



Intermediate and deep water paleoceanography of the northern North Atlantic over the past 21,000 years

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[1] Benthic foraminiferal stable isotope records from four high-resolution sediment cores, forming a depth transect between 1237 m and 2303 m on the South Iceland Rise, have been used to reconstruct intermediate and deep water paleoceanographic changes in the northern North Atlantic during the last 21 ka (spanning Termination I and the Holocene). Typically, a sampling resolution of ~100 years is attained. Deglacial core chronologies are accurately tied to North Greenland Ice Core Project (NGRIP) ice core records through the correlation of tephra layers and changes in the percent abundance of *Neogloboquadrina pachyderma* (sinistral) with transitions in NGRIP. The evolution from the glacial mode of circulation to the present regime is punctuated by two periods with low benthic $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values, which do not lie on glacial or Holocene water mass mixing lines. These periods correlate with the late Younger Dryas/Early Holocene (11.5–12.2 ka) and Heinrich Stadial 1 (14.7–16.8 ka) during which time freshwater input and sea-ice formation led to brine rejection both locally and as an overflow exported from the Nordic seas into the northern North Atlantic, as earlier reported by Meland et al. (2008). The export of brine with low $\delta^{13}\text{C}$ values from the Nordic seas complicates traditional interpretations of low $\delta^{13}\text{C}$ values during the deglaciation as incursions of southern sourced water, although the spatial extent of this brine is uncertain. The records also reveal that the onset of the Younger Dryas was accompanied by an abrupt and transient (~200–300 year duration) decrease in the ventilation of the northern North Atlantic. During the Holocene, Iceland-Scotland Overflow Water only reached its modern flow strength and/or depth over the South Iceland Rise by 7–8 ka, in parallel with surface ocean reorganizations and a cessation in deglacial meltwater input to the North Atlantic.

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1. Introduction

[2] The most recent glacial termination (Termination I) occurred between ~19 and 7 ka, following a maximum in global ice volume at ~21 ka (Last Glacial Maximum (LGM)). It is widely accepted that Termination I was initiated by gradual changes in solar insolation, due to orbital configuration, while rising atmospheric carbon dioxide concentrations and albedo changes provided a strong positive feedback [e.g., Imbrie et al., 1992; Shackleton, 2000; Cheng et al., 2009]. Ice core records from Greenland show that the climate amelioration in the North Atlantic and surrounding regions was not smooth, but included a series of abrupt step-like changes and reversals. The role of the oceans in globally redistributing heat and partitioning carbon dioxide, and their ability to switch between different states of equilibrium, has led to the proposal that changes in the Atlantic meridional overturning circulation (AMOC), and associated

modes of deepwater formation, are responsible for the sharp climate transitions seen in the Greenland ice core records [e.g., Broecker and Denton, 1989; Sarnthein et al., 2000; Rahmstorf, 2002].

[3] Numerous studies have since confirmed that glacial and stadial intervals are associated with a shoaling and/or weakening of deep convection in the North Atlantic and a related shift in the location of open ocean convection from north to south of the Greenland-Scotland Ridge (GSR) [e.g., Dokken and Jansen, 1999; Sarnthein et al., 2000; McManus et al., 2004; Piotrowski et al., 2004; Lynch-Stieglitz et al., 2007, and references therein; Meland et al., 2008]. Sites proximal to the GSR have proven particularly sensitive for reconstructing changes in the convective mode of the North Atlantic. Recent studies from this region have suggested that oceanic reorganizations throughout Termination I may be more complicated than previously envisaged: Meland et al. [2008] used benthic foraminiferal oxygen isotopes to infer episodic overflow over the GSR of brines from the Nordic seas during Heinrich Stadial 1 (HS-1) and the Younger Dryas (YD), while Rickaby and Elderfield [2005] provide foraminiferal Cd/Ca and $\delta^{13}\text{C}$ evidence that instead suggests the incursion of Antarctic Intermediate Water to the high-latitude North Atlantic during HS-1 and the YD. Yu et al. [2008] have used foraminiferal B/Ca and $\delta^{13}\text{C}$ measurements to infer the

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Table 1. South Iceland Rise Core Locations

Site	Core Number	Core Type	Latitude (N)	Longitude (W)	Water Depth (m)	Core Length (m)
CD159-10	RAPiD-10-1P	Piston	62°58.53'	17°35.37'	1237	6.16
CD159-12	RAPiD-12-1K	Kasten	62°05.43'	17°49.18'	1938	6.39
CD159-15	RAPiD-15-4P	Piston	62°17.58'	17°08.04'	2133	11.21
CD159-17	RAPiD-17-5P	Piston	61°28.90'	19°32.16'	2303	14.41

presence of a third water mass (likely brine overflow from the Nordic seas; in addition to Glacial North Atlantic Intermediate Water (GNAIW) and Glacial Antarctic Bottom Water (GAABW)) during the LGM. It is clear that further investigation is required to constrain the oceanic changes that occurred in this critical region during the last deglaciation. Moreover, many existing deglacial records do not have sufficient resolution to clearly resolve the timing and phasing of oceanic changes during abrupt climate reversals such as the YD, undermining efforts to document the role ocean circulation played during such events.

[4] In this study, we provide high-resolution paleoceanographic records from locations adjacent to the GSR, in order to determine the nature and timing of changes in deep convection in the northern North Atlantic throughout Termination I. Our work builds upon regional syntheses of benthic isotopes by *Sarnthein et al.* [1994, 2000] and *Meland et al.* [2008] by providing a series of high-resolution cores from south of the GSR that can clearly resolve the abrupt changes of the deglaciation. Four ocean sediment cores from the South Iceland Rise (SIR) are used to form a depth transect between 1237 m and 2303 m and provide typical sampling intervals of between 50 and 100 years. The presence of tephra layers and changes in the percent abundance of the polar species *N. pachyderma* (sinistral) within the sediment cores allows accurate correlation to the North Greenland Ice Core Project (NGRIP) ice core, producing tightly constrained age models. Intermediate and deep water mass properties are reconstructed using benthic foraminiferal stable isotope data, alongside planktonic stable isotope records for comparison. The study site's close proximity to the primary source of terrigenous sediment (Iceland), the regular input of volcanic detritus, and strong Holocene currents severely hampered the use of the sortable silt mean grain size method [McCave et al., 1995; McCave and Hall, 2006]. Therefore grain size analysis is restricted to establishing the onset of the modern circulation regime during the early Holocene through the identification of winnowed deposits caused by strong bottom currents.

2. Oceanographic Setting and Study Location

[5] In the modern North Atlantic, deep water formation sites are largely supplied by the northward flow of warm, saline, surface and thermocline waters from the low-latitude North Atlantic, via the North Atlantic Current (NAC). Deep water formation occurs in the Nordic and Labrador seas, where atmospheric cooling promotes deep convection. Deep water formed in the Nordic seas overflows the GSR entraining thermocline and subthermocline water to produce Iceland-Scotland, Denmark Straits and Wyville-Thompson Overflow

Water (ISOW, DSOW and WTOW, respectively) [Hansen and Østerhus, 2000], which flow southward as deep return flows, with respective fluxes of 3.5, 2.9, 0.3 Sv [Dickson and Brown, 1994]. These are the main constituents of the “deep northern boundary current” of McCartney [1992], which becomes Lower North Atlantic Deep Water (LNADW). This has a flux around the south of Greenland of ~13 Sv [Dickson and Brown, 1994], to which is added a shallower component formed by deep convection in the Labrador Sea. The latter, Labrador Sea Water (LSW), overlying and mixing with LNADW, flows along the Labrador margin as North Atlantic Deep Water (NADW) [Schmitz and McCartney, 1993].

[6] North Atlantic circulation differed significantly from the modern regime during the LGM: overflows were diminished, a shallow convection cell formed south of Iceland producing GNAIW, and there was a replacement at depth of NADW with southern sourced Glacial Antarctic Bottom Water (GAABW) which filled the North Atlantic up to ~2 km [Oppo and Lehman, 1993; Lynch-Stieglitz et al., 2007].

[7] The four cores sites examined in this study (Table 1) were obtained during cruise CD-159 of RRS *Charles Darwin* during July 2004 [McCave, 2005]. The cores are located on the flanks of the East and West Katla Ridges, two thick sediment deposits striking ~210° from the continental slope on the South Iceland Rise, formed by the Neogene denudation of Iceland [Lonsdale and Hollister, 1979; Shor, 1980] (Figure 1). Sediment is delivered to the continental slope by turbidity currents and is subsequently entrained and transported southwestward by the flow of Iceland-Scotland Overflow Water.

[8] The core sites are situated under the flow paths of the deep ISOW and the surface NAC. The three deepest cores (RAPiD-12-1K, RAPiD-15-4P and RAPiD-17-5P; hereafter 1K, 4P, and 5P, respectively) are situated within the core of ISOW, in which CTD data taken onboard CD-159 recorded typical temperatures of 2.0°C and salinity (*S*) of 35.0 practical salinity unit (psu). The temperature and salinity at the shallowest core, RAPiD-10-1P (hereafter 1P), was 4.5°C and 35.0 psu, with the core site being bathed by Labrador Sea Water (LSW) and Intermediate Water (IW), the origin of which is uncertain (either aged Arctic Intermediate Water or northward flowing water from the European and African margin) [van Aken and de Boer, 1995]. Modern deep winter mixing events homogenize the water column to depths of 600 m, producing Subpolar Mode Water (SPMW, 8–9°C, 35.2–35.3 psu).

[9] The main flow axis of ISOW over the Katla Ridges lies between 1400 and 1800 m [Shor, 1980], while further west on Björn Drift it is located between 1626 and 1827 m [Bianchi and McCave, 2000]. Long-term current meters situated ~100 km upstream (east) of the Katla Ridges record

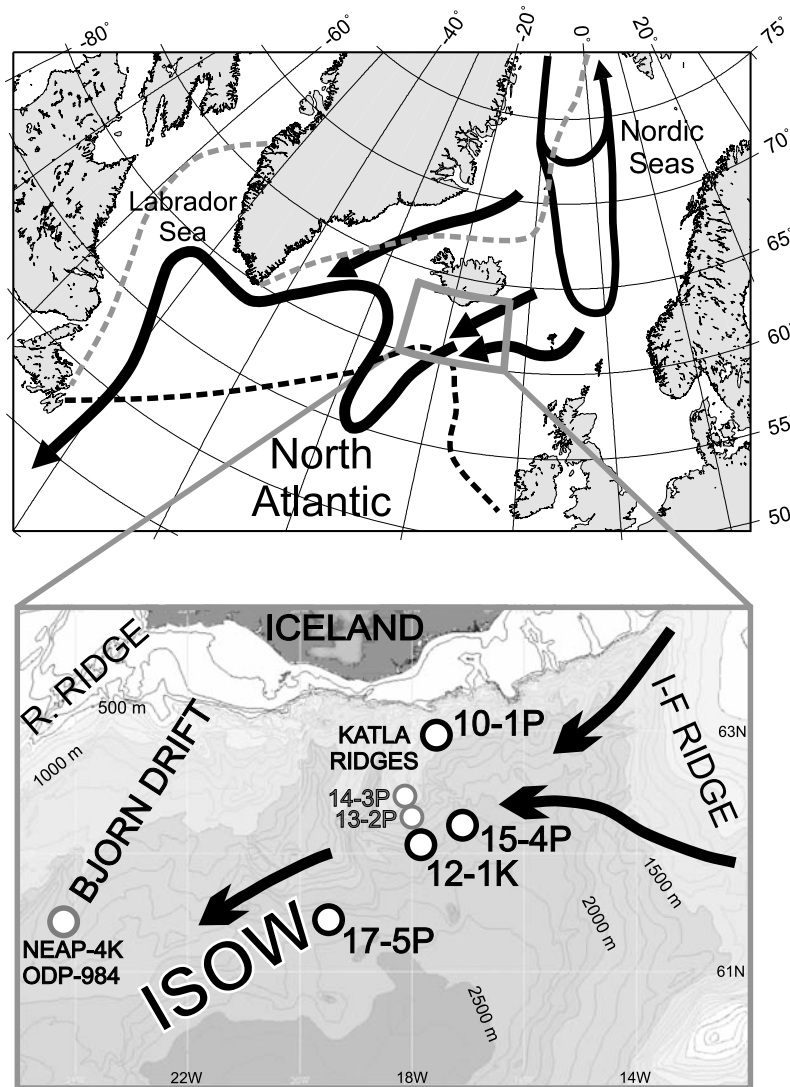


Figure 1. (top) Schematic map of the major deep water currents (black arrows) of the northern North Atlantic. Also shown are approximate positions for the winter sea ice margin during the LGM (black dashed line, based on Glacial Atlantic Ocean Mapping 2000 [Pflaumann *et al.*, 2003]) and modern (gray dashed line). (bottom) Enlarged section of the northern Icelandic Basin shows the core locations (white dots with bold black outline are the main study sites and smaller dots with gray outline are the core sites not studied in detail). RAPiD prefix has been omitted from the labels for RAPiD-10-1P, RAPiD-12-1K, RAPiD-13-2P, RAPiD-14-3P, RAPiD-15-4P, and RAPiD-17-5P. The locations of NEAP 4K [Hall *et al.*, 2004; Rickaby and Elderfield, 2005] and ODP Site 984 [Praetorius *et al.*, 2008] are shown for reference. Black arrows in Figure 1 (bottom) indicate the pathway of the main flow of Iceland-Scotland Overflow Water.

near-bottom flow speeds of 15–20 cm/s between 1600 and 2000 m [Saunders, 1996] and persistent flow of 20 cm/s between 1400 and 1800 m on the Katla Ridges [Shor, 1980], sufficient to resuspend lithic grains and move foraminifera in excess of 125 μm diameter. Evidence for this vigorous flow of ISOW can be found in cores 3P (62°27.26'N, 18°05.49'W, 1475 m) and 2P (62°12.93'N, 17°59.58'W, 1751 m) (see Figure 1 for locations), in which the upper meter of sediment contains iron-oxide-stained glacial foraminifera and lithic fragments, and no Holocene sediment. Current

speeds on the flanks of the main flow axis are sufficient to produce winnowed foraminiferal sand deposits in the mid to late Holocene sections of cores 1K (1938 m) and 4P (2133 m).

3. Methods

3.1. Sample Preparation

[10] Sediment samples were disaggregated on a rotating wheel for 24 h, before being sieved at 63 μm using deionized water. The coarse (>63 μm) and fine (<63 μm) fractions were

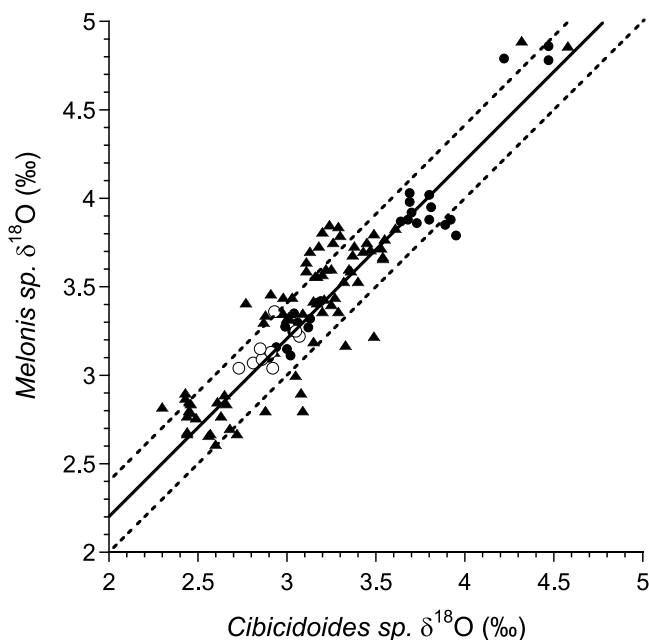


Figure 2. Comparison of stable oxygen isotope analyses from the same depth intervals for *Cibicidoides* species and *Melonis* species. Analyses from RAPiD-10-1P (solid triangles) were on *M. barleanuum* and mixed *Cibicidoides*; analyses from RAPiD-15-4P (large open circles) and RAPiD-17-5P (solid circles) were on *M. pompilioides* and *C. wuellerstorfi*. The data is best described by a 1:1 relationship with an offset of $0.24 \pm 0.2\text{‰}$, shown by the solid line ($n = 115$, $r^2 = 0.86$).

then weighed to obtain the percent sand fraction (wt. % sand). Eighty to one hundred tests of the planktonic foraminifera *Neogloboquadrina pachyderma* sinistral (*Nps*) were picked from the 150–250 μm fraction. Benthic foraminifera were picked from the $>212 \mu\text{m}$ size fraction ($>250 \mu\text{m}$ size fraction where possible), avoiding any damaged or discolored tests. Samples typically consisted of between 3 and 40 shells. The epifaunal species *Cibicidoides wuellerstorfi* was analyzed from cores 1K, 4P and 5P, while in the shallower core, 1P, where *C. wuellerstorfi* was not available, *Cibicidoides pachyderma* and mixed *Cibicidoides* were used. Specimens of the infaunal species *Melonis pompilioides* (abundant in the deeper cores, 4P and 5P) and *Melonis barleanuum* (abundant in 1K and 1P) were also analyzed to support the *Cibicidoides* data. Large samples were crushed, homogenized and split to produce $\sim 100\text{--}150 \mu\text{g}$ for stable isotopic analysis.

3.2. Stable Isotope Analysis

[11] The stable isotopic compositions of the benthic foraminifera were used to reconstruct the intermediate and deep water paleoceanography of the South Iceland Rise. Stable isotope measurements were performed using the Godwin Laboratory VG Prism mass spectrometer attached to a Micromass Multicarb Sample Preparation System. Measurements of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were determined relative to the

Vienna Peedee belemnite (VPDB) standard, with an analytical precision better than 0.08‰ and 0.06‰ , respectively. To aid interpretation, benthic oxygen isotope data were corrected for the global shifts in seawater $\delta^{18}\text{O}$ caused by the decay of continental ice sheets (hereafter termed “ $\delta^{18}\text{O}$ ice volume corrected,” $\delta^{18}\text{O}_{\text{ivc}}$), assuming a 1‰ glacial maximum shift scaled to the Fairbanks [1989] sea level curve.

[12] *C. wuellerstorfi* and other species of *Cibicidoides* have been widely used to reconstruct past changes in seawater $\delta^{13}\text{C}$. Coexisting *C. wuellerstorfi* and other species of *Cibicidoides* exhibit no significant differences in $\delta^{13}\text{C}$ [Curry et al., 1988]. The *Cibicidoides* taxonomic group reliably records deep water properties [Curry and Oppo, 2005]; although the epibenthic *C. wuellerstorfi* and most other *Cibicidoides* species secrete calcite not in isotopic equilibrium with the ambient bottom water $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$, but in a constant 1:1 relationship [Mackensen and Bickert, 1999]. Traditionally, *C. wuellerstorfi* $\delta^{18}\text{O}$ values have been adjusted by $+0.64\text{‰}$ to be placed on a *Uvigerina* equivalent scale, although it has since been argued that *C. wuellerstorfi* secretes calcite in close isotopic equilibrium with seawater $\delta^{18}\text{O}$ [Bemis et al., 1998]. In this study, no vital effect correction has been applied to the *Cibicidoides* data. Similarly, no vital effect correction has been applied to the *Nps* $\delta^{18}\text{O}$ values. (The presence of a vital effect is likely although it is poorly constrained and may be highly variable, depending upon the degree of secondary encrusting calcite [Kozdon et al., 2009].)

[13] Based on Atlantic studies, *Melonis* spp. are infaunal, abundant in the top 4–5 cm of sediment, and do not actively migrate within the sediment column [Tachikawa and Elderfield, 2002; Mackensen, 2004]. The stable carbon isotope compositions of many infaunal species are unreliable recorders of bottom water $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$. Furthermore, significant species offsets for $\delta^{18}\text{O}$ values exist [Shackleton and Opdyke, 1977; Duplessy et al., 1980; Vidal et al., 1998]. Previous studies have found constant offsets between $\delta^{18}\text{O}$ values recorded by *C. wuellerstorfi* and *Melonis* spp. [e.g., Labeyrie et al., 1992; Vidal et al., 1998; Kristjansdottir, 2005]. $\delta^{18}\text{O}$ measurements on *Cibicidoides* spp. and *Melonis* spp. from the same sample depths confirm a near constant offset of $0.24 \pm 0.2\text{‰}$ (Figure 2), and throughout this work *Melonis* spp. $\delta^{18}\text{O}$ values are corrected to *Cibicidoides* spp. by subtracting 0.24‰ .

4. Age Models

[14] Dating by ^{14}C accelerator mass spectrometry (AMS) of monospecific planktonic foraminifera samples (*Globigerina bulloides* for the Holocene, *Nps* for the LGM) produced age models for the cores throughout the Holocene and the LGM using a surface radiocarbon reservoir age of 400 years and 800 years, respectively. Highly variable surface radiocarbon reservoir ages throughout the deglaciation [Waelbroeck et al., 2001] add a large uncertainty (several hundred years) to ^{14}C based deglacial age models. The age models through this interval are therefore constructed by correlating tephra layers found in both the sediment cores and the annual layer counted NGRIP ice core, as well as correlating abrupt coolings and

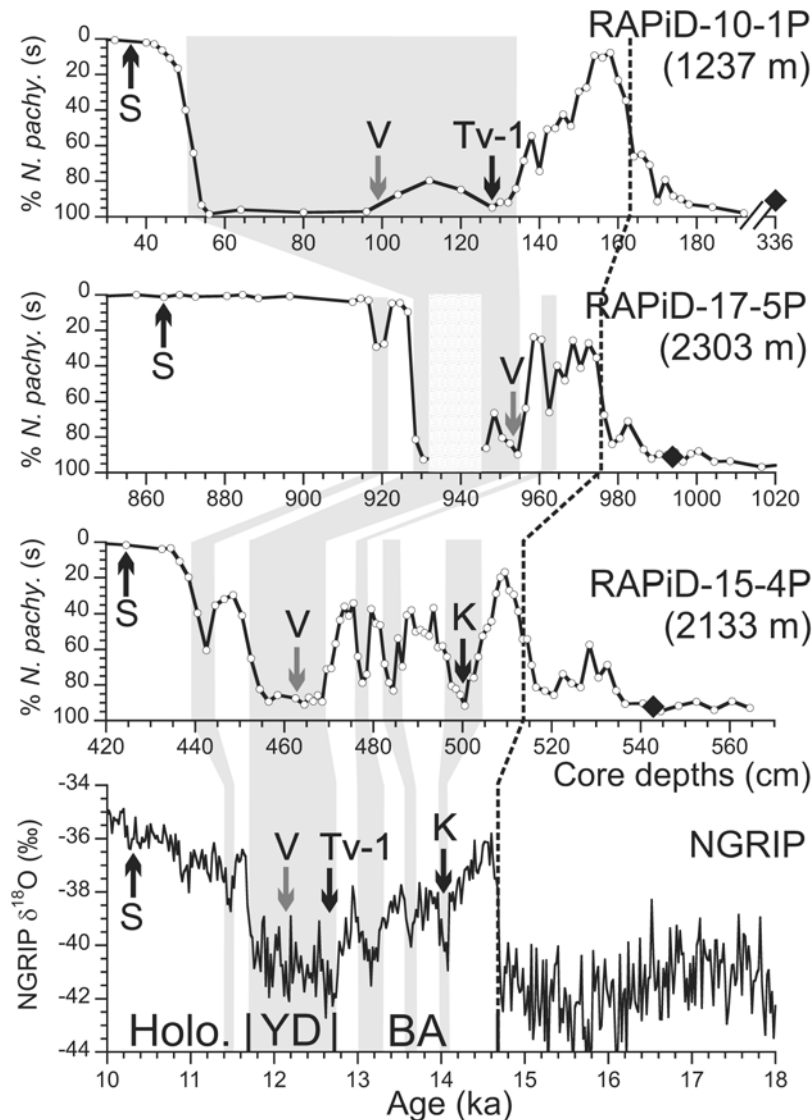


Figure 3. Correlation between the RAPiD cores and the NGRIP ice core [North Greenland Ice Core Project Members, 2004] used to produce the deglacial age models. Gray shading and the dashed line (HS-1 to BA transition) indicate correlation between changes in the percent abundance of *Nps* (150–250 μm fraction) and cooling or warming as indicated by NGRIP. The base of tephra layers are indicated by the arrows (gray arrows are rhyolite, black arrows are basaltic, S is Saksunarvatn Ash, V is Vedde Ash, Tv-1 is Tv-1 tephra, and K is Katla-derived tephra layer [Mortensen *et al.*, 2005]). The black diamonds mark the initial deglacial decreases in benthic foraminifera $\delta^{18}\text{O}$, tied at 17.2 ka. Cross-hatched area shown for RAPiD-17-5P indicates the presence of a turbidite; analyses were not conducted on this section of core.

warmings, as indicated by the percent abundance of the polar species *Nps*, with similar events in the NGRIP ice core [North Greenland Ice Core Project Members, 2004; Rasmussen *et al.*, 2006] (Figure 3). This established technique assumes cooling over Greenland is tightly coupled to North Atlantic sea surface temperatures, which has been previously demonstrated in studies throughout the northern North Atlantic [e.g., Lehman and Keigwin, 1992; Bond *et al.*, 1993; Haflidason *et al.*, 1995; Peck *et al.*, 2006; Knutz *et al.*, 2007], as well as in RAPiD-15-4P [Thornalley *et al.*, 2010].

Discrete tephra layers identified from detrital counts were examined visually and by electron probe microanalysis [Thornalley, 2008] and correlated with tephra layers recorded in NGRIP [Mortensen *et al.*, 2005]. Counts were conducted on splits of the 150–250 μm size fraction ensuring over 300 grains/tests were counted where possible. An additional age constraint between the LGM and the onset of the Bølling-Allerød (BA) was provided by tying initial deglacial decreases in benthic $\delta^{18}\text{O}$ in the RAPiD cores to the same event in the well dated North Atlantic core SU81-18 (38°N, 10°W)

Table 2. Age Model for RAPiD-10-1P^a

	Sediment Depth (cm)	¹⁴ C Age ^b (years B.P.)	±1σ Error (years)	Calendar Age (ka before year A.D. 2000)
¹⁴ C AMS SUERC 14094 ^c	8.5	1.364	35	1.390
Saksunarvatn ash	38.5	—	—	10.347
Tied to NGRIP δ ¹⁸ O	51.5	—	—	11.703
Vedde ash top ^d	70.5	—	—	12.170
Vedde ash base	98.5	—	—	12.171
Tv-1 tephra	128.5	—	—	12.65
Tied to NGRIP δ ¹⁸ O	135.5	—	—	12.76
Tied to NGRIP δ ¹⁸ O	158.5	—	—	14.6
Tied to NGRIP δ ¹⁸ O	192.5	—	—	14.74
Tied benthic δ ¹⁸ O	312.5	—	—	17.2

^aHere ¹⁴C ages are radiocarbon years before A.D. 1950, calendar ages are before A.D. 2000 (in accordance with the NGRIP GICC05 age).

^bNo reservoir applied.

^cSUERC (Scottish Universities Environmental Research Centre) is the prefix assigned to ¹⁴C dates run by the UK NERC radiocarbon facility in Glasgow.

^dThick deposit associated with Vedde ash, including ash at base and top with a muddy turbidite containing no foraminifera between. Assumed near instantaneous deposition.

[Waelbroeck *et al.*, 2001], located at a latitude where surface reservoir age variations are expected to be minor. Correlation of the cores to NGRIP is illustrated in Figure 3, while details of the tie points used are given in Tables 2–4. Previously published age models are used for cores 1K [Thornalley *et al.*, 2009] and 4P [Thornalley *et al.*, 2010]. (The age model for 1K, spanning the Holocene, is based on 13 *G. bulloides* ¹⁴C AMS dates.) Because age models have been tuned to the Greenland Ice-Core Chronology 2005 (GICC05) timescale, dates presented in this study are given in ky before A.D. 2000.

5. Results

5.1. Last Glacial Maximum and Early Deglacial (21–17 ka)

[15] Figure 4 shows that the stable oxygen isotope records (Figure 4) all vary by more than the whole ocean change of

1 ± 0.1‰ across Termination I [Schrag *et al.*, 1996], confirming the established view that during the LGM intermediate and deep water were relatively colder and/or more saline than at present. Figure 5 shows that during the LGM, all three cores record benthic δ¹⁸O_{ivc} values of ~3.5–3.7‰, in agreement with other North Atlantic LGM data [e.g., Curry and Oppo, 2005]. These values compare with Holocene δ¹⁸O_{ivc} values of 2.3‰ for the shallowest core (1237 m), and 2.8‰ for the deeper cores (1938–2303 m). Attributing the heavier δ¹⁸O_{ivc} values solely to temperature change (~0.25‰ ≡ 1°C [Shackleton, 1974]) implies cooling from 4.5°C to −0.7°C at 1237 m, and 2°C to −1.2°C at ~2000 m. This is consistent with a glacial North Atlantic temperature at 2.1 km of −1°C, and a salinity of ~36.1 psu (the salinity is 1.1 psu higher than the modern (~35 psu) because of the global sea level lowering of ~120 m) [Adkins *et al.*, 2002].

[16] A strong gradient in δ¹³C existed during the LGM between 1.5‰ at 1237 m, and 0.6‰ at 2303 m. This is

Table 3. Age Model for RAPiD-15-4P^a

	Sediment Depth (cm)	¹⁴ C Age ^b (years B.P.)	±1σ Error (years)	Calendar Age (ka b2k)
¹⁴ C AMS SUERC 14081	168.5	6.520	37	7.139
¹⁴ C AMS SUERC 14084	368.5	7.906	36	8.469
¹⁴ C AMS SUERC 14085	416.5	8.796	37	9.569
Saksunarvatn ash	426.5	—	—	10.347
Tied to NGRIP δ ¹⁸ O	442.5	—	—	11.46
Tied to NGRIP δ ¹⁸ O	452.5	—	—	11.703
Vedde ash	462.5	—	—	12.171
Tied to NGRIP δ ¹⁸ O	469.5	—	—	12.78
Tied to NGRIP δ ¹⁸ O	476.5	—	—	13.02
Tied to NGRIP δ ¹⁸ O	479	—	—	13.32
Tied to NGRIP δ ¹⁸ O	482	—	—	13.58
Tied to NGRIP δ ¹⁸ O	487	—	—	13.7
Tied to NGRIP δ ¹⁸ O	496.5	—	—	13.98
Katla basalt	500.5	—	—	14.02
Tied to NGRIP δ ¹⁸ O	504.5	—	—	14.12
Tied to NGRIP δ ¹⁸ O	513.5	—	—	14.7
Tied benthic δ ¹⁸ O	542.5	—	—	17.2
¹⁴ C AMS SUERC 14092	565.5	18.666	66	21.292

^aFrom Thornalley *et al.* [2010]. Here ¹⁴C ages are radiocarbon years before A.D. 1950, calendar ages are before A.D. 2000 (in accordance with the NGRIP GICC05 age).

^bNo reservoir applied.

Table 4. Age Model for RAPiD-17-5P^a

	Sediment Depth (cm)	¹⁴ C Age ^b (years B.P.)	±1σ Error (years)	Calendar Age (ka b2k)
¹⁴ C AMS SUERC 14101	750.5	8,823	39	9.590
Saksunarvatn ash	862.5	—	—	10.347
Tied to NGRIP δ ¹⁸ O	919.5	—	—	11.46
Tied to NGRIP δ ¹⁸ O	927.5	—	—	11.703
Vedde ash	954.5	—	—	12.171
Tied to NGRIP δ ¹⁸ O	956.5	—	—	12.76
Tied to NGRIP δ ¹⁸ O	962.5	—	—	13.16
Tied to NGRIP δ ¹⁸ O	976.5	—	—	14.7
Tied benthic δ ¹⁸ O	994.5	—	—	17.2
Tied to NGRIP δ ¹⁸ O	1072.5	—	—	23.38

^aNote that the turbidite between 932.5 and 944.5 cm was omitted from the age model, assuming near instantaneous deposition. Here ¹⁴C ages are radiocarbon years before A.D. 1950, calendar ages are before A.D. 2000 (in accordance with the NGRIP GICC05 age).

^bNo reservoir applied.

consistent with nutrient-depleted GNAIW overlying nutrient-rich GAABW, with a boundary at ~2 km in this region [Oppo and Lehman, 1993; Bertram et al., 1995; Meland et al., 2008]. To maintain this strong δ¹³C gradient, there must have been

either a strong density gradient between the water masses, or fairly rapid renewal of one, or both, water masses. Previous studies suggest active renewal of GNAIW, while GAABW flowed more sluggishly [Boyle, 1995; Manighetti and

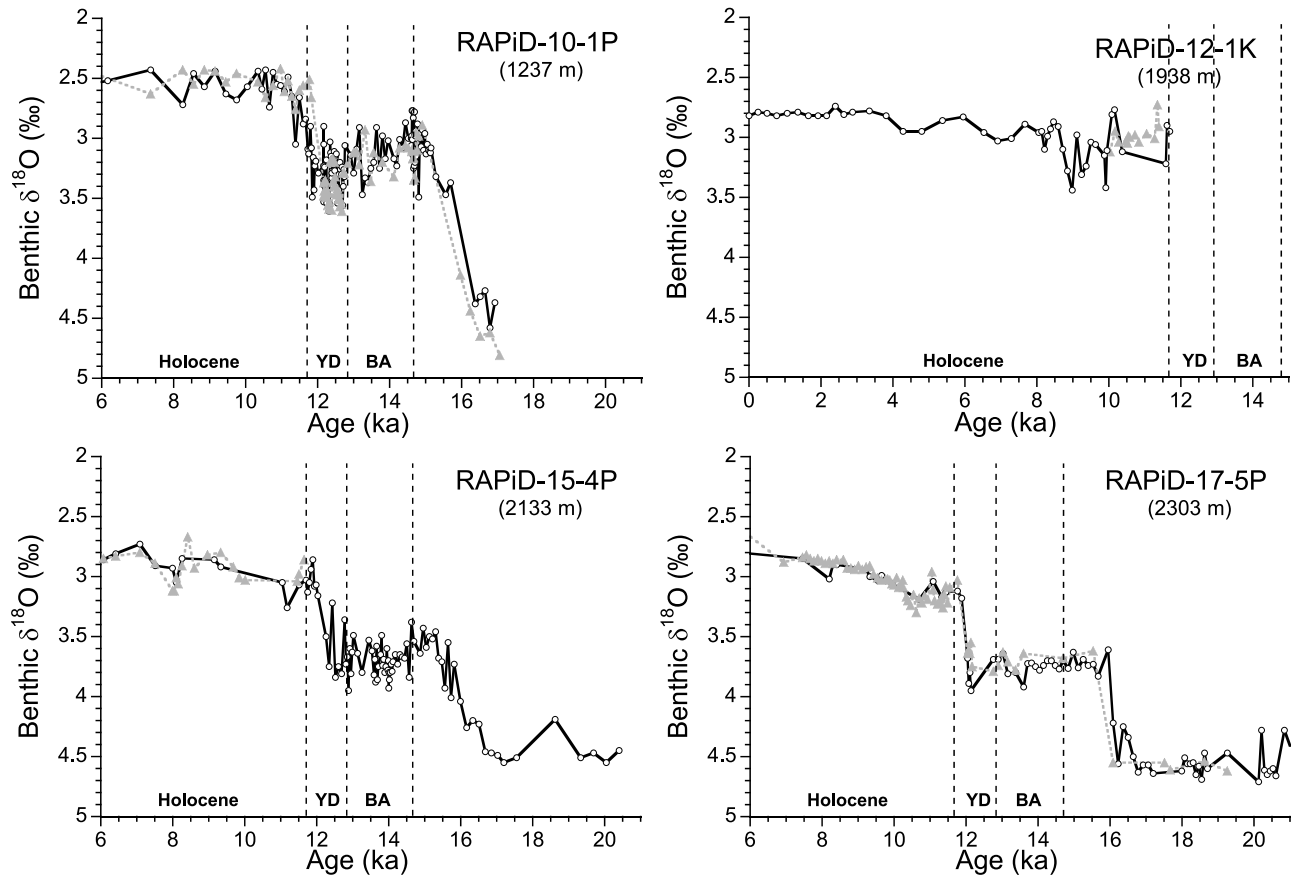


Figure 4. Benthic stable oxygen isotope data (no “ice volume correction”). *C. wuellerstorfi* (RAPiD-12-1K, RAPiD-15-4P, and RAPiD-17-5P) and mixed *Cibicidoides* (RAPiD-10-1P) are represented by black solid lines and open circles. *M. barleanum* (RAPiD-12-1K and RAPiD-10-1P) and *M. pompilioides* (RAPiD-15-4P and RAPiD-17-5P) are represented by gray dashed lines and solid triangles. In core RAPiD-12-1K, between 8 and 10 ka, *C. wuellerstorfi* were sparse and inferred current speeds were fast. Therefore, the high δ¹⁸O values may be the result of reworked foraminifera being included in analyses and biasing the record.

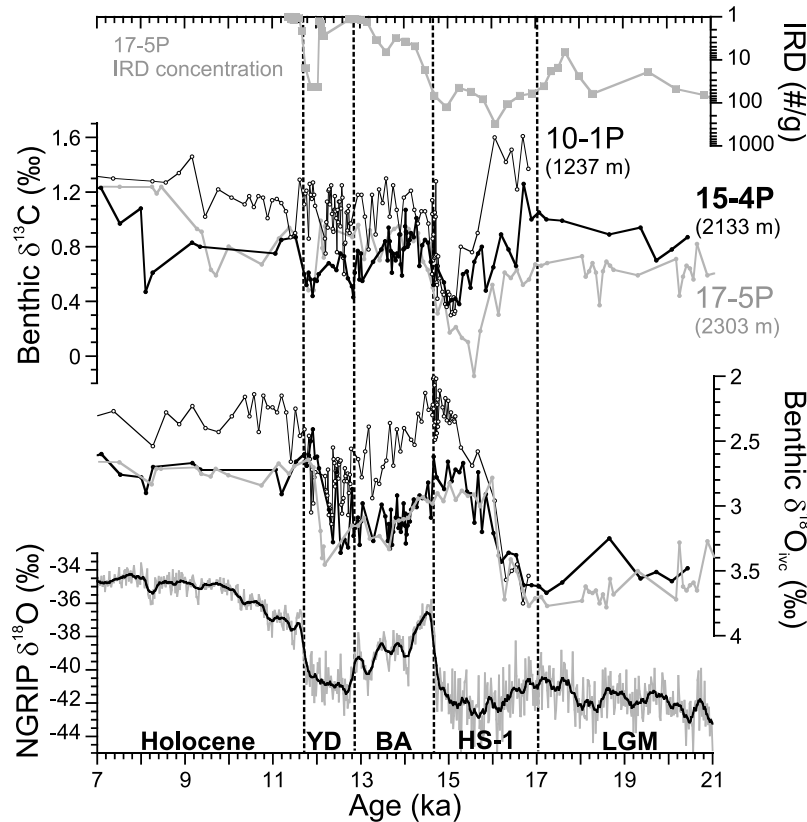


Figure 5. Benthic stable isotope records for the RAPiD cores covering Termination I (note that the RAPiD prefix has not been included on labels). Oxygen isotopes have been ice volume corrected. RAPiD-10-1P (mixed *Cibicidoides* is represented by the thin black line with open black circles), RAPiD-15-4P (*C. wuellerstorfi* is represented by the thick black line with solid circles), and RAPiD-17-5P (*C. wuellerstorfi* is represented by the thick gray line with solid circles) are shown. Also shown at the top are counts of ice-rafted debris grains (150–250 μm fraction), excluding volcanics, per gram of dry sediment in RAPiD-17-5P. To enable zero values to be plotted on a logarithmic scale, a value of 1 has been added to all the counts. NGRIP $\delta^{18}\text{O}$ 20 year values (gray [North Greenland Ice Core Project Members, 2004]); 200 year running mean (black).

McCave, 1995; Sarnthein et al., 2000; Curry and Oppo, 2005; Praetorius et al., 2008].

[17] Similar benthic and *Nps* $\delta^{18}\text{O}_{\text{ice}}$ values observed during the LGM (Figure 7), combined with the evidence for well-ventilated intermediate waters, suggests surface waters south of Iceland were dense enough to convect to intermediate depths during winter cooling and form GNAIW.

5.2. Heinrich Stadial 1 (17–14.7 ka)

[18] A decrease in $\delta^{13}\text{C}$ values occurring at 16.8 ka coincides with the onset of Heinrich Event 1 [Hemming, 2004]. During HS-1, intermediate and deep waters are characterized by low $\delta^{13}\text{C}$ and light $\delta^{18}\text{O}_{\text{ice}}$ values. Minimum $\delta^{13}\text{C}$ values are 0.4‰ for cores 1P and 4P and –0.1‰ for 5P. All cores show decreasing $\delta^{18}\text{O}_{\text{ice}}$ values through HS-1 and minima in $\delta^{18}\text{O}_{\text{ice}}$ and $\delta^{13}\text{C}$ approximately coincide, with the lowest values occurring between ~15 and ~16 ka.

5.3. Bolling Allerød Interstadial (14.7–12.9 ka)

[19] All cores show an abrupt recovery in $\delta^{13}\text{C}$ values to between 0.9 and 1.2‰ at the onset of the BA. There is only a ~0.3‰ gradient between the deep and intermediate sites throughout the BA, suggesting intermediate and deep waters were bathed by similar water masses. During the BA, intermediate water $\delta^{13}\text{C}$ was lower than at the LGM. This difference can be attributed to poorer ventilation, increased air-sea exchange (invasion of ^{12}C into the ocean at deep water formation sites), or lower $\delta^{13}\text{C}$ values of entrained waters in the BA.

[20] $\delta^{18}\text{O}_{\text{ice}}$ values increase through the BA, with deep water sites consistently recording ~0.4‰ heavier $\delta^{18}\text{O}_{\text{ice}}$ values than the intermediate site. Increasing $\delta^{18}\text{O}_{\text{ice}}$ values through the BA indicate decreasing temperature, increasing salinity, or perhaps a decreasing influence of light $\delta^{18}\text{O}$ brine (see section 6).

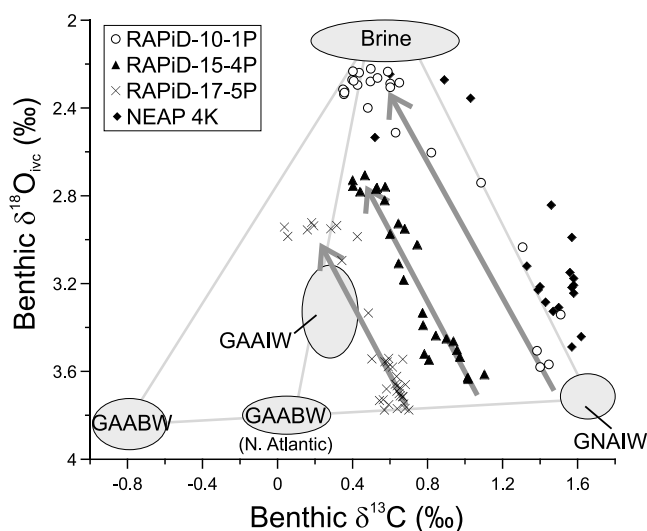


Figure 6. Crossplot of benthic (*Cibicidoides*) carbon and ice-volume-corrected oxygen isotopes between 21 and 14.7 ka, using three-point running mean values. Also shown is unsmoothed data from NEAP 4K [Rickaby and Elderfield, 2005] spanning the same interval. Glacial water mass end-members are indicated by gray regions and are derived from Curry and Oppo [2005], except for the brine end-member, which is estimated using data from this study and a deep Nordic Sea $\delta^{13}\text{C}$ value of 0.8‰ during HS-1 [Veum et al., 1992; Bauch et al., 2001]. (Note the isotopic composition of brines will vary considerably depending upon the isotopic composition of the surface water from which they form.) End-member values for both GAABW from the South Atlantic and North Atlantic are shown. All cores describe a shift from an LGM mixing line between GAABW and GNAIW to a HS-1 mixing line between a southern end-member (GAABW or GAAIW) and brine. The deviation of data from NEAP 4K, and to a lesser extent 1P, away from the brine-GNAIW mixing line may indicate that another water mass is affecting these cores, a nutrient depleted, light $\delta^{18}\text{O}_{\text{ice}}$ water mass ($\delta^{13}\text{C} \sim 1.6\text{‰}$, $\delta^{18}\text{O}_{\text{ice}} \sim 2.6\text{‰}$), possibly entrained surface or mode waters.

5.4. Younger Dryas Stadial (12.9–11.7 ka) and Early Holocene

[21] The high-resolution cores 1P and 4P both indicate an abrupt interval of lowered $\delta^{13}\text{C}$ values (0.3‰ decrease) at the onset of the YD (centered at 12.7 and 12.8 ka, respectively). The decreases in $\delta^{13}\text{C}$ values are not coupled to any clear changes in $\delta^{18}\text{O}_{\text{ice}}$ values. Subsequently, $\delta^{13}\text{C}$ values in 1P increase to $\sim 1.2\text{‰}$ through the YD, suggesting increased intermediate ventilation, and following a maximum at 12.3 ka, $\delta^{18}\text{O}_{\text{ice}}$ values decrease gradually, reaching modern values by 11.2 ka. In contrast, both of the deep cores record a second period of lower $\delta^{13}\text{C}$ values during the late YD. This latter interval of low $\delta^{13}\text{C}$ is associated with a sharp 0.8‰ decrease in $\delta^{18}\text{O}_{\text{ice}}$ values, particularly prominent in 4P beginning at ~ 12.3 ka. The 200 year apparent lag between the decrease in $\delta^{18}\text{O}_{\text{ice}}$ in 5P and 4P may be caused by differential bio-

turbation between the cores, particularly given the presence of a 10 cm turbidite layer in 5P immediately below the decrease in $\delta^{18}\text{O}_{\text{ice}}$, preventing downward mixing of younger material [Manighetti et al., 1995]. These observations suggest that during the late YD different water masses occupied the intermediate and deep water sites, with 4P recording lighter $\delta^{18}\text{O}_{\text{ice}}$ values than the shallower site, 1P. Light $\delta^{18}\text{O}_{\text{ice}}$ values persist in the deep cores into the early Holocene, only increasing to typical deep water Holocene values by ~ 11.5 ka. Deep water $\delta^{13}\text{C}$ values do not stabilize and reach modern values until ~ 8 ka.

6. Discussion

6.1. Brine Influence During Heinrich Stadial 1

[22] The most striking features of the benthic stable isotope records are the low $\delta^{13}\text{C}$ and light $\delta^{18}\text{O}_{\text{ice}}$ values found in both the intermediate and deep cores during HS-1, and in the deep cores, to a lesser extent, during the YD. Traditionally, low $\delta^{13}\text{C}$ values in the North Atlantic have been interpreted as nutrient-rich water of southern origin, i.e., GAABW [e.g., Curry and Oppo, 2005]. During the LGM, benthic stable isotope data from the South Iceland Rise depth transect forms a mixing line between GNAIW and GAABW (Figure 6). (Although it should also be noted that Yu et al. [2008] have recently argued that a low $\delta^{13}\text{C}$ water mass, originating from the Nordic seas, occupied depths between 2 and 2.5 km, sandwiched between GNAIW and GAABW.) Figure 6 illustrates that during HS-1, the benthic stable isotope data do not lie on a mixing line between GNAIW and GAABW. Instead they form a mixing line between a southern sourced end-member (GAABW or Glacial Antarctic Intermediate Water (GAAIW)) and an end-member water mass characterized by relatively low $\delta^{13}\text{C}$ and light $\delta^{18}\text{O}_{\text{ice}}$ values, which is likely brine formed during sea ice formation. Similar trends are observed in benthic records from nearby cores NEAP 4K [Rickaby and Elderfield, 2005] and Ocean Drilling Program (ODP) Site 984 [Praetorius et al., 2008].

[23] Light $\delta^{18}\text{O}$ typical of surface waters influenced by freshwater input can be transferred to the deep ocean during sea ice formation. Freshwater input stratifies the upper ocean and subjects the surface to intense, prolonged cooling and consequent sea ice formation. During sea ice formation, fractionation of oxygen isotopes is insignificant, but salt is rejected from the newly forming sea ice [Dokken and Jansen, 1999]. This process produces dense brines, which sink to the deep ocean, transferring the light $\delta^{18}\text{O}$ surface water signal to the deep ocean. Numerous authors have suggested brine formed in the Nordic seas and overflowed the GSR during Dansgaard-Oeschger stadials and/or Heinrich stadials, and influenced the northern North Atlantic [Jansen and Veum, 1990; Vidal et al., 1998; Dokken and Jansen, 1999; Sarnthein et al., 2000; Raymo et al., 2004; Labeyrie et al., 2005; Meland et al., 2008]. A comparison of benthic oxygen isotope records from the northern North Atlantic and Nordic seas [Meland et al., 2008] records brine overflow during HS-1, affecting the northern North Atlantic above 2200 m. It is likely that this brine overflow contributed toward the light benthic $\delta^{18}\text{O}$ signal observed south of Iceland during

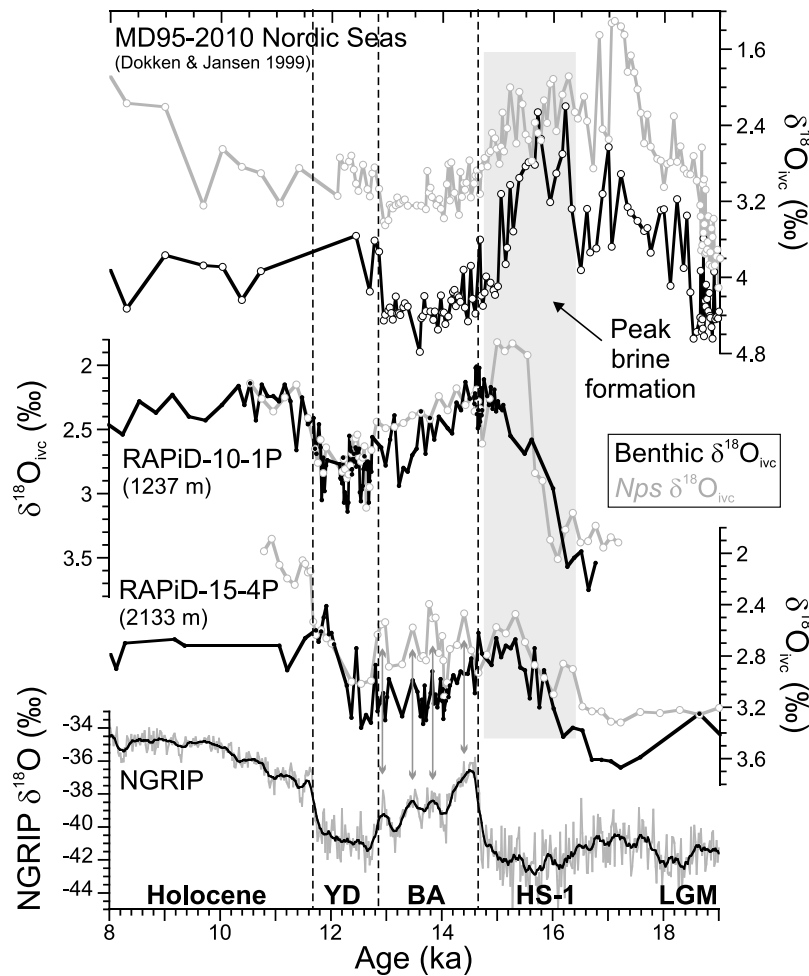


Figure 7. Illustration of the timing of brine formation in the Nordic seas [Dokken and Jansen, 1999] and its export to the South Iceland Rise (RAPiD cores). For all cores, benthic data is shown in black and planktonic (*Nps*) data is shown in gray. Peak brine formation in the Nordic seas (gray-shaded region) coincides with the sharp decrease in benthic $\delta^{18}\text{O}_{\text{ice}}$ south of Iceland, indicating export of the brine out of the Nordic seas. Minima in *Nps* $\delta^{18}\text{O}_{\text{ice}}$ within the high sedimentation rate core RAPiD-15-4P correlate with minima in the percent abundance of *Nps* and warm intervals indicated by NGRIP are highlighted by the gray arrows. NGRIP $\delta^{18}\text{O}$ 20 year values (gray [North Greenland Ice Core Project Members, 2004]); 200 year running mean (black).

HS-1. Figure 7 shows peak brine formation in the Nordic seas coincides with the sharp decrease in benthic $\delta^{18}\text{O}_{\text{ice}}$ south of Iceland.

[24] Increased ice-rafted debris (IRD) abundance (lacking characteristic detrital carbonate grains associated with the Laurentide ice sheet) is observed during HS-1 within cores from the Nordic seas [e.g., Haflidason et al., 1995; Dokken and Jansen, 1999] and south of Iceland (Figure 5). This suggests that the increased freshwater input responsible for enhanced sea ice formation and brine rejection was primarily sourced from the surrounding Fennoscandinavian and Icelandic ice sheets, although the advection of freshwater from the Labrador Sea to the Nordic seas during Laurentide ice sheet surging may have also contributed toward surface freshening.

[25] Peak brine formation in the Nordic seas coincides with a large decrease in RAPiD-10-1P *Nps* $\delta^{18}\text{O}_{\text{ice}}$ values, starting at 16 ka and persisting through to 15 ka (Figure 7). A smaller magnitude but similar trend is observed in RAPiD-15-4P, located ~100 km further south from the Icelandic shelf. Mg/Ca data for the *Nps* records rule out that the decrease in $\delta^{18}\text{O}_{\text{ice}}$ is caused by warming (D. J. R. Thornalley et al., Reconstructing North Atlantic deglacial surface hydrography and its link to the Atlantic overturning circulation, submitted to *Global and Planetary Changes*, 2010) and therefore strong surface freshening, concentrated closer to Iceland, is implied. However, on the basis of *Nps*'s intolerance to low salinities, Hillaire-Marcel and de Vernal [2008] have argued that very light *Nps* $\delta^{18}\text{O}$ values indicate the injection of locally formed brines below the halocline.

Brines were, therefore, probably also forming locally on the Icelandic shelf or open ocean (similar to those reported by *Hillaire-Marcel and de Vernal* [2008] in the Labrador Sea), in addition to those forming in the Nordic seas and overflowing the GSR.

6.2. Brine Formation, Deep Water Warming, or Freshening?

[26] It has been argued that large decreases in benthic $\delta^{18}\text{O}_{\text{ivc}}$ values (equivalent to 5–11°C warming) from the intermediate depth Nordic seas during HS-1 cannot be caused by warming or freshening since water masses with such high temperatures or low salinities would not be dense enough to sink to below 1 km depth and replace water previously formed during earlier convection events [*Dokken and Jansen*, 1999]. Similarly, it can be argued that the decrease in benthic $\delta^{18}\text{O}_{\text{ivc}}$ values observed south of Iceland during HS-1 (0.9–1.3‰) cannot be caused by freshening. Yet, the smaller decrease in benthic $\delta^{18}\text{O}_{\text{ivc}}$ values observed south of Iceland during HS-1 (equivalent to increasing water temperature from –1°C to ~3–5°C) can feasibly be attributed to warming because there is minimal change in the density of seawater over this temperature range. Moreover, based on benthic faunal assemblages, light benthic $\delta^{18}\text{O}_{\text{ivc}}$ values recorded at intermediate depths on the north side of the Iceland-Scotland ridge, during HS-1, have been interpreted as an inflow of Atlantic water and warming by ~4°C [*Rasmussen et al.*, 1996; *Bauch and Bauch*, 2001]. However, in this study a potential source for the proposed warming south of Iceland, that is consistent with other evidence, cannot be found. (1) The expansion of sea ice and a decrease in air temperatures in the northern North Atlantic region during HS-1 [e.g., *Rasmussen and Thomsen*, 2008] excludes the possibility that warmer deep and intermediate waters were derived from locally convected waters. (2) An incursion of warm Atlantic water, presumable sourced from warmer, low latitude, shallow or thermocline waters, would likely be nutrient depleted and therefore associated with high $\delta^{13}\text{C}$, yet the opposite trend is observed, with benthic $\delta^{13}\text{C}$ shifting to lower values. (3) Temperatures of between 0 and –1°C are recorded off the Iberian Margin at 3.1 km water depth [*Skinner et al.*, 2003] during HS-1, implying deep southern sourced water did not warm considerably during this interval. (4) Mg/Ca-based temperature reconstructions conducted on the thermocline dwelling *Nps* within the south Iceland RAPiD cores reveal no warming of a comparable magnitude [*Thornalley*, 2008]. Therefore, the light benthic $\delta^{18}\text{O}_{\text{ivc}}$ values recorded south of Iceland during HS-1 are primarily attributed to the influence of brines that have formed in the Nordic seas and overflowed the GSR, and the addition of brines formed locally.

6.3. Carbon Isotope Composition of Brine

[27] The $\delta^{13}\text{C}$ signature of brine is not well known, although it is expected to have a low $\delta^{13}\text{C}$ signature during HS-1 for the following reasons:

[28] 1. Sea ice formation and brine rejection is thought to have occurred on the shallow shelf areas [*Dokken and Jansen*, 1999; *Meland et al.*, 2008]. Modern studies of planktonic foraminifera in Arctic shelf regions relate low $\delta^{13}\text{C}$ values

to the remineralization of organic matter and the input of low $\delta^{13}\text{C}_{\text{DIC}}$ river water [*Spielhagen and Erlenkeuser*, 1994; *Volkmann and Mensch*, 2001].

[29] 2. Comparatively low $\delta^{13}\text{C}$ values in the Nordic seas during stadials may have been caused by enhanced sea ice cover of shelf areas [*Spielhagen and Erlenkeuser*, 1994]. Isotopic equilibrium between the atmosphere and surface ocean at low temperatures causes high $\delta^{13}\text{C}$ values of surface water [*Lynch-Stieglitz et al.*, 1995]. Enhanced sea ice cover of the shelves limits air-sea exchange of CO_2 and could have allowed metabolic CO_2 from bacterial respiration of organic matter (with low $\delta^{13}\text{C}$ values) to accumulate in the water [*Spielhagen and Erlenkeuser* 1994].

[30] 3. *Lekens et al.* [2006] suggest low glacial $\delta^{13}\text{C}$ values are partly caused by meltwater, which reduces ventilation by the development of a freshwater lid [*Spielhagen et al.*, 2004]. Meltwater from glaciers also has a $\delta^{13}\text{C}_{\text{DIC}}$ signature similar to the atmosphere (–7 to –8‰) and may directly contribute to low surface water $\delta^{13}\text{C}$ values [*Veum et al.*, 1992; *Bauch and Weinelt*, 1997].

[31] 4. Methane releases from the East Greenland Shelf during stadials [*Millo et al.*, 2005] and Termination I [*Smith et al.*, 2001] and probably also the Barents Sea [*Damm et al.*, 2007; *Wollenburg and Mackensen*, 2009] may have contributed to the low $\delta^{13}\text{C}$ values in the Nordic seas during HS-1.

[32] Reliable epibenthic $\delta^{13}\text{C}$ data are very limited in the Nordic seas. Existing records indicate a lowering of benthic $\delta^{13}\text{C}$ in the deep Nordic seas (2,700 m) from ~1.2–1.3‰ during the LGM to 0.8‰ during HS-1 [*Veum et al.*, 1992; *Bauch et al.*, 2001], while intermediate waters (~1 km) in the Icelandic Sea record values as low as 0.4‰ [*Hagen*, 1999], but these may be influenced by methane release [*Millo et al.*, 2005]. In Figure 6, the mixing line described by HS-1 data from 1P, 4P and 5P constrains the $\delta^{13}\text{C}$ of brine to between 0.4 and 1.0‰, likely ~0.6–0.8‰. This is relatively low compared to typical $\delta^{13}\text{C}$ values of northern sourced waters formed by open ocean convection (GNAIW, ~1.5‰; NADW, ~1.2‰). Therefore the benthic $\delta^{13}\text{C}$ decrease recorded in 1P, 4P and 5P during HS-1 does not necessarily reflect an increased influence of southern sourced water, but rather a change in the water mass with which it is mixing, i.e., GNAIW or brine.

[33] The presence in the NE Atlantic of a northern sourced end-member that has relatively low $\delta^{13}\text{C}$ complicates the traditional interpretation that low $\delta^{13}\text{C}$ values imply increased incursion of nutrient enriched southern source water. Instead, a decrease in $\delta^{13}\text{C}$ (associated with light $\delta^{18}\text{O}$) may be recording the switch from southern source water mixing with high $\delta^{13}\text{C}$ GNAIW, to mixing with brine with a $\delta^{13}\text{C}$ of 0.6–0.8‰. On the basis of benthic $\delta^{13}\text{C}$ and Cd/Ca data, low $\delta^{13}\text{C}$ values recorded during HS-1 in NEAP 4K were originally interpreted as an incursion of GAAIW [*Rickaby and Elderfield*, 2005]. Yet the light $\delta^{18}\text{O}_{\text{ivc}}$ values that also occur during this interval (Figure 6) cannot be attributed to GAAIW and are unlikely to be caused by intermediate water warming (as previously discussed). Therefore, given previously reported evidence for brine overflows, it seems probable that NEAP 4K was influenced by a mixture of brine and southern source water (GAAIW, according to *Rickaby*

and Elderfield [2005]) during HS-1. To confirm this interpretation, typical Cd/Ca values for the brine overflows need to be constrained.

6.4. Paleocceanography During the Bølling Allerød Interstadial (14.7–12.9 ka)

[34] The abrupt increase of intermediate and deep water $\delta^{13}\text{C}$ values at the onset of the BA suggests the replacement of brine water south of Iceland, with a water mass formed by open ocean convection, exhibiting similar benthic $\delta^{18}\text{O}_{\text{ivc}}$ and $\delta^{13}\text{C}$ values to modern ISOW. Benthic $\delta^{13}\text{C}$ values ($\sim 1.1\%$ at 1237 m; $\sim 0.8\%$ at 2133 and 2303 m) were, however, slightly lower than typical late Holocene ISOW values (1.2–1.3%). This suggests one or all of the following: ISOW formation was less vigorous than today and there was a significant proportion of southern sourced water, or remnant brine entrained into water below ~ 2000 m in the northern North Atlantic; the preformed value of ISOW was lower during the BA; the whole ocean carbon inventory was still 0.2% lower than at present.

[35] The scenario most consistent with existing data for the onset of the BA is that deep and intermediate waters remained cold (-1 – 0°C) throughout HS-1, consistent with the proposed LGM temperatures south of Iceland and those reconstructed on the Iberian Margin at 3.1 km depth [Skinner *et al.*, 2003], while benthic $\delta^{18}\text{O}_{\text{ivc}}$ values decreased throughout HS-1 due to brine influence. Benthic $\delta^{18}\text{O}_{\text{ivc}}$ values remained low at the onset of the BA because deep and intermediate water temperatures increased abruptly by $\sim 3^\circ\text{C}$ [Skinner *et al.*, 2003], counterbalancing a reduction in brine formation and overflow. Benthic $\delta^{18}\text{O}_{\text{ivc}}$ then increased gradually throughout the Bølling caused by either the continued decline in the formation and entrainment of brines, or cooling of intermediate and deep waters into the Allerød.

6.5. Deep and Intermediate Water Changes During the YD

[36] The abrupt decrease in benthic $\delta^{13}\text{C}$ recorded at the onset of the YD in the intermediate depth core 1P and the deep core 4P suggests a brief reduction in ventilation south of Iceland and a likely incursion of southern sourced water between 12.9 and 12.6 ka. The absence of light benthic $\delta^{18}\text{O}_{\text{ivc}}$ or low planktic $\delta^{13}\text{C}$ values at this time argue against any influence of brine, or a significant change in the preformed signal, respectively. Pa/Th data reveals a reduction in the strength of the AMOC during the YD [McManus *et al.*, 2004], which may have resulted in an increased influence of nutrient depleted southern sourced water [e.g., Rickaby and Elderfield, 2005]. It has been postulated that this reduction in AMOC was triggered by either rerouting of freshwater from the Laurentide ice sheet [Broecker and Denton, 1989] or a discharge of freshwater to the Arctic Ocean [Spielhagen *et al.*, 2005; Tarasov and Peltier, 2005], although as yet, compelling paleoceanographic data to support these hypotheses has not been found. The rapid recovery of benthic $\delta^{13}\text{C}$ values at ~ 12.6 ka suggests that the decrease in ventilation was transient, with intermediate waters indicating stronger ventilation throughout the remainder of the YD.

[37] During the late YD, the two deep cores both show light benthic $\delta^{18}\text{O}_{\text{ivc}}$ and low $\delta^{13}\text{C}$ values, which based on earlier evidence presented for HS-1, are characteristic of brine. Meland *et al.* [2008] reconstruct the formation and sinking of brines to intermediate depths within the Nordic seas during the mid to late YD. It is likely the overflow of these brines that is being detected south of Iceland in the deep cores 4P and 5P. The lack of any strong depletions in benthic $\delta^{18}\text{O}_{\text{ivc}}$ in 1P during the YD suggests that intermediate water south of Iceland was not significantly affected by brines. This disagrees with the conclusions of Meland *et al.* [2008] who infer the overflows above 1500 m consisted of a mixture of brine and water formed by open ocean convection.

[38] Bakke *et al.* [2009] provide evidence for a flickering climate in the Nordic seas during the late YD, in which incursions of warmer Atlantic waters led to episodic ice sheet melting and enhanced sea ice growth. Data from south of Iceland support this hypothesis: planktonic (*Globorotalia inflata* and *Globigerina bulloides*) Mg/Ca data from 4P indicate warmer Atlantic water incursions from the mid-YD onward [Thornalley *et al.*, 2010]; the increased abundance of IRD (not containing detrital carbonate) within 4P [Thornalley *et al.*, 2010] and 5P (Figure 5) during the late YD suggests local ice sheet instability. The late YD export of brines from the Nordic seas may therefore have been a result of repeated intervals of sea ice growth and hence increased brine formation during this flickering climate, possibly assisted by episodic open ocean convection, occurring during incursions of warmer water, flushing out previously formed brines.

[39] The late YD is also associated with Laurentide ice sheet surging (Heinrich Event 0 (HE-0)) and likely brine formation in the Labrador Sea [Hillaire-Marcel and de Vernal, 2008]. HE-0 may possibly have also provided freshwater to south of Iceland and the Nordic seas; but given the absence of a strong brine signal in the intermediate depth core 1P, it seems unlikely that substantial local brine formation occurred south of Iceland during the YD, hence the major sites of brine formation were probably centered on the Nordic and Labrador seas.

6.6. Strengthening of ISOW During the Early Holocene

[40] Benthic $\delta^{18}\text{O}_{\text{ivc}}$ values suggest that water at depths between 2100 m and 2300 m contained a small contribution of brine between 11.7 and 11.5 ka, in agreement with Meland *et al.* [2008]. Intermediate and deep water $\delta^{13}\text{C}$ only reached modern values by ~ 7 – 8 ka (Figure 8). The simultaneous change in the benthic and planktic $\delta^{13}\text{C}$ records during the early Holocene suggests that a significant component of the benthic shift may be caused by changes in the preformed $\delta^{13}\text{C}$ value of ISOW or the entrainment of low $\delta^{13}\text{C}$ surface and thermocline waters into the overflow. Alternatively, low benthic $\delta^{13}\text{C}$ values may indicate an increased influence of southern sourced waters during weaker/shallower ISOW flow. A switch from weak/shallow ISOW on the South Iceland Rise to a strong flow at modern depths at ~ 7 – 8 ka is supported by large and abrupt increases in the wt. % sand at ~ 8 ka in core RAPiD-12-1K (1938 m); and at ~ 7 – 7.5 ka in the deeper core RAPiD-15-4P (2133 m), indicating winnowing of fine material by strong bottom water currents. Increased

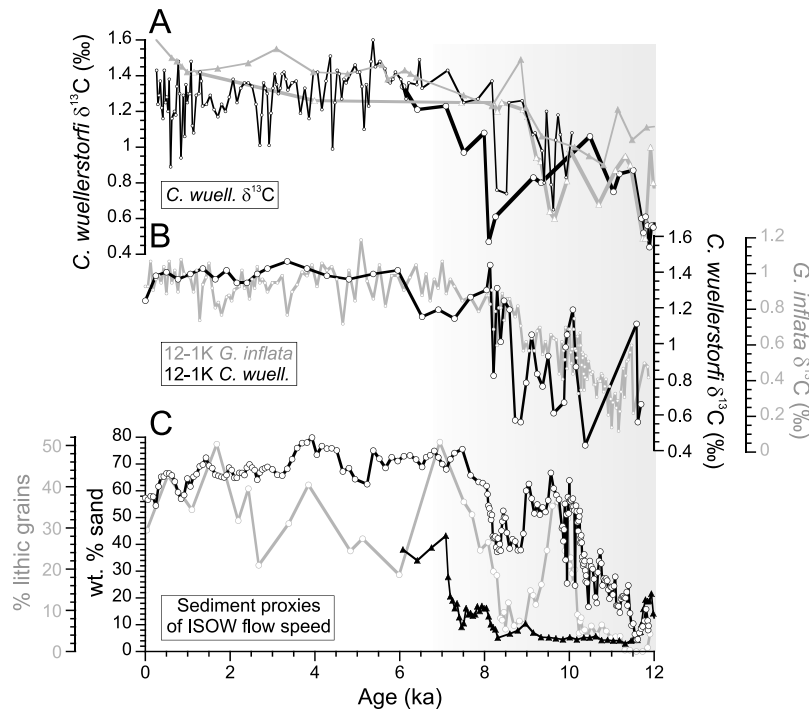


Figure 8. Proxies illustrating the Holocene establishment of strong ISOW flow. (a) *C. wuellerstorfi* $\delta^{13}C$ from RAPiD-15-4P (thick black line with open black circles), RAPiD-17-5P (thick gray line with gray open triangles), NEAP 4K [Hall *et al.* [2004] shown by thin black line with black open squares and Rickaby and Elderfield [2005] shown by thin gray line with gray solid triangles). (b) Comparison between benthic and planktic carbon isotopes from RAPiD-12-1K. (c) Weight percent sand fraction for RAPiD-12-1K (1938 m water depth shown by black line with open circles) and percent abundance of lithic grains (150–250 μm) within RAPiD-12-1K (gray line with open circles), which indicate the establishment of strong bottom current activity over these sites by ~ 7 –8 ka. (The upper 1.4 m of RAPiD-15-4P, spanning the last ~ 6 ka, has not been sampled but consists of foraminiferal sand similar to that sampled between 6 and 7 ka.)

abundance of reworked lithic grains ($>150 \mu m$ diameter) in RAPiD-12-1K at ~ 8 ka, and throughout the remainder of the Holocene, confirms that the increase in wt. % sand is caused by increased flow speed at the core site.

[41] Flow speed records on a slope close to the source are problematic because of movement up or down the slope of the main flow axis and the very local supply of material. Sortable silt mean grain size data from the nearby cores ODP 984 [Praetorius *et al.*, 2008] and NEAP 4K [Hall *et al.*, 2004] (note this record only extends back to 10 ka) do not suggest increasing ISOW strength during the early Holocene; although the increase in percent sortable silt in ODP 984, up to a maximum at 7.5 ka, that occurs in unison with increasing wt. % sand in 1K and 4P, is consistent with increasing ISOW strength over Bjorn drift. In a hydrodynamically sorted deposit, there should be a well-defined relationship between sortable silt mean grain size and the percent sortable silt fraction [McCave and Hall, 2006]. The lack of such a relationship during the early Holocene in ODP 984 suggests the fidelity of the sortable silt mean grain size method on Bjorn drift during this interval may be compromised. This is likely due to the proximity to Iceland

and large changes in sediment source and routing associated with the deglaciation of Iceland and the input and reworking of tephra from the Saksunarvatn eruption (10.38 ka).

[42] Evidence presented here, suggesting that ISOW did not reach its modern flow strength and/or depth over the South Iceland Rise until ~ 7 –8 ka, is consistent with records of NADW intensity based on Pa/Th data [McManus *et al.*, 2004; Gherardi *et al.*, 2009] and water mass proportions based on sediment ϵNd [Piotrowski *et al.*, 2004], and benthic foraminiferal Cd/Ca ratios [Marchitto *et al.*, 1998]. Strengthening/deepening of ISOW also coincides with major surface water changes in response to the cessation of deglacial meltwater input at ~ 7 –8 ka, including the onset of deep convection in the Labrador Sea [Hillaire-Marcel *et al.*, 2001], a reorganization of the subpolar and subtropical gyres [Thornalley *et al.*, 2009], and a reduction in the albedo cooling effect of the Laurentide ice sheet [Renssen *et al.*, 2009]. Furthermore, Bianchi and McCave [1999] inferred a reorganization of the deep circulation around 7.5 ka, which they attributed to the end of significant glacial meltwater input. Deepening of ISOW flow throughout the early Holocene was likely a result of a reduction in deglacial meltwater

flux, such that ISOW became denser and the main flow axis migrated downslope. Finally, it is noted that the lag of ~4 ka between the onset of the Holocene and modern strength/depth ISOW flow is remarkably similar to recent findings from Gardar drift during the initiation of the last interglacial (Marine Isotope Stage 5e, ~128 ka), which demonstrate a 3.5 ka lag between initial climate warming and strong ISOW flow as detected by sortable silt mean grain size and benthic $\delta^{13}\text{C}$ data [Hodell *et al.*, 2009].

7. Conclusions

[43] Benthic stable isotope records from a depth transect on the South Iceland Rise have been used to reconstruct the changes in deep and intermediate water paleoceanography in the northern North Atlantic during the deglaciation. The main conclusions of this study are as follows:

[44] 1. During the LGM, benthic $\delta^{18}\text{O}$ values were constant between 1.2 km and 2.3 km and suggest the water column between these depths cooled to -0.5 to -1.0°C . In contrast, there is a strong $\delta^{13}\text{C}$ gradient between 1.2 km and 2.3 km, supporting previous work that has identified well ventilated GNAIW in the northern North Atlantic above ~2 km water depth.

[45] 2. Benthic stable isotope records reveal that brines were formed both locally and exported from the Nordic seas to the northern North Atlantic during HS-1 and the late YD, producing a mixing line between a brine end-member and southern sourced water.

[46] 3. The relatively low $\delta^{13}\text{C}$ values of brines complicates the traditional interpretation that decreases in $\delta^{13}\text{C}$ indicate an increased incursion southern sourced water. Instead, the $\delta^{13}\text{C}$ decrease may indicate a shift from southern sourced water mixing with GNAIW to mixing with brine; the presence of brine can potentially be determined by examining benthic $\delta^{18}\text{O}$ values. Additionally, the similarity of broad trends observed between planktonic and benthic $\delta^{13}\text{C}$ records during the Holocene suggests that Holocene variations in North Atlantic deep water $\delta^{13}\text{C}$ are significantly affected by their preformed $\delta^{13}\text{C}$ signature, or the $\delta^{13}\text{C}$ composition of entrained water during overflow of the GSR, and therefore do not necessarily imply an increased influence of southern source water.

[47] 4. The onset of the YD is accompanied by a brief (~200–300 year) decrease in intermediate and deep water $\delta^{13}\text{C}$ values (with no simultaneous change in benthic $\delta^{18}\text{O}$, indicative of brine). This suggests a decrease in ventilation in response to a reduction in AMOC strength, possibly triggered by a freshwater discharge event.

[48] 5. ISOW flow only reached its modern strength and/or depth over the South Iceland Rise by ~7–8 ka, consistent with existing records of NADW intensity, and coinciding with the cessation of deglacial meltwater input to the North Atlantic and major surface ocean reorganizations.

8. Implications and Future Work

[49] Confirming the presence of brine in the North Atlantic through the use of benthic foraminifera stable

isotopes requires calcification temperatures to be constrained. Yet benthic Mg/Ca paleothermometry at low temperatures (less than 5°C) conducted on commonly used *Cibicides* species has recently been shown to be insensitive to temperature, with the Mg/Ca ratio being more strongly controlled by carbonate ion saturation state [Yu and Elderfield, 2008]. Buffering of pore water carbonate ion concentrations may allow infaunal species to better record calcification temperatures [e.g., Skinner *et al.*, 2003], although further investigation is required. Epifaunal and infaunal benthic foraminifera from the deep South Iceland Rise RAPiD cores were analyzed for trace metal concentration but contamination of the deglacial samples by manganese carbonate overgrowths prevented any paleoceanographic interpretation.

[50] Recently, a distinct third water mass (in addition to GNAIW and GAABW), likely brine overflow from the Nordic seas, has been identified in the NE Atlantic between 2 and 2.5 km water during the LGM, through the use of paired benthic foraminiferal $\delta^{13}\text{C}$ and B/Ca analysis [Yu *et al.*, 2008]. In our study no brine overflow was detected during the LGM through stable isotope analysis alone. This raises the possibility that brines may only be detected by stable isotope analysis when particularly light $\delta^{18}\text{O}$ fresh water is released into the surface ocean, the light $\delta^{18}\text{O}$ acting as a water mass tracer. However, the most extensive periods of sea ice and subsequent brine formation are likely to be associated with periods of increased input of glacial meltwater, which does contain a distinctive light $\delta^{18}\text{O}$ signature. In the future, combined benthic foraminiferal B/Ca and stable isotope analysis ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) may be used to investigate further the formation of brine and its export to the North Atlantic, with the added advantage that B/Ca ratios are less susceptible to contamination from phases such as manganese carbonate.

[51] Presently, there is considerable uncertainty regarding the spatial extent of brine overflow during HS-1 and the YD. Consistent with earlier studies, it seems highly likely that brine overflows entrained water throughout much of the NE Atlantic during periods of enhanced freshwater input and reduced open ocean convection in the region. Further to this though, a global compilation of cores by Labeyrie *et al.* [2005] has suggested that the input of deglacial light $\delta^{18}\text{O}$ (along with the associated low $\delta^{13}\text{C}$ signature) into the ocean via brine formation in the North Atlantic can be traced throughout all major basins of the world's ocean during HS-1 and the YD, with faster transmission at intermediate depths (~few hundred years) compared to deep water (~1500 years). If this observation is correct, brine formation and export from the Nordic seas is an important pathway for transferring ^{16}O stored in glacial ice sheets to the deep ocean. However, records of sea level rise [e.g., Fairbanks, 1989] suggest that less than half the input of meltwater to the ocean occurred during the identified periods of brine formation in the North Atlantic, implying there are other, likely more significant, pathways for transferring glacial meltwater to the deep ocean (such as diffusive mixing or entrainment at sites where open ocean convection can occur).

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