Early Holocene cooling events in Malangenfjord and the adjoining shelf, north-east Norwegian Sea

Katrine Husum & Morten Hald

A high resolution study of early Holocene climate and palaeoceanography has been performed on two combined sediment cores from Malangenfjord, northern Norway. The fjord provides a regional oceanographic climatic signal reflecting changes in the North Atlantic heat flux at this latitude because of its deep sill and the relatively narrow adjoining continental shelf. Fauna and stable oxygen and carbon isotopes indicate cool, meltwater-depleted water masses in the fjord from 12 000 to 11 400 cal. yr BP followed by a warming between 11 400 - 10 300 cal. yr BP. The climatic variability can be explained partly by freshwater forcing hampering the North Atlantic heat conveyor, and partly by changing solar irradiance. A major cooling event at 11 500 - 11 400 cal. yr BP, followed by a rapid warming, is correlated to the Preboreal Oscillation, a widespread signal in the North Atlantic region which is probably linked to the increased meltwater flux to the northern North Atlantic at this time. Brief and small-scale cooling events between 10 300 and 10 100 cal. yr BP, correlated to the onset of increased ^{10}Be flux in the Greenland ice cores, suggest a response to solar forcing.

K. Husum & M. Hald, Dept. of Geology, University of Tromsø, N-9037 Tromsø, Norway.

During the end of the last deglaciation large influxes of meltwater around the Nordic seas hampered thermohaline circulation, continuing into the early Holocene (Björck et al. 1996; Hald & Hagen 1998). Two climatically unstable periods—the so-called Preboreal Oscillation at 11 300 - 11 150 cal. yrs BP and a cooling event at 10 300 cal. yrs BP—are well documented around the Nordic seas (e.g. Björck et al. 2001). It has been suggested that solar-related changes have a greater effect on the climate system than previously believed and that these changes may be an important underlying mechanism for sub-Milankovitch climate fluctuations (Björck et al. 2001). To elucidate the causes of the cooling events it is crucial to obtain more detailed studies of the cooling events during the early Holocene. Here we present a high resolution record from Malangenfjord, on the fringe of the north-east Norwegian Sea. Communicating well with the Norwegian Sea on account of its deep sill and a relatively narrow adjoining continental shelf (Fig. 1), the fjord provides a regional oceanographic climatic signal.

Study area

Malangenfjord consists of an outer and inner basin of 450 and 250 m depths, respectively (Fig. 1). The fjord continues into Malangsdjupet, a transverse glacial trough on the shelf. A deep sill at 200 m marks the boundary between the fjord and the shelf trough (Fig. 1c). The river Målselv empties into the fjord head, draining a catchment area of approximately 6145 km^2 (Fig. 1b).

The Norwegian Atlantic Current transports relatively warm (> 8 °C), saline (> 35 psu) waters

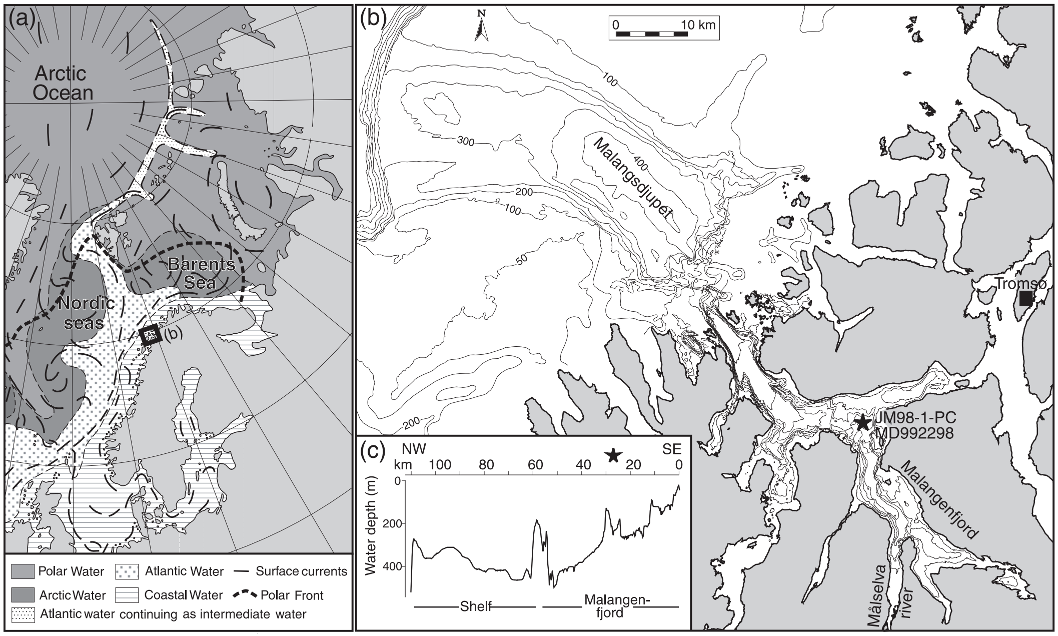

Fig. 1. (a) Index map showing surface waters in the Norwegian Sea and adjoining seas (modified from Mosby 1968). (b) Location and bathymetry of Malangenfjord, showing the position of the investigated piston cores. Bathymetric contours are given in metres. (c) A NW–SE depth profile of the fjord and adjoining shelf.

into the Norwegian Sea (Norwegian Atlantic Water) (Hopkins 1991) (Fig. 1a). It flows north along the Norwegian coast together with the Norwegian Coastal Current (Coastal Current Water) (Fig. 1a). The water masses of the Norwegian Coastal Current are characterized by temperatures of 2 - 13 °C and salinities of 32 - 35 psu (Hopkins 1991). Wedge-shaped in cross-section, the Norwegian Coastal Current becomes broader and shallower during the summer than during the winter (Sætre et al. 1988).

The Malangenfjord is stratified almost the entire year (Svendsen 1995). The main part of the water mass exchange is driven by changes in the vertical water mass density at the coast caused by coastal winds (Svendsen 1995). Local agents such as wind and run-off within the fjord (Svendsen 1995) drive only a minor part of the exchange. The bottom-waters of the fjord are renewed during the summer, when deeper and denser Atlantic water (from the Norwegian Atlantic Current) expands onto the shelf, mixes with the overlying water masses and flow into the fjord (Leth 1995). The bottom-waters of the fjord have temperatures between 4 and 7 °C and salinities between 34 and 35 (Normann 2001).

Materials and methods

The study is based on two overlapping piston cores (Figs. 1, 2). Core MD99-2298 (69° 29.72' N, 18° 23.03' E; water depth 210 m) is 35.85 m long. Core JM98-1-PC (69° 29.9' N, 18° 23.6' E, water depth 213 m) is 7.45 m long.

The size fraction 1 mm - 100 µm was used for foraminiferal analyses. There were fewer than 10 planktonic foraminifera in each sample, so they were only registered in connection with the benthic quantitative analysis. The total benthic foraminiferal flux was calculated assuming a constant average sediment density of 2 g/cm^3 for both cores.

Stable isotope measurements were carried out at the University of Bergen's GMS-Laboratory using a Finnigan MAT 25 isotope ratio mass spectrometer. The laboratory operates with a reproducibility of 0.07 ‰ for $\delta^{18}O$ measurements. The benthic stable isotope ratio was performed on *Cassidulina neoteretis* and *C. reniforme. C. neoteretis* calcifies its test in equilibrium with the ambient water in terms of oxygen isotope composition, and in terms of the carbon isotope composition the test is 1.5 ‰ lighter than equilibrium (Poole 1994). *C. reniforme* calcifies its test in disequilibrium in terms of oxygen isotope composition, with an offset of +0.13‰ (heavier) than equilibrium (Austin & Kroon 1996). The stable oxygen isotope values have also been corrected for the global ice volume effect following Fairbanks (1989) and Bard (1990). Some reservations must be expressed regarding the $\delta^{13}C$ values, as they have been measured on infaunal foraminiferal species. This implies that $\delta^{13}C$ reflects the pore water of the sediment which depends on both the bottom-water $\delta^{13}C$ values and the amount of depleted carbon that has been added to the pore water by organic matter decomposition (McCorkle et al. 1990). However, $\delta^{13}C$ measured on infaunal benthic foraminifera have proven useful as a bottom-water indicator in various settings (e.g. Hald et al. 2001).

Accelerator mass spectrometry (AMS) radiocarbon dating was performed on fossil molluscs, and measured at the Radiocarbon Laboratory in Trondheim, Norway, and the Leibniz Laboratory for Radiometric Dating and Isotope Research, Christian Albrechts University, Kiel, Germany (Table 1). All dates were calibrated using the INTCAL98 database (Stuiver et al. 1998). We are aware that the reservoir ages have ranged from ca. 800 to 700 years during the Younger Dryas and the Preboreal (e.g. Haflidason et al. 2000). A thorough estimate of reservoir ages has not been carried for our region, so we use an average delta R value of 464±35 for northern Norway (Stuiver et al. 1986). All ages are reported as calendar years BP, unless otherwise indicated.

Results

Lithostratigraphy

The two cores are divided into five and two lithostratigraphic units; for further details refer to Husum (2002). This study focuses on the early Holocene sediments belonging to the lithostratigraphic units II and III (Fig. 2). Unit III, only found in core MD99-2298 (ca. 22-11 m), is massive mud (clay and silt) with scattered clasts (Fig. 2). Unit II is found in both MD99-2298 (11-ca. 6 m) and JM98-1-PC (7.45-4.6 m) (Fig. 2), and is characterized by massive, bioturbated mud (clay and silt).

Depth–age models and correlation

The chronology of the compiled record is based on 11 AMS radiocarbon dates (Table 1) (Husum

Table 1. Radiocarbon dates from cores JM98-1-PC and MD99-2298. The radiocarbon dates have been calibrated with INTCAL98 (Stuiver et al. 1998) using a reservoir age of 464±35 for northern Norway (Stuiver et al. 1986). Where there is more than one intercept the middle value of the intercepts has been given, unless it is from an uncertain region or a linear extension of the calibration curve—marked with brackets: [ ].

Lab. code	Core ID	Depth (cm)	^{14}C age	Calibrated age (cal. BP)	2σ maximum cal. age (cal. age intercepts) minimum cal. age (cal. BP)	Material	Weight (mg)
TUa 2111	JM98-1-PC	488	9025±55	10128	10570 (10128) 9715	Fragments of 1 bivalve shell	180
TUa 2110	JM98-1-PC	739	9145±65	10281	10583 (10281) 9825	Fragments of 1 bivalve shell	100
TUa 2597	MD99-2298	768	9015±105	10117	10574 (10117) 9623	Fragment from gastropod	196.8
TUa 2598	MD99-2298	1180	9850±95	11138	11901 (11138, [10953], [10867]) 10649	1 bivalve shell: *Nuculana pemula*	106.4
TUa 2601	MD99-2298	1240	9310±70	10318	11122 (10318) 10116	1 bivalve shell: *Abra nitida*	8.5
TUa 2599	MD99-2298	1464	9605±65	11007	11297 (11015, 11007, 10819) 10343	Fragment of 1 bivalve shell (taxodont)	69
TUa 2600	MD99-2298	1767	9770±70	11111	11605 (11111, [10965], [10855]) 10551	1 bivalve shell: *Macoma pemula*	169.2
TUa 2602	MD99-2298	1867	10075±80	11359	12264 (11359) [10848]	1 bivalve shell: *N. pemula*	35.3
TUa 2594	MD99-2298	1875	10115±105	11495	12272 (11618, 11495, 11388) [10849]	Pair bivalve shells: *N. pemula*	282.8
TUa 2593	MD99-2298	2087	10220±90	11681	12307 (11681) [10870]	Pair bivalve shells: *Yoldiella lenticula*	14.5
KIA 9335	MD99-2298	2150.5	10315±55	11816	12801 (11926, 11816, 11749) 11278	Pair bivalve shells: *Y. lenticula*	15.3

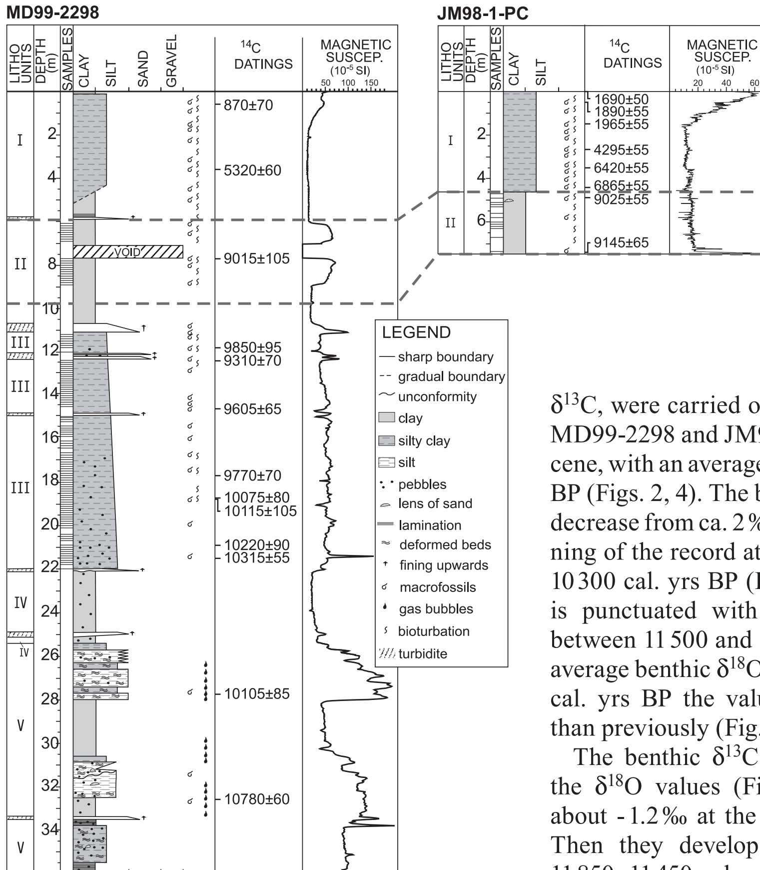

Fig. 2. Lithostratigraphy and magnetic susceptibility of the investigated piston cores JM98-1-PC and MD99-2298. The benthic stable isotope sample levels are also indicated. The dashed lines show the correlation between the investigated intervals in the two cores.

2002). Depth–age models were developed for the two cores following recommendations by Andrews et al. (1999) (Fig. 3). Further, the turbidites were assumed to represent zero time.

The early Holocene sediments, unit II from core JM98-1-PC and unit III and II from core MD99-2298, are correlated on basis of the lithology, magnetic susceptibility and the calendar age as shown in Fig. 2 (Husum 2002). The investigated unit II from core JM98-1-PC ranges from 10 285 to 10 111 cal. yrs BP (7.45 - 4.6 m) (Figs. 2, 3). The investigated units III and II from MD99-2298 range from 11 995 to 10 042 cal. yrs BP (ca. 22 - 6 m) (Figs. 2, 3). The sedimentation rates vary between 2.5 and 23.5 mm/yr during the early Holocene (Fig. 3), but these values should be treated with caution because of the limited age–depth control through some intervals.

Stable isotopes

Analyses of benthic stable isotopes, $\delta^{18}O$ and $\delta^{13}C$, were carried out on 130 samples from core MD99-2298 and JM98-1-PC from the early Holocene, with an average time resolution of 15 cal. yrs BP (Figs. 2, 4). The benthic $\delta^{18}O$ values generally decrease from ca. 2 ‰ to ca. 1.5 ‰ from the beginning of the record at ca. 12 000 cal. yrs BP to ca. 10 300 cal. yrs BP (Fig. 4). The decreasing trend is punctuated with a sudden heavier interval between 11 500 and 11 400 cal. yrs BP, where the average benthic $\delta^{18}O$ value is 2.3 ‰. At ca. 10 300 cal. yrs BP the values become slightly heavier than previously (Fig. 4).

The benthic $\delta^{13}C$ values fluctuate more than the $\delta^{18}O$ values (Fig. 4). The $\delta^{13}C$ values are about - 1.2 ‰ at the beginning of the Preboreal. Then they develop a broad minimum at ca. 11 850 - 11 450 cal. yrs BP, decreasing by 0.74 ‰. The $\delta^{13}C$ values are relatively stable until 11 300 cal. yrs BP, when they rapidly increase to a new stable value of ca. - 0.8 ‰ before decreasing again to - 1.5 ‰ at 10 200 cal. yrs BP. The values are low until another peak appears at 10 150 cal. yrs BP. After this point the $\delta^{13}C$ values are around -1 ‰.

Benthic foraminiferal fauna

A total of 31 foraminiferal samples from the early Holocene interval from core MD99-2298 and JM98-1-PC were analysed (Fig. 4). The resolution is between 3 and 285 (average ca. 60) cal. yrs BP (Fig. 4). The benthic foraminifera were well preserved and showed no signs of dissolution. 73 benthic calcareous species were registered. The frequencies of the most abundant benthic species are shown in Fig. 4.

Between ca. 12 000 and 11 400 cal. yrs BP the fauna is dominated by *Cassidulina reniforme* (Fig. 4). Important secondary species are *Cibicides lobatulus* and *Elphidium excavatum* f. *clavata*. The benthic flux rises from 15 to 150 species/

cm²/yr in this interval. After 11 400 cal. yrs BP *C. lobatulus* becomes the most frequent species. *Astrononion gallowayi*, *Trifarina angulosa*, and *Cassidulina neoteretis* also increase strongly (Fig. 4). The benthic flux has an average value of 220 species/cm²/yr until 10 300 cal. yrs BP.

The frequencies of the species fluctuate more after 10 300 cal. yrs BP, partly due to a better time resolution. *C. lobatulus* and *A. gallowayi* increase and both *C. reniforme* and *C. neoteretis* decrease (Fig. 4). The benthic flux rapidly changes between ca. 30 and ca. 1800 species/cm²/yr (Fig. 4).

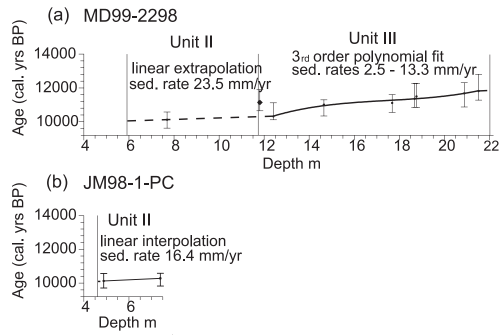

Fig. 3. Depth–age models for (a) core MD99-2298 and (b) core JM98-1-PC. The black diamond at 1180 cm (MD99-2298) is a reworked radiocarbon date. The lithostratigraphic units and their sedimentation rates are also indicated.

Discussion

Early Holocene palaeoenvironment in Malangenfjord

During the early Holocene relative sea level in Malangenfjord was less than 10 m above present (Corner & Haugane 1993; Hald & Vorren 1983), and therefore probably did not affect the hydrodynamics of the fjord, with its deep threshold. The relatively heavy oxygen isotope ratios suggest that the bottom-water in Malangenfjord was cold in the early Holocene. The $\delta^{18}O$ values have an average of 1.72 ‰ between 12 000 and 10 000 cal. yrs BP, which are lighter than the benthic $\delta^{18}O$ values (average 2.09 ‰) from a shelf record near Malangenfjord in the same time interval (Hald & Hagen 1998); this indicates that the bottom-water was either somewhat depleted by fresh meltwater or slightly warmer, or a combination of both. Despite the lower time resolution of the foraminiferal samples than the isotopic samples, they also point to cold bottom-water masses probably influenced by meltwater between 12 000 and 11 400 cal. yrs BP. The dominating species *C. reniforme* and *E. excavatum* f. *clavata* (Fig. 4) are common in Arctic and glacier-influenced fjords today (Hald & Korsun 1997). The $\delta^{18}O$ maximum between 11 500 - 11 400 cal. yrs BP, following the $\delta^{13}C$ depletion (Fig. 4), could be caused by either a cooling or reduced flux of $\delta^{18}O$-depleted meltwater. A decreased influence of meltwater could have resulted in an increased value of $\delta^{18}O$, as bottom-water became more saline.

The marked benthic $\delta^{18}O$ depletion (ca. 1 ‰) at ca. 11 400 cal. yrs BP indicates a temperature increase or increased meltwater flux. An increase in temperature is supported by the increased abundance of *C. lobatulus*, *T. angulosa*, *C. neoteretis* and *A. gallowayi* at 11 400 cal. yrs BP and their continued high abundance upwards (Fig. 4). These species reflect an increased influence of temperate Atlantic Water and stronger bottom currents (Mackensen et al. 1985; Hald & Steinsund 1996).

The two $\delta^{13}C$ minima at about 11 850 - 11 450 cal. yrs BP and 10 300 - 10 000 cal. yrs BP (Fig. 4) may be caused by, for example, increased stratification of the water column or an increase in biogenic production and organic decomposition with subsequent utilization of oxygen (Berger & Vincent 1986). The oldest and broad $\delta^{13}C$ minimum is most likely caused by increased stratification, as the oxygen isotopes indicate an influence of meltwater, and the abundance of both foraminifera and dinocysts indicate a minor biogenic production at this time (Husum 2002). Foraminiferal sample resolution is very low in this interval, and care must be taken into any interpretation of such data (Fig. 4). The youngest $\delta^{13}C$ minimum may partly be caused by a local production maximum, rather that meltwater stratification, as the dinocyst abundance shows the highest value for the entire Holocene in this interval (Husum 2002).

Early Preboreal cooling

The benthic $\delta^{18}O$ record derived from the present study is compared to other proxy records in Fig. 5. The $\delta^{18}O$ record from the GRIP ice core is largely a proxy record of the air temperature above the ice surface in Greenland (Dansgaard et al. 1993; Johnsen et al. 1995). The shelf record

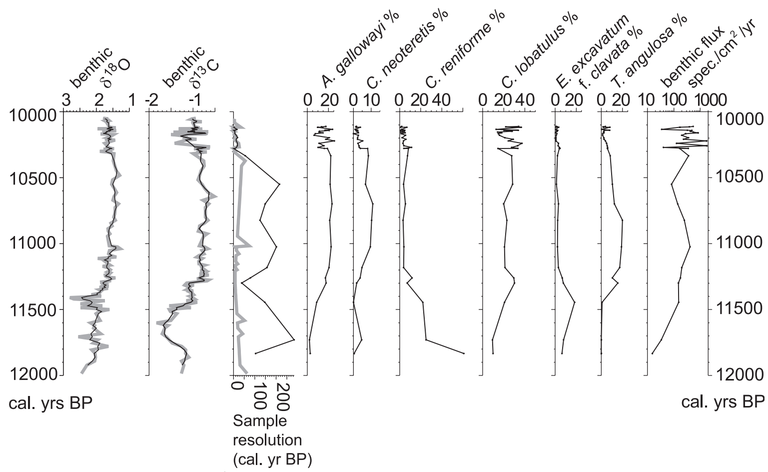

Fig. 4. The benthic stable isotope records, percentages of the most common benthic foraminiferal species and the benthic foraminiferal flux shown versus calibrated age. The isotope record has been corrected for global ice volume effect and disequilibrium effects. In addition, the time resolution of the isotopic (grey line) and foraminiferal samples (black line) is shown emphasizing the different resolution of the two proxies.

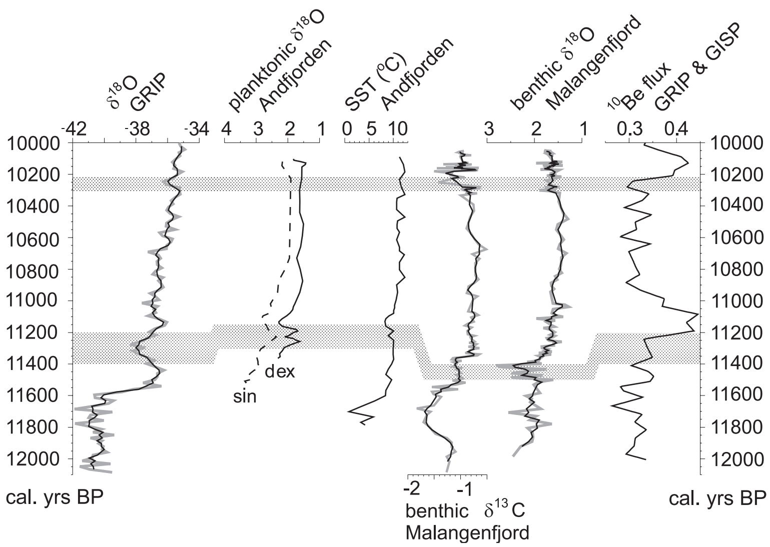

Fig. 5. The benthic $\delta^{18}O$ record from Malangenfjord compared to the $\delta^{18}O$ record from GRIP (Greenland Ice-core Project) (Dansgaard et al. 1993; Johnsen et al. 1995), the planktonic and benthic $\delta^{18}O$ record with SST from Andfjorden (Hald & Hagen 1998), and the ^{10}Be flux (10^6 atoms/(cm^2)*yr) from GRIP and GISP2 (Greenland Ice Sheet Project 2) (Bond et al. 2001 and references therein). The age scales for each proxy record are based on independent dates. When data are presented with both a grey and a black line, the black line is a three point running average of the data (grey line). The stippled grey areas show the early Holocene cooling events.

near Malangenfjord (Hald & Hagen 1998) represents water mass characteristics of Atlantic Water and meltwater on the shelf. The ^{10}Be flux from GRIP and GISP2 (Bond et al. 2001 and references therein) is regarded as a proxy record of the solar irradiance above Greenland. This flux is considered to be proportional with the global ^{10}Be flux (Muscheler et al. 2000).

We correlate the $\delta^{18}O$ maximum from 11 500 to 11 400 cal. yrs BP and subsequent warming in Malangenfjord with the so-called Preboreal Oscillation (PBO), meaning that the $\delta^{18}O$ maximum must be caused by a cooling. PBO is a widespread and robust signal in the Nordic seas region, both in marine records (Koç Karpuz & Jansen 1992; Lehman & Keigwin 1992; Bergsten 1994; Sarnthein et al. 1994; Conradsen & Heier-Nielsen 1995; Hald & Hagen 1998) and terrestrial records (e.g. Vorren & Moe 1986; Becker et al. 1991; Ingólfsson & Norddahl 1994; Andersen et al. 1995; Björck et al. 1996), and in Greenland ice cores (Johnsen et al. 1995). However, dating the PBO is problematic as it occurs in a time interval with radiocarbon plateaus (Björck et al. 1997). In the age model of the present study (Fig. 3), the age of the PBO cooling (11 500 - 11 400 cal. yrs BP) appears older by 100 - 200 years compared, for example, to the estimation of Björck et al. (1996)

of an age of 11 300 - 11 150 cal. yrs BP. However, the similar structure and widespread distribution of this signal suggest that it is synchronous throughout the Northern Hemisphere.

Hald & Hagen (1998) identified a meltwater event associated with the PBO on the shelf of northern Norway (Fig. 5). They argued that the PBO cooling was caused by a meltwater event hampering the North Atlantic heat conveyor. The data from Malangenfjord seem to support this theory by showing an increased stratification most likely due to increased meltwater flux in association with the PBO.

On the other hand, Björck et al. (2001) suggest that, due to its correlation with the onset of the first ^{10}Be rise in the early Holocene, the PBO cooling may have been caused by reduced solar forcing. However, because most PBO cooling in the ice core record occurred prior to the ^{10}Be flux rise (Fig. 5) we favour a meltwater event as the main forcing factor causing the early Preboreal cooling.

Minor cooling event at 10 300 - 10 100 cal. yrs BP

The marked δ^{13}C depletion in the Malangenfjord material at 10 300 - 10 100 cal. yrs BP coincides with a δ^{13}C depletion observed in the North Atlantic during a secondary early Holocene cooling event around 10 300 cal. yrs BP (Bond et al. 1997). In the Malangenfjord data this minimum is associated with an apparent local biogenic production maximum (Fig. 4). However, it may also be linked to a disturbance in the thermohaline circulation, as suggested by Bond et al. (1997). The δ^{18}O "flickering" around 10 300 - 10 100 most likely indicates small-scale cooling periods as they coincide with a cooling identified from lacustrine, tree ring, ice core and marine records in the North Atlantic region (Björck et al. 2001). There are no direct meltwater indicators for this cooling in the Malangenfjord data. However, this cooling also correlates with a rise in ^{10}Be flux (Fig. 5). In this case there is also a slight delay between the ^{10}Be increase and the cooling event observed in the ice core (Fig. 5), which cannot yet be explained (Björck et al. 2001). Despite this and the obvious reservation regarding the chronology we speculate that variations of solar irradiance do not appear to have a large impact in Malangenfjord, unless they are "amplified" by meltwater events.

Conclusions

Both the benthic foraminiferal faunas and the δ^{18}O record show a glaciomarine, meltwater δ^{18}O-depleted environment in Malangenfjord from ca. 12 000 to 11 400 cal. yrs BP. At 11 400 cal. yrs BP the fjord experienced stronger bottom currents and a rapid warming due to increased influx of Atlantic Water to the fjord.

The so-called Preboreal Oscillation / cooling is reflected in the fjord with reduced δ^{13}C values possibly caused by meltwater, followed by an abrupt bottom-water cooling between 11 500 - 11 400 cal. yrs BP. We suggest that the main forcing factor causing this cooling was increased meltwater flux hampering the North Atlantic heat conveyor.

A secondary, regional, early Holocene cooling event at 10 300 - 10 100 cal. yrs BP is only reflected by minor temperature variations. A δ^{13}C minimum at this time is associated with an apparent local biogenic production maximum.

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