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THE BARENTS SEA ICE SHEET — A MODEL OF ITS GROWTH AND DECAY DURING THE LAST ICE MAXIMUM

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On the basis of geomorphological and sedimentological data, we believe that the entire Barents Sea was covered by grounded ice during the last glacial maximum. ^{14}C dates on shells embedded in tills suggest marine conditions in the Barents Sea as late as 22 ka BP; and models of the deglaciation history based on uplift data from the northern Norwegian coast suggest that significant parts of the Barents Sea Ice Sheet calved off as early as 15 ka BP. The growth of the ice sheet is related to glacioeustatic fall and the exposure of shallow banks in the central Barents Sea, where ice caps may develop and expand to finally coalesce with the expanding ice masses from Svalbard and Fennoscandia.

The outlined model for growth and decay of the Barents Sea Ice Sheet suggests a system which developed and existed under periods of maximum climatic deterioration, and where its growth and decay were strongly related to the fall and rise of sea level.

INTRODUCTION

Various models have been presented for the maximum extension and timing of deglaciation of the last Barents Sea Ice Sheet. Jones and Keigwin (1988) suggested, on the basis of light isotope influx in the Fram Strait, a massive calving of the Barents Sea at approximately 15 ka BP. A rapid and massive deglaciation of the marine-based Barents Sea Ice Sheet has also been proposed by Denton and Hughes (1981). Alternatively, on the basis of sedimentological and geomorphological information from the Barents Sea, combined with onshore data, a step-wise deglaciation has been suggested by Vorren *et al.* (1988a); Hald *et al.* (1989) and Elverhøi *et al.* (1990). It is important to note that this latter model suggests that a significant proportion of the Barents Sea Ice Sheet calved off at an early stage: approximately 15 ka BP. Data from the eastern parts of the Barents Sea also seem to support the idea of an early phase of deglaciation of the Barents Sea (Polyak, *pers. commun.*, 1990).

So far, only general ideas about the onset of the Late Weichselian glaciation in the Barents Sea have been presented (Kvasov, 1978; Denton and Hughes, 1981). Recently ages of shell fragments and formaminifera from glacigenic sediments in the Barents Sea suggest marine conditions in the area as late as 23 ka BP (Hald *et al.*, 1990; Berg, 1991). Based on the observations of a low relative sea level in central parts of Svalbard at least 20 ka BP before the last glacial readvance, Mangerud *et al.* (1992) also conclude that the Barents Sea was ice free prior to the last glacial maximum. Thus, for the first time, it is now possible to discuss the onset of glaciation in the Barents Sea. A main objective of this paper is to summarize present knowledge on the timing of the onset as well as withdrawal of the Barents Sea Ice Sheet. The latter portion of the paper provides a test of the deglaciation history using uplift data from the area. This consists primarily of the observed present rate of uplift for northern Norway, which is characterized by coastal parallel isobases (Ekman, 1989). These data are assumed to reflect the rebound of the Fennoscandian Ice Sheet, and an important aspect of the study has been to determine a time frame for the deglaciation of the southern and central Barents Sea which is isostatically compatible with the present day uplift pattern along the northern Norwegian coast.

GEOMORPHOLOGICAL AND SEDIMENTOLOGICAL EVIDENCE OF THE BARENTS SEA ICE SHEET

The idea of a Late Weichselian ice sheet in the Barents Sea has been based largely on observations of geomorphological features and the general pattern of sediment distribution (Elverhøi and Solheim, 1983; Solheim *et al.*, 1988; Vorren *et al.*, 1988a; Elverhøi *et al.*, 1990; Solheim *et al.*, 1990) and Holocene raised beaches in eastern Svalbard (Salvigsen, 1981). The principal geomorphological features include: (1) sediment ridges; (2) thick glaciomarine accumulations; and (3) glacial flutes. The sediment ridges, which have an elevation of 20–50 m above the surrounding topography and an areal extension of 1–10 km, have been localized: (i) along the shelf edge in the mouth of the Bjørnøyrenna and farther northward; (ii) in troughs radiating out from Spitsbergenbanken; and (iii) on the outer part of Storbanken (central Barents Sea) (Fig. 1). Based on morphology, a glacial origin seems likely; however, at present, datable material from these features has not been obtained, and there is no final proof of a Late Weichselian formation. However, the thin cover (< 3 m) of soft sediments above and adjacent to these features suggests a relatively 'recent' formation, and a Late Weichselian age has been suggested by most authors (Vorren *et al.*, 1988a; Elverhøi *et al.*, 1990).

Thick (30–40 m) acoustically transparent lenses localized in the inner part of the major embayments in the northern

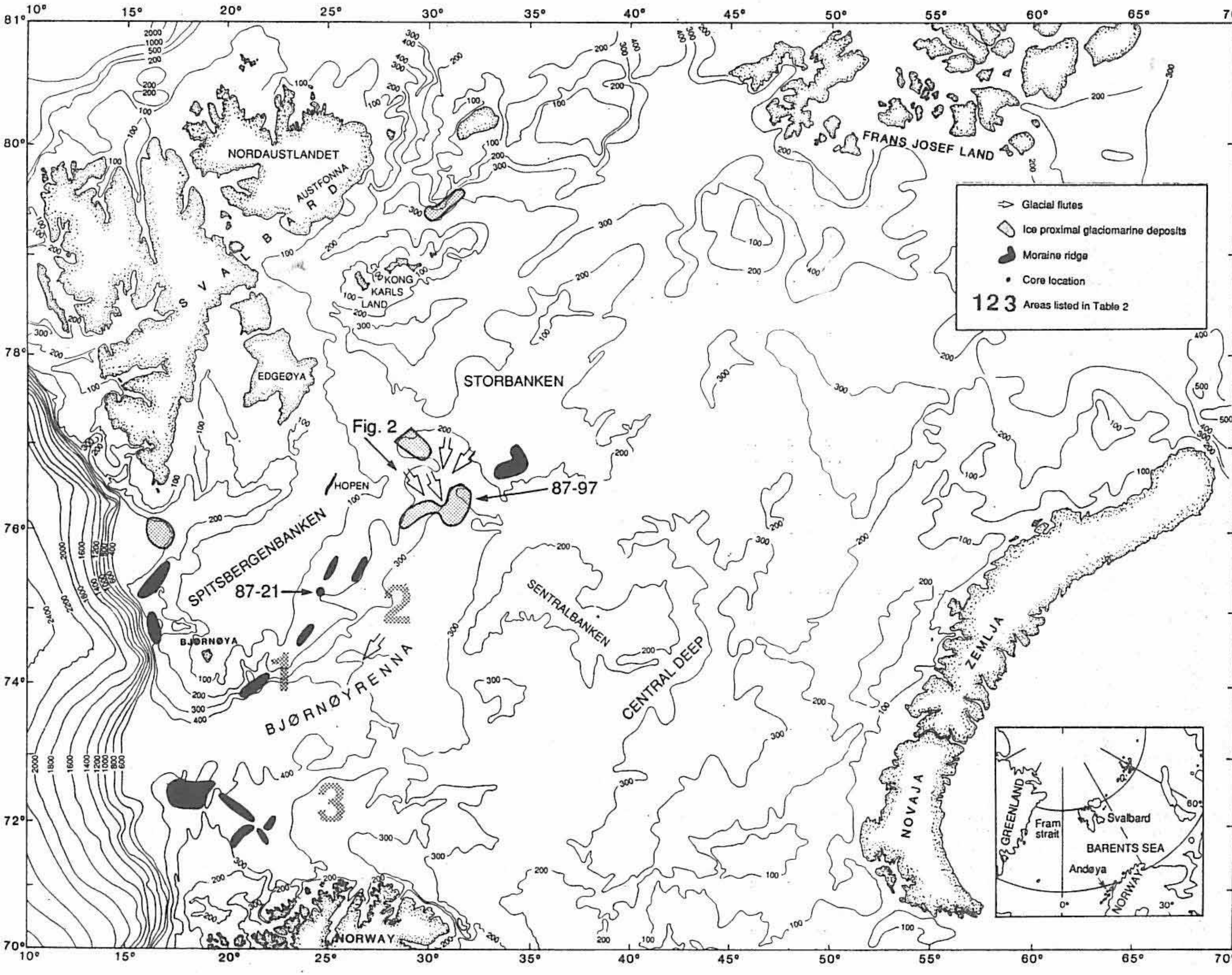

FIG. 1. Bathymetric map of the Barents Sea showing areas with glacial flutes and flute directions. The map also shows the locations of moraine ridges and thick glaciomarine sediments deposited proximally to an ice front (Elverhøi and Solheim, 1983; Vorren and Kristoffersen, 1986; Solheim *et al.*, 1988, 1990; Elverhøi *et al.*, 1990; Russwurm, 1990; Berg, 1991).

Barents Sea (Bjørnøyrenna, Storfjordrenna and in the trough north of Kong Karls Land (Fig. 1) have been used to reconstruct ice recession (Kvasov, 1978; Elverhøi and Solheim, 1983). On the basis of shallow sediment coring, acoustic character and comparison with modern deposits outside ice caps, which terminate in the ocean (e.g. Austfonna, Fig. 1), the transparent deposits are interpreted as ice proximal glaciomarine accumulations (Elverhøi *et al.*, 1990). All of the deposits are found close to a common present day water depth of 300 m, and are suggested to represent contemporaneous formations, with the ice margin located at this depth. They are interpreted to reflect a major halt in the ice recession (Elverhøi *et al.*, 1990). Similarly as for the sediment ridges along the shelf edge, the age of these deposits has not been verified; however, as the sediments of these deposits grade into the Holocene top sediments, a Late Weichselian age has been suggested.

The observation of a 4000 km^2 glacially fluted region covered only by a thin veneer (< 2 m) of late and postglacial sediments in the central Barents Sea (Figs 1 and 2) clearly demonstrates the existence of a grounded Late Weichselian ice sheet in the area (Solheim *et al.*, 1990). The generally southerly direction of the flutes indicates an ice stream flowing down the inner part of Bjørnøyrenna (Fig. 1). An extension of the ice sheet farther westward towards the shelf edge is indicated by isolated observations of glacial flutes farther west in the Bjørnøyrenna (Fig. 1). It is likely that the fluted areas have a wider distribution, but the possibility of their observation is hampered by the thickness of the overlying soft sediments. The flutes disappear when the soft sediment cover becomes more than 2–3 m thick. The fluted surface is also disrupted by intensive iceberg ploughing to the north and towards shallower regions (Solheim *et al.*, 1990).

In general, the northern and central Barents Sea has a thin (< 10–15 m) cover of Quaternary sediments above the underlying Mesozoic bedrock. The thickness of the Quaternary sediments increases in deeper waters (> 300 m of water depth), towards the margins and within the Bjørnøyrenna (Fig. 1), and at the shelf edge the thickness may exceed 400 m (Solheim and Kristoffersen, 1984). The general sediment stratigraphy consists of stiff diamicton (Fig. 3), overlain by softer sediments, which consist of glaciomarine sediments deposited during ice recession (1–15 m thick) and a thin veneer (< 1 m) of fine-grained sediments below 100 m waterdepth and coarser deposits in regions shallower than 100 m. In general the sediments in the upper unit (Fig. 3) are largely Holocene in age. The distribution of stiff diamicton, interpreted as basal till (Elverhøi and Solheim, 1983; Elverhøi *et al.*, 1990), has been used as a basic piece of evidence in ice reconstruction. 'Stiff' diamictons have been widely reported from the northwestern Barents Sea, where these deposits have been found to outcrop at the sea bed or to be covered by a thin veneer of soft sediments (Fig. 3).

The 'stiff' diamicton has previously been interpreted as basal till (Elverhøi and Solheim, 1983). However, consolidation tests have shown that these sediments are overconsolidated, but with estimated preconsolidation stresses (200–500 kPa) far below those resulting from full consolidation below a thick ice sheet. In addition, the recent investigations have demonstrated the existence of two units of overconsolidated diamicton in the northern Barents Sea (Fig. 3). The uppermost, Unit C, is associated with glacial flutes and a subglacial origin is, therefore, considered most likely. For small ice caps, Boulton and Jones (1979) demonstrated that a highly porous layer existed beneath the ice, and that a significant portion of the glacier movement took place within this layer. Recently, a porous deformable layer with low effective stresses has also been identified beneath the fast-flowing Ice Stream B in West Antarctica (Alley *et al.*, 1989). If we apply a similar model to the marine based Barents Sea Ice Sheet, the relatively low preconsolidation stresses do not contradict the possibility of an extensive ice sheet in the area. We therefore conclude that the slightly overconsolidated diamicton of Unit C represents a basal till formed and deformed beneath an extensive ice sheet.

The S_u values of Unit D are higher than for Unit C: in the range of 100 kPa, with measured values up to 300 kPa. But the calculated preconsolidation pressures are too low for full compensation by an inland ice sheet, and, similarly as for Unit C, a significant portion of the ice load must have been carried by the hydrostatic pressure under deposition of Unit D. Microfabric analyses of Unit D (from core 87-21, Fig. 1) at a sediment depth of 5 m show preferential orientation of coarser fragments and rotation of the finer grades around the coarser fragments (Berg, 1991). This type of microfabric has

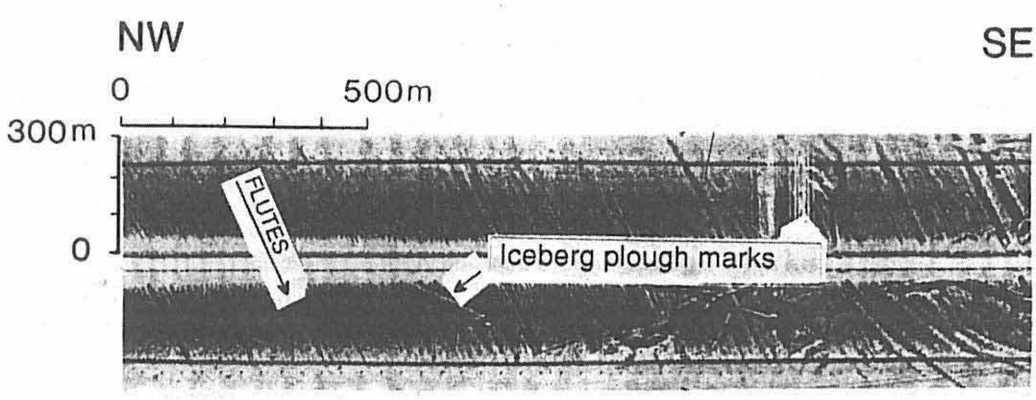

FIG. 2. Side scan sonar record showing glacial flutes from the central Barents Sea (for location, see Fig. 1).

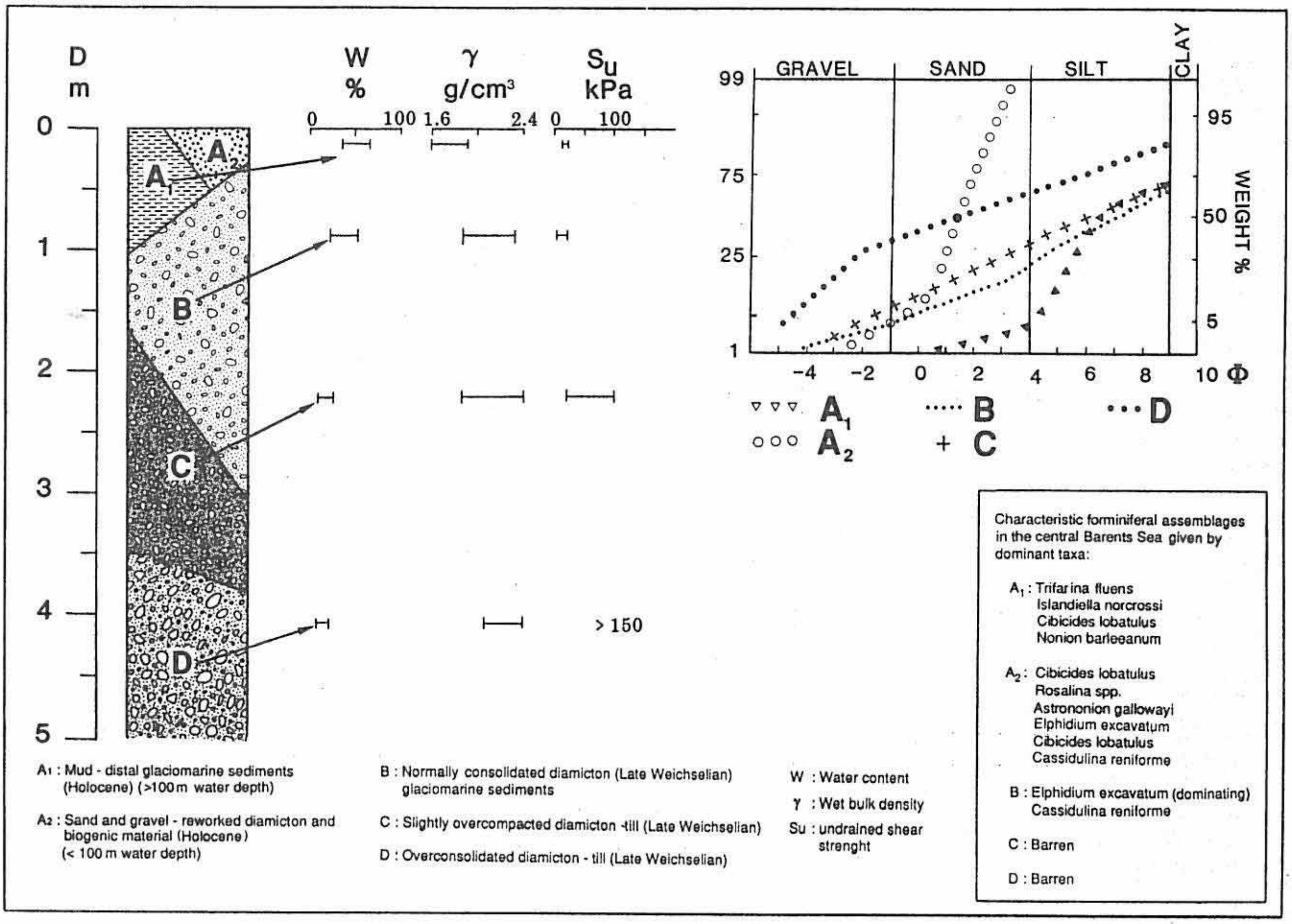

FIG. 3. Generalized lithological section of the Late Cenozoic sediments in the central and northern Barents Sea (Elverhøi and Solheim, 1983; Solheim *et al.*, 1988, 1990; Elverhøi *et al.*, 1990; Russwurm, 1990; Berg, 1991).

been interpreted to reflect shearing and rotation of the sediments, characteristic for lodgement till deposits (Van der Meer, 1987; Lagerlund and Van der Meer, 1990). Plasmic fabric analyses (Brewer, 1976; Van der Meer, 1987) also show a fabric pattern characteristic for shearing and rotation, with a distinctive **unistral** and also **skelsepic** fabric (unistral fabric: parallel orientation of domains of the fine-grained matrix; skelsepic fabric: parallel orientation of domains of the matrix parallel to coarser grains). The plasmic fabric observed in Unit D has also been found in lodgement till deposits (Van der Meer, 1987; Lagerlund and van der Meer, 1990), and we therefore conclude that Unit D, as found in the central and western Barents Sea, represents a lodgement till deposit. The homogeneous and unimodal composition of the clasts in Unit D deposits reflects a local origin and relatively short transport distance. As demonstrated from Unit D sediments from the inner part of Bjørnøyrenna (Antonsen *et al.*) 1991), distinctive lithologies of early Cretaceous or, alternatively, Late Jurassic rocks are found in the various site locations. In the case of a deformation till such as that described from Ice Stream B in Antarctica, a more long distance transport is likely to result in the mixing of various lithologies. The available Unit C sediments demonstrate a more variable clast composition and in the sediments from the inner part of Bjørnøyrenna, the Permo-Carboniferous clasts demonstrate input of materials from remote areas such as Edgeøya or Nordaustlandet and its adjacent regions (Fig. 1), which are the only areas where rocks of Permo-Carboniferous ages are exposed in the area. Even though the available till material is limited compared to the size of the region, the overall characteristics of Unit C and Unit D seem to correspond to deformation and lodgment till origins, respectively.

THE EARLY PHASE OF LATE WEICHSELIAN GLACIATION

Relatively thick (50–100 m) sediments are present above the upper regional unconformity (URU) in the outer part of the Bjørnøyrenna (Solheim and Kristoffersen, 1984). Based on ^{14}C dates and amino acid analyses, a Mid to Late Weichselian age has been proposed for the entire unit (Hald *et al.*, 1990; Sættem *et al.*, 1991). The ^{14}C analyses showed ages in the range of 21 to 35 ka BP and infinite ages (Hald *et al.*, 1990) (Fig. 5). However, younger material has been found to be overlain by older samples, and sediment reworking of older glaciomarine sediments has been suggested as the principal cause (Sættem *et al.*, 1991). In general, the sediments above URU in the outer part of Bjørnøyrenna are nonstratified and consist of silty, sandy clay with scattered pebbles (diamicton), and the presence of overconsolidated deposits with typical lenticular fissility, suggests shearing and deformation from a grounded Late Weichselian Ice Sheet (Sættem *et al.*, 1991).

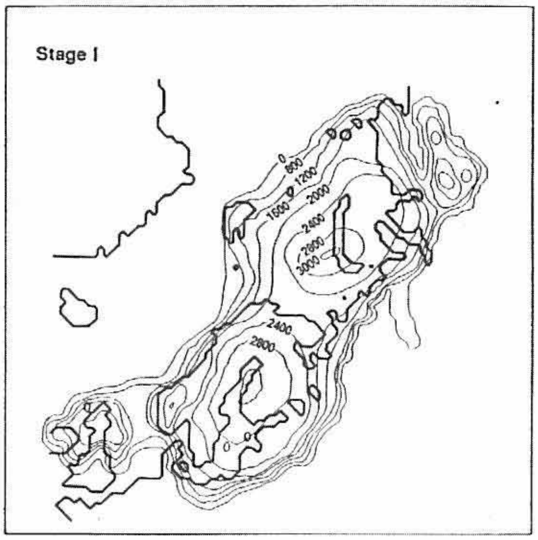

Fig. 4a

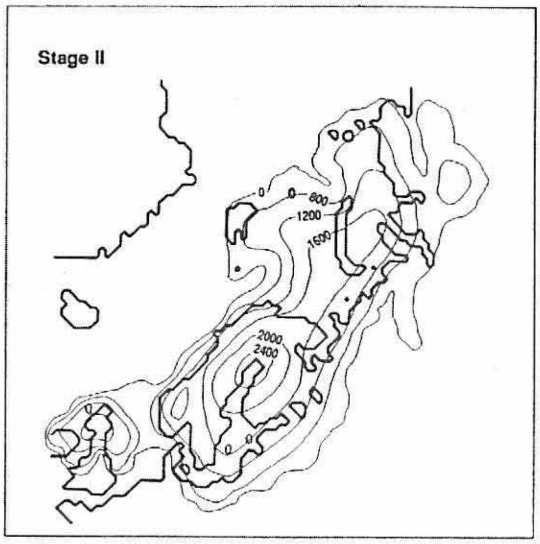

Fig. 4b

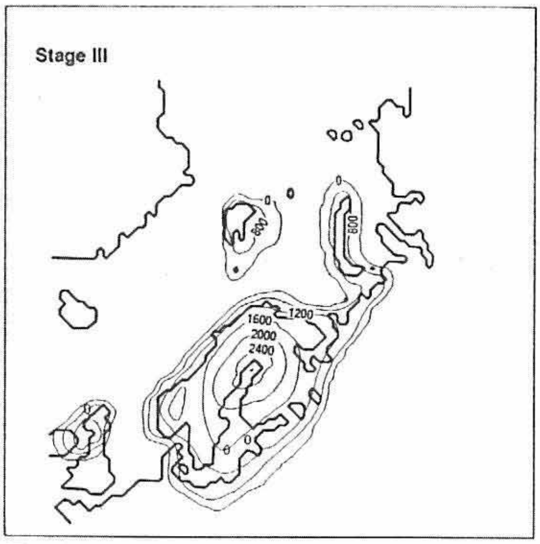

Fig. 4c

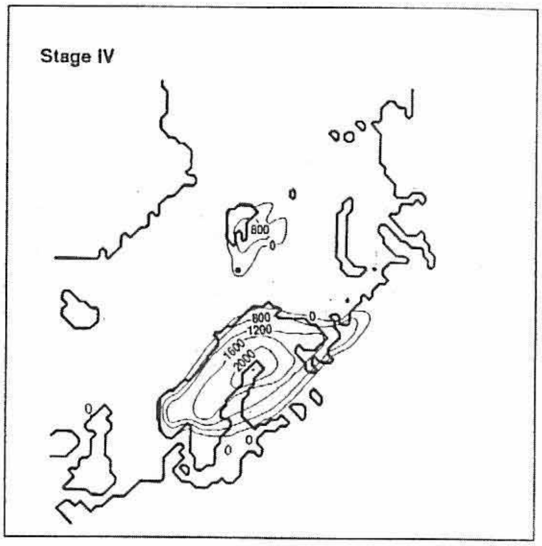

Fig. 4d

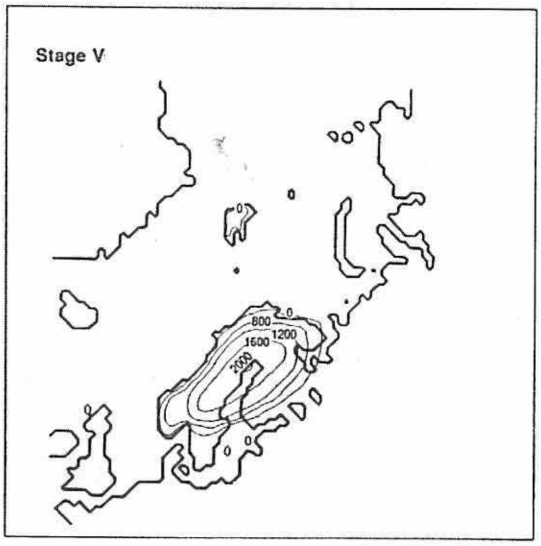

Fig. 4e

FIG. 4a-e. The extent and thickness of the Barents Sea Ice Sheet (BSIS) as used in the computations. Ice reconstructions based on data from Andersen (1981), Elverhøi and Solheim (1983), Vorren *et al.* (1988a), Solheim *et al.* (1990). Tentative ages of the various stages are: Stage I: glacial maximum, stage II: 17 ka BP, stage III: 15 ka BP, stage IV: 11.5 ka BP, stage V: 10 ka BP.

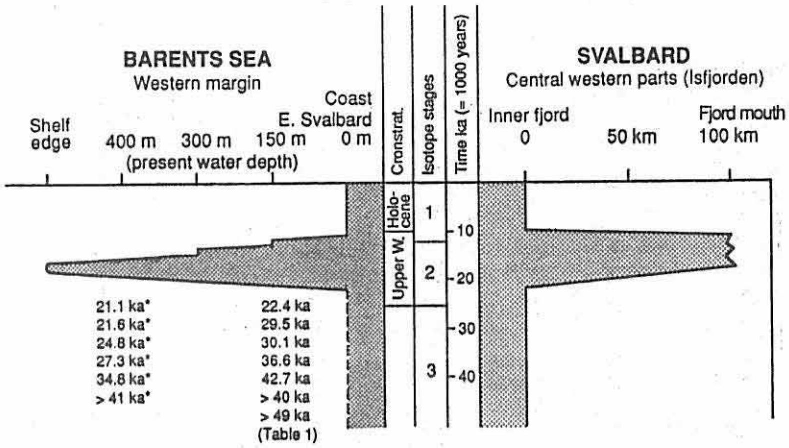

FIG. 5. Time–distance diagram for the central, western Barents Sea. The glaciation curve from Svalbard (central western coast) is adopted from Mangerud *et al.* (*in press*). *Data based on Hald *et al.* (1989). Data from Table 1 refer to ^{14}C-dates from a sediment depth of 0.75 m (6 datings) and a depth of 2.90 m (1 dating). The stepwise ice recession curve is based on the occurrence of ice marginal features (see Fig. 1).

Bioclastic material also has been obtained from a local sediment accumulation (55 m thick) 100 km south of Hopen (Fig. 1, core 87-21). The bioclastic material has been taken from 0.75 m and 2.90 m sediment depths within lithofacies Unit C and from 5.20 m sediment depth within lithofacies Unit D (Fig. 3). According to our interpretation, the bioclastic material has been glacially reworked in the two types of basal till.

In Table 1, two groups of ages can be seen: (1) infinite ages, and (2) ages in the range of 22–30 ka BP. Similarly as for the samples from the Bjørnøyrenna, the amino acid analyses show younger and older ratios. Because of the small amount of material available, only one of the fragments has been identified (*Mya truncata*). Therefore, the amino acid analyses are of limited value. However, if one applies a diagenetic temperature of –1°C for the last 100 ka, the age of the *Mya truncata* fragment is approximately 80 ka BP. The diagenetic temperature of –1°C reflects the present day temperature of the overlying cold polar water masses and is also a relevant temperature estimate for an overlying grounded ice sheet which is at the pressure melting point at its base (Løvø *et al.*, 1990).

Bioclastic material is abundant in the adjacent shallow Spitsbergenbanken, where presently up to 90% of the sediments consists of carbonates (Bjørlykke *et al.*, 1978), and similar conditions may also have existed earlier. Assuming that these dates are correct (Table 1), we can conclude that marine conditions prevailed in the central Barents Sea in the early phase of the Late Weichselian, and that the growth of grounded ice took place relatively late compared to the growth of the other major Late Weichselian ice masses, such as the Fennoscandian Ice Sheet (Andersen and Mangerud, 1989). The timing of the onset of glaciation in the Barents Sea seems to correlate much better with the time of the onset of ice growth on Svalbard (Fig. 5).

The growth of a marine ice sheet like the Barents Sea Ice Sheet represents a major glaciological problem. According to Denton and Hughes (1981), sea ice may gradually thicken and grow into a thin ice shelf. Thickening, combined with sea level lowering, may cause the ice shelf to ground, and then represent an initial loci for ice build-up. A basic problem with this model is the isolation effect of sea ice and the convection of the underlying water masses, thus the theory of Denton and Hughes (1981) is controversial. Alternatively, ice growth may have occurred in response to falling sea level and concurrent expansion of ice caps located on the islands surrounding the Barents Sea, which continued to grow until they joined to form the Barents Sea Ice Sheet (Kvasov, 1978). In addition, the shallow banks such as Spitsbergenbanken and Storbanken (Fig. 1) may represent

TABLE 1. ^{14}C ages and amino acid analyses of shell fragments from core 87-21, south of Hopen (75°19'N 24°43'E; for location, see Fig. 1)

Station nr.	Sediment depth	Material type	Lab. ref. no.	Age	Amino acid D/L ratio*
87-21	0.76 m	shell fragment	Ua 1136	22400 ± 1095	
87-21	0.76 m	shell fragment	Ua 1137	29510 ± 1260	
87-21	0.76 m	shell fragment	Ua 1138	30140 ± 925	
87-21	0.76 m	shell fragment	Ua 1270	36600 ± 1900	
			BAL 1749		0.058
87-21	0.76 m	shell fragment	Ua 940	42700 ± 1500	
		Mya truncata	BAL 1717		0.072
87-21	0.76 m	shell fragment	Ua 1229	> 40000	
			BAL 1748		0.029
87-21	2.90 m	barnacles	Ua 939	> 45000	
87-21	5.20 m	shell fragment	Ua 1810	37480 ± 900	

*D/L is alloisoleucine/icoleucine for the total amino acid population.

Ua: Tandem accelerator, University of Uppsala, Sweden, BAL: Bergen Amino Acid Laboratory, University of Bergen, Norway (Note: the results have not been corrected for reservoir age, approximately 400 years).

sites for ice cap formation. According to this concept, the initial development of the Barents Sea Ice Sheet would be strongly related to sea level and the subaerial exposure of parts of the Barents Sea. Conflicting views exist regarding the exact timing and amount of sea level lowering during the first stage of the last glaciation, 30–20 ka BP (Shackelton, 1987). However, a gradual reduction in ocean water volume, amounting to an approximately 125 m fall in global eustatic sea level, is believed to have peaked at around 18 ka BP (Fairbanks, 1989). This may have left the shallowest parts of Spitsbergenbanken and Storbanken subaerially exposed. Today, the northernmost islands are covered with grounded ice, and 1/3 of the Austfonna (Fig. 1) is grounded below sea level. It is reasonable to assume that the northerly ice masses of 25–22,000 years ago may have expanded and joined with ice caps formed at the bank areas further to the south in the Barents Sea.

Essential to such a theory is the availability of precipitation. The recently studied sediment record from Andøya indicates that the early stage of the Late Weichselian was characterized by a low to middle Arctic maritime climate, probably with a seasonally open ocean, from which moisture could have been exported to feed ice sheet growth (Vorren *et al.*, 1988b). If such conditions prevailed in the Andøya region, a similar setting may also have existed in the southern Barents Sea, and this would have contributed to the growth of the Barents Sea Ice Sheet.

LATE WEICHSELIAN ICE DISTRIBUTION

The later isostatic calculations are based on the entire ice sheet in northern Europe, detailed knowledge of ice distribution and deglaciation in areas peripheral to the Barents Sea is not essential. Thus we focus only on the Barents Sea and its adjacent areas. Most models indicate ice extension to the shelf edge (Vorren *et al.*, 1988a; Elverhøi *et al.*, 1990), even though at present we do not have any dates to confirm such a model (Fig. 4a, stage I). Based on evidence from the island of Andøya (Vorren *et al.*, 1988b) (Fig. 1), the ice margin, of at least the southwestern parts of the Barents Sea, remained stable up to 16 ka BP (Vorren *et al.*, 1988a). A minor withdrawal is indicated as the first step of the ice recession (Fig. 4b, stage II), based on the data from Andøya. The western boundary in Bjørnøyrenna follows the most westerly observation of glatial flutes in the Bjørnøyrenna (Fig. 1); hence, the outline of stage II can also be considered as a conservative estimate of the Late Weichselian ice distribution.

The first major oxygen isotope excursion indicating the first significant ice recession, has been dated by Duplessy *et al.* (1981) as lasting from 15.5 to 13.3 ka BP. Ruddiman and MacIntyre (1981) also suggest a major influx of meltwater and ice-rafted material into the North Atlantic during the period from 16 to 13 ka BP, also identified in the Fram Strait between 15 and 13 ka BP (Jones and Keigwin, 1988). Because of its proximity to the Barents Sea, Jones and Keigwin (1988) relate this influx to a significant deglaciation of the Barents Sea. Based on a combination of lithostratigraphy and oxygen isotope variations in benthic foraminifera, Vorren *et al.* (1988a) presented a time–distance deglaciation diagram for the shelf north of northern Norway showing significant ice recession in the shelf region during the interval between 16 and 14 ka BP.

In the central and northern Barents Sea, the well-defined glaciomarine accumulations at about 300 m present water depth (Fig. 1) indicate a major halt in the ice recession at this depth. Furthermore, there are no marginal features in deeper water between these features and the shelf edge. We conclude, therefore, that significant parts of the ice sheet calved off during the first phase of deglaciation of the Barents Sea (Fig. 4c, stage III). Reconstructions of the eastern extension of the ice sheet are based on the work of Polyak (Polyak, *pers. commun.*, 1990), who indicates that the ice cap in the central, northern Barents Sea decoupled from the Novaja Zemlja ice cap between 16 to 13 ka BP. Because of water depths exceeding 300 m in the Central Deep of the eastern Barents Sea (Fig. 1), it seems likely that these eastern parts of the Barents Sea were also deglaciated during the first major phase of ice recession.

Because of the low concentration of foraminifera (5 individuals per 100 g sediment) in Unit B, which also show evidence of dissolution, and intensive iceberg reworking of the glaciomarine sediments (lithofacies Unit B), we have not been able to establish a geochronological record and date the ice recession from the central and northern Barents Sea (Elverhøi and Solheim, 1987; Russwurm, 1990). A minimum age for the deglaciation of the area, however, can be estimated from biostratigraphy and its correlation. Undisturbed sediments, as reflected in side scan sonar records, have been obtained in limited areas, and Table 2 shows the foraminiferal zonation in one core from such an area in the inner part of the Bjørnøyrenna (Fig. 1). Correlation with earlier investigations in the western Barents Sea suggests that the lowermost zone seems to correspond to the *Elphidium/Cassidulina* fauna described by Elverhøi and Bomstad (1980) and Østby and Nagy (1982) (Table 2). In the southern Barents Sea, the *Elphidum*-dominated fauna disappeared at 12 ka BP (Hald *et al.*, 1989), and if the two *Elphidum*-dominated faunas are correlated, a minimum age for deglaciation of the inner part of the Bjørnøyrenna of 12 ka BP is obtained. It may be argued that there has been time-transgressive development northward, but with simultaneous deglaciation of the northern and southern parts of Bjørnøyrenna, there should not be a significant time lag between the development of the two areas.

At about 11,500 BP, the ice margin was located in mid-fjord positions along northern Norway (Fig. 4d) (Vorren *et al.*, 1988a), and ice marginal features at the southern tip of Storbanken (Solheim *et al.*, 1990) (Fig. 1) are believed to correspond to such a stage. At about 10,500 BP (the Younger Dryas stadial), significant ice masses were still present in Fennoscandia (Andersen and Mangerud, 1990) (Fig. 4e). In our reconstructions we have used a minimum age for the deglaciation of the northern Barents Sea of 10 ka BP, which is based on ^{14}C dates of foraminifera and molluscs in glaciomarine sediments south of Nordaustlandet (Elverhøi and Solheim, 1987) (Fig. 1). This corresponds with the age proposed by Forman (1990) and Mangerud *et al.* (1992),

TABLE 2. Correlation of benthic foraminiferal stratigraphy from the western Barents Sea. (AZ = assemblage zone)

^{14}C-ages (ka BP)	87-97[1]	Area 1[2]	Area 2[3]	Area 3[4]
— 5	Nonion barleeanum	Trifarina-Islandiella (AZ)		Epistominella-Trifirna (AZ A)
	Bucella frigida		Nonion-Cassidulina (AZ)	
— 10	Cassidulina laevigata	Cassidulina-Cibicides (AZ)		Nonion barleeanum (AZ B)
— 11	Nonion barleanum			
— 12	Elphidum excavatum	Elphidum-Cassidulina (AZ)	Elphidum-Cassidulina (AZ)	Elphidum excavatum (AZ C) (Upper)
— 13				
— 14				AZ C (Lower)
Max Weichsel?				

[1]Russwurm (1990); [2]Østby and Nagy, 1982; [3]Elverhøi and Bomstad (1980); [4]Hald *et al.* (1989).

who suggested that the major deglaciation of Svalbard, including the eastern parts of the Archipelago (Edgeøya, Kong Karls Land, Nordaustlandet), occurred around 10 ka BP. The large increase in glacio-isostatic uplift at 10 ka BP in eastern Svalbard supports the idea of a late final deglaciation of the northernmost Barents Sea (Mangerud *et al.*, 1992). According to this concept, the ice distribution shown in Fig. 4d for 11,500 BP may have occurred towards the very end of the Pleistocene.

UPLIFT MODELLING

Glacial isostacy, movements of the solid earth to re-establish isostatic equilibrium during changes of ice load, has been investigated to test the deglaciation history described above. This approach is based on raised shorelines along the northern Norwegian coast that are parallel to the coast. The present uplift pattern also shows the same orientation (Fig. 6). Thus, there is no isostatic signal which suggests the existence of the Barents Sea Ice Sheet along its southern margins. An important aspect of the modelling, therefore, is to determine the time interval during which the Barents Sea Ice Sheet must have calved off, without leaving evidence of isostatic adjustment along its southern margin. As noted, the calculations are based on the entire ice sheet of northern Europe, and the speed of melting from one ice sheet configuration to the next is modelled uniformly. Previous modelling of the Fennoscandian and Barents Sea uplift (Fjeldskaar and Cathles, 1991a,b) assumed a 6-step deglaciation (with ice sheets of 20,000, 15,000, 11,500, 10,500, 9300, and complete deglaciation of the entire region at 8500 BP). Because the present study is focused on the deglaciation of the Barents Sea, we have added one stage to the deglaciation: a step between the maximum stage and the fully deglaciated Barents Sea, here termed stage II (see Fig. 4b).

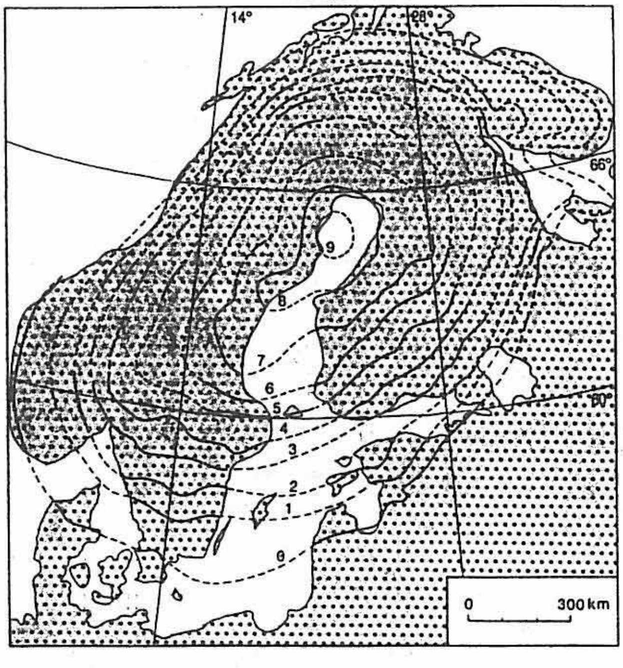

FIG. 6. Observed present rate of uplift in Scandinavia (Ekman, 1989).

The earth is modelled as a nonspherical viscoelastic fluid in which the viscosity may vary with depth, because it is overlain by a uniformly thick elastic lithosphere. With this flat earth model, we are able to treat the isostatic problem analytically, using the Fourier transform technique. The mantle is treated as a layered Newtonian half-space, causing the rate of displacement to vary with the wavelength of Fourier load harmonics. The elastic lithosphere is treated as a low-pass filter, because loads of small size tend to be balanced by the lithosphere itself, and not by buoyancy. The lithosphere flexure also speeds the isostatic response. A detailed mathematical description of the model is given in Fjeldskaar and Cathles (1991a).

A high resolution study based on data from tilting of raised shorelines at particular locations and the present rate of uplift in Fennoscandia, using the approach described above, indicates a mechanical lithosphere less than 50 km thick, a mantle viscosity of 1.0×10^{21} Pas and an asthenosphere of 75 km with a viscosity of 1.3×10^{19} Pas (Fjeldskaar and Cathles, 1991a).

Hydro-isostasy, the isostatic compensation due to changes in water-load, is included in the calculations. The change in water-load is accounted for indirectly by the Fourier transform technique, because the technique requires a load redistribution. With adjustment of the appropriate computational box, the modelled melting of the ice gives a sea level curve (Fig. 7) roughly in accordance with published eustatic curves (Fairbanks, 1989).

The present rate of uplift relative to mean sea level in Scandinavia increases from 0 mm year^{-1} along the western coast of Norway to 9 mm year^{-1} in central parts of Sweden (Fig. 6). For the uplift of the crust relative to the geoid, the observed rate of uplift has to be corrected for eustatic sea level change, which would probably add approximately 1 mm to the numbers given in Fig. 6 (Lambeck and Nakiboglu, 1984).

In our model we have used data from Denton and Hughes (1981) and ice thickness values obtained from theoretical profiles (Paterson, 1981), which are based on rocks as substratum. A marine based ice sheet, however, will expand over older glacial and younger marine deposits, which form a soft, deformable substratum. Paleoclimatic considerations imply that the base of most of the ice sheet was at the pressure melting point (Løvø *et al.*, 1990), and the existence of glacial flutes shows clearly that the substratum was deformed. From field experiments on Iceland, Boulton and Jones (1979) found that as much as 90% of the glacier movement could take place in a thin, subglacial deformable till layer, and that this created ice profiles considerably lower than theoretical ones. The results from Ice Stream B in Antarctica indicate that the entire glacial advance takes place in the substratum, with no relative movement between the ice and the sediment (Alley *et al.*, 1986), and hence confirm the findings of Boulton and Jones (1979). If the Barents Sea Ice Sheet was drained by fast-flowing ice streams, parts of the ice sheet may have been relatively thin, compared to most theoretical ice sheet profiles. Any estimate of total thickness would be speculative, given the present data, but, as will be seen from the following discussion, the total time for maximum ice extension is too brief for thickness estimates to be a significant source of error in calculations of the isostatic response.

To test the timing of the deglaciation, we have varied the ages of deglaciation stages II and III, and kept the ages of the additional stages constant. The theoretical uplift pattern based on ages of 15 ka and 13 ka for stages II and III, respectively, is shown in Fig. 8a. The deglaciation model assumes an almost complete deglaciation of the Barents Sea at 13 ka BP. This deglaciation history requires a mantle of viscosity 1.0×10^{21} Pas overlain by a 75 km asthenosphere with a viscosity of 1.2×10^{19} Pas. There are clearly significant discrepancies between observed and theoretical present rates of uplift. At the northernmost point of

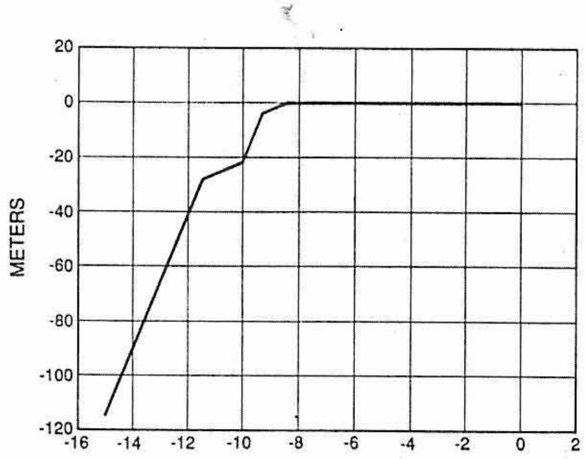

FIG. 7. The eustatic curve used in the calculations.

(a)

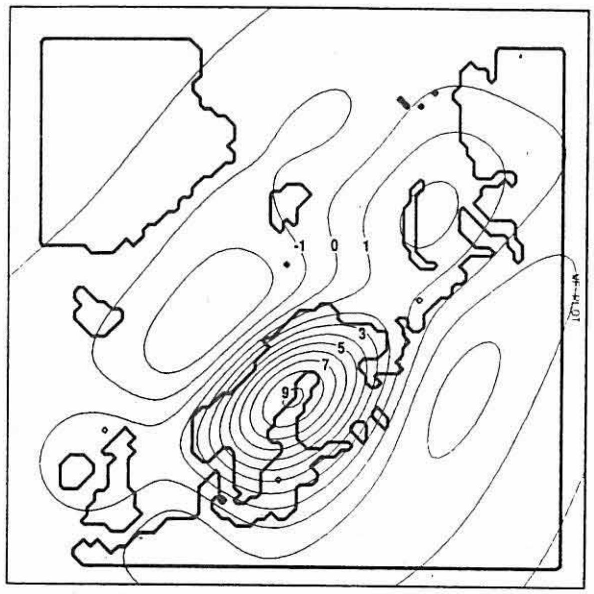

FIG. 8a. Theoretical present rate of uplift, assuming ages of 15 ka BP and 13 ka BP for stages II and III, respectively.

(b)

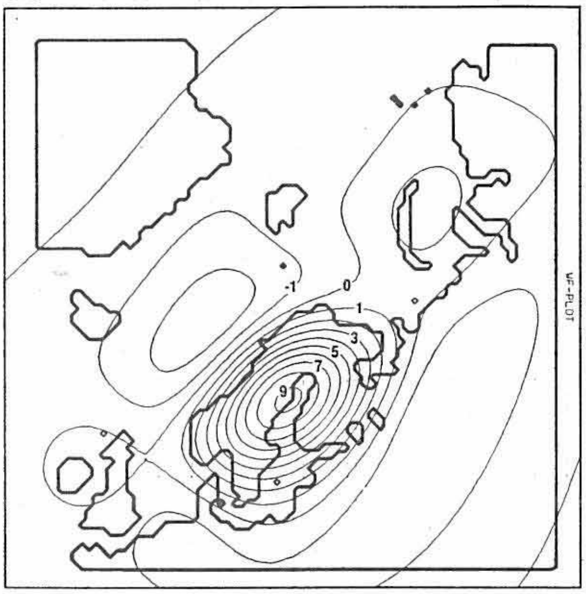

FIG. 8b. Theoretical present rate of uplift, assuming ages of 17 ka BP and 15 ka for stages II and III, respectively.

(c)

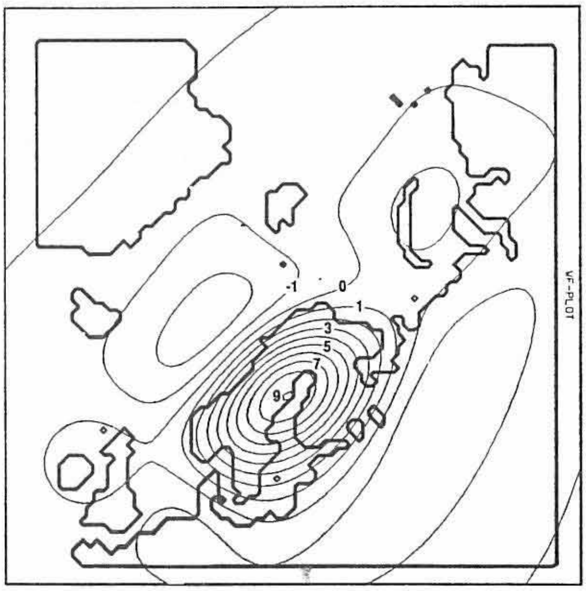

FIG. 8c. Theoretical present rate of uplift, assuming ages 17 ka BP and 15 ka BP for stages II and III, respectively, and a load cycle of 10 ka.

Fennoscandia, the observed 0-isoline for the uplift bends towards the east, while the theoretical isoline has a northerly direction. This indicates an earlier deglaciation of the Barents Sea than modelled.

The deglaciation history modelled in the following is changed slightly to optimize the fit between observed and theoretical present rates of uplift in northern Norway. During the calculations, the maximum uplift rate at Fennoscandia was kept at 9.0 mm year^{-1}, while the viscosity profile of the mantle was allowed to vary. The new deglaciation model has ages of 17 ka and 15 ka for stages II and III, respectively. This deglaciation history implies an almost fully deglaciated Barents sea as early as 15 ka BP. This new deglaciation history requires a mantle with a viscosity of 1.3×10^{21} Pas, overlain by a 75 km thick asthenosphere of viscosity 1.3×10^{19} Pas, and it gives a much better fit for northern Norway (Fig. 8b). Because the period of glaciation in the Barents Sea was probably of short duration, less than 10,000 years, the above deglaciation history was also modelled using a load cycle of 10,000 years (for model description, see Fjeldskaar and Cathles, 1991b). This means that the maximum ice sheet extension was applied to the surface only for this period of time. The resulting present rate of uplift pattern is very similar to that with no load cycle (Fig. 8c).

This reconstruction model seems to correspond best with observations of ice recession outlined on the basis of geological data. Thus, the isostatic modelling seems to support the idea of significant deglaciation of the Barents Sea as early as 15 ka BP.

CONCLUDING REMARKS

Compared to the Fennoscandian Ice Sheet, which may have existed in inland Scandinavia for most of the Weichselian (Andersen and Mangerud, 1989), the duration of the Barents Sea glaciation was relatively short. The available radiocarbon dates indicate glacial growth after 23 ka BP (Fig. 5), and according to the various models of ice recession, significant parts of the Barents Sea were deglaciated at 15 ka BP. The entire cycle of the Late Weichselian Barents Sea Ice Sheet was in the range of 10 ka, which corresponds to the duration of the last glaciation on Svalbard. According to the model outlined, the Barents Sea Ice Sheet represents a glacial system which developed and existed under periods of maximum climatic deterioration. The growth was controlled by sea level lowering and exposure of land, while the deglaciation and the pattern of ice recession was a function of sea level rise. So far, no data have been presented which support the idea of a rapid deglaciation of the entire Barents Sea. At this stage, it seems more reasonable to assume a step-wise ice recession.

The established model of glaciation has been tested by means of theoretical and observed present rates of uplift in the Finnmark region, northern Norway. The theoretical present rate of uplift, assuming a load cycle of 10,000 years, gives an uplift pattern very similar to that attained when isostatic equilibrium at ice maximum is assumed. Observed data and calculated glacial isostatic response suggest an early deglaciation of the Barents Sea: that ice, that significant parts of the Barents Sea were ice-free at about 15 ka BP. A later deglaciation gives a higher present rate of uplift than observed. This geophysical test thus shows that a glaciation model which incorporates significant deglaciation at about 15 ka BP accords best with the present rate of uplift.

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