Chapter 3

Paper II: Modelling Late Cenozoic isostatic elevation changes in the Barents Sea and their implications for oceanic and climatic regimes: preliminary results

Modelling Late Cenozoic isostatic elevation changes in the Barents Sea and their implications for oceanic and climatic regimes: preliminary results

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Abstract Late Cenozoic isostatic changes in the elevation of the Barents Sea are simulated using a numerical model. Isopach maps of the deposits off present-day Bear Island and Storfjorden troughs made earlier are used to calculate the thickness of sediment cover removed from the respective drainages basins at various time intervals during the last 2.3 Ma. Results indicate that Barents Sea was subaerially exposed at 2.3 Ma and major parts of it became submarine after 1 Ma. Barents Sea today receives around 40 % of the warm and saline North Atlantic waters flowing into the Greenland-Scotland Ridge and about half of the Atlantic water entering the Arctic Ocean. It thus has an important role to play in the present-day ocean circulation pattern in the Polar North Atlantic region and water-mass transformations that take place in the Greenland-Iceland-Norwegian Sea and the Arctic Ocean. The effects of an uplifted Barents Sea on the oceanic regime and the Arctic sea-ice cover under the present-day forcings fields are studied using the Miami Isopycnic Coordinate Ocean Model. Preliminary results indicate that a subaerial Barents Sea causes an increased input of warm Atlantic waters into the Arctic Ocean through the Fram Strait which results in warming of the Atlantic water masses in the Arctic Ocean, followed by a reduction in the sea-ice cover. The obtained findings can be used to explain the apparent discrepancy in the late Cenozoic record of the sub-Arctic and Arctic regions whereby Fennoscandia, Iceland and Greenland are envisaged to have been covered by major ice sheets during late Pliocene whereas high Arctic areas such as Svalbard and NE Greenland were apparently free of any major ice.

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3.1 Introduction

The present-day climate at high northern latitudes is regulated by the advection of warm North Atlantic waters into the Nordic Seas (Fig. 3.1) and the water-mass transformations that take place therein. The Barents Sea receives about 40 % of these northward flowing waters and is thus an important part of the system (e.g., Simonsen and Haugan 1996). It has since long been recognised that Barents Sea was subaerially exposed during pre-Quaternary times (e.g., Nansen 1904, 1920). However, the transition from a continental region to an epicontinental shelf sea, in terms of oceanographic patterns and climatic evolution in the Polar North Atlantic, has received little attention.

In the present study, temporal shifts in the elevation of the Barents Sea are simulated using an isostatic model by Dimakis et al. (1998). The impact of an exposed Barents Shelf on the general ocean circulation and the thermodynamics in the region is then studied by a high latitude version (Drange 1999) of the prognostic, three-dimensional ocean general circulation model (MICOM–Miami Isopycnic Coordinate Ocean Model; Bleck et al. 1992).

Such a study has become all the more relevant in view of the apparent discrepancies in the palaeoclimatic reconstructions of the Polar North Atlantic region during the Plio-Pleistocene. Major parts of Fennoscandia, Iceland and Greenland are envisaged to have been covered by large ice masses around 2.6 Ma (e.g., Jansen and Sjøholm 1991) whereas high-Arctic areas, such as NE Greenland and Svalbard, were probably free of any major ice accumulations at this time (e.g., Funder et al. 1985; Butt et al. 2000).

Probably, the most significant aspect of the ocean circulation pattern in high northern latitudes is the formation of deep waters in the Greenland-Iceland-Norwegian (GIN) Sea, and this convection is believed to drive much of the global overturning cell in the ocean (Aagard and Carmack 1994). While the actual mechanism remains uncertain, it has been widely recognised that the magnitude and consistency of the North Atlantic Current (NAC), transporting warm and saline waters into the GIN Sea through the Faroe-Shetland Channel, is essential to the thermohaline balance and circulation patterns in the Polar North Atlantic (Hopkins 1991) In moving from the Faroe Shetland Channel to Svalbard, the temperature of NAC drops by 4-5 $^{\circ}$C (Johannessen 1986). This drop in temperature results in the release of $\sim$59 TW of heat to the atmosphere (Simonsen and Haugan 1996) which is instrumental in moderating the present-day climate of Europe (Broecker et al. 1985).

Any perturbations in North Atlantic forcing are likely to affect the thermohaline circulation and consequently, the climate in the region. A weakening of NAC is likely to lead to cooling and a southward shift in convective regions as well as a southward expansion in Arctic ice. On the other hand, if the strength of the NAC were to increase, then it will cause heat transport farther north and a consequent decrease in ice cover (Aagard and Carmack 1994).

Estimates of the average Atlantic water inflow into the Barents Sea vary by a factor of two, ranging from less than 3 Sv to more than 5 Sv (Simonsen and Haugan 1996). Reported values of the northward volume transport of Atlantic water through the Fram Strait with the West Spitsbergen Current (WSC) are also rather uncertain, varying from 2 to 6 Sv (Simonsen and Haugan 1996). Two recent estimates indicate that the transport is in the range 2.9-3.7 Sv (Jónsson and Foldvik 1992; Rudels et al. 1994). This means that the Barents Sea may receive about one-third of the total North Atlantic waters that flow into the GIN Sea over the Greenland-

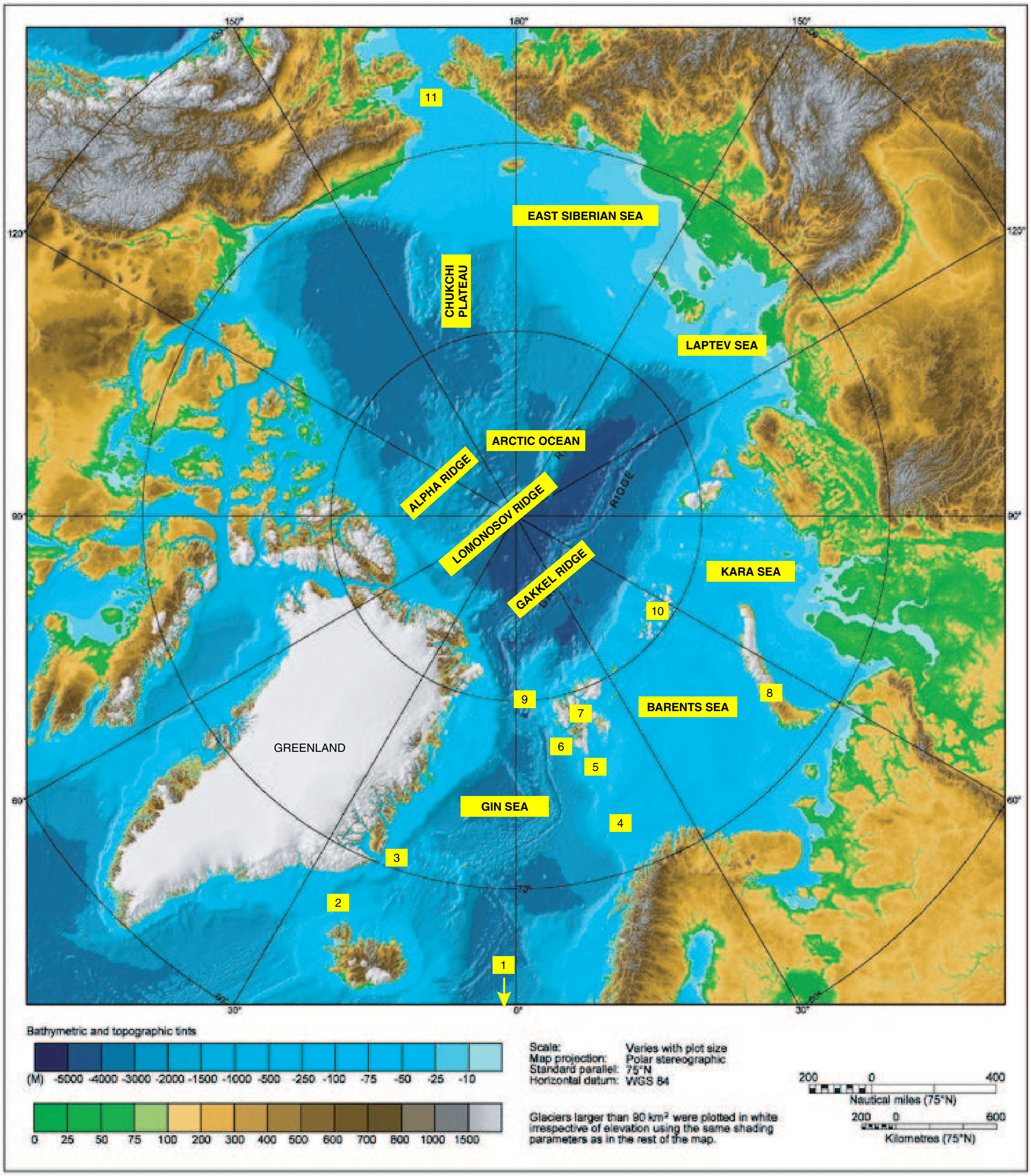

Figure 3.1: Map over the Nordic Seas and the Arctic Ocean. 1, Faroe-Shetland Channel (to the south as indicated by the arrow); 2, Denmark Strait; 3, ODP Site 987 (approximate location); 4, Bear Island Trough; 5, Storfjorden Trough; 6, ODP Site 986 (approximate location); 7, Svalbard; 8, Novaya Zemlya; 9, Fram Strait; 10, Franz Josef Land; 11, Bering Strait (modified from IBCAO 2000).

Scotland Ridge in the present-day situation (8 Sv; Hansen and Østerhus 2000) and about half of the Atlantic water entering the Arctic Ocean. The Barents Sea thus, represents a significant heat and salt sink for the region. Furthermore, heat loss and brine rejection during winter leads to an increase in the density of the Atlantic waters in the Barents Sea. This modified Atlantic water enters the Arctic Ocean from the Barents Sea partly through the opening between Novaya Zemlya and Franz Josef Land, but mostly via the Kara Sea (Pfirman et al. 1994) (Fig. 3.1). The Barents Sea, therefore, acts as a mediator for the formation of dense water in the region.

Outflow of cold, dense bottom waters formed in the Barents Sea back into the GIN Sea takes place in the northern parts of the Bear Island Trough (Swift 1986; Midttun and Loeng 1987; Blindheim 1989; Midttun 1989) and through Storfjorden (Quadfasel et al. 1988). The effect of Barents Sea inputs on GIN Sea water masses and hence the WSC is not clear. But, if these inputs, including the export of sea-ice, are considered to modify the heat and salt content of the WSC, then the Barents Sea, besides contributing directly to the Arctic Ocean, also has an important indirect effect on the water masses of the Arctic Ocean (Hopkins 1991).

In view of the role of the Barents Sea in the present-day ocean circulation pattern and based on some initial estimates, it is suggested that the late Cenozoic elevation changes in the Barents Sea could have played a key role in the evolution of the Polar North Atlantic palaeoclimate.

3.2 The Barents Sea

3.2.1 Location and sequence stratigraphy

The Barents Sea today is a shallow epicontinental sea with an average water depth of around 230 m and an area of about 1.2×10^6 km^2 (Fig. 3.1). It is bound in the north and west by Tertiary rift and shear margins (e.g., Faleide et al. 1993). The Novaya Zemlya region forms the eastern boundary whereas the Norwegian coast and the Kola Peninsula mark the southern boundary (Fig. 3.1). The bathymetry is characterised by banks and troughs.

The western Svalbard-Barents Sea Margin is characterised by large accumulations of sediments on the continental slope, believed to have resulted from late Cenozoic glaciations (e.g., Solheim et al. 1998). Seven regional seismic reflectors have been identified along this margin (Faleide et al. 1996) dividing the glacial sediments into six sequences (Fig. 3.2). The ages assigned to these reflectors (Tab. 3.1) follow Channel et al. (1999) and are based on palaeomagnetic data from Ocean Drilling Program (ODP) Site 986 (Fig. 3.1) with support from biostratigraphy (Eidvin and Nagy 1999). The age estimates assume that the sequence boundaries i.e., reflectors R1-R7, represent time lines. Main sediment accumulations are found off Storfjorden and Bear Island troughs.

3.2.2 Modelling late Cenozoic elevation changes in the Barents Sea

At least two significant phases of uplift and erosion in the northern North Atlantic region are known to have occurred during the Cenozoic: one during late Palaeogene and the second around the Quaternary. The first event is attributed to the emplacement of magma from the Iceland plume in and below the crust while isostacy associated with glaciations is believed to have been important in the second event (Japsen and Chalmers 2000). Whether or not glacio-isostacy was

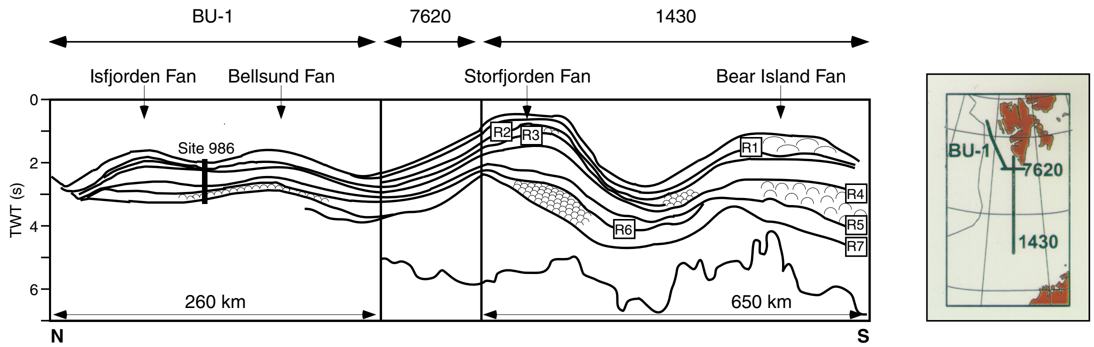

Figure 3.2: A cartoon showing the seismic stratigraphy of the Svalbard-Barents Sea Margin (modified from Solheim et al. 1998).

the only mechanism responsible for the late Cenozoic elevation changes is open to discussion. In the Barents Sea region, estimates of total Cenozoic erosion from different parts of the Barents Sea range from 500 to 3500 m with greatest erosion estimated for the northwestern parts (Hjelstuen et al. 1996; Rasmussen and Fjeldskaar 1996; Dimakis et al. 1998). Two-thirds of this erosion took place during the late Cenozoic (Eidvin and Riis 1989), the time of frequent glacial-interglacial changes in the region. Glacial erosion has thus been the most dominant even if not the only mechanism during the late Cenozoic. Consequently, in the current work, relief changes in the Barents Sea during the last $\sim$2.5 Myr are assumed to have taken place principally due to glacially driven exhumation.

For this work, a uniform grid consisting of 262 square cells, each measuring 50×50 km^2, was laid over a bathymetric map (drawn at 20 m contour intervals) of the Barents Sea (Fig. 3.3). The grid covers areas interpreted as drainage areas for sedimentary deposits off Storfjorden (Hjelstuen et al. 1996) and Bear Island Trough (Fiedler and Faleide 1996). The total area covered by the grid measures 655×10^3 km^2. The contour value at the centre of each grid cell was used to define the average contour value for the whole cell and based on this, the present-day bathymetry of the presumed drainage area was reconstructed (Fig. 3.4). The bathymetry so constructed is a fair representation of the actual present-day bathymetry (cf. Figs. 3.3 and 3.4).

Hjelstuen (2000) and Hjelstuen et al. (1996) have constructed isopach maps for deposits off Storfjorden while Fiedler and Faleide (1996) have presented the same for the sedimentary wedge off Bear Island Trough. Based on these estimates and dating by Channel et al. (1999), sediment volumes for each time interval between R7 and R1 were calculated using linear interpolation for intervals where estimates are not available (Tab. 3.1).

An isostatic model by Dimakis et al. (1998) was used for relief simulations. The model is based on Airy's isostatic theory and response of the lithosphere to loading/unloading (deposition/erosion) is described by Airy's isostatic equations. For a complete description of the equations and the way they are used, the reader is referred to the original paper by Dimakis et al. (1998).

Sediment volumes calculated for each of the seismic sequences in each of the two drainage basins (i.e., Storfjorden and Bear Island) were added together and divided by the respective drainage area. The resulting total sediment thickness for each basin was laid over its drainage area, as a slab of uniform thickness, to construct the relief of the Barents Sea at 2.3 Ma (R7

Interval	Age (Ma)	Sed. Vol. (km^3) depositional	Sed. Vol. (cm^3) depositional	DBD (g/cm^3) depositional	Sed. mass (g)	DBD (g/cm^3) drainage	Sed. Vol. (cm^3) drainage	Sed. thickness (m) drainage
R7-R6	2.3-1.6	3.40×10^{4}	3.4×10^{19}	1.9	6.6×10^{19}	2.2	3.0×10^{19}	432
R6-R5	1.6-1.4	2.24×10^{4}	2.2×10^{19}	1.7	3.8×10^{19}	2.2	1.7×10^{19}	249
R5-R4	1.4-1.0	2.04×10^{4}	2.0×10^{19}	1.6	3.2×10^{19}	2.6	1.2×10^{19}	181
R4-R3	1.0-0.8	1.02×10^{4}	1.0×10^{19}	1.4	1.5×10^{19}	2.6	5.6×10^{18}	82
R3-R2	0.8-0.5	8.60×10^{3}	8.6×10^{18}	1.4	1.2×10^{19}	2.7	4.3×10^{18}	62
R2-R1	0.5-0.2	8.60×10^{3}	8.6×10^{18}	1.3	1.1×10^{19}	2.7	4.0×10^{18}	59
R1-SF	0.2-	5.70×10^{3}	5.7×10^{18}	1.0	5.9×10^{18}	2.7	2.2×10^{18}	32

(a)

Interval	Age (Ma)	Sed. Vol. (km^3) depositional	Sed. Vol. (cm^3) depositional	DBD (g/cm^3) depositional	Sed. mass (g)	DBD (g/cm^3) drainage	Sed. Vol. (cm^3) drainage	Sed. thickness (m) drainage
R7-R6	2.3-1.6	7.70×10^{4}	7.7×10^{19}	1.8	1.4×10^{20}	1.8	7.7×10^{19}	134
R6-R5	1.6-1.4	2.20×10^{4}	2.2×10^{19}	1.8	4.0×10^{19}	1.8	2.2×10^{19}	38
R5-R4	1.4-1.0	6.40×10^{4}	6.4×10^{19}	1.6	1.0×10^{20}	2.0	5.1×10^{19}	89
R4-R3	1.0-0.8	3.20×10^{4}	3.2×10^{19}	1.6	5.1×10^{19}	2.0	2.6×10^{19}	44
R3-R2	0.8-0.5	4.80×10^{4}	4.8×10^{19}	1.6	7.7×10^{19}	2.0	3.8×10^{19}	67
R2-R1	0.5-0.2	4.80×10^{4}	4.8×10^{19}	1.6	7.7×10^{19}	2.0	3.8×10^{19}	67
R1-SF	0.2-	1.06×10^{5}	1.1×10^{20}	1.5	1.6×10^{20}	2.1	7.6×10^{19}	131

(b)

Table 3.1: Estimates used for calculating elevation changes in (a) the Storfjorden drainage basin and (b) the Bear Island drainage basin. DBD=dry bulk density, SF=seafloor.

time) (Fig. 3.5a).

In the next step, the estimated sediment volume for the 2.3-1.6 Ma (R7-R6) interval, for each basin, was removed from the projected R7 relief and the model was run again. This procedure was repeated for each of the time intervals. Results for selected intervals are shown in Fig. 3.5.

The results show that the Barents Sea region was totally subaerial at 2.3 Ma and significant part of it became submarine after 1.0 Ma (Fig. 3.5).

The weaknesses associated with this modelling approach and sensitivity of the model output to input parameters have been discussed earlier in a similar modelling study (Butt et al. 2001). Sensitivity analyses were carried out to account for possible errors in estimates of drainage area, sediment densities, which in turn affect the estimates of sediment volume/thickness added or removed from the drainage area. While the elevations at different time intervals varied with changes in these input parameters (greater sediment thickness ADDED=higher resulting topography, etc.) the timing of the transition from a primarily subaerial platform to a shelf sea was constrained to a relatively short time window of 200 kyr.

Figure 3.3: The modelled area along with the overlain grid. The Storfjorden drainage area is outlined by heavier lines. The rectangle marked by the dashed line covers the area shown in Figs. 3.4 and 3.5. X, Y and Z are marked for orientation with Figs. 3.4 and 3.5.

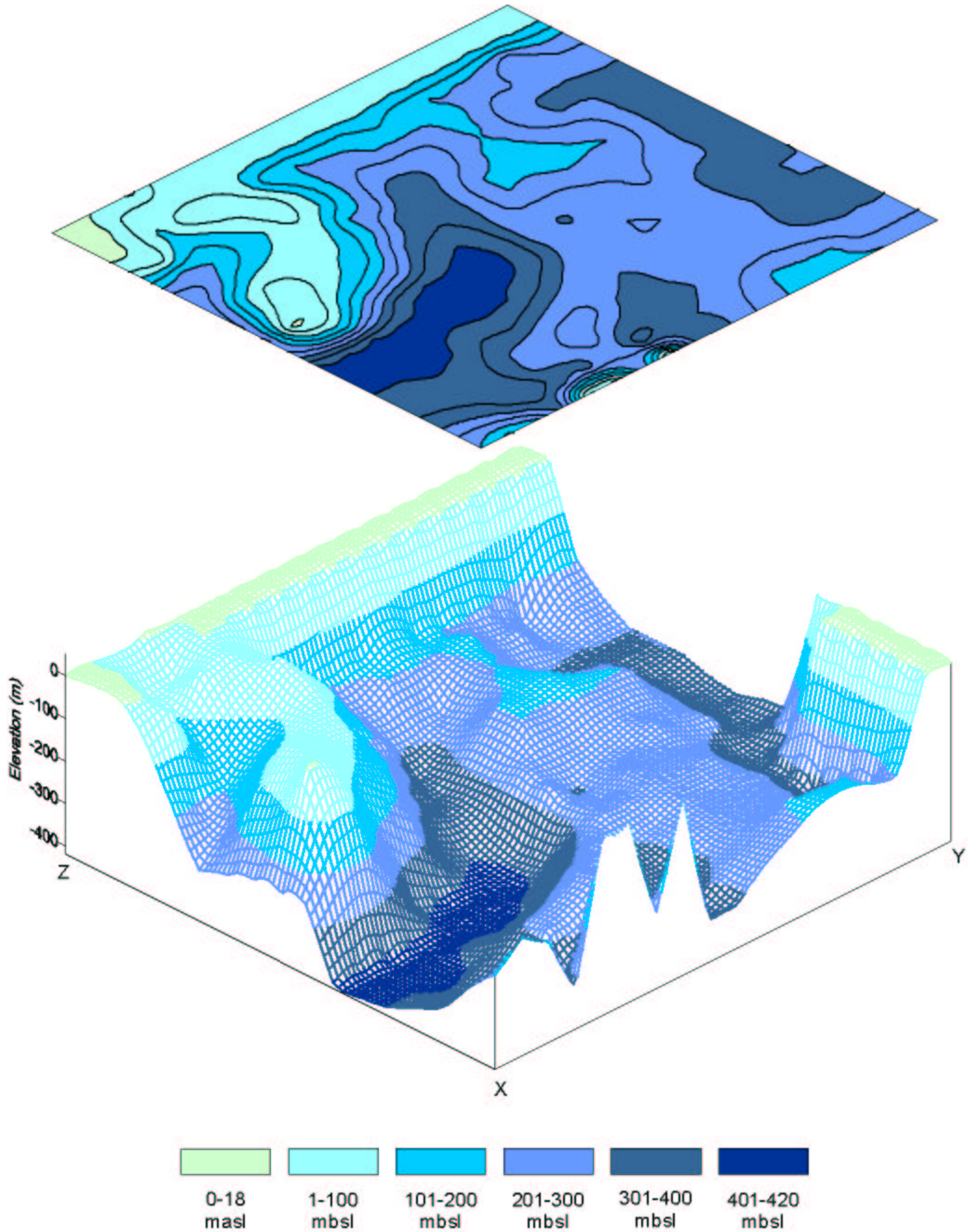

Figure 3.4: Present-day bathymetry of the modelled area as constructed from the grid. Areas lying within the dashed rectangle but outside the grid are kept at zero elevation for the plot.

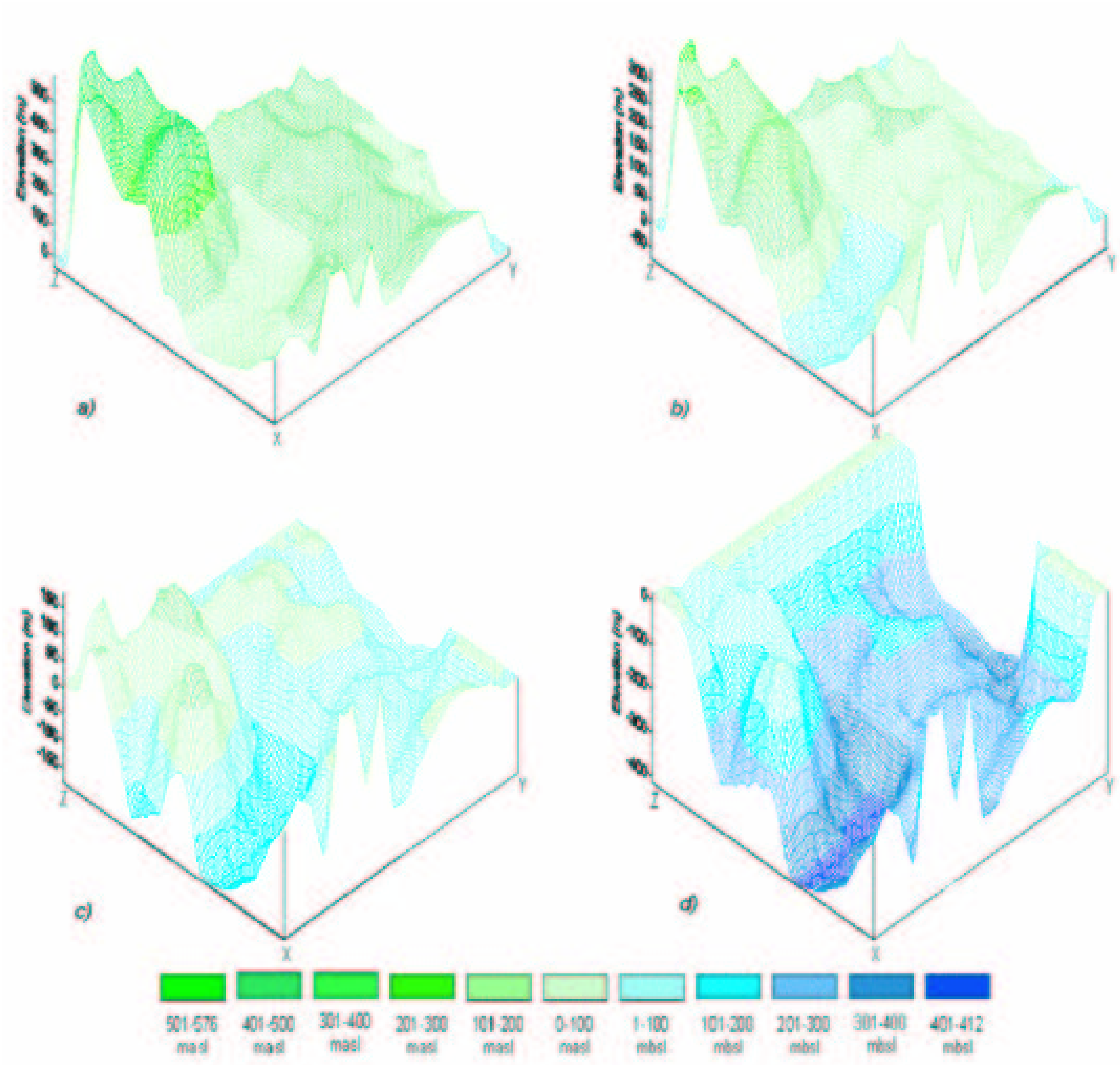

Figure 3.5: Simulated topography of the Barents Sea at selected time intervals. (a) 2.3 Ma, (b) 1.8 Ma, (c) 1.0 Ma and (d) present day (simulated). Areas lying within the dashed rectangle but outside the grid are kept at zero elevation for the plots.

3.3 Exposed Barents Shelf and the oceanic regime in the Nordic Seas

Based on this work and earlier studies (e.g., Rasmussen and Fjeldskaar 1996), a subaerially exposed Barents Sea during the Pliocene (i.e., before ∼2 Ma) is most likely. According to the model results, the Barents Sea dominantly became a shelf sea around 1 Ma (Fig. 3.1). It may thus be assumed that ocean circulation pattern similar to the present-day one is a phenomenon of the past ca. 1 Ma.

In addition to the elevation of the Barents Sea, there are several other parameters that need to be determined before an attempt can be made to ascertain the Pliocene oceanographic and climatic regimes. These include the palaeodepths of the openings that regulate exchange of water between the Arctic Ocean and the North Atlantic (Faroe-Shetland Ridge, Fram Strait, Denmark Strait) and the Arctic and the Pacific oceans (the Bering Strait), solar radiation balance, sea-level changes, paleobathymetry, slope angles of the continental margins bordering the GIN Sea, etc.

As a first step, we have chosen to simulate the effect of an emergent Barents Shelf given the present-day oceanic and climatic forcing fields whose values are comparatively well known (e.g., Simonsen and Haugan 1996; Hansen and Østerhus 2000). Simulating the Pliocene climate with a global coupled ocean-sea ice-atmosphere model (Furevik et al. 2000) is planned as the next step.

3.3.1 The ocean model

MICOM is an ocean general circulation model that utilises surfaces of constant density (isopycnic surfaces) as the vertical coordinate (Bleck et al. 1992). The density of the upper layer is, however, free to evolve according to changes in the heat and freshwater fluxes caused by the thermodynamic forcing and mixing of the surface water. Prognostic variables in all of the layers are the horizontal velocity components, the layer thickness, and temperature. In addition, salinity is a prognostic variable in the upper surface layer, whereas salinity is diagnosed from the equation of state of seawater for all of the subsurface layers.

In the configuration used in this study (Drange 1999), the reference pressure is set to zero, and the layer densities are expressed in σ_θ-units. There are 23 layers in the vertical, with σ_θ-values ranging from 24.05 to 28.11, and with a density spacing of 0.01 σ_θ-units for the most dense water masses. The layer densities have been chosen in order to describe the water mass characteristics of the major Atlantic-Arctic water masses.

The model domain covers the Atlantic Ocean from about 20°S and northwards, including the Arctic Ocean. The dynamic sea-ice model of Harder (1996) and the thermodynamic of Drange and Simonsen (1996) have been coupled to the ocean circulation model. The horizontal grid system is local orthogonal (Bentsen et al. 1999) with grid focus in the GIN Sea. The horizontal resolution in the GIN Sea varies from 55 to 65 km, whereas the grid spacing is about 250 km near the southern and northern model boundaries. The model topography represents present-day conditions and for each model grid cell, the ocean depth has been calculated as the arithmetic mean of the 5-min resolution topography database ETOPO5 (NOAA 1988).

The circulation model is initialised by observed hydrography (Levitus and Boyer 1994; Levitus et al. 1994)), and spun up from rest. As forcing fields, climatological monthly mean 10 m wind (ECMWF 1988), 2 m surface air temperature (ECMWF 1988; Simonsen and Haugan 1996), cloudiness (Huschke 1969; Oberhuber 1988), precipitation (Legates and Willmott 1990), and relative humidity (Maykut 1978; Oberhuber 1988) are used. The sea surface salinity (SSS) and temperature (SST) fields are relaxed towards observed monthly mean SSS and SST (Levitus et al. 1994; Levitus and Boyer 1994) with a relaxation time scale of 1 month for a 100 m deep mixed layer (New et al. 1995). Temperature and salinity relaxations are switched off in the part of the Arctic Ocean where ice is present at maximum ice extent (typically occurring in March). The lateral boundaries at 20°S and south of the Bering Strait are treated as walls with vanishing water transports.

For the first 5 years, the ocean dynamics and thermodynamics evolve according to inconsistencies between the initial density structure and the applied surface forcing. After an integration time of about 10 years, the evolution of the simulated dynamics and thermodynamics approaches an annually repeated cycle. As an example, the net northward and southward transport of water through the Barents Opening between Svalbard and Northern Norway is displayed in Fig. 3.6. It follows from the figure that the net northward and southward volume transports are fairly stable year by year, and that there is a mean, northward transport of about 3.25 Sv and a southward transport of about 0.7 Sv through the opening. The simulated value of the inflow compares well with the Blindheim (1989) estimate of 3.1 Sv, whereas the outflow is underestimated compared to the value of 1.2 Sv given by Blindheim (1989).

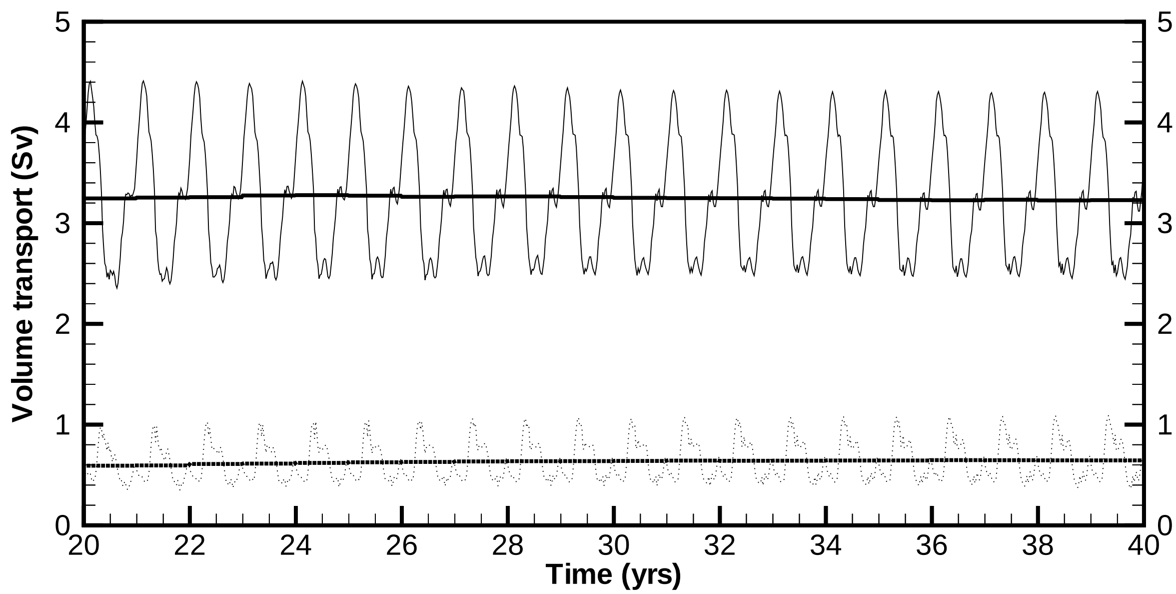

Figure 3.6: Time series of the simulated northward (solid line) and southward volume transport (in Sv) through the Barents Opening for the years 20-40 of the control run.

Computations of the volume transported through some of the standard sections in the North Atlantic and the GIN Sea show that the major current systems are in general agreement with the large-scale circulation in the region (Nansen 1906; Johannessen 1986; Hopkins 1991; Hansen and Østerhus 2000). One exception is that the net southward transport of water through the Denmark Strait (about 2 Sv) is at the lower end of current estimates (e.g., Worthington 1970; Hopkins 1991). The reason for the somewhat weak flow through the Denmark Strait is not clear, but could be linked to insufficient horizontal grid resolution in the strait, or that the climatological surface wind-forcing fields are too weak. Comparison of model and satellite derived sea-ice concentrations in the Arctic shows that the simulated sea-ice field picks up many of the observed features (Lisæter et al. 2000). We, therefore, conclude that the ocean model applied in this study, despite problems like the weak southward transport through the Denmark Strait, describes the present-day circulation regime in a fairly realistic way.

The ocean model experiment reported here has been performed as a quasi-twin experiment, starting from year 20 of the spin-up integration. As a base case (or control) integration, the spin-up integration was continued for 90 years. Over this time period, negligible model drifts were observed in the simulated fields (Fig. 3.6). In the second integration, the Barents Sea region, bounded by Tromsø in Northern Norway and the islands, Spitsbergen, Franz Josef Land and Novaya Zemlya, was defined as land. The differences between the two simulations can then be attributed to the effect of closing the Barents Sea for water transports. In reality, the climate system consists of a series of non-linear coupling mechanisms between the atmosphere, the land surface and the ocean-sea ice environment, so the obtained results should only be viewed as a first attempt to examine possible climate effects of a closed Barents Sea (see Sec. 3.4).

3.3.2 Results

When the Barents Sea is closed i.e., subaerially exposed, the northward flowing Atlantic water is forced to follow the continental margin between northern Norway and Svalbard, leading to an intensification of the WSC. In the control simulation using the present-day scenario, about 2 Sv (Fig. 3.7) of the Atlantic water enters the Arctic Ocean with the WSC, whereas the remaining water recirculates in the strait and flows southward towards the GIN Sea. The situation with a closed Barents Sea is similar, but now the inflow almost doubles to a value of about 3.85 Sv (Fig. 3.7). As this water enters the Arctic Basin, it subducts under the cold and fresh (and consequently low density) surface water.

The increased transport of the Atlantic water into the Arctic Ocean leads to heating of the subsurface waters. In a vertical section through the Fram Strait and towards Alaska (Fig. 3.8), maximum heating of more than 1.5 K occurs at depths between 200 and 400 m. The heating of the subsurface waters increases with time. This is seen from Fig. 3.8, and is evident from the evolution of the vertical temperature profiles at a station north of Greenland and at the North Pole (Fig. 3.9). An extended integration period would lead to even stronger heating of the polar waters, but such experiments should be performed by global models without the artificial lateral boundaries applied in the regional model used here.

The horizontal distribution of the Atlantic water in the upper part of the Arctic Ocean is not uniform. Figure 3.10 indicates strongest influence of this water, and consequently strongest heating, in a branch extending from the Fram Strait towards the Laptev Sea, and in a branch just north of Greenland. It is the core of the latter branch that is depicted in Fig. 3.8.

The simulated sea-ice evolves according to the heat balance at the top of the ice pack, and the horizontal and vertical inputs of heat into the upper ocean mixed layer. If ice is present and if the temperature of the surface mixed layer exceeds the melting temperature of sea-ice, ice is melted and the surface water temperature is reset to the melting temperature. This means that the temperature trace of the Atlantic water is removed from the surface water if ice melts. This, together with the subduction of the inflowing Atlantic water, is the reason for the near vanishing temperature change of the surface waters in Fig. 3.8.

It appears from Fig. 3.10 that the upper 50-100 m of the central Beaufort Gyre, located in the Canada Basin, is only weakly affected by the warm Atlantic water. However, the simulated sea-ice thickness indicates significant (about 1 m) melting of ice in the Beaufort Gyre after 90 years (Fig. 3.11). The reason for the melting is the persistent clockwise (anticyclonic) motion

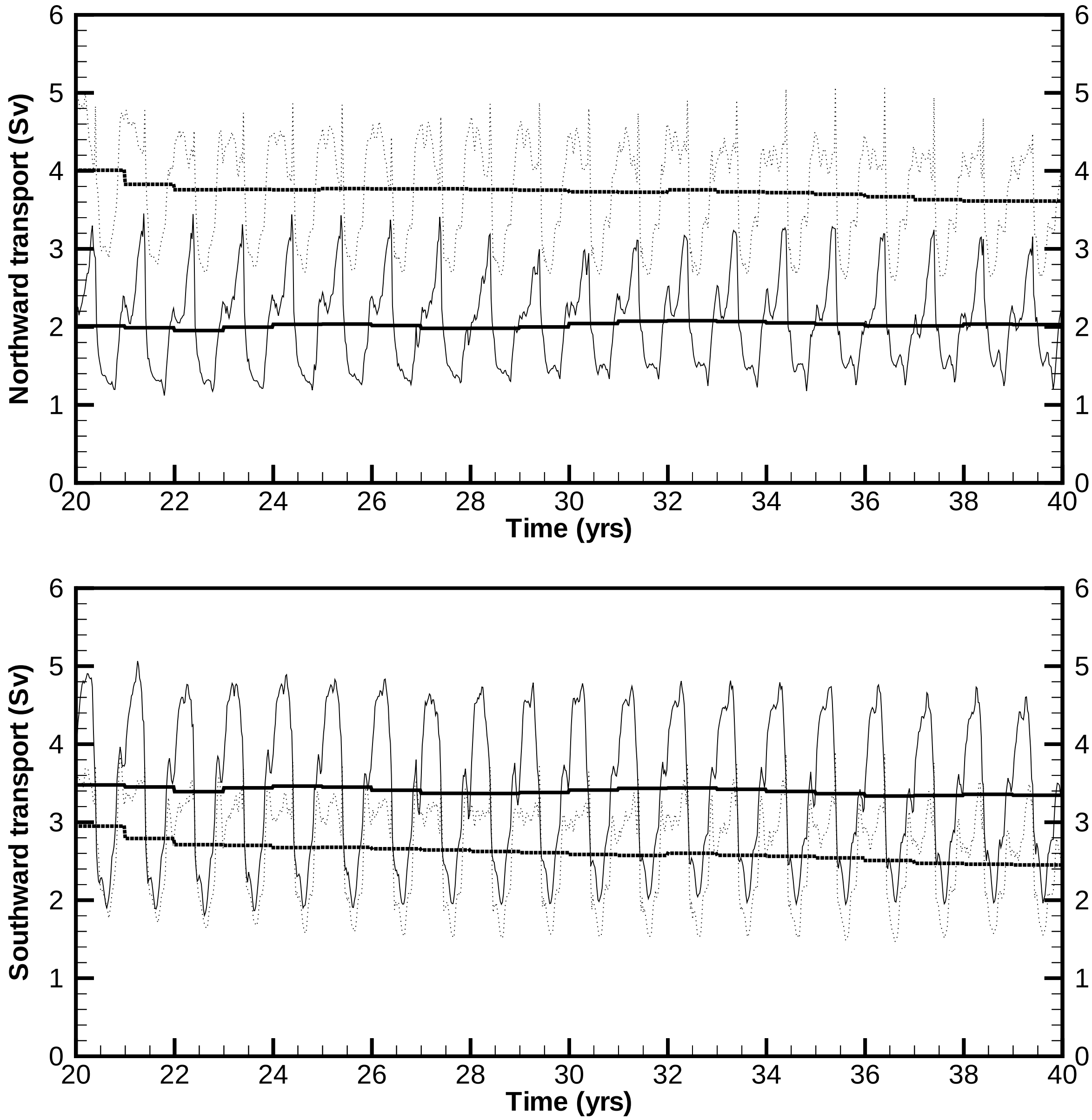

Figure 3.7: Simulated northward (upper panel) and southward (lower panel) transports (in Sv) of water through the Fram Strait. Solid (dotted) lines represent open (closed) Barents Sea.

of the ice and surface waters in the Beaufort Gyre (e.g., Barry et al. 1993), leading to transport of Atlantic waters from the periphery towards the centre of the gyre, followed by melting of ice, resulting in a sea surface temperature equal to the melting temperature of ice.

On the contrary, if the presence of the Atlantic water was confined to the Arctic Ocean surface mixed layer, melting of the Polar Ice Cap would occur and the temperature of the surface layer could exceed the melting temperature of ice. In the present simulation, this is the case in and north of the Fram Strait and along the coast of East Greenland throughout the year.

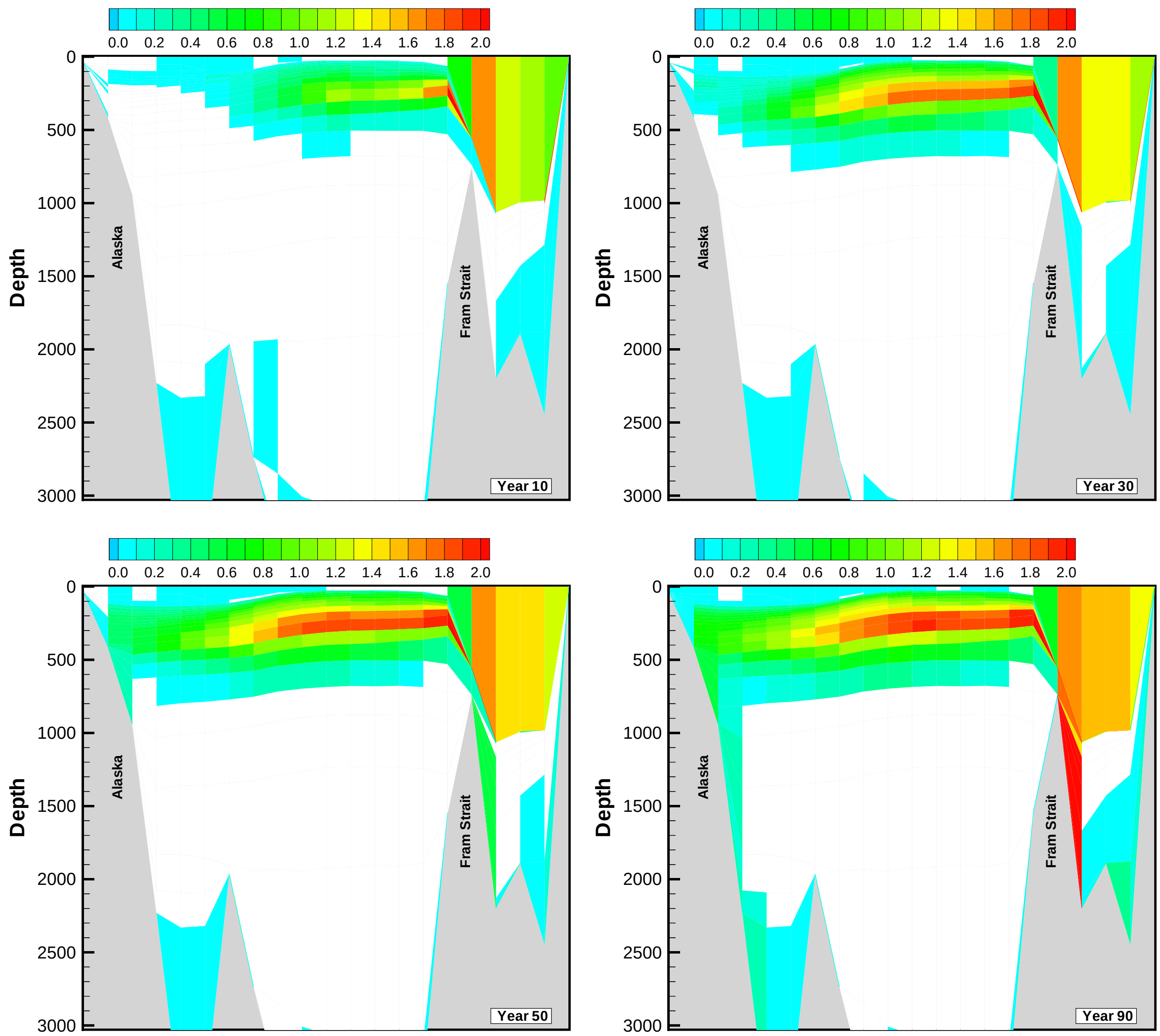

Figure 3.8: Heating (K) of the waters in the Arctic Ocean along a vertical section through the Fram Strait towards Alaska in years 10, 30, 50 and 90 of the simulation with closed Barents Sea.

The open water regions, particularly during the cold winter and spring seasons, will feed large amounts of heat into the atmosphere. The climate effect of such heat sources is hard to assess without incorporating it into an atmosphere circulation model. As an indication of the amount of heat released from open waters in Polar regions, flux measurements from polynias in the Arctic Basin show a heat loss of more than 500 Wm^{-2} (Dethleff 1994). The climate effect of ice-free waters in the Arctic Ocean may, therefore, be of importance for the local climate.

The modelled sea-ice response to the increased inflow of Atlantic water to the Arctic Basin is a general reduction in the thickness and extent of the ice. The only exception to this is a region north of Franz Josef Land where sea-ice accumulates due to pile-up of ice towards the new coastline. It follows from Fig. 3.11 that the reduction in ice thickness is larger in summer (represented by September) than in winter (March).

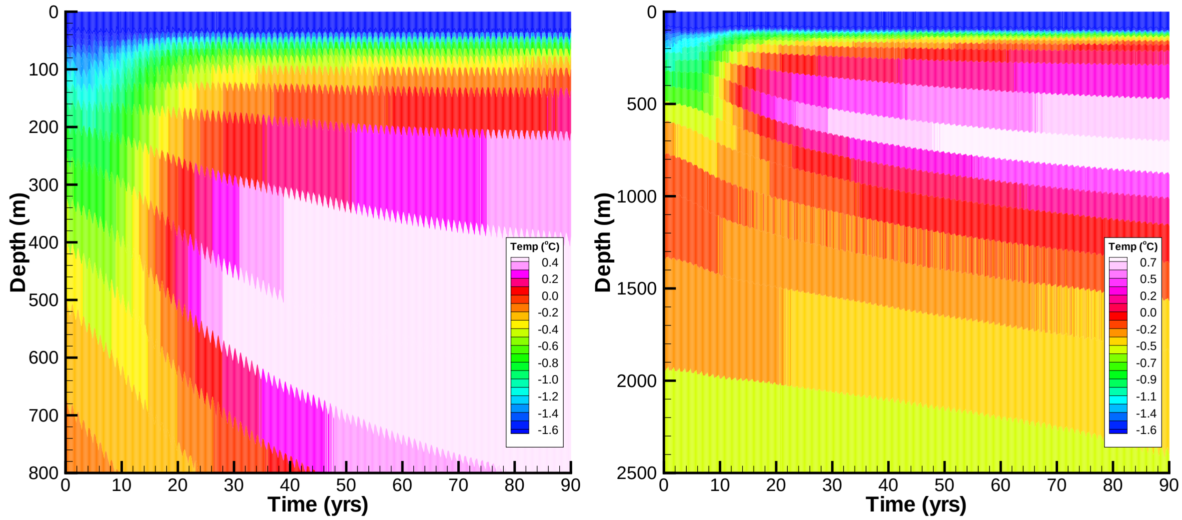

Figure 3.9: Evolution of the vertical temperature (°C) profile at a location north of Greenland (left panel: 85°N 118°W) and in the Central Arctic (right panel: 86°N 13°W).

3.4 Discussion

The modelled elevation changes in the Barents Sea and their impact on ocean circulation in high northern latitudes are relevant for heat transport and mode of deepwater formation in the Nordic Seas and on the extent of sea-ice formation in the Arctic Ocean.

The role of Arctic Ocean in regulating global climate, particularly, during the Cenozoic when the Earth's supercontinents were tectonically rearranged, has been a topic of intense debate for the past several decades. The theories have varied from scenarios where the Arctic Ocean is deemed to exert primary control on Northern Hemisphere climate through regulation of moisture supply from an ice-free Arctic (Ewing and Donn 1956) or alternatively through the presence of an Antarctic-size ice cap regulating the sea-level fluctuations and δ^{18} variations observed during the Pleistocene (Mercer 1970; Broecker 1975; Keigwin 1982). In other instances, the Arctic Ocean is suggested to have initially merely responded to orbital forcing just like other parts of the Earth (e.g., Boyd et al. 1984) and then by a series of feedback mechanisms made a significant impact on maintaining world climate.

Likewise, the initial formation of the Arctic ice cover and its temporal variations have been a controversial topic and this issue is important because both the existence and extent of Arctic sea-ice cover are vital for processes such as heat exchange with the atmosphere and determining the site of deepwater formation which in turn can impact climate on regional/global scale. Clarke (1982) predicts a more or less consistent ice cover during the past 4 Ma and while conditions may have varied through glacial, deglacial and interglacial periods (Clarke and Morris 1985), the Arctic Ocean is not believed to have been ice-free following the initial freezing (Untersteiner 1969). On the other hand, the Pliocene-Holocene interval is suggested to have comprised of three main phases (Herman and Hopkins 1980; Margolis and Herman 1980; Worsley and Herman 1980) including a cold but ice-free Arctic from ca. 4.0-2.7 Ma, stratified saline and fresh water masses from 2.7-0.7 Ma and the 0.7 Ma-present interval characterised

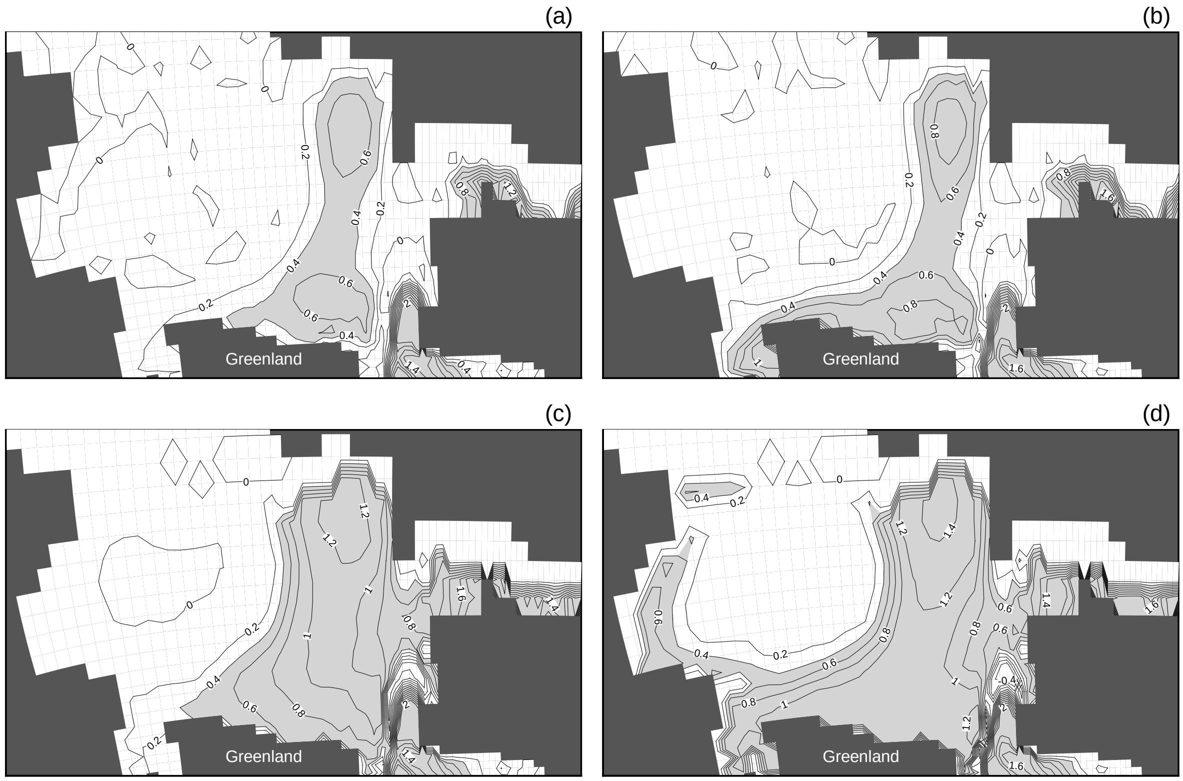

Figure 3.10: Heating (K) of the waters in the Arctic Ocean at 50 m depth after (a) 20 years, (b) 90 years, and at 100 m depth after (c) 20 years and (d) 90 years. Contour intervals 0.2 K, shaded area = heating > 0.4 K.

by the formation of first ice cover.

The major drawback in the interpretation of the Arctic data is the lack of reliable dating. Climatic fluctuations have been recognised in the Arctic sedimentary record based on the changes in texture, composition, structures, palaeontological and chemical content (Clarke et al. 1980; Morris et al. 1985; Mudie 1985; Spielhagen et al. 1997; Jackobsson et al. 2000). Chronological models used have been based upon palaeomagnetic reversal stratigraphy (e.g., Schneider et al. 1996), and dating methods such as ^{10}Be (e.g., Aldahan et al. 1997), amino acids (Sejrup et al. 1984) or lithostratigraphy (e.g., Clarke et al. 1980) and interpolation between these and ^{14}C dating (Spielhagen et al. 1997). Clarke (1990) suggested that changes in the nature of the Arctic ice cover might be correlated with climatic events in the North Atlantic. However, even recent interpretations of data from the same area within the Arctic Ocean differ widely owing to different age estimates (e.g., Spielhagen et al. 1997; Jackobsson et al. 2000).

Furthermore, interpretations of the Plio-Pleistocene record from the Arctic Ocean and marginal areas around it show markedly different climatic regimes than has been suggested from studies in the areas further to the south. There are reports of milder, sub-Arctic environments in NE Greenland around 2.45 Ma (Funder et al. 1985; Simonarson et al. 1998). Data from ODP Site 986 (off western Svalbard) do not bear any evidence of major glacial presence on Svalbard around this time (Smelror 1999; Butt et al. 2000). On the other hand, intensifica-

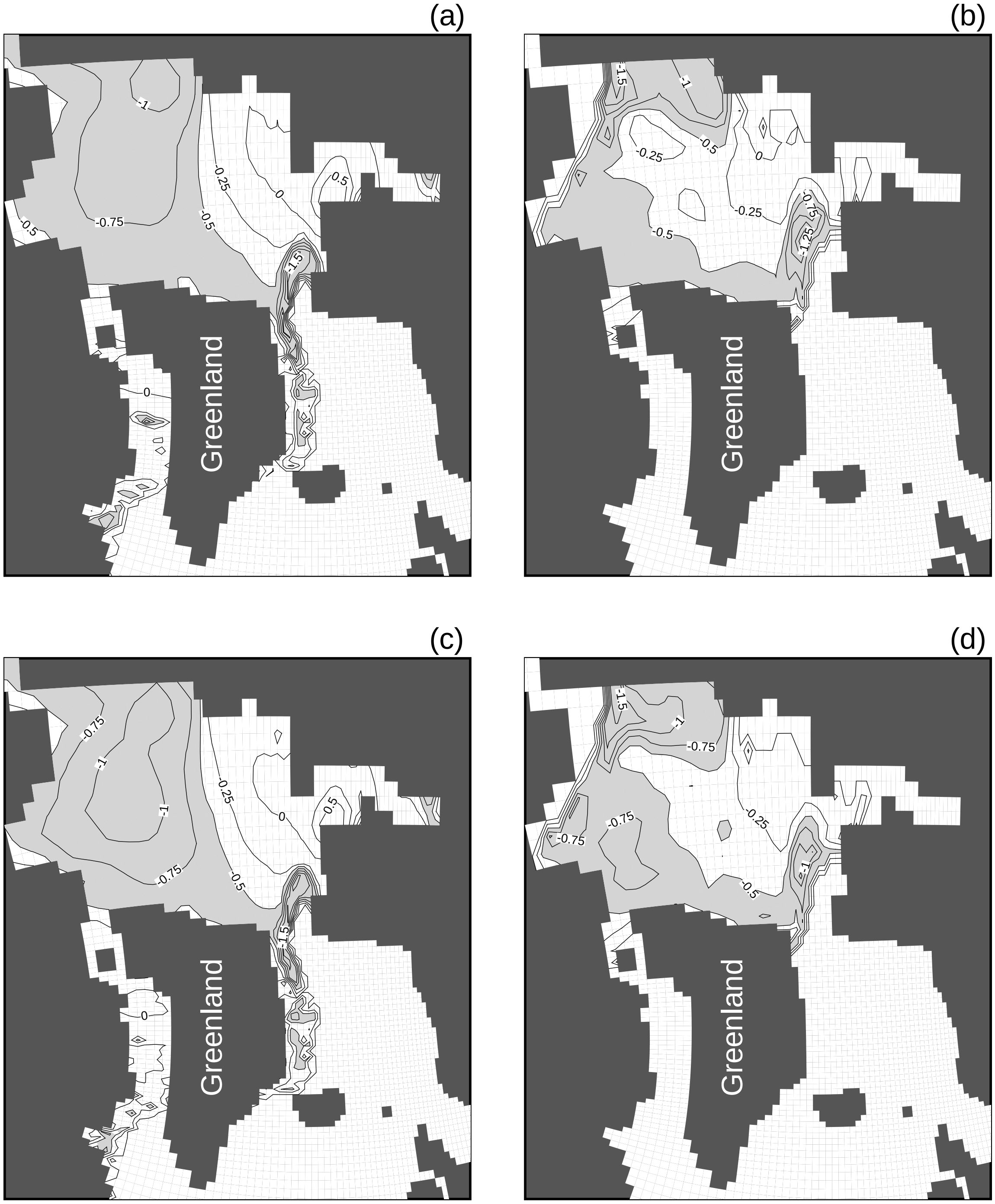

Figure 3.11: Changes in the simulated ice thickness (m) after 20 years in (a) March, (b) September, and after 90 years in (c) March and (d) September. Contour intervals: 0.25 K, shaded area = ice thickness reduction > 0.5 m.

tion in the late Cenozoic Northern Hemisphere glaciations is dated to around 2.6 Ma (Jansen and Sjøholm 1991; Tiedemann et al. 1994; Shackleton et al. 1995) whereby major ice sheets are envisaged to have occupied large areas over Greenland, Fennoscandia, Iceland, Svalbard and the Barents Sea.

This work enables us to comment on some earlier observations made in the Arctic and sub-Arctic regions that in our view are significant for interpreting the Plio-Pleistocene Arctic Ocean.

The modelling results presented herein indicate increased water and heat transport through the Fram Strait under a subaerial Barents Sea scenario (Fig. 3.8 and Fig. 3.8). This leads to warming of Arctic waters along two distinct paths; one along North Greenland and the other towards the Laptev Sea (Fig. 3.10) causing a reduction in ice thickness (Fig. 3.11). The current results thus provide a mechanism for explaining the findings of Funder et al. (1985) who, based on fauna and flora indicative of forest tundra environments found preserved in Kap København Formation, NE Greenland, proposed high-energy, marine environments during late Pliocene which Simonarson et al. (1998) date to 2.45 Ma. Funder et al. (1985) ruled out the possibility of a continuous perennial ice cover since the middle Cenozoic as proposed by Clarke (1982) and the modelling results tend to support that (Fig. 3.11).

The oceanographic simulations cover a period of ca. 100 years with a subaerial Barents Sea. The effects on a longer timescales are unclear but these preliminary findings suggest a seasonal, if any, Arctic sea-ice cover under such a scenario.

Butt et al. (2000) proposed onset of major glaciations on Svalbard around 1.6 Ma based on the data from ODP Site 986 off western Svalbard. Fig. 3.5 shows the development of the Bear Island Trough at this time allowing for water intake into the Barents Sea region. In theory, this would cause a reduction in the strength of the WSC flowing towards the Fram Strait and at the same time provide a moisture source for inner parts of the Barents Shelf. Major parts of the Barents Sea are seen to become submarine around 1 Ma and a major shift in Northern Hemisphere glaciations, the Mid-Pleistocene Revolution, is dated to around 0.9 Ma (Berger and Jansen 1994). There is thus room to speculate that late Cenozoic elevation changes in the Barents Sea may have played an important role in the palaeoclimatic evolution of the Polar North Atlantic region.

3.5 Concluding remarks

The modelling results presented here raise important issues regarding the late Cenozoic palaeoclimatic evolution in the high northern latitudes and highlight the potentially important role the elevation of the Barents Sea and temporal changes therein, may play in this regard. As mentioned earlier, the concept of a once subaerial Barents Sea is not new, but it has not been incorporated into the palaeoclimatic reconstructions for the Northern Hemisphere. The Barents Sea covered with a major ice sheet has been discussed (e.g., Siegert and Dowdeswell 1996) but not as a subaerial shelf.

Mullen and McNeil (1995), based on foraminiferal evidence from the central Arctic Ocean, indicate probable connection between the Arctic and North Atlantic oceans during Pliocene and suggest a cold but milder palaeoclimate in the Arctic Ocean during early Pliocene than that which existed during Pleistocene glacial intervals. Results from ODP Leg 151 similarly indicate at least periodic intrusions of warm Atlantic waters into the Fram Strait during Late Pliocene (Cronin and Whatley 1996; Osterman 1996; Spiegler 1996).

There is thus enough evidence to suggest inflow of warm waters in the Arctic and lower Arctic around 2.3 Ma. An uplifted Barents Sea at this time is a possible explanation for these observed warmer conditions. As results of ocean modelling indicate, such a scenario is likely to promote heat transfer to the Arctic Ocean with a consequent reduction in ice cover. The climatic consequences of a reduced ice cover have not been incorporated in the current work. However, in view of the role of the existing polynias in the Arctic Ocean and their role in heat exchange with the atmosphere (Aagard and Carmack 1994), open water conditions will cause increased heat exchange with the atmosphere over the Arctic Ocean and may lead to local, and possibly large scale, changes in climate.

The results, nevertheless, can only be regarded as preliminary at this stage. Several important related issues remain unanswered. These include, but are not limited to, the palaeodepths of the gateways, the atmospheric fields during late Pliocene (wind patterns, solar irradiance, etc.), the morphology of a subaerial Barents Shelf and its impact on atmospheric circulation and its drainage impact on the oceanic water masses, etc. Even the full climatic impacts under the present forcing fields cannot be understood without coupling ocean and sea-ice models with an atmospheric circulation model.

The final answers will, however, have to be provided by data from the Arctic Ocean itself. For this reason, it is important to come up with a reliable stratigraphy for the sedimentary record of the Arctic Ocean. Biostratigraphy holds the key in this regard. Lithostratigraphy and chemical properties of sediments are important, but reliable dating is a pre-requisite for a valid palaeoclimatic reconstruction of the Arctic Ocean and its comparison with the record from the Nordic Seas and elsewhere. Only then can the impacts of various processes be understood.

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