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Sedimentbeförderungen vom Fjord zur Tiefsee entlang des nordöstlichen kanadischen Kontinentalsaums: Modelle und Daten

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Résumé de l'article

Dans les fjords de l'Alaska les taux de sédimentation sont élevés. Au cours d'une avancée glaciaire, les sédiments d'un bassin-fjord sont transportés vers le front du glacier pour former un banc qui contribue à réduire le taux de vêlage. Ainsi, au cours de cycles glaciaires successifs les sédiments sont d'abord emmagasinés dans les fjords puis retirés. Dans les fjords de l'est de l'île de Baffin, les taux de sédimentation sont, et étaient, beaucoup moins élevés (<1000 kg m2 ka) et leur comblement peut s'étendre sur plusieurs cycles glaciaires. Cette hypothèse concorde avec l'observation de taux de sédimentation relativement faibles sur le plateau adjacent (50 à 500 kg/m2 ka) et la plaine de haute mer (< = 50 kg/m2 ka). L'avancée de langues émissaires à travers ces fjords arctiques peut s'expliquer par la croissance in situ d'une plate-forme de glace flottante, ancrée à l'embouchure du fjord. La limite de la Glaciation de Foxe dans les fjords McBeth et ltirbilung peut être tracée grâce aux deltas marins soulevés (50-85 m anm) pour lesquels on a obtenu des datations au 14C sur les coquillages et les os de baleine in situ de > 54 ka. Les plages soulevées holocènes sont plus basses et datent de <10 ka. Ces données ainsi que l'absence de till dans les sections marines soulevées le long de la côte externe permet difficilement de concevoir une glace ancrée jusqu'au plateau pendant le maximum glaciaire de 18 ka. Les carottes recueillies dans les fjords Tingin, ltirbilung et McBeth varient entre 4 et 11 m de longueur, mais ne représentent l'échantillonnage que d'une partie du comblement des bassins.

FIORD TO DEEP SEA SEDIMENT TRANSFERS ALONG THE NORTH-EASTERN CANADIAN CONTINENTAL MARGIN: MODELS AND DATA*

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ABSTRACT   In Alaskan fiords, sedimentation rates are high; during a glacial advance fiord-basin sediments are transported to the ice front to form a shoal which reduces the calving rate. Thus, during successive glacial cycles, sediment is initially stored and then removed from fiord basins. In the fiords of eastern Baffin Island sedimentation rates are, and were, much lower (<1000 Kg/m² ka), and fiord-basin fills may span several glacial cycles. This hypothesis is in keeping with the relatively low sedimentation rates on the adjacent shelf (50 to 500 kg/m² ka) and deepsea plain (< = 50 kg/m² ka). The advance of outlet glaciers through these arctic fiords may be explained by the in situ growth of a floating ice-shelf, grounded at the mouth of the fiord. The extent of late Foxe Glaciation in McBeth and Itirbilung fiords can be delimited by raised marine deltas (50-85 m asl) with ¹⁴C dates on *in situ* shells and whalebone of >54 ka. Holocene raised beaches are lower and date <10 ka. These data, plus the absence of tills in raised marine sections along the outer coast, make it difficult to extend grounded ice onto the shelf during the 18 ka global maximum. Piston cores from Tingin, Itirbilung and McBeth fiords vary between 4 and 11 m in length, but only sample a proportion of the total basin-fills.

RÉSUMÉ   *Transfert de sédiments des fjords à la haute mer, le long de la marge continentale du nord-est du Canada: modèle et données.* Dans les fjords de l'Alaska les taux de sédimentation sont élevés. Au cours d'une avancée glaciaire, les sédiments d'un bassin-fjord sont transportés vers le front du glacier pour former un banc qui contribue à réduire le taux de vêlage. Ainsi, au cours de cycles glaciaires successifs les sédiments sont d'abord emmagasinés dans les fjords puis retirés. Dans les fjords de l'est de l'île de Baffin, les taux de sédimentation sont, et étaient, beaucoup moins élevés (<1000 kg/m² ka) *et leur comblement peut s'étendre sur plusieurs cycles glaciaires. Cette hypothèse concorde avec l'observation de taux de sédimentation relativement faibles sur le plateau adjacent (50 à 500 kg/m² ka) et la plaine de haute mer (< = 50 kg/m² ka). L'avancée de langues émissaires à travers ces fjords arctiques peut s'expliquer par la croissance in situ d'une plate-forme de glace flottante, ancrée à l'embouchure du fjord. La limite de la Glaciation de Foxe dans les fjords McBeth et Itirbilung peut être tracée grâce aux deltas marins soulevés (50-85 m anm) pour lesquels on a obtenu des datations au ¹⁴C sur les coquillages et les os de baleine in situ de > 54 ka. Les plages soulevées holocènes sont plus basses et datent de <10 ka. Ces données ainsi que l'absence de till dans les sections marines soulevées le long de la côte externe permet difficilement de concevoir une glace ancrée jusqu'au plateau pendant le maximum glaciaire de 18 ka. Les carottes recueillies dans les fjords Tingin, Itirbilung et McBeth varient entre 4 et 11 m de longueur, mais ne représentent l'échantillonnage que d'une partie du comblement des bassins.

ZUSAMMENFASSUNG   *Sedimentbeförderungen vom Fjord zur Tiefsee entlang des nordöstlichen kanadischen Kontinentalsaums: Modelle und Daten.* In den Fjords von Alaska sind die Sedimentbildungsraten hoch; während eines glazialen Vorstosses werden die Sedimente des Fjord-Beckens zur Eisfront transportiert und bilden dort eine Schwelle, welche die Kalbungsrate verringert. So wird Sediment während der sukzessiv aufeinanderfolgenden glazialen Zyklen zunächst angesammelt und dann von den Fjordbecken weggetragen. In den Fjords im Osten der Baffin-Insel sind und waren die Sedimentbildungsraten viel niedriger (<1000 kg/m² ka), und die Füllung der Fjordbecken kann sich über mehrere glaziale Zyklen erstrecken. Diese Hypothese stimmt mit den relativ niedrigen Sedimentbildungsraten auf dem angrenzenden Plateau (50 bis 500 kg/m² ka) und der Tiefseeebene (< = 50 kg/m² ka) überein. Das Vorrücken von Gletscherausläufern durch diese arktischen Fjords kann durch das Anwachsen in situ eines schwimmenden Eisplateaus, welches an der Mündung des Fjords verankert ist, erklärt werden. Die Ausdehnung der späten Foxe-Vereisung in den Fjords von McBeth und Itirbilung kann durch gehobene marine Deltas (50-85 m asl) mit ¹⁴C-Daten von Muscheln und Walfischbein in situ von > 54 ka bestimmt werden. Die im Holozän angehobenen Küstenstreifen sind niedriger und stammen von < 10 ka. Diese Daten und das Fehlen von Grundmoränen in den angehobenen marinen Gebieten entlang der äusseren Küste lassen wohl kaum die Annahme zu, dass verankertes Eis während des globalen Maximums von 18 ka sich bis zum Plateau ausdehnte. Die in den Fjords von Tingin, Itirbilung und McBeth gewonnenen Bohrkerne sind zwischen 4 und 11 m lang, belegen jedoch nur einen Teil der ganzen Beckenauffüllungen.

INTRODUCTION

Fiords are deep, narrow arms of the sea that characterize glaciated landscapes (*e.g.* Syvitski *et al.*, 1987). They are commonly cited as prime examples of the efficiency of glacial erosion (Sugden and John, 1976). During the development of ice sheets over northern Canada (*cf.* Ives *et al.*, 1975) there would come a time when outlet glaciers would attempt to move seaward through the fiords. However, recent research on tidewater glaciers in Alaska (*e.g.* Meier and Post, 1987; Brown *et al.*, 1982) indicates that the calving rate increases as a function of water depth. For the glacier to advance the calving rate has to be less than the forward velocity of the ice. In Alaska, observations indicate that during glacier advances fiord-basin sediments are transported to the ice front to form a moraine shoal (Fig. 1) which has the effect of reducing the calving rate (Post, 1975). Thus during a glacial cycle, sediment is removed from the fiord; coarse sediments are stored along the maximum ice-front limit as moraines, but the finer-grained sediments are deposited on the shelf and in adjacent deep-sea basins.

The Alaskan model (Fig. 1) suggests that there should be distinct pulses in the offshore rate of sediment accumulation associated with fiord-sediment evacuation. Furthermore, it is implicit that fiord basin-fills only span a single glacial cycle. In this scenario (Fig. 1), the ice can continue to erode bedrock once the sediment has been transported to the ice front. Such a process might explain the suggested high rates of glacial erosion for the eastern sectors of the Laurentide Ice Sheet which were derived from a sediment balance analysis of the northwest North Atlantic (Laine, 1980; Bell and Laine, 1985).

During interglacials and interstadials, that is during periods when glaciation is normally reduced, fiord basins act as major sediment traps in the terrestrial to deep-sea sediment pathway. These nearshore basins hinder the "normal" relationship of sediment transport from rivers onto the continental shelves and to the deep-sea (*e.g.* Hay and Southam, 1977). However, a question of interest in this paper is the rate and nature of sedimentation in Baffin Bay. The Ocean Drilling Site (ODP) 645 was located at the foot of the slope east of Clyde Fiord (Fig. 2). The section ODP drilled consisted of sedimentary cycles with an estimated 7.7 ka periodicity; in the "glacial" section these were dominated by detrital carbonates. The carbonate sources are most probably associated with major ice masses which moved toward Baffin Bay through the Arctic Channels (Kravitz, 1982; Klassen, 1985; Andrews *et al.*, 1985b) and suggests that the delivery of sediment from Canadian Shield sources was restricted. Why?

The question addressed in this paper is the applicability of the Alaskan scheme (Fig. 1) to an Arctic setting, where even today fast-ice exists for 9-10 months of the year, where the mean annual temperature on land is *ca.* − 12° C, and the water temperatures are frequently below 0° C (*cf.* Gilbert, 1982, 1983; Syvitski, 1986; Jacobs *et al.* 1985). Following a comparison between the Eastern Canadian Arctic and Alaska, field and piston core data are presented from fiords in east-central Baffin Island (Tingin, Itirbilung, McBeth, and Inugsuin) to: 1) delimit the late Foxe glacial maximum; 2) define Holocene sedimentation rates; and 3) attempt to delimit the age of the fiord-basin fills.
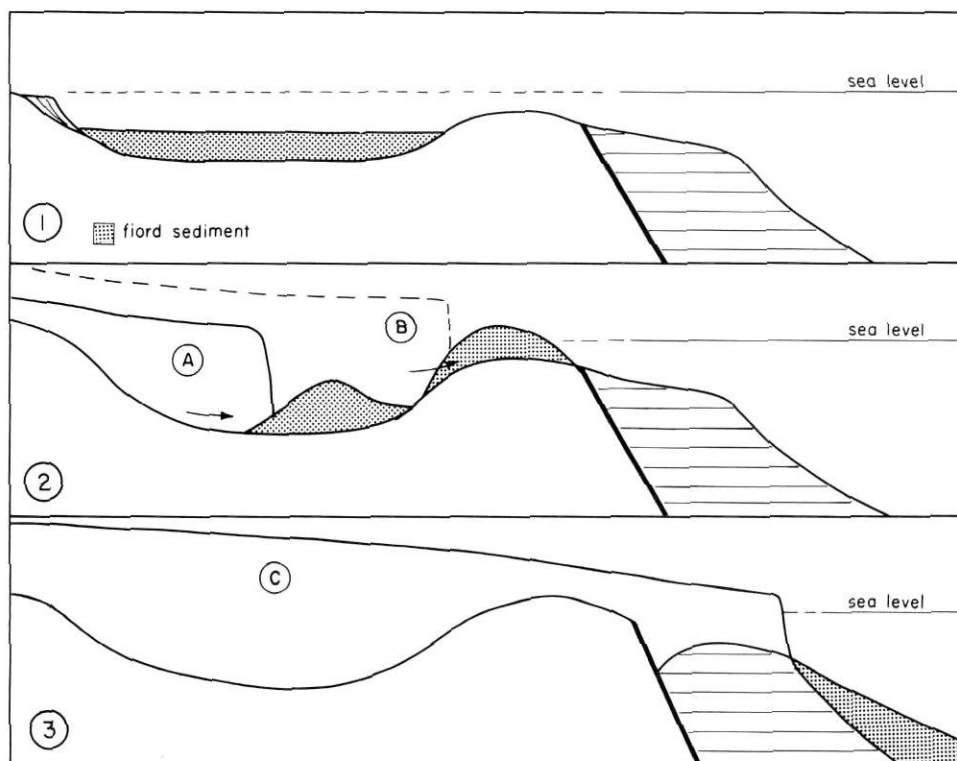

FIGURE 1. Schematic illustration of the "Alaskan" model for an advance of a glacier down a fiord from a period of no glaciers to an advance of the ice during intervals, A, B, and C. To reduce the calving rate the glacier can only advance if a moraine shoal is constructed at the glacier front. The horizontaly shaded area on the shelf is meant to represent offshore Tertiary sediments.

Modèle «alaskien» schématisé de l'avancée d'un glacier vers l'amont d'un fjord à partir d'une période sans glacier jusqu'à celle d'une avancée glaciaire au cours des intervalles A, B et C. Pour qu'il y ait diminution du taux de vêlage, le glacier ne progressera que s'il y a formation d'un banc de moraine au front du glacier. La zone en grisé qui apparaît sur le plateau représente les sédiments littoraux tertiaires.

AN ARCTIC FIORD GLACIER MODEL

As part of the Canadian Sedimentology of Arctic Fiords Experiment (S.A.F.E.) project (Syvitski and Schafer, 1985) a series of fiords along the eastern margin of Baffin Island was surveyed for bathymetry, sediment thickness, and acoustic facies; piston cores were obtained from one or more basins in ten fiords (Syvitski and Blakeney, 1983; Syvitski, 1984; Syvitski and Praeg, 1987). A critical part of this project was the determination of sedimentation accumulation based on accelerator mass spectrometry (AMS) radiocarbon dates (e.g. Jennings, 1986; Andrews, 1987a & 1987b; Andrews et al., 1989). The problems of evaluating such dates are discussed in Fillon et al. (1981), Mudie and Guilbault (1982), Andrews et al. (1985a, 1986) and Short et al. (in press).

The gradient in the Holocene sediment flux, from the outer fiord basins to the deep-sea, is illustrated as Figure 3. The rate of sediment accumulation in Baffin Bay and the northern Labrador Sea is currently under debate (De Vernal et al., 1987; Srivastava et al., 1987; cf. Aksu, 1985; Aksu and Mudie, 1984; Thouveny, 1988). The seafloor flux in outer fiord basins is close to 10,000 kg/m^{-2} (Fig. 3) and these data are replicated in at least six other cases (this paper). The bulk of the sediment is silt or clay (Fig. 3), but the percentage of sand increases onto the shelf and into the deep-sea. The seafloor sediment flux shows a nonlinear decrease from the fiords, onto the shelf, and to the deep-sea (Fig. 3). Much of the shelf is <200 m in depth and is scoured by icebergs; sediment thickness is often less than 10 m (e.g. Praeg et al., 1986; MacLean, 1985). However, sediment does accumulate on the floors of the few deep troughs that extend across the SE Baffin Shelf (HU78-37, Fig. 3). However, at these sites the available information (Andrews, 1987a; Praeg et al., 1986) indicates ca. 5 m or less of sediment has accumulated in the last 10 ka, and in the Resolution Basin, NE of the entrance of Hudson Strait, some 3 to 5 m of sediment has accumulated in the last 25 ka. Unless these sites are being largely by-passed, these rates of accumulation should set upper limits on the probable amount of sediment that has accumulated in the adjacent deep-sea basins. Indeed this appears to be true because cores in the NE Labrador Sea (Hess et al., 1987) have radiocarbon dates which indicate < 1 m of sediment has accumulated in the last 6-8 ka and less than 3 m in the last 25 ka (in Andrews et al., 1989).
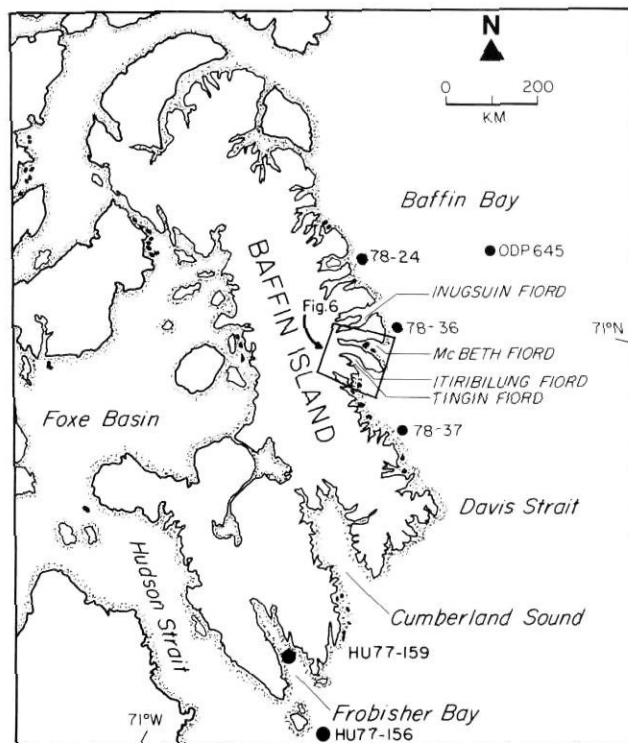

FIGURE 2. Location map of the area of study, showing the location of the fiords, piston cores from the shelf, and the location of the ODP 645 drill site. Figure 4 is located between McBeth and Tingin fiords.

Carte de localisation de la région à l'étude montrant l'emplacement des fjords, des sites de carottage et du site de forage ODP 645. Le territoire de la figure 4 est situé entre les fjords McBeth et Tingin.
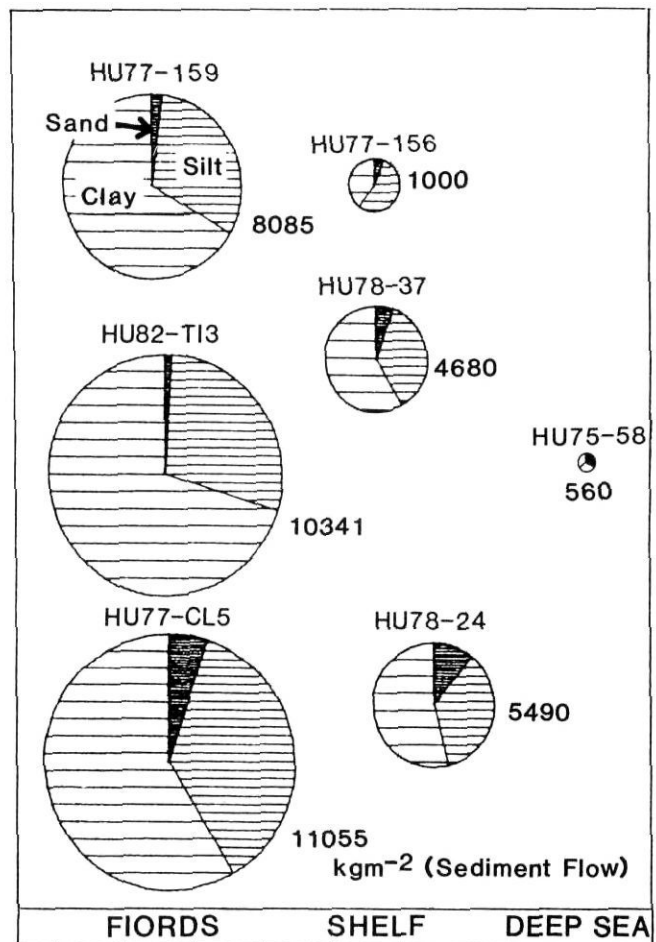

FIGURE 3. Sediment accumulation at sites along the eastern continental margin of Baffin Island showing the proportion of sand/silt/clay in the fiord, shelf, and deep-sea environments. The locations of cores HU82-TI3 and HU78-37 are shown on Figure 2, the other sites are located in Andrews (1987a).

Diagrammes de l'accumulation de sédiments dans les sites le long de la marge continentale de l'île de Baffin montrant les proportions de sable, de silt et d'argile dans les fjords, sur le plateau et en haute mer. La localisation des sites HU82-TI3 et HU78-37 apparaît à la figure 2; les autres sites sont traités dans Andrews (1987a).

In the fiord basins the piston cores only penetrate <10% of the total sediment thicknesses (Gilbert, 1985), but they extend into, and sometimes through, the last deglacial pulse of sedimentation associated with massive meltwater discharges from the retreating Laurentide Ice Sheet (cf. Dyke, 1974). Thus the rate of sediment accumulation was an order of magnitude higher between 9 and 6 ka than at present (Andrews et al., 1985a, 1986) and was slower between ca. 9 and 12 ka. During the late Foxe Glaciation, many fiord glaciers in northern Baffin Island did not extend to the outer coast, but terminated within the fiords (Loken, 1966; Miller 1976, 1985; Klassen, 1985; Locke, 1987). In these cases, the basins would trap sediment inshore of the shelf, and sediment delivery to the deep-sea would be restricted but without drilling through the fiord basin fill, it is difficult to know the age of the basal sediment. The record of glaciation on the outer east coast of Baffin Island suggests that outlet glaciers extended a short distance onto the shelf during late marine isotope stage 5 or stage 4 (Mode et al., 1983; Miller, 1985; Feyling-Hanssen, 1985).

DISCUSSION

One hypothesis would suggest that the sediment fill was removed during the early Foxe Glaciation; this would allow consideration of the Alaskan model (Fig. 1); conversely, a different model for the advance of a fiord glacier across deep basins is feasible (Figs. 4 and 5). In this model the growth of an ice sheet over the high uplands of Baffin Island (cf. Ives et al., 1975) leads to the eventual advance of glaciers into the fiords. As the glaciers advance into deeper water the calving rate increases and may be limiting. However, the climate of Baffin Island is arctic and the fiords are frozen solid for 9 months of the year. Thus iceberg drift within the fiords and across the shelf is restricted when compared with Alaskan

cases. In fiords with sills, deep-draught icebergs may be confined within the fiords. Even if this were not the case I suggest that the combination of icebergs and sea-ice causes the development of a floating mass of ice that constitutes a floating ice-shelf (Fig. 5). In a typical eastern Baffin Island fiord (Dowdeswell and Andrews, 1985), with a calving rate of 0.2 km/ a the fiord would be jammed with icebergs in a matter of a few hundred years. If there was abrupt climatic deterioration, the land-fast sea ice might be the seed for an ice shelf that would buttress the ice front and reduce calving, leading to an advance of the grounding line.

If the ice shelf grounded on a sill, or on the shallow shelf, prior to the grounding line advancing out of the deep fiord basins (Fig. 5) then the question is: what happens to the sediment fill and confined water? One possibility is that the back-pressure allows the grounding line to advance across the deep basin and reach the outer coast and scour the basins. One negative argument for extensive sediment transportation is that the amount of re-transported sediment that has been recognized is limited. Loken (1966) and other workers have noted shelly tills along the outer fiord walls, frequently with ages >30 ka, but the total volume of such sediment is small. In addition, there have been few morainal banks mapped on the Baffin Island shelf that might represent accumulation of recycled fiord sediment. An alternative hypothesis is that the water in the fiord and in the sediment cannot escape during the glacier advance, hence the glacier has little or no contact with its bed. Evidence for this is seen in the low shear stresses that have been calculated for former fiord glaciers on eastern Baffin Island using lateral moraine elevations (cf. Mode, 1980; Locke, 1980). Although normal shear stresses at the base of glaciers are around 100 kPa several reconstructions on fiord glaciers give values of 50 kPa or much
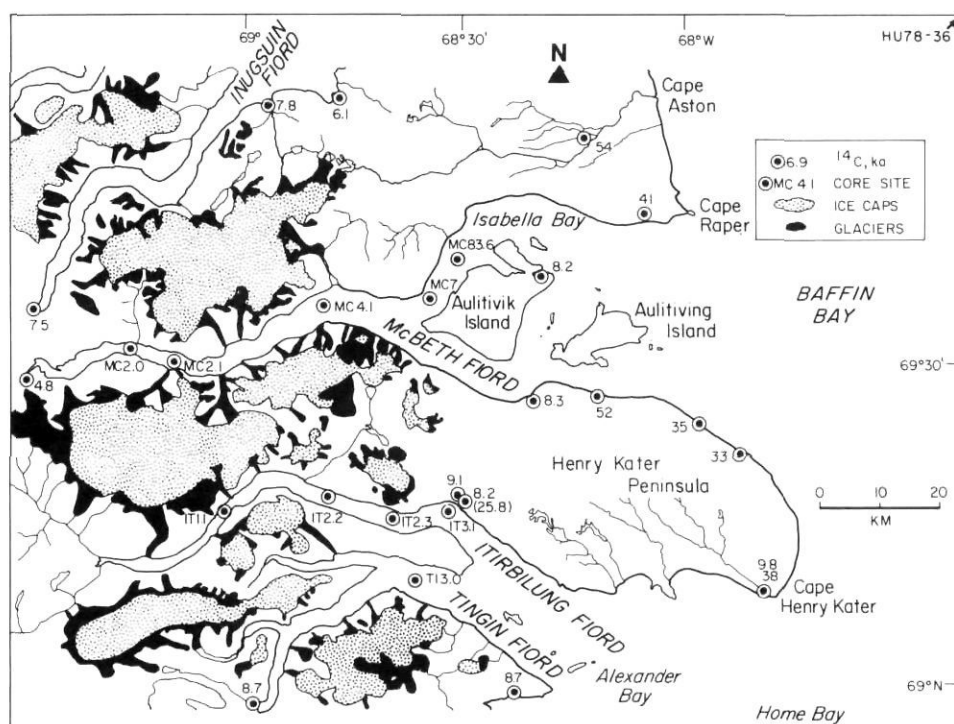

FIGURE 4. Location map of the fiord area of east-central Baffin Island showing the extent of local ice caps and glaciers, the location of ^{14}C dates piston cores (see Table II).

Carte de localisation de la région des fjords du centre-est de l'île de Baffin qui montre l'étendue des calottes et des glaciers locaux, ainsi que l'emplacement des sites de carottage où l'on a obtenu les dates au ^{14}C (voir le tableau II).

lower. Such low basal shear stresses suggest that the glaciers were flowing across deformable beds or were "floating" on a thin layer of water (*cf.* Boulton and Jones, 1979; Blankenship *et al.*, 1986; Fisher *et al.*, 1985).

An order of magnitude estimate of the total sediment volume stored in Baffin Island fiords, based on a survey of the fiords (Dowdeswell and Andrews, 1985; Gilbert, 1985), suggests that 2500 km³ ± 600 of sediment is trapped (Table I). This represents a surface lowering of *ca.* 3.5 m. The critical question is: does this storage represent the last 10 ka (*i.e.* the Alaskan model) or does it represent a longer-term storage of sediment and the restriction of sediment inputs to the deep-sea?

SEDIMENTATION AND GLACIAL EVENTS IN EAST-CENTRAL BAFFIN ISLAND FIORDS

The area of interest are four fiords that lie on the northern edge of Home Bay (Fig. 5) namely (from south to north): Tingin, Itirbilung, McBeth, and Inugsuin. These fiords were chosen because of the variations in the bedrock geology (Fig. 6), particularly the variations in the outcrop of Archean granite gneisses and the younger metasediments of the Foxe Fold Belt (Tippett, 1984). Table II lists data on the area and surface bedrock of the drainage basins that supply water and sediment to the three main fiords (see Figs. 6 and 7). Other information on these fiords can be found in the one or more of the S.A.F.E. Data reports (Syvitksi and Blakeney, 1983; Syvitski, 1984; Syvitski and Praeg, 1987). Bathymetric data from the NOAA marine data base indicate that a deep trough angles at N45E degrees from Home Bay. No deep trough connects McBeth Fiord and Baffin Bay.

GLACIAL GEOLOGY

The surficial landforms, stratigraphy, and chronology of terrestrial glacial and raised marine sediments for the area are outlined in Harrison (1966), Loken (1966, 1967), King (1969), Andrews *et al.* (1970), Andrews (1980). Miller (1979) also visited the area and made key collections of samples for ¹⁴C dating. The stratigraphy and chronology of the wave-cut cliffs, just to the north, along the Clyde Foreland, is also relevant (*cf.* Miller, 1985; Feyling-Hanssen, 1985; Mode, 1985; Mode *et al.*, 1983).

Figure 4 shows the location of available dates, and Figure 8 is a simplified shoreline diagram of the age/elevation re-

TABLE I

Order of magnitude estimate of sediment volume stored in the fiords of eastern Baffin Island and amount of glacial erosion

Number of fiords: *ca.* 250 (Dowdeswell and Andrews, 1985)

Average surface area of these fiords: 100 km² (Dowdeswell and Andrews, 1985)

Maximum sediment thickness: <200 m (Gilbert, 1985)
Average sediment thickness: *ca.* 100 m (Gilbert, 1985)

Total sediment volume: 2,500 km³

Area of Baffin Island: 450,000 km²

Thickness of sediment if spread evenly across the island: < = 5.5 m

Bedrock lowering (assumed density of 2700 kg m⁻³): < = 3.4 m

Rate of bedrock lowering if this associated with late Foxe = 3400 mm/*ca.*15 ka = *ca.* 220 mm/ka

FIGURE 5. An "Arctic" model for fjord glacier re-advance in which rapid iceberg calving, seasonal fast-ice, and the jamming of icebergs within the sill results in an ice shelf and the possible preservation of sediments in the fiord basin.

Modèle «arctique» de récurrence d'un glacier de fjord dans lequel le vêlage rapide, la formation saisonnière de banquises et les embâcles par les icebergs à l'intérieur du verrou-barre aboutit à la formation d'une plate-forme de glace et probablement au stockage des sédiments dans le fjord.
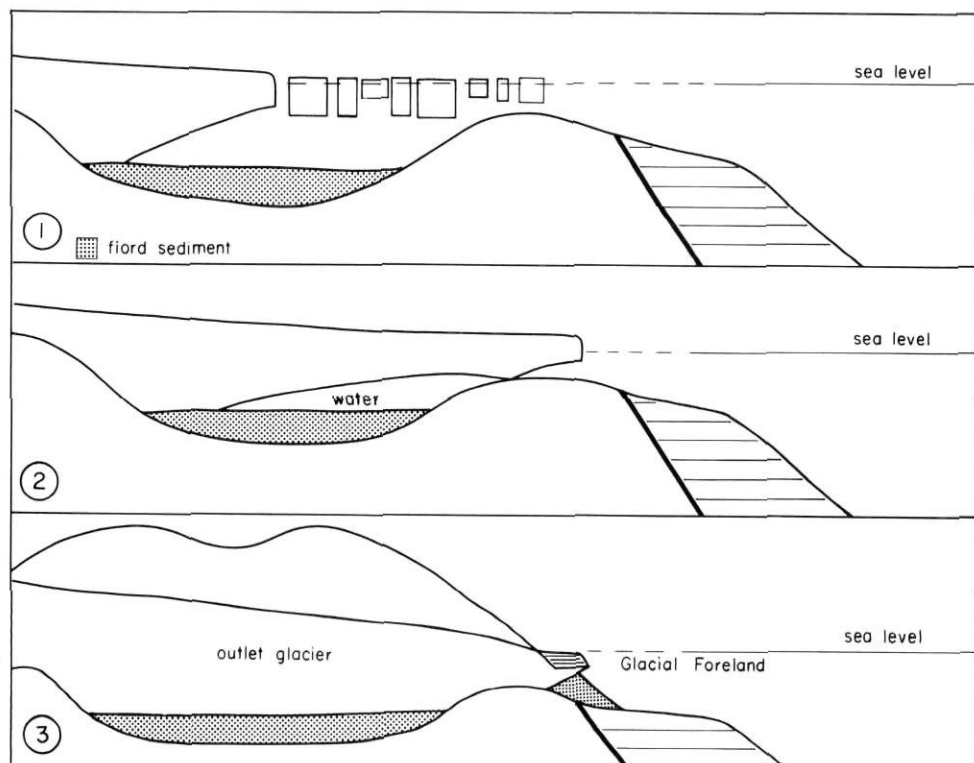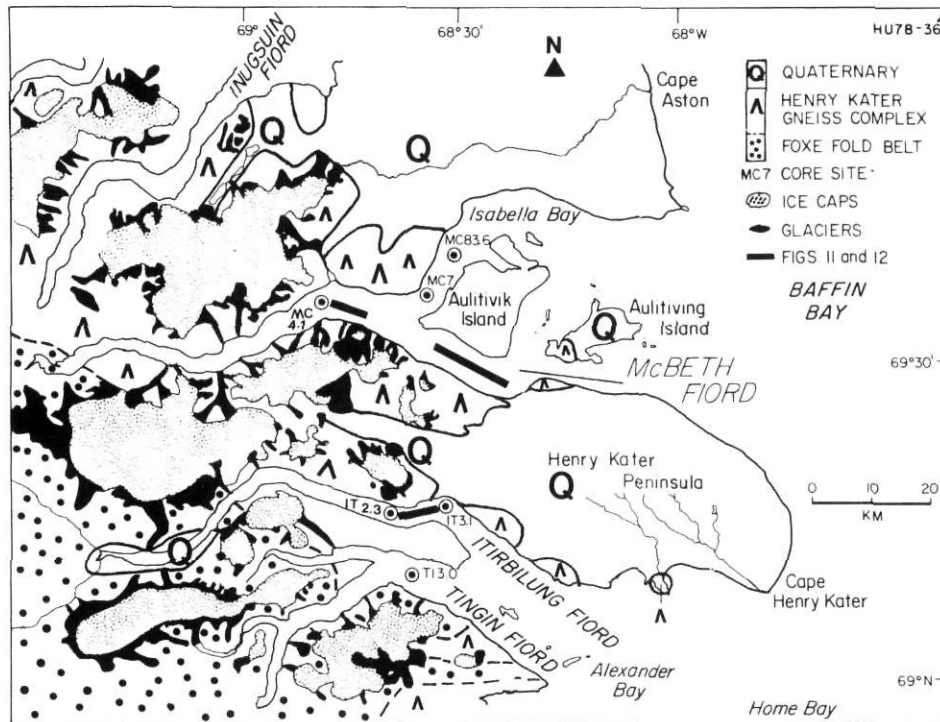
FIGURE 6. Generalized bedrock map of the east-central area showing the extent of the Foxe Fold Belt and the Henry Kater gneissic complex. The transects for Figures 10 and 11 are shown in addition to the core locations.

Carte généralisée du sous-bassement rocheux de la région des fjords du centre-est montrant l'étendue de la zone de plissements de Foxe et du complexe gneissique Henry Kater. Les transects des figures 10 et 11 sont indiqués.

TABLE II

Information on drainage basin areas (km²) (see Fig. 7)
(from Gilbert and MacLean, 1983 in Syvitski and Blakeney, 1983, and in Syvitski, 1984)

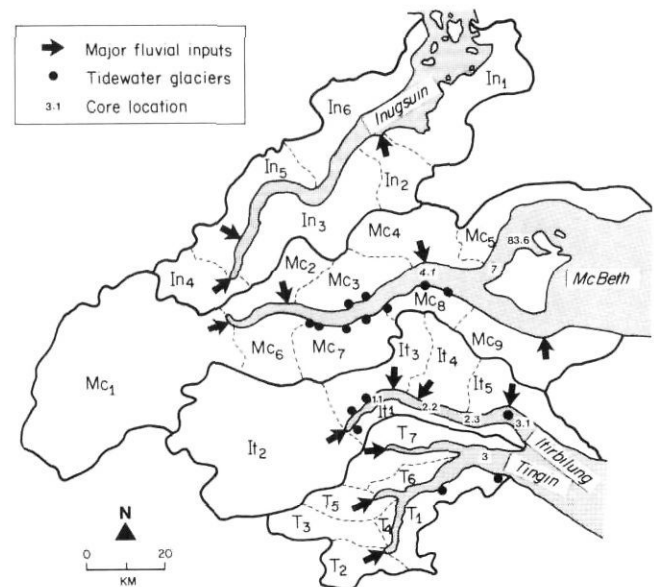| Basin No. (Fig. 7)	Tingin	Itirbilung	McBeth	% Foxe Fold Belt in the three areas
Total drainage	1261	2242	4038	
Fiord area	218	162	402	
1	302	141	1664	100, 26, 89
2	140	1243	278	100, 98, 2
3	179	327	247	100, 29, 0
4	35	243	347	100, 0, 0
5	143	286	335	100, 0, 0
6	92		234	100, 0, 71
7	370		362	100, 0, 16
8			103	0
9			468	0

FIGURE 7. Drainage basins, locations of tidewater glaciers, and major fluvial inputs for the east-central fiords (from Gilbert and MacLean, 1983; Syvitski, 1984).

Bassins versants, localisation des glaciers «littoraux» et des principaux apports fluviatiles de la région des fjords du centre-est (de Gilbert et MacLean, 1983; Syvitski, 1984).

lationships along a plane approximately parallel to the long axes of the fiords. These data: 1) delimit the probable extent of late Foxe (Wisconsin) ice in the fiords; 2) provide some constraints on the nature of ice on the shelf; 3) provide minimum dates for the fiord basin-fills; and 4) provide information on the rate of ice retreat in the fiords. The data break into three age groups (Figs. 4 and 8). First there are the dates from Cape Aston, a massive raised marine delta (Loken, 1966), which indicate that it was deposited >54 ka BP (Y-1703). This, plus the >52 ka date (QL-976) on whalebone from a delta on the south side of Mcbeth Fiord (Miller, 1979), and a number of "old" dates from wave-cut sections between McBeth and Itirbilung fiords (King, 1969), suggests that ice reached

the outer coast >50 ka. Amino acid racemization data on shells from several of these sites indicate that most of these deposits are coeval with the Kogalu aminozone or the Ayr Lake stadial (*i.e.* early Foxe Glaciation) which is currently considered to date from *ca.* 80 ka (*cf.* Andrews *et al.*, 1985b; Miller, 1985).
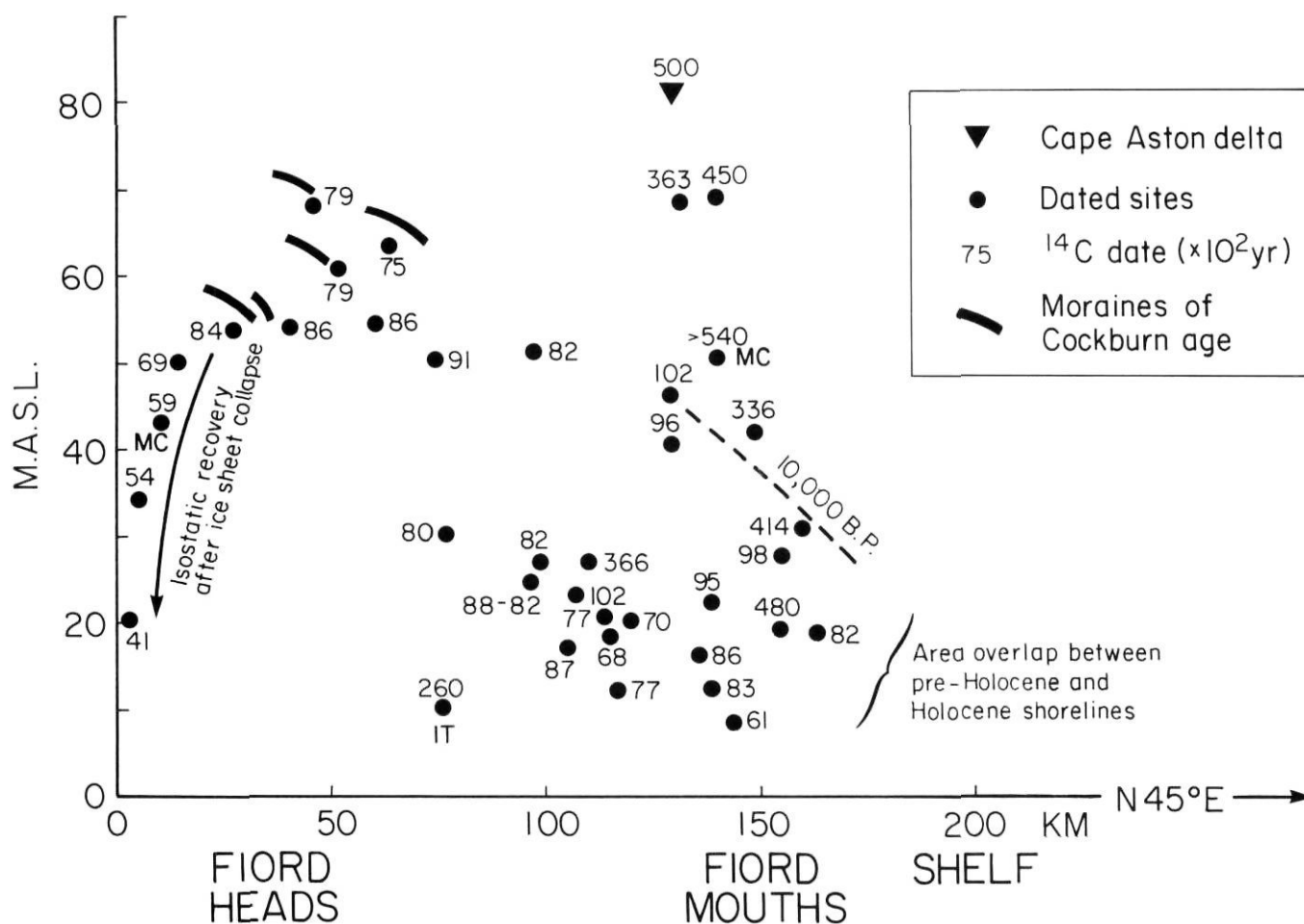
FIGURE 8. Simplified shoreline diagram (from Andrews, 1980) from the east-central fiord area showing the elevation and age of dated marine deposits and the location of moraines of Cockburn age. Note the 26,000 date within Itirbilung (IT) fiord ice (ca. 75 km distance from the 0 km origin).

Diagramme simplifié des lignes de rivage (tiré de Andrews, 1980) de la région des fjords du centre-est qui montre l'altitude et l'âge des dépôts marins datés et la localisation des moraines de l'épisode de Cockburn. À noter la date de 26 000 ans dans les glaces du fjord Itirbilung (IT) (à environ 75 km du 0 km d'origine).

There is a major depositional hiatus in the terrestrial sequence between 10 and 80 ka, however, one unit on the outer coast between McBeth and Itirbilung may be equivalent to the Loks Land aminozone (Miller, person. commun. 1988; Miller, 1985) which probably dates close to 50 ka. King (1969) reported a date of 25,860 ± 800 (I-3212) from shells in a delta on the north side of Itirbilung near core site IT3.1 (Fig. 4). It is unclear how to interpret this date. There are several ^{14}C dates in the 10-11 ka range (Figs. 4 and 8) which suggest the marine emergence and deglaciation started in the latest Pleistocene. However, the sites are scattered, and the relative scarcity of shells in this age range may imply that nearshore marine conditions were severe, with limited sediment input for the construction of deltas and beaches.

In outer McBeth and Itirbilung, and at the head of Tingin, dates on the major raised deltaic sequences are between 9.1 and 8.7 ka (NOTE: all shell dates have been corrected for reservoir effects), whereas the delta at the head of Inugsuin is dated at 7.5 ka (Loken, 1967), but deglaciation of the head of McBeth is undated. The marine limit at the head of Inugsuin is 65 m (Loken, 1967) and the marine limit on the north side

of McBeth, near the fiord head, is ca. 60 m asl. Glacial tectonized sediment was discovered on the south side of McBeth, unfortunately the enclosed shell samples were contaminated by an unknown source (Andrews et al., 1989); samples close by, but at a lower elevation, were collected by Stravers in 1987 and dated ca. 5 ka. At the head of Itirbilung, massive clays, probably marine, extend to a maximum of 43 m asl.

Age estimates for the deglaciation of the heads of Itirbilung and McBeth can be developed from the regional shoreline isobases (Loken, 1967; King, 1969; Andrews et al., 1970). Such estimates indicate that the ice retreated onto land between 7 and 8 ka. In all these fiords, massive valley-fills of marine sediments and outwash sands and gravels extend to the local marine limit. During subsequent glacial isostatic rebound, rivers have eroded these valley-fills and moved the sediment onto and across the pro-delta.

Local glaciers and ice caps developed during the Neoglaciation (Andrews, 1982; Davis, 1985). Small glaciers reach tidewater along these fiords (Fig. 7) and built ice-cored moraine complexes. During the 1983 *Hudson* cruise I visited some of these glaciers and estimate the age of the readvance(s) by

lichenometry; previously, Harrison (1966) discussed the chronology of neoglaciation in the valley north of core site HU83-MC4.1 (Fig. 4) (Siward Glacier) and King (1969) had measured lichens on moraines near Itirbilung fiord. Harrison's work documented a series of proglacial lake shorelines and end moraines; the latest advance occurred between 1790 and 1820 AD with one older neoglacial advance, bearing *Rhizocarpon geographicum* s.l. diameters of 45 mm (*ca.* 900-1150 BP). On the south side of the fiord, tidewater glaciers were fronted with moraines bearing lichens dating from about 900, 300 and 100 BP.

In summary, the glacial geology and raised marine evidence indicates that the very outermost part of McBeth Fiord was not glaciated during the late Foxe Glaciation. Ice may have been in the outer fiords *ca.* 10-11 ka and by 8.7 ka copious sediment was being deposited in ice-proximal deltaic sequences. The retreat of the outlet glacier along McBeth was 30 m/a. This figure is higher than estimates from Home Bay (Andrews *et al.*, 1970) of between 10 and 20 m/a.

FIORD AND SHELF CORES AND CHRONOLOGIES

Figure 4 shows the location of the piston cores. In addition, I will also discuss two cores from the adjacent shelf, namely HU78-36 and HU78-37. The former core occurs east of Cape Aston (Figs. 2 and 4) whereas HU78-37 is located on the shelf, just south of Home Bay.

TABLE III

Uncorrected and corrected radiocarbon dates from cores (Lab. AA is an AMS facility)

Core No. (Fig. 5)	Depth (cm) in core	Material*	Lab No.	Reported Date	Corrected Date
Hu-					
78-36	62-68	S	GX-6280	11,770	11,420
78-37	215	O	GX-8755	8,285	5,680
	437	O	GX-8756	12,035	8,100
	569	O	GX-6608	16,360	10,900
83-MC83.6	85-88	O	AA-1012	12,970	8,730
	292	O	AA-0654	19,200	12,800
83-MC4.1	80-82	O	AA-1011	2,819	2,410
	325	S	AA-1801	4,780	4,370
	782-786	O	AA-0653	16,700	11,150
83-IT1.1	593	S	AA-1917	3,920	3,510
83-IT2.3	100-102	O	AA-2276	5,084	3,670
	370-375	O	AA-2275	8,390	5,750
	841-845	O	AA-1523	15,800	10,560
83-IT3.1	102-105	O	AA-3260	15,080	9,740
	445-452	O	AA-0935	13,500	9,080
82-TI3	137-142	O	GX-11335	5,185	3,670
	364-384	O	GX-9434	10,430	7,080
	1077-1108	O	AA-0190	12,890	8,680

* S = shell; O = acid-insoluble organic matter (Kihl, 1975).

There are two significant issues to be considered when discussing sediment rates from cores. These are: 1) how reliable are the ^{14}C dates; and 2) does the piston core top date from present?). These are two separate questions and neither can be answered with 100% confidence. Question #1 has been examined by Fillon *et al.* (1981), Andrews *et al.* (1985a); this paper does not add new insight into the problem of evaluating ^{14}C dates on the acid-insoluble total oganic matter (AIOM) fraction. Shells were very rare to absent in the cores, but in three cases shells were used for dating (Table III). The palynology of HU83-MC4.1 and MCIT3.1 (Short *et al.*, in press; Andrews, 1987b) indicated that the amount of pre-Quaternary pollen in the cores was very low. Thus contamination by "old" carbon is not as easy to document as it was for two other cores (Andrews *et al.*, 1985a), nevertheless, I feel that the AIOM dates are too old, or at best "less than or equal to". For this paper I use reservoir corrected shell dates and corrected AIOM dates (Table III).

Examination of 6 cores along the shelf and fiords indicated a peak in Dinoflagellates (Short *et al.*, in press) that was dated *ca.* 5 ka; the synchrony of this event, and others (Jennings, 1986), suggests that the corrected AIOM dates are "reasonable" estimates.

There are two measures of the recovery of the upper section of sediment. One is the difference in length between the recovered sediment and the length of penetration of the coring barrel (Table IV), and the other is the similarity between the upper 1-2 m of sediment in the piston core, compared with a 11 cm diameter Lehigh gravity core from the same station. A rigorous analysis of the problem is not conducted in this paper, suffice to note that sediment has been by-passed. It is assumed in Figure 9 that the surface of the piston cores dates from 500 BP.

The ^{14}C dates (Table III) indicate that deposition was occurring on the shelf between 11 and 12 ka. The corrected date from the base of HU83-MC83.6 suggests that deposition in the outer north arm of McBeth was underway by *ca.* 13 ka and the corrected dates from the distal fiord basins all suggest

TABLE IV

Information on the core (see Fig. 4)

Core No.	Lat. and Long.	Water depth (m)	Core penetration (A)	Core length (B)	Ratio B/A
78-36	70°08'08", 66°48'07"	99	NA	99	NA
MC83.6	69°40'07", 69°09'08"	429	450	306	0.68
MC7	69°37'05", 69°16'00"	497	NA	1121	NA
MC4.1	69°31'04", 69°57'00"	549	810	610	0.75
IT3.1	69°17'06", 68°12'03"	365	810	483	0.6
IT2.3	69°17'05", 68°27'00"	410	853	610	0.72
IT1.1	69°20'00", 69°03'08"	256	900	793	0.88
TI3	69°11'05", 68°23'05"	487	NA	1108	NA
78-37	68°15'05", 65°12'09"	457	NA	593	NA

N.B. Cores MC7 and TI3 were collected in 1982, whereas MC83.6, MC4.1, IT3.1, IT2.3 and IT1.1 were raised in 1983.
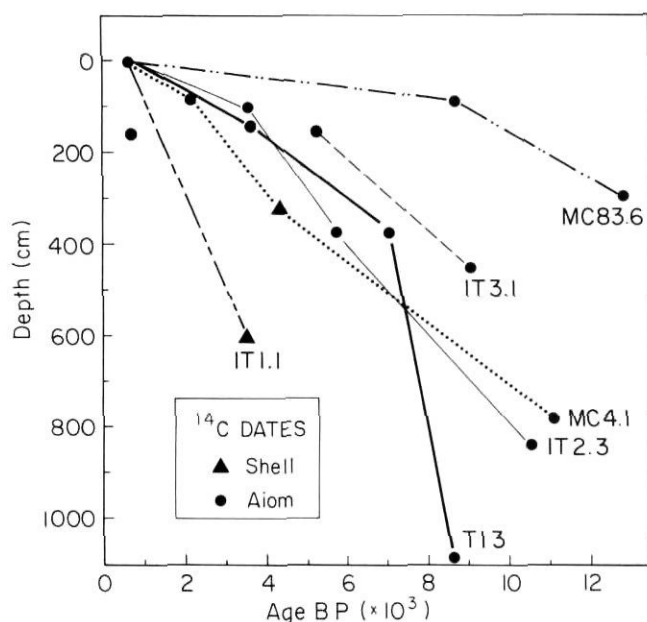
FIGURE 9. Age/depth diagram for piston cores from the east-central fiords. Location of the piston cores from the east-central fiords. Location of the piston cores is shown on Figure 4. The sedimentation rate for IT3.1 is based on a date and paleomagnetic correlation.

Diagramme de la profondeur sur l'âge des carottes provenant de la région des fjords du centre-est. L'emplacement des sites de carottages apparaissent à la figure 4. Le taux de sédimentation de IT3.1 est fondé sur une date et une corrélation paléomagnétique.

that these sediments were deposited < 10.5 ka. Figure 9 shows depth/age data for the fiord cores. These curves indicate that there is high sediment inputs near the fiord heads (IT1.1) which decrease to the outer fiord (MC83.6). Average net sedimentation rates vary between 1.5 m/ka to 0.23 m/ka. At ODP Site 645, just to the NE (Fig. 2), the Pleistocene sedimentation rate is uncertain, and is either *ca.* 0.15 m/ka or 6-9 times smaller (*cf.* de Vernal *et al.*, 1987).

A critical question is whether we can reasonably extrapolate back in time using the sediment accumulation curves to the base of the seismic section (Fig. 9)? The concept of paraglaciation (Church and Ryder, 1972), and our knowledge of melt rates of the Laurentide Ice Sheet, suggests that the highest rates of sediment accumulation would occur during deglaciation, that is between *ca.* 9 and 6 ka. The data (Fig. 9) and other published sedimentation rates from the region (Andrews *et al.*, 1985a; Jennings, 1986; Horvath, 1986) indicate maximum accumulation rates occurred in the 6-9 ka interval. Thus a *conservative* measure of the age of the base of the section may be derived by extrapolating the deglacial rates.

ACOUSTIC STRATIGRAPHY OF SOME FIORD BASINS

High resolution Huntec deep-tow system (DTS) and airgun records were run in along the axes of the fiords (Gilbert, 1985; Syvitski, 1984; Jennings, 1986). Interpretation of these records is not straight-forward because of interference from fiord-walls, and the lack of a 3-D perspective on the geometry of

the acoustic units. In particular, the influence of sediment sources from valley walls and side-valleys is difficult to evaluate. The two basic questions of interest in this paper are: 1) what is the total thickness of sediment in the fiords, particularly at the piston cores sites?; and 2) what is the nature of the acoustic records and how do they translate into the core lithostratigraphy? The critical question is the age and nature of the basal sediment fill.

In 1983 CSS *Hudson* entered McBeth Fiord in daylight. There was little evidence of basinal sediments in the north arm until we passed over a bedrock sill that was aligned with a series of lateral moraines and deltaic deposits that were described and dated by King (1969). On the proximal side of this feature the Huntec record (Fig. 10) suggests that sediment was overridden by ice (in Syvitski, 1984). Up-fiord, the sediment is ponded in a series of basins with little evidence for sediment draping. At HU83-MC4.1 (Fig. 10) at least 50 m of acoustically laminated sediment lies beneath the base of the core. The frequency of parallel acoustic laminations increases toward the surface. HU82-MC7 appears to penetrate a slide, although the paleomagnetic record (Andrews *et al.*, 1986) indicates that the upper 4 m is primarily *in situ*. HU83-MC83.6 (Fig. 4) was cored on a rise in the fiord and should thus have a low sedimentation rate. The Huntec record suggests the occurrence of 10-15 m of rather weakly laminated sediments, possibly overlying a diamicton. Thus, in the outer reaches of the McBeth there is little evidence for diamictons within the basinal sequences and it is difficult to trace major seismic reflectors from one basin to the next.

Syvitski (1984) interpreted the Huntec DTS records from Itirbilung. Seaward from IT3.1 (Figs. 4 and 11) the seismic stratigraphy is complex and suggests that the base of the section consists of diamictons, overlain by acoustically laminated sediments with a drape of more transparent sediments completing the sequence. There are some well defined reflectors within Itirbilung that have some axial continuity. Close to HU83-IT2.3 Syvitski (1984, Fig. 16-13) suggests that "Disturbed (overridden)?" sediments underlie "debris flows?" with a thick sequence of acoustically laminated sediments forming the upper 15-20 m. Radiocarbon dates on shells from the north shore of Itirbilung Fiord (Fig. 4) imply that a minimum date for the till/glacial-marine contact near HU83-IT3.1 should be >9.1 ka.

A paleomagnetic and radiocarbon study of IT3.1 (Table III; Andrews and Jennings, subm.) indicates the importance of both these techniques in studies of sedimentation rates. This core is the only one in which there is a reversal in the stratigraphic position of the ^{14}C dates and the paleomagnetics indicated that between 1.5 and 0.5 m core depth the inclination of the remnance magnetism was reversed or low. Andrews and Jennings (subm.) proposed that about 1 m of sediment was emplaced by a gravity flow. Study of the paleomagnetics of the other cores used in this study indicates that this problem does not apply to them and continuous deposition can be supported.

A major change in the rate of sedimentation should have occurred when the retreating outlet glaciers reached the heads
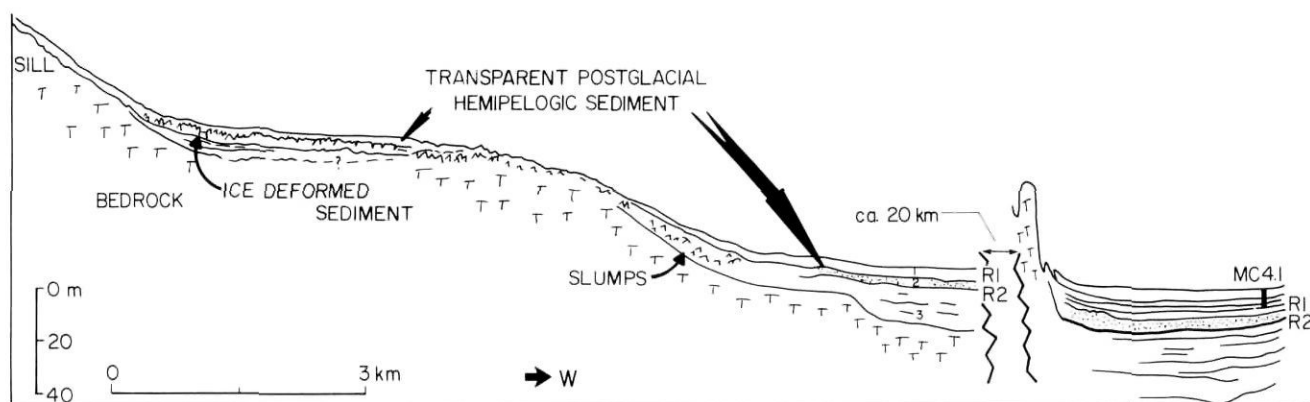
FIGURE 10. Interpreted Huntec DTS record from the outer, southern branch of McBeth Fiord. The location of this profile is shown on Figure 6. Note the break in the profile. R1 and R2 are two reflectors that can be traced throughout the outer fiord. Note the position of core MC4.1 to the extreme right of the diagram.

Profil obtenu à partir de l'interprétation de données obtenues par dispositif Huntec sur le bras extérieur sud du fjord McBeth (localisation à la fig. 6). À noter la cassure dans le profil. R1 et R2 sont deux réflecteurs qui peuvent être retracés partout dans le fjord externe. À noter l'emplacement du trou de forage MC4.1, à l'extrême droite.
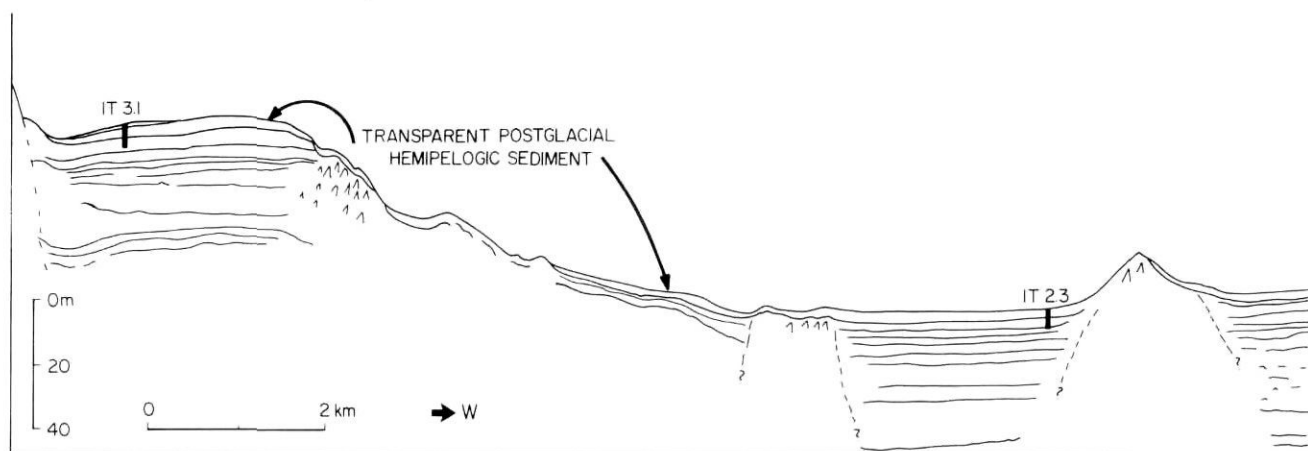

FIGURE 11. Interpreted Huntec DTS record from Itirbilung Fiord between HU83-IT3.1 and HU83-IT2.3 (Fig. 6). Location of the cores is shown relative to the DTS profiles.

Profil obtenu à partir de l'interprétation de données obtenues par dispositif Huntec entre HU83-IT3.1 et HU83-IT2.3 dans le fjord Itirbilung (fig. 6). La localisation des trous de forage est montrée en rapport avec les profils.

of the fiords, that is as early as 8.7 ka for Tingin and *ca.* 7.5 ka in Itirbilung, McBeth, and Inugsuin (Fig. 4). The evidence from McBeth and Itirbilung fiords is not conclusive, but there is some suggestion that sediment was overridden in McBeth (Fig. 10) and till may interfinger with glacial marine sediments in part of Itirbilung (Syvitski, 1984).

Analysis of the seismic data (Gilbert, 1985) indicates that the outer fiord basins contain about 50 m of sediment (Figs. 10 and 11). At this point two hypotheses can be advanced (*e.g.* Figs. 1 and 5): in the first we can *assume* that sediment from earlier deglacial intervals has been removed, hence the basins began operating as sediment traps upon deglaciation some 8 to 10 ka; in the second scenario some proportion of the basin fills was not eroded during the late Foxe advance (Fig. 10). There is not enough existing data to argue strongly for one or other of these alternative, although I argue that the weight of available evidence favours the suggestions embodied in Figure 5.

One factor that needs to be taken into account is the influence of basin geometry on the rate of sediment accu-

mulation. For example, if the flux of sediment (kg/m^2 a) to the seafloor was constant the rate of sediment accumulation (m/a) would vary between a U-shaped and V-shaped cross section. Calculations indicate that this is of no concern when considering the upper 10-20 m of sediment, but at some point it would be a factor. It is thus important that future research determine the 3-dimensional geometry of fiord-basin fills. Because fiord basin do narrow with depth this means that the extrapolation of basal sediment ages, using data on Figure 9, results in maximum age estimates.

If the fiord glaciers advanced to the fiord mouths during the last glaciation (*e.g.* Fig. 8), and in doing so excavated their basins of the pre-existing sediment fill, as in Figure 1, the question is: Where is the sediment? Huntec DTS records do not show massive moraine-like features at the mouth of either McBeth or Itirbilung. This fact supports the model portrayed as Figure 5. In addition, analysis of the Huntec DTS records from McBeth and Itirbilung fiords (Figs. 10 and 11) indicate that the acoustic stratigraphy within the *outer* basins consists of a series of strong, parallel reflectors that show

little evidence of up-fiord thickening toward a glacial sediment source. There is little doubt that sedimentation rates should decrease exponentially down fiord from an ice front, hence the nature of the acoustic records (Figs. 10 and 11) suggests deposition in distal basins, some distance from a point sediment source.

Existing data can not resolve the question of the rate of sediment infilling. What is required is the operation of the "long-coring facility" in the outer fiord basins. With cores of 20-30 m in length, and with the AMS ^{14}C facilities, it is probable that sufficient shell and/or foraminifera can be recovered for more reliable dates. In addition, Huntec DTS profiles are required *across* the fiords to obtain better data on sediment geometries.

REGIONAL VERSUS LOCAL SEDIMENT SOURCES

The last section of this paper deals with the style and source of fiord sediments in the outer basins and how this might be used to provide a check on the age estimates of the cores. We can investigate the hypothesis that the basal core dates do indeed relate to regional glaciation from the Foxe Basin sector of the Laurentide Ice Sheet by examining the lithostratigraphy of the cores and indicators of sediment provenance, keeping in mind the broad regional bedrock geology (Fig. 6) and the area of drainage basins (Table II, Fig. 7). Previously, Andrews and Jennings (1987) noted that the magnetic susceptibility (MS) of sediments derived from the Foxe
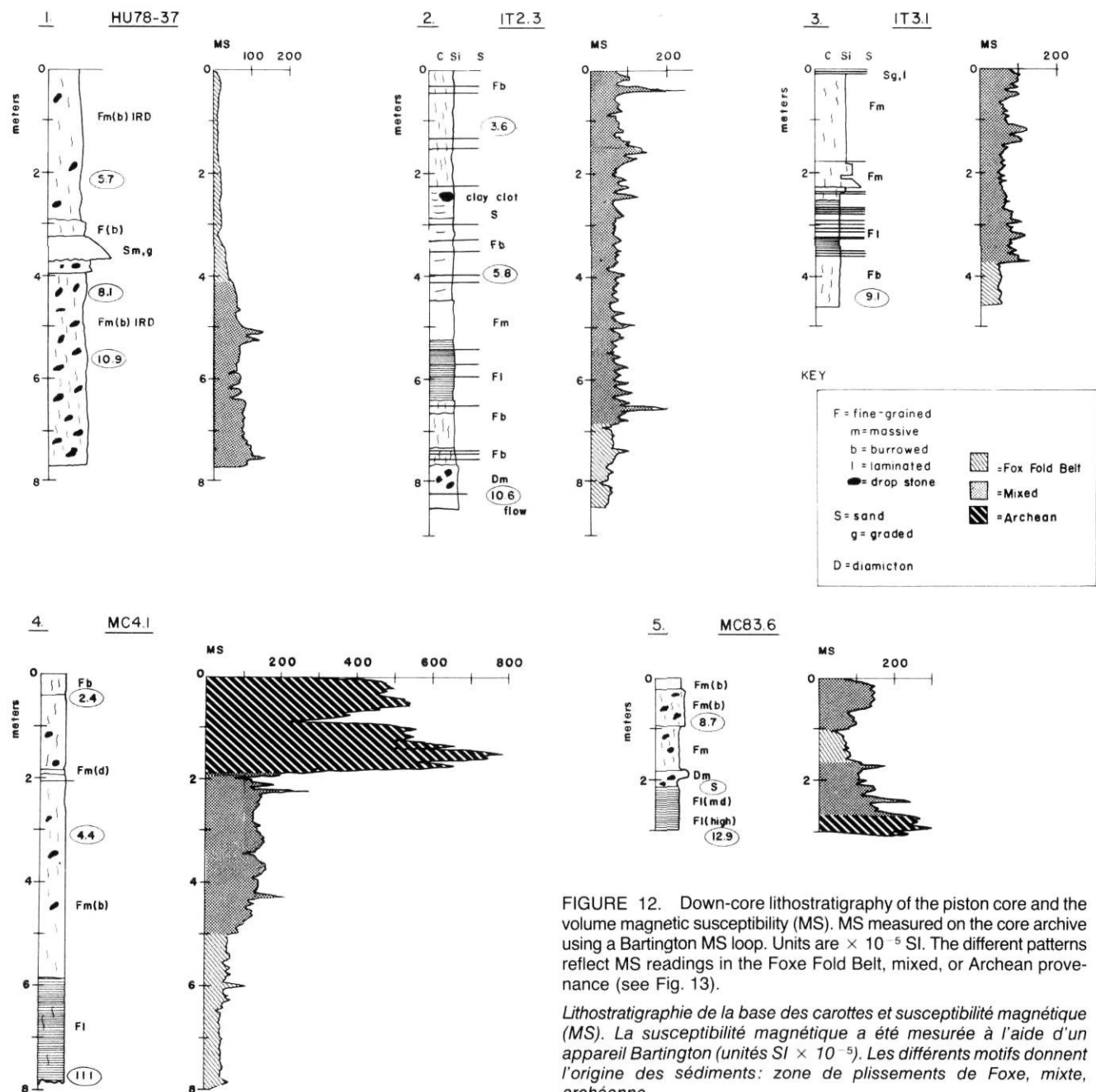

FIGURE 12. Down-core lithostratigraphy of the piston core and the volume magnetic susceptibility (MS). MS measured on the core archive using a Bartington MS loop. Units are $\times 10^{-5}$ SI. The different patterns reflect MS readings in the Foxe Fold Belt, mixed, or Archean provenance (see Fig. 13).

Lithostratigraphie de la base des carottes et susceptibilité magnétique (MS). La susceptibilité magnétique a été mesurée à l'aide d'un appareil Bartington (unités SI $\times 10^{-5}$). Les différents motifs donnent l'origine des sédiments: zone de plissements de Foxe, mixte, archéenne.

Fold Belt were very low in comparison with sediments derived from the Archean region of the shield. Table II lists the basin areas which supply water and sediment to the fiords (Fig. 7) and the percentages of Foxe Fold Belt or Archean bedrock within each drainage (Fig. 6).

Core descriptions from the area are given by Hein and Longstaffe (1985), Andrews et al. (1986), Hein (1987) and in various chapters of the S.A.F.E. reports. Figure 12 illustrates the core stratigraphy and the downcore log of the volume magnetic susceptibility (MS) (Andrews and Jennings, 1987). The core descriptions are based on X-radiography and visual inspection. Terminology follows Eyles et al. (1985).

The cores consist primarily of fine-grained, burrowed to massive (Fm(b)) sediments with frequent small sand beds, often graded. Toward the base of most cores (see also Jennings, 1986) there is a fine-grained laminated (Fl) unit (Fig. 12). In Clark Fiord, Jennings (1986) noted that the number of beds in the Fl unit was approximately equal to the number of radiocarbon years allowed for deposition. The cores show a general upward sequence of Fl followed by Fm(b). Farrow et al. (1984) and Syvitski et al. (1989) discuss the present-day conditions that promote burrowing of the sediment by a variety of organisms, this is related to nutrients and the well-oxygenated bottom waters in the fiords. The presence of bioturbation also places some upward limits on the rate of sediment accumulation. The conditions under which the Fl units were deposited are not clear. Ekdale and Mason (1988) suggest that one possible scenario would be anoxic conditions. Such conditions may indeed prevail in polar areas where either a permanent ice cover (sea-ice or ice shelf) or lack of meltwater may restrict the typical fiord/estuarine circulation. If such an environmental model is appropriate, then core stratigraphies (Fig. 11) suggest that such an interval was followed by increased oxygenated conditions and the establishment of an active infauna (Syvitski et al., 1989). The frequency of sand beds varies among fiords, and decreases seaward. They represent individual events associated with sediment transport along the axis of the fiord or from the fiord walls.

The question next addressed is the change(s) in sediment provenance. If the cores represent rapid sedimentation under present-day conditions then the sediment provenance should remain relatively constant with depth (= time). Conversely, major changes in provenance may be associated with changes in extent and volume of ice. The parameter chosen to investigate this question was the magnetic susceptibility of the sediment (Thompson and Morton, 1979, Bradshaw and Thompson, 1985; Dearing et al., 1981; Tarling, 1983).

The magnetic susceptibility (MS) of surface sediments in the fiords (Andrews and Jennings, 1987) shows a large difference between sediments at the head of fiords that lie all or mainly within the Foxe Fold Belt. Values (all values quoted as SI units $\times$ 10^{-5}; Figs. 12 and 13) are between 10 and 20 SI at the head of Itirbilung and Tingin fiords compared with >200 SI in Inugsuin and further north in Cambridge Fiord. Although MS is moderately correlated with grain-size (Andrews and Jennings, 1987), such that finer-grained sediments have lower MS readings, variations in grain-size do not swamp the order of magnitude difference between these two major bedrock types. In Figure 13 a "mixing line" is drawn that joins 100% Foxe Fold Belt MS values with those totally within the Shield granites/gneisses. The median and range of MS readings from cores (Fig. 4) are plotted along this line, individual downcore MS logs are shown on Figure 12.

On the basis of present surface samples three categories of sediment source are delimited (Fig. 13); switches from one mode to another are shown on Figure 12. An important aspect of IT2.3, IT3.1, and MC4.1 is the Foxe Fold Belt signature at the base of the cores. In HU83-MC83.6, this signature occurs in the middle of the section (Fig. 12), with an age at the top of the section of ca. 9 ka, approximately the same age at the low MS readings in the other cores. I suggest that these data must be associated with the period of maximum glacial ex-
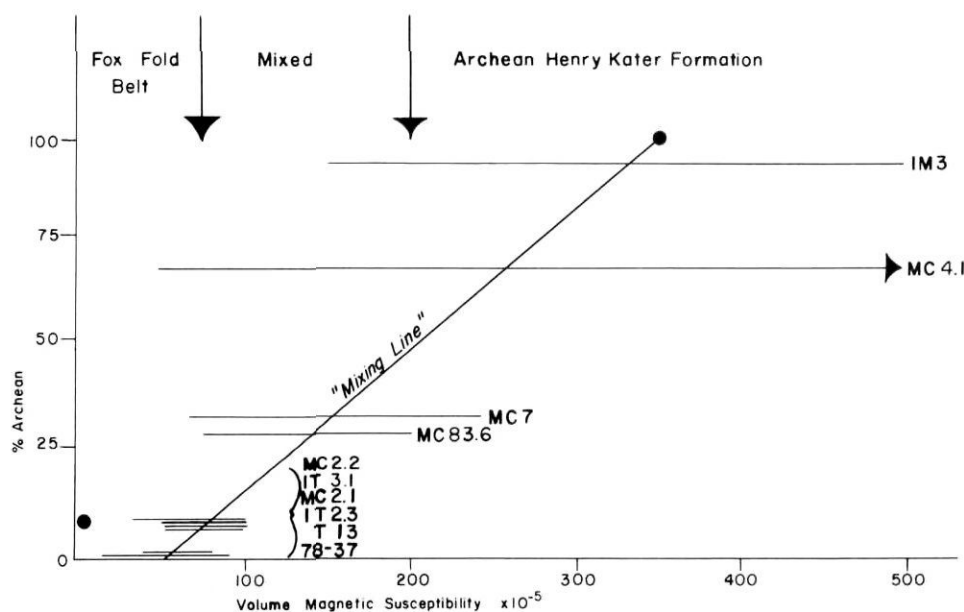

FIGURE 13. Diagram of the variations in the magnetic susceptibility of piston cores with respect to a "mixing line" that indicates the trend from 100% Foxe Fold Belt derived sediment to 100% Archean.

Diagramme de la variation de la susceptibilité magnétique des carottes en rapport avec une ligne de «transition» qui relie les sédiments à 100 % de la zone de plissements de Foxe aux sédiments à 100 % d'origine archéenne.

pansion where ice was being channelled into the fiords after a sustained flow across the interior outcrop of the Foxe Fold Belt (Fig. 6). This episode was followed in Itirbilung and McBeth fiords by the input of sediment that had a more Archean MS signature (Figs. 12 and 13). This indicates that the contribution of sediment from local mountain glaciers and ice caps (Fig. 7) increased proportionally during the middle and late Holocene. Such a change would only occur after 7 ka, at the time when the ice sheet lay landward of the fiord heads (cf. Dyke, 1974; Andrews et al., 1970). The southern shelf core, HU78-37, lies in an area where there is a plume of low MS (on surface sediments; Andrews and Jennings, 1987). Figure 12 indicates that this site was influenced by sediment from the Foxe Fold Belt for several thousands of years, although some of the low MS in the core may also reflect increase in organic carbon and diatom numbers (Williams, 1988).

This analysis of the core lithologies and provenance strongly suggests that the interval covered by the piston cores is indeed of the order of 10 ka. Thus, it is indeed possible that some fiord basins retained sediment during the last glacial advance (see Figs. 10 and 11).

CONCLUSIONS

The landscapes of Baffin Island partly reflect glaciation, but at the regional scale, large areas (42% of the total) show little evidence of glacial erosion (Andrews et al., 1985c). Areas of intense glacial scour, as shown by the density of small, structurally controlled rock-basins (Sugden, 1977; Andrews et al., 1985c) lie toward Foxe Basin or are concentrated at the heads of fiords. Aerially sustained average rates of denudation of 0.2 m/a (Table I) are difficult to reconcile with the regional geomorphology. In addition, Harbour et al. (1988) have shown that a reasonable glaciological model of glacial erosion requires 10^5 yr for creation of a U-shaped valley, i.e. rather a long period of time.

The morphology of the Baffin Island continental shelf is significantly different from those along the Labrador coast or off West Greenland. A striking feature of the Baffin Island shelf is the absence of the coast-parallel troughs that mark the Tertiary/basement contact. These deep marginal troughs imply the presence of grounded and erosive glacial ice on the shelf (e.g. Josehans et al., 1986). I propose that the absence of these features on the Baffin Island shelf implies the contrary condition, i.e. that the shelf was not covered by thick grounded ice.

The lack of deep troughs connecting every fiord to the shelf slope indicates that the sediment supply from the fiords to the deep-sea would be restricted to either a few major routes, or to times when grounded ice lay at the shelf edge. In the terrestrial record, the evidence for the latter is not at all clear, as few basal tills have not been recognized in the wave-cut outcrops along the outer coast (Mode et al., 1983; Feyling-Hanssen, 1985). However, the analysis of the ODP 645 data (Srivastava et al., 1987) indicates that the detrital carbonate-rich sediments are the "glacial" indicators. This implies that during the glacial intervals the supply of sediment to the deep sea was not dominated by sediment transfers

from the east-central fiords because the carbonate content of these sediments is usually < 2% (e.g. Jennings, 1986).

The low shear stresses ascribed to some former outlet glaciers on Baffin Island indicates that the ice may have moved on a water layer or by deformation of fiord sediments with a relatively high moisture content (e.g. Boulton and Jones, 1979). Whether any of the structures or reflectors in the fiords represent such units is an important question that this paper cannot answer (e.g. Fig. 10). However, a deforming bed would be a method of transporting sediment toward the ice front, yet it is the absence of a thick sediment wedge in the outer fiords and inner shelf proximal to the ice limit (Fig. 8) that is currently a puzzle.

REFERENCES

Aksu, A. E., 1985. Climatic and oceanographic changes over the past 400,000 years: Evidence from deep-sea cores in Baffin Bay and Davis Strait, p. 181-209. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Aksu, A. E. and Mudie, P. J., 1985. Late Quaternary stratigraphy and paleoceanography of northwest Labrador Sea. Marine Micropaleontology, 9: 537-557.

Andrews, J. T., 1980. Progress in relative sea level and ice sheet reconstructions Baffin Island, N. W. T., for the last 125,000 years, p. 275-200. In N.-A. Morner (ed), Earth rheology, isostasy, and eustasy. Wiley & Sons, London.

———— 1982. Holocene glacier variations in the Eastern Canadian Arctic. Striae, 18: 9-14.

———— 1987a. Late Quaternary marine sediment accumulation in Fiord-Shelf-Deep transects, Baffin Island to Baffin Bay. Quaternary Science Reviews, 6: 231-243.

———— 1987b. Downcore variations in the carbon content of fiord piston cores and association with sedimentation rates. In J. P. M. Syvitski and D. B. Praeg (compilers), Sedimentology of Arctic Fjords Experiment: Data Report, Volume 3. Canadian Data Report of Hydrography and Ocean Sciences No. 54, Chapter 12, 19 p.

Andrews, J. T. and Jennings, A. E., 1987. Influence of sediment source and type on the magnetic susceptibility of fiord and shelf deposits, Baffin Island and Baffin Bay, N. W. T. Canadian Journal Earth Sciences, 24: 1386-1401.

—— subm. Geomagnetic secular variations in high latitude fiord sediments.

Andrews, J. T., Laymon, C. A. and Briggs, W. M., 1989. Radiocarbon Date List VI, Baffin Island and III Labrador-Ungava. University of Colorado, Boulder, INSTAAR Occasional Paper No. 46, 85 p.

Andrews, J. T., Buckley J. T. and England, J. H., 1970. Late-glacial chronology and glacio-isostatic recover, Home Bay, east Baffin Island. Geological Society of America Bulletin, 81: 1123-1148.

Andrews, J. T., Jull, A. J. T., Donahue, D. J., Short, S. K., and Osterman, L. E., 1985a. Sedimentation rates in Baffin Island fiord cores from comparative radiocarbon dates. Canadian Journal of Earth Sciences, 22: 1827-1834.

Andrews, J. T., Aksu, A. E., Kelly, M., Klassen, R., Miller, G. H., Mode, W. N., and Mudie, P. J., 1985b. Land/ocean correlations during the Last Interglacial/Glacial transition, Baffin Bay, Northwestern North Atlantic: A review. Quaternary Science Reviews, 4: 333-355.

Andrews, J. T., Clark, P. U. and Stravers, J. A., 1985c. The patterns of glacial erosion across the eastern Canadian Arctic, p. 69-92. In J. T. Andrews (ed.), Quaternary environments: Baffin Island, Baffin Bay and West Greenland. Allen and Unwin, Boston.

Andrews, J. T., Mothersill, J. S. and Tabrez, A., 1986. Paleomagnetic record, texture, and mineralogy of Late Quaternary fiord sediments, Baffin Island. Arctic and Alpine Research, 18: 361-376.

Bell, M. and Laine, E. P., 1985. Erosion of the Laurentide Ice Sheet by glacial and glaciofluvial processes. Quaternary Research, 23: 154-174.

Blankenship, D. D., Ronney, S. T., Alley, R. B. and Bentley, C. R., 1986. Seismic evidence for a deforming and eroding till beneath an active Antarctic ice stream. EOS, 67: 946.

Bonifay, D. and Piper, D. J. W., 1988. Probable Late Wisconsinan ice margin on the upper continental slope off St. Pierre bank, eastern Canada. Canadian Journal of Earth Sciences, 25: 853-865.

Boulton, G. S. and Jones, A. S., 1979. Stability of temperate ice caps and ice sheets resting on beds of deformable sediment. Journal of Glaciology, 24: 29-43.

Bradshaw, R. and Thompson, R., 1985. The use of magnetic measurements to investigate the mineralogy of Icelandic lake sediments and to study catchment processes. Boreas, 14: 203-215.

Brown, C. S., Meier, M. F. and Post, A., 1982. Calving speed of Alaskan tidewater glaciers, with application to Columbia Glacier. U.S. Geological Survey Professional Paper 1258-C, 13 p.

Church, M. and Ryder, J. M., 1972. Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. Geological Society of America Bulletin, 83: 3059-3072.

Davis, P. T., 1985. Neoglacial moraines on Baffin Island, p. 682-718. In Andrews, J. T. (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Dearing, J. A., Elner, J. K. and Happey-Wood, C. M., 1981. Recent sediment flux and erosional processes in a Welsh Upland Lake-Catchment based on magnetic susceptibility measurements. Quaternary Research, 16: 356-372.

De Vernal, A., Hillaire-Marcel, C., Aksu, A. E. and Mudie, P. J., 1987. Palynostratigraphy and chronostratigraphy of Baffin Bay deep sea cores: Climatostratigraphic implications. Palaeogeography, Palaeoecology, Palaeoclimatology, 61: 97-105.

Dowdeswell, E. K. and Andrews, J. T., 1985. The fiords of Baffin Island: Description and classification, p. 682-718. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Dyke, A. S., 1974. Deglacial chronology and uplift history: northeastern sector, Laurentide Ice Sheet. Occasional Paper No. 12, INSTAAR, University of Colorado, Boulder, 73 p.

Ekdale, A. A. and Mason, T. R., 1988. Characteristic trace-fossil associations in oxygen-poor sedimentary environments. Geology, 16: 720-723.

Eyles, N., Clark, B. M., Kaye, B. G., Howard, K. W. F. and Eyles, C. H., 1985. The application of basin analysis techniques to glaciated terrains: An example from the Lake Ontario Basin, Canada. Geoscience Canada, 12: 22-32.

Farrow, G. E., Atkinson, J. and Moore, G., 1984. Macrobenthos and bioturbation. In J. P. M. Syvistky (compiler), Sedimentology of Arctic Fiords Experiment: HU83-028 Data Report, Volume 2. Canadian Data Report of Hydrography and Ocean Sciences No. 28, Chapter 6, 35 p.

Feyling-Hanssen, R., 1985. Late Cenozoic marine deposits of East Baffin Island and E. Greenland: Microbiostratigraphy, Correlation, Age, p. 354-393. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Fillon, R. H. et al., 1981. Labrador shelf: shell and total organic matter C-14 date discrepancies. In Current Research, Part B. Geological Survey of Canada, Paper 81-1B: 105-111.

Fisher, D. A., Reeh, N. and Langley, K., 1985. Objective reconstructions of the Late Wisconsinan Ice Sheet and the significance of deformable beds. Géographie physique et Quaternaire, 39: 229-238.

Gilbert, R., 1982. Contemporary sedimentary environments on Baffin Island, N.W.T., Canada. Glaciomarine processes in fiords of eastern Cumberland Peninsula. Arctic and Alpine Research, 14: 1-12.

—— 1983. Sedimentary processes of Canadian Arctic fiords. Sedimentary Geology, 36: 147-175.

—— 1985. Quaternary glaciomarine sedimentation interpreted from seismic surveys of fiords on Baffin Island, N.W.T. Arctic, 38: 271-280.

Gilbert, R. and MacLean, B., 1983. Geophysical studies based on conventional shallow and Huntec high resolution seismic surveys of fiords on Baffin Island. In J. P. M. Syvitski and C. P. Blakeney (compilers), Sedimentology of Arctic Fjords Experiment: HU 82-031 Data Report, Volume 1. Canadian Data Report on Hydrography and Ocean Sciences No. 12, Chapter 15, 90 p.

Harbor, J. M., Hallet, B. and Raymond, C. F., 1988. A numerical model of landform development by glacial erosion. Nature, 333: 347-349.

Harrison, D. A., 1966. Recent fluctuations of the snout of a glacier at McBeth Fiord, Baffin Island, N.W.T. Geographical Bulletin, 8: 45-58.

Hay, W. W. and Southam, J. R., 1977. Modulation of marine sedimentation by the continental shelves, p. 569-604. In N. R. Anderson and A. Malahoff (eds), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum, New York.

Hein, F. J., 1987. Core logs for HU83-028 piston cores and PA85-062 Lehigh cores from Itirbilung Fjord. In J. P. M. Syvitski and D. B. Praeg (compilers), Sedimentology of Arctic Fjords Experiment:

Data Report, Volume 3. Canadian Data Report of Hydrography and Ocean Sciences No. 54, Chapter 9, 44 p.

Hein, F. J. and Longstaffe, F. J., 1985. Sedimentologic, mineralogic, and geotechnical descriptions of fine-grained slope and basin deposits, Baffin Island fiords. Geo-Marine Letters, 5: 11-16.

Henderson, J. R., 1985. Geology, McBeth Fiord- Cape Henry Kater District of Franklin Northwest Territories. Geological Survey of Canada Map 1605A.

Hess, R., Chough, S. K. and Rakofsky, A., 1987. The Northwest Atlantic Mid-Ocean Channel of the Labrador Sea. V. Sedimentology of a giant deep-sea channel. Canadian Journal of Earth Sciences, 24: 1595-1624.

Horvarth, V. V., 1986. Glaciomarine sedimentation in Cambridge Fiord, Baffin island, N.W.T., Canada. MSc thesis, Queen's University, Kingston, 225 p.

Ives, J. D., Barry, R. G. and Andrews, J. T., 1975. Growth and decay of the Laurentide Ice Sheet and comparisons with Fenno-Scandinavia. Naturwissenschaften, 62: 118-125.

Jacobs, J. D., Andrews, J. T. and Funder, S., 1985. Environmental background, p. 26-68. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Jennings, A. E., 1986. Late Quaternary marine sediments from a transect of fiord and continental shelf environments: A study of piston cores from Clark Fiord and Scott Trough, Baffin Island. MSc thesis, University of Colorado, Boulder, 245 p.

Josenhans, H. W., Zevenhuizen, J. and Klassen, R. A., 1986. The Quaternary geology of the Labrador Shelf. Canadian Journal of Earth Sciences, 23: 1190-1214.

King, C. A. M., 1969. Glacial geomorphology and chronology of Henry Kater Peninsula, east Baffin Island, N. W. T. Arctic and Alpine Research, 1: 195-212.

Kravitz, J. H., 1982. Sediments and sediment processes in Kane Basin, a high arctic glacial marine basin. INSTAAR Occasional Paper no. 39, University of Colorado, Boulder, 184 p.

Klassen, R. A. 1985. An outline of Glacial History of Bylot Island, District of Franklin, p. 428-460. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Laine, E. P., 1980. New evidence from beneath the western North Atlantic for the depth of glacial erosion in Greenland and North America. Quaternary Research, 14: 188-198.

Locke, W. W. III, 1980. The Quaternary geology of the Cape Dyer area, southeasternmost Baffin Island, Canada. PhD thesis, University of Colorado, Boulder, 331 p.

—— 1987. The late Quaternary geomorphic and paleoclimatic history of the Cape Dyer area, easternmost Baffin Island, N.W.T. Canadian Journal of Earth Sciences, 24: 1185-1198.

Loken, O. H., 1966. Baffin Island refugia older than 54,000 years. Science, 153: 1378-1380.

—— 1967. Postglacial emergence at the south end of Inugsuin Fiord, Baffin Island, N.W.T. Geographical Bulletin, 7: 243-258.

MacLean, B., 1985, Geology of the Baffin Island shelf, p. 154-177. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay, and Western Greenland. Allen and Unwin, London.

Meier, M. M. and Post, A., 1987. Fast tidewater glaciers. Journal of Geophysical Research, 92: 9051-9058.

Miller, G. H., 1976. Anomalous local glacier activity, Baffin Island, Canada: Paleoclimatic implications. Geology, 4: 502-504.

—— 1979. Radiocarbon Date List IV: Baffin Island, N.W.T., Canada. University of Colorado, INSTAAR Occasional Paper No. 29, 61 p.

—— 1985. Aminostratigraphy of Baffin Island shell-bearing deposits, p. 394-427. In Andrews, J. T. (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Mode, W. N., 1980. Quaternary stratigraphy and palynology of the Clyde Foreland, Baffin Island, N.W.T., Canada. PhD Dissertation, University of Colorado, Boulder, 219 p.

—— 1985. Pre-Holocene pollen and molluscan records from Eastern Baffin Island, p. 502-519. In J. T. Andrews (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland. Allen and Unwin, London.

Mode W. N., Nelson, A. R. and Brigham J. K., 1983. A facies model of Quaternary glacial-marine cyclic sedimentation along eastern Baffin Island, Canada p. 495-534. In B. F. Molnia (ed.), Glacial-marine Sedimentation. Plenum Press, New York.

Mudie, P. A. and Guilbault, J.-P., 1982. Ecostratigraphic and paleo-magnetic studies of Late Quaternary sediments on the northeast Newfoundland Shelf. In Current Research, Part B. Geological Survey of Canada Paper 82-1B: 107-116.

Piper, D. J. W., 1988. Glaciomarine sedimentation on the Continental Shelf off Eastern Canada. Geoscience Canada, 15: 23-28.

Post, A. 1975. Preliminary hydrology and historic terminal changes of Columbia Glacier, Alaska. U.S. Geological Survey, Hydrological Investigation Atlas.

Praeg, D. B., MacLean, B., Hardy, I. A. and Mudie, P. J., 1986. Quaternary geology of the southeast Baffin Island continental shelf. Geological Survey of Canada Paper 85-14, 13 p. (with maps).

Short, S. K., Andrews, J. T. and Mode, W. N., 1989. Modern and late Quaternary pollen spectra from fiord sediments, eastern Baffin Island, Arctic Canada. Marine Micropaleontology, 15: 181-202.

Srivastaava S. P. et al., 1987. Proceedings of the Ocean Drilling Program. Part A- Initial Report, Baffin Bay and Labrador Sea. ODP 105, Texas A & M University, College Station, Texas, Volume 105, 917 p.

Sugden, D. E., 1977. Reconstruction of the morphology, dynamics and thermal characteristics of the Laurentide Ice Sheet at its maximum. Arctic and Alpine Research, 9: 21-47.

Sugden, D. A. and John, B. S., 1976. Glaciers and Landscapes. Edward Arnold, London.

Syvitski, J. P. M. (compiler), 1984. Sedimentology of Arctic Fjords Experiment: HU 83-028 Data Report, Volume 2. Canadian Data Reports on Hydrography and Ocean Sciences No. 28, 20 chapters, various paginations.

—— 1986. Estuaries, deltas and fjords of Eastern Canada. Geoscience Canada, 13: 91-100.

Syvitski, J. P. M. and Blakeney, C. P. (compilers), 1983. Sedimentology of Arctic Fjords Experiment: HU 82-031 Data Report, Volume 1. Canadian Data Report on Hydrography and Ocean Sciences No. 12. 935 p.

Syvitski, J. P. M. and Praeg, D. B., 1987. Sedimentology of Arctic Fjords Experiment: Data Report, Volume 3. Canadian Data Report of Hydrography and Ocean Sciences No. 54, various paginations, 15 chapters.

Syvitski, J. P. M. and Schafer, C., 1985. Sedimentology of Arctic Fjords Experiment (S.A.F.E.): Project Introduction. Arctic, 38: 264-270.

Syvitski, J. P. M., Burrell, D. C. and Skei, J. M., 1987. Fjords. Processes and Products. Springer-Verlag, New York, 379 p.

Syvitski, J. P. M., Farrow, G. E., Atkinson, R. J. A., Moore, P. G. and Andrews, J. T., 1989. Baffin Island fjord macrobenthos: bottom communities and environmental significance. Arctic, 42: 232-247.

Tarling, D. H., 1983. Paleomagnetism: Principles and Applications in Geology, Geophysics and Archeology. Chapman and Hall, London, 379 p.

Thompson, R. and Morton, D. J., 1979. Magnetic susceptibility and particle size distribution in recent sediments of the Loch Lomond drainage basin. Journal of Sedimentary Petrology, 49: 801-818.

Thouveny, N., 1988. High-resolution paleomagnetic study of Late Pleistocene sediments from Baffin Bay: first results. Canadian Journal of Earth Sciences, 25: 833-843.

Tippett, C. R., 1984. Geology of a transect through the southern margin of the Foxe Fold Belt (mainly NTS 27B), central Baffin Island, District of Franklin. Geological Survey of Canada, Open File 1110 and 1111: 73 p. and 83 p.

Williams, K. M., 1988. Late Quaternary Paleo-oceanography of the Baffin Bay Region, Based on Diatoms. Ph.D. dissertation, University of Colorado, Boulder, 263 p.