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Last glacial benthic foraminiferal $\delta^{18}O$ anomalies in the polar North Atlantic: A modern analogue evaluation

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Abstract. Modern processes are evaluated to understand the possible mechanisms behind last glacial benthic foraminiferal $\delta^{18}O$ anomalies that occurred concurrent with meltwater events in the polar North Atlantic; such anomalies in the Nordic seas were recently interpreted to be caused by brine formation. Despite intensive sea-ice production on circumarctic shelves, modern data show that brines ejected from sea-ice formation containing low $\delta^{18}O$ water do not significantly contribute to deep waters in the Arctic Ocean today. Assuming that this process was, nevertheless, responsible for $\delta^{18}O$ anomalies in Nordic seas deep water during the last glaciation, a broad, shallow shelf area adjacent to the Nordic seas, such as the Barents Sea, had to be seasonally free of sea-ice in order to serve as an area for brine formation. Another process which may explain $\delta^{18}O$-depleted water at depth is found in the Weddell Sea today, where a low $\delta^{18}O$ signal in deep waters originates from ice shelf interactions. If temperature were considered the main mechanism for the low benthic $\delta^{18}O$ values, an increase of 4°C must have occurred in the deep water. An analogous situation with a reversed water temperature pattern due to a subsurface inflow of warm Atlantic water is found today in the eastern Arctic Ocean, and deep water warming is observed in the Greenland Gyre in the absence of deep convection. Because paleoproxy data also indicate an Atlantic water inflow into the Nordic seas during such benthic $\delta^{18}O$ anomalies, temperature as a principal mechanism of changing $\delta^{18}O$ cannot be excluded.

1. Introduction

Substantial oceanographic interest is focused on the Greenland, Norwegian, and Iceland Seas (Nordic seas) because deep and bottom waters form there today [e.g., *Aagaard and Carmack*, 1994]. It is believed that salinity changes at the surface influence the rate of deep water formation and, consequently, the global climate [*Rahmstorf*, 1995]. Many paleoceanographic studies in the Nordic seas have investigated the water mass circulation of the last glacial period because lowering of surface water salinities was likely then due to input of freshwater from melting glacier ice [e.g., *Sarnthein et al.*, 1995]. Because of the close vicinity of the area of deep water formation to heavily glaciated landmasses, paleoceanographic records from the Nordic seas are expected to deviate from the global average, and study of these records offers the chance to identify processes connected with variations in deep water formation.

It was formerly believed that the Nordic seas were ice-covered all year round during the last glaciation [e.g. *CLIMAP, Project Members*, 1976]. However, more recently faunal and isotope data imply that the Nordic seas were at least seasonally free of sea-ice [*Bauch*, 1994; *Weinelt et al.*, 1996], and some authors suggest that the modern thermohaline circulation was replaced by a mode driven by brine formation during certain periods of the last glaciation [*Dokken and Jansen*, 1999].

As shown by recent studies in the Nordic seas, glacial benthic foraminiferal $\delta^{18}O$ records covering the period 60 to 15 kyr show anomalously high-amplitude depletions in both benthic and planktic $\delta^{18}O$, particularly during the so-called "Heinrich events", which strongly deviate from the glacial global average [*Rasmussen et al.*, 1996a; *Dokken and Jansen*, 1999] (Figure 1). Heinrich events are deposits rich in ice-rafted detritus (IRD) and are related to intensive iceberg discharge and subsequent low $\delta^{18}O$ glacial meltwater release into the North Atlantic [e.g. *Bond et al.*, 1992]. It is suggested that brine formation was the mechanism responsible for transporting surface $\delta^{18}O$ depletions into the deep waters of the Nordic seas during such events [*Vidal et al.*, 1998; *Dokken and Jansen*, 1999]. Based on the time-coeval occurrence of the planktic and benthic $\delta^{18}O$ depletions it is further suggested that the mode of thermohaline water mass convection shifted during these times to a mode driven by brine formation [*Dokken and Jansen*, 1999]. While we consider this explanation of deep water formation during glacial times to be innovative, we also believe that it is an oversimplification that does not take into account other mechanisms producing low benthic $\delta^{18}O$.

In order to interpret glacial benthic $\delta^{18}O$ anomalies and to identify the processes responsible, comparison with modern analogues is essential. In this paper we discuss the process of brines ejected by sea-ice formation as well as other mechanisms capable of producing low benthic $\delta^{18}O$ in more detail.

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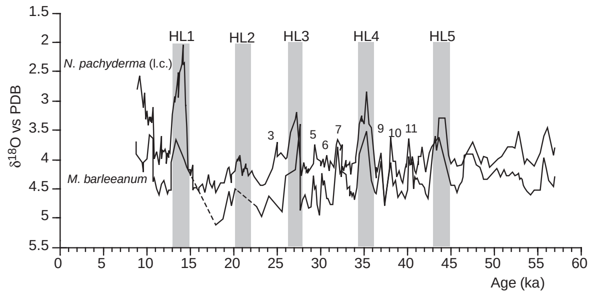

Figure 1. Planktic and benthic δ^{18}O records versus age from core ENAM 93-21 (62°N, 4°W, 1020 m) [*Rasmussen et al.,* 1996a]. The shaded areas indicate the Heinrich events, and the numbers refer to the stadials defined in the Greenland ice record. Figure 1 is adapted from *Vidal et al.* [1998].

2. Modern Analogues of Mechanisms Producing Low Benthic δ^{18}O

The δ^{18}O signal of foraminiferal calcite is influenced by the temperature (T) and the δ^{18}O composition of the water (d_w) in which it is formed. For inorganic calcite (d_c) precipitated under equilibrium conditions, this can be described by $T = 16.9 - 4.38(d_c - d_w) + 0.1(d_c - d_w)^2$ [*O'Neil et al.,* 1969]. This equation can be applied to foraminiferal calcite in conjunction with species-dependent offsets, so called "vital effects" [e.g., *Bemis et al.,* 1998]. Therefore low benthic δ^{18}O signals can be caused by low δ^{18}O bottom water as well as by increased bottom water temperatures.

During the formation of sea-ice, salt is released via channels from the ice as so-called "brines." The observed iso-tope fractionation effects during sea-ice formation are small (α = 1.002 to 1.0025 [*Melling and Moore,* 1995; *Macdonald et al.,* 1995]) and will not lower the δ^{18}O content of the remaining water significantly. Therefore only sea-ice forma-tion in conjunction with admixture of isotopically low water can generate low δ^{18}O water dense enough to descend in the water column. This isotopic ratio could then be taken up as a low δ^{18}O signal by benthic foraminifera.

We will evaluate the following processes and coherent modern situations as possible mechanisms causing low benthic δ^{18}O in the glacial Nordic seas:

1. Brine release in conjunction with low δ^{18}O surface water on a shallow shelf area is a possible mechanism. A well-studied and representative modern analogue is sea-ice forma-tion on the shelf areas of the Arctic Ocean.

2. Brine release in conjunction with low δ^{18}O surface water in an open-ocean setting is a process to consider. A suitable modern analogue related to deep convection is the Weddell Polynya in 1973-1976.

3. Injection of low δ^{18}O water at depth below an ice shelf is a process to consider. A suitable modern analogue is the formation of deep and bottom waters in the Weddell Sea below the Filchner/Ronne Ice Shelf involving δ^{18}O depleted meltwater from the ice shelf.

4. Warming of deep water is a possible mechanism. The isolation of warm subsurface Atlantic waters below cold Arctic Ocean halocline waters and the slight warming of Greenland Sea Deep Water between 1993 and 1996 due to the cessation of deep convection is suggested as being a possible modern analogue.

Slope convection driven by a density excess from fine-grained suspended sediments has been investigated through modeling studies and experiments [e.g. *Garcia and Parker,* 1993; *Kämpf et al.,* 1999]. These dense plumes enriched in sediments might be another possible process to transport low δ^{18}O water to depth, but very little is known about this process and its capacity to transport water to depth; therefore this process is not discussed further.

2.1 Sea-Ice Formation on Shallow Shelves

The Arctic Ocean interior is permanently covered by sea-ice and only part of the shelf areas are seasonally free of ice. These shelf regions, covering ~ 1/3 of the Arctic Ocean area but representing only ~ 2% of its volume, receive ~ 0.1 Sv river runoff [*Aagaard and Carmack,* 1989] having an average δ^{18}O composition of about -21‰ [*Östlund and Hut,* 1984]. Because sea-ice only grows slowly after initial formation, the major part of sea-ice is produced on the shelf areas, where enhanced rates of sea-ice production is facilitated by flaw leads and polynyas. Polynyas are cooled and kept open by cold and steady winds, thus allowing for quasipermanent new ice production at high rates despite their limited areal extent [*Dethleff et al.,* 1998].

The dense water expelled by sea-ice formation mixes with the low δ^{18}O and relatively low salinity water on the shelves. These shelf waters, with a wide range of salinities, feed into the Arctic Ocean halocline [*Aagaard et al.,* 1981] (Figure 2).

Deep profiles of δ^{18}O in the Arctic Ocean show no apparent decrease in δ^{18}O composition towards the sea floor [*Bauch et al.,* 1995] (Figure 3). A comparison of salinity and δ^{18}O composition of the bottom waters of the Arctic Ocean, the Nordic seas, and the Atlantic Inflow as the main source water mass (Figure 4) reveals a slight influence of δ^{18}O-depleted water only in the Makarov Basin [*Bauch et al.,* 1995]. The salinity shift of ~ 0.15 reveals an overall freshwater content of ~ 0.5% and the δ^{18}O shift of ~ 0.03‰ indicates that ~ 0.2% of the bottom water in the Makarov Basin is derived from low δ^{18}O river water [*Bauch et al.,* 1995]. Even though ~ 4500 km^3 of sea-ice are formed on the circum-arctic shelves annually [*Dethleff et al.,* 1998], very little δ^{18}O-depleted

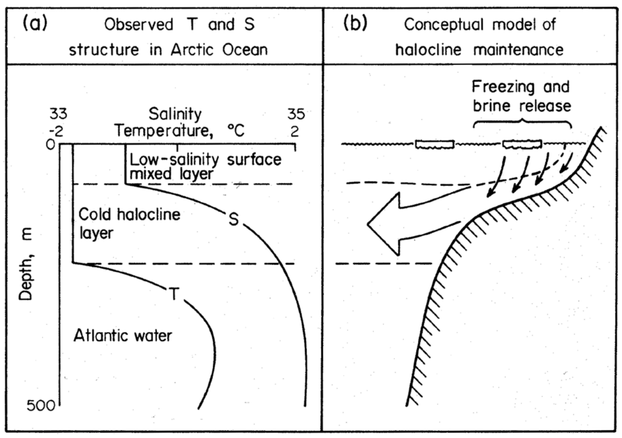

Figure 2. Schematic drawing to illustrate the maintenance of the Arctic Ocean halocline from the shelves. Figure 2 is adapted from *Aagaard et al.* [1981]. Reprinted with permission from Elsevier Sciences. Copyright © 1981.

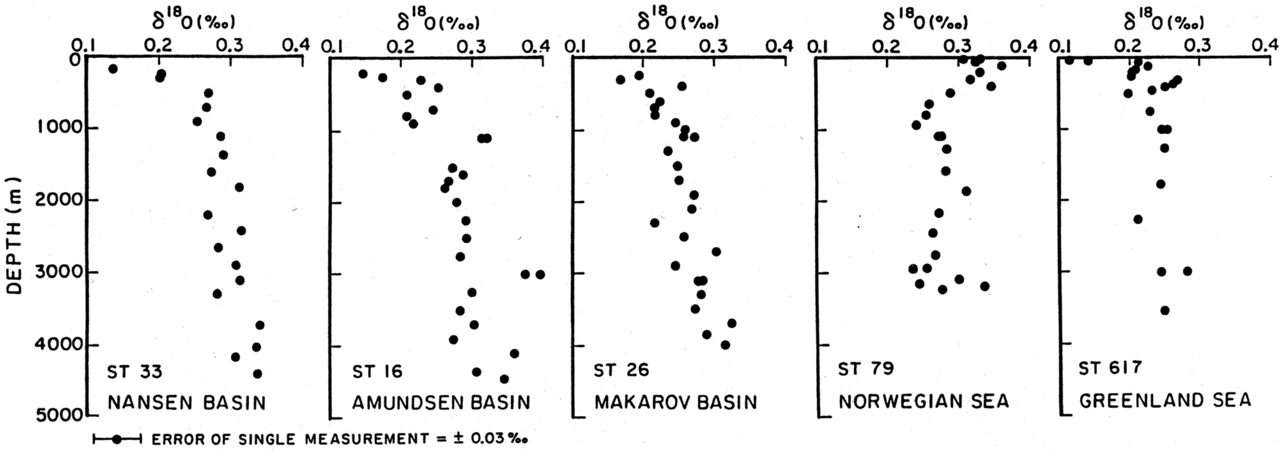

Figure 3. Measurements of $\delta^{18}O$ in the water column versus depth for deep stations in different basins of the Arctic Ocean and the Norwegian and Greenland Seas: station 33 at 86.8°N, 10.3°E, 4382 m depth; station 16 at 87.6°N, 69.7°E, 4451 m depth; station 26 at 88.0°N, 163.7°E, 4451 m depth; station 79 at 70°N, 0°E, 3280 m depth; station 617 at 74.8°N, 5.5°W, 3528 m depth. Figure 3 is taken from *Bauch et al.* [1995]. Reprinted with permission from Elsevier Sciences. Copyright © 1995.

freshwater is transported via brines into the bottom waters of the Arctic Ocean. The major part of the low $\delta^{18}O$ river water mixing with brines is not dense enough to descend into greater depth and remains in the Arctic halocline.

2.2. Sea-Ice Formation and Open Ocean Convection

In the Southern Ocean most of the pack ice is newly formed every year. Averaged density profiles from regions around Antarctica (Figure 5, adapted from *Martinson et al.* [1981]) show a deep mixed layer, underlain by a thin pycnocline (dominated by salinity) and nearly homogenous deep water down to the seafloor (~ 4.5 km deep). The Weddell region shows the thinnest (~ 20 m) and shallowest pycnocline, which makes it most susceptible to overturn as a result of vertical heat exchange [*Martinson,* 1990]. Each year a seasonal sea-ice cover develops, and salt ejection associated with sea-ice growth further weakens the stability of the water column by eroding the pycnocline. Extensive ice growth can destabilize the water column, inducing deep convection. The heat flux associated with such convection is sufficient to completely eliminate the sea-ice cover [*Martinson,* 1990; *Martinson and Iannuzzi*, 1998]. Presumably, the presence of the Weddell Polynya in 1973-1976 [*Zwally and Gloersen,* 1977] was the surface manifestation of localized deep convection [*Gordon, 1982; Martinson et al.*, 1981]. Open ocean convection triggered by brine release leads to strong

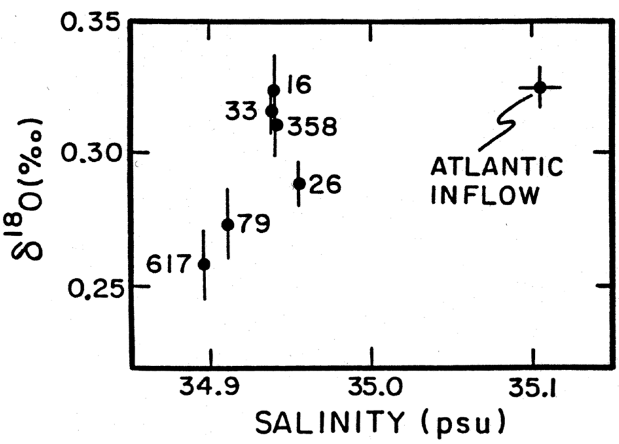

Figure 4. The $\delta^{18}O$ versus salinity for bottom waters (2600 m to bottom depth) and Atlantic Inflow (main source water mass). Station 358 at 84.04°N, 30.63°E, 2752 m depth; other stations as in Fig. 3. Figure 4 is taken from *Bauch et al.* [1995]. Reprinted with permission from Elsevier Sciences. Copyright © 1995.

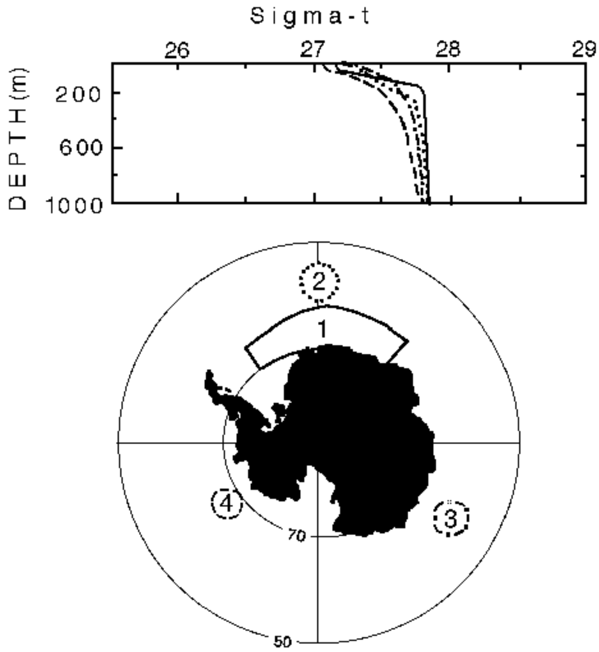

Figure 5. Averaged density profiles from the indicated regions around Antarctica. Figure 5 is adapted from *Martinson et al.* [1981]. Reprinted with permission from American Meteorological Society. Copyright © 1981.

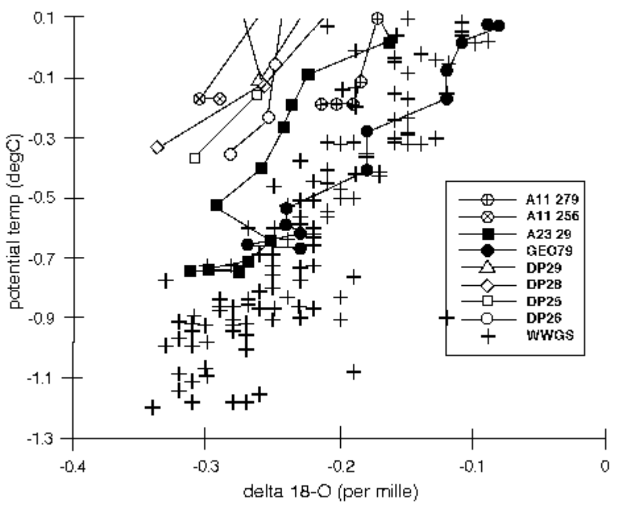

Figure 6. Potential temperature versus $\delta^{18}O$ plot for the Weddell Sea Deep Water (potential temperature above -0.7°C) sampled in the Drake Passage (DP25, DP26, DP28, and DP29) and the outflowing part of Weddell Sea Deep Water and Weddell Sea Bottom Water (potential temperature above -0.7°C) sampled within the Weddell Sea during the Winter Weddell Gyre Study (WWGS). Additional data are Geochemical Ocean Section Study station 79 (GEO79) and World Ocean Experiment (WOCE) A23 section station 29 (A23 29) just north of the Weddell Sea and WOCE A11 section stations 256 and 279 (A11 256 and A11 279) located in the South Atlantic at 45°S. Figure 6 is taken from *Meredith et al.* [1999].

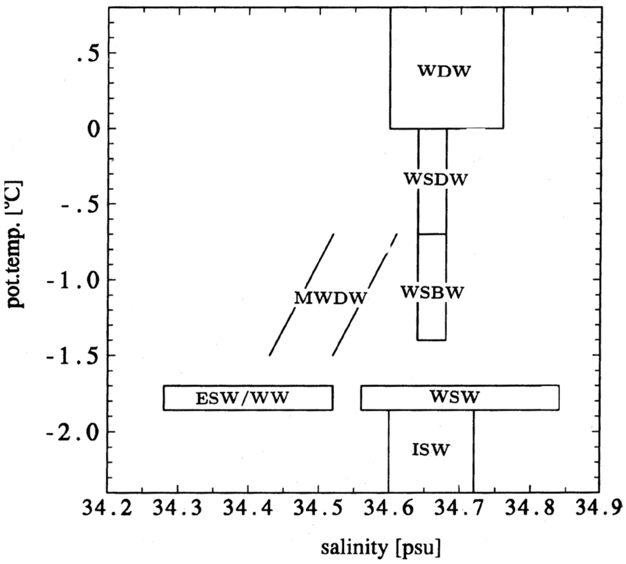

Figure 7. Schematic diagram outlining the mixing schemes that can lead to the formation of Weddell Sea Bottom Water (WSBW) and Weddell Sea Deep Water (WSDW). Other abbreviations are as follows: WDW, Weddell Deep Water or Warm Deep Water; MWDW, Modified Warm Deep Water; WW, Winter Water; ESW, Eastern Shelf Water; WSW, Western Shelf Water; ISW, Ice Shelf Water. Figure 7 is adapted from *Weppernig et al.* [1996].

dilution of potentially $\delta^{18}O$ depleted surface waters; no detectable low $\delta^{18}O$ signal is transported into the deep waters by this process in the Weddell Sea. The observed depletion in $\delta^{18}O$ with depth (see Figure 6 *Meredith et al.,* [1999]) is caused by processes other than open ocean convection, as discussed in section 2.3.

2.3. Melting of Low $\delta^{18}O$ Water Below an Ice Sheet

The Antarctic Ice Sheet, with a $\delta^{18}O$ composition of about -40‰ [*Morgan*, 1982], moves slowly out from the interior towards the coastline where it ends as a vertical ice wall called the "Barrier." This ice shelf floats in the water over large areas. In the Weddell Sea the floating ice shelf covers an area comparable to that of the Greenland Basin, and the depth of the ice submerged in the water at the Barrier may exceed 400 m [*Foldvik and Gammelsrød,* 1988].

Warm Deep Water (WDW) is the main water mass found in the Weddell Gyre with a $\delta^{18}O$ composition of about -0.1‰. Eastern Shelf Water and Western Shelf Water are formed near the Barrier of the Ice Shelf east and west of the Filchner Depression, respectively, with a $\delta^{18}O$ of about -0.45‰, caused by admixture of low $\delta^{18}O$ precipitation [*Weppernig et al.,* 1996].

Two processes are responsible for deep water formation in the Weddell Sea. Modified by shelf waters, WDW can mix with Winter Water (WW) (a remnant of the winter mixed layer), and Western Shelf Water (WSW) to form Weddell Sea Bottom Water (WSBW) [*Foster and Carmack*, 1976] (Figure 7). An alternative process of deep water formation involves the formation of supercooled Ice Shelf Water (ISW) from WSW beneath the Filchner Ice Shelf (Figure 8). Measurements of $\delta^{18}O$ in ISW (potential temperature below -1.8°C) show a clear correlation with the potential temperature (Figure 9). ISW has $\delta^{18}O$ values as low as about -0.7 to -0.8‰, caused by the admixture of glacial meltwater and is observed locally in plumes at depth [*Weppernig et al.,* 1996; *Schlosser et al.,* 1990]. Subsequent entrainment of WDW leads to the formation of Weddell Sea Deep Water (WSDW) and WSBW (Figure 7). The proportions of ISW and WSW in WSDW and

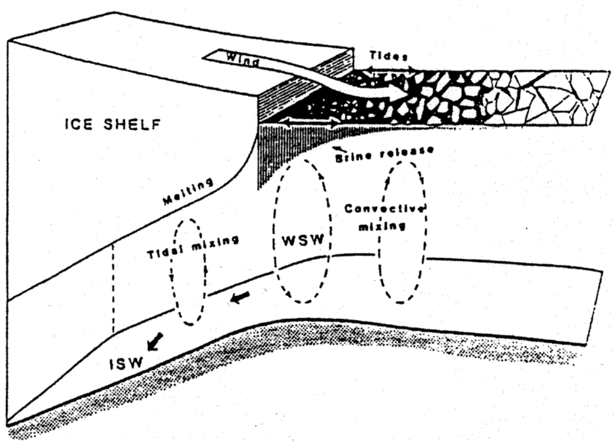

Figure 8. Sketch indicating the formation of Ice Shelf Water (ISW). Western Shelf Water (WSW) is cooled and densified due to brine release near the barrier. ISW is formed from WSW by further cooling due to the melting of low $\delta^{18}O$ glacial water under the ice shelf. The organized flow of ISW due to topography is sketched. Figure 8 is adapted from *Foldvik and Gammelsrød* [1988]. Reprinted with permission from Elsevier Sciences. Copyright © 1988.

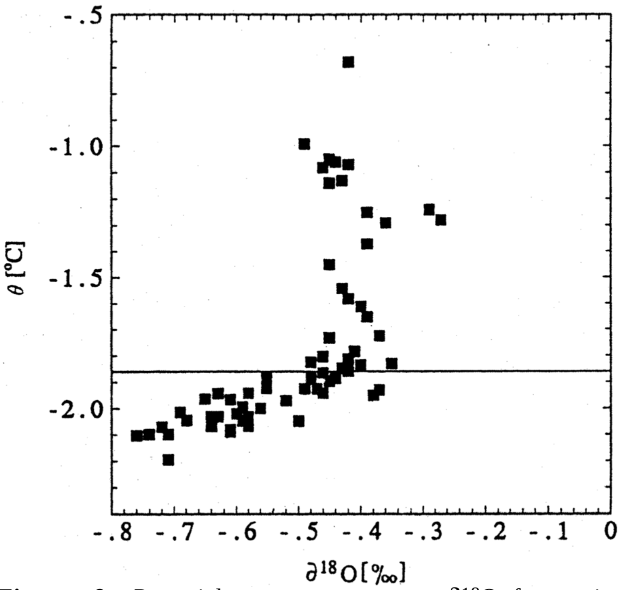

Figure 9. Potential temperature versus $\delta^{18}O$ for stations located in the Filchner Depression close to the ice shelf front [*Schlosser et al.*, 1990]. The horizontal line indicates the freezing point of seawater at surface pressure. Figure 9 is adapted from *Weppernig et al.* [1996].

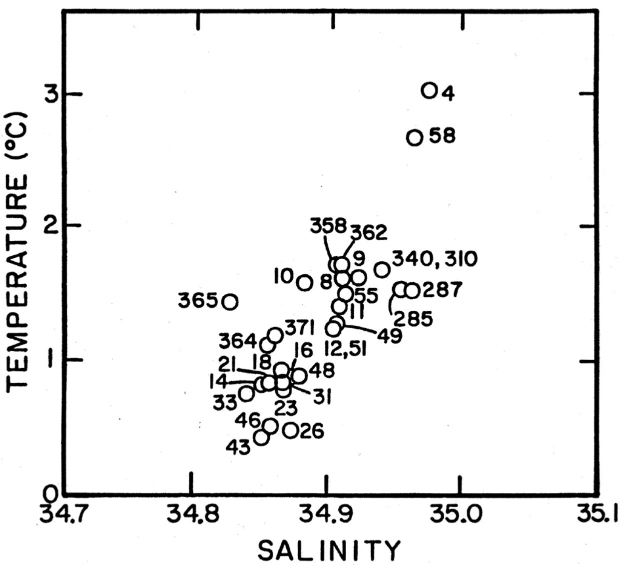

Figure 11. Potential temperature versus salinity plot for the Atlantic core. Station numbers are from ARKIV/3 and Project ARCTIC'91. Figure 11 is taken from *Bauch et al.* [1995]. Reprinted with permission from Elsevier Sciences. Copyright © 1995.

WSBW are roughly ~ 10% and 30%, respectively, and are estimated from a balance of $\delta^{18}O$ and ^4He concentrations [*Weppernig et al.,* 1996]. The concurrent $\delta^{18}O$ depletion in WSDW and WSBW is ~ -0.2 to -0.3‰ relative to the main source water mass WDW (Figure 6) [*Meredith et al.,* 1999; *Weppernig et al.,* 1996].

2.4. Deep Water Warming

With the absence of deep reaching winter convective events the temperature of the deeper waters of the Greenland Gyre increased by ~ 0.03 K between 1993 and 1996 [*Budeus et al.*, 1998]. This temperature increase is explained by a large-scale downward movement of water, in agreement with chemical tracer observations. While deep winter convection freshens and cools the bottom waters, downward movement of warmer, saltier water steadily raises the temperature of the bottom waters (Figure 10). Winter convection acts mainly as a mixing agent throughout the convective layer and rapidly redistributes water masses. The large-scale convection mechanism induces a replacement of waters in the Greenland Sea which, if extrapolated, would flush the Greenland Sea within 20-30 years [*Budeus et al.*, 1998]. It is not clear if the proposed downward movement of water represents a permanently active process masked by superimposed winter convection or if it alternates with phases of deep convection [*Budeus et al.,* 1998].

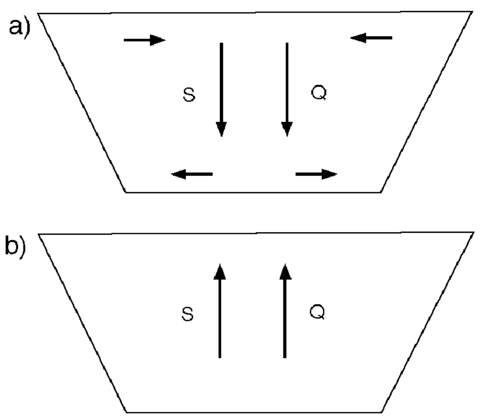

Figure 10. (a) Concept of deep water replacement due to large scale decadal convection. S and Q denote salinity and heat fluxes, respectively, in the deep water resulting from today's salinity and temperature distributions. (b) Effect of deep reaching winter convective events in contrast to scenario Figure 10a. Figure 10 is adapted from *Budeus et al.* [1998].

It is important to recall that warm bottom waters do not imply warm surface waters on a local scale. A strong surface stratification and sea-ice cover in the Arctic Ocean today leads to isolation of relatively warm Atlantic-derived waters below cold halocline waters (-1.8°C). Atlantic Water flowing north in the West Spitzbergen Current loses heat rapidly (e.g., 8°C west of Spitzbergen compared with up to 4°C north of Spitz-bergen [*Bauch et al.,* 1995; *Pfirman et al.,* 1994]). On submerging below the Arctic Halocline, this rapid cooling of the warm Atlantic Water stops and the temperature of the Atlantic temperature maximum stays between 1° and 2°C within the Eurasian Basin of the Arctic Ocean [*Coachman and Barnes,* 1963] (see also Figure 11).

3. Paleoceanographic Implications

In an open ocean setting, brine release associated with sea-ice formation can lead to deep convection as observed in the Weddell Polynya today. However, open ocean convection or convection on a deep shelf area is not effective as a mechanism to produce deep water with a low $\delta^{18}O$ signal, because the dilution is too high. When sea-ice formation occurs on a shallow shelf, the dilution of the surface $\delta^{18}O$

signal is considerably smaller because of the limited mixing depth. Dense but low $\delta^{18}O$ water can then descend from the shelf as boundary currents along the slope without considerable further mixing.

During the last glaciation, when sea level dropped significantly, ice sheets developed on Scandinavia and on the Barents Sea shelf [e.g., *Svendsen et al.*, 1999]. The presence of ice sheets must have induced persistent katabatic winds which facilitated the occurrence of polynyas in which large amounts of sea-ice could be produced. The shallow parts of the Greenland and Norwegian shelves are relatively narrow and might have been covered by ice sheets, or, otherwise, they were probably not seasonally free of sea-ice. Therefore the Barents Sea, with an average water depth of 160 m today, is the only broad shelf area adjacent to the Nordic seas. Assuming that this area was partly free of ice sheets and seasonally free of sea-ice, enhanced sea-ice formation might have occurred there as it does on the broad and shallow Siberian shelf areas in the Arctic Ocean today.

If we speculate that glacial deep water in the Nordic seas was influenced by brines from the Barents Sea shelf, we have to assume that relative to the Arctic shelves today a smaller amount of low $\delta^{18}O$ freshwater was present (in order to produce higher salinities and densities) and/or a larger amount of brine formed. Shelf waters dense enough to contribute significantly to the bottom waters might have formed under such circumstances. This hypothesis has been adopted by some authors for the glacial Nordic seas during meltwater pulses such as the Heinrich events [e.g., *Vidal et al.*, 1998; *Dokken and Jansen*, 1999].

The following scenario is a first-order approximation to demonstrate the sea-ice formation rates and the rates of exchange between shelf waters and deep waters required to produce a low benthic $\delta^{18}O$ spike of about -1‰. Past conditions of the Barents Sea (BS) and the Norwegian and Greenland Seas Deep Waters (NGSDW) are estimated by adding 1 unit to present day salinities (average salinity of BS shelf waters 34.7 [*Bauch*, 1995] and average salinity of NGSDW 34.9) and 1.1‰ to the present $\delta^{18}O$ (BS and NGSDW both ~ 0.3‰; [*Bauch et al.*, 1995]), which correspond approximately to a sea level drop of 120 m [*Fairbanks*, 1989]. At low temperatures, density is dominated by salinity, and, for simplification, salinity rather than density is discussed here: If about 15% of glacial meltwater was added to the Barents Sea during the Heinrich events, average conditions would have changed from a salinity of 35.7 to ~ 30 and from a $\delta^{18}O$ of ~ 1‰ to -5‰, assuming a $\delta^{18}O$ of ~ -40‰ for glacial meltwater (see also Figure 12). Shelf water has to reach a salinity of ~ 36 in order to resemble past NGSDW (with a salinity of ~ 34.9+1 as estimated above) and to be able to sink to depth. Depending on the amount of sea-ice assumed to be produced and contributing to deep water formation each year, a certain amount of Barents Sea water could contribute to the deep waters formed in the glacial Nordic seas. Assuming that an amount of sea-ice equivalent to that formed in the entire Laptev Sea flaw lead today (258 km^3/yr, i.e., ~ 10% of the Siberian branch of the Transpolar Drift and therefore one of its main contributors [*Dethleff et al.*, 1998]) contributed to deep water formation, ~ 1300 km^3/yr of Barents Sea shelf water could achieve a salinity of 36 and descend to the deep waters of the Nordic seas. This supply of 1300 km^3/yr (~ 0.4 Sv) of Barents Sea shelf water with an $\delta^{18}O$ of -5‰ would lower the $\delta^{18}O$ of Norwegian and Greenland Seas Deep Waters (today a volume of about 1.5 x 10^3 km^3 [*Heinze et al.*, 1990]) by -1‰ within a little more than 2 centuries if it is assumed that no further exchange occurred.

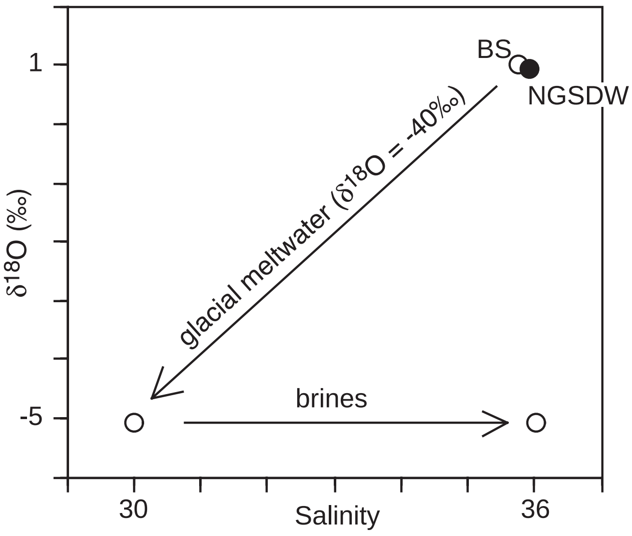

Figure 12. Schematic $\delta^{18}O$ versus salinity plot for the transformation of Barents Sea water (BS, open circles) by glacial meltwater during meltwater events and the addition of brines in order to reach a salinity high enough to contribute to the Norwegian and Greenland Sea Deep Water (NGSDW; solid circle). As a simplification salinity changes instead of density changes are shown. For further explanation, see text.

On the basis of these assumptions it is indeed possible to produce a benthic $\delta^{18}O$ signal of -1‰ via transport of low $\delta^{18}O$ surface water by brines within an appropriate time interval. However, this scenario would only work if the Barents Sea was seasonally free of sea-ice and served as a production area of what we believe to be unrealistically high amounts of sea-ice contributing brines to the deep waters without further dilution. Additionally the assumptions used in the scenario, a sea level drop of 120 m (reached during the last glacial maximum only) and a relatively low $\delta^{18}O$ value for glacial meltwater of -40‰ were chosen to favor the possibility of brine formation transporting low $\delta^{18}O$ to depth.

Low benthic $\delta^{18}O$ signals are also observed during times when ice sheets were largest (e.g. at about 21 kyr and 14 kyr) and could have been derived from the interaction with ice shelves in analogue to the processes causing WSDW and WSBW to be about -0.2 to -0.3‰ more depleted than the main source water mass (WDW). We consider this process to be more likely than brine formation on the Barents Sea shelf because simple budget considerations reveal that a small increase of pure glacial meltwater would be enough to decrease the $\delta^{18}O$ signal of the bottom waters sufficiently. Taking the present situation in the Weddell Sea, an additional amount of ~ 2% of pure glacial meltwater ($\delta^{18}O$ of -40‰) in the bottom water would lower the $\delta^{18}O$ at depth by about 1‰. This means that slightly less than 3 times the amount of pure glacial melt-water found in WSBW today had to be involved. The relative increase is considerable but still seems easily attainable, for example, by assuming a water mass relatively warmer than WSW (-1.8°C) to exchange with an ice shelf.

However, it is also difficult to explain how the formation of ice shelf water, as known from the Weddell Sea today, could have produced more than a basin-wide $\delta^{18}O$ decrease in the deep waters, for example during early Termination I (i.e. the end of the last glaciation) in the Nordic seas as well as in the North Atlantic [*Veum et al.,* 1992; *Rasmussen et al.,* 1996a; *Costello and Bauch*, 1997; *Bauch et al*., 2001]. It is likely that an alternative process replaced or supplemented these more locally restricted processes during certain glacial intervals.

A process that could explain benthic $\delta^{18}O$ changes on a broader scale is a general temperature increase of the bottom waters. Records of benthic $\delta^{18}O$ across the last glacial-interglacial transition suggest that glacial NE Atlantic bottom water temperatures were ~ 4°C cooler than today [*Labeyrie et al.*, 1987; *Bauch et al.,* 2000, *Schrag et al.,* 1997]. Because a similar phenomenon is observed worldwide for glacial-interglacial transitions [e.g. *Shackleton et al.,* 1983] it is generally accepted that bottom water temperatures changed even though the actual range might appear unreasonably high as a face value. The 1‰ $\delta^{18}O$ decrease observed during Heinrich events 2 to 5 (HL2 to HL5; see Fig. 1) would be equivalent to a temperature increase of about 4°C. Higher benthic $\delta^{18}O$ spikes are observed during HL1 and HL6 (see Figure 1 and *Dokken and Jansen* [1999]), which both follow icevolume maxima (i.e. oxygen stages 4 and 2). It is most probable that such large anomalies were the combined result of several factors caused by the collapse of entire ice sheets.

Changes in benthic foraminiferal assemblages in association with the low benthic $\delta^{18}O$ spikes in core ENAM 93-21 (see Figure 1) indicate changes of water masses and have been interpreted as the intrusion of relatively warm water from the North Atlantic [*Rasmussen et al.*, 1996a; 1996b]. This possibility has been excluded by others citing benthic foraminiferal $\delta^{13}C$ data of infaunal species [*Vidal et al.*, 1998; *Dokken and Jansen*, 1999]. These authors note that the $\delta^{13}C$ values in the Nordic seas are relatively low during the low $\delta^{18}O$ spikes, which seem to disagree with much higher epifaunal $\delta^{13}C$ values in the North Atlantic at intermediate depths at the same time [*Oppo and Lehman*, 1993]. However, $\delta^{13}C$ data obtained from infaunal benthic species may not be representative of the $\delta^{13}C$ of the Dissolved Inorganic Carbon (DIC) in the bottom waters. Differences in pore water and seawater $\delta^{13}C$ of DIC have been reported (*Zahn et al*., 1986]. Also, while the offset in oxygen isotopes between benthic infaunal and epifaunal species is rather constant during interglaciations, this consistency does not hold for glacial intervals in the Nordic seas [*Bauch et al.*, 2000; *Bauch et al.*, 2001]. Moreover, data from the epifaunal species *C. wuellerstorfi* from a core close to ENAM 93-21 show consistently high $\delta^{13}C$ values during HL2 [*Veum et al.*, 1992], underlining the uncertainties occurring when comparing infaunal and epifaunal $\delta^{13}C$ records. We therefore believe that the reported low benthic $\delta^{13}C$ data from infaunal species does not exclude the possibility of a high $\delta^{13}C$ water mass. Also the faunal evidence for the intrusion of warm water from the North Atlantic should not be dismissed [*Rasmussen et al.,* 1996a; 1996b; *Bauch et al.,* 2000].

From planktic foraminiferal assemblage studies there is now ample evidence for an almost continuous inflow of Atlantic water into the Nordic seas during late oxygen isotope stages 3 and 2, i.e. the maximum of the last glacial period [*Bauch*, 1994; *Bauch et al.,* 2001], reaching as far north as the Fram Strait [*Hebbeln et al.,* 1994; *Dokken and Hald*, 1994] and even the Arctic Ocean (N. Nørgaard-Pedersen, oral communication, 1999]. While Atlantic water entering the Nordic seas during the last glaciation might have participated in a form of modern-type vertical convection, the circulation style must have changed dramatically during Heinrich events, when low benthic and planktic $\delta^{18}O$ spikes are observed synchronously [*Dokken and Jansen*, 1999]. Meltwater released in the Nordic seas during the Heinrich events might have facilitated a situa-tion in which Atlantic water was submerged below an enhanced cold and stratified surface layer as in the Arctic Ocean today [*Bauch et al.,* 2001]. Under such circumstances, deep winter convection, as well as extensive heat loss to the atmosphere from the underlying warm water mass, is interrupted. In the absence of deep vertical convection the sinking of the entire water column can cause gradual warming of the bottom water, as the observed 0.03 K in the Greenland Gyre between 1993 and 1996 [*Budeus et al.,* 1998]. This is one possible mechanism of deep water warming. Although the time series of observations is too short to call this recent case from the Greenland Sea fully representative of past conditions, a similar rate of heat input to the deep waters of the Nordic seas during the last glaciation could explain the low benthic $\delta^{18}O$ spikes as found in core ENAM 93-21 (HL2-HL5; see Figure 1).

If we assume that the density of the waters remains constant during such a temperature change (~ 4°C), a salinity increase of ~ 0.3 would be required. On the basis of global ice volume changes, only rough estimates of salinity changes during the last glacial period can be derived, and it seems probable that these average salinity differences were not evenly distributed. Therefore, even though a salinity increase of 0.3 seems hard to envisage under present conditions, this might be different assuming a probably stronger salinity gradient between the Nordic seas and the North Atlantic during glacial times. However, in order to clarify if temperature were responsible for low benthic $\delta^{18}O$ values in the Nordic seas during the last glacial period, a sophisticated modelling approach beyond the scope of this study is necessary. A modeling study shows that certain circumstance involving melwater in the Southern Ocean and enhanced formation of deep water in the North Atlantic can lead to a general increase in deep water tempera-tures [*Seidov et al.*, in press]. As it stands, deep water warming is one mechanism not to be neglected as a contributor to low benthic $\delta^{18}O$ signals during major meltwater events of the last glaciation.

4. Summary

Low benthic $\delta^{18}O$ signals associated with Heinrich events in the Nordic seas have been interpreted previously as caused by brine production related to sea-ice formation. Discussion of modern analogues implies that this interpretation may be difficult to accept as the sole cause. Open ocean convection would introduce too much dilution to produce a low $\delta^{18}O$ signal, so brine formation processes must have occurred on shallow shelf areas such as the Barents Sea. This shelf must have been only partially covered by an ice sheet and was seasonally free of sea-ice during the last glaciation. An alternative mechanism for generating low benthic $\delta^{18}O$ is the intrusion of glacial meltwater with low $\delta^{18}O$ beneath a floating ice shelf as found in the Weddell Sea today, where Ice Shelf Water below the Filchner/Ronne Ice Shelf is involved in the formation of Weddell Sea Deep and Bottom Waters leading to a $\delta^{18}O$ decrease of ~ 0.2 to 0.3‰. For the Nordic seas such a scenario would imply a floating ice sheet grounded in the Barents Sea and the other shelf areas. On the basis of simple budget considerations this ice sheet interaction seems a more likely process of producing low benthic $\delta^{18}O$ signals than brine formation. A mechanism to produce a low $\delta^{18}O$ signal on a broader scale could be a temperature change of the bottom waters. A stratified surface layer might have caused the cessation of deep winter convection and isolated the inflowing warm waters. This may have led to gradual warming by lateral advection of warm waters as observed in the Arctic Ocean today, and an overall downward movement, as observed in the Greenland Sea during 1993-1996 in the absence of deep convection.

Evaluation of possible bottom water temperature changes in the Nordic seas requires further investigations. This seems especially important because the alternate mechanisms involving glacial meltwater below an ice shelf and brine release by sea-ice formation are more likely to be processes of locally restricted impact.

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