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Stable isotope clue to episodic sea ice formation in the glacial North Atlantic

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Abstract

Sea ice is a determinant parameter of the climate system, which acts as amplifier through positive ice-albedo feedbacks and induces deep-water formation through brine production. It is also one of the most elusive parameters in the paleoclimate record. Most proxies of sea ice provide information on its presence and extent of (e.g., diatoms or dinocysts), but do not permit assessment on its rate of formation. However, seemingly off-equilibrium, low δ^{18} O values in mesopelagic planktic foraminifer species, as observed today in the Arctic Ocean, are thought to relate to the production of isotopically light brines during sea ice formation, and might thus provide a clue on the rate of sea ice growth. With reference to the isotopic properties of modern planktic foraminifers in the Arctic Ocean and using multi-proxy data sets from the last glacial stage in the northwest North Atlantic, we re-examined some of the current interpretations of planktic isotope records in the glacial NW North Atlantic. The large amplitude light isotopic excursions recorded in Neogloboquadrina pachyderma left coiling during Heinrich events 1, for example, correspond to extensive sea ice cover as reconstructed from dinocysts, and do not seem unequivocally linked to low-salinity pulses. Such light isotopic excursion more likely responded to enhanced rates of sea ice formation resulting in the production and sinking of isotopically light brines. On the contrary, the isotopically heavy planktic foraminifers of the last glacial maximum (LGM) interval stricto sensu would rather suggest relatively low rates of brine production, thus low sea ice growth rates in the area. This would imply that the LGM distribution of sea ice in the North Atlantic was primarily linked to spreading and drifting from marginal and Nordic source areas and not from enhanced in situ production. At the scale of glacial-interglacial cycles, isotopic distillation processes relating to sea ice production at high latitudes could be one of the factors controlling the ¹⁸O-salinity relationship in deep ocean water masses that might deserve closer examination, as it would be independent of continental ice volume changes.

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1. Introduction

In view of the importance of sea ice in the climate system, much effort has been made by the community of paleoceanographers and paleoclimatologists for the development of proxies, which would permit to reconstruct past sea ice cover (e.g., Fisher et al., 2006). Amongst biogenic remains recovered in the deep-sea sediments, organic-walled dinoflagellate cysts (or dinocysts) are most useful tracers of sea ice cover (e.g., de Vernal

* Corresponding author. E-mail address: chm@uqam.ca (C. Hillaire-Marcel). and Hillaire-Marcel, 2000). They have been used to reconstruct the seasonal duration of sea ice in the northern North Atlantic during the last glacial maximum (de Vernal et al., 2000, 2005a) and the Holocene (de Vernal et al., 2005b; de Vernal and Hillaire-Marcel, 2006). However, dinocyst-based estimates of paleo-sea ice cover do not provide indication on the dynamics of sea ice, notably on its rate of formation. Recent studies of isotopic properties of planktic foraminifers from the Arctic Ocean suggest that their light ¹⁸O values, far out of isotopic values expected under the cold Arctic conditions, might well relate to the production of isotopically light brines resulting from sea ice formation (e.g., Bauch et al., 1997; Hillaire-Marcel et al., 2004). Therefore, they might indirectly respond to rates of sea ice

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production. From these observations we revisited planktic isotope records from the glacial northern North Atlantic, with special attention to the last glacial maximum (LGM), and to light isotopic excursion associated with Heinrich events that could relate to episodes of enhanced sea ice formation, linked to cooling and/or freshening events, rather than to a direct effect of dilution of paleo-sea surface waters by ¹⁸O-depleted meltwater and/or deepening of the surface water layer as often proposed (e.g., Clarke et al., 1999; Hillaire-Marcel and Bilodeau, 2000; Grousset et al., 2001; Rachid and Boyle, 2007).

2. Isotopic impact on Arctic foraminifers of sea ice formation

Planktic foraminifer data from the Arctic show practically monospecific assemblages composed of *Neogloboquadrina pachyderma*, with the left-coiled form being the same genetic species than the right-coiled form (Darling et al., 2006). In the Arctic Ocean and adjacent seas, both forms are characterized by large offset from theoretical equilibrium with regard to the oxygen isotopic composition of their carbonate shell (Fig. 1), assuming that carbonate precipitation occurs at approximately

mid-pycnocline depth between the cold and dilute surface water layer and the underlying warm and saline water mass (Fig. 2), originating from the North Atlantic-NAW-(cf. Also Aksu and Vilks, 1988; Kohfeld et al., 1996; Volkmann and Mensch, 2001; Simstich et al., 2003; Hillaire-Marcel et al., 2004). The offset of δ^{18} O values ranges from near -1% in the northern Greenland Sea to about -3% in the western Arctic Ocean (e.g., Bauch et al., 1997; Fisher et al., 2006). Nevertheless, despite this offset, size dependent δ^{18} O-gradients in *N. pachyderma* still preserve some temperature information. δ^{18} O-gradients between small (light) and large (dense) specimens are opposite to those observed in the North Atlantic, in response to the reverse temperature gradients of the pycnocline between the cold surface and warm underlying NAW (see Jones, 2001), as illustrated by Hillaire-Marcel et al. (2004). This observation led to attribute N. pachyderma a near isopycnal distribution with narrow linkage with density conditions (see also Simstich et al., 2003).

In almost all situations, δ values and potential densities (σ_{θ}) vary together, either in near co-linearity when salinity is determinant, or according to a polynomial relation when large δ and



Fig. 1. Sketch-map of the distribution of isotopic offsets between planktic foraminiferal (*Neogloboquadrina pachyderma* left-coiled — Npl) calcite δ^{18} O values and equilibrium conditions for a calcite precipitated at mid-depth along the pycnocline between the cold and dilute surface water layer of the Arctic Ocean and the underlying North Atlantic Water mass. The isotopic measurements in Npl populations are compiled from several sources. They include data from samples collected in the water column through plankton tow and in surface sediments (Spielhagen and Erlenkeuser, 1994; Kohfeld et al., 1996; Bauch et al., 1997; Volkmann and Mensch, 2001; Hillaire-Marcel et al., 2004). Note that both plankton tow and core top populations of Npl yield similar isotopic compositions (e.g., Bauch et al., 1997). The isotopic offsets are tentatively linked to the rate of production and accumulation of isotopically light brines along the pycnocline, due to sea ice formation. The Arctic topography map is from the International Bathymetric Chart of the Arctic Ocean (IBCAO; http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/ currentmap.html).

temperature gradients in the water column are involved (cf. Hillaire-Marcel et al., 2001a,b).

with
$$\delta^{18} O = \left[\frac{\binom{18}{0}}{\binom{18}{0}}_{\text{sample}} - 1 \right] \times 10^3$$

and $\sigma_{\theta} = \left[\frac{\rho_{\text{sample-water}}}{\rho_{\text{pure-water}}} - 1 \right] \times 10^3$,

where ρ stands for density.

From the above equation, it can be seen that σ_{θ} and δ^{18} O are both a direct function of salinity and a reverse function of temperature and should thus vary accordingly, with some departure from this pattern below 4 °C for freshwater that shows a maximum density at this temperature. However, sea ice formation may result in peculiar situation with respect to the δ vs. σ_{θ} relationship. In the Arctic, the overall isotopic offset of planktic foraminifers, as defined above and which ranges approximately -1 to -3% east-west, cannot be attributed to dilution with freshwater and lowering of salinity, notably because of ecological characteristics of N. pachyderma that would include stringent minimum salinity requirements (\geq 34.5; Hilbrecht, 1996), at least for Atlantic taxa. N. pachyderma has however a very large tolerance to high-salinity conditions and seems adapted to brine conditions. In vitro experiments demonstrate that it can still add chambers with salinities above 50 and survive for a few days with salinity above 80 (Spindler, 1996).

In the Arctic, brine extrusion occurring during sea ice growth is accompanied by isotopic fractionation that results in characteristic signatures in sea water (e.g., Tan and Strain, 1980; Bédard et al., 1981). The freezing of low-salinity surface water produces isotopically light brines sinking in the water column to the halocline where they mix with subsurface waters, whereas sea ice melting results in the addition of isotopically heavy, low-salinity water to the surface layer. One must note here that the isotopic enrichment of sea ice with a separation factor of near +3% under equilibrium conditions (O'Neil, 1968), although generally unattained here, due to kinetic and boundary layer effects when sea ice forms (see discussion in Ekwurzel et al., 2001), should result in a further depletion in ¹⁸O of the residual brines.

As a consequence of these converging processes, sea ice formation provides the only situation leading to opposite trends of the salinity and δ values relationship in oceanic water, thus in biogenic carbonates formed in such waters (Figs. 2-3). On these grounds, we have interpreted the large scale isotopic offset in Arctic N. pachyderma shells as the result of sea ice formation processes, with higher rates of sea ice growth yielding enhanced rates of brine extrusion and the production of an isotopically lighter "sea water" for a given salinity. This is consistent with the Holocene record of the Chukchi Sea in the western Arctic, which suggests maximum sea ice cover from dinocysts concomitant with particularly low δ^{18} O in planktic foraminifers during the early Holocene (cf. de Vernal et al., 2005b). Worth of mention is the fact that such distillation processes could lead to the formation within sea ice itself of high-salinity, low $\delta^{18}O$ micro-sites or "pockets" compatible with the ecological requirements of N. pachyderma (cf. Spindler and Dieckmann, 1986) and to the sinking of brines deeper in the water column, according to their density.

3. Records from the glacial North Atlantic

The above interpretation of isotopic compositions of planktic foraminifers, under Arctic conditions, leads us to revisit isotopic



Fig. 2. Conceptual linkage between *N. pachyderma* and salinity conditions in the Arctic. Left: salinity and temperature in the upper water column of the Western Arctic (data from Schmidt, 1999; Bigg and Rohling, 2000). Right: brines resulting from sea ice formation accumulate along the halocline (adapted from Aagaard, 1981) where *N. pachyderma* is likely to encounter suitable salinity conditions (above approximately 34.5; see text). *N. pachyderma* thus records, during its growth stages, the reverse temperature gradients characterizing the interface between the cold and dilute Arctic surface water layer, and the underlying warmer but more saline North Atlantic water mass.



Fig. 3. δ^{18} O vs. salinity relationships in surface waters (A and B) and deep waters (C). The δ^{18} O vs. salinity relationship in surface waters of the Western Arctic and North Atlantic is illustrated based on data from the Goddard Institute database (available at http://www.giss.nasa.gov/data/o18data/; cf. Schmidt, 1999; Bigg and Rohling, 2000). In the western Arctic (A) and the NW Atlantic (B), it defines linear relationships with an apparent freshwater end-member having an isotopic composition of about -40% and -20%, respectively, which are incompatible with freshwater signatures of Arctic Rivers (mean modern weighted values of approximately -17%; cf. Hélie et al., 2006). This provides independent evidence for the impact of sea ice brine distillation processes on the isotopic composition of Deep North Atlantic Water masses. Other evidence of the effect of sea ice formation on the ¹⁸O vs. salinity relationship comes from data obtained through the water column in western Fram Strait which indicate an "apparent" freshwater end-member of about -60% for the more saline component of the water mass flowing out through the Strait (cf. Volkmann and Mensch, 2001), which is certainly incompatible with any freshwater input from fluvial of glacial origin and rather suggests determinant impact of sea ice formation and brine addition to subsurface waters in the Arctic Ocean. Complementary evidence for large scale effects of sea ice formation on the ¹⁸O vs. salinity relationship is found from deep-water data of the modern southern Ocean and pore water data from the last glacial maximum (LGM) as shown in Figure C redrafted from Adkins et al. (2002).

records from the last glacial stage in the northwest North Atlantic.

During the last glacial maximum (LGM) as defined from EPILOG convention (Mix et al., 2001), sea ice cover was more extensive than at present in the North Atlantic, with sea ice spreading seasonally at least down to 40°N off North America and at about 55°N in the central and eastern North Atlantic (cf. Fig. 4). The LGM was characterized by Npl- δ^{18} O values of about +4.5‰ in most North Atlantic records (e.g., Weinelt

et al., 1996, 2003; de Vernal et al., 2002), which would discard any significant impact of sea ice formation at regional scale, except possibly during very brief excursions and/or along icesheet margins. On the contrary, all Heinrich events (H-events) show distinctive large amplitude light isotopic shifts of up to 2.5‰ near the major source area for ice-surging: the Hudson Strait (e.g., Clarke et al., 1999; Hillaire-Marcel and Bilodeau, 2000; see Fig. 5). In opposition to current interpretations, including ours' (Hillaire-Marcel and Bilodeau, 2000; see also



Fig. 4. Map showing the location of core HU91-045-094 and MD95-2024 (cf. Figs. 5–7) and the distribution of the restricted-perennial but variable seasonal sea ice cover in the northern North Atlantic during the LGM as derived from dinocyst assemblages (cf. de Vernal et al., 2005a). Each "x" sign corresponds to LGM data point. Note that the perennial sea ice limit in the Nordic Seas is compatible with the one that can be estimated from the distribution of planktic foraminifers (cf. Kucera et al., 2005; de Vernal et al., 2006). The continental ice limits are delimited after Peltier (1994a). The isobaths correspond to water depths of 200 and 1000 m.



Fig. 5. CaCO₃ data and δ^{18} O in planktic foraminifers from core HU91-045-094 (50° 12.26 N, 45° 41.14 W; Water depth=3448 m; see location in Fig. 4) in the northwest North Atlantic. Detrital carbonate pulses originating from Hudson Strait area highlight Heinrich layers at this site (e.g., Stoner et al., 1998).

Rachid and Boyle, 2007), such variations cannot be associated straightforwardly with low-salinity pulses and deepening of the halocline. For most of these isotopic excursions, theoretical paleosalinities calculated from ¹⁸O-data, assuming a linear or polynomial relationship between sea water salinity and $\delta^{18}O$ values in foraminifers not unlike that estimated for the Last Glacial Maximum (e.g., de Vernal et al., 2002), would yield a 32-33 range, i.e., a range likely too low for an "isopycnal" species such as N. pachyderma. These light isotopic excursions, with Npl- δ^{18} O near late Holocene values (Fig. 5), cannot be associated with higher sea surface temperature either, given the micropaleontological and IRD content of Heinrich layers suggesting particularly harsh conditions (Bond et al., 1992; de Vernal et al 2000). Adding the fact that the sea water was enriched in ¹⁸O, in comparison with the modern ocean during the last glacial stage, even higher temperatures than at present would be required to account for the almost similar Npl- δ^{18} O values characterizing the H-layer assemblages, which is an improbable scenario. However, model experiments from Mignot et al. (2007) suggest the possibility of some warming in the subsurface intermediate water layer during H-events (linked to heat accumulation in a "capped" layer), a situation not unlike that of the modern Arctic Ocean, with the warmer North Atlantic Water underlying, below a reverse thermocline, the cold and dilute surface water layer. This could indeed result in some temperature effect on the isotopic composition of deep dwelling foraminifers, but would not invalidate necessarily the above hypothesis of sea ice growth with production of isotopically light brines. However, isotopic measurements on subpopulations of Npl, according to their size, in the H-layers examined here (Fig. 6), yielded heavier isotopic composition in large shells vs. smaller ones, as in the late Holocene North Atlantic assemblages, but a trend opposite to that of the Arctic Ocean (Hillaire-Marcel et al., 2004). This suggests a negative temperature gradient along the thermocline between surface waters and the underlying water layer during H-events, thus weakening the hypothesis of higher temperatures in the deep habitat of Npl.

We thus suggest that sea ice might well be the "joker" here, and account for the addition of isotopically light brines along the pycnocline and deeper in the water column (cf. Tan and Strain, 1980) inducing at least part of the large isotopic shift recorded by *N. pachyderma* during these events.

An independent evidence for the above hypothesis is provided by sea ice reconstructions from dinocyst data in a core from Orphan Knoll in southern Labrador Sea (cf. Fig. 7). At this site, H1 is characterized by an isotopic excursion in N. pachyderma exceeding -2% that is concomitant with maximum sea ice cover as estimated from dinocyst assemblages. Of particular interest is the fact that the sea ice peak is decoupled from salinity as reconstructed from dinocyst assemblages that does not show significant change. This suggests that the isotopic shift does not relate to a major salinity change at regional scale, but to another process, which could be the production of isotopically light brines according to our hypothesis. Also notable is the fact that both the isotopic shift and sea ice maximum occur above the initial peak of detrital calcium carbonate and extent above the Heinrich layer as defined from sedimentological evidence. The observed features are consistent with the current interpretation of Heinrich events that are associated with ice surges in Hudson Strait leading to massive iceberg discharge in the North Atlantic (cf. Bond et al., 1992). They show that the ice surge acted as trigger mechanisms for a very severe cooling, which was accompanied by the formation of dense sea ice cover



Fig. 6. Size-dependant isotopic composition of Npl shells. Late Holocene assemblages from the Western Arctic show a negative δ^{18} O-trend between small and large specimens linked to a reverse thermocline between the surface and underlying warm North Atlantic Water masses, whereas assemblages from H-layers in the northwest North Atlantic core MD2024 (50°12.40 N, 45°41.22 W; Water depth: 3539; cf. Hillaire-Marcel and Bilodeau, 2000) show a positive trend, practically identical to that of late Holocene assemblages in the area, thus a normal thermocline between surface and subsurface water masses during the corresponding events ("Modern NW Atlantic", i.e., late Holocene data are from Hillaire-Marcel et al., 2004). The relatively light δ^{18} O values of the H-layers assemblages are interpreted here as resulting from some addition of isotopically depleted brines, along the halocline, assuming colder temperatures and an ¹⁸O-enriched sea water during H-events, in comparison with late Holocene conditions.



Fig. 7. Blow up on H1 and H0 in core HU-91-045-094. Calibrated ¹⁴C ages are from *N. pachyderma* (left-coiled) populations, with the exception of the age of 10.48 ± 0.17 yielded by *Globigerina bulloides* shells (Hillaire-Marcel et al., 1994). Dinocyst-based reconstructions of sea ice cover, and sea surface salinity are from de Vernal et al. (2005a): the thin lines correspond to estimates calculated from the 5 best modern analogues, and the thick lines correspond to 3-point running averages; uncertainties in paleosalinity reconstructions average ± 1 (de Vernal et al., 2005a,b). Note i) the decoupling between salinity conditions in the surface water layer and the "light" isotopic event recorded by *N. pachyderma* (filled circles = oxygen isotopes; open circles = carbon isotopes) and ii) the low-salinity range of surface waters during the whole interval suggesting that due to its salinity requirements, Npl should have developed deeper along the halocline, where higher salinities prevailed. Isotopic data in Npl are from Hillaire-Marcel and Bilodeau (2000).

in the Labrador Sea, as suggested from both δ^{18} O in planktic foraminifers and dinocyst data.

Interestingly, in the Orphan Knoll core, H0 follows a slightly contrasted overall pattern (Fig. 6), with a smaller isotopic shift of approximately -1‰. A low-salinity peak in surface waters, reconstructed from dinocyst data, corresponds to the base of the carbonate layer and to enhanced sea ice cover. It however precedes the light ¹⁸O-peak, which matches a trend toward higher salinity, possibly due to brine addition. In this case, the ice surge was apparently accompanied by meltwater outflow and iceberg discharge, which led to cooling and regional sea ice formation then responsible for subsequent production of isotopically light brines recorded deeper in the water column by *N. pachyderma*.

The isotopic shift towards lighter ¹⁸O values characterizing all H-events in the Labrador Sea (Fig. 4) and to a lesser extent the Nordic Seas (e.g., Dokken and Jansen, 1999) could thus, at least partly, respond to the production of isotopically light brines formation linked to high rates of sea ice formation fostered by cooling and freshening of surface waters due to the spreading of icebergs into the North Atlantic from surging areas. In such an hypothesis, the reduction of the Atlantic Meridional Overturning Circulation (AMOC) that has been proposed, in particular during H1 or H4 (e.g., Zahn et al., 1997), and which is illustrated here by a decrease in δ^{13} C values in planktic foraminifers (Fig. 7; cf. also Hillaire-Marcel and Bilodeau, 2000) would not relate exclusively to salinity and sea surface density variation in the northern North Atlantic, but also likely to the spreading of dense sea ice cover that pushed further south the location of sites where deep-water formation could possible have occurred. However, the sinking of brines could have

maintained a sluggish AMOC, during the interval and possibly contributed to the resumption of convective processes and of deep North Atlantic water formation that followed rapidly Heinrich events. Nonetheless, the cooling that has characterized large areas of the Northern Hemisphere during H-events would have been fostered primarily by the sea ice spreading in the Northern North Atlantic as suggested by dinocyst data.

4. Discussion

Recent studies on the calcification depths and specific ecological behaviours of planktic foraminifers (e.g., Bauch et al., 1997; Spero et al., 2003; Hönisch et al., 2003) have considerably modified our interpretations of their stable isotope compositions with respect to the reconstruction of paleo-sea surface salinities and temperatures (LeGrande et al., 2004). Their bathymetric distribution, the presence or not of symbiotic algae, their seasonal development, the shell density and many other factors are now considered prior to any attempt at interpreting biometrical and geochemical data from fossil populations (cf. Waelbroeck et al., 2005; Ravelo and Hillaire-Marcel, 2007).

The present paper also leads to question deeper the implicit (linear or polynomial) relationship often postulated between oxygen isotopes in foraminiferal calcite and the ocean salinity. The relationship is obviously not valid in the Arctic Ocean due to the distillation effect of sea ice formation (cf. Fig. 3A). As illustrated in Fig. 3B, this relationship is not straightforward either in the northwest North Atlantic where surface waters apparently preserve an "isotopic memory" of processes linked to sea ice formation in the Arctic. In the northwest North

Atlantic, the linear distribution of δ^{18} O vs. salinity values corresponds to an "apparent" freshwater end-member with an isotopic composition below -20%, whereas the weighted isotopic composition of Arctic rivers is approximately -17%(Hélie et al., 2006). Addition of isotopically light brines partly accounts for the slope of this δ^{18} O-salinity linear regression line (0.57% per salinity unit — su). This "forcing" of the $\delta^{18}O$ salinity relationship is also illustrated, in the Western Arctic data set of Fig. 3A, by the cluster of data points representing conditions along the pycnocline between surface waters and the NAW, around a regression line with a slope of 1.14‰/su, whereas shallower waters show a scatter of values influenced by the melting of the low-salinity-high δ^{18} O sea ice balancing the low δ^{18} O-brines that sank deeper in the water column when formed (cf. Bédard et al., 1981). This pattern is even more critical, when examining the δ^{18} O–salinity relationship of Deep North Atlantic Water (DNAW) masses (e.g., Adkins et al., 2002), which shows a slope of about 0.82‰/su. Therefore, in addition to salt and ¹⁸O enrichment of surface ocean waters in the equatorial ocean, the contribution of isotopically light brines in the polar oceans possibly obscures the linkage between ¹⁸O changes in ocean water masses and the overall ocean salinity and isotopic composition. Variations of sea ice formation rates have certainly occurred during the late Cenozoic, thus resulting in variable distillation rates of isotopically heavy sea ice meltwater in surface water masses, and of isotopically light brines added to deeper water masses. As a consequence, transcription of paleo-sea water δ^{18} O values, either derived from pore water studies (e.g., Adkins et al., 2002) or from benthic foraminiferal calcite analysis (e.g., Chappell and Shackleton, 1986; Marsiat and Berger, 1990; Lindsey, 1996), into paleosalinity and paleo ice volumes (and subsequently into paleo-sea levels) may require some deeper examination. Some conflicting observations and inferences, for example about LGM low sea-levels, as highlighted by Peltier (1994b), might found here some explanation.

5. Conclusion

In the present paper, we have examined the isotopic properties of planktic foraminifers from the Arctic under late Holocene climatic conditions. We interpret their ¹⁸O-depletion with respect to equilibrium conditions as the result of sea ice production processes leading to the distillation of isotopically light brines. These brines eventually accumulate along the halocline and deeper. They thus contribute to increasing salinity of the subsurface Arctic water that flow southward into North Atlantic, thus providing a characteristic δ^{18} O vs. salinity signature to the resulting North Atlantic deep-water masses.

On these grounds, we revisited the isotopic properties of planktic foraminifers from the last glacial maximum and Heinrich layers in the northwest North Atlantic. The high δ^{18} O values of *Neogloboquadrina pachyderma* during the last glacial maximum do not provide evidence for significant sea ice production, whereas the light δ^{18} O values characteristic of Hevents suggest intense production of sea ice, with addition of isotopically light brines at halocline level and below it. Thus, H- events would represent the only intervals when Arctic-type conditions settled in the glacial northern North Atlantic.

Shifts in oxygen isotope values of foraminiferal calcite, particularly in high latitude records, must have been partly controlled by rates of sea ice formation. Thus, they should not be unequivocally interpreted in terms of paleosalinities and/or paleotemperature changes. Because deep waters are formed in high latitude sites, where sea ice production occurs and has likely varied through time, the isotopic vs. salinity properties of deep ocean water masses may show some departure from any given straightforward linear relationship. This might have to be taken into account when attempting to estimate variations in ice volume (and sea-levels) from δ^{18} O in benthic foraminifers.

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