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Progressive reduction in NE Atlantic intermediate water ventilation prior to Heinrich events: Response to NW European ice sheet instabilities?

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[1] We present high-resolution benthic $\delta^{13}C$ records from intermediate water depth core site MD01-2461 (1153 m water depth), from the Porcupine Seabight, NE Atlantic, spanning 43 to 8 kyr B.P. At an average proxy time step of 160 ± 56 years this core provides information on the linkage between records from the Portuguese Margin and high-latitude North Atlantic basin, allowing additional insights into North Atlantic thermohaline circulation (THC) variability during millennial-scale climatic events of the last glacial. Together, these records document both discrete and progressive reductions in Glacial North Atlantic Intermediate Water (GNAIW) formation preceding Heinrich (H) events 1, 2, and 4, recorded through the apparent interchange of glacial northern and southern-sourced intermediate water signatures along the European Margin. Close coupling of NW European ice sheet (NWEIS) instability and GNAIW formation is observed through transient advances of SCW along the European margin concurrent with pulses of ice-rafted debris and meltwater release into the NE Atlantic between 27 and 16 kyr B.P., when the NWEIS was at maximum extent and proximal to Last Glacial Maximum convection zones in the open North Atlantic. It is such NWEIS instability and meltwater forcing that may have triggered reduced North Atlantic THC prior to collapse of the Laurentide ice sheet at H1 and H2. Precursory reduction in GNAIW formation prior to H4 may also be inferred. However, limited NWEIS ice volume prior to H4 and convection occurring in the Norwegian-Greenland Sea require that if a meltwater trigger is invoked, as appears to be the case at H1 and H2, the source of meltwater prior to H4 is elsewhere, likely higher-latitude ice sheets. Clarification of the sequencing and likely mechanisms of precursory decrease of the North Atlantic THC support theories of H event initiation relating to ice shelf growth during cold periods associated with reduced North Atlantic THC and subsequent ablation through subsurface warming and sea level rise associated with further reductions in meridional overturning.

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Theme: Past Ocean Circulation
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1. Introduction

[2] An increasing number of theories concerning the initiation of periodic collapse of the Laurentide ice sheet (LIS), so-called Heinrich (H) events call for improved understanding of phase relationships within the North Atlantic ice-ocean-climate system. A key unknown and critical component to several H event initiation theories [e.g., *Moros et al.*, 2002; *Flückiger et al.*, 2006] is the phasing of Atlantic Meridional Overturning Circulation (AMOC) perturbation with respect to LIS collapse.

[3] Advance of Southern Component Water (SCW) into the North Atlantic during the last glacial is apparent from extensive mapping of benthic foraminiferal carbon isotopes from North Atlantic core sites, with high nutrient, low $\delta^{13}C$ Antarctic Bottom Water (AABW) bathing the North Atlantic basin at depths >2000 m as far north as the Rockall Trough [e.g., *Oppo and Lehman*, 1993; *Sarnthein et al.*, 1994; *Curry and Oppo*, 2005]. The glacial mode of North Atlantic thermohaline circulation (THC) was highly sensitive to density contrasts between Glacial North Atlantic Intermediate Water (GNAIW) and SCW, most likely Antarctic Intermediate Water (AAIW) [e.g., *Broecker et al.*, 1990; *Weaver et al.*, 2003; *Rickaby and Elderfield*, 2005]. Significantly reduced ventilation of waters bathing intermediate depth North Atlantic sites during episodes of northern hemisphere ice sheet instability, particularly the H events, suggests reduced GNAIW formation, likely due to increased freshwater flux obstructing North Atlantic deep/intermediate water formation [e.g., *Cortijo et al.*, 1997; *Vidal et al.*, 1997; *Ganopolski and Rahmstorf*, 2001], and invasion of SCW to high northern latitudes (at depths <2000 m) [*Sarnthein et al.*, 1994; *Zahn et al.*, 1997; *Willamowski and Zahn*, 2000; *Elliot et al.*, 2002; *Rickaby and Elderfield*, 2005]. However, this apparently straightforward cause-effect relationship between H event meltwater and intermediate water formation is questioned, with North Atlantic benthic $\delta^{13}C$ records suggesting that weakening of the North Atlantic THC may have begun up to 2.5 kyr prior to LIS instability [*Zahn et al.*, 1997; *Willamowski and Zahn*, 2000]. Furthermore, recent investigation into AMOC variability using $^{231}Pa_{xs}/^{230}Th_{xs}$ suggest a substantial decrease in the rate of overturning prior to deposition of LIS-sourced ice-rafted debris (IRD) [*McManus et al.*, 2004; *Hall et al.*, 2006], possibly driven by NW European ice sheet (NWEIS) sourced meltwater [*Hall et al.*, 2006]. *Flückiger et al.* [2006] suggest that subsurface warming and relative sea level rise likely resulted from reduced AMOC prior to the H events, ablating and destabilizing northern hemisphere ice sheet margins and initiating ice sheet collapse. Conversely, *Moros et al.* [2002] suggest that enhanced subsurface heat flux driven by a strengthened AMOC may have melted ice shelves and forced the retreat of grounding lines. These divergent theories require further investigation into the records of AMOC variability and phasing with ice sheet-surface ocean conditions.

[4] We present benthic $\delta^{13}C$ records from sediment core MD01-2461 from the Porcupine Seabight, an intermediary for pre-existing NE Atlantic records, concentrated on the Iberian margin [*Zahn et al.*, 1997] and Rockall Trough-Irminger Sea [*van Kreveld et al.*, 2000; *Elliot et al.*, 2002; *Knutz et*

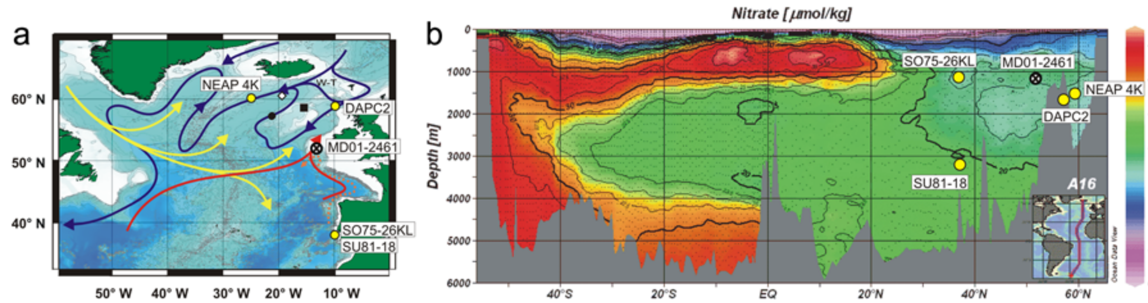

Figure 1. Positioning of core sites discussed in this study and modern circulation scheme. (a) Locations of cores discussed in this study and the modern North Atlantic circulation scheme. Blue arrows illustrate the pathways of the NADW across the Wyville-Thomson Ridge (W-T) and recirculation in the eastern North Atlantic basin at ~2000–3000 m water depth. Yellow arrows represent LSW at 1500–2000 m. The dashed orange shows the path of MOW at ~1000 m depth, and red arrows show surface North Atlantic Current and ENAW (to depths of ~750 m). GEOSECS (http://ingrid.ldeo.columbia.edu/SOURCES/.GEOSECS/) Station 23 (black, open diamond); V28-73 (black, filled circle) and VM29-198 (black, filled square) both from *Oppo and Lehman* [1993]. Figure adapted from *Frank et al.* [2004]. (b) eWOCE (http://www.ewoce.org/) meridional section A16 showing modern-day nitrate concentrations, with nutrient-rich AAIW clearly expressed at intermediate depth reaching as far north as ~20°N. Nutrient-depleted NADW fills the basin and flows south at depth. Cores discussed in this study are plotted on this meridional section to show their relation to modern-day hydrography.

al., 2002; *Hall et al.*, 2006; *Rickaby and Elderfield*, 2005]. Combined with previously published records of ice sheet instability and surface ocean conditions from core MD01-2461 [*Peck et al.*, 2006], we investigate at multicentennial to millennial time steps the distribution of intermediate waters along the European Margin to determine sequencing of North Atlantic ice sheet vs. THC events.

2. Materials and Methods

[5] Sediment core MD01-2461 was recovered from a seismically laminated sequence on the NW flank of the Porcupine Seabight (51°45′N, 12°55′W) at a water depth of 1153 m (Figure 1b). The Porcupine Seabight represents a N-S trending embayment in the continental shelf SW of Ireland, gradually deepening to the south, to ~3000 m water depth and opening onto the Porcupine Abyssal Plain. Thermohaline convection in the Labrador Sea and Norwegian-Greenland Sea drives present-day circulation, through the formation of intermediate level Labrador Seawater (LSW) and deeper North Atlantic Deep water (NADW). The latter water mass comprises three components that enter the North Atlantic Basin through the Wyville-Thomson, Iceland-Faeroe and Greenland-Iceland ridges [*McCartney and Talley*, 1984] to mix with the overlying LSW and recirculate over the Porcupine Abyssal Plain [*van Aken*, 2000]. This southward penetration of deep-water masses is compensated for by the northward flow of warm, saline surface waters to the Norwegian Sea via the North Atlantic Drift (NAD) and Eastern North Atlantic Waters (ENAW). Today, a pole-ward flow occurs at all depth levels along the eastern slope of the Porcupine Seabight [*Rüggeberg et al.*, 2005], however, the morphology of the Porcupine Seabight prevents direct throughflow of waters below ~500 m water depth where currents are topographically steered in a cyclonic direction [*Frank et al.*, 2004]. ENAW overlies and mixes with Mediterranean Outflow Water (MOW), which currently bathes the site of MD01-2461 and is characterized by a salinity maximum and oxygen minimum at a depth of 1000–1200 m [*Frank et al.*, 2004].

[6] The 20.2 m core recovered olive-gray silty-clay sediments with frequent drop stones. Visual and x-radiograph evidence present no evidence of core disturbance or turbidite sequences within the interval studied. Geochronology of MD01-2461 is based on 25 monospecific foraminifera (*Globigerina bulloides* or *Neogloboquadrina pachyderma* sinistral) AMS ^{14}C dates and fine-tuning of the relative abundance of *Neogloboquadrina pachyderma* sinistral within the planktonic assemblage (*N. pachyderma* sin.%) with the GISPII δ^{18}O record [*Peck et al.*, 2007]. The GISPII-tuned age model (r^2 = 0.83, 27–8 kyr B.P.) suggests significantly enhanced ^{14}C-marine reservoir ages, exceeding 2 kyr during the deglacial [*Peck et al.*, 2006] (Figure 2a). Sedimentation rates range between 12 and 60 cm kyr^{-1}.

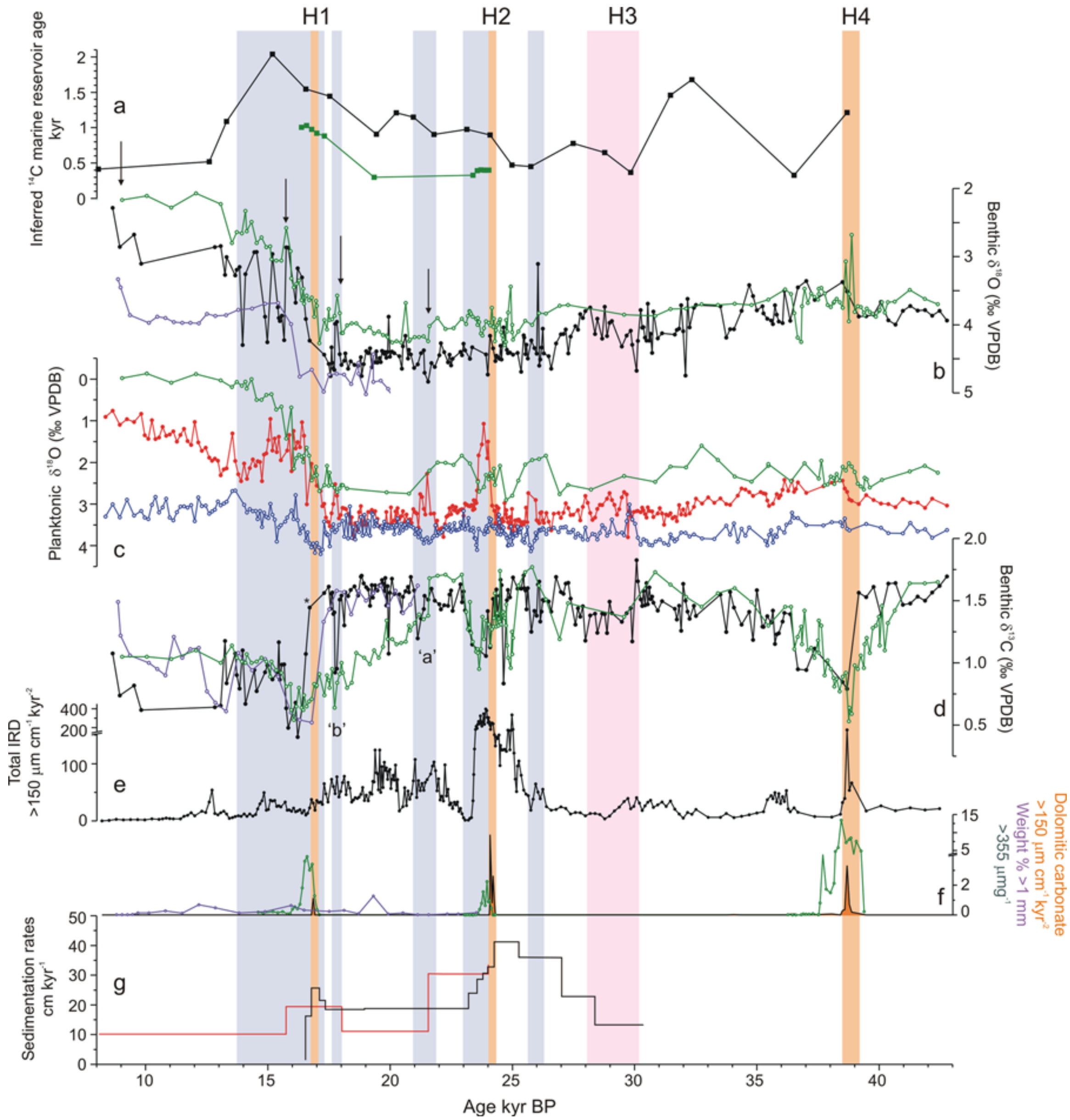

Figure 2. Benthic and planktonic stable isotope and IRD stacks from MD01-2461, SO75-26KL, and NEAP4K. Unless otherwise stated, records from MD01-2461 are black, closed circles; SO75-26KL are green, open circles; and NEAP4K are purple, open circles. (a) Inferred ^{14}C marine reservoir ages for MD01-2461 [*Peck et al.*, 2006] and SO75-26KL (see text). (b) δ^{18}O *C. wuellerstorfi*. Vertical arrows indicate tie-points synchronizing the record of SO75-26KL to MD01-2461 (SO75-26KL age model as *Zahn et al.* [1997] prior to 24 kyr B.P.). (c) δ^{18}O planktonic. Surface-dwelling *G. bulloides* (MD01-2461, red; SO75-26KL, green) and subsurface-dwelling *N. pachyderma* sin. from MD01-2461 (blue). (d) δ^{13}C *C. wuellerstorfi*. (e) Flux of total IRD at MD01-2461 (>150 μm cm^{-2} kyr^{-1}). (f) Flux of dolomtic carbonate at MD01-2461 (orange); concentration of total IRD at SO75-26KL (>355 μm g^{-1}); weight% >1 mm at NEAP4K. (g) Radiocarbon (black) and tuned (red) sedimentation rates at SO75-26KL. Blue vertical bars highlight episodes of surface ocean stratification. Orange vertical bars highlight H layers 4, 2, and 1, with H3 located by the purple bar.

[7] Lithic grains >150 μm embedded within the silty clay sediment are considered to be ice rafted, and lithological and geochemical examination has been used to determine the phasing of circum-North Atlantic ice sheet instability and iceberg rafting to the core site [*Peck et al.*, 2007]. Vertical structure of the upper water column is inferred from δ^{18}O records of surface- and subsurface-dwelling planktonic foraminifera *Globigerina bulloides* and *Neogloboquadrina pachyderma* sinistral, calcifying

at an average depth of 30 m and 150 m respectively [see *Peck et al.*, 2006]. Benthic, epifaunal species *Cidicidoides wuellerstorfi* has been found to accurately record the δ^{13}C of dissolved inorganic carbon in ambient bottom water, making down-core benthic δ^{13}C records a useful proxy for the reconstruction of past bottom water mass changes [*Duplessy et al.*, 1984]. Youngest benthic δ^{13}C values from MD01-2461 of +1.1‰ VPDB compare well with core top values derived from nearby core sites at similar water depths (1.07‰ VPDB at V29–198, 1139 m water depth; 1.17‰ VPDB at V28–73, 2063 m water depth) [*Oppo and Lehman*, 1993] and closely correlates with δ^{13}C of ambient water total CO_2 of +1.0 to +1.1‰ VPDB (between 1000–2000 m water depth, GEOSECS station 23; Figure 1a). Between 1 and 4 specimens of *C. wuellerstorfi* were selected from the >250 μm size fraction wherever present, providing an average time step of 160 ± 56 kyr for benthic isotope records. Stable isotope analyses were made using a Thermo-Finnigan MAT 252 with an external reproducibility of $\leq$0.08‰ for δ^{18}O and 0.03‰ for δ^{13}C. Following *Shackleton and Opdyke* [1973] a disequilibrium correction of +0.64‰ is applied to $\delta^{18}O_{C.\ wuellerstorfi}$ to allow direct comparison with other benthic records.

[8] Whole ocean δ^{13}C changes [e.g., *Curry et al.*, 1988; *Duplessy and Shackleton*, 1985; *Duplessy et al.*, 1988] will have affected the epibenthic δ^{13}C records from the different core sites used in this study equally and therefore are not considered an obstacle in our documentation of regional δ^{13}C gradients. Correction of our record for changes of the global ocean carbon reservoir demand robust constraints on mean ocean δ^{13}C variations that occurred during the last 40 kyr at high temporal resolution, which are currently unavailable.

3. Results and Discussion

3.1. Records From MD01-2461

[9] Elevated benthic δ^{13}C values of 1.65‰ VPDB during the last glacial reflect the influence of a well-ventilated mid-depth water mass, GNAIW [*Oppo and Lehman*, 1993] or Upper North Atlantic Deep Water (UNADW) [*Boyle and Keigwin*, 1987]. H events recorded in MD01-2461, identified through the occurrence of IRD deposition with contribution of distinctive Hudson Strait-sourced dolomitic carbonate (Figure 2f) [*Peck et al.*, 2006], coincide with benthic δ^{13}C depletions of 0.6‰ to 1.25‰. A smaller negative δ^{13}C anomaly, of $\sim$0.3‰, is associated with an enhanced flux of IRD at $\sim$30 kyr B.P. likely recording the occurrence of H3. This suggestion is supported by the apparent lack of LIS-sourced IRD in this interval, which has been considered a type-characteristic of H3 along with a strong volcanic component to the IRD assemblage [e.g., *Grousset et al.*, 1993; *Gwiazda et al.*, 1996], likely of East Greenland/Icelandic origin [*Peck et al.*, 2007]. Decreased benthic δ^{13}C can potentially be ascribed to high vertical fluxes of organic carbon [*Mackensen et al.*, 1993]. However, the negative benthic δ^{13}C anomalies associated with H events at MD01-2461 are considered to reflect, principally, bottom water mass changes [cf. *Willamowski and Zahn*, 2000], most feasibly advances of SCW, likely a glacial version of AAIW, to our core site in response to decreased or halted GNAIW formation, confirming restriction of deep-intermediate convection in the northern North Atlantic in the course of the events [e.g., *Cortijo et al.*, 1997; *Vidal et al.*, 1997; *Zahn et al.*, 1997; *Rickaby and Elderfield*, 2005].

[10] An additional control on the ventilation of deep-intermediate waters in the North Atlantic during meltwater events might involve a change in the mode of North Atlantic deep water formation, from open-ocean convection under glacial conditions to buoyancy loss due to brine formation in the course of sea-ice formation in the Nordic Seas following meltwater surges [*Dokken and Jansen*, 1999]. Recent studies have suggested that brine formation proximal to ice margins persisted throughout the last glacial in the northern North Atlantic concurrent with open-ocean convection in the North Atlantic [*Labeyrie et al.*, 2005]. The most likely preformed poorly ventilated signal [*Vidal et al.*, 1997; *Elliot et al.*, 2002] and low δ^{18}O signature of these intermediate brine waters being masked during ambient glacial conditions by the well-ventilated GNAIW and only becoming apparent during episodes of reduced MOC, when simultaneous negative anomalies in both planktonic and benthic δ^{18}O and benthic δ^{13}C are observed in North Atlantic records [*Dokken and Jansen*, 1999; *Labeyrie et al.*, 2005; *Waelbroeck et al.*, 2006]. Such an interpretation challenges the interpretation of North Atlantic benthic δ^{13}C anomalies as documenting the northward penetration of southern sourced waters in response to reduced MOC as the rapid transmission of the North Atlantic-derived brine signal to both the Indian and Pacific Ocean necessitates active intermediate water formation [*Labeyrie et al.*, 2005; *Waelbroeck et al.*, 2006].

Table 1. Site Locations Discussed in This Study

Core	Location	Water Depth, m	Reference
NEAP 4K	Björn Drift 61°29′N, 24°10′W	1627	*Rickaby and Elderfield* [2005]
DAPC2	Rockall Trough 58°58′N, 09°37′W	1709	*Knutz et al.* [2002]
SO75-26KL	Iberian Margin 37°49′N, 20°30′W	1099	*Zahn et al.* [1997]
SU81-18	Iberian Margin 37°46′N, 10°11′W	3135	*Gherardi et al.* [2005]

[11] At site MD01-2461 we also see some evidence for the injection of low δ^{18}O intermediate waters, characteristic of formation via brine rejection, notably at meltwater event "b" (18–17.7 kyr B.P.) with a well defined decrease of benthic δ^{18}O by 0.7‰ associated with a transient negative benthic δ^{13}C anomaly. Further benthic δ^{18}O (brine) anomalies are hinted at during H4, H3 and meltwater events at ~26 kyr B.P. and "a" (21.8–20.9 kyr) although these anomalies are much less distinct and appear absent at H2. This lack of a clear brine signal together with the recently reported covariance of $^{231}Pa_{xs}/^{230}Th_{xs}$ and benthic δ^{13}C records from core DAPC2 [*Hall et al.*, 2006] provides convincing evidence that changes in bottom water ventilation along the northern European margin (inferred from benthic δ^{13}C) relate to changes in the rate of MOC and the interchange of poorly ventilated southern and well ventilated northern sourced water masses (at least up to ~1700 m water depth) rather than the dominant injection of poorly ventilated brine waters from the Nordic Seas [*Labeyrie et al.*, 2005]. This interpretation of benthic δ^{13}C data provides the basis for the following discussion.

[12] While initial benthic δ^{13}C decreases in MD01-2461 directly coincide with the onset of H layer deposition, peak minima in benthic δ^{13}C coincide with maximum divergence in δ^{18}O from surface- and subsurface-dwelling planktonic foraminifers (Figure 2b) that we interpret as indications of meltwater stratification of the upper water column [*Peck et al.*, 2006]. Benthic δ^{13}C values fall to their lowest values within the entire record, +0.4‰ VPDB, at 16.2 kyr B.P., some 0.6–1.1 kyr after the incursion of the detrital carbonate peak indicative of H1 deposition at this site, coincident with offset between δ^{18}O *G. bulloides* and *N. pachyderma* sin. of >2.5‰ indicating prominent meltwater stratification. Deglacial meltwater forcing is evident from persistent upper ocean stratification through 16.4–14.0 kyr B.P., associated with significant sea level rise at this time [*Bard et al.*, 1990, 1996]. Benthic δ^{13}C values do not return to their elevated glacial levels but vary between 0.7–1.2‰ VPDB throughout the deglacial, plausibly reflecting a combination of changes involving the terrestrial biosphere and its influence on marine carbon reservoir δ^{13}C and establishment of the modern mode of North Atlantic THC. A decrease in benthic δ^{13}C of 0.5‰ at 13 kyr B.P. is captured by two data points only, but potentially reflects a convection slow-down in the North Atlantic during the Younger Dryas cold period [*Boyle and Keigwin*, 1987; *Sarnthein et al.*, 1994; *Rickaby and Elderfield*, 2005]. In addition to the H events, short-lived negative δ^{13}C anomalies are observed at 26–25.7 kyr, 21.8–20.9 kyr ("a") and 18–17.7 kyr ("b") B.P., and directly coincide with intermittent meltwater stratification of the upper water column, and enhanced deposition of NWEIS-sourced IRD. Our timings of the two most recent events, "a" and "b" support dates of British ice sheet (BIS) ice sheet retreat determined from cosmogenic nuclide (^{36}Cl) surface-exposure dating of glacial erratic boulders and glacially smoothed bedrock sampled around the former ice margins in Ireland [*Bowen et al.*, 2002]. IRD associated with meltwater event "b" and an IRD peak with a similar NWEIS-signature prior to H2 (at 25.0 kyr B.P.) likely correspond to the well-documented "European precursor" events [e.g., *Grousset et al.*, 2000; *Scourse et al.*, 2000]. Freshwater surging associated with these episodes of NWEIS instability conceivably reached the area of GNAIW formation and caused transient, 200–500 year, reduction in northern sourced intermediate water flux, allowing brief northward penetration of SCW to the MD01-2461 site [*Peck et al.*, 2006].

3.2. Intermediate Water Ventilation Changes From NE Atlantic Records

[13] We compare the benthic δ^{13}C record of MD01-2461 with similar records of other North Atlantic sites (Table 1) to determine regional variability of intermediate water masses. In particular, we use the high-resolution record from core SO75-26KL from the Portuguese Margin at a water depth of 1099 m, similar to that of MD01-2461 [*Zahn et al.*, 1997]. Data records from SO75-26KL have

been published on a radiocarbon timescale applying a constant ^{14}C-marine reservoir age correction of 400 years [*Zahn et al.*, 1997; *Willamowski and Zahn*, 2000]. Assessment of multiple ^{14}C data [e.g., *Voelker et al.*, 1998; *Waelbroeck et al.*, 2001], in agreement with age modeling of the MD01-2461 records [*Peck et al.*, 2006], suggest that NE Atlantic marine ^{14}C reservoir ages were highly variable during the glacial period such that the radiocarbon timescale of SO75-26KL does not allow for a detailed comparison with MD01-2461. To attempt a synchronized timescale for SO75-26KL we graphically correlate the benthic δ^{18}O records of both cores by "tuning" the record of benthic δ^{18}O record from SO75-26KL to that of MD01-2461 across the deglaciation (24–8 kyr B.P.) (Figure 2b). This procedure receives independent support from the observation that an IRD peak at SO75-26KL, on the synchronized timescale, is concurrent with the lithologically and geochemically distinct H layer 1 at MD01-2461, consistent with synchronous deposition of H layers 4 and 2 at the two sites (Figure 2f). Additionally, the onset of deglacial warming/surface ocean freshening in the planktonic δ^{18}O records is simultaneous at the two sites (Figure 2c). The synchronized timescale of SO75-26KL suggests marine ^{14}C-reservoir ages between 0.4 kyr to 0.9–1.0 kyr in the period 17.3–16.4 kyr B.P. at this site (Figure 2a). A latitudinal gradient of reservoir ages during the deglaciation from ~1 kyr at 37°N (SO75-26KL) up to ~2 kyr at 52°N (MD01-2461) [*Peck et al.*, 2006] is comparable with the findings of *Waelbroeck et al.* [2001].

[14] The lower resolution benthic δ^{13}C record from core NEAP 4K at Björn Drift (1627 m water depth; Figure 1) [*Rickaby and Elderfield*, 2005] is used as reference for comparison with mid-depth ventilation changes in the high-latitude North Atlantic. The age model of NEAP 4K (>13 kyr B.P.) is based on stratigraphical correlation (benthic and planktonic δ^{18}O) with core BOFS 5K [*Barker et al.*, 2004], which has a radiocarbon-based age model incorporating the elevated marine reservoir ages of *Waelbroeck et al.* [2001] (1.9 kyr) for this time period. Temporal resolution at this site averages 500 years and does not allow for a detailed correlation of the benthic δ^{18}O record with that of MD01-2461. A broad peak in weight% of the >1 mm size fraction, spanning ~20–9 kyr B.P. is of little use for correlation to the IRD events at the European Margin sites. Therefore we have no firm control on the timing of isotope patterns along the benthic isotope records of NEAP 4K in relation to MD01-2461 and SO75-26KL and will use the record for a qualitative assessment of regional gradients only.

[15] During the last glacial, an eastern branch of GNAIW was advected along the European Margin toward the Portuguese Margin, while northward flowing SCW (likely AAIW) penetrated as far north as the Moroccan Margin as is indicated in paired benthic δ^{13}C-Cd/Ca profiles (Figure 1a) [*Willamowski and Zahn*, 2000]. GNAIW therefore maintained the well-ventilated ambient bottom water conditions (δ^{13}C > 1.4‰ VPDB) recorded at both MD01-2461 and SO75-26KL during mean glacial conditions. Conversely, the offset, of up to 0.5‰, between the MD02-2461 and SO75-26KL benthic δ^{18}O records over the glacial interval suggests the additional influence of a warmer and/or low-δ^{18}O glacial mid-depth water mass at the upper Portuguese margin [*Zahn et al.*, 1997].

[16] Similar absolute values of benthic δ^{13}C are recorded at both SO75-26KL and MD01-2461 associated with and following IRD deposition at H4, 2 and 1. However, decline into these benthic δ^{13}C anomalies starts considerably earlier at the Portuguese margin and appears more gradual, notably in the periods prior to H4 and H1. That is, mid-depth ventilation appears to deteriorate at the Portuguese margin up to 5 kyr prior to these two H events, and importantly, before the collapse of mid-depth ventilation at the site of MD01-2461. This contrast in mid-depth ventilation plausibly reflects the proximity of the northerly MD01-2461 to the site of mid-depth convection that provided the site with well-ventilated mid-depth waters even though the production of these waters was in decline. If so, the early decrease in benthic δ^{13}C at the Portuguese margin suggests that the production of mid-depth waters started to deteriorate well before H1 and H4 presumably because of a gradual built-up of meltwater surging before the LIS destabilized and large-scale iceberg calving occurred. The benthic δ^{13}C proxy close to the centers of convection would not resolve such an early decline in mid-depth convection.

3.3. Precursory Meltwater Forcing From the NWEIS

[17] GNAIW convection in the open North Atlantic at the LGM [*Sarnthein et al.*, 1994; *Vidal et al.*, 1997], proximal to the fully advanced NWEIS, facilitated close coupling of NWEIS meltwater and overturning circulation. Within the H2 to H1 interval, benthic δ^{13}C at MD01-2461 suggests that bottom waters at this site remained well ventilated

by GNAIW until ~16.6 kyr B.P., the exception being the apparent transient advance of the GNAIW/SCW hydrographic front north of ~52°N at meltwater events "a" and "b." At SO75-26KL, benthic δ^{13}C fell steadily over a ~5 kyr period approaching H1, following an abrupt shift by −0.5‰ at 21.4 kyr B.P. Initiation of this trend of decreasing ventilation at SO75-26KL is coincident with event "a" at site MD01-2461 when the apparent advance of SCW as displayed in a likewise brief episode of benthic δ^{13}C depletion centered on 21.2 kyr B.P. On the basis of an IRD assemblage at MD01-2461 dominated by BIS-derived lithologies, this event has been suggested to constitute a reduction in GNAIW formation during an episode of NWEIS instability and associated meltwater surging into the NE Atlantic [*Peck et al.*, 2006]. The end of this freshwater surge allowed intermediate water production to resume, reverting benthic δ^{13}C values at MD01-2461 back to elevated/GNAIW values at 20.9 kyr B.P. However, continuing decrease of benthic δ^{13}C values recorded at SO75-26KL suggest GNAIW production did not fully recover and that North Atlantic THC was progressively weakening prior to the incursion of H1 icebergs and meltwater to the NE Atlantic. As no further meltwater forcing is evident in the planktonic δ^{18}O records of MD01-2461 until event "b," at 17.8 kyr B.P., convection was plausibly reduced by freshwater forcing at higher latitudes [e.g., *Elliot et al.*, 2002] or further to the west. $^{231}Pa_{xs}/^{230}Th_{xs}$ ratios at DAPC2 (1709 m water depth) in the Rockall Trough, increase toward production values at ~18.0 kyr B.P., suggest substantially reduced rates of overturning [*Hall et al.*, 2006] concurrent with event "b" which may represent either a second transitory advance of SCW or an episode of brine injection to MD01-2461 preceding H1 by ~0.9 kyr. However, *Gherardi et al.* [2005] use $^{231}Pa_{xs}/^{230}Th_{xs}$ records from the Iberian Margin to suggest that "shallow" overturning in the NE Atlantic basin was vigorous until 16.5 kyr B.P. Core SU81-18 used in their study is located at 3135 m water depth, some 2000 m deeper than both MD01-2461 and SO75-26KL. It may be possible that ^{231}Pa export at depths below cores SO75-26KL, MD01-2461 and DAPC2 is recorded at SU81-18 accounting for these divergent signals, supporting a reduction in shallow overturning (≤1700 m), while deeper convection was perhaps maintained, accounting for the lower $^{231}Pa_{xs}/^{230}Th_{xs}$ values recorded at SU81-18 at this time.

[18] The low-resolution record of NEAP-4K closely follows the structure of the benthic δ^{13}C record of MD01-2461 and documents elevated δ^{13}C values around 1.4‰ throughout the interval between H2 and H1, suggesting persistent bathing of this site with GNAIW. This pattern confirms our contention of continued production of GNAIW, albeit at lower rates in response to NWEIS instabilities, at "a" and "b."

[19] Benthic δ^{13}C values of <0.7‰ VPDB at all three intermediate water depth sites reflect the prominence SCW between 16.2–15.4 kyr B.P., implying large-scale collapse of GNAIW production following H1. Unlike the coupled H layer-benthic δ^{13}C collapse records of SO75-26KL, NEAP-4K and DAPC2 [*Knutz et al.*, 2002; *Hall et al.*, 2006], significantly reduced ventilation at MD01-2461 is not recorded until several hundred years after H1. Geochemical (^{40}Ar/^{39}Ar dates of individual hornblende grains, Sr-Nd isotopic composition of carbonate-free IRD), magnetic susceptibility and lithological classification (dolomitic carbonate) each suggest that this is the only horizon within the deglacial interval (20–10 kyr B.P.) of MD01-2461 that contains a notable contribution of LIS-derived debris [*Peck et al.*, 2006]. The H1 layer in MD01-2461 spans a few hundred years, whereas deposition at both DAPC2 and SO75-26KL is in the order of 1.0 kyr. While it appears plausible that MD01-12461 may have witnessed the earlier stages of H1 deposition in the NE Atlantic only, perhaps reflecting changing surface current patterns, it remains an issue to explain why benthic δ^{13}C at MD01-2461 does not decrease until after the H1 layer in this core. One possible explanation is that the elevated δ^{13}C value recorded immediately following H1 layer deposition and measured from a single specimen of *C. wuellerstorfi* (asterisk on Figure 2d) is not representative of ambient bottom water ventilation at this time and ventilation reduction at MD01-2461 was effectively simultaneous with H1. Alternatively, taking the benthic δ^{13}C record at MD01-2461 at face value across H1, the elevated benthic δ^{13}C levels may reflect a time-transgressive shoaling of reduced ventilation that reached the deeper sites in the north (NEAP-4K, DAPC2) first, before affecting MD01-2461.

[20] At H2, a "precursory" reduction in benthic δ^{13}C is observed at SO75-26KL at 25 kyr B.P., ~1 kyr prior to H layer deposition, accompanied by a brief anomaly at MD01-2461 at 24.6 kyr B.P. The abrupt decrease in benthic δ^{13}C at SO75-26KL

is synchronous with the equally abrupt increase in IRD from the NWEIS at MD01-2461, representing instability of the BIS immediately preceding H2 [*Peck et al.*, 2007]. Surface and subsurface planktonic δ^{18}O display an only minor divergence at this time (up to 0.5‰) indicating freshwater forcing at a smaller scale than during the "a" and "b" events, but its influence on GNAIW formation appears significant enough to produce the reduction in ventilation observed at SO75-26KL. Mid-depth ventilation recovers abruptly after this event before large-scale ventilation collapse occurs in the course of H2. Minimum ventilation of intermediate waters along the European Margin is seen in the benthic δ^{13}C record of both SO75-26KL and MD01-2461 that immediately follows the H2 IRD deposition in MD01-2461. Significant destabilization of the BIS is suggested by a substantial increase in the flux of BIS-derived debris immediately following deposition of H2 at MD01-2461, perhaps triggered by sea level rise of up to 15 m associated with LIS-collapse [*Yokoyama et al.*, 2001; *Chappell*, 2002]. Consistently the maximum (>2.5‰) offset between δ^{18}O of the *G. bulloides* and *N. pachyderma* sin. which lags H layer deposition by ~300 years is thought to be derived principally from NWEIS meltwater.

[21] Similarly, to H1, a progressive decrease in benthic δ^{13}C preceding H4 by ~2 kyr is observed in SO75-26KL, while benthic δ^{13}C depletion at MD01-2461 occurred abruptly associated with H layer deposition. The short-lived increase in IRD flux associated with H4 plausibly reflects the juvenile state of the BIS at this time [*Peck et al.*, 2007], a contention that appears to be confirmed by the lack of a coeval freshwater signal in planktonic δ^{18}O (Figure 2c). The limited extent of the NWEIS, coupled with North Atlantic deep-intermediate water convection likely occurring in a similar location to the contemporary ocean prior to the stage 3-2 boundary [*Sarnthein et al.*, 1994; *Vidal et al.*, 1997] suggest the NWEIS was an unlikely source of meltwater for triggering the reduction in mid-depth ventilation prior to H4. Rather, initial reduction in GNAIW production likely reflects other meltwater sources [*van Kreveld et al.*, 2000; *Elliot et al.*, 2002].

3.4. Ice Sheet Instability in Response to North Atlantic THC Changes

[22] Several hypotheses attempting to explain the occurrence of H events, incorporating a range of internal and external forcing factors, have been proposed [e.g., *MacAyeal*, 1993; *Johnson and Lauritzen*, 1995; *Marshall and Clarke*, 1997; *Hunt and Maslin*, 1998; *Arbic et al.*, 2004]. Recent concepts have used the disintegration of ice shelves fringing the Antarctic Peninsula as a modern analogue for the sudden iceberg releases during H events [*Hulbe*, 1997; *Hulbe et al.*, 2004] and consider ocean subsurface temperatures, coupled with North Atlantic THC variability, as a factor that may have destabilized the LIS through their effect on ice shelves and fringing ice margins [e.g., *Moros et al.*, 2002; *Shaffer et al.*, 2004; *Flückiger et al.*, 2006].

[23] The concept of recurrent meltwater release from the NWEIS and surface ocean stratification, promoting transient weakening of the North Atlantic THC and regional cooling particularly appears to apply for the period preceding H1. Such cool conditions may also have promoted the growth of a LIS-fringing ice-shelf, perhaps priming the LIS for H event collapse [*Hulbe*, 1997; *Hemming*, 2004; *Hulbe et al.*, 2004]. Subsequent subsurface warming and sea level rise (0.3–0.5 m), associated with the THC reduction, may then have played a role in undermining the ice shelf thus removing the buttressing support exerted on the feeder ice streams, leading to large-scale surging of the ice sheet [*Flückiger et al.*, 2006].

4. Conclusions

[24] Benthic δ^{13}C records from NE Atlantic core MD01-2461 document interchange of well-ventilated northern-sourced and poorly ventilated southern-sourced intermediate waters during the last glacial. Enhanced ventilation at MD01-2461 documented by δ^{13}C benthic values of >+1.6‰ VPDB, compared to early Holocene values of ≤1‰ VPDB suggests that well-ventilated GNAIW was bathing this core site, along with sites on the Björn Drift and Portuguese Margin at similar water depths during ambient LGM conditions. Frequent meltwater incursions, associated with instability of the NWEIS, appear coincident with weakened GNAIW formation leading to transient advances of SCW along the European margin. Such a scenario is likely to account for precursory reductions in North Atlantic THC prior to both H2 and H1 when NWEIS was at maximum extent and convection was more proximal. Progressive reduction in North Atlantic THC may also be inferred prior to H4; however, the limited extent of the NWEIS coupled with the likely location of convection

centers at higher latitudes precludes meltwater release from the NWEIS as the trigger.

[25] LIS destabilization therefore appears to have consistently occurred following progressive (H1 and H4) or discrete (H2) reductions in the North Atlantic THC. The role that North Atlantic THC variability may have played in H event initiation is still debatable, yet a scenario where climatic cooling in response to surface ocean stratification and reduced North Atlantic THC promoted ice shelf growth, which were subsequently ablated through accompanying subsurface warming, triggering ice sheet collapse is consistent with the data reported here.

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References

Arbic, B. K., D. R. MacAyeal, J. X. Mitrovica, and G. A. Milne (2004), Palaeoclimate—Ocean tides and Heinrich events, *Nature*, *432*, 460–460.

Bard, E., B. Hamelin, and R. G. Fairbanks (1990), U-TH ages obtained by mass-spectrometry in corals from Bardados–sea level during the past 130,000 years, *Nature*, *346*, 456–458.

Bard, E., B. Hamelin, M. Arnold, L. Montaggioni, G. Cabioch, G. Faure, and F. Rougerie (1996), Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge, *Nature*, *382*, 241–244.

Barker, S., T. Kiefer, and H. Elderfield (2004), Temporal changes in North Atlantic circulation constrained by planktonic foraminiferal shell weights, *Paleoceanography*, *19*, PA3008, doi:10.1029/2004PA001004.

Bowen, D. Q., F. M. Phillips, A. M. McCabe, P. C. Knutz, and G. A. Sykes (2002), New data for the Last Glacial Maximum in Great Britain and Ireland, *Quat. Sci. Rev.*, *21*, 89–101.

Boyle, E. A., and L. D. Keigwin (1987), North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature, *Nature*, *330*, 35–40.

Broecker, W. S., G. Bond, and M. A. Klas (1990), A salt oscillator in the glacial Atlantic? 1. The concept, *Paleoceanography*, *5*, 469–477.

Chappell, J. (2002), Sea level changes forced ice breakouts in the Last Glacial cycle: New results from coral terraces, *Quat. Sci. Rev.*, *21*, 1229–1240.

Cortijo, E., L. Labeyrie, L. Vidal, M. Vautravers, M. Chapman, J. C. Duplessy, M. Elliot, M. Arnold, J. L. Turon, and G. Auffret (1997), Changes in sea surface hydrology associated with Heinrich event 4 in the North Atlantic Ocean between 403N and 603N, *Earth Planet. Sci. Lett.*, *146*, 29–45.

Curry, W. B., and D. W. Oppo (2005), Glacial water mass geometry and the distribution of $\delta^{13}C$ of ΣCO_2 in the western Atlantic Ocean, *Paleoceanography*, *20*, PA1017, doi:10.1029/2004PA001021.

Curry, W. B., J. C. Duplessy, L. D. Labeyrie, and N. J. Shackleton (1988), Changes in the distribution of $\delta^{13}C$ of deep water CO_2 between the last glaciation and the Holocene, *Paleoceanography*, *3*, 317–341.

Dokken, T. M., and E. Jansen (1999), Rapid changes in the mechanism of ocean convection during the last glacial period, *Nature*, *401*, 458–461.

Duplessy, J. C., and N. J. Shackleton (1985), Response of global deep-water circulation to Earth's climatic change 135,000–107,000 years ago, *Nature*, *316*(6028), 500–507.

Duplessy, J.-C., N. J. Shackleton, R. K. Matthews, W. Prell, W. F. Ruddiman, M. Caralp, and C. Hendy (1984), 13-C record of benthic foraminifera in the last interglacial ocean: Implications for the carbon cycle and the global deep water circulation, *Quat. Res.*, *18*, 218–235.

Duplessy, J.-C., N. J. Shackleton, R. G. Fairbanks, L. Labeyrie, D. Oppo, and N. Kallel (1988), Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation, *Paleoceanography*, *3*(3), 343–360.

Elliot, M., L. Labeyrie, and J. C. Duplessy (2002), Changes in North Atlantic deep-water formation associated with the Dansgaard-Oeschger temperature oscillations (60–10 kyr), *Quat. Sci. Rev.*, *21*, 1153–1165.

Flückiger, J., R. Knutti, and J. W. C. White (2006), Oceanic processes as potential trigger and amplifying mechanisms for Heinrich events, *Paleoceanography*, *21*, PA2014, doi:10.1029/2005PA001204.

Frank, N., M. Paterne, L. Ayliffe, T. van Weering, J. P. Henriet, and D. Blamart (2004), Eastern North Atlantic deep-sea corals: Tracing upper intermediate water Delta C-14 during the Holocene, *Earth Planet. Sci. Lett.*, *219*, 297–309.

Ganopolski, A., and S. Rahmstorf (2001), Rapid changes of glacial climate simulated in a coupled climate model, *Nature*, *409*, 153–157.

Gherardi, J. M., L. Labeyrie, J. F. McManus, R. Francois, L. C. Skinner, and E. Cortijo (2005), Evidence from the Northeastern Atlantic basin for variability in the rate of the meridional overturning circulation through the last deglaciation, *Earth Planet. Sci. Lett.*, *240*, 710–723.

Grousset, F. E., J. A. Labeyrie, A. Sinko, M. Cremer, G. Bond, G. Duprat, E. Cortijo, and S. Huon (1993), Patterns of ice-rafted detritus in the glacial North Atlantic (40–55N), *Paleoceanography*, *8*, 175–192.

Grousset, F. E., C. Pujol, L. Labeyrie, G. Auffret, and A. Boelaert (2000), Were the North Atlantic Heinrich events triggered by the behavior of the European ice sheets?, *Geology*, *28*, 123–126.

Gwiazda, R. H., S. R. Hemming, and W. S. Broecker (1996), Provenance of icebergs during Heinrich event 3 and the contrast to their sources during other Heinrich episodes, *Paleoceanography*, *11*(4), 371–378.

Hall, I. R., S. B. Moran, R. Zahn, P. C. Knutz, C.-C. Shen, and R. L. Edwards (2006), Accelerated drawdown of meridional overturning in the late-glacial Atlantic triggered by transient pre-H event freshwater perturbation, *Geophys. Res. Lett.*, *33*, L16616, doi:10.1029/2006GL026239.

Hemming, S. R. (2004), Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global

climate imprint, *Rev. Geophys.*, *42*, RG1005, doi:10.1029/2003RG000128.
Hulbe, C. L. (1997), An ice shelf mechanism for Heinrich layer production, *Paleoceanography*, *12*, 711–717.
Hulbe, C. L., D. R. MacAyeal, G. H. Denton, J. Kleman, and T. V. Lowell (2004), Catastrophic ice shelf breakup as the source of Heinrich event icebergs, *Paleoceanography*, *19*, PA1004, doi:10.1029/2003PA000890.
Hunt, A. G., and P. E. Maslin (1998), Possible triggering of Heinrich events by ice-load-induced earthquakes, *Nature*, *393*, 155–158.
Johnson, R. G., and S. E. Lauritzen (1995), Hudson Bay-Hudson Strait jökulhlaups and Heinrich events: A hypothesis, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *117*, 123–137.
Knutz, P. C., I. R. Hall, R. Zahn, T. L. Rasmussen, A. Kuijpers, M. Moros, and N. J. Shackleton (2002), Multidecadal ocean variability and NW European ice sheet surges during the last deglaciation, *Geochem. Geophys. Geosyst.*, *3*(12), 1077, doi:10.1029/2002GC000351.
Labeyrie, L., C. Waelbroeck, E. Cortijo, E. Michel, and J. C. Duplessy (2005), Changes in deep water hydrology during the Last Deglaciation, *C. R. Geosci.*, *337*, 919–927.
MacAyeal, D. R. (1993), Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic's Heinrich events, *Paleoceanography*, *8*, 775–784.
Mackensen, A., H.-W. Hubberten, T. Bickert, G. Fischer, and D. K. Fütterer (1993), The δ^{13}C in benthic for a foraminiferal tests of *Fontbotia Wuellerstorfi* (Schwager) relative to the δ^{13}C of dissolved inorganic carbon in southern ocean deep water: Implications for glacial ocean circulation models, *Paleoceanography*, *8*(5), 587–610.
Marshall, S. J., and G. K. C. Clarke (1997), A continuum mixture model of ice stream thermomechanics in the Laurentide Ice Sheet: 2. Application to the Hudson Strait Ice Stream, *J. Geophys. Res.*, *102*, 20,615–20,637.
McCartney, M. S., and L. D. Talley (1984), Warm water to cold water conversion in the northern North Atlantic Ocean, *J. Phys. Oceanogr.*, *14*, 922–935.
McManus, J. F., R. Francois, J. M. Gherardi, L. D. Keigwin, and S. Brown-Leger (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, *428*, 834–837.
Moros, M., A. Kuijpers, I. Snowball, S. Lassen, D. Backstrom, F. Gingele, and J. McManus (2002), Were glacial iceberg surges in the North Atlantic triggered by climatic warming?, *Mar. Geol.*, *192*(4), 393–417.
Oppo, D. W., and S. J. Lehman (1993), Mid-depth circulation of the subpolar North-Atlantic during the last glacial maximum, *Science*, *259*, 1148–1152.
Peck, V. L., I. R. Hall, R. Zahn, H. Elderfield, F. E. Grousset, and J. D. Scourse (2006), High resolution evidence for linkages between European ice sheet instability and Atlantic meridional overturning circulation, *Earth Planet. Sci. Lett.*, *243*, 476–488.
Peck, V. L., I. R. Hall, R. Zahn, F. E. Grousset, and J. D. Scourse (2007), The relationship of Heinrich events and their European precursors over the past 60 kyr BP: A multi-proxy ice rafted debris provenance study in the North East Atlantic, *Quat. Sci. Rev.*, in press.
Rickaby, R. E. M., and H. Elderfield (2005), Evidence from the high-latitude North Atlantic for variations in Antarctic Intermediate water flow during the last deglaciation, *Geochem. Geophys. Geosyst.*, *6*, Q05001, doi:10.1029/2004GC000858.
Rüggeberg, A., C. Dullo, B. Dorschel, and D. Hebbeln (2005), Environmental changes and growth history of a cold-water carbonate mound (Propeller Mound, Porcupine Seabight), *Int. J. Earth Sci.*, doi:10.1007/s00531-005-0504-1.
Sarnthein, M., K. Winn, S. J. A. Jung, J.-C. Duplessy, L. Labeyrie, H. Erlenkeuser, and G. Ganssen (1994), Changes in east Atlantic deepwater circulation over the last 30,000 years: Eight time slice reconstructions, *Paleoceanography*, *9*(2), 209–268.
Scourse, J. D., I. R. Hall, I. N. McCave, J. R. Young, and C. Sugdon (2000), The origin of Heinrich layers: Evidence from H2 for European precursor events, *Earth Planet. Sci. Lett.*, *182*, 187–195.
Shackleton, N. J., and N. J. Opdyke (1973), Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 105 year and 106 year scale, *Quat. Res.*, *3*, 39–55.
Shaffer, G., S. M. Olsen, and C. J. Bjerrum (2004), Ocean subsurface warming as a mechanism for coupling Dansgaard-Oeschger climate cycles and ice-rafting events, *Geophys. Res. Lett.*, *31*, L24202, doi:10.1029/2004GL020968.
van Aken, H. M. (2000), The hydrography of the mid-latitude Northeast Atlantic Ocean: I, The deep water masses, *Deep Sea Res., Part I*, *47*, 757–788.
van Kreveld, S., M. Sarnthein, H. Erlenkeuser, P. Grootes, S. Jung, N. J. Nadeau, U. Pflauman, and A. Voelker (2000), Potential links between surging ice sheets, circulation changes, and the Dansgaard-Oeschger cycles in the Irminger Sea, 60–18 kyr, *Paleoceanography*, *15*, 425–442.
Vidal, L., L. Labeyrie, E. Cortijo, M. Arnold, J. C. Duplessy, E. Michel, S. Becqué, and T. C. E. van Weering (1997), Evidence for changes in the North Atlantic Deep Water linked to meltwater surges during the Heinrich events, *Earth Planet. Sci. Lett.*, *146*, 13–27.
Voelker, A. H. L., M. Sarnthein, P. M. Grootes, H. Erlenkeuser, C. Laj, A. Mazaud, M. J. Nadeau, and M. Schleicher (1998), Correlation of marine C-14 ages from the Nordic seas with the GISP2 isotope record: Implications for C-14 calibration beyond 25 kyr BP, *Radiocarbon*, *40*, 517–534.
Waelbroeck, C., J. C. Duplessy, E. Michel, L. Labeyrie, D. Paillard, and J. Duprat (2001), The timing of the last deglaciation in North Atlantic climate records, *Nature*, *412*, 724–727.
Waelbroeck, C., C. Levi, J.-C. Duplessy, L. Labeyrie, E. Michel, E. Cortijo, F. Bassinot, and F. Guichard (2006), Distant origin of circulation changes in the Indian Ocean during the last deglaciation, *Earth Planet. Sci. Lett.*, *243*, 244–251.
Weaver, A. J., O. A. Saenko, P. U. Clark, and J. X. Mitrovica (2003), Meltwater pulse 1A from Antarctica as a trigger of the bolling-allerod warm interval, *Science*, *299*, 1709–1713.
Willamowski, A., and R. Zahn (2000), Upper ocean circulation in the glacial North Atlantic from benthic foraminiferal isotope and trace element fingerprinting, *Paleoceanography*, *15*, 515–527.
Yokoyama, Y., T. M. Esat, and K. Lambeck (2001), Coupled climate and sea level changes deduced from Huon Peninsula coral terraces of the last ice age, *Earth Planet. Sci. Lett.*, *193*, 579–587.
Zahn, R., J. Schönfeld, H. Kudrass, M. Park, H. Erlenkeuser, and P. Grootes (1997), Thermohaline instability in the North Atlantic during meltwater events: Stable isotope and ice-rafted detritus records from core SO75-26KL, *Portuguese margin, Paleoceanography*, *12*(5), 696–710.