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THE NORWEGIAN–GREENLAND SEA CONTINENTAL MARGINS: MORPHOLOGY AND LATE QUATERNARY SEDIMENTARY PROCESSES AND ENVIRONMENT

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Abstract—The continental margins surrounding the Norwegian–Greenland Sea are to a large degree shaped by processes during the late Quaternary. The paper gives an overview of the morphology and the processes responsible for the formation of three main groups of morphological features: slides, trough mouth fans and channels.

Several large late Quaternary slides have been identified on the eastern Norwegian–Greenland Sea continental margin. The origin of the slides may be due to high sedimentation rates leading to a build-up of excess pore water pressure, perhaps with additional pressure caused by gas bubbles. Triggering might have been prompted by earthquakes or by decomposition of gas hydrates.

Trough mouth fans (TMF) are fans at the mouths of transverse troughs on presently or formerly glaciated continental shelves. In the Norwegian–Greenland Sea, seven TMFs have been identified varying in area from 2700 km^2 to 215 000 km^2. The Trough Mouth Fans are depocentres of sediments which have accumulated in front of ice streams draining the large Northwest European ice sheets. The sediments deposited at the shelf break/upper slope by the ice stream were remobilized and transported downslope, mostly as debris flows. The Trough Mouth Fans hold the potential for giving information about the various ice streams feeding them with regard to velocity and ice discharge.

Two large deep-sea channel systems have been observed along the Norwegian continental margin, the Lofoten Basin Channel and the Inbis Channel. Along the East Greenland margin, several channel systems have been identified. The deep-sea channels may have been formed by dense water originating from cooling, sea-ice formation and brine rejection close to the glacier margin or they may originate from small slides on the upper slope transforming into debris flows and turbidity currents. © 1998 Elsevier Science Ltd. All rights reserved.

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INTRODUCTION

Objectives

This paper aims to explain the morphology and late Quaternary sedimentary processes and palaeoenvironment of the Norwegian–Greenland Sea continental margin. In particular it concentrates on the North Norwegian and Barents Sea continental margin.

Physiographic setting

The first general description of the morphology of the continental margin, particularly the continental shelves, of the Norwegian–Greenland Sea was given by Nansen (1904). A comprehensive review on the morphology, sediments and paleoenvironments of the Norwegian–Greenland deep-sea as well as the continental margins was given by Vogt (1986).

The continental margins surrounding the Norwegian–Greenland Sea (Fig. 1) have been shaped by tectonic and sedimentary processes throughout the Cenozoic. Morphodynamic patterns have evolved during the late Cenozoic under the influence of the waxing and waning Fennoscandian-, Barents Sea- and Greenland ice sheets. An adequate understanding of the development of the continental margin surrounding the Norwegian–Greenland Sea relies on thorough knowledge of the morphodynamic features of the present sea floor as well as the hinterland.

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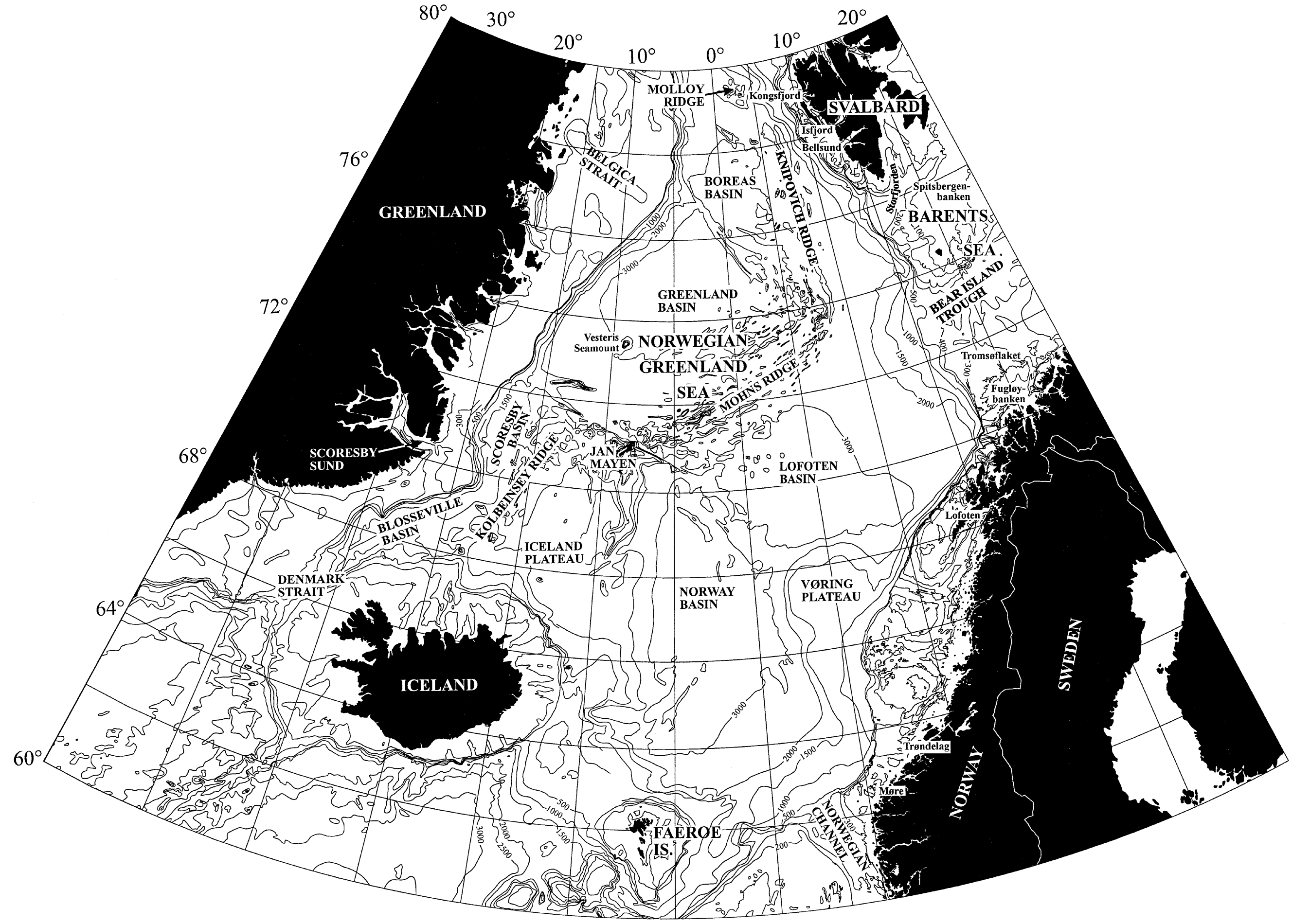

FIG. 1. Bathymetric map of the Norwegian Greenland Sea adapted from Perry *et al.* (1980). Water depth in meters.

The development of the Norwegian–Greenland Sea ocean basins (Fig. 1) is governed by plate-tectonic evolution (Talwani and Eldholm, 1977; Myhre *et al.*, 1992). In the Norwegian–Greenland Sea the mid-ocean ridge comprises four active segments, the Kolbeinsey Ridge, the Mohns Ridge, the Knipovich Ridge and the Molloy Ridge. The ridges are often interrupted by transverse valleys created by fracture zones cutting across their topography. The mid-ocean ridge and fracture zones separate the deep Norwegian–Greenland Sea into four major basins (Norway, Lofoten, Greenland and Boreas Basins), and two minor basins (the Scoresby and Blosseville Basins) (Perry, 1986) (Fig. 1).

A key element in the understanding of the Late Cenozoic evolution of the continental margin is an understanding of the Cenozoic uplift of Greenland, Scandinavia, the Barents Sea and Svalbard. A wide variety of hypotheses as to the timing and causes of the Cenozoic uplift have been proposed. Recent studies from the Barents Sea have suggested that much of the uplift occurred during the Plio–Pleistocene (Riis and Fjeldskaar, 1992; Eidvin *et al.*, 1993; Henriksen and Vorren, 1996; Stuevold and Eldholm, 1996). During and after the uplift large amounts of sediment were transferred from the hinterland to the continental slope. The outer part of the Norwegian–Barents Sea continental margin has a wedge of mostly clastic Late Cenozoic sediments up to 7 km (5 s twt) thick (Fig. 2).

The bathymetry of the East Greenland Continental Shelf is characterised by shallow banks and local depressions, with large troughs cut into the shelf bedrock. The maximum shelf width is 280 km at 77°25′N and the minimum width is 30 km at 81°50′N. Shelf depths range between 100 and 300 m, but shoals of less than 100 m exist (Perry, 1986). The continental slope off East Greenland varies in gradient from 1:38 at 70°N to 1:13 at 82°N. The slope ranges in depth from 400 to 1600 m in the south and from 400 to 3400 m in the north. A prominent feature of this margin is the fan-like feature offshore of Scoresby Sund (Fig. 1).

The continental shelf off Norway (Fig. 1) has great variations with regard to width and depth. Two regions are particularly narrow, namely offshore Møre (minimum width less than 65 km), and offshore Lofoten (minimum width less than 10 km). The shelf area between has a width exceeding 200 km (maximum 260 km). The narrowest areas are also the shallowest, with depths less than 200 m, while the shelf between is generally more than 250 m deep and has maximum depths of ca 500 m. Several troughs cross the shelf (Fig. 1). These transverse troughs often have a steep

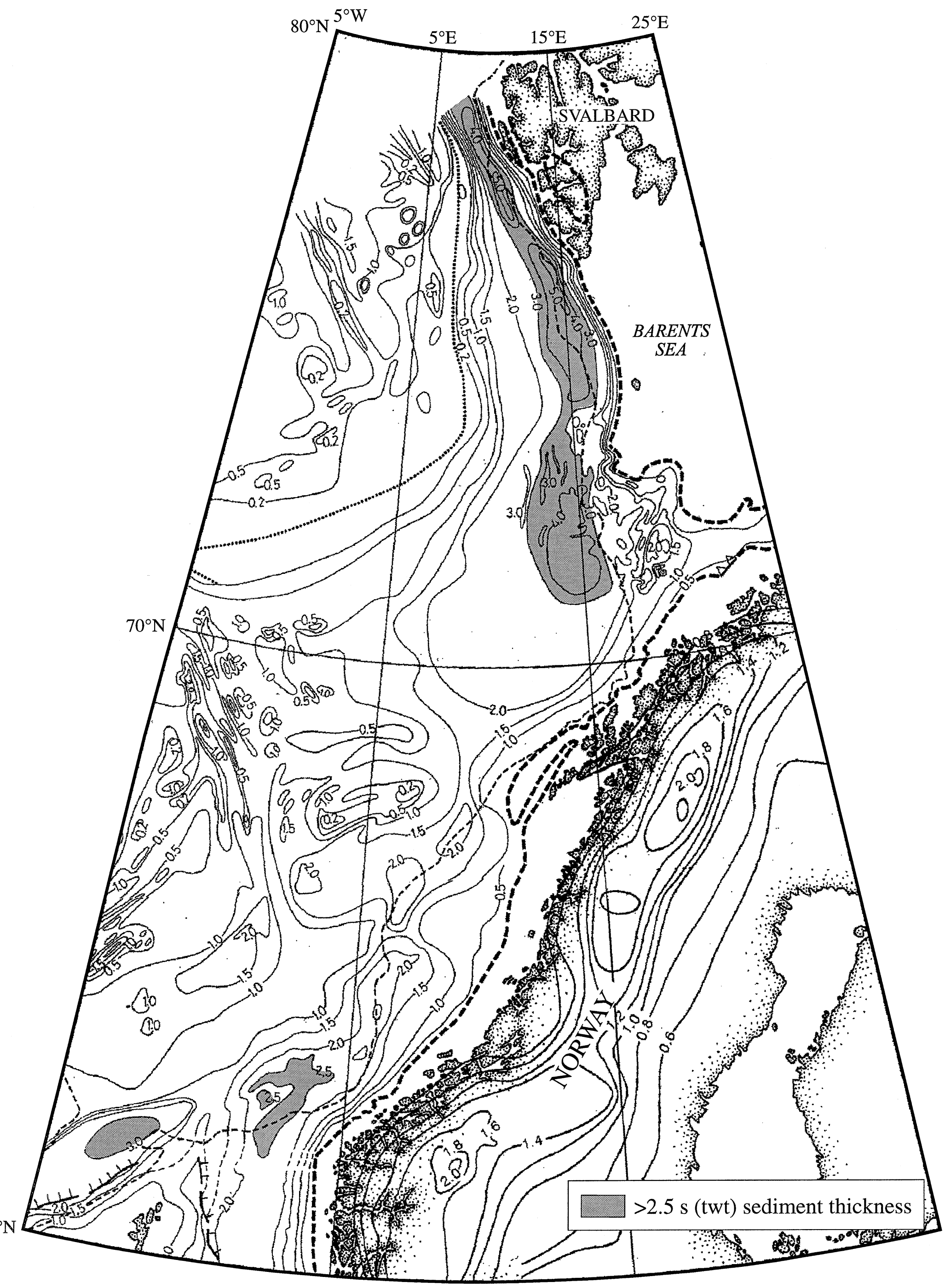

FIG. 2. Envelope surface of Norway (m × 1000) (Gjessing, 1967) and isopach map of the Cenozoic sediments in the Norwegian Sea (s twt) (adapted from Vorren *et al.*, 1991). Note the pronounced depocenters on the Barents Sea continental margin.

backwall aligned more or less along the longitudinal trough situated at the boundary between the landward crystalline rocks and the seaward sedimentary rocks (Holtedahl, 1971). The continental slope is steepest off the narrowest shelf areas. Offshore mid Norway the continental margin broadens due to a marginal plateau, the Vøring Plateau which exhibits depths mainly between 1200 and 1600 m.

The Barents Sea covers one of the widest continental shelves in the world (Fig. 1). At 40°E the shelf reaches a width of 1500 km. The western sea floor of the Barents Sea comprises a series of broad troughs of which the Bear Island Trough is the most prominent. This trough ranges between 300 and 500 m in depth. It is bordered by a series of shallower banks and platforms like Spitsbergenbanken to the north (minimum depth of less than 20 m) and Tromsøflaket (200 m) and Fugløybanken (200 m) to the south. The shelf west of Svalbard is narrow (30–60 km) and split into shallow banks by several transverse troughs.

The western continental margin of the Barents Sea is characterised by fan-shaped protrusions. The most prominent is off the mouth of the Bear Island Trough extending more than 280 km beyond the shelf break. Further north are fans off the Storfjorden and smaller fans off Bellsund, Isfjorden and Kongsfjorden on Svalbard (Fig. 1).

SLIDES

Norwegian margin

Several smaller and larger slides have been identified on the Norwegian continental margin (Fig. 3). Holtedahl (1971) seems to have been the first to indicate sliding activity on the Norwegian continental margin by writing: "off Møre the slope down to 1000 m is highly irregular and partly very steep, and this particular topography may be due to sliding activity, and perhaps also due to current activity" (translated from Norwegian). The existence of several slides on the continental margin was indicated by Damuth (1978).

During the 1980s, a detailed research programme organised by the IKU (Continental Shelf and Petroleum Technology Research Institute A/S, Norway) focused on the Storegga Slide region (Fig. 3). This resulted in a series of papers (Bugge, 1983; Bugge *et al.*, 1987, 1988; Jansen *et al.*, 1987).

Based on GLORIA long-range side-scan sonar data, Bugge (1983) and Kenyon (1987) gave an overview of slides occurring along the Northwest European margin up to 73°N. In the area between the Storegga Slide and 67°N, Bugge (1983) found "no real sliding or slumping of any reasonable size". Between 67° and 73°N, several larger and smaller slides were observed, noticeably the Trænadjupet and Bjørnøyrenna Slides (Fig. 3). Recent GLORIA mapping (Dowdeswell *et al.*, 1996) has indicated another large slide, the Andøya Slide and many small slides in this region. In the following we shall briefly describe these slides, and in Table 1 we have compiled available data of the largest slides.

The Storegga Slide

The Storegga Slide is one of the world's largest known submarine slides. The 290-km long headwall is located at the shelf break 100 km off the coast of Møre–Trøndelag (Fig. 4). About 5600 km^3 was involved in the sliding (Bugge *et al.*, 1988). Bugge (1983) distinguished three slide events; the first probably took

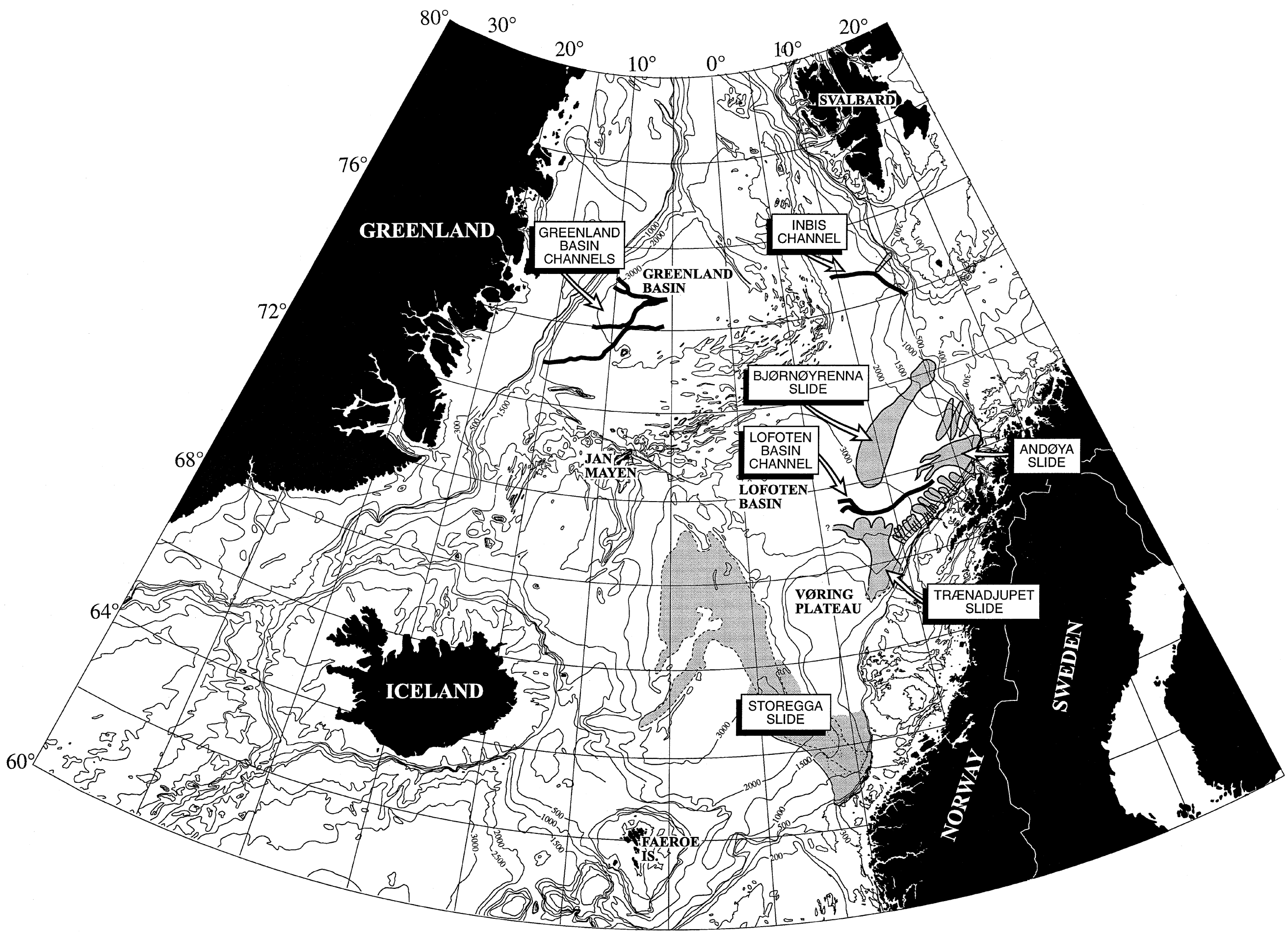

FIG. 3. Bathymetric map showing location and extent of slides, small slides and large channels on the Norwegian–Greenland Sea continental margin. The figure is compiled from several sources, e.g. Bugge *et al.* (1987), Mienert *et al.* (1993), Laberg and Vorren (1993), Dowdeswell and Kenyon (1995), Dowdeswell and Kenyon (1997) and Dowdeswell *et al.* (1996).

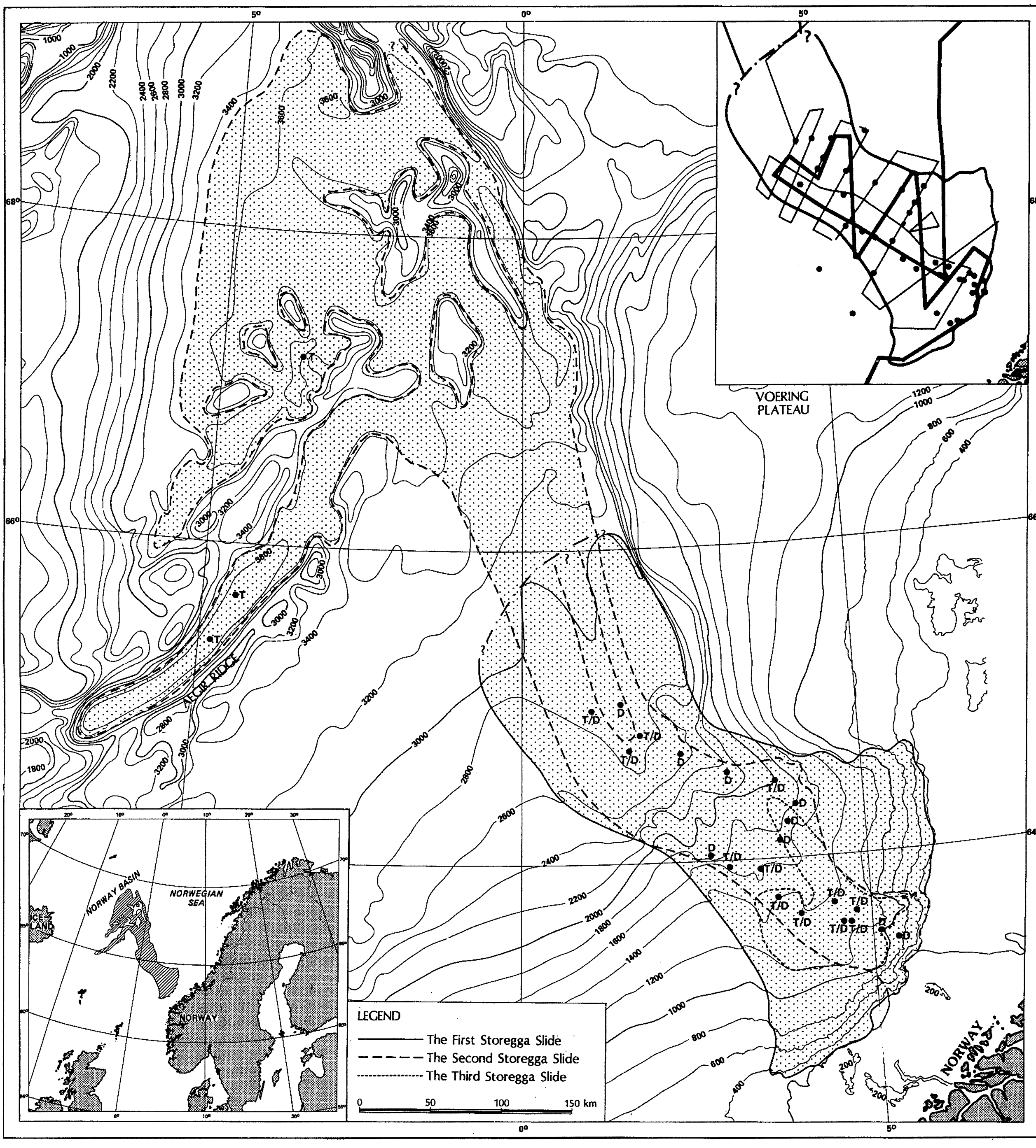

FIG. 4. Bathymetric map of the Storegga Slide area (after Bugge *et al.*, 1987). Water depth in meters.

place at 30–50 ka, and the two younger events seem to have occurred in close succession about 6–8 ka (Bugge *et al.*, 1987, 1988; Jansen *et al.*, 1987). The first event affected almost the whole area of the present slide scar; the second slide cut up to 200 m into the former slide scar and formed a new headwall some 5–8 km further into the shelf; the last slide was the smallest and was confined to the upper part of the second slide. Large areas of the Norway Basin seems to be covered by various types of slide deposits (Bugge *et al.*, 1987; Vogt, 1997).

The Trænadjupet Slide

The existence of the Trænadjupet Slide (Fig. 3) was first indicated by Damuth (1978). Bugge (1983) and Kenyon (1987) found that the headwall is at a water depth of about 300 m and the upper part of the slide covers an area of about 4000 km^2. The sediments were transported 200 km down a slope of about 1° to a depth of 3000 m. The slide deposits include many large blocks, which are up to kilometres wide and have a relief of up to 100 m (Dowdeswell *et al.*, 1996).

The total slide affected area is about 12 000 km^2 (Dowdeswell *et al.*, in 1996).

The Andøya Slide

The Andøya Slide is located within the steepest part of the eastern Norwegian–Greenland Sea continental margin (Fig. 3). It originated on the upper continental slope (Kenyon, 1987) where the present gradient is 5–18°. From GLORIA side-scan sonar records it is apparent that there is a complex of different kinds of slide deposits including three readily identifiable thin, distal flow deposits (Fig. 3). The total slide affected area is about 8000 km^2 (Dowdeswell *et al.*, 1996).

The Bjørnøyrenna Slide

The Bjørnøyrenna Slide, located on the southern flank of the Bear Island Trough Mouth Fan (Fig. 3), was first described by Kristoffersen *et al.* (1978) and later investigated by Bugge (1983), Vorren *et al.* (1989) and Laberg and Vorren (1993). According to Laberg and Vorren (1993), the slide has a run out distance of about 400 km, is up to 400 m deep and a total volume of about 1100 km^3 has been displaced. The boundary between the erosional and depositional area appears to be at a depth of about 2500–2700 m. Kristoffersen *et al.* (1978) suggested an age of about 200 ka and Laberg and Vorren (1996a) found the failure to have occurred sometime between Isotope Stage 8 (313 ka) to Isotope Stage 6 (194 ka). Hemipelagic sediments, in excess of 160 m thick infill the slide scar (Fig. 5), and indicate that the slide occurred prior to the last glacial–interglacial cycle.

Small slides

Several small slides have been identified from side-scan sonar and 3.5-kHz records between the Trænadjupet and Andøya Slides, immediately north of the Andøya Slide and on the continental margin west of Bjørnøya (Bugge, 1983; Kenyon, 1987; Crane and Solheim, 1995; Dowdeswell *et al.*, 1996) (Fig. 3). Little is known about these small slides, which seem to be most frequent along the steepest part of the continental margin.

Buried slides

Recently, Evans *et al.* (1996) and King *et al.* (1996) found indications of two or more earlier sliding events on the North Sea Trough Mouth Fan (TMF: see below for details), probably of Late Cenozoic age. On the Bear Island TMF, Knutsen *et al.* (1992) mapped a Plio–Pleistocene slide covering an area of 12 000 km^2 and involving a total volume of 5100 km^3. A later slide, probably of Mid/Late Pleistocene age, was identified by Laberg and Vorren (1996a). This slide also occurred on the central fan. It covered a comparable area, but seems to have been shallower, thus involving less volume of sediment. Recently, Kuvaas and Kristoffersen (1996) identified at least five separate failure events within the Pliocene and Pleistocene succession, documenting the importance of large-scale mass movements within the Bear Island TMF area.

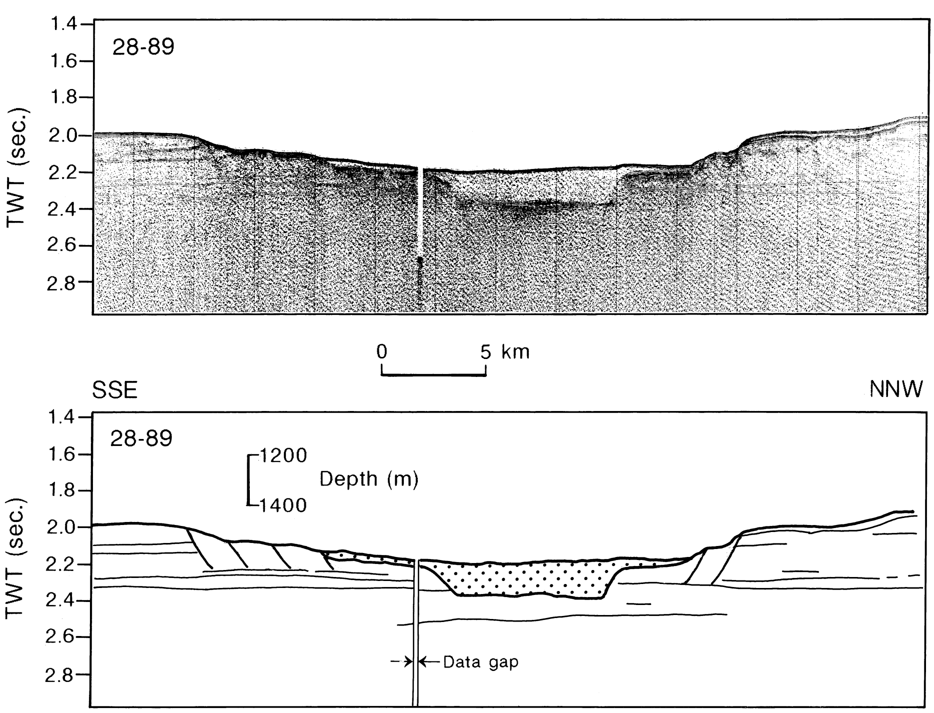

FIG. 5. Seismic sparker profile across the Bjørnøyrenna Slide at about 1500 m water depth showing the infilling of the depression (dotted area) enclosed by a zone of several tilted blocks separated by faults. Location is given in Fig. 7. From Laberg and Vorren (1993).

East Greenland margin

Compared to the Norwegian/Barents Sea continental margin, relatively few detailed studies have so far been undertaken on the East Greenland continental margin. However, much of the East Greenland margin from 70–80° (Mienert *et al.*, 1993, 1995; Pfirman and Kassens, 1995) and 68–70° (Dowdeswell *et al.*, 1997) has been mapped by GLORIA long-range side-scan sonar and PARASOUND/3.5 kHz data (see below). These data reveal no large scale sliding. Thus, within much of the East Greenland continental slope no large-scale sliding events seem to have occurred during Mid/Late Pleistocene time. Mass wasting does, however, seem to have been pronounced on the Southeast Greenland continental margin during the Oligocene (Clift, 1996).

Origin of the slides

Three of the large slide complexes on the Norwegian–Barents Sea margin (the Storegga, Trænadjupet and Bjørnøyrenna Slides) occurred in areas characterised by high sediment input (Fig. 2 and Fig. 3). Thus, relatively high sediment supply leading to unstable sediments may have been important for the initiation of slope failures in these areas. Another important factor may be the presence of shallow gas, suggested by Knutsen *et al.* (1992) as important for the instability of sediments found on the central part of the Bear Island TMF.

Studies of the present seismicity in the eastern Norwegian–Greenland Sea has shown relatively high activity along older fault systems (Kvamme and Hansen, 1989; Bungum *et al.*, 1991). From other continental margin areas, earthquakes have been recognised as one of the most likely triggering mechanisms for submarine slides (e.g. Hampton *et al.*, 1996). Thus, for the triggering of the large slides along the eastern Norwegian–Greenland Sea continental margin, earthquakes have probably been an important mechanism (Bugge, 1983; Bugge *et al.*, 1987; Kenyon, 1987; Knutsen *et al.*, 1992; Laberg and Vorren, 1993, 1996a; Evans *et al.*, 1996).

Decomposition of gas hydrates was also suggested as a possible triggering mechanism for the Storegga Slides (Bugge *et al.*, 1987). Andreassen and Hansen (1995) using a phase boundary diagram for a methane hydrate system infer that gas hydrates may be present under the Norwegian/Barents Sea continental slope today. Increased bottom water temperature and/or lowering of sea level would cause destabilization of the hydrate zone (McIver, 1982).

To summarise, high sedimentation rates, which in turn may have led to a build-up of excess pore water pressure, and perhaps with additional pressure caused by gas bubbles, probably led to unstable or metastable sediments within relatively large parts of the eastern Norwegian–Greenland Sea continental margin. Destabilizing and triggering may have been prompted by earthquakes or perhaps by decomposition of gas hydrates.

TROUGH MOUTH FANS AND DEBRIS FLOWS – GENERAL REMARKS

On the continental margins surrounding the Norwegian–Greenland Sea, fan or delta-like protrusions occur in front of many of the glacial troughs or channels crossing the continental shelf and ending on the shelf break (Fig. 6). Many of these protrusions were noted by Nansen (1904). Vogt and Perry (1978) pointed out that these protrusions are probably prograded deltas and attached fans. Vorren *et al.* (1988, 1989) proposed naming these features, which also occur on other glaciated margins, "trough mouth fans" (TMF). Not all troughs ending on the shelf break have fans at their mouths, but on the eastern margin of the Norwegian–Greenland Sea they are particularly frequent: the North Sea TMF, the Bear Island TMF, the Storfjorden TMF, and smaller TMFs off Svalbard (the Bellsund TMF, the Isfjorden TMF and the Kongsfjorden TMF). On the Greenland continental margin, the Scoresby Sund TMF is well developed (Dowdeswell *et al.*, 1997), but it is in fact difficult to identify other TMFs on this margin from the available bathymetric, seismic and long-range side-scan sonar data. The TMFs vary in size and shape (Table 2). In recent years the fans have been investigated with regard to their architecture, sedimentological processes and origin, and their potential as palaeoclimatic archives.

At least during the late Quaternary, the TMFs have been the sites of intense debris flow activity. Damuth (1978) was the first to indicate their presence. From 3.5-kHz records he noted that large areas of the Bear Island and North Sea TMFs had a characteristic echo type comprising "continuous, sharp bottom echoes with one or two sharp, unconformable, wedging subbottom reflections which generally persist for less than 10–15 km". He suggested they may represent some type of downslope progradation via gravity-controlled mass flows, and that "sediments deposited by the ice on the continental slope and upper rise may have moved further downslope by localised debris flows, turbidity currents or other gravity-controlled bottom flows."

Vorren *et al.* (1988, 1989) using sparker data found that the debris flows, seen in cross section, are bundled in *sets* of lenses separated by high amplitude reflections, and pointed to their potential as a palaeoclimatostratigraphic proxy. They suggested that in most cases each of the lobes in the lens set represents a single debris flow originating on the upper slope, emplaced during periods when the glacier grounding line was near the shelf break. Then "there was probably a very high sediment input all along the trough mouth. This situation caused oversteepened slopes and/or excess pore fluid pressures in the accumulated sediments, creating instability and mass movement. The high amplitude reflections between each lens-set probably

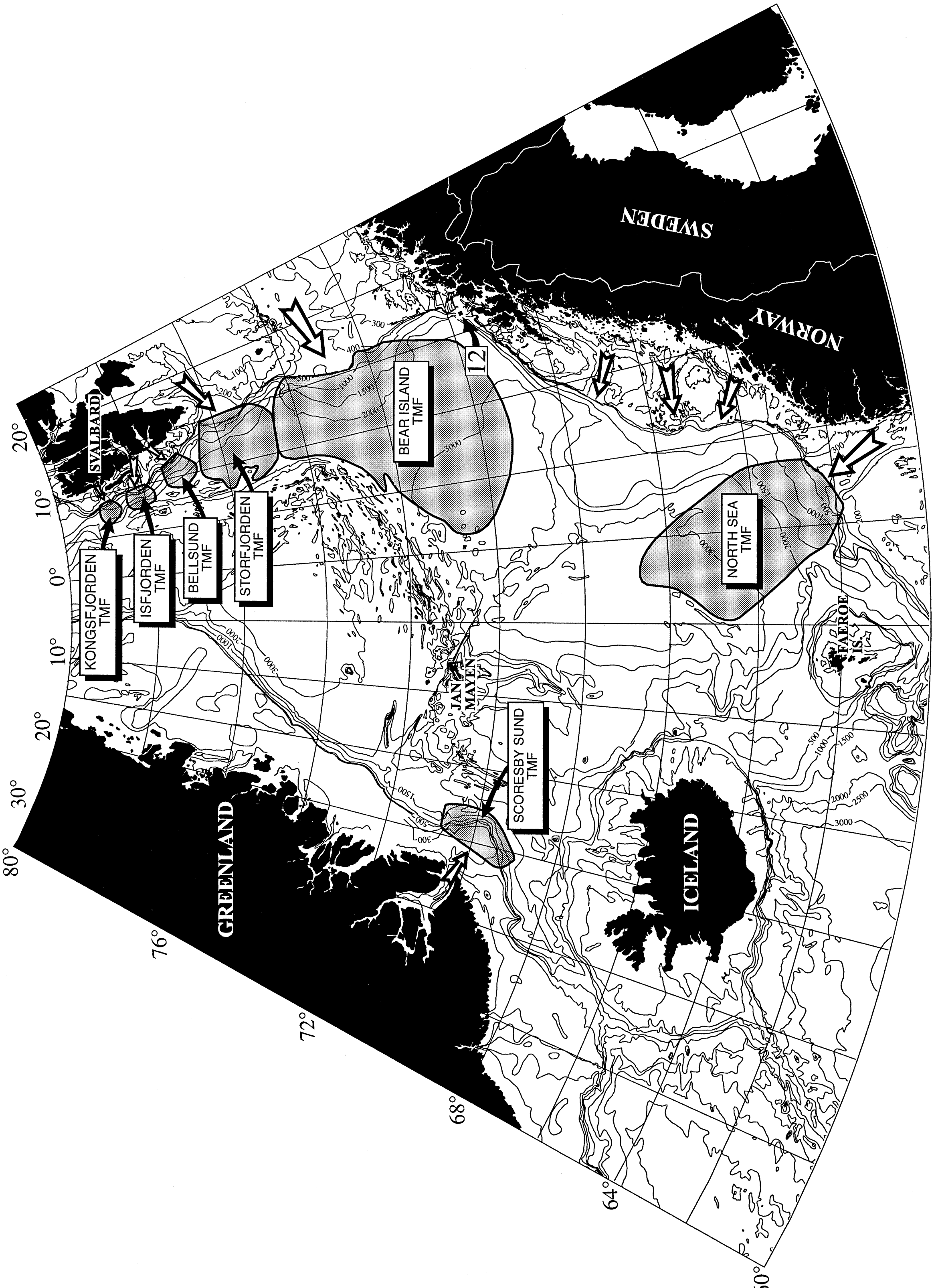

FIG. 6. Bathymetric map of the Norwegian–Greenland Sea continental margin showing location and extent of trough mouth fans. Arrows indicate large shelf troughs.

TABLE 1 Dimensions of large scale slides on the Norwegian continental margin

	Storegga Slide	Trænadjupet Slide	Andøya Slide	Bjørnøyrenna Slide
Run-out distance (km)	850	200		400
Slide scar area (km^2)	34 000			
Total slide influenced area (km^2)	112 500	12 000	8000	12 500
Maximum thickness (m)	430			
Volume (km^3)	5580			1100

Storegga Slide dimensions are from Bugge *et al.* (1988); Bjørnøyrenna Slide dimensions are from Laberg and Vorren (1993); Trænadjupet Slide dimensions are from Bugge (1983) and Dowdeswell *et al.* (1996) and Andøya Slide dimensions are from Dowdeswell *et al.* (1996) and this study.

represent periods of low sediment input/erosion during interstadials or interglacials" (Vorren *et al.*, 1989).

Later mapping by seismic and side-scan sonars has confirmed that the debris flows are the main building blocks, and palaeoclimatic recorders, of the younger part of the TMFs in the Norwegian–Greenland Sea (Vogt *et al.*, 1993; Laberg and Vorren, 1995, 1996a,b; King *et al.*, 1996; Sejrup *et al.*, 1996; Dowdeswell *et al.*, 1996, 1997). Similar flows have been reported from the continental margin off Northwest Britain (Stoker, 1995) and the eastern Canadian continental margin (Aksu and Hiscott, 1992; Hiscott and Aksu, 1994).

The trough mouth fans along the eastern Norwegian–Greenland Sea continental margin south of Svalbard all seem to have acquired their modern shape later than the early Mid-Pleistocene. This indicates that the north European ice sheets (i.e. The British Ice Cap, the Fennoscandian Ice Sheet and the (southern) Barents Sea Ice Sheet), did not extend to the shelf-break for any appreciable time before the Mid Pleistocene. This conforms with deep-sea as well as terrestrial data which show that the Mid and Late Pleistocene was the time of the largest glaciations in northern Europe (Vorren and Laberg, 1997).

MORPHOLOGY AND STRATIGRAPHY OF THE TROUGH MOUTH FANS

The North Sea TMF

Morphology

The North Sea TMF is deposited in front of the Norwegian Channel (Fig. 6), and comprises a thick wedge extending into the Norway Basin. A N–S axial dip line along the fan body shows a uniform dip of about 0.7° flattening out to about 0.3° in the distal fan area (King *et al.*, 1996).

Stratigraphy

According to King *et al.* (1996), the fan stratigraphy indicates two main depositional phases. The early phase, of which a significant part may be preglacial is represented by steeply dipping (1–2.5°) progradational strata comprising much of the most proximal fan, together with correspondingly thick uniform deposits extending to the Faeroe–Shetland Escarpment. The late phase comprising an overlying wedge, extends beyond the escarpment exhibiting overall slopes less than 0.5° and is of Quaternary age and mainly of glacigenic origin.

The Quaternary part of the fan was subdivided by King *et al.* (1996) into 12 seismic sequences. The seismic sequences of the fan body are interpreted as comprising three major depositional styles, namely terrigenous hemipelagic, debris flow aprons and disturbed slide sediments. The glacially-related debris flow aprons represent five of the seven uppermost sequences. A middle Pleistocene age is suggested for the first debris flow apron (sequence 7) and a (Late) Weichselian age for the youngest debris flow apron (sequence 1).

The Bear Island TMF

Morphology

The Bear Island TMF is situated in front of the Bear Island Trough (Fig. 6). This fan extends from the shelf break to depths greater than 3000 m, where it merges with the abyssal plain of the Lofoten Basin. To the west and Northwest the fan is bounded by the Mohns and Knipovich spreading ridges (Fig. 1). The Bear Island TMF has a width of about 250 km at its proximal part and approximately 440 km at 2200 m water depth. It covers an area of approximately 215 000 km^2 (Table 2). The morphology of the fan is influenced by the Bjørnøyrenna Slide scar on its southern flank (Fig. 1). Based on slope gradient and seismic facies distribution, Laberg and Vorren (1995) subdivided the fan into three morphological zones. At about 1500 m a minor break in slope occurs which separates the upper from the middle fan. The upper fan extends from the shelf break, the average depth of which is 500 m, and has a slope gradient of about 0.8° along the central fan axis. The surface of the upper fan is generally hummocky. The surface of the middle fan, between 1500 and 2200 m, is both hummocky and ridged on the northern and central fan, whilst smooth on the southern part. The average gradients of the mid- and lower fan are about 0.5° and 0.2°, respectively. The slightly convex-upward surface of the lower fan is uneven on the northern fan flank and smooth near the slide scar.

TABLE 2 Dimensions and slope gradients of through mouth fans along the Norwegian–Greenland Sea continental margin

	Kongsfjorden TMF	Isfjorden TMF	Bellsund TMF	Storfjorden TMF	Bear Island TMF	North Sea TMF	Scoresby Sund TMF
Radius (km)	55	50	70	190	590	560	110
Width upper (km)	40	45	55	130	250	165	180
Width lower (km)	60	75	85	210	550	300	240
Depth upper (km)	0.2	0.25	0.15	0.4	0.5	0.4	0.3
Depth lower proximal (km)				2.4	3.0	2.7	1.5
Depth lower distal (km)	2.0	3.0	2.3	2.7	3.2	3.5	1.5
Area (km^2)	2700	3700	6000	35 000	215 000	142 000	19 000
Gradient (upper)				1.8°	0.8°	0.6°	
Gradient (middle)	1.9°	3.2°	1.8°	1.0°	0.4°	0.8°	2°
Gradient (lower)				0.2°	0.2°	0.3°	

Radius = radius along the longest axis; Width upper = width at shelf break; Width lower = maximum width of the lower fan; Depth upper = depth at the shelf break; Depth lower proximal = depth at the base of the proximal part of the fan; Depth lower distal = depth of the base of the distal part of the fan; Area = total fan area including shelf and distal areas (except for the Scoresby Sund TMF the shelf area constitutes a negligible part of the fan area and has not been included). Gradients of the upper, middle, and lower slope along the longest fan axis of the Storfjorden and Bear Island TMF according to Laberg and Vorren (1996b) and from King *et al.* (1996) for the North Sea TMF. Average gradients of the whole fan slopes are given for the Kongsfjord TMF, the Isfjord TMF, the Bellsund TMF, and the Scoresby Sund TMF.

Stratigraphy

The overall seismic stratigraphy of the clastic wedge of sediments comprising the fan has been studied by many workers (Spencer *et al*., 1984; Richardsen *et al*., 1991, 1992; Vorren *et al*., 1991; Knutsen *et al*., 1992; Eidvin *et al*., 1993; Sættem *et al*., 1994; Faleide *et al*., 1996) (Table 3). The age of the fan progradation and also the origin of the sediment comprising the fan is disputed. Recent dating suggests that most of the fan sediments are of (Late) Pliocene–Pleistocene age (Eidvin *et al*., 1993; Sættem *et al*., 1994; Faleide *et al*., 1996).

Much of the sediment in the fan contains clasts of glacial origin (Eidvin *et al*., 1993; Sættem *et al*., 1994). The origin of the sediments comprising the lower part of the fan has to a large degree been a matter of discussion about the extent of the sediment-feeding glaciers through time. Many favour extensive glaciations of the Barents Sea throughout the deposition of the fan/clastic wedge, e.g. "… it is likely that the shelf was covered by ice sheets during a major part of this period" (Eidvin *et al*., 1993), "We argue that glaciers reached the position of the westernmost Barents shelf already at the very earliest stage of the deposition of the clastic wedge in front of Bjørnørenna" (Sættem *et al*., 1992). Vorren *et al*. (1991) favoured a gradually increasing glacial dominance, culminating with total glaciation of the Barents Sea shelf during the deposition of the youngest sequence (i.e. unit TeE) only. Sættem *et al*. (1994) also indicate, in accordance with Vorren *et al*. (1991), that the earliest glaciations were centred on the Svalbard area. However, Sættem *et al*. (1994) indicate that glaciers reached the shelf break in the Bear Island Trough, from the north, at the very earliest time of the clastic wedge deposition.

In this work we will concentrate on the upper seismostratigraphic unit, corresponding to Unit IV of Spencer *et al*. (1984), Unit TeE of Vorren *et al*. (1991), and Unit G III of Faleide *et al*. (1996) (Table 3) and also corresponding to the following units identified on the continental shelf and uppermost slope: namely units I to IV of Solheim and Kristoffersen (1984); units 1W to 4W of Vorren *et al*. (1990) and units B to G of Sættem *et al*. (1992). There is a general consensus that this upper part of the clastic wedge/fan was deposited during a period when the whole of the Barents Sea shelf underwent repeated glaciations.

Eight seismostratigraphic units, I (oldest) to VIII were identified by Laberg and Vorren (1996a) in this upper part of the clastic wedge. All these units were dominated by debris flows assumed to be deposited during glacials (Fig. 7). The reflections separating these units are interpreted to be formed during interglacials/interstadials (see below).

Age estimates

The Holocene interglacial sedimentation rate has been about 15 cm/1000 years within the uppermost part of slide scar C (Laberg and Vorren, 1993). This is relatively low compared to the up to 124 cm/1000 years inferred during the Late Weichselian glacial maxima (Laberg and Vorren, 1996a). Hence the key reflections which separate the seismic units were inferred to reflect palaeo-fan surfaces formed during sediment-starved interstadials and/or interglacials. The seismic correlation of the debris flow units (I–VIII) to the glacigenic sedimentary units on the continental shelf, shows that the diamicton in the debris flows was deposited during glacials (Laberg and Vorren, 1996a).

Based on the assumption that: (1) all the identified seismic units on the fan comprise mainly sediments deposited during glacial maxima; (2) the Barents Sea Ice Sheet advanced to the shelf break during each Middle and Late Pleistocene glacial maximum; and (3) the Middle and Late Pleistocene fan stratigraphic record is complete, then 'back-counting' from the last glacial maximum (Stage 2) indicates that seismo-stratigraphic

TABLE 3 Suggested ages and correlation of seismostratigraphic units of the western Barents Sea and Svalbard continental margin

Barents Sea margin (70–74°N)				Barents Sea margin (70–77°N)	Svalbard margin (77–80°N)	
Spencer *et al.* (1984)	Eidvin and Riis (1989); Eidvin *et al.* (1993)	Vorren *et al.* (1991); Knutsen *et al.* (1992); Richardsen *et al.* (1992)	Sættem *et al.* (1992, 1994)	Faleide *et al.* (1996); Fiedler and Faleide (1996); Hjelstuen *et al.* (1996)	Schlüter and Hinz (1978); Myhre and Eldholm (1988)	Andersen *et al.* (1994)
Plio/ Pleisto IV		TeE	B - G	G III	SP I	
	Reflector 1	0,8	< 0.44	0.44		0.15-0.3
III B Oligo l. Plio	Late Pliocene/ Pleistocene	TeD	A Late Pliocene/ Pleistocene	G II		A
						5.5
						B
	Reflector 2	3.0 —(5.5)		1.0		
III A		TeC	Ao	G I	SP II	
	Reflector 3	15.5		2.3	5.5	
Late Pale. Oligo I-IIA-IIB	55	TeA-TeB		G 0	SP III	

unit VI was deposited mainly during Isotope Stage 6, V during Stage 8, IV during Stage 10, III during Stage 12, II during Stage 14 and seismostratigraphic unit I during Isotope Stage 16 (Fig. 8). This correlation indicates that the studied sediments on the fan are younger than 622 ka (Laberg and Vorren, 1996a). This age estimate is somewhat younger than the 0.8 Ma suggestion of Vorren *et al.* (1990), and somewhat older than the maximum 440 ka amino acid dating of Sættem *et al.* (1992). By using the isotope time scale, the interval during which the individual seismo-stratigraphic units were deposited can be bracketed more precisely (Fig. 8).

However, the suggested ages may be maximum ages, because more than one unit can be deposited during each glacial maximum, as exemplified by units VII and VIII. On the other hand, some of the isotope glacial maxima may not be represented by an ice sheet extending to the shelf break. Therefore, it must be stressed that the suggested datings should be considered as a first approximation for the seismic units (Laberg and Vorren, 1996a).

The Trough Mouth Fans as proxies for ice stream velocities and ice sheet mass balance

Vorren and Laberg (1996) estimated the velocities of the Bear Island Ice Stream as well as the iceberg calving flux and the snow accumulation of the drainage area of this ice stream during the last glacial maximum (18–15 ka). Their estimate was based on the measured amount of sediments in the debris flows transported to the Bear Island TMF (1500 km^3), and estimates and assumptions on the origin of the sediments (deformation till), thickness of the deforming basal till layer (4 m), duration of input (3 ka), and width (150 km) and thickness (500 m) of the feeding ice stream. They concluded that the speed of the basal ice was 2.5 km/year at the front and that the ice discharge to the ocean was about 200 km^3/year.

When more data are available one could, by using the same type of reasoning, estimate the ice velocities and calving flux of other ice streams feeding trough mouth fans during the last glacial maximum and possibly older glacials.

THE STORFJORDEN TMF

Morphology

The Storfjorden TMF which covers an area of about 40 000 km^2 and has a radius of about 190 km, developed concentrically off the Storfjorden trough (Fig. 6). The water depth is mostly between 400 and 2500 m. The fan extends from the shelf break westward to the damming Knipovich Ridge. The upper and middle fan gradients along the central fan axis are 1.8° and 1.0°, respectively, while the lower fan gradient is about 0.2° or less (Laberg and Vorren, 1996b).

Stratigraphy

The Storfjorden TMF is an integral part of the Cenozoic sediment wedge which has been deposited at the western Barents Sea margin (Myhre and Eldholm, 1988; Vorren *et al.*, 1991). The Cenozoic sediment in the Storfjorden TMF area has been divided into three

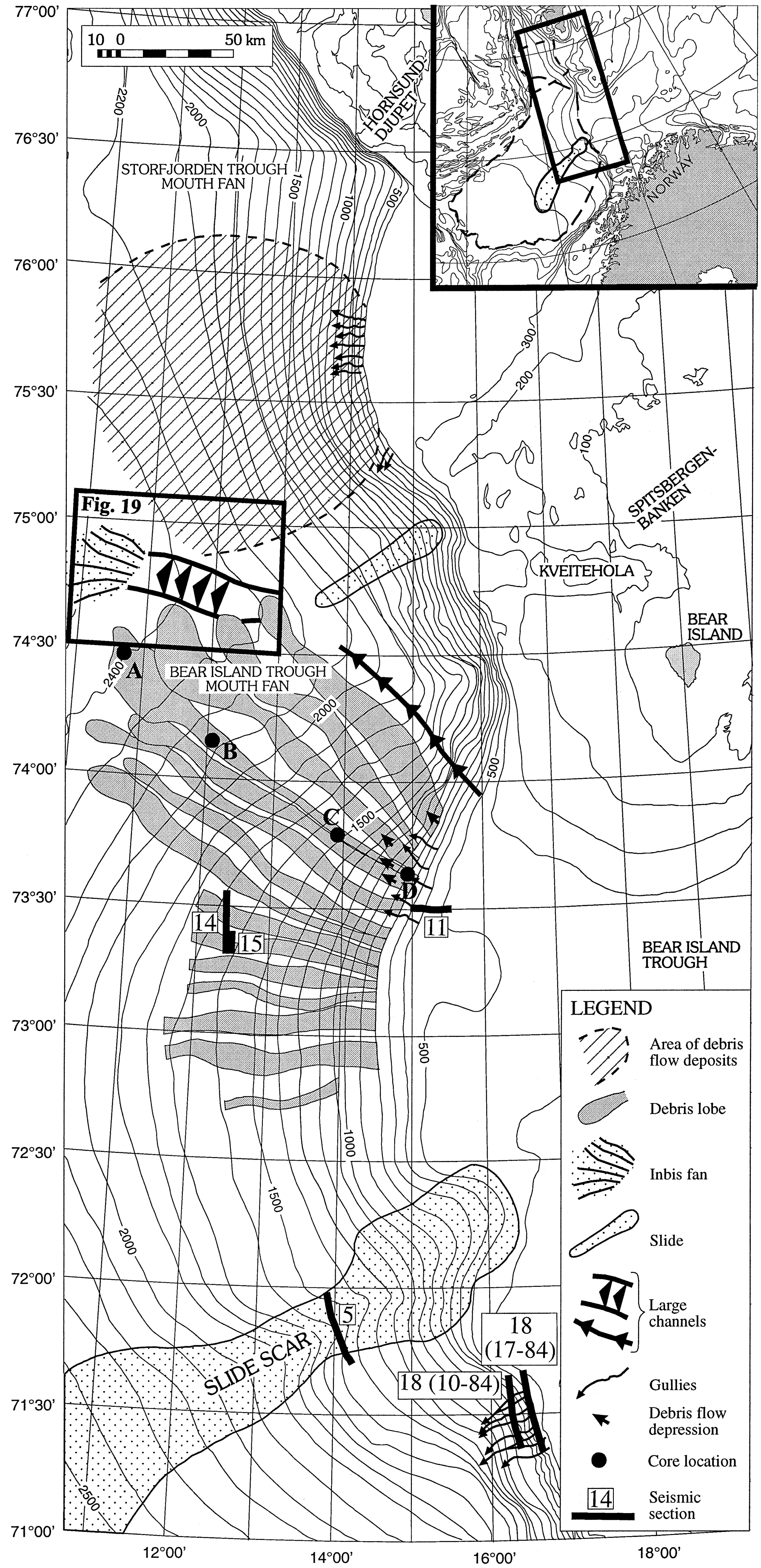

FIG. 7. Map of the Bear Island TMF and the Storfjorden TMF indicating the extent of the youngest debris lobes and location of large channels and gullies. Location of seismic profiles (Fig. 5, Fig. 11, Fig. 14, Fig. 15 and Fig. 18) are given. Positions of gravity cores (Fig. 10) are given by heavy dots. Bold frame outlines Fig. 19.

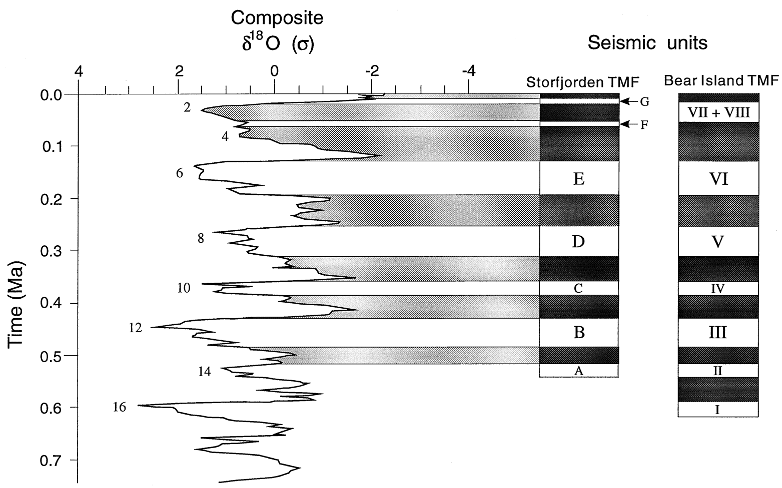

FIG. 8. Tentative correlation of seismic (debris flow) units from the Bear Island and Storfjord TMFs to the oxygen isotope stratigraphy (Williams *et al.*, 1988). Isotope Stages 2–16 are indicated. Dating of units E, F and G are based on correlation to the Svalbard land record (Mangerud and Svendsen, 1992). The figure has been adapted from Laberg and Vorren (1996b).

seismic sequences by Schlüter and Hinz (1978), SPI-I, II and III, and four sequences by Hjelstuen *et al.* (1996), G0 (oldest), GI, II and III. In this study, we will focus on the youngest part of seismic sequence SPI-I, which comprises sequence GIII and the youngest part of GII. This part of the fan sequence is dominated by debris flows, and tentatively dated to be of Mid and Late Quaternary age (Laberg and Vorren, 1996b).

Seven seismic units have been identified within the study area of the Storfjorden TMF; A (oldest) to G (Laberg and Vorren, 1996b). The units are bounded by regional key seismic reflections; a (oldest) to g, of which reflection b correlates to reflection R1 of Hjelstuen *et al.* (1996). However, due to the data coverage and penetration, some of the deepest reflections (a–d) could not be traced with confidence on the middle and lower fan, preventing a precise correlation in these areas.

Age estimate

There are no published dates from the studied part of the Storfjorden TMF or the nearby shelf. The only results available are from the DSDP Site 344 which was located on the distal part of the fan, and which showed that sequence SPI-I of Schlüter and Hinz (1978) comprises sediments of Pliocene–Pleistocene age (Talwani *et al.*, 1976). By seismic correlation to commercial wells in the southwestern Barents Sea, Hjelstuen *et al.* (1996) propose that their sequence GIII comprises sediments younger than 440 ka.

Correlation with the Svalbard land record led Laberg and Vorren (1996b) to suggest that the three major glaciations known to have extended beyond the coast of western Svalbard were probably mirrored by ice advances through the Storfjorden Trough from the southeastern part of Svalbard. Based on the above considerations, and the dating results of Mangerud and Svendsen (1992), Laberg and Vorren (1996b) tentatively suggest that there may have been two ice advances to the shelf break in the outer Storfjorden Trough during the Weichselian; in Middle (65–55 ka) and Late Weichselian (20–10 ka) respectively, and one advance during the Late Saalian. During these ice advances, units E (Late Saalian), F (Mid Weichselian) and G (Late Weichselian) may have been deposited.

As a first approximation, the older units (A–D) were tentatively correlated to the oxygen isotope curve (Fig. 8) (Williams *et al.*, 1988). Based on this correlation, Laberg and Vorren (1996b) suggest that seismic unit D can be correlated to Isotope Stage 8, unit C to Stage 10, unit B to Stage 12 and seismic unit A to Stage 14. Hence the studied succession is found to comprise sediments younger than about 600 ka. In this correlation we have made similar assumptions as for the dating of the Bear Island TMF (see above).

The Svalbard TMFs

Morphology

Along the western shelf-margin of Svalbard three trough mouth fans can be discerned on bathymetric maps, Bellsund TMF off Bellsund, Isfjorden TMF off Isfjorden and Kongsfjorden TMF off Kongsfjord (Fig. 6). The three fans are of much the same size and shape (Table 2), and partly merge in their distal areas (Andersen *et al.*, 1994). These fans cover an area of about 4000 km^2 each, have radii of 50–70 km and relatively steep slopes, i.e. on average between 1.8° and 3.2° (Table 2).

Stratigraphy

Schlüter and Hinz (1978) recognised three sequences off western Svalbard, SPI-I, II and III, separated bytwo unconformities, U1 and U2. The U2-unconformity, separating the oldest SPI-III from the SPI-II, has been taken to correspond to the onset of glacigenic sedimentation (Myhre and Eldholm, 1988). Andersen *et al.*

(1994) have indicated that the Late Cenozoic succession consists of two primary sequences (A and B) which are separated by a regional unconformity, R4. Sequence A corresponding to the youngest part of seismic sequence SPI-I, and is responsible for the fan growth. Andersen *et al.* (1994) suggest that this took place at about 1–0.8 Ma ago and that this marks the advent of more extensive glaciations.

The Scoresby Sund TMF

Morphology

On the East Greenland continental margin between 69° and 71°N, the presence of a major submarine fan offshore of the 14 000 km^2 Scoresby Sund fjord system is indicated by the crescent-shape of the shelf break and the contours of the continental slope (Dowdeswell *et al.*, 1997) (Fig. 6). The fan has a gradient of about 1.8–2° and terminates in about 1500 m of water to the west of the Kolbeinsey Ridge. The glacier-influenced fan volume is about 15 000 $\pm$ 5000 km^3 (Dowdeswell *et al.*, 1997), based on seismic reflection studies (Larsen, 1990; Vanneste *et al.*, 1995). The seismic evidence suggests a smaller total volume for underlying non-glacial sediments of possible riverine origin (Vanneste *et al.*, 1995). Thus, the total volume of the Scoresby Sund TMF is unlikely to exceed 30 000 $\pm$ 10 000 km^3.

Stratigraphy

A seismic reflection profile across the southern part of the Scoresby Sund TMF, published by Vanneste *et al.* (1995), together with a west to east profile from just north of the mouth of Scoresby Sund (Larsen, 1990), provide information on fan thickness and internal structure. Larsen (1990) also gives a total isopach map of sediment thickness above basement for most of the fan and adjacent basin. There are three major units above oceanic basement rocks, with a total sediment thickness of up to about 2000 m (Fig. 9; Vanneste *et al.*, 1995). However, the lower units, I and II in Fig. 9, are interpreted by Vanneste *et al.* (1995) to be of preglacial origin and to be related at least in part to riverine processes. The uppermost unit (III) is on the order of 275 m thick on the prograding continental shelf, and up to 1000 m in thickness on the outer shelf and upper slope, thinning to less than 140 m on the Iceland Plateau (Fig. 9). The base of Unit III is represented by a clearly identified erosion surface on the continental shelf, and this is interpreted to indicate the first ice-sheet advance across the shelf. Unit III also has five erosional unconformities within it, probably related to further ice advances, and the entire unit is inferred to be of predominantly glacial origin. The age of the transition from preglacial to glacial sedimentation at the base of Unit III is still in doubt, but it may correspond to the inception of glaciation reaching to sea-level in East Greenland at about 7 million years ago (Larsen *et al.*, 1994).

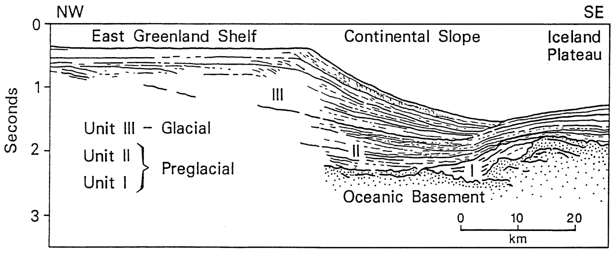

FIG. 9. Interpretation of a seismic reflection profile (100 km in length) across the East Greenland continental margin on the southern part of the Scoresby Sund TMF (Dowdeswell *et al.*, 1997; redrawn from Vanneste *et al.*, 1995). The profile was acquired by Vanneste *et al.* during RV 'Polarstern' cruise ARKVII/3b.

Sedimentary patterns and processes on the Scoresby Sund TMF

The character and spatial distribution of acoustic facies provide a view, in plan from GLORIA imagery, and in section from 3.5-kHz records, of Late Pleistocene sedimentation on the Scoresby Sund TMF and in the adjacent ocean basin truncated by the Kolbeinsey Ridge and Iceland Plateau to the east (Dowdeswell *et al.*, 1997). A series of acoustically transparent features, with irregular surface topography and elongate downslope, are interpreted as debris flows of varying width (0.5–2 km) and thickness (5–15 m). This acoustic facies makes up the bulk of relatively recent sedimentation on the upper fan, and marks the area of most recent active fan growth. However, on the northernmost one-third of the fan, a draping unit is present close to the surface. This is underlain by irregular topography, similar in form to the debris flow described above. This northern region is interpreted to represent the least active area of the Late Pleistocene fan, in which predominantly hemipelagic sediments with a contribution from ice-rafted debris (Nam *et al.*, 1995) overlie older debris-flow deposits (Dowdeswell *et al.*, 1997). These debris flows, interpreted from GLORIA and 3.5-kHz data, are the basic building blocks in the long-term development of the Scoresby Sund TMF. The more distal area of the ocean basin has a flat floor with parallel sub-bottom reflections. This facies is interpreted as an area of predominantly low-energy hemipelagic sedimentation, punctuated by occasional ice rafting and turbidity current activity which serves to produce the internal reflections within the records (Dowdeswell *et al.*, 1997). It lies beyond a break of slope which presumably represents the toe of the prograding Scoresby Sund TMF and the limit of debris-flow activity. It is possible that the debris flows active on the upper slope are related to, and in some cases trigger, turbidity currents which travel beyond this topographic and acoustic stratigraphic boundary into the most distal parts of the basin (Hampton, 1972). During full glacials in East Greenland, the inland ice sheet advanced to fill the Scoresby Sund fjord system and extended across the shelf to reach the shelf break in some glacial cycles (Dowdeswell *et al.*, 1994; Funder *et* *al.*, 1994). Debris flows formed in areas of most rapid sediment flux, which were probably related to the delivery of glacial sediments to the ice margin in an active deforming subglacial layer. The Scoresby Sund TMF is relatively similar to the Storfjorden TMF on the eastern Polar North Atlantic margin, but differs from the larger Bear Island TMF in having a steeper fan gradient, much smaller debris flows and no large-scale slides.

DEBRIS FLOWS — DIMENSIONS AND COMPOSITIONS

Dimensions

The dimensions of the debris flows vary from fan to fan. The width varies between 0.5 and 40 km, thickness between 5 and 60 m, length from less than 10 km up to 200 km, the areas covered are up to 1880 km^2 and the volumes from 0.5 to 50 km^3 (Table 4). There is a clear tendency towards larger fans having the larger and more voluminous debris flows (Table 4).

Lithology

King *et al.* (1996) report that the debris flow material of the North Sea TMF is an extremely homogenous diamicton with 38% clay, 34% silt, 28% sand and usually less than one percent greater than 1 mm in diameter. It is remarkably consistent in texture both with depth and across the fan.

From the Isfjorden TMF, Elverhøi *et al.* (1997) report that the debris flow sediments comprise 35–40% clay, 35–40% silt and 20–30% sand.

Four cores were recovered from one specific Bear Island TMF-debris lobe at different depths (Fig. 10). A massive dark grey diamicton was found to represent the basal unit in all cores (Laberg and Vorren, 1995). The diamicton has a sharp upper boundary. The sediment is poorly sorted and has high clast content, derived from both crystalline and sedimentary rocks. The grain-size distribution of the diamicton is 30–55% clay, 30–50% silt, 10–30% sand and 1–10% gravel. There is no systematic variation in grain-size distribution between the upper and lower fan. The undrained shear strength of the diamicton increases downcore; values between 3.3–16 kPa have been measured. Water contents between 19 and 30% (of wet weight) have been found in most of the unit comprising debris flow sediments in the cores, increasing to 30–35% at the top of the unit. The grain-size distribution and water content are compatible with the youngest glacigenic diamicton on the outer Barents Sea shelf (Sættem *et al.*, 1992; Poole *et al.*, 1994). Foraminiferal analysis of core JM93-6/1 (Fig. 10) shows that the benthic fauna is dominated by Arctic species like *Cassidulina teretis* and *Cassidulina reniforme*. A shelf provenance of the sediments is indicated by species related to shelf and upper slope environment (e.g. *Elphidum excavatum f. clavata*) being very frequent (Laberg and Vorren, 1995). The frequent appearance of the planktic foraminifer *Neogloboquadrina pacyderma* (*sin.*) indicates that some hemipelagic sediments were also entrained into the debris flow deposits.

Seismic signature

Here we will describe and interpret the seismic signature as revealed on the Bear Island and Storfjorden TMFs. Published information from the North Sea and Isfjorden TMFs appear to be quite similar to this.

Key seismic reflections

Key seismic reflections separate the identified seismostratigraphic units. The reflections have medium to high amplitude. On the upper part of the fan, they are partly discontinuous, whereas they are continuous further downslope. The reflections define the base and top of units dominated by debris flows, but also comprise channel erosion as well as slide deposits. By analogy to the present interglacial, the key seismic reflections probably represent the presence of thin inter-bedded units of hemipelagic interglacial/interstadial sediments.

Shelf break–upper fan

The upper part of the fans have a chaotic seismic signature indicating reworking by sliding activity. In some places remains of slide deposits can be identified (Fig. 11), while other places show a prograding delta-like feature having more gently dipping foresets than the general slope gradient (Fig. 12).

The middle part of the fans are dominated by a mounded signature in cross section (Fig. 13 and

TABLE 4 Dimensions of individual debris flows on trough mouth fans in the Norwegian–Greenland Sea

	Isfjorden TMF	Storfjorden TMF	Bear Island TMF	North Sea TMF	Scoresby Sund TMF
Width (km)	2–5	1–5	3–24	2–40, (5)	0.5–2
Thickness (m)	10–30	15	5–50	< 60, (15–30)	5–15
Length (km)	10–20	50–100	100–200		
Area (km^2)	< 100	< 500	850–1880		
Volume (km^3)	0.5–1	3–8	10–32	10–50	

Data from this work and King *et al.* (1996) and Elverhøi *et al.* (1997) (North Sea TMF); Laberg and Vorren (1995) (Bear Island TMF); Laberg and Vorren (1996b) (Storfjorden TMF); Elverhøi *et al.* (1997) (Isfjorden TMF) and Dowdeswell *et al.* (1997) (Scoresby Sund TMF).

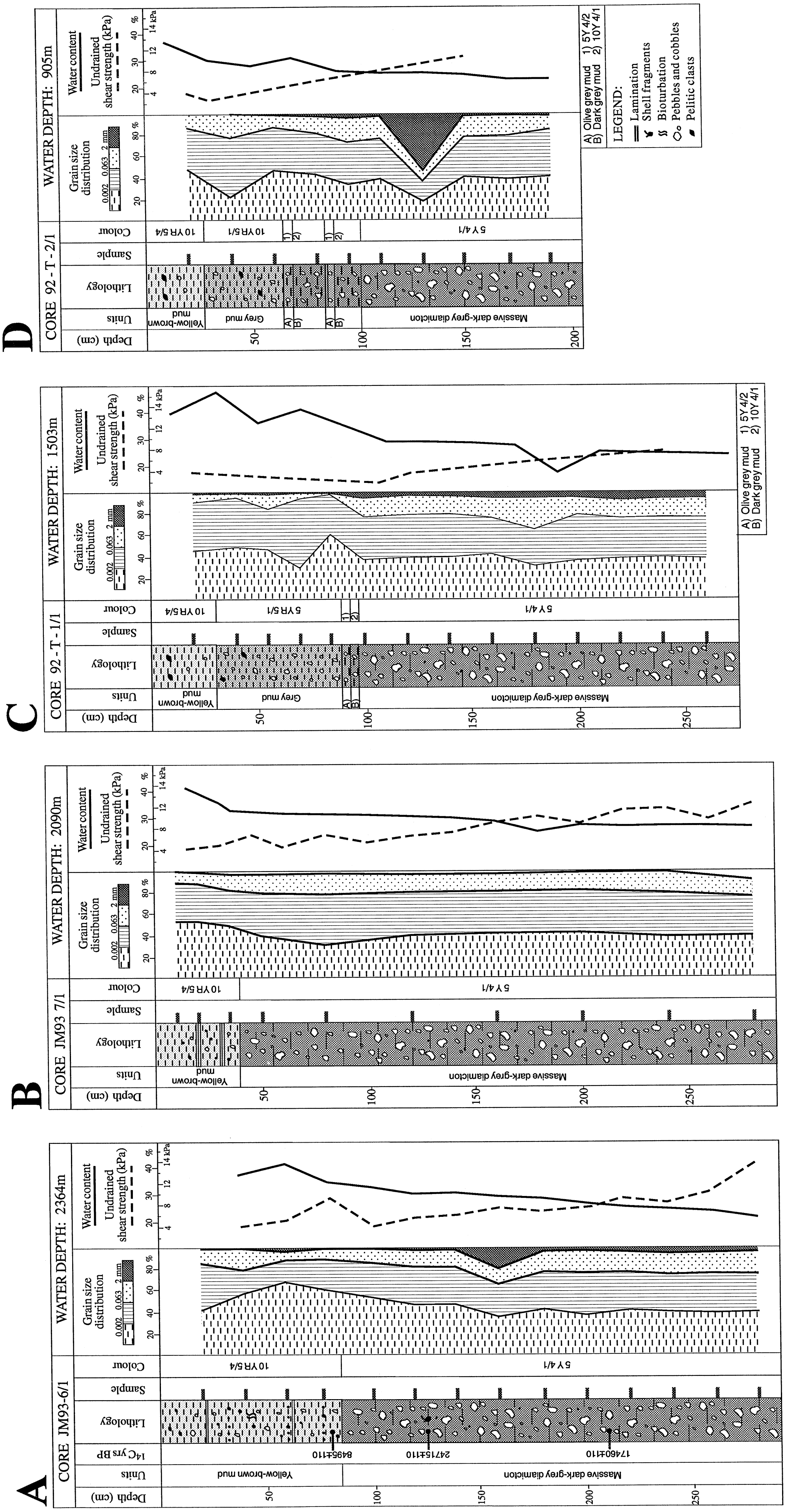

FIG. 10. Stratigraphy and lithology of four gravity cores recovered along a depth transect of one debris lobe on the Bear Island TMF (see Fig. 7 for location). From Laberg and Vorren (1995).

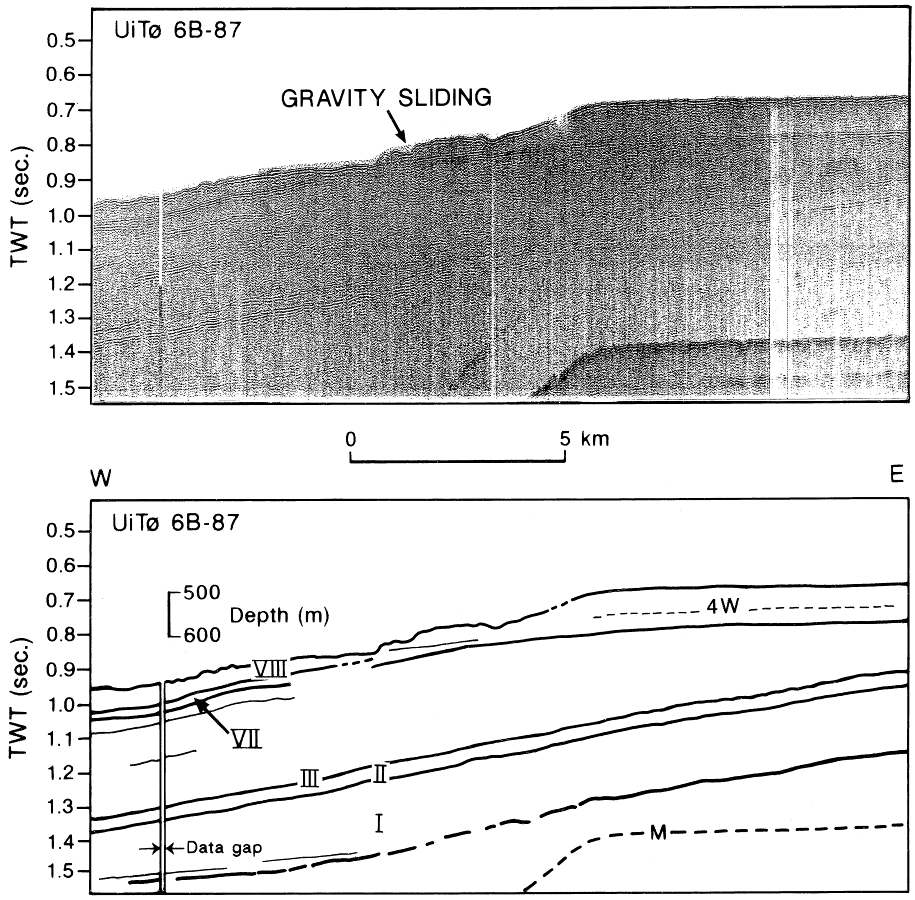

FIG. 11. Part of sparker profile UiTø 6B-87 across the shelf break of the Bear Island TMF (see Fig. 7 for location). Gravity sliding just below the shelf break is evident. Units I, II, III, VII and VIII refer to the seismic units of Laberg and Vorren (1996a) and unit 4W to the seismic units of Vorren *et al.* (1990). M = multiple (slightly modified from Vorren *et al.*, 1990).

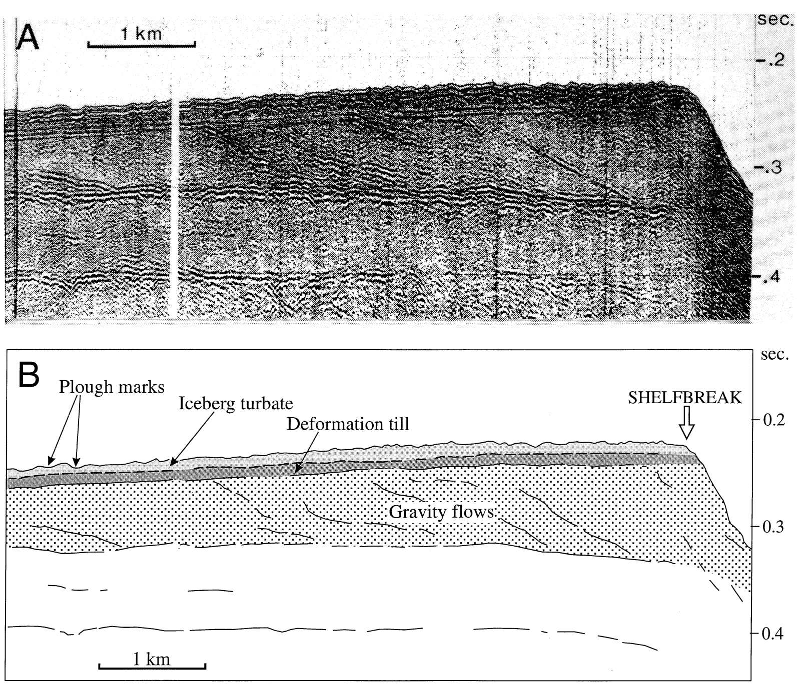

FIG. 12. Sparker profile along Andfjorden, a transverse trough north of Andøya, north Norway (see Fig. 6 for location). The sparker-record indicates a prograding delta-like unit which ends at the steep shelf break to the right. We suggest that this feature may be a till delta. In this area the upper slope is so steep that sediments transported to the shelf break probably were directly transported further downslope by gravity flows (modified from Vorren *et al.*, 1984).

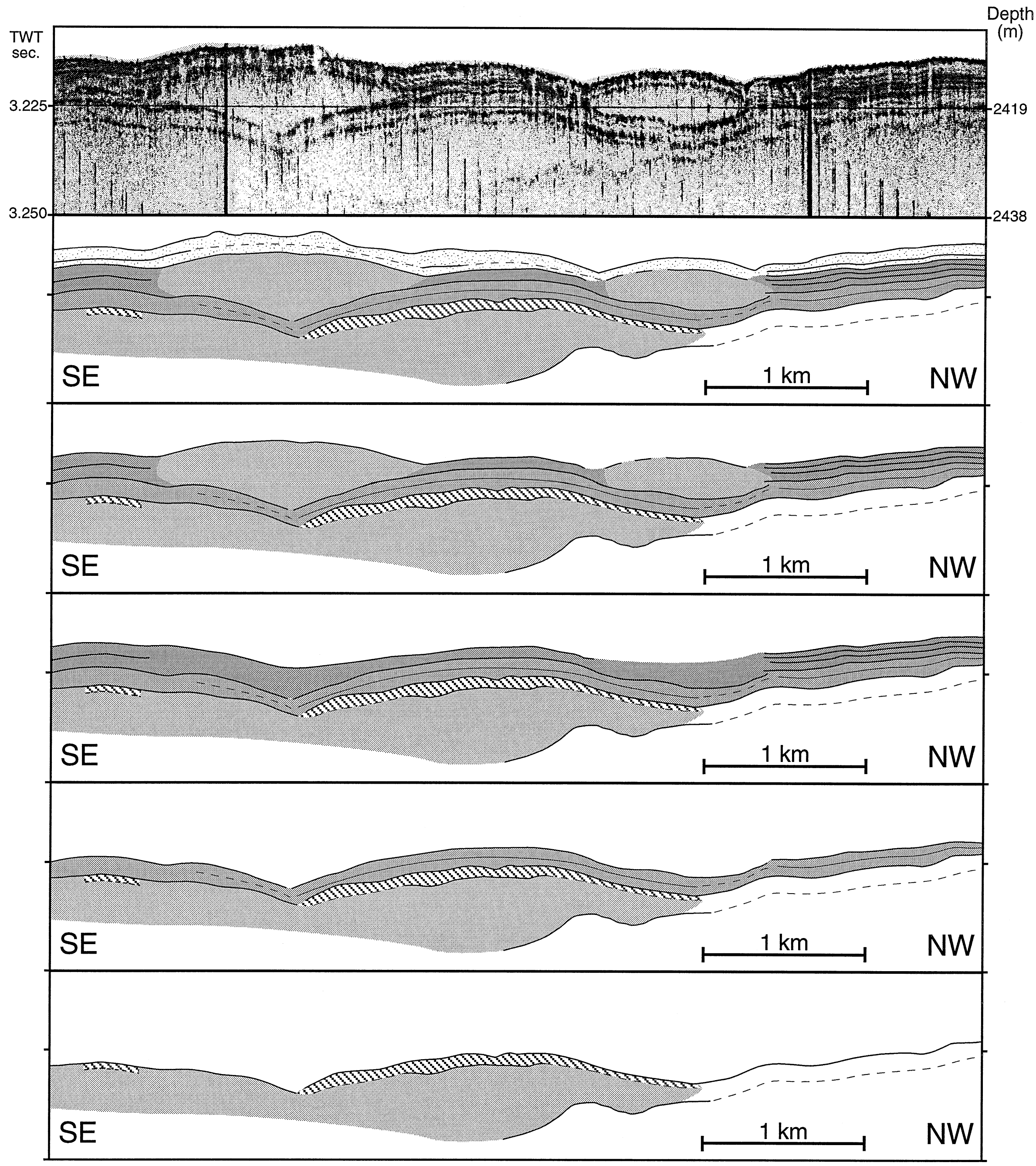

FIG. 13. Cross profile ME72-31 (3.5 kHz) of debris lobes on the lower slope of the Storfjorden TMF (see Fig. 19 for location). Vertical exaggeration is approximately 1:50. The profile was acquired during RV 'METEOR' cruise 7-2, 1988.

Fig. 14), representing sections through debris flows. There is a central depression along the top of some of the debris flows (Vogt *et al*., 1993; Laberg and Vorren, 1995). As revealed from seismic profiles, the depressions are up to 250 m wide and 1.5 m deep (Fig. 15). The debris flows are deposited into bathymetric lows between older deposits. The flows end on the lower fan. The snouts of the debris flows are mostly very well delineated, as can be observed both from seismic (Fig. 13) and side-scan sonar data (Vogt *et al*., 1993) (Fig. 16). The distinct downslope margin of the debris lobes indicate a sudden halt on the lower fan.

The base reflections are not commonly found on the upslope part of the debris lobes. This, together with the fact that the regional reflections immediately below the mounds have in some places an irregular character, indicates disturbance and probably incorporation of underlying sediments during downslope flow. Aksu and Hiscott (1992) and Stoker (1995), investigating debris flows on the Newfoundland continental slope and Hebrides slope, respectively, also found that the basal reflection of a small number of lenses exhibited down-cutting into the underlying acoustically stratified unit.

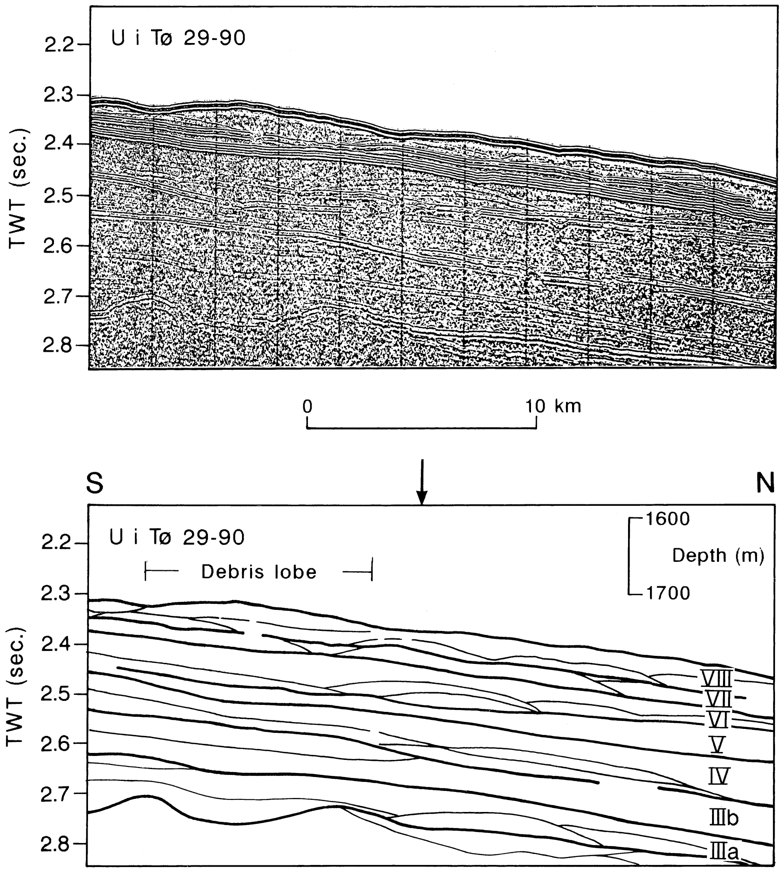

FIG. 14. Cross profile (sparker) of stacked debris lobe sets. See Fig. 7 for location. After Laberg and Vorren (1996a).

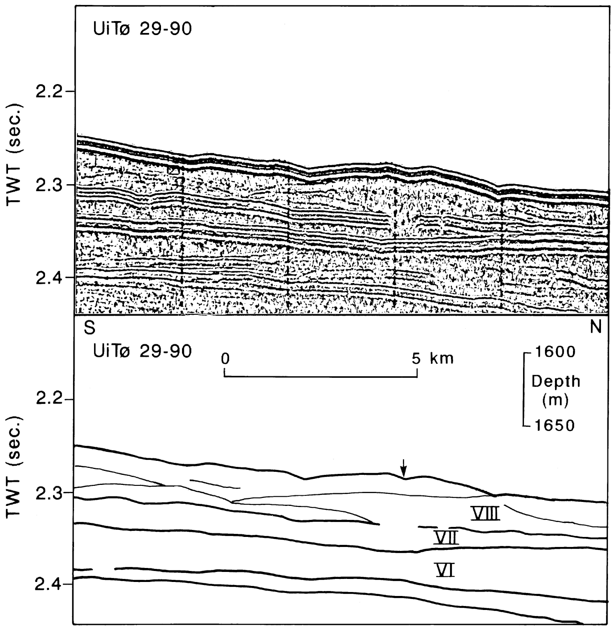

FIG. 15. Sparker profile illustrating the stacked, mounded signature found within seismic unit VIII on the Bear Island TMF. A central depression is found on top of one of the mounds (arrow). Water depth is indicated (see Fig. 7 for location) (after Laberg and Vorren, 1995).

ORIGIN OF THE DEBRIS FLOWS

In discussing the origin of the debris flows we aim to elucidate the following questions:

What was the source area of the material involved in the slump/slide that evolved into the debris flow?

How and when was this material transported to the shelf break/upper slope?

How and when was this material accumulated at the shelf break/upper slope?

What release factors caused the material to slump/slide?

Why did the slide evolve into a debris flow?

Did the debris flow erode the base and incorporate material?

What was the flow behaviour of the debris flows?

Source area

Laberg and Vorren (1995) pointed to the similarities in textural composition and water content of the debris flow diamicton to the youngest glacigenic diamicton on the outer shelf (Sættem *et al.*, 1992; Poole *et al.*, 1994). The lithology as well as the fossil content indicated clearly that *most* of the debris flow sediment was derived from glacigenic shelf diamictons (see below). Radiocarbon dating also strongly indicated that the material of the debris flow was deposited during peak glaciations (Laberg and Vorren, 1995). Thus, the data from the Bear Island TMF strongly suggest that the adjoining shelf is the source area, and that the material was accumulated when the ice sheet was at the shelf break. Similar conclusions are also reached for the debris flows on the North Sea TMF (King *et al.*,

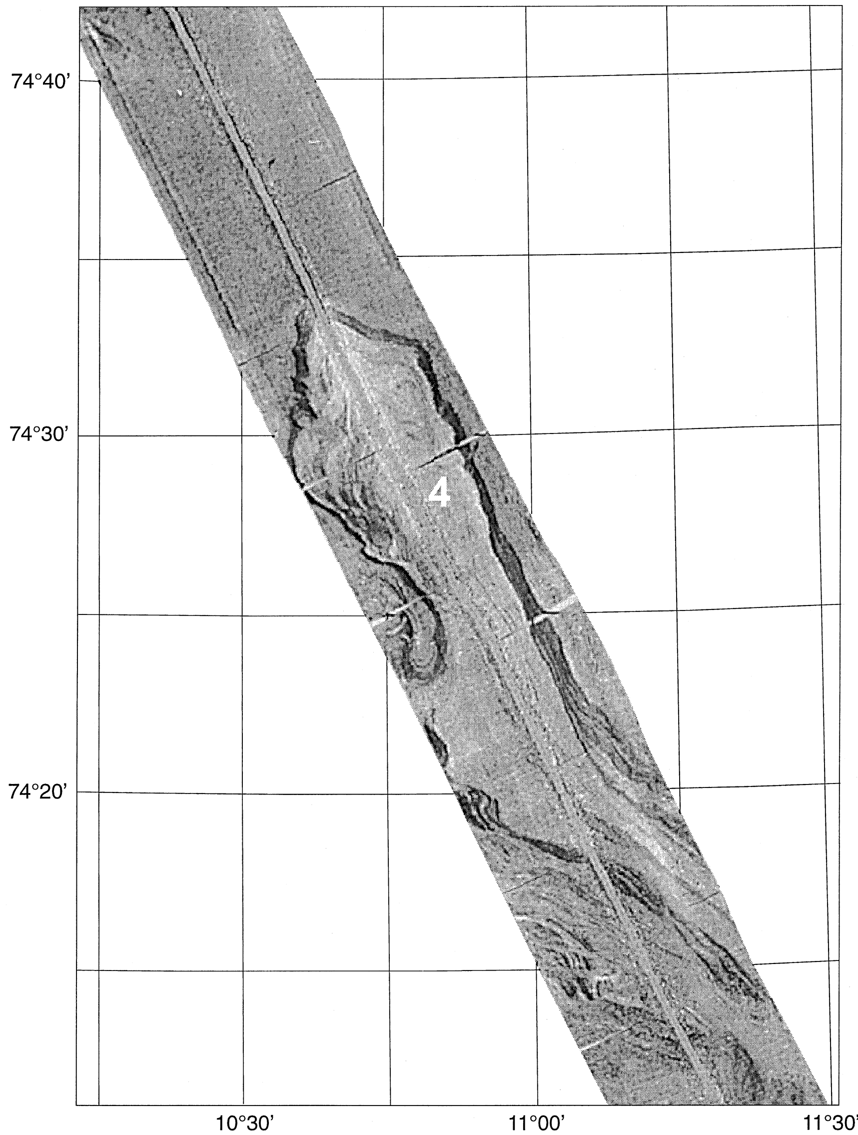

FIG. 16. SeaMARC II sidescan data from the northern, lower Bear Island TMF showing the lower part of one debris lobe (adapted from Vogt *et al.*, 1993) (see Fig. 7 for location).

1996), the Storfjorden TMF (Laberg and Vorren, 1996b) and the Isfjorden TMF (Elverhøi *et al.*, 1997).

Transport to and accumulation at the shelf break

It has been suggested that the diamicton on the outer continental shelf of the Barents Sea was first deposited as deformation till at the glacier bed (Sættem *et al.*, 1992; Laberg and Vorren, 1995). We favour this model (Fig. 17), which implies that glacigenic sediments were brought to the grounding-line as a deforming till-layer (Boulton, 1979). This probably resulted in a build up, either of 'till-deltas' according to the model of Alley *et al.* (1989) or as 'diamict aprons' (Hambrey *et al.*, 1992) along the glacier terminus. The glacigenic sediments could also have continued directly downslope. The sediments deposited in the till deltas are probably inherently unstable, and not preserved on a sloping subsurface. However, there are examples on the seismic records across the shelf break which could be interpreted as till deltas (Fig. 12).

The relatively high water content of the outer shelf diamicton (Sættem *et al.*, 1992) indicates relatively little glacier compaction of the source sediments. Thus the sediments deposited along the grounding line were relatively unstable, and the water content was probably high enough for the sediments to flow downslope without incorporation of additional water (Laberg and Vorren, 1995, 1996b).

Release factors

The morphology, showing slide scars on the present uppermost slope surface and a chaotic seismic facies on the upper fan, indicates that several sediment slides were released near the shelf break (Laberg and Vorren, 1995). The slides may have been triggered by; (1) build up of excess pore pressure due to high sediment input;

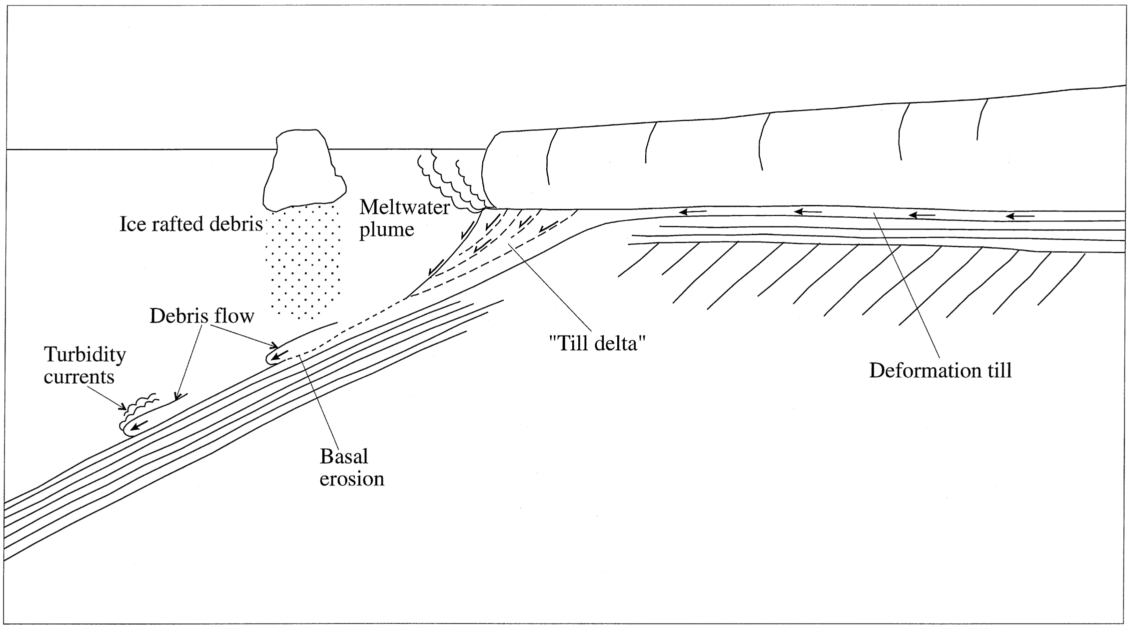

FIG. 17. Schematic model showing the main sedimentary processes on the shelf break and upper slope during the presence of an ice sheet at the shelf break (from Laberg and Vorren, 1995).

(2) earthquakes; (3) oversteepening; (4) ice loading (e.g. Mulder and Moran, 1995); and/or (5) seepage of shallow gas. The repeated release of slides on a regional scale suggests a trigger mechanism common to these particular settings. Thus, most likely the slides were released by a build-up of excess pore pressure and oversteepening, and perhaps sometimes ice oscillation and iceberg calving.

Basal erosion?

Laberg and Vorren (1996b) observed that the base reflections are not often found on the upslope part of the debris flows on Bear Island and Storfjorden TMF. This, together with the fact that the regional reflections immediately below the mounds have in some places an irregular character, indicates disturbance and probably incorporation of underlying sediments during downslope flow. King *et al.* (1996) report that occasional basal indentations, presumably troughs, occur in the North Sea TMF, indicating the possibility of minor erosion at the base. Aksu and Hiscott (1992), investigating debris flows on the Newfoundland continental slope, also found that the basal reflection of a small number of lenses exhibited down-cutting into the underlying acoustically stratified unit. Hiscott and Aksu (1994) showed that the snout of a large debris lobe had erosional power and that a remoulded zone occurred beneath the lobe. Thus, it appears safe to conclude that the debris lobes comprise sediments both from the original shelf diamicton and, additionally, in some cases, sediments from the more distal continental slope.

Flow behaviour

The debris flows moved downslope following bathymetric lows between older deposits (Aksu and Hiscott, 1992; Laberg and Vorren, 1995). Flows that moved further downslope than their forerunners spread out laterally, resulting in a width and thickness increase on the lower fan. Generally, the debris flows containing the largest sediment volume have the longest run out distance (Laberg and Vorren, 1995).

On several of the debris lobes on the Bear Island TMF and on Storfjorden TMF, Laberg and Vorren (1995, 1996b) observed central depressions on the debris lobes (Fig. 15). They concluded that these depressions most likely originated from freezing of the marginal parts of the flow. This may leave a central mobile core which, in turn, caused a channel-like collapse on the crest of the debris lobe.

According to Elverhøi *et al.* (1997), the high clay content of the debris flows indicates that the lobes consist of cohesive material, which favours a plug flow mechanism for their origin. Plug flow is characterised by no internal movement in the sediments during the flow. The original structure of the sediments is thus retained from the release of the slide to its final position.

Many of the observed debris flows have a large run out distance on low gradient slopes, particularly on the Bear Island and North Sea TMF. This indicates low viscosity behaviour. The mobility of debris flows involves an important contribution from excess pore fluid in allowing long run out distances. As indicated above, the water content of the diamicton transported to the shelf break was probably high enough to cause flow. Furthermore, in the marine environment, excess pore fluid pressures are probably generated both during remoulding of porous and saturated materials following slope failure, and during flow, as wet muds are overridden and incorporated into the base of the flows (Aksu and Hiscott, 1992). The latter process may have been of some importance on the upper slopes, as documented here, because sediments deposited rapidly from suspension just beyond the glacial margin would have been extremely rich in pore water.

Textural variations might also control the spatial distribution of the debris flows as discussed by Hiscott and Aksu (1994). They suggested that 'high efficiency' debris flows with a long runout distance are more mud rich than 'low efficiency' debris flows. The higher shear strength of the mud-rich sediments, caused by high cohesive strength (Hampton, 1972), results in the build-up of larger amounts of sediment on the upper slope prior to sediment release. However, the scanty data at hand so far indicate that the observed debris flows on the Norwegian Sea continental margin have very similar textural composition. Thus, the size of the feeding slide and/or the slope gradient seem to be the decisive factors regarding the run out distance.

CHANNELS

Norwegian margin

The observed channels on the Norwegian margin can be grouped into two types according to their size and morphology: (1) Gullies on the upper and middle continental slope; (2) Large channels on the continental slope, rise, and in the deep sea.

Gullies

On the upper continental slope south of Bear Island TMF, there are several gullies (Bugge, 1983; Kenyon, 1987; Vorren *et al.*, 1989). Many of these begin at the shelf break and merge together further downslope. Some of the gullies terminate at a depth of 1000 m while others continue to greater depths (Bugge, 1983). The gullies are up to 150 m deep and 1 km wide. Buried fossil gullies of similar size also occur (Vorren *et al.*, 1989) (Fig. 18). In interpreting the gullies there are two important features: (1) The gullies seem to be eroded into older glacial deposits and they are cutting into primary bedding planes; (2) The gullies have a 'fresh' relief and do not appear to contain any sediment infill (Fig. 18). These observations indicate that the gullies probably did not form during full glaciations. Erosion by density currents/cold dense water formed on the shelf was suggested by Bugge (1983) and Vorren *et al.* (1989). Gullies which are probably of similar origin have also been found on the upper Bear Island TMF (Laberg and Vorren, 1995) and the Storfjorden TMF (Laberg and Vorren, 1996b) (Fig. 7).

Large channels

Two channel systems have been observed along the Norwegian–Barents Sea continental margin, one on either side of the Bear Island TMF (Fig. 3 and Fig. 7). We here propose to name the southern channel system

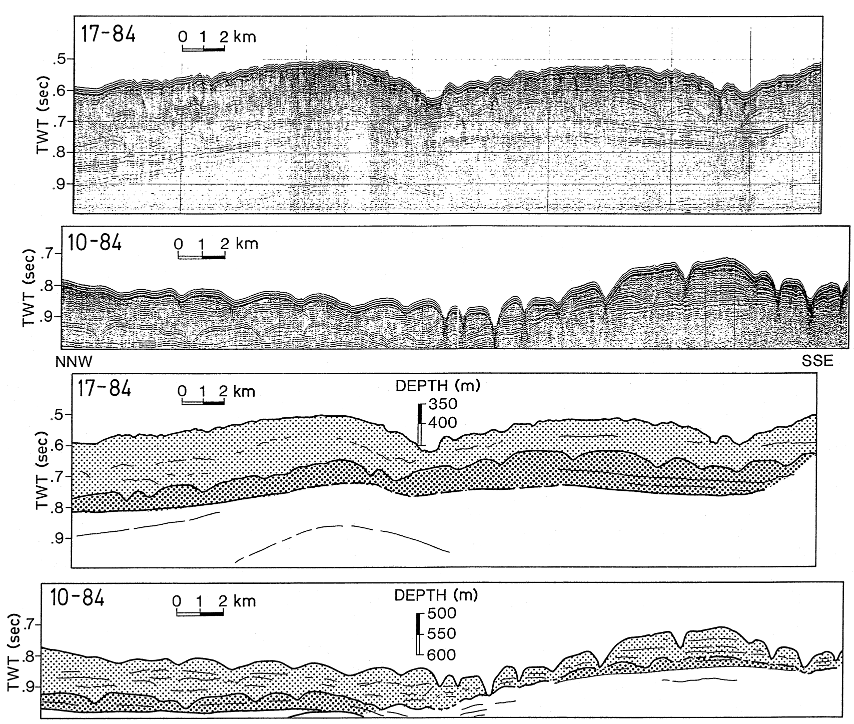

FIG. 18. Sparker profiles along the upper continental slope south of the Bear Island TMF in water depths of ca 400 and 600 m (see Fig. 7 for location). The profiles cross gullies up to 150 m deep. Note the buried gullies below the lightly stippled upper seismic unit (after Vorren *et al.*, 1989).

'the Lofoten Basin Channel', and the northern 'the Inbis Channel' (the Interfan Bear Island TMF and Storfjorden TMF channel system).

The Lofoten Basin Channel, located in the axis of the deep embayment south of the Bear Island TMF was observed and described from long-range side-scan sonar data by Dowdeswell *et al.* (1996). This channel system, which leads into the Lofoten Basin, increases in depth to about 30 m and the width, which is about 3 km, increases greatly where it reaches the basin plain. The fairly flat area beyond the channel has discontinuous, parallel bedding and medium 3.5-kHz penetration of up to 25 m. Recent coring results lend support to the idea that this is a sandy channel mouth lobe (Dowdeswell *et al.*, 1996). Dowdeswell *et al.* (1996) interpret the channel system as being formed either by cascading dense cold water from the Barents Sea, formed through brine rejection during winter sea-ice formation, and/or by turbidity currents generated by the many debris flows on the slope above.

The Inbis Channel. On the continental rise west of Bear Island in water depths of 2360–2520 m, a 60 km long and 5–15 km wide nearly flat bottomed east–west oriented channel exists. To the south and the north, the channel is bordered by the Bear Island and Storfjorden TMFs, respectively (Fig. 19).

The upslope part of the channel is buried under debris flows having their origin on the upper slope in front of Kveitehola Trough (Fig. 7 and Fig. 13). The southern limit is partly bordered by debris flows of the Bear Island TMF (Laberg and Vorren, 1995, Fig. 7 and Fig. 19).

At water depths of 2450–2500 m, the northern channel flank is built up by a set of overbank deposits forming a 10–15 m high natural levee (Fig. 20). A few hundred meters to the north the overbank deposits change into an area (up to 10 km wide) of well developed asymmetric mud waves, having amplitudes of 2–4 m and spacings of 200–400 m.

The steepest slopes and crest displacements of the mud waves adjacent to the Inbis Channel (Fig. 20), suggest upslope migration against the sediment transporting currents (Allen, 1982; Blumsack and Weatherly, 1989). Deep-towed high-resolution side-scan sonar profiles displayed parallel linear features (width of the stripes are about 10 m) on the surface of the mud waves. The stripes apparently have the character of current lineations, i.e. linear bedforms of extremely low relief oriented parallel to a strong downslope current. This interpretation is corroborated by the occasional appearance of obstacle marks within the lineations. The orientation of the current lineations radiate fanlike downslope from the Inbis Channel (Fig. 19). Thus, both the mudwaves and the current lineations show that density currents in the Inbis Channel have frequently been overflowing the channel, depositing sediments in the overbank areas.

The sediments in a 8.16-m long core (GIK 23257) from the northern levee (Fig. 19 and Fig. 21) comprise two sequences of very fine grained (95 wt.% of < 20 µm) turbidite mud; an upper sequence from 17 to 475 cm, and a lower sequence from 655 cm to the base of the core (Fig. 21). Single sets up to 10 cm thick consist of dark basal cross-bedded silt laminae

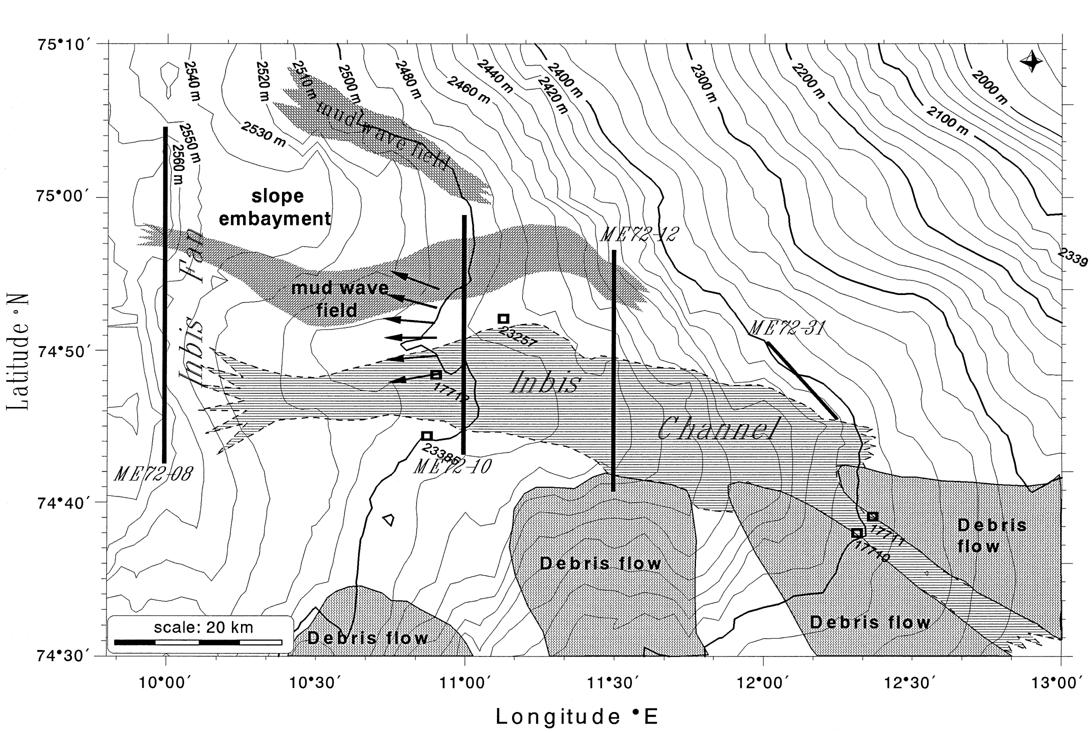

FIG. 19. Map of the Inbis Channel on the western Barents Sea continental slope and rise (see Fig. 7 for location) with adjacent mud wave fields and current lineations (bold arrows). North–south shaded lines indicate 3.5-kHz profiles shown in Fig. 13 and Fig. 20. Location of sediment cores is shown by circles.

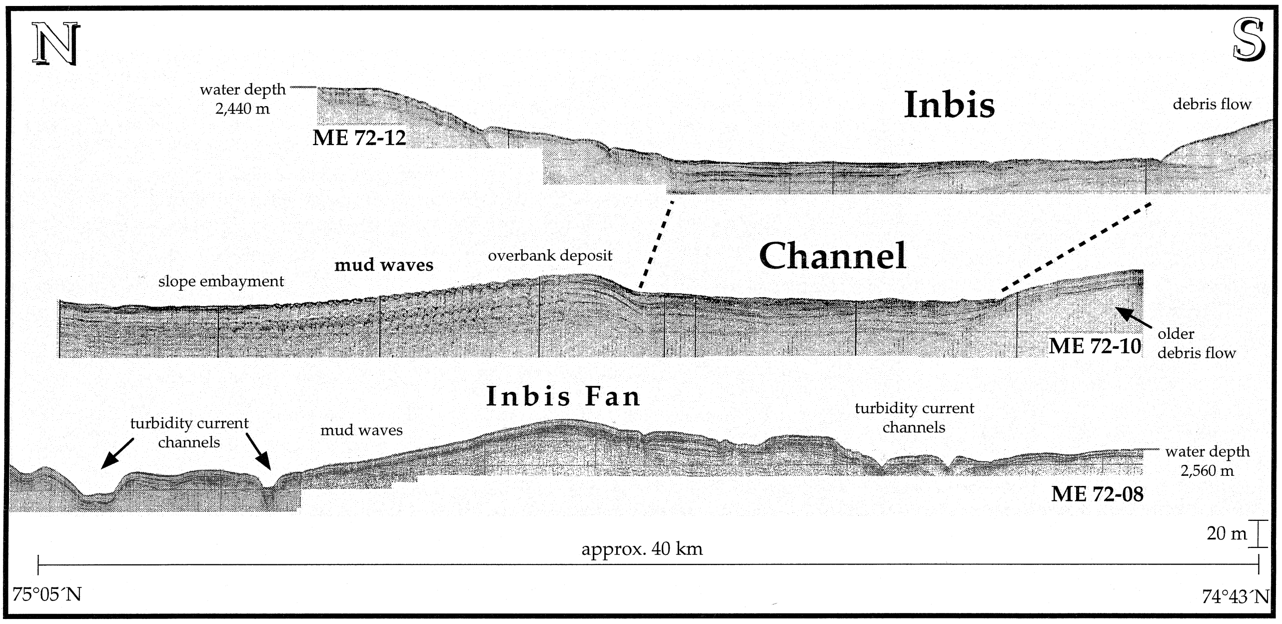

FIG. 20. Three 3.5-kHz-profiles (RV 'METEOR' cruise 7-2, 1988) crossing the Inbis Channel, looking from the deep-sea floor upslope to the east. Vertical exaggeration is approx. 1:50. See Fig. 19 for location.

changing to brighter, very fine ungraded mud (Fig. 21). The units have a typical distal turbidite character as described by Stow and Shanmugam (1980). Scattered ice rafted debris (IRD) in all levels of the laminated mud indicates that the depositional time for one set has lasted longer than for a 'normal' distal turbidite and/or that the IRD activity was high during the deposition of the turbidites. Both of the turbidite sequences are probably of Late Weichselian age.

The sediment-echo character (3.5-kHz sub-bottom profiling) of the relative flat channel bed differs clearly from the surrounding uneven and ridged sea bottom by its hard reflecting irregular seismically laminated sediment infill (Fig. 20). The deepest part of the channel is along the southern side, whereas the sediment bodies in the channel are thicker along the northern side.

Beyond the upslope limit of the Inbis Channel (Fig. 19) a branched system of small V-shaped valleys between the debris flow deposits can be traced as tributary channels by higher reflectivity and small morphological irregularities in their deepest part. Also older valleys filled up by debris flows are dominated by this echo-character (cf. Fig. 13).

At the mouth of the Inbis Channel there is a >50 km wide fan-like accumulation, the Inbis Fan. A fan-like accumulation area is also deduced from the GLORIA and 3.5-kHz profiler data (Dowdeswell *et al.*, 1996). On profile ME72-08 (Fig. 20), sharp erosive channel incisions 0.2–1 km wide and 5–10 m deep bordered by natural levees can be observed flanking the Inbis Fan.

The location of the Inbis Channel indicates that its main source area is the southern flank of the Storfjorden TMF, the northern flank of the Bear Island TMF and the interfan area in front of Kveitehola Trough (Fig. 7). Core GK 23257 (Fig. 21) indicates that turbidity currents were quite active during episodes with high IRD input, i.e. during glaciations. Also, the partial infilling of the Inbis Channel by debris flows, which had their main activity during peak glaciations (Laberg and Vorren, 1995; Vorren and Laberg, 1996), suggests that the density current activity mainly took place during glaciations.

The origin of the density currents running along the Inbis Channel may have been dense water formed by cooling, sea-ice formation and brine rejection or they may originate from small slides on the upper slope transforming into turbidity currents. Along with decreasing gravity energy of the turbidity current, the distal portion of the suspension load was retarded and deflected to the right by the Coriolis force, where the levee accumulated (Fig. 20). The Coriolis force, affecting predominantly the upper fine-grained level of the density current, may explain the greater thickness of the laminated mud along the northern side of the channel.

The Holocene succession at the lower slope (>2100 m water depth) is generally only a few centimetres thick (Blaume, 1992). Sandy turbidites have been identified within the Holocene sediments, but the thickness and frequency of the turbidity currents seem to have been far less than during glacial periods.

East Greenland margin

GLORIA long-range side-scan sonar and PARASOUND subbottom-profiler records allow us to determine the characteristic acoustic backscatter patterns along the East Greenland continental margin, and to derive from it a better understanding of the interaction

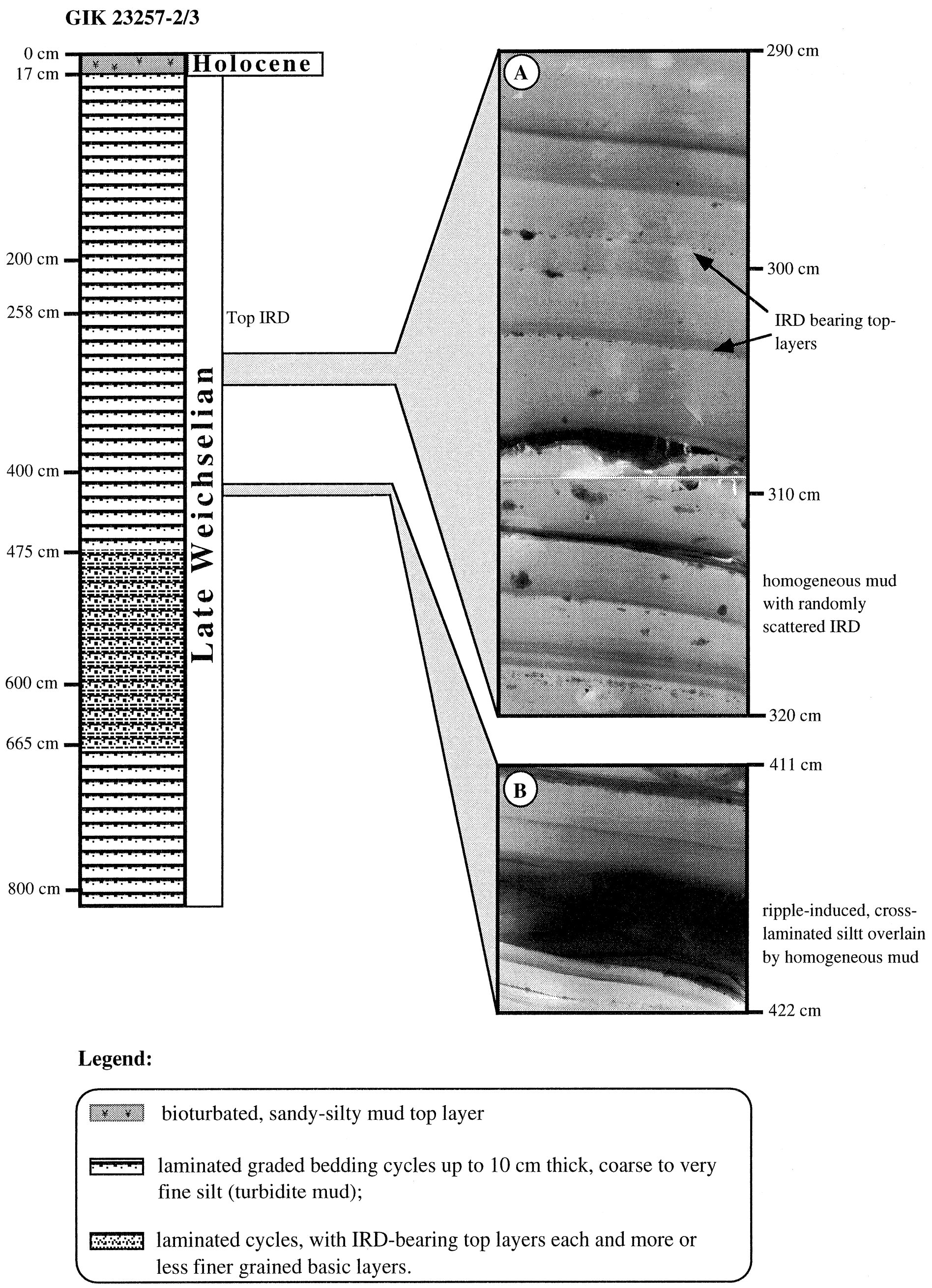

FIG. 21. (a) Schematic lithostratigraphy of sediment core GIK 23257, located on the northern levee near the mud wave field (for location, see Fig. 19). The upper 17 cm comprise Holocene bioturbated sandy, silty mud. The Late Weichselian sediments comprise two sequences of muddy turbidites, an upper from 17 to 475 cm and a lower from 655 cm to the base of the core. Two zoomed X-ray radiographs show core interval 290–320 cm (A) of the upper turbidite sequence with repeating couplets of laminated, layered silts to nearly homogenous muds with IRD (mud pellets) in the upper parts of a turbidite set, and a core interval 411–422 cm (B) with a set of cross/parallel-laminated silty muds indicating turbidity current origin.

between ice sheets, bottom-water currents and sediment-transport processes (Mienert *et al.*, 1993).

Along the East Greenland Margin, southward flowing currents with a velocity of 2.3 cm/s at approximately 1400 m water depth (Aagaard *et al.*, 1985) carry cold water masses from the Arctic Ocean through the Denmark Strait to the North Atlantic Ocean. These currents may cause an intense along-slope transport of fine-grained sediments. In contrast, the calving of icebergs from the Greenland Ice Sheet and the production of melt water may transport large amounts of coarse- and fine-grained sediments to the shelf, which are finally carried and deposited along the margin and in the deep sea by down-slope gravity flows. Additionally, cold water may cascade from the margin and drive the down-slope transport of sediments. Strong evidence of down-slope processes was found in the sonographs from the northern and southern Greenland Basin, but it was not evident in the Boreas Basin (Mienert *et al.*, 1993; Hollender, 1996) (Fig. 1).

The southern Greenland Basin shows four distinct channel systems which may have experienced several events of active and inactive phases of sediment transport. These channels (Fig. 22) have a low sinuosity and a length of up to 300 km. The longest channel originates from the upper slope (about 1500 m water depth) at ca 73°10′N 16°W and is possibly connected to an east–west trending transverse trough leading into Kaiser Franz-Josef Fjord. However, the semi-permanent sea-ice cover on the shelf prevents bathymetric surveys and, therefore, the bathymetric control on the shelf is poor and allows only preliminary interpretations. Also, the uppermost part of the channels was not seen, due to sea ice and distortions from water stratification that limit the range of the side-scan sonar system. However, it is clearly seen that there is a well-developed tributary pattern of shallow (< 10 m deep) channels with a high order of branching on the upper slope (Mienert *et al*., 1993). Downstream from the tributary channels, a deep (< 100 m) single channel develops which turns north and runs towards the deepest part of the basin (3800 m water depth) (Fig. 22). This channel gradually changes from a U-shape to a V-shape, with characteristic steep slopes and a bulge of sediments at the levee on the eastern side of the wall (Fig. 23). PARASOUND profiles across the channel show that it can be narrow (2 km) and fairly straight as well as wide (10 km), with a weak sinuosity where the depth of the channel increases from 50 m in the U-shaped section to 90 m in the V-shaped part. Sediment box-cores collected inside and outside the channel (Mienert *et al*., 1993) show ash layers and also sandy sediments at their base (approximately 40 cm below sea-floor) and undisturbed sediments above. A pronounced ash layer in the cores may be the Vedde ash, which has a ^{14}C age of 10.3 ka (Bard *et al*., 1994). This age may indicate that the channel system has been inactive since the Younger Dryas. On some outside walls of the channels, regularly spaced bands of contrasting backscatter exist, which possibly represent sediment waves with wavelengths from 2 to 3 km. This feature in deep-sea channelized systems is usually attributed to moulding of the muddy overbank sediments by turbidity currents. Turbidity-current activities may be related to large sediment supply rates which possibly occurred during times of ice-sheet advances towards the shelf edge.

All four channels in both the southern and northern Greenland Basin show a similar pattern in that they finally turn north and then run towards the deepest part of the basin (Fig. 22). Thus the channels run against the flow of the modern southward directed deep-water currents. While these currents may cause a vigorous southward directed transport of fine-grained sediments, the turbidity currents in the channels may dominantly transport coarse-grained sediments northward. Not only the erosional channel areas show characteristic backscatter patterns. The mainly depositional regions in the distal part of the channel systems also have characteristic patterns. The pattern

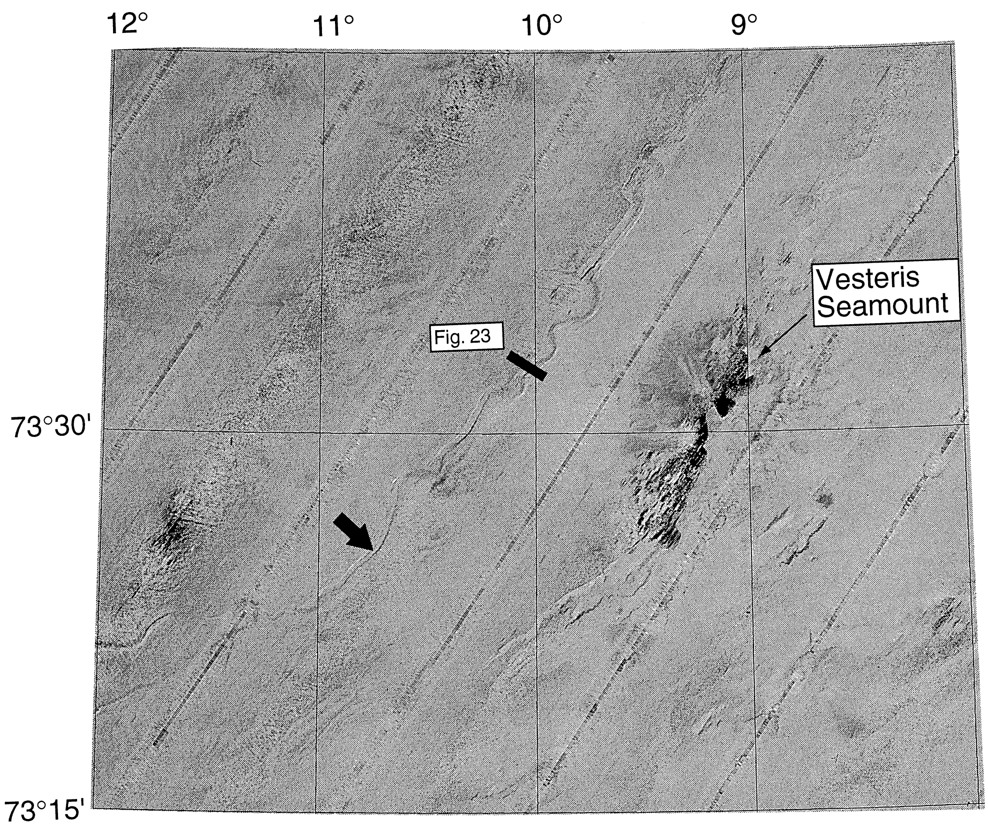

FIG. 22. GLORIA sonograph mosaic showing a major channel (black arrow) running with a low sinuosity towards the 3800 m deep Greenland Basin. It passes the approximately 3000 m high Vesteris Seamount located east of the channel (modified from Mienert *et al*., 1993). Location of Fig. 23 is given.

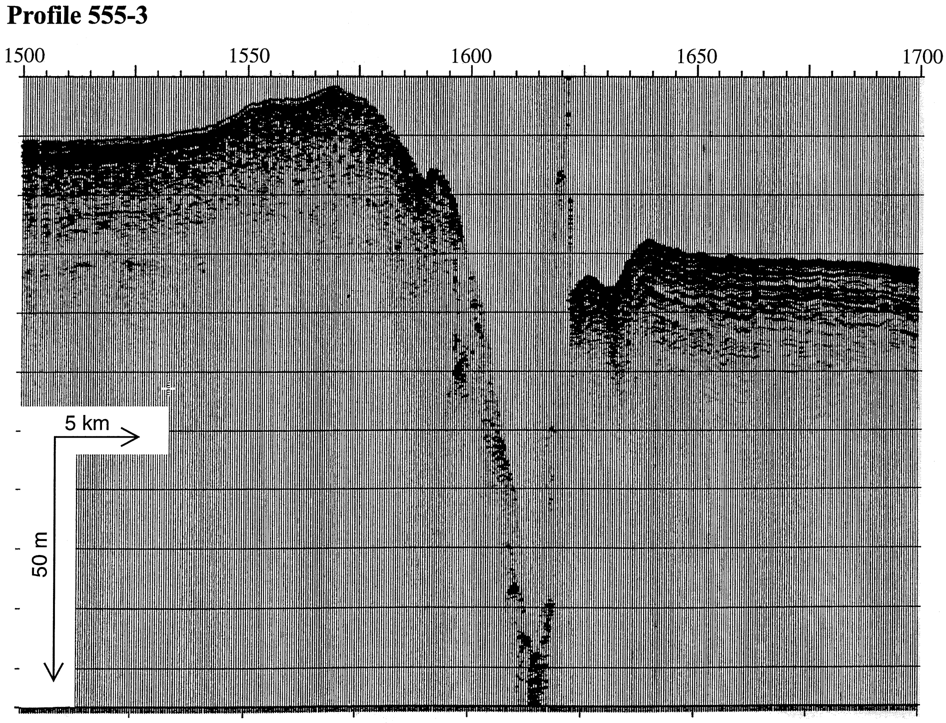

FIG. 23. Parasound profile showing a depth transect across the Greenland Basin channel at about 3040 m water depth (73°34′N, 10°03′W). For location see Fig. 22 (modified from Hollender, 1996).

is braid-like, with overall higher backscattering that has been attributed elsewhere to deposits of sand, for example beyond a channel on the Orinoco deep-sea fan (Belderson *et al.*, 1984) and beyond channels on fans in the Bering Sea (Kenyon and Millington, 1995). Observed features on the East Greenland margin are probably very shallow channels and sandy lobes. Overall, the four major sandy depositional lobes identified in the Greenland Basin cover an area of about 3000–5000 km^2.

The northern Greenland Basin shows a contrasting pattern, seemingly devoid of channelized features (Fig. 1). This non-channelized pattern extends downslope in front of an approximately 300 km long and 400 m deep transverse trough known as Belgica Strait (Perry, 1986). What kind of environmental conditions favour the development of channel systems on polar continental slopes is still not clear. One possible explanation may be related to the position of the ice sheet on the shelf which influences the mass transport across the shelf to the slope region.

CONCLUSIONS

Several large late Quaternary slides have been identified on the eastern Norwegian–Greenland Sea continental margin (the Storegga, Trænadjupet, Andøya and Bjørnøyrenna slides). Buried slides have also been identified in the Storegga slide area and on the western Barents Sea continental margin, indicating a long period of slope instability.

The high sedimentation rate may have led to a build-up of excess pore water pressure, perhaps with additional pressure caused by gas bubbles, and may in turn have led to unstable sediments within relatively large parts of the Norwegian–Greenland Sea continental margin. Triggering might have been prompted by earthquakes or by decomposition of gas hydrates.

Trough mouth fans (TMF) are fans at the mouths of transverse troughs/channels on presently or formerly glaciated continental shelves. In the Norwegian–Greenland Sea, seven TMFs have been identified (the North Sea, Bear Island, Storfjorden, Bellsund, Isfjorden, Kongsfjorden and Scoresby Sund TMFs). The area of the TMFs varies from 2700 km^2 (Kongsfjord TMF) to 215 000 km^2 (the Bear Island TMF).

The Trough Mouth Fans are depocentres of sediments accumulated in front of ice streams draining the large Northwest European ice sheets (Barents Sea, Fennoscandian and British ice sheets) and the Greenland Ice Sheet. The sediments deposited at the shelf break/upper slope by the ice stream were remobilized and transported downslope, mostly as debris flows.

Besides being a locus for sediment deposition, the Trough Mouth Fans were also the main sites of fresh water supply to the oceans in front of glaciated margins (mainly in the form of icebergs) during the mid/late Pleistocene ice ages. The Trough Mouth Fans hold the potential for giving information about the various ice streams feeding them with regard to velocity and ice discharge.

Two large deep-sea channel systems have been observed along the Norwegian continental margin, the

Lofoten Basin Channel and the Inbis Channel. Along the East Greenland margin, several channel systems have been identified.

The deep-sea channels may have been formed by dense water originating from cooling, sea-ice formation and brine rejection close to the glacier margin or they may originate from small slides on the upper slope transforming into debris flows and turbidity currents.

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