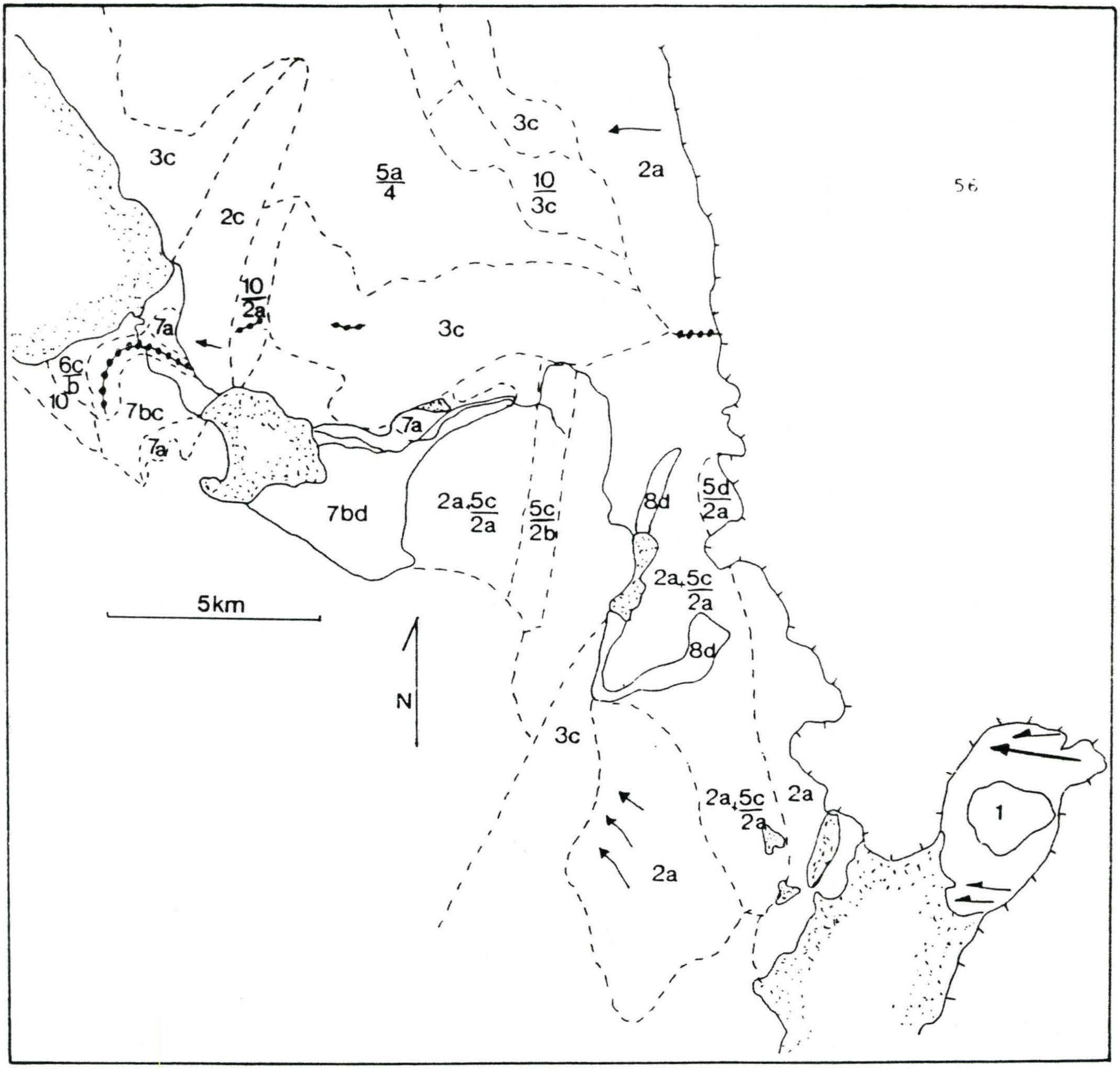


Figure 20. Map of surficial materials (Hodgson, 1979)

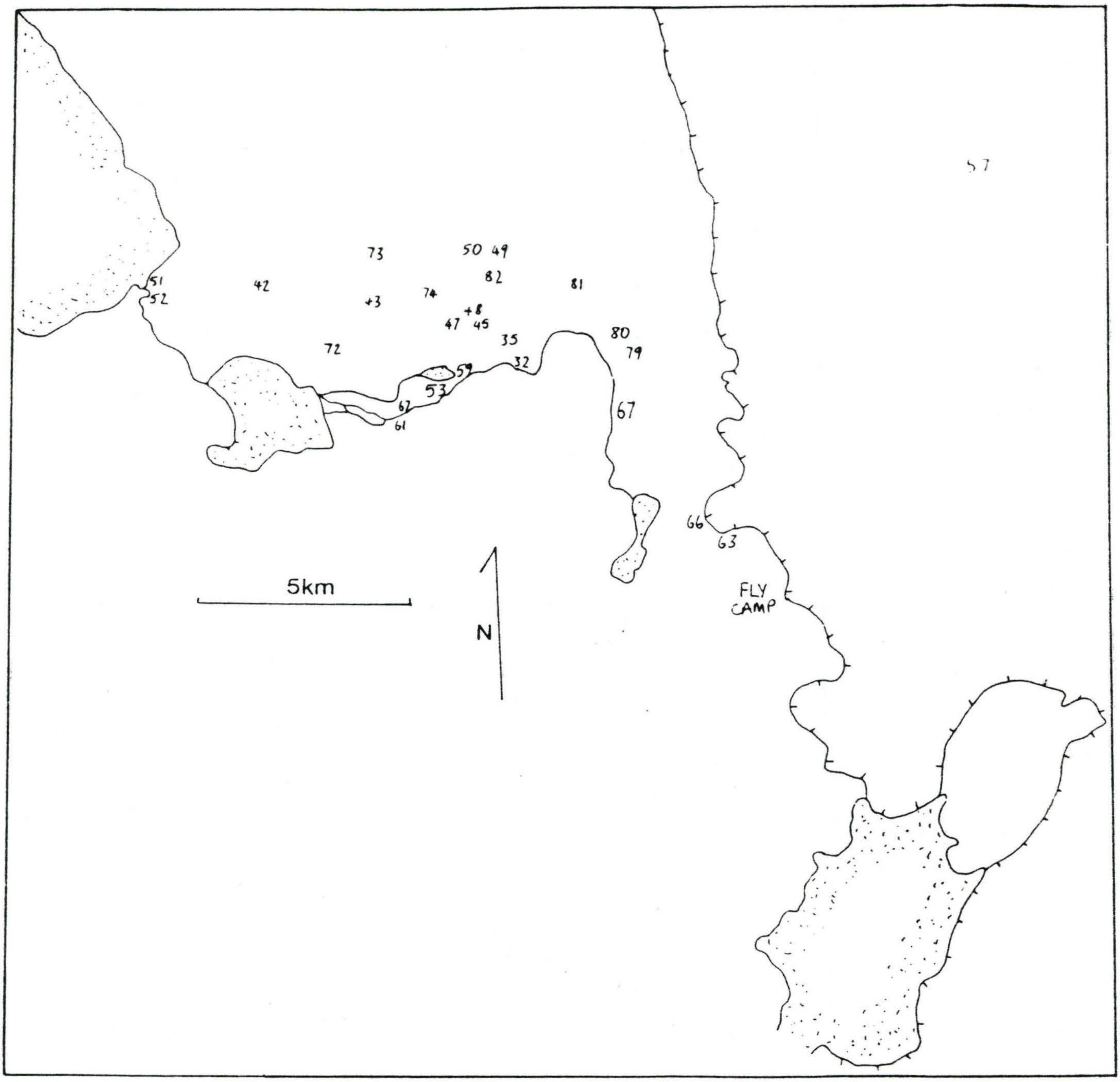


Figure 21. Location map of the Strathcona Fiord field area

Bay Fiord, however, Hodgson did report a thick, unweathered till overlain by Holocene marine deposits but accepted that it could not be determined how old the till was.

Kame deltas on Strathcona Fiord are graded to a 100-102m sea level and are overlapped to at least 85m by Holocene beaches. These mark no more than pauses in retreat for remnants of kame terraces are present several kilometres downfiord, and upslope bedrock is scoured (see figure 8; Hodgson, 1985). Dates for deglaciation are provided by two dates reported by Hodgson. Surface shells collected from the distal side of the end moraine at the head of Strathcona Fiord dated 6780 ±80 years B.P. (GSC-3765) and a terrestrial sample of Salix sp. twig collected from beneath outwash at the 100m kame deltas (site "100", figure 8) dated 7280 ±90 years B.P. (GSC-3728). These dates, then, give some idea of the timing of deglaciation for the area.

Hodgson (1979), in open file 635 - the surficial materials classification - identified some glacial features and deposits in the area. His data are shown in figure 20. Hodgson identified a number of moraine ridges in the area as well as areas of glacially scoured rock and what he defined as "coarse-grained Holocene advance till" (Hodgson, 1979, pl). These data suggest a simple westwards expansion of the ice sheet; the field investigations reported here hope to expand on this view.

4.2: Original field data.

A. Ice marginal zone.

Direct observations at the ice margin were made difficult by the location of that margin (see figure 44). The present day ice margin is perched at the top of a cliff at approximately 750m asl. whilst a valley running parallel to the margin (reflecting a major fault line), less than 3km to the west, lies 530m lower. A number of outlet glaciers flow almost 300m down this cliff. All the observations reported here were made from a fly camp, of which the location is shown on figure 21.

Site 63 is located at the southern corner of one of these lobes of ice. It flows the furthest of all the outlet glaciers in the area down to the narrow step/plateau, on which the fly camp was situated, approximately 200m above the valley floor. At site 63 the lobe can be seen to have passed through a small lake basin and out the other side of it. Bands of debris can be seen clearly which have been frozen to the glacier base and sheared up into the basal debris layer which is 2.5m thick at the snout and approximately 1-2m thick upslope of the lake. There is no evidence of any meltwater at the base of the ice, although a dry streambed could be traced running south from the lake. The angle of the slope down which the ice is flowing (in

Figure 22. Terminal moraines and debris cover at site 66.

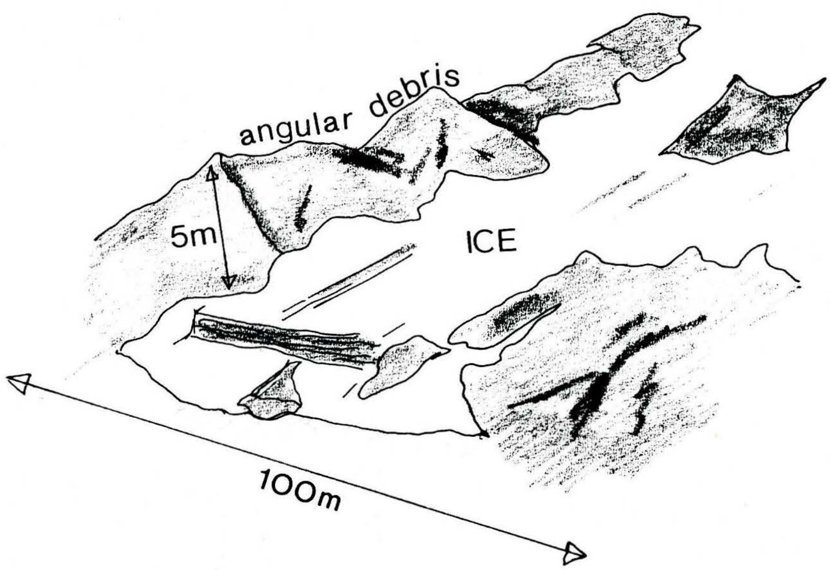


Figure 23. Field sketch of ice lobe at site 66.

excess of 30^{o}) and the fact that the ice surface was not broken up or crevassed indicated that it was probably frozen to its base.

To the southwest of the lake are anumber of moraine-type features which are extensions of those described at site 66. Figure 22 shows these features and figure 23 is a sketch of the supraglacial debris at this location. The zone immediately in front of the ice here can be divided into three zones:

i) the scarp slope of Ordovician strata dipping towards the glacier which is covered by a hummocky, bouldery diamict. This is for the most part unorganised but there is a semblance of semi-continuous ridges within the debris running parallel to the ice front. The southern crest of this slope is a distinct, sharp-crested ridge. The largest boulder on this slope is in excess of 2m long and many boulders are seen over 1m in length. Some areas of the diamict are more clast-supported than others; the hummocks tend to be clast-supported while the depressions show accumulations of finer debris.

ii) A zone of stagnant, debris covered ice with a relief of up to 4m. Seperated by a depression from the active ice front, the stagnation is probably due to the ice front retreating to a higher structural bench. The debris covering the ice is very disorganised - much more so than that seen in zone 1. Discrete mounds can be recognised which are predominantly matrix-supported yet other areas are clast-supported. All sizes and shapes of clast are seen.

iii) The third zone is the very edge of the ice front itself. In one place the basal debris bands are exposed-half a metre can be seen. Elsewhere the ice is covered with gravelly to bouldery debris which at the northern edge of the lobe rises into mounds up to 5m high. These mounds run parallel to flow and appear to be ice-cored (i.e. ice can be seen exposed in the lower part of one whose debris has flowed off).

These three zones are believed to be stages in recession of the ice as the ice front has retreated approximately 20-25m since 1959 when the air photographs of the region were taken. This is in accordance with the view stated by Blake (1981) that glaciers in the region are at, or have retreated slightly from, the maximum position they have reached in several thousand years. The significance of the forms and deposits will be discussed in chapter five.

B. Proximal proglacial zone.

It is very difficult to delimit the divide confidently between proximal proglacial and distal proglacial zones, so any distinctions made herein are to serve clarity in the presentation of results rather than major changes in environment. Having said that, the proximal zone will normally coincide with the glaciated valley landsystem defined by Eyles (1983) while the distal zone is, by and large, his glaciolacustrine and glaciomarine landsystem.

The topography in the Strathcona Fiord area together with, presumably, the nature of the last glacial advance has resulted in a different pattern of glacial sedimentation to the Vendom Fiord field area. Identifiably glacial deposits are more sporadically distributed, preserved by chance in corners or pockets in the substrate, and fluvial systems have not developed into large, braided sandar but, rather, have tended to incise relatively restricted channels. Frequent subaerial exposure of glacially scoured surfaces illustrates that deposition in some areas has been limited or completely absent whilst at other localities continuous sections of Holocene sediment in excess of 10m thick were observed.

The deposits have been subdivided into those thought to be ice contact; stagnant ice, reworked and undifferentiated glacial; and fluvioglacial, although these categories are to facilitate presentation of data rather than provide exhaustive interpretation.

i) Ice-contact features and deposits.

In the Strathcona Fiord area the preservation of these features is sporadic and it is often difficult to relate one feature to another in the absence of any dates or datable horizons.

Marginal moraines are found in a number of sites in the area but the best preserved examples are found on the slopes just below and to the south of the "Walden Plateau". Examples were examined in detail at sites 40, 49, 50, and 74; from a distance at sites 43, 81, and 82; and from the air due south of the eastern end of "Foxtrap Lake". These features can be identified from air photographs and on the ground by their linear shape parallel to valley orientation. The ridges are composed of a diamict containing all clast sizes from silt up to boulders, the larger clasts being matrix-supported. A 40cm deep pit dug into the ridge at site 40 showed the deposit to be cobble-rich and that most of the clasts were rounded to subrounded; no fabric or orientation could be discerned. Figures 24 and 25 show the distribution of deposits at sites 50 and 74 respectively. From sites 43, 49, 81, and 82 a line of these ridges can be traced along the valley wall to the south of the "Walden Plateau". These ridges are thought to be lateral moraines composed of glacial diamict, possibly flow tills from the margins of an ice lobe. They presumably mark a temporary stage in retreat but site 74 shows that they do not record a simple story, as there are two or three generations of

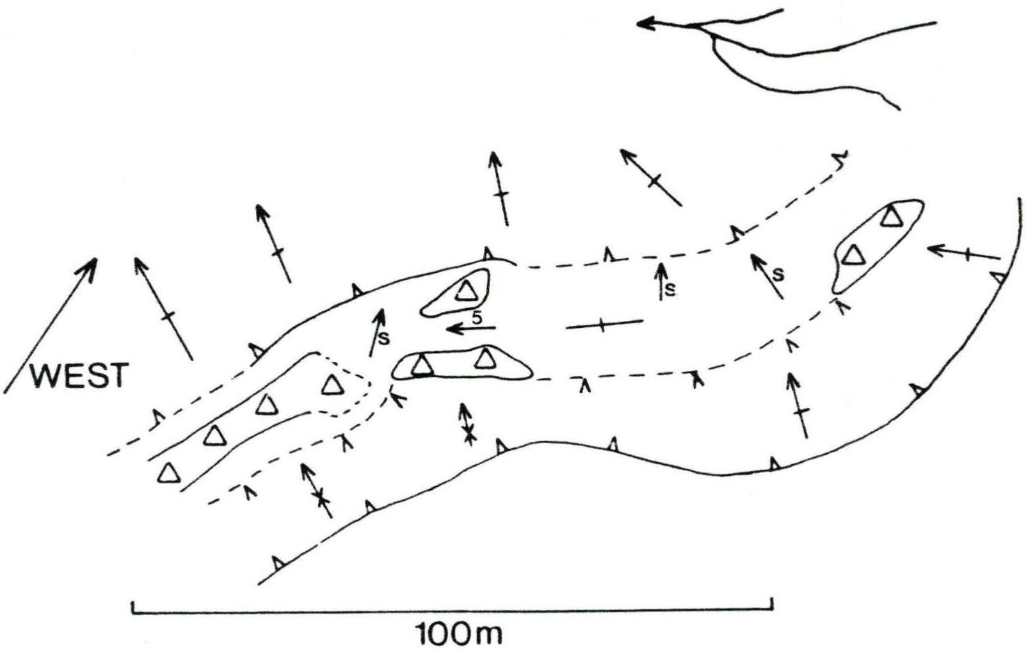


Figure 24. Geomorphological map of moraines at site 50.

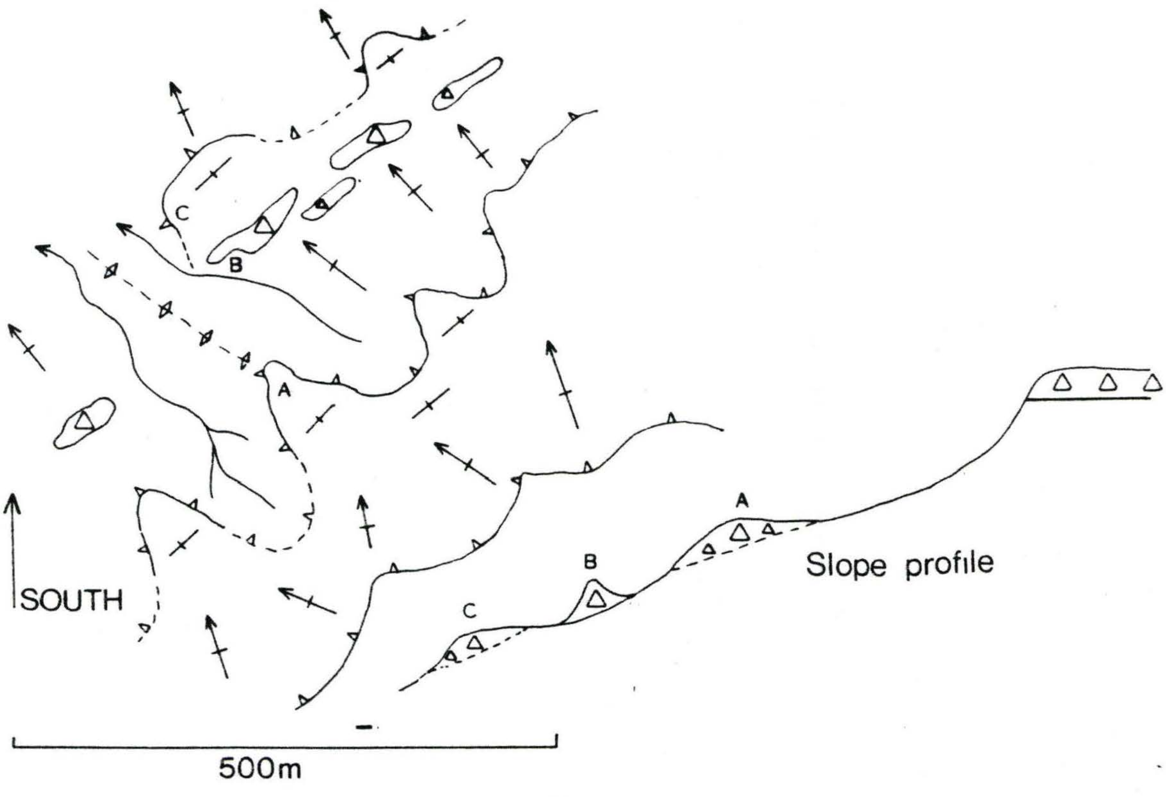


Figure 25. Geomorphological map of moraines at site 74.

moraines separated by 64m of altitude. It is not possible to determine to which generation - the upper or the lower- the features observed from a distance belong.

As the lateral moraines record a pause in retreat, perhaps terminal moraines may be identified which can be related to them. Two sets of terminal moraines have been identified in the Strathcona Fiord area. One relates to a 100m sea level and appears as site "100" in figure 8; the other, at the head of Strathcona Fiord (see figure 20) relates to a sea level of between 70 and 82m (Hodgson, 1985, figure 2). Calculations using the simplified flow law equation:

$$h^2 = (2\tau_o / \rho g)(L-x) \qquad \text{(equation 1)}$$

where: h is the ice thickness at distance (L - x) from the edge of the ice, ρ is density of ice, g is acceleration due to gravity, and τ_o is basal shear stress (Paterson, 1972, 1977; Hutter, 1983; Drewry, 1986). Using a shear stress of 50 kPa which simulates ice sheet conditions with some basal sliding, the upper generation of moraines at site 74 confirm well to the terminal moraine at the head of the fiord and a sea level of 70m. This would give them an age of approximately 6780 ±80 years B.P. (G.S.C. -3765; Hodgson, 1985) which is the approximate age of the terminal moraine. The form of the moraines also tells us something about the nature of the ice sheet at the time of their deposition. Site 50 (figure 24) shows what appears to be a marginal channel cutting through the moraine (its present position making post-depositional dissection unlikely). This implies a similarity to the present ice margin which is also characterised by marginal channels, formed because of the impermeability of "cold" ice (Sugden and John, 1978). The major thing that these features record, however, is that during retreat the ice was confined to the present valleys. This, in turn, suggests that the ice in this area came from a simple westwards expansion of the Central Ellesmere Ice Cap.

The other ice contact landforms seen in the field area are the terminal moraine at the head of the fiord and the kame delta to the north of it. Hodgson (1985) claimed that the terminal moraine records coarse outwash which was then advanced over by floating ice. Hodgson (personal communication, 1986) indicated that exposures at the river section cut through the terminal moraine were better when he examined them in 1983 than when they were examined in 1986. Poor exposures in this section show coarse gravel and pebble beds interbedded with coarse sands which are sometimes laminated or show well-preserved fabric parallel to bedding; matrix-supported pebble beds; coarse sand / fine gravel beds with silty beds included within it; and many more similar combinations. Overlying this are chaotic deposits of matrix-supported, rounded cobbles; highly contorted sands; graded gravel beds; and fine gravel beds showing moderate to good fabric parallel to bedding. Dips and strikes of all of these beds show no consistent orientation either within-bed or between beds.

One explanation may be that the lower deposits are kame delta deposits and the upper sequences show structure imposed by ice push during overriding by a later advance. The whole sequence is then capped with a waterlain diamict deposited by the floating tongue. At first sight, this hypothesis may seem to cause problems for the linking of the lateral moraines described previously to the terminal moraine. A possible scenario might be similar to, but at a far smaller scale than, that proposed by Andrews et al. (1983) for deglaciation of Hudson Bay. In this scenario a glacier (or larger ice body) will advance to the coast and eventually ground below mean sea level. Subglacial incursion of seawater causes floating, surging, and then catastrophic collapse. This would explain why all the kame deposits were not destroyed and why there is not a more widespread cover of waterlain diamicts. The kame delta on the distal side of the moraine and to the north presumably formed quickly during that catastrophic collapse. This is only a tentative hypothesis, however, and further fieldwork would need to be conducted at other sites on the shores of Strathcona Fiord to verify it completely.

Deposits which are undeniably ice-contact are distributed sporadically throughout the field area. The lateral moraines preserve no recognisable structure within them so are identified by their morphology and situation as is the kame delta at the coast within which no sections are exposed. The only distinctive ice-contact deposits are the three exposures of diamict identified at sites 51, 53, and 56 (figures 26, 27, 28, 30a).

The diamict in all three exposures is a massive, matrix-supported deposit. Matrix is silty and clasts vary in size from coarse gravel to boulders 52 x 33cm in exposed plane. Clasts are composed of various lithologies; the largest tend to be subrounded igneous rocks, and the most angular clasts seen are sedimentary cobbles. Some clasts have been frost-shattered *in situ*. With the exception of the very largest boulders, there are no significant differences in rounded/angular or sedimentary/igneous ratios between size ranges.

Within the diamict at site 51 (figures 27, 30a) some structure can be seen and this provides useful clues to its provenance. Just above the debris obscuring the lower part of the diamict is a band of stratified coarse sand and gravel in which, at one point, a large clast squarely sits.

Figure 26. Location map of sites around the delta on the "Muskox River".

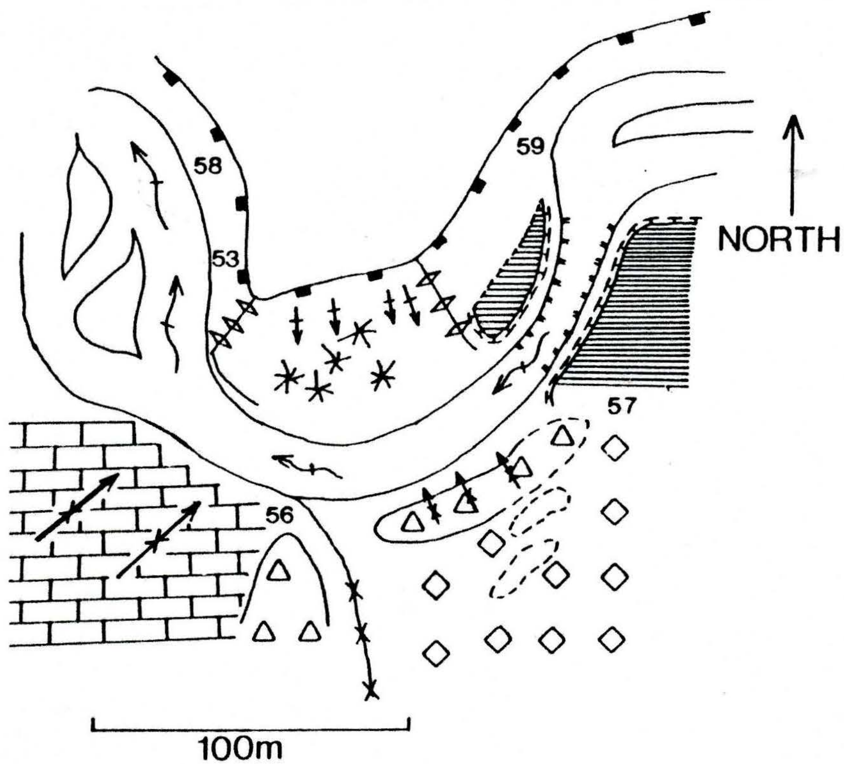


Figure 27. Photograph of the diamict at the base of section 53

This band is well defined both above and below and is laterally traceable for 5 meters until the face is obscured by fallen debris from above. In the middle of the diamict is an enclosed lens 3m long and 0.8m at its maximum height. This consists of wispy-laminated silts and very fine sands, laminated / bedded fine gravels and some coarse gravels, with bedding, that grade upwards into the diamict. At about two-thirds of its length there is a fault within the silts and fine gravels with displacement of approximately 3-5cm.

This enclosure is thought to be a portion of unlithified substrate incorporated into the basal debris layer by freezing as described by Weertmann (1961, 1972; see chapter two). Such unlithified clasts have been described by many authors (Boulton, 1979; Shaw, 1982, 1985; Harris and Bothamley, 1984) and are described as a common attribute of melt-out tills. Formation of the beds during melt-out was rejected as the shape of the lens is not as distorted as one would expect from deposition in an ablating ice/debris mixture, and also because the limits of the finer sediments are so well-defined at the sides and bottom. The fault within the clast must have occurred as loads were increased above the diamict causing differential stresses on the linear-shaped clast.

The presence of the banded gravel bed and the incorporated clast suggest very strongly that this diamict is a stagnant ice melt-out till. The banded gravel resulted from flow within the diamict during ablation and deposition and, yet, the unlithified incorporated clast was not deformed to too great an extent during re-sedimentation.

ii) Reworked and undifferentiated glacial features and deposits

Hodgson (1979) illustrated most of the Strathcona Fiord area as being covered with "coarse-grained Holocene advance till" in a glacially scoured landscape. Much of the land surface is covered with an admixture of debris but it is difficult to tell what is glacial and what is the result of post-glacial weathering. The high ground to the north of "Muskox River" and east of "Kettle Lake" provides an ideal environment for preserving small pockets of glacial sediment, should any exist. On a first reconnaissance it was very difficult to see any pattern to deposition in the area - any sediment on a slope was subject to solifluction and frost heave which, during the Holocene had served to redistribute deposits into topographic hollows. Many bare rock surfaces still exhibited glacial striae preserving a mean flow direction between 65°/245° and 84°/264° (figure 33). Erratic boulders were seen on the surface at many locations - sometimes in what may have been boulder stripes or, more likely, were post-glacial redistributions of glacial deposits.

At site 45, on the top of a hill, a pit was dug into the surficial deposit which revealed that the top 2cm of sediment contained nearly all the clasts (presumably heaved to the surface by frost action) and below this, to a depth of 56cm, at which the permafrost table was reached, was homogeneous fine sand, orange in colour. This was thought to show that much of the surface covering was simply weathering products from the relatively soft Tertiary Eureka Sound formation rocks. Site 35, however, in a similar topographically high situation and with a similar surface appearance told a very different story. Again the top 5cm showed an accumulation of clasts below which a similarly homogeneous orange-brown sediment was encountered. Two small angular clasts which had been shattered in situ were seen at 19cm and 21cm depth. At 42cm depth, just as the sediment was beginning to become frozen, bedrock was reached which showed fresh glacial striae orientated at 84°/264°. The most important piece of evidence was that the deposit was very similar to that seen at site 45, yet the bedrock was very different: that underlying site 35 being Ordovician Allen Bay and Read Bay formation dolomites and limestones.

We see, therefore, a veneer of glacially deposited sediment having been emplaced over striated bedrock over much of the topography and, although many of the clasts have been subject to periglacial downslope movement, the deposit is still relatively fine and matrix-rich. The similarity between the matrix seen in diamicts at Vendom Fiord and this deposit cannot be overlooked - they both consist of a mixture of clays, silts, and sand sizes to produce a loamy texture.

A similar, but more clast-rich, deposit is seen to overlie most of the "Walden Plateau". Observations at site 73 show this covering to be 10-12m thick at its margins and aerial photographs show its hummocky surface relief. Clasts exposed at the surface are very large, up to 2m in length, and all examined were of igneous lithologies. It appears that this deposit is a stagnant ice-laid diamict but it is not clear how old this is. It may be of the same age as the deposits described above which lie below and to the south of it or, more likely given its different appearance and greater thickness, it may relate to an earlier ice advance or an earlier stage of the same advance.

iii) Fluvioglacial features and deposits.

In comparison to the Vendom Fiord field area the fluvioglacial regime is far less prominent in the environment of Strathcona Fiord. Examination of maps and aerial photographs of Vendom Fiord show extensive sandar formed by the "Sverdrup" and "Schei" Rivers. Deposits in front of the "Schei Glacier" also show that in addition to the presently active "Upper Schei River" sandur there are a number of relict, dissected sandar (Ballantyne, 1975; Hodgson, 1973, 1979); sections also show fluvioglacial deposits in that area (Chapter two, section 3.3B). In the Strathcona Fiord field area, by contrast, sandar are conspicuous by their absence. The only fully developed sandar are those of the "Muskox River" east of "Foxtrap Lake" and that 3km northwest of site 66 at site 67. Neither of these are active today in contrast to the continued activity seen on the "Sverdrup and "Schei" Rivers.

Signs can be seen, however, of considerable fluvial activity in the past in the Strathcona Fiord area. The deeply incised gorge of the "Muskox River" east of "Foxtrap Lake" attests to a far higher discharge than at present. Sites 36 and 37 within the gorge show remnants of two stage channel fill preserved in pockets along the river bank. At its deepest point the river gorge is over 100m deep. 8.5m above present river level is an upper terrace surface which is discontinuous but found at the same height at three different locations. The deposit consists of rounded boulders and cobbles in a gravel and coarse sand matrix. 2.5m below the surface of this upper terrace is the surface of a lower terrace which is an incised rock step with a planed or sawtooth surface describing an upper level which varies around 6m above present river level. It is difficult to assign an age to this gorge but it is likely that its beginnings pre-date the last glacial advance (perhaps being formed by meltwater from an earlier ice advance). The gorge was infilled with debris when base level was higher and discharge and sediment load also considerably higher than at present. The lower rock terrace must relate to an earlier stage of incision when base level was higher than today but not at full glacial high - perhaps during a previous interglacial. Final incision has resulted from continued lowering of base level during the Holocene.

Site 79 also shows the results of incision due to lowering base level. Here an alluvial fan, considerably older than two modern alluvial fans which are on either side of it, has been extensively dissected by now disappeared streams. The deposit forming the fan has the same size distribution as diamicts everywhere except that larger clasts have been cryoturbated and the upper surfaces winnowed. No diagnostically fluvial deposits can be seen but sections are poor and the active layer is very thin. Features marking it as an alluvial fan are the correspondence in height amongst dissected segments which falls from the northwest to southeast, its fan-shaped morphology, and its position relative to other relict river channels.

The most impressive fluvial deposits in the area were, unfortunately, only observed from a distance. These are large wedges of colluvial / fluvial sediment in gullies draining southwards from the "Walden Plateau", observed from sites 38, 74, 81, and 82. The deposits consist of very coarse sediment (predominantly cobbles and boulders) roughly bedded in wedges up to 20m thick. Maximum clast size is in the order of metres yet the streams flowing in the gulllies at present have channels only 1-2m in width and only centimetres deep. These wedges must have resulted from very large-scale flows emanating from a large mass of stagnating ice on the plateau. They have provided coarse debris for smaller channels to transport but for the most part the clasts sit at the foot of the slope down which they were transported because of the absence of any subsequent flows competent to transport them further.

C. Distal proglacial zone.

Features and deposits in this area result from glaciolacustrine and glaciomarine sedimentation and a number of excellent sections exposing these sediments were found (figures 28-32, 34-38). The sedimentary logs and photographs in these figures show the deposits well and there is no need to describe the sediments in great detail [1]. The deposits at site 52 correspond to those which overlie section 51 although it is not known how much is missing between the two measured sections. Similarly, section 58 corresponds to the units overlying section 53 with only 2-3m of laminated sands missing between the two. Section 59 corresponds to similar units to those in section 58 but is described from a different exposure and orientation. Both sequences show melt-out till (see chapter four, section 4.2Bi) overlain by laminated silts which gradually coarsen upwards to sand and then are overlain by gravel and cobble beds. Both sections are thought to represent large deltas in a very quiet glaciomarine environment.

The boundary between the basal till and overlying laminated silts is shown in figures 35 and 36. This shows that the till must have been deeply immersed when deltaic sedimentation began. It appears likely that sediment discharges were high and accumulation rates rapid as sediment was brought in by high discharge rivers during stagnation of the last ice (see previous section). Laminations are not thought to represent annual varves but can be ascribed to events of higher sediment influx

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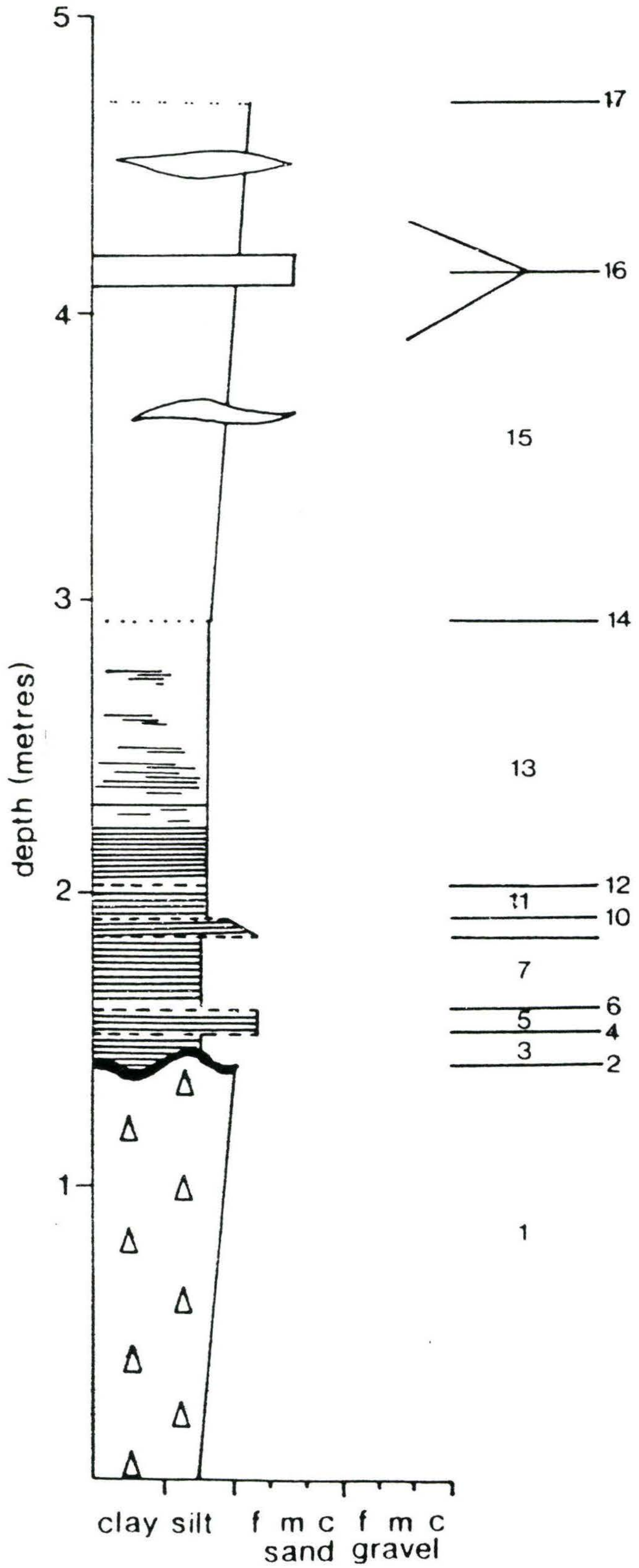
[1] A full description of all the delta beds will appear elsewhere (Miller and McCann, 1987 in preparation).



17. Highest point exposed in dug section - 4.69m. above high water mark. This does not represent a significant boundary.

16. Thick sand layers (up to 80mm.) appear above 3.5m. Representation is schematic as these are sporadic and pinch out laterally.

15. Silt-rich layer coarsening gradually upwards. Bedding/lamination becomes less distinct and thickness of sand beds increases.

13. Silt-rich layer, silt bands back to 8-9mm. in thickness becoming less well defined upwards. Gradation between layers less distinct than underlying beds.

11. Three silt beds 35-40mm. thick defined by sand partings.

9. Sand-rich bed, thickness of sand partings decreases upwards.

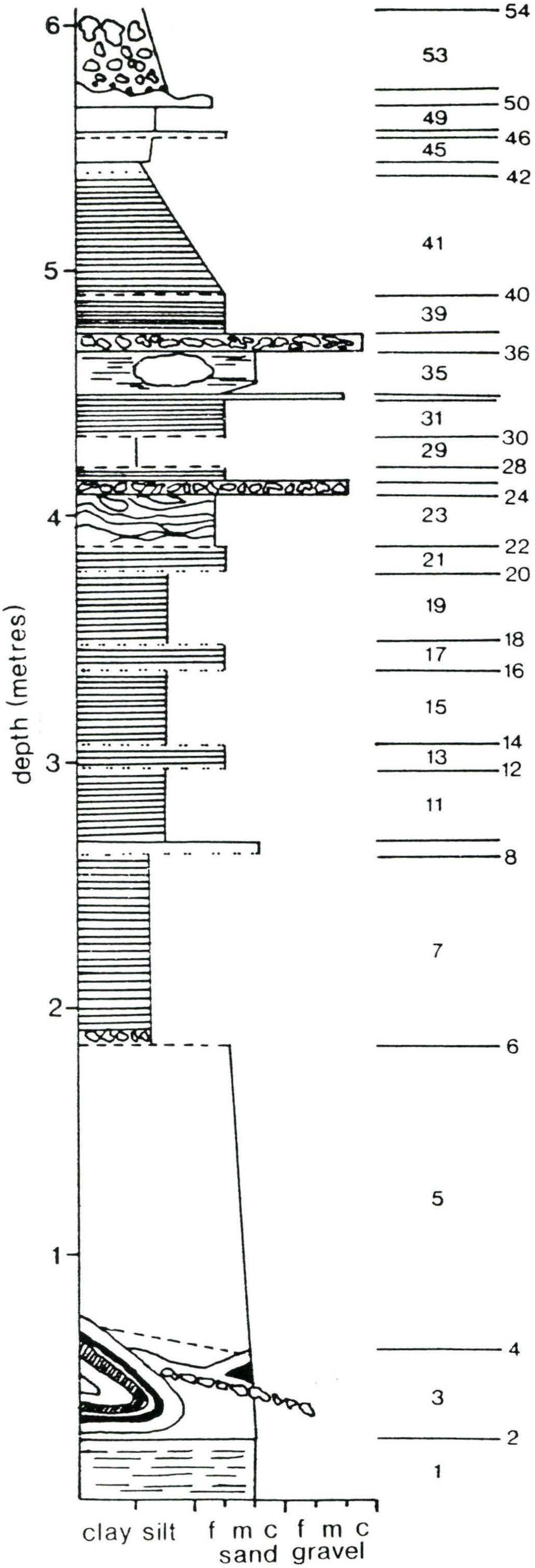
7. Silt-rich unit, sand partings are clearer than those in 3.

5. Fine sand-rich layer - silt bands are much thinner than average for section. Sand divisions are up to 2mm. thick and more numerous.

3. Silt-rich unit with bands averaging 8-9mm. in thickness - very thin fine sand partings.

1. Massive diamict.

Figure 28. Section 51.



53. Matrix-supported gravel and large pebbles. Matrix fines upward, clasts coarsen upward from 4-5mm. at base to 50-60mm. at top.

52. Abrupt undulating boundary. 53. loaded into 51.

51. Fine sand and silt layer, laminations shown by iron staining and fine silt beds.

49. Massive clay/v. fine silt band. Thickness varies laterally.

45. Fine silt/clay bed, varies in thickness laterally. Massive within but coal clasts up to 15 by 6mm. seen on horizontal lower boundary.

43. Clay unit with fine sand laminations.

41. Laminated sand and silt/clay beds with sand beds increasing in thickness downwards.

39. Fine to medium sand unit laminated/interbedded with fine silt/clay.

35. Faintly laminated sand unit which coarsens up for bottom 50mm. and then remains as coarse sand. Contains one large clast (250 by 150mm.) lying flat within sands parallel to laminations.

33. Thin gravel layer, thinner and finer than 37.

31. Same as 39.

29. Clay/silt unit containing three clasts up to 15mm. long.

27. Finely laminated fine to medium sand with fine silt laminae.

25. Thin gravel layer matrix-supported in fine sand.

23. Well-sorted fine sand bed, structure very deformed. Convolutions well defined by fine laminated silt. Pinches out to nothing 1m. to right.

21. Fine/medium sand unit with clay/fine silt laminations. Laminations are mildly deformed. Occasional smallfaults with displacement 10-20mm., some beds pinch out laterally and then re-appear. No large-scale disturbances.

19. Laminated silt bed. Laminae composed of thin, fine sand beds mostly distinguished by colour.

17. Same as 21.

15. Same as 21.

13. Same as 21.

11. Same as 19.

9. Coarse sand matrix containing fine to medium gravel and some large clasts up to 25mm. long. This unit is loaded into the top of 7.

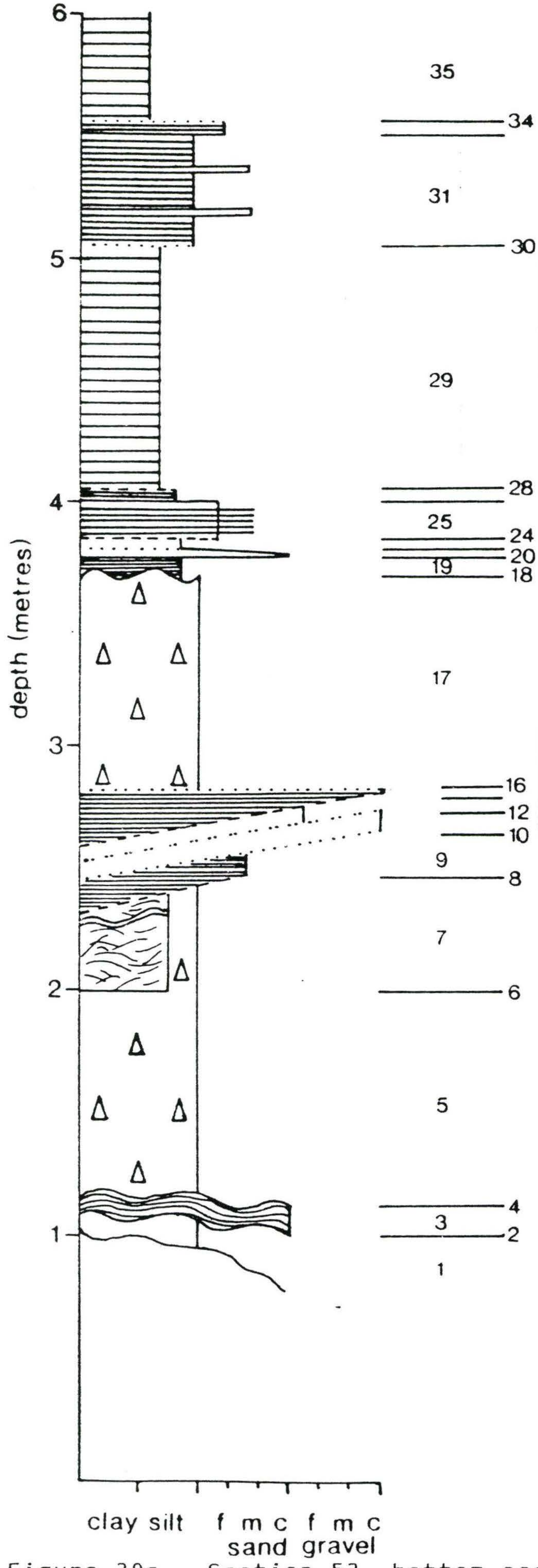
7. Laminated fine silt/clay, mildly deformed. Laminations mostly defined by colour. In bottom 60-100mm.subrounded to sub angular clasts 20-60mm. long can be seen sitting in fine silt matrix on lower boundary - not a separate unit.

5. Very disturbed sand unit. Intense contortions marked by dark silt in sand.

3. Tightly folded silt/clay/sand laminated unit. Abutting the nose is a silty/sandy unit which has medium gravel at its base - all deformed. Below this, deformed laminatedsilts/fine sands. Some silt beds contain gravelly clasts.

1. Well-sorted medium sand unit - some faint lamination perhaps discernible. Permafrost table reached within this unit.

Figure 29. Section 52.



33. Contains clay rip-up clasts.

31. Silt/clay rip-up clasts seen in sand beds of 20-50mm. thickness but not in thinner beds. Depressions can be seen in top of clay bands from where rip-ups came. Beds coarsen up over 20-60mm. cycle.

29. Same as 35.

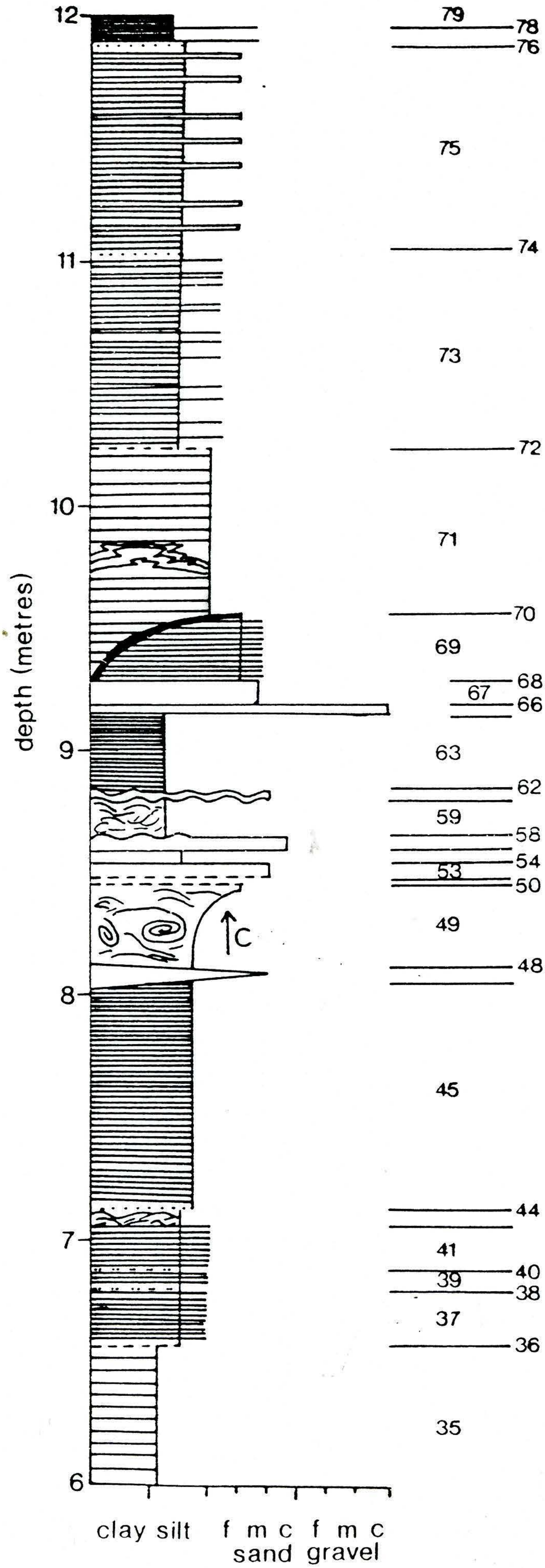
19. Bands of silt and fine sand draped over undulating diamict surface. Some small-scale faulting.

7.-15. ?Clast incorporated into diamict consisting of laminated silts and sands and gravels

1., 5., 17. Massive diamict. Matrix-supported with largest clasts up to 520 by 330mm.

3. Stratified coarse sand and fine gravel.

Figure 30a. Section 53, bottom section.



79. and 77. Very finely laminated silt - contains two thin medium sand beds.

75. Fine sand beds are also laminated with coarse silt - Interbeds of silt are laminated with clay.

73. As for 75. but sands are more minor part of unit - sand beds are thinner and locally discontinuous.

71. Contains large folded and faulted section- anticlincrium shape.

70. Major, arched discontinuity.

67. Well-sorted medium sand bed.

65. Bed of rounded pebbles up to 50mm. sub-rounded.

63. Very finely bedded fine/coarse silt - 0.5-0.6mm. per band. Each band fines upwards.

61. Silt rip-up clasts seen on lower boundary.

59. Loading structures and incorporated clasts of sandy material in silt matrix.

49. Loading and "rolled-up" clasts of silt in sand background. Coarsens up.

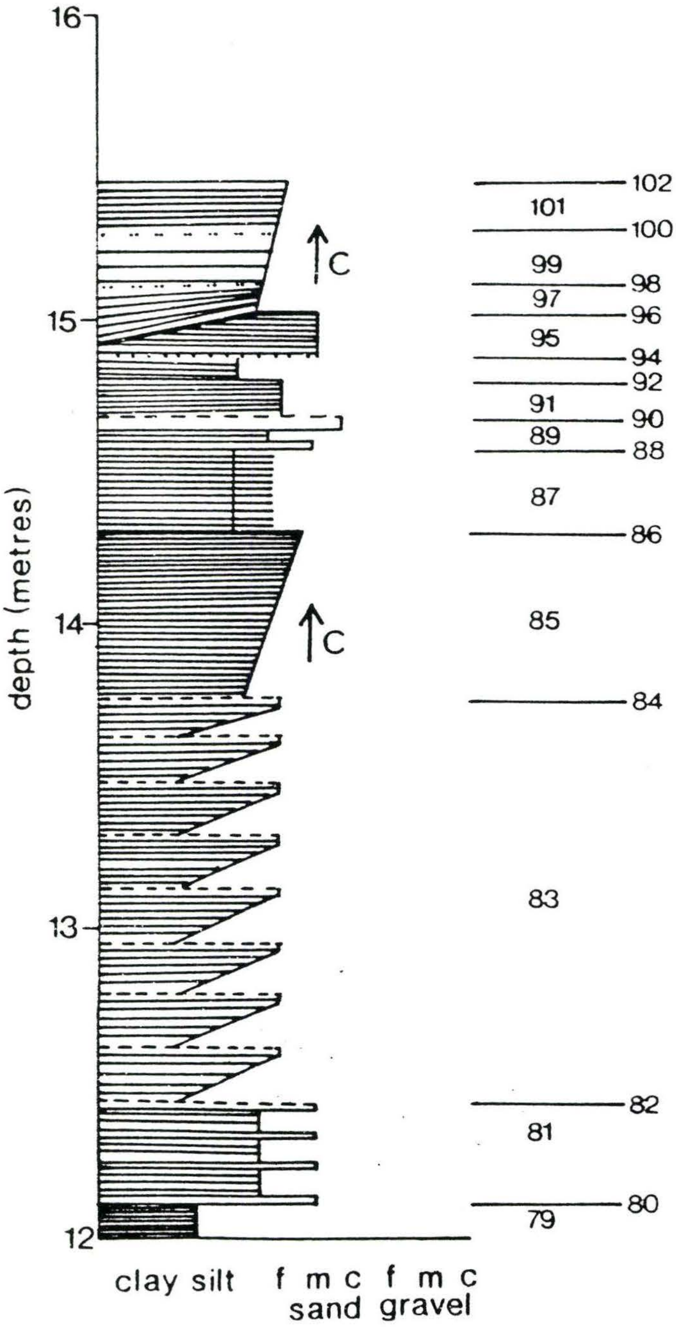
45. Dark, fine silt layer with fine bleached sand laminae usually less than 2mm. thick - may be three or four cycles per 10mm. Laminae often pinch out laterally.

41. and 37. Silty bed with fine sand laminae - up to 10 in each bed.

39. Fossiliferous bed - many shells found including one whole paired set of valves ?Portlandia arctica.

35. Clay bed with fine silt laminae - distinguished by colour cycle changes over 30-50mm.

Figure 30b. Section 53, middle section.



97-101. Clean, bleached well-sorted sand beds which show little silt content. Lamination frequency varies.

96. Discontinuity.

85. Some pinching out of sand beds and loading of silt into sand can be seen.

Figure 30c. Section 53, top section.

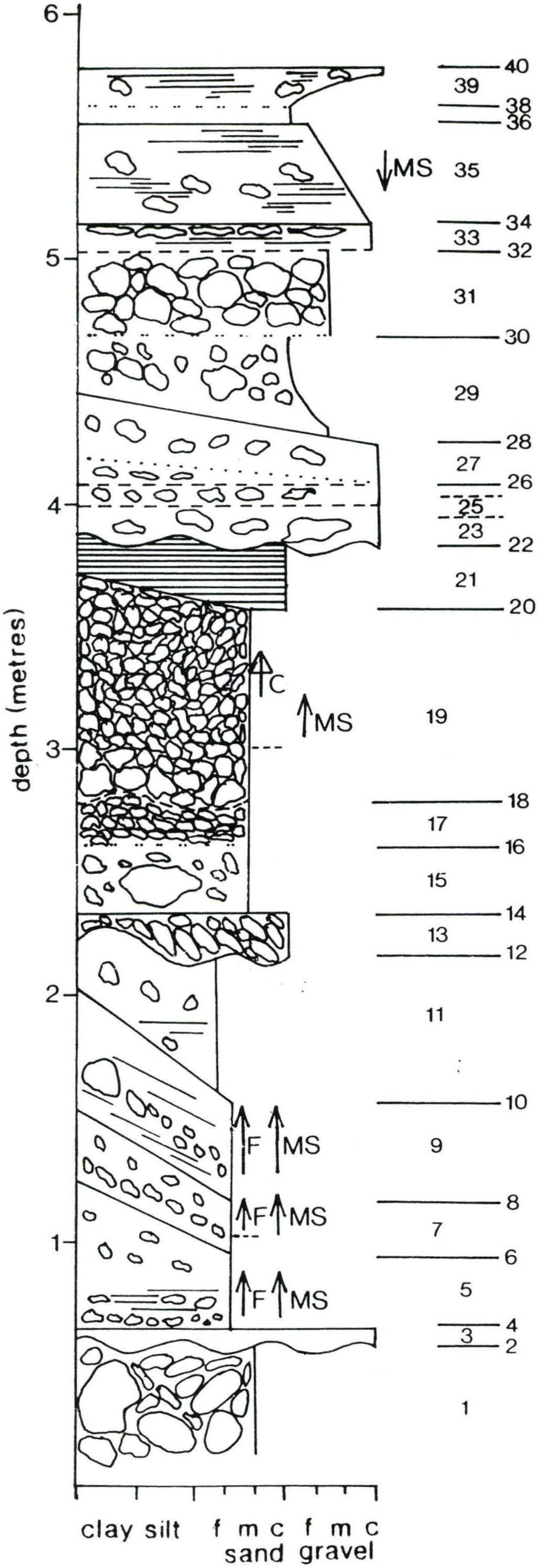


Figure 31. Section 58.

39. Angular, coarse gravel barely matrix-supported. Clasts vary from 45 by 30mm. to coarse sand. Strong fabric parallel to base of bed.

37. As 39. except more matrix-supported and no fabric evident. Locally coarse sand-rich pockets.

35. Just matrix-supported at top, matrix content increases downwards. Clasts vary from rounded to angular. Strong fabric parallel to base seen in gravel-sized clasts.

33. Clast-supported gravel- upper surface defined by platey clasts bedded parallel to bedding in 33.

31. Nearly clast-supported, vague fabric parallel to bedding - seen in finer gravels. Clasts up to 110 by 80mm. rounded.

29. As above but generallly more clast-supported with matrix fining slightly upwards. Base of unit is a "pad" of fine gravel/coarse sand with no clasts.

27. Gravel with clasts up to 60mm. Clast-supported.

23. Matrix-supported coarse gravel (+clasts up to 210 by 160mm.) containing clast-supported but otherwise identical layer.

21. Coarse sand bound with some silt - exhibits very strong fabric.

19. Very open clast-supported gravel/pebbles (modal clast size 40-60mm.). Coarsens upwards above layer of clasts 150-200mm. in bottom 250mm of bed. Towards bottom matrix content increases but bed is still clast-supported.

17. Clast-supported deposit of pebbles/coarse gravel. Moderately good fabric parallel to bedding. Matrix of medium sand and amalgam of all sand sizes is present.

15. Matrix-supported bed - matrix as for 17. although fine sand may be dominant. Clasts mainly 20-30mm. One large clast 230 by 70mm.

13. Just matrix-supported, locally clast-supported, bed of imbricated clasts - vary in size from 60mm. to 2mm. Well-sorted matrix.

11. Matrix-supported bed - clasts 10-80mm. angular to sub-rounded. Platey clasts show weak fabric parallel to bedding - more noticeable in lower half of bed. Matrix is an amalgam of all sand sizes.

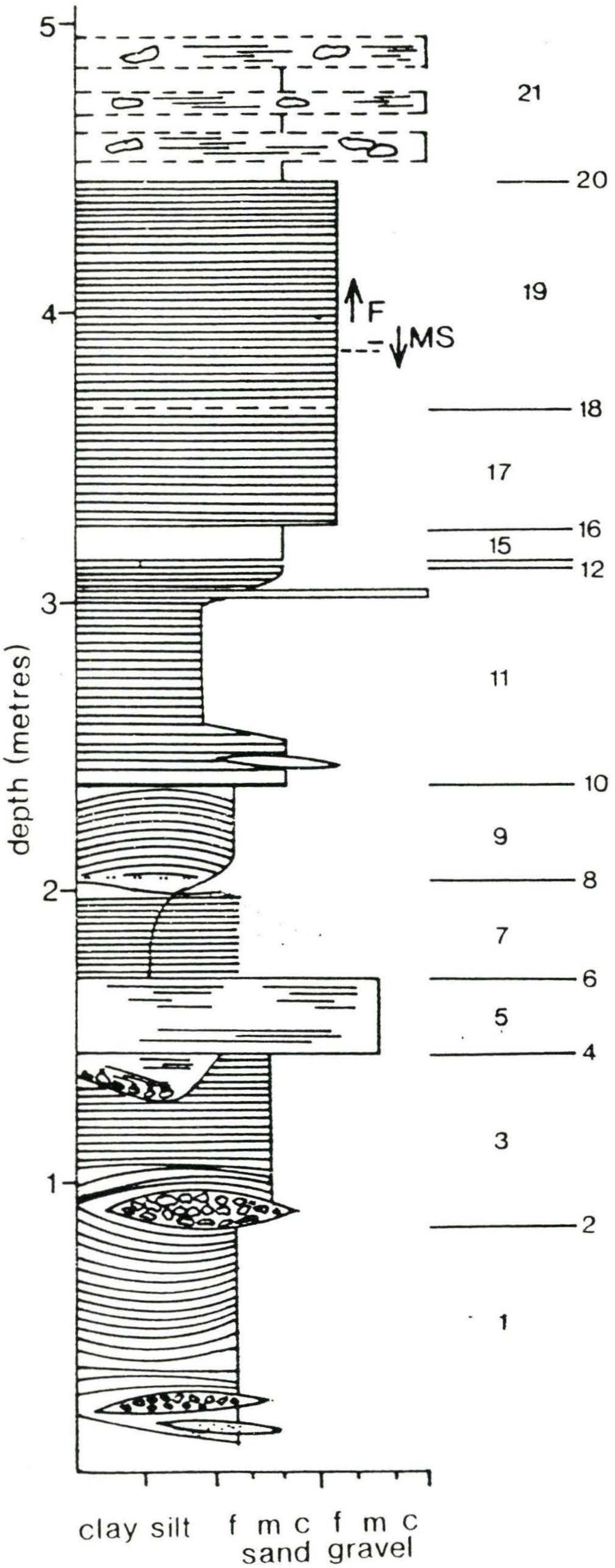
9. Fining-upwards bed. Largest clasts at 2/3 depth - up to 110 by 70mm. Clast size then fines from 30-60mm up to 20mm. Moderate to weak fabric parallel to bedding seen in fine clasts.

7. As for 9. except finer overall and grading is less clear.

5. Again fines upwards but less clear again because of smaller variation in clast size. Lower 100mm. of bed shows good fabric parallel to bedding.

3. Matrix-supported coarse gravel. Pinches in and out in undulating boundary.

1. Coarse unit - noticeably coarser clasts than all above units. Poorly sorted 230 by 160mm. to 60 by 20mm. Matrix is amalgam of all sand sizes. Base not seen



21. Medium/coarse sand unit within which are four gravel-rich. Within coarser beds, clasts are mostly coarse gravel and pebbles up to 60mm. with weak fabric parallel to bedding. Clasts are rare in finer layers.

19. Bedded coarse sand/fine gravel bed. Matrix supported with comonents of medium to fine sand sizes. Also clasts 10-20mm. and up to 60mm. Fines upwards from maximum clast size at 1/3 height. Below this remains well-sorted pebbles. Matrix-support increases downwards.

17. Matrix-supported, bedded gravel, clasts coarser than 19. Good fabric parallel to bedding. Modal clast size 20-30mm. (compared to 10-20mm. for 19.).

15. Featureless fine/medium sand.

13. Thin clay bed.

11. Sand and silt bed with laminae of silt. Coarse angular gravel bed near top and lens of coal fragments near bottomm of unit. Some clasts seen which pierce laminae (therefore dropped?) up to 40mm.

10. Major erosive contact.

9. Fine sand/coarse silt with dark-coloured, wavy coarse silt laminae. No clasts at all.

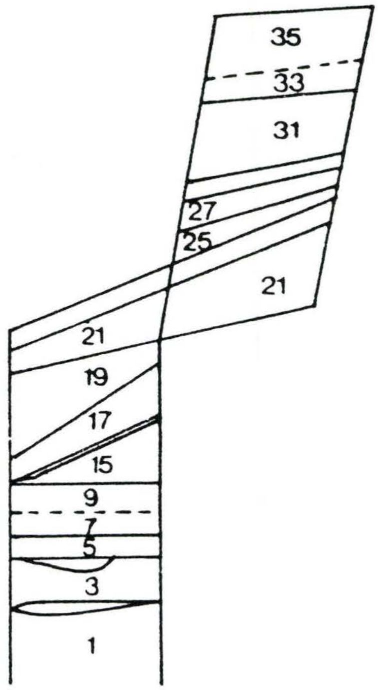
7. Bed gradually fines to clay but laminae are of fine sand - less disturbed and wavy than 9., some coal fragments and other clasts seen rarely - occur in sand layers or at transition from sand to clay.

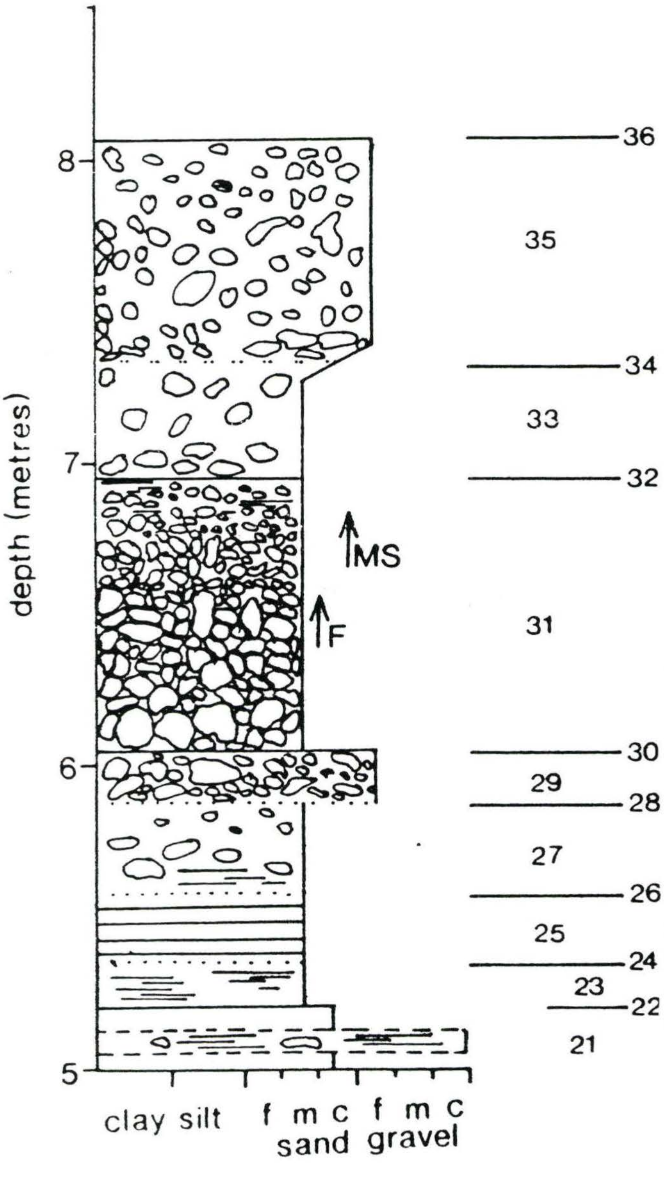
5. Clast-supported medium to fine gravel set in coarse sand matrix. Some fabric parallel to bedding - strongest at base.

3. Fairly well-sorted medium sand which is laminated parallel to bedding and contains two lenses of coarser material. One cuts across bedding (therefore erosive), the other sits conformably within laminae.

1. Predominantly fine sand containing wavy silty laminae and some coarser, lensoid beds of sandy gravel with maximum clast size of 50mm.

Figure 32a. Section 59, bottom half.





36. Arbitrary beginning at top of first clearly identifiable bed.

35. Matrix-supported, poorly sorted gravel and pebble bed although some medium to fine sand is seen. Modal clast size 50-60mm. rarely up to 150mm. Clasts subrounded to subangular.

33. Clasts as for 35. but finer matrix consisting of amalgam of all sand sizes. Bottom 100mm. of bed shows more clasts than upper part of bed.

31. Top 150mm. barely matrix-suported clasts in amalgam of sand sizes. Clasts up to 80 by 100mm but majority are smaller and sub-rounded. Weak fabric parallel to bedding seen here. Below this section unit becomes clast-supported and there is a noticeable increase in open void space. Majority of clasts are sub-rounded medium pebbles 10-20mm. Maximum clast size at base is 230 by 130mm. and cobbles are common.

29. Clast-supported, poorly sorted unit - all sizes from coarse sand to 80mm. are represented.

27. Rounded to sub-angular gravel and clasts up to 70mm. set in well sorted, dry medium sand matrix. Weak fabric parallel to bedding in lower part of unit.

25. Well-sorted medium sand unit with rare rounded finegravel clasts - bedded into four beds - fabric parallel to bottom contact.

23. Similar to 25. but with addition of fine sand component. No clasts at all, weakly bedded.

Figure 32b. Section 59, top half.

interspersed with quiter periods (Hakanson and Jansson, 1983; Ashley, 1975; Church and Gilbert, 1975; Domack, 1983; Smith and Ashley, 1985). The coarsening upwards through the laminated sediments is due to delta progradation (figure 37) and coarser and even overturned beds as seen in section 52 are due to large-scale flows or slumps on the delta front (Miall, 1984; Coleman et al., 1983). These laminated beds all represent prodelta deposition and delta bottomsets whilst the coarser cobbles and gravels are foresets of fluvioglacially transported debris (Leckie and McCann, 1982). The foresets in particular indicate very large-scale flows and turbidity currents depositing 50-70m of sediment in a single event. This period of high sediment input and correspondingly large flows must be contemporaneous with the catastrophic fluvial flows described previously.

Similar deposits were described by Hodgson (1985) from Canon Fiord and Irene Bay so this pattern of deglaciation was areally widespread during the Holocene deglaciation of the region. The delta whose deposits are contained in sections 53, 58, and 59 is truncated by another delta flowing from the south, presumably originating from the river which forms the other major input (along with the "Muskox River") to "Foxtrap Lake". This truncation has caused the preservation of the "clinoform" depositional surface of the foresets (Miall, 1984) (figure 38) which sed

Figure 33. Striated boulder.

Figure 34. View west along "Muskox River" showing sections 53 and 58

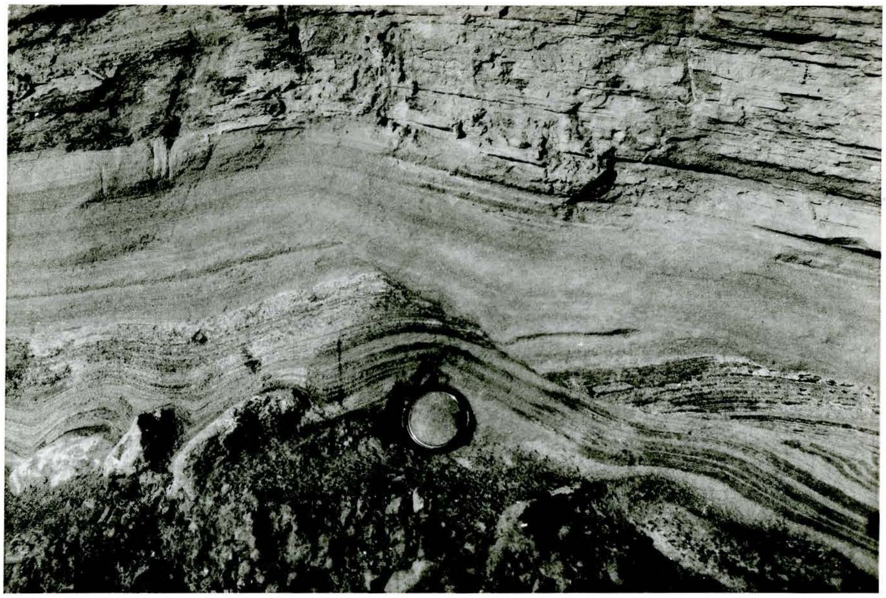


Figure 35. Top surface of diamict at base of section 53 showing draping of laminae and small-scale faulting.

Figure 36. Alternative view of top surface of diamict.

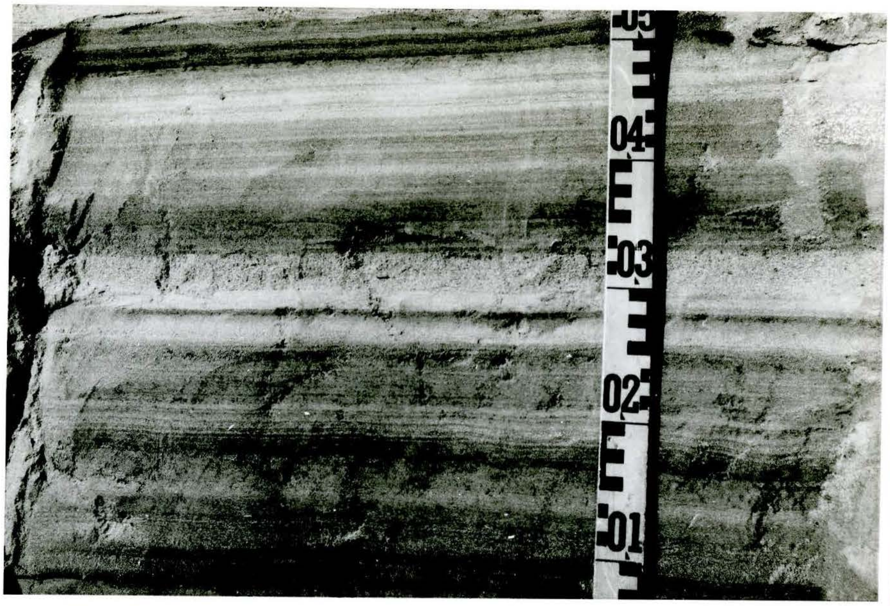


Figure 37. Laminated sand beds in middle part of section 53.

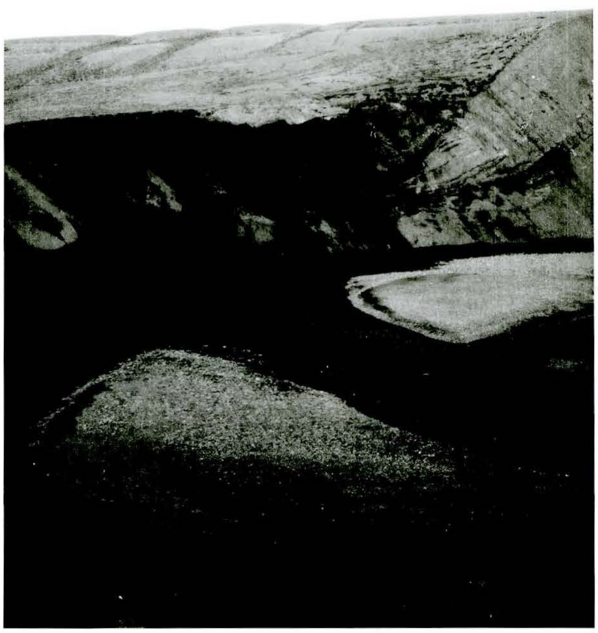


Figure 38. Delta foreset wedge and "clinoform" west of 53.

forms a topographic feature. Overlying the foresets are approximately one metre of fluvial sandur topsets compodelta of similar sediment to the foresets but exhibiting faint horizontal lamination within the matrix of the deposit. This delta was deposited in a stagnant ice environment, shown by the presence of a large kettle lake within the delta. This may have been responsible for some of the flows into the water body but none of the sections examined showed deformation which could be ascribed to deposition in an ice cored basin (Eyles and Miall, 1984). In the high arctic environment it is possible that the deltaic sediments were beginning to be permafrozen before the ice block in the kettle hole had completely melted.

Data from Hodgson (1985) shows that glaciofluvial deltaic lithofacies are common and important in this area. Rhythmites forming under glaciomarine conditions form the most continuous records of sedimentation to be found in the area yet preserve little dateable matter - marine fauna are not well preserved outside the prodelta environment. The water body into which the deltas identified prograded was, most likely, an extension of Strathcona Fiord during times of higher sea level and the fine sediments between the bedrock ridge and the terminal moraine are glaciomarine silts which must be contemporaneous deposits.

4.3: Summary.

Deposits identified in Strathcona Fiord were very different to those in the Vendom Fiord field area, yet some similarities can be seen. In Strathcona Fiord tills were identified as were lateral and terminal moraines which could be related to each other. A thin veneer of fine sediment was found to cover much of the area even though it had been subject to solifluction processes throughout the Holocene.

Very thick fluvioglacial sequences wee also identified which record periods of much higher discharge and sediment load than today. All these environments of deposition provide clues for reconstruction of a late glacial history of the area as well as an insight into glacial sedimentation in a High Arctic environment.

CHAPTER FIVE: MODERN ICE MARGIN

5.1: Introduction and previous work.

The investigation of the modern ice margin was conducted by helicopter with three landings at sites marked on figure 39 as H1, H2, and H3. Observations of the nature of the ice front from Makinson Inlet to Augusta Bay were made from the air. Obviously, any conclusions drawn from such a reconnaissance flight can be only tentative but, nevertheless, a number of interesting and useful observatioins were made.

The only previous literature referring specifically to the margin of the Central Ellesmere Ice Cap is Lorrain et al. (1981) in which the authors described basal debris distribution in Hook Glacier, at the head of Makinson Inlet. Hook Glacier is an outlet glacier flowing westwards from the ice cap which divides into two tongues at its snout. The southern tongue was investigated which ends 1.5km from the fiord. Push moraines up to 30m high record a recent advance and, from the descriptions provided, the sediment making up the push moraine is deltaic rhythmites similar to those described previously (chapter four, section 4.2C). The

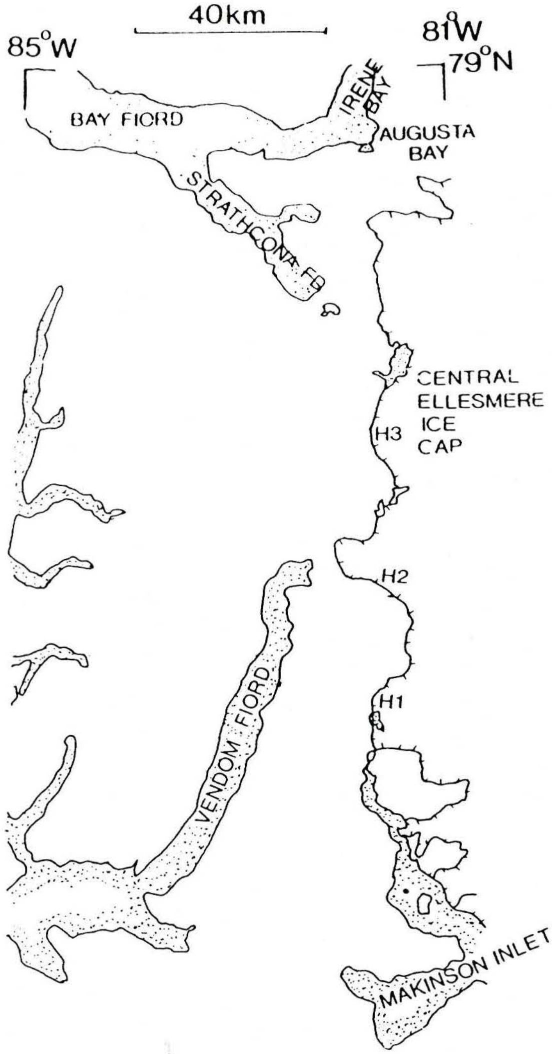


Figure 39. Location of the ice margin sites.

Figure 40. Ice cliffs on the present ice margin.

basal layer of the ice was described as containing many pink orthoclase feldspar chips and layers of greyish limestone chips. Each set was a few metres thick and most of the debris was sand or silt-sized - the grey bands being much finer. Recumbent folds were also observed in the basal layers (Lorrain _et al_., 1981; their figure 17.4) and it is hypothesized that these folds account for increased thickness of the basal layer in the frontal zone.

Blake (1981) also described sections from the proglacial zone of Hook Glacier and a number of other sites, but it is his figures which are of most use as they illustrate areas of the western margin of the Central Ellesmere Ice Cap which were not visited.

5.2: Nature of the present margin.

Thw western margin of the Central Ellesmere Ice Cap consists mostly of steep ice cliffs overriding the substrate with the minimum of effects (figures 40-43). The most common feature observed was a marginal channel running parallel to the ice front as is seen in figure 41 at the bottom left corner of the photograph, and in the same position in figure 42. Nowhere was a margin seen which had the form of a perfect parabola as is the case in Sverdrup Pass (Blake, 1981; his figure 11) or on Baffin Island at the Aktineq Glacier (Lorrain _et al_., 1981, their figure 17.6).

Figure 41. Ice cliffs on the present margin.

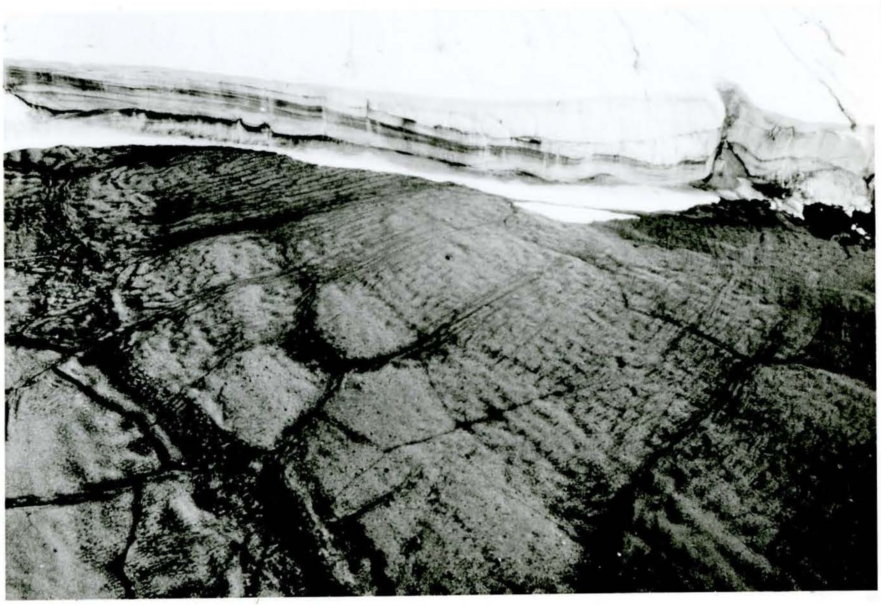


Figure 42. Ice cliffs on the present ice cap margin.

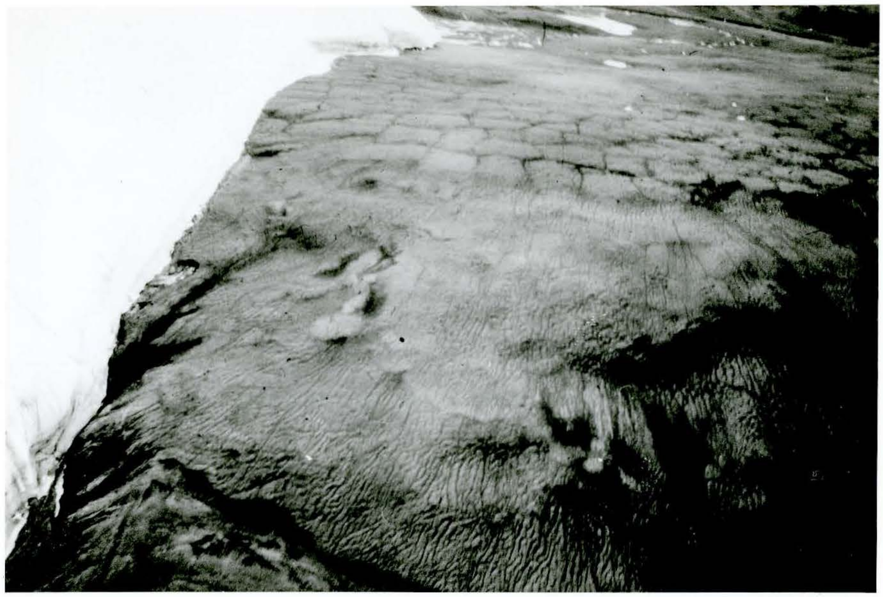


Figure 43. Ice cliffs at the present margin.

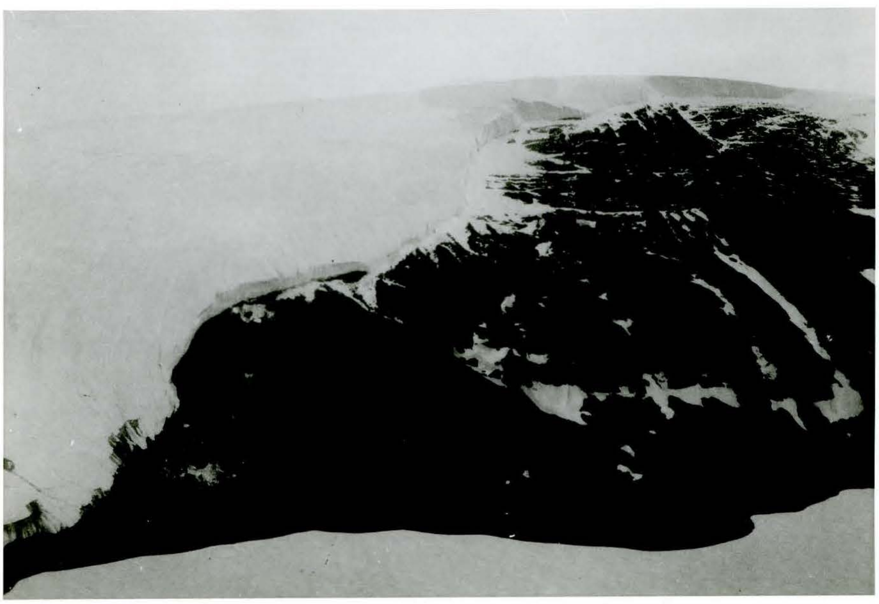


Figure 44. Location of the present margin on a topographic high and lobe flowing downhill from it.

In many places an apron of debris lies at the foot of the cliffs (figure 43) and this is thought to be an important source of sediment for advancing glaciers (see chapter two, section 2.1; Shaw, 1977, 1985).

At other locations, particularly where the ice margin lies on a topographic high, outlet glaciers flow out from the ice cap. Figure 44 shows an example of this which is 5km south of the fly camp, and illustrates the topography in the area which triggers this outward flow. The "Schei Glacier" (7B-8) is the largest of these outlet glaciers; it extends 12km west from the ice cap. Within these glaciers, debris banding can be seen from the air (figure 45), as it can in the "passive margins" (figure 40), so landings were made at three of these outlet glaciers to examine more closely these patterns and proglacial deposits.

5.3: Site H1 - Makinson Inlet

This is the southernmost of the landing sites, at the head of Makinson Inlet north of Hook Glacier. The Glacier terminates approximately 200m from an incised river channel and the level bench between the river and the ice front has the appearance of having been recently deglaciated. This level bench is covered with a deposit consisting of small, rounded cobbles and boulders of sedimentary lithologies, some large angular granitic boulders, and a matrix of angular pink granitic and feldspathic chips (coarse sand / fine gravel size). Very little meltwater drains from under the glacier - only enough to make the ground slightly damp and feed a stream less than 1 metre in width at the northern end of the snout. There is no englacial or supraglacial drainage, the stream is fed by rill-flow from the melting ice margin.

At the edges of the lobe, lateral moraines can be seen which have ablated, and are ablating, out of debris bands in the ice. These moraines consist of mostly local lithologies (Ordovician and Devonian limestones and sandstones) set in a matrix with a loamy texture consisting of a complete range of sand sizes, silts and clays.

It seems likely that the debris being deposited in these moraines has only been transported from the beginning of the lobe where extending flow and increased velocity lead to debris entrainment. Because of the small size and flow pattern of the glacier this debris has remained on the outer edges whilst retreat of the centre of the lobe reveals that it contains a very different debris load. This central load consists of chips of rock and mineral from igneous and metasedimentary rocks which outcrop in the accumulation zone of the ice cap, combined with a much smaller amount of sedimentary debris derived locally.

5.4: Site H2 - Glacier 7B-8.

This site is located at the edge of the 7B-8 lobe, known as the "Schei Glacier" at Vendom Fiord. Figure 46 shows the ice margin at the site with J. Fisher standing next to the snowbank at bottom left for scale. The ice here is approximately 40m high and debris is concentrated in the bottom 11m of the section.

The lowest 1.0-1.2m of the section shows ice with dispersed debris (less than in the overlying layers) covered with washed and melted-out dirt from the ice and perhaps from the marginal channel. Above this, from 1-7m in height is the darkest, dirtiest band of ice in which sediment is concentrated in many fine bands. Between approximately 7 and 11m height there is less debris; the sediment is still concentrated in fine bands but there is more clean ice between the bands. At approximately 11m height there is a very distinct change to almost entirely debris-free ice although there is a band of debris, perhaps 10-20cm thick, at approximately 20m height. For 50cm below this band there may be be diffuse dirt banding or this may be just staining from the bed above. Over its entire height the ice shows foliation and banding which is shown by bubbles and cracks.

A close examination of the debris bands at 1.5-2.0m height shows that the englacial debris consists only of grey silt and coarse sand / fine gravel-sized angular chips of

Figure 45. Glacier lobe with ablating debris bands on surface.

Figure 46. Ice margin at H2 showing debris distribution.

pink granite or orthoclase feldspar. No clasts were seen larger than 3-5mm anywhere in the section examined.

This section is interesting because it shows how debris is transported in this large outlet glacier. The flow direction is parallel to this margin and from the air no crevasses can be seen which would suggest a sizable lateral flow component, so it is unlikely that there is marginal thickening of the basal debris layer. The ice appears to be completely frozen to the substrate as there was no water visible. A marginal channel runs along the ice margin but this was completely dry.

5.5: Site H3 - "Active Glacier"

Between the Vendom Fiord and Strathcona Fiord field areas, approximately 10km north of the "McMaster Lakes", an ice lobe has advanced southwards over a relict sandur. This glacier is a short distance south of that shown in figure 45 which shows active ablation and this site was chosen for a landing because there appeared to be high debris concentrations in the ice.

The ice margin at this site shows intense thrusting and folding of debris bands through the entire thickness of the glacier (figures 47 and 48). The debris apron at the foot of the ice cliffs is 10m high in some places and large amounts of sediment have been deposited. Dirty ice has been thrust at right angles to glacier flow showing a textbook example of this structure (figure 47). Most interesting of all are the debris concentrations shown in figure 48. These are a circular (approximately 3m diameter) and a lensoid-shaped (3m x 1.5m) exposure of very clast-rich ice or frozen sediment. These may be incorporated frozen clasts which have just happened to be exposed in the same vertical plane, or they may be infilled englacial channels. If they are the latter this causes a number of questions to be raised. Under what conditions will englacial drainage occur in cold ice? From where did the sediment originate? And, how common are these features? The implications if these <u>are</u> englacial channnels are far-reaching yet it is difficult to see how they could be . At every other site visited on the western Central Ellesmere Ice Cap margin, drainage has been entirely supraglacial. Whatever questions were raised concerning the thermal conditions at the base of the ice there was never any question that the upper surface of the ice was cold and impermeable - dry snow zones may even be present (Paterson, 1981). These enigmatic features must await further investigation.

This section shows a very different character to that at H2. Far from being a "passive margin", the amount of deformation of ice and resulting incorporation and deposition of debris is impressive. This may be due to unusual flow characteristics within the ice - a meeting of westerly and southerly flowing ice streams perhaps - or to topographic control on ice flow which has caused thrusting and deformation within the ice. Whatever the mechanism, the end result is an active and faulted glacier front where deposits reflect both these characteristics in their distribution, volume, and structure.

Figure 47. "Active Glacier" showing debris apron and thrust faults within the ice.

Figure 48. "Active Glacier" showing posible englacial drainage

CHAPTER SIX: DISCUSSION AND CONCLUSIONS.

6.1: Glacial lithofacies in west-central Ellesmere Island.

Models of sedimentation previously proposed for cold-based or complex thermal regime ice bodies have been outlined in chapter two and the deposits found in the field area have been described in chapters three, four, and five. The different environments in the two field areas have produced different dominant lithofacies. At Vendom Fiord the dominant lithofacies in the proglacial zone is a massive matrix-supported diamict which is interpreted as glacial diamict, probably melt-out till, which has been resedimented in a stagnant ice environment and also later reworked by solifluction and small-scale debris flow. Cryoturbation and overland flow have also modified the deposit. Secondary to this major lithofacies are fluvioglacial and glaciolacustrine deposits. In the sandur deposits it is interesting to note that quiet water facies have been preserved contrary to the belief of Rust and Koster (1984) who stated that these deposits have very low preservation potentials. The glaciolacustrine deposits also preserve debris-flow inputs as well as rhythmites from more regular input variations.

In Strathcona Fiord, by contrast, glacial diamicts are much more restricted in their extent; they are found only in lateral and terminal moraines and a few isolated pockets of melt-out till which have been fortuitously preserved beneath deltaic sequences. Glacially scoured bedrock is seen in many locations and the whole area has been covered with a fine deposit of an amalgam of silt, clay and sand. Fluvial lithofacies are seen but these are thought to be reworkings of glacial deposits by short duration catastrophic flows.

In the distal proglacial zone in both field areas (and elsewhere, see Hodgson, 1985; Blake, 1981) extensive rhythmically bedded prodelta and deltaic lithofacies are preserved. Marine silts are also a common landscape unit. Nowhere is a distinctive deposit such as a sublimation till seen. The thermal regime of the ice bodies responsible for the deposits is not so much reflected in the types of deposits (apart from the absence of lodgement till) as in the relative abundance of different types of deposits, degree of reworking after glacial deposition, and actual amounts of glacial sediment.

All in all the actual amounts of glacial sediment deposited are not large, and are less than one would expect in a temperate or continentally glaciated area. It would appear that despite the thickness of basal debris layers, actual amounts of debris carried in these glaciers are less than in temperate environments (Boulton, 1970, 1971, 1972b; Andrews, 1971, 1972). Of major importance in the sediment cycle is the time during stagnation of regional ice cover when large volumes of meltwater are released. It is these high discharges of both water and sediment that lead to the formation of extensive sandar where there are now only small streams, emplacement of thick wedges of fluvial and colluvial debris at the foot of hillslopes, and formation of thick, extensive deltaic sequences. The almost clast-free deposits found even on topographic highs are thought to be composed of the sediment which is carried dispersed through the ice rather than concentrated in the basal layers. Stagnant ice ablating on plateau areas will form hummocky terrain as seen on "Walden Plateau" and west of the "Hogsback Ridge", and a portion of this deposit will also be transported by meltwater. The subsequent aridity of the climate leaves these meltwater features largely intact as does the rapid fall in base level which encourages fluvial incision rather than reworking of former deposits.

The texture of the matrix in the glacial deposits and reworked glacial deposits in both field areas has been a source of puzzlement for some time. The matrix has a loamy texture as a result of its being an amalgam of clay, silt and sand. This is unusual in glacial sediments in which the matrix is usually well-sorted by meltwater. It is thought that in this environment, the glacial regime largely being meltwater free , the products of glacial abrasion and crushing are preserved along with the original debris rather than being washed away (Whillans, 1978). During deposition all the sediment sizes are deposited together and, because there is a relatively short time during which meltwater is available for reworking, the deposit tends to be moved only short distances <u>en masse</u> rather than sorted efficiently. The subsequent aridty of the climate comnbined with permafrost tends to preserve this texture in all but the topmost layers.

An alternative explanation is that the orthoclase feldspar chips seen in the ice and reported in basal layers by Lorrain <u>et al</u>. (1981) break down to kaolinites after sands have been deposited. This would provide a fine fraction which would not be winnowed below the top layers in the absence of precipitation.

The over-representation of sedimentary lithologies in glacial debris considering the underlying geology of the ice cap accumulation zone is due to the complex thermal regime at the base of the ice sheet. In all places where the margin was observed it was found to be frozen to the substrate, yet debris was seen incorporated in the ice which must have frozen onto the bottom through regelation or ice-debris accretion (Weertmann, 1961, 1972). Thrusting was also seen in the ice which, it has been suggested (Boulton, 1972a), is a result of warm-based ice moving by basal sliding backing up against the margin which is frozen to its bed. This seems likely in the lobes flowing down very steep slopes from the highlands on which the Central Ellesmere Ice Cap is situated. It is hypothesized, then, that much of the debris entrainment takes place here at the margins of the ice cap where the substrate geology changes to predominantly sedimentary lithologies. This is in accord with observations at site H1 and would explain why lateral moraines are so much more common than other moraine deposits.

6.2: Glacial regimes in west-central Ellesmere Island

Chapter two outlined various aspects of the morphology and deposits of cold-based and complex thermal regime ice bodies and it was assumed from the literature that the Central Ellesmere Ice Cap was indeed cold-based (Lorrain *et al*., 1981; Ballantyne, 1975; Hodgson, 1973, 1985; Blake, 1981). Observations from the air and on the ground proved that the margins of the ice cap were frozen to the substrate (figures 40-46) and deposits indicate that sections of the glaciers' bases must be above pressure melting point (see previous section) as does theory, in order for debris to accumulate in basal layers.

Shaw (1977) has suggested that there is an a priori bias in glaciology and glacial geomorphology because many of the studies of large ice bodies were made in the temperate sub-polar and polar regimes of Iceland and Spitsbergen. He suggested that much of the work of Boulton (1970, 1972a) was not widely applicable to polar areas because, despite its high latitude, Spitsbergen enjoyed a relatively moist climate with warm summers. Shaw suggested that the glaciers in Victoria Land, Antarctica were more typical of an arid, polar regime. Most of the world's high latitude ice bodies which are not part of major icesheets (i.e. Greenland and Antarctica) are found in the north of Canada and the Soviet Union, so it is interesting to compare the applicability of Shaw's Antarctic model to a high latitude northern site. Chapters three and four and section 6.1 have shown that the deposits found do not tally with Shaw's model and it was hypothesized that perhaps fundamental differences in climate were the cause. Appendix 2 shows climatic data for Eureka (Maxwell, 1982), Svalbard Airport (Spitsbergen)(Government of Norway, 1986), and for Vanda Station (Wright Valley, Antarctica)(Schwerdtfeger, 1984). It can be seen that Spitsbergen is indeed warmer and wetter than the other two areas but Eureka and Vanda are very similar. Both have the same mean annual daily temperature to within 0.5°c, Eureka has only slightly more precipitation than Vanda, and the Canadian station is actually consistently colder than the Antarctic station in the winter whilst its summer is slightly warmer. The difference may lie in more subtle climatic differences or in other environmental factors such as geological history (back to the Tertiary period), local lithology, weather patterns during glacial periods, etcetera.

6.3: Late glacial history of Strathcona Fiord

The late glacial history of Strathcona Fiord cannot be reconstructed as fully as those presented in the literature for other areas because of the short duration of the field season (four and one-half weeks) and the lack of C14 dates to provide a chronology. Nevertheless, a tentative reconstruction can be presented to be modified later as data permits.

At some stage during the late Quaternary the entire area of the Strathcona Fiord field region and the "Walden Plateau" to the north was covered by ice flowing westwards from the Central Ellesmere Ice Cap. A component may also have been contributed by an ice centre over the Braskeruds Plain as proposed by Hodgson (1985) but there is no evidence in the field area for this. The maximum westwards limit was beyond the 100m kame deltas at site "100" on figure 8, over 30km west of the present ice margin. Ice may have floated in the fiords (Hodgson, 1985)

The kame deltas at site "100" were formed some time before 7590±80 years B.P. (GSC -3823, Hodgson, 1985) and the terminal moraine before 6780±80 years B.P. (GSC -3765, Hodgson, 1985). At the time of the latter date sea level was 70m above present level; the lateral moraines reported in section 4.2Bi are thought to be contemporaneous with this terminal moraine. Then it is thought that some instability, perhaps a small rise in sea level or some threshold being exceeded by the glacier, caused the floating tongue of ice to surge out into the fiord and then collapse, stagnating. This would isolate ice on the "Walden Plateau" to the north of the "Muskox River" valley and "Foxtrap Lake" basin and preserve marginal ice features on the valley sides. The surging of the glacier would deform the terminal moraine at the head of the fiord but not completely rework it and the kame delta to the north of the moraine would begin to form soon after the surge, fed by meltwater from the "Walden Plateau", came through the gap in the bedrock ridge. Moving inland, marine silts were being deposited inside the terminal moraine to the west of "Foxtrap Lake" and a large delta was prograding into the fiord head further inland, with remnants of the ice being incorporated into the deposit which would later melt to form kettles such as that now occupied by "Kettle Lake".

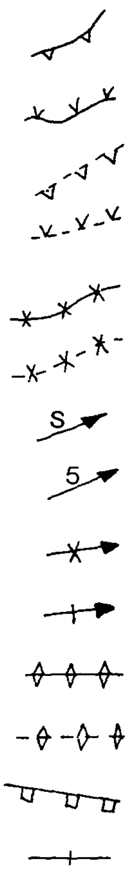
The melting of the stagnant ice would proceed fairly slowly but it is likely that meltwater would accumulate on the surface of the stagnant ice - in ice-covered basins as proposed by Eyles and Miall (1984) - until a threshold was exceeded that allowed sudden release of large quantities of meltwater. In this way, large volumes of sediment would be transported in catastrophic flows whilst elsewhere, away from these flows, other deposits were preserved. Shortly after a major influx of sediment from the northeast into the large delta (figure 34), another delta flowing into the fiord head from the south truncated the older and larger delta (figure 38).

Continued uplift led only to fluvial incision as the geomorphic activity quietened down during the late Holocene. Only periglacial processes were active, modifying surface layers of deposits and causing some downslope mass movement. At some time during the last thousand years (Blake, 1981) the ice margin advanced to just beyond its present limit and deposited a number of terminal moraines. In the past twenty-seven years the ice margin has retreated approximately 20m to its present position, depositing a relatively thin veneer of bouldery, hummocky diamict.

APPENDIX ONE

Basic morphological mapping symbols

(after Gardiner and Dackombe, 1983)



angular convex } break of slope
angular concave

convex } smooth change of slope
concave

sharp } valley axis
rounded

angle of slope
(degrees or slight ($<3^{\circ}$))

convex slope unit

concave slope unit

sharp } ridge crest
rounded

cliffs (40° or more)

level ground

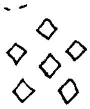
Geomorphological (genetic) mapping symbols

(after Cooke and Doornkamp, 1974)

 rapids

 erosion terrace

 temporary lake

 talus

(after Blachut and Muller, 1966)

 ablation mound (debris covered)

APPENDIX TWO

CLIMATOLOGICAL DATA FOR THREE HIGH LATITUDE STATIONS

EUREKA

80° 00′ N 85° 56′ W

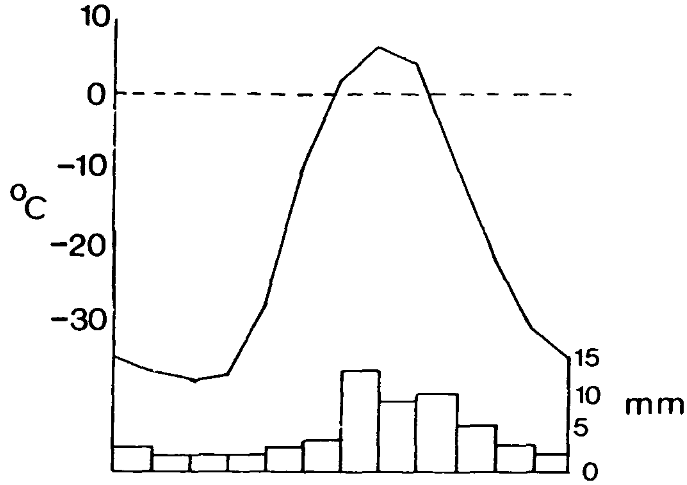
Elevation 10m asl.

Mean annual daily temp.

-19.4°C

Mean annual precipitation

58.4mm



SVALBARD AIRPORT

78° 14′ N 15° 31′ E

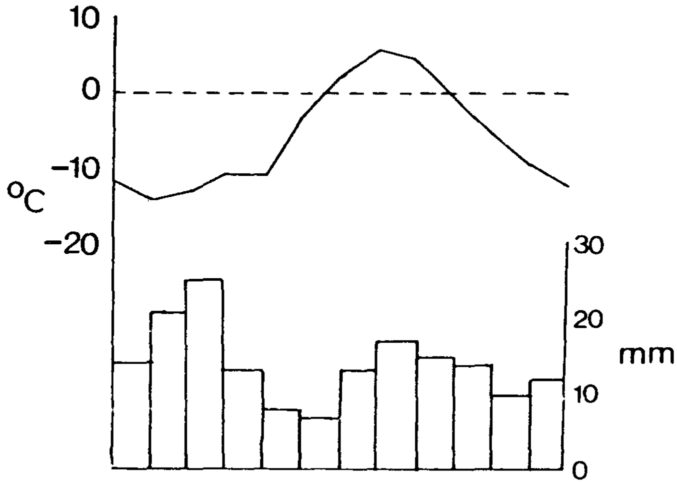
Elevation 15m asl.

Mean annual daily temp

-5.4°C

Mean annual precipitation

165.8mm



VANDA STATION (WRIGHT VALLEY, ANTARCTICA)

77° 30′ S 161° 36′ E

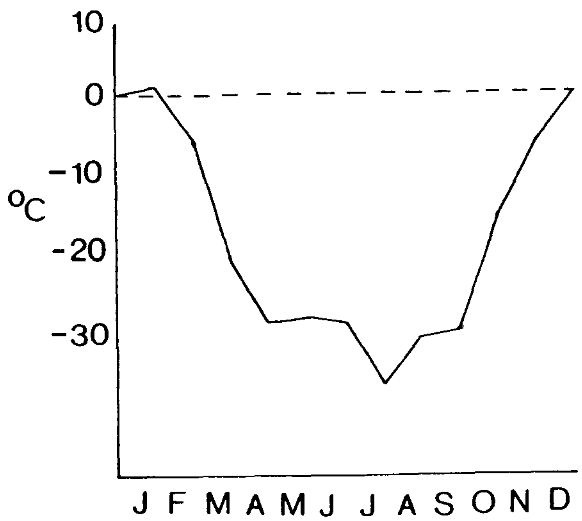
Elevation 75m asl.

Mean annual daily temp

-19.8°C

Mean annual precipitation

<50mm



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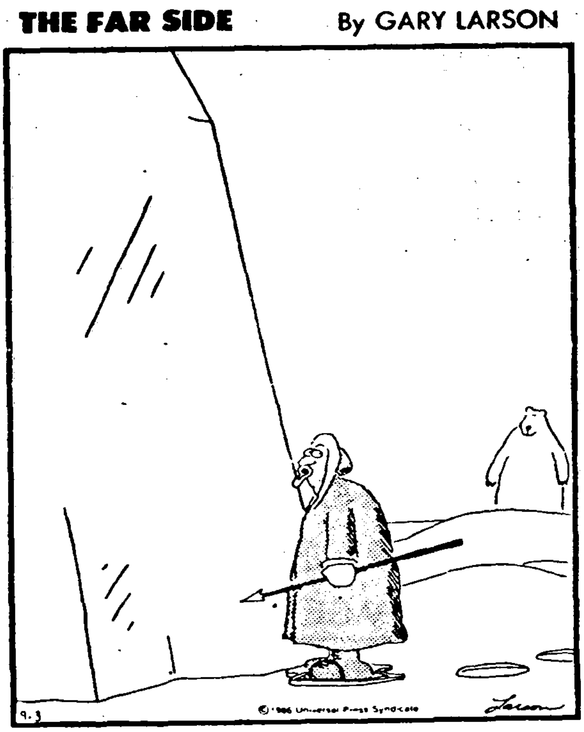

Never put your lip on a glacier.