The timing of deglacial circulation changes in the Atlantic

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[1] Well-dated benthic foraminifer oxygen isotopic records (δ^{18}O) from different water depths and locations within the Atlantic Ocean exhibit distinct patterns and significant differences in timing over the last deglaciation. This has two implications: on the one hand, it confirms that benthic δ^{18}O cannot be used as a global correlation tool with millennial-scale precision, but on the other hand, the combination of benthic isotopic records with independent dating provides a wealth of information on past circulation changes. Comparing new South Atlantic benthic isotopic data with published benthic isotopic records, we show that (1) circulation changes first affected benthic δ^{18}O in the 1000–2200 m range, with marked decreases in benthic δ^{18}O taking place at ~17.5 cal. kyr B.P. (ka) due to the southward propagation of brine waters generated in the Nordic Seas during Heinrich Stadial 1 (HS1) cold period; (2) the arrival of δ^{18}O-depleted deglacial meltwater took place later at deeper North Atlantic sites; (3) hydrographic changes recorded in North Atlantic cores below 3000 m during HS1 do not correspond to simple alternations between northern- and southern-sourced water but likely reflect instead the incursion of brine-generated deep water of northern as well as southern origin; and (4) South Atlantic waters at ~44°S and ~3800 m depth remained isolated from better-ventilated northern-sourced water masses until after the resumption of North Atlantic Deep Water (NADW) formation at the onset of the Bølling-Allerod, which led to the propagation of NADW into the South Atlantic.

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1. Introduction

[2] In order to decipher causal links between changes in climate and ocean circulation, deep-sea marine records have to be placed in the same time frame as polar ice and continental records. Ideally, all these climatic records should be dated in calendar years B.P., so that phasing between the climatic response and insolation forcing can be assessed. The last deglaciation period is especially interesting in that respect, because it belongs to the time span covered by ^{14}C dating technique. The ^{14}C dates can be converted into calendar ages using calibration curves describing the variations in the atmospheric $^{14}C/^{12}C$ ratio over the past 30 to 50 kyr [*Reimer et al.*, 2009]. The ^{14}C dating of marine records is however problematic in certain regions where the surface reservoir age has not remained constant over time [*Siani et al.*, 2000; *Sikes et al.*, 2000; *Waelbroeck et al.*, 2001]. This can be corrected in oceanic regions surrounding polar ice sheets by synchronizing sea surface temperature (SST) recorded in marine cores with air temperature recorded in polar ice cores, based on the assumption of thermal equilibrium between surface water and overlying air [*Govin*, 2008; *Govin et al.*, 2009; *Skinner and Shackleton*, 2004; *Skinner et al.*, 2010; *Waelbroeck et al.*, 2001].

[3] Oxygen stable isotopic composition ($^{18}O/^{16}O$) of foraminifer tests has been used as a chronostratigraphical tool for the last 40 years [*Imbrie et al.*, 1984; *Lisiecki and Raymo*, 2005, 2009; *Shackleton and Opdyke*, 1973]. On orbital time scales, and averaged over the whole ocean, planktonic and benthic $^{18}O/^{16}O$ (δ^{18}O, expressed in ‰ versus VPDB) reflect changes in global ice volume. However, because δ^{18}O of foraminifer calcite is a function of the temperature and isotopic composition of the water in which calcification takes place, it also records changes in local temperature and water δ^{18}O. As deep water temperature and δ^{18}O vary less than surface water temperature and δ^{18}O, benthic foraminifer δ^{18}O has been preferentially used as a global correlation tool [*Lisiecki and Raymo*, 2005, 2009]. However, a significant fraction of the benthic δ^{18}O signal does not derive from changes in global ice volume and hence is not a global signal [*Labeyrie et al.*, 2005; *Skinner and Shackleton*, 2005; *Waelbroeck et al.*, 2002, 2008]. Downcore pore fluid δ^{18}O measurements have shown that only 50% to 70% of the total benthic δ^{18}O glacial-interglacial amplitude can be attributed to the mean ocean δ^{18}O signal

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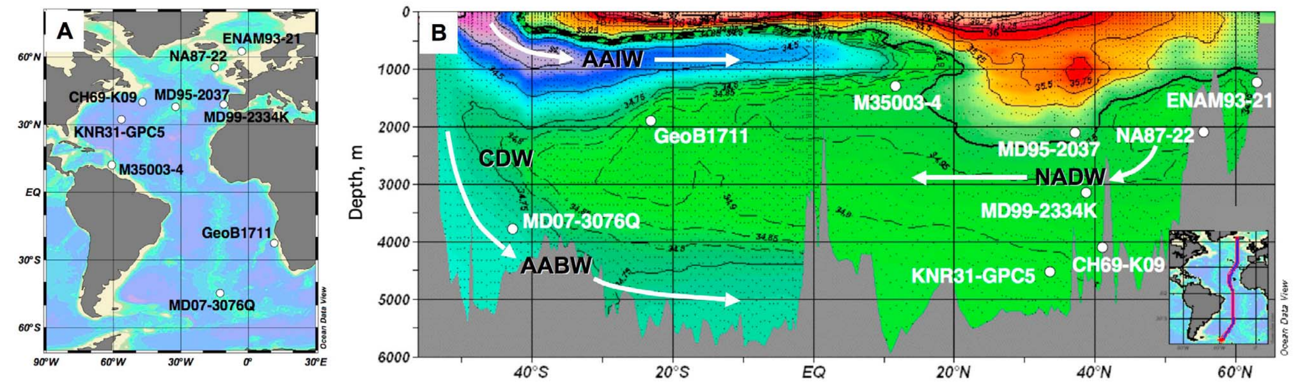

Figure 1. (a) Cores location map. (b) Salinity section of the Atlantic Ocean showing the principal water masses at present: white arrows indicate the approximate water flow direction of North Atlantic Deep Water (NADW), Antarctic Intermediate Water (AAIW), and Antarctic Bottom Water (AABW). Circumpolar Deep Water (CDW) is a mixture of NADW, AABW, and AAIW recirculating within the Antarctic Circumpolar Current [*Orsi et al.*, 1999].

resulting from changes in global ice volume, while the remainder of the benthic δ^{18}O glacial-interglacial amplitude is explained by changes in local bottom water temperature and δ^{18}O [*Adkins et al.*, 2002; *Schrag et al.*, 1996]. Furthermore, the timing of deglacial changes in these two components need not be in phase at a given site, or synchronous globally.

[4] Here we present a new well-dated benthic isotopic record from a deep South Atlantic site and compare it with published Atlantic benthic isotopic records versus calendar age from different water depths and latitudes. We demonstrate that not only the amplitude, but also the timing of the benthic δ^{18}O glacial-interglacial transition greatly varies from one depth range to another within the Atlantic Ocean basin. Using additional information from benthic foraminifer carbon isotopic composition, we interpret the differences in benthic δ^{18}O glacial-interglacial amplitude and timing in terms of changes in the distribution of different water masses within the Atlantic basin.

2. Material and Methods

2.1. Study Area

[5] We present new data from core MD07-3076Q (44°09.19′S, 14°13.70′W, 3770 m water depth), retrieved on the eastern flank of the Mid-Atlantic Ridge, in the Atlantic sector of the Southern Ocean. Site MD07-3076 is currently bathed in a predominantly southward flowing mixture of modified North Atlantic Deep Water (NADW) and Lower Circumpolar Deep Water (LCDW), which feeds into the core of eastward flowing Circumpolar Deep Water (CDW) [*Talley*, 2008] (Figure 1). We compare core MD07-3076 benthic isotopic records with well-dated published benthic isotopic records from various depths between 1000 and 5000 m in the Atlantic Ocean and Nordic Seas (Figure 1 and Table 1). We selected *Cibicides* isotopic records with sufficiently high temporal resolution and low dating uncertainty over the 10–21 ka time interval to allow assessment of the phasing between the different records over that time interval. Site ENAM93-21 is located in the Nordic Seas, north of the Faeroe-Shetland Channel, close to a major area of outflow of Norwegian Sea Deep Water into the North Atlantic [*Rasmussen et al.*, 1998]. Site NA87-22 is situated in the northern North Atlantic, on the southern side of the Iceland-Scotland Ridge and is currently bathed by a mixture of Upper NADW and Norwegian Sea Deep Water [*Vidal et al.*, 1997]. Site MD95-2037 is located further south on the eastern slope of the Mid-Atlantic Ridge. At present, site MD95-2037 is bathed at depth by the southward flow of the Upper NADW [*Gherardi et al.*, 2009]. Site MD99-2334K, at 3146 m depth on the Iberian Margin, is currently bathed in northward flowing Northeast Atlantic Deep Water, a mixture of about 50% NADW and 50% Lower Deep Water which is derived from Antarctic Bottom Water (AABW) [*van Aken*, 2000]. Site CH69-K09 is located at 4100 m depth and ~42°N in the Western Atlantic, and is currently bathed by the

Table 1. Cores Coordinates and Data References

Site	Ocean	Latitude	Longitude	Depth (m)	References: Benthic isotopes	References: Dating
ENAM93-21	Atl N	62°44.31′N	3°59.92′W	1020	*Rasmussen et al.* [1996]	*Rasmussen et al.* [1998]
NA87-22	Atl N	55°29.8′N	14°41.7′W	2161	*Vidal et al.* [1997]	*Waelbroeck et al.* [2001, 2006]
MD95-2037	Atl N	37°05.23′N	32°01.87′W	2159	*Labeyrie et al.* [2005]	*Gherardi et al.* [2009]
M35003-4	Atl N	12°05′N	61°15′W	1299	*Rühlemann et al.* [2004], *Zahn and Stüber* [2002]	*Rühlemann et al.* [1999]
GeoB1711	Atl S	23°18.9′S	12°22.6′E	1967	*Little et al.* [1997], *Vidal et al.* [1999]	*Vidal et al.* [1999], *Waelbroeck et al.* [2006]
CH69-K09	Atl N	41°45.4′N	47°21.0′W	4100	*Labeyrie et al.* [1999]	*Waelbroeck et al.* [2001]
MD99-2334K	Atl N	37°48′N	10°10′W	3146	*Skinner and Shackleton* [2004]	*Skinner et al.* [2003]
KNR31 GPC-5	Atl N	33°41.2′N	57°36.9′W	4583	*Keigwin et al.* [1991]	*Keigwin and Jones* [1994]
MD07-3076Q	Atl S	44°09.19′S	14°13.70′W	3770	this study	*Skinner et al.* [2010]

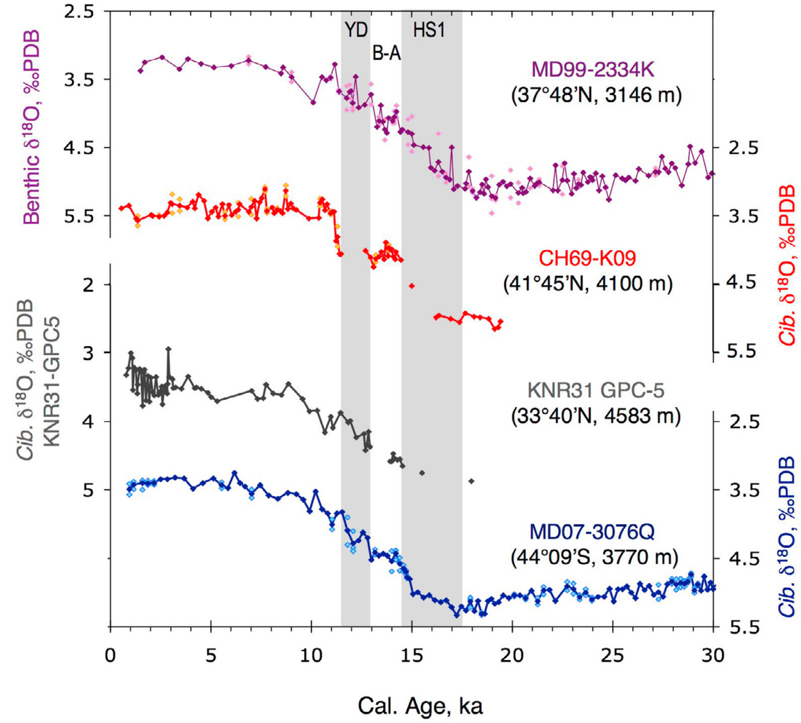

Figure 2. Atlantic benthic $\delta^{18}O$ records from the 3100–4500 m depth range versus calendar age. Core CH69-K09, KNR31 GPC-5, and MD07-3076 *Cibicides* records have been corrected by +0.64‰ in order to account for their equilibrium offset with respect to *Uvigerina* [*Duplessy et al.*, 1984]. Core MD99-2334K benthic $\delta^{18}O$ signal is based on measurements on *Globobulimina affinis* and *Cibicides wuellerstorfi* corrected for their respective equilibrium offsets [*Shackleton et al.*, 2000]. Light filled symbols indicate replicate measurements. Vertical gray stripes indicate the Heinrich Stadial 1 (HS1), Bølling-Allerod (B-A), and Younger Dryas (YD) North Atlantic chronozones. Complete references for isotopic measurements and age models are given in Table 1.

Lower NADW contour current [*Labeyrie et al.*, 1999]. Site KNR31GPC-5 located at 4583 m depth on the Bermuda Rise is currently also bathed in Lower NADW [*Talley*, 2008]. Site M35003 is located on the Atlantic side of the Caribbean sill and is currently in the transition zone between Antarctic Intermediate Water (AAIW) and Upper NADW [*Rühlemann et al.*, 2004]. Core GeoB1711 was taken from the Cape Basin, but above the Walvis Ridge water depth, so that the site is currently under the influence of the southward flow of Upper NADW [*Vidal et al.*, 1999].

2.2. Dating

[6] Core MD07-3076 age model is based on 49 monospecific planktonic radiocarbon dates that have been corrected for variable surface reservoir age effects using ice core and marine chronostratigraphic tie points [*Skinner et al.*, 2010]. Age models of the other Atlantic cores used in this study were published in previous studies (see references in Table 1). They all rely on ^{14}C dates, with corrections in case of variable reservoir ages for cores located north of 40°N [*Waelbroeck et al.*, 2001] and south of 40°S [*Skinner et al.*, 2010]. In the latter cases, the average uncertainty on the final calendar age over the 10–21 ka time interval is larger, ranging from 450 to 950 when a correction for variable reservoir age has to be applied, whereas it is comprised between 270 and 410 years for cores that do not necessitate variable reservoir age corrections (Table S1 in the auxiliary material).[1]

2.3. Isotopic Analyses

[7] Benthic foraminifera *Cibicides kullenbergi* were picked in the >150 μm fraction. *C. kullenbergi* $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios ($\delta^{18}O$ and $\delta^{13}C$ respectively, expressed in ‰ versus Vienna Pee-Dee Belemnite (VPDB)) were measured at LSCE (Gif-sur-Yvette) on Finnigan Δ+ and Elementar Isoprime mass spectrometers. VPDB is defined with respect to NBS-19 calcite standard ($\delta^{18}O$ = −2.20‰ and $\delta^{13}C$ = +1.95‰) [*Coplen*, 1988]. The mean external reproducibility (1σ) of carbonate standards is ±0.05‰ for $\delta^{18}O$ and ±0.03‰ for $\delta^{13}C$; measured NBS-18 $\delta^{18}O$ is −23.2 ± 0.2‰ VPDB and $\delta^{13}C$ is −5.0 ± 0.1‰ VPDB.

[8] The $\delta^{13}C$ in the epifaunal benthic foraminifer *Cibicides* has been shown to record the bottom water dissolved inorganic carbon (DIC) $\delta^{13}C$ [*Duplessy et al.*, 1984]. Deep water $\delta^{13}C$ values reflect the physical mixing of different water masses, their initial ^{13}C content, controlled by biological processes in the surface ocean and air-sea CO_2 exchanges [*Lynch-Stieglitz et al.*, 1995], and the remineralization at depth of ^{13}C-depleted organic matter raining from the surface

[1]Auxiliary materials are available in the HTML. doi:10.1029/2010PA002007.

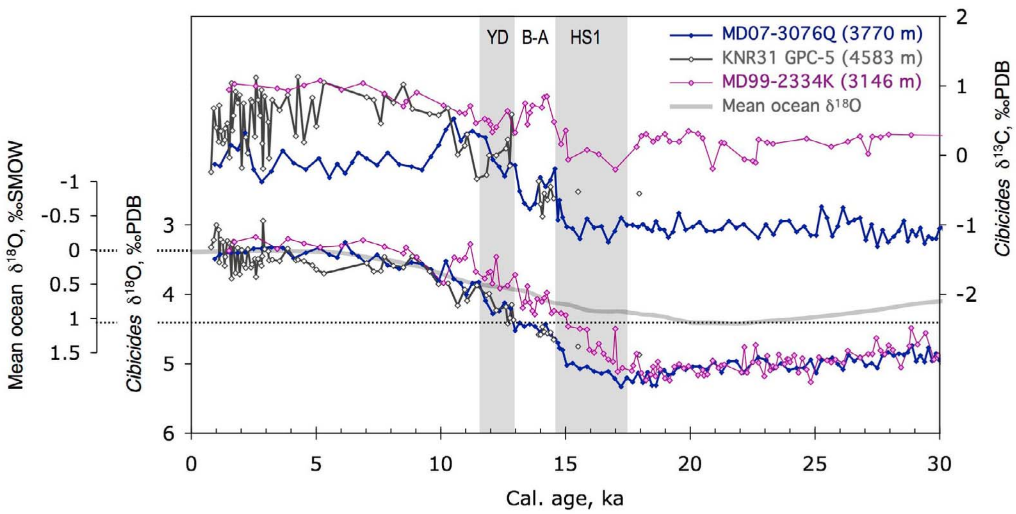

Figure 3. Comparison of South Atlantic core MD07-3076 and North Atlantic core KNR31 GPC-5 and MD99-2334K isotopic records. Core MD07-3076, KNR31 GPC-5, and MD99-2334K benthic δ^{18}O records from Figure 2 and mean ocean δ^{18}O versus calendar age are shown on the left axis. Ice-volume equivalent sea level [*Lambeck and Chappell*, 2001] has been translated into mean ocean δ^{18}O by multiplying ice-volume equivalent sea level by 1.05‰/145 m. Core KNR31 GPC-5 published *Cibicides* δ^{18}O values have been further corrected by +0.4‰ to allow core top *Cibicides* δ^{18}O values to be in isotopic equilibrium with the World Ocean Atlas bottom temperature at that site (2.3 ± 0.1°C [*Locarnini et al.*, 2006]). Core MD07-3076, KNR31 GPC-5, and MD99-2334K *Cibicides* δ^{13}C records versus calendar age are shown on the right axis.

[*Kroopnick*, 1985]. A decrease in *Cibicides* δ^{13}C thus generally corresponds to an increase in bottom water nutrient content. As nutrient content largely follows water mass structure and circulation in the modern ocean, *Cibicides* δ^{13}C can be used to trace water masses [*Duplessy et al.*, 1988]. A decrease in *Cibicides* δ^{13}C may thus be interpreted as a decrease in bottom water ventilation, and conversely. [*Mackensen et al.*, 1993] have proposed that increased seasonally variable productivity driven rain rates of organic matter to the ocean floor produce low *Cibicides* δ^{13}C values in the South Atlantic. However, the fact that enhanced glacial-interglacial benthic δ^{13}C amplitudes are observed across the whole South Atlantic, independently of sedimentary environments, primary productivity regimes, or chosen benthic foraminifera species, has been used to argue that extremely low glacial benthic δ^{13}C values in the Southern Ocean reflect changes in ocean circulation [*Hodell et al.*, 2003; *Mackensen et al.*, 2001; *Michel et al.*, 1995; *Ninnemann and Charles*, 2002].

3. Asynchrony Between North and South Atlantic Benthic δ^{18}O Signals Below 3000 m Depth

3.1. Observed Phase Shifts

[9] In order to define the timing of the beginning of the last glacial-interglacial transition in deep benthic δ^{18}O signals, we computed the average benthic δ^{18}O over the period 20–17 calendar kyr B.P. (ka) in cores located below 3000 m and determined at what time benthic δ^{18}O became significantly lighter than its 20–17 ka value (i.e., lighter than the mean 20–17 ka value minus 1 standard deviation). Applying this approach to the two high-resolution continuous benthic δ^{18}O records of Figure 2 yields an onset of the last glacial-interglacial transition at 17.0 ka (±0.4 kyr) in North Atlantic core MD99-2334K, and at 16.0 ka (±1.4 kyr) in South Atlantic core MD07-3076. Although these two dates are not significantly different due to relatively large dating uncertainties induced by the uncertainty on South Atlantic surface reservoir age correction [*Skinner et al.*, 2010], the two records differ strongly in the subsequent evolution of the benthic δ^{18}O signal: benthic δ^{18}O in Iberian Margin core MD99-2334K decreases by ~0.6‰ from 17 to 15 ka, whereas the decrease in South Atlantic core MD07-3076 benthic δ^{18}O is of only ~0.2‰ over the same time interval, before markedly accelerating at about 14.9 ka (Figure 2 and 3). A similar marked decrease in benthic δ^{18}O has been dated at about 15.8 ka in core RC11-83 (41°36′S, 9°48′E, 4718 m) [*Charles et al.*, 1996]. The younger age of our record results from the increase in surface water reservoir age over that period, as interpolated by *Skinner et al.* [2010]. However, significant uncertainties in the magnitude and variability of surface reservoir ages both in MD07-3076 and RC11-83 remain, in particular over the 17–13 ka time interval. It is thus not possible at present to determine if phase shifts exist between the different deep South Atlantic benthic δ^{18}O records, or if the discrepancies instead represent poorly constrained surface reservoir age variability across the South Atlantic. These issues will only be resolved in future through concerted efforts to further constrain surface reservoir age variability in the region.

[10] Core MD99-2334K 0.6‰ benthic δ^{18}O decrease is much larger than the change in mean ocean δ^{18}O resulting from sea level rise: sea level had risen by only one third of its total glacial-Holocene rise by 15 ka [e.g., *Lambeck and Chappell*, 2001], which corresponds to a decrease in mean ocean δ^{18}O of about 0.3‰ [*Adkins et al.*, 2002; *Schrag et al.*, 1996] (Figure 3). This early decrease in benthic δ^{18}O at

~3150 m depth on the Iberian Margin therefore reflects changes in this site bottom water temperature and/or δ^{18}O. In particular, nonuniform mixing of meltwater could induce large local water δ^{18}O signals.

[11] In contrast to the situation prevailing in Iberian Margin core MD99-2334K, the benthic δ^{18}O record of South Atlantic core MD07-3076Q does not indicate any large change in bottom water properties at that site until the HS1/B-A transition (Figures 2 and 3). Changes in bottom water δ^{18}O arising from changes in sea level might have accounted for a portion of core MD07-3076Q *Cibicides* δ^{18}O signal. The heaviest *Cibicides* δ^{18}O values likely reflect very cold and salty waters, in agreement with pore fluid measurements in ODP core 1093 located at ~3630 m water depth in the South Atlantic, a few degrees South and East from site MD07-3076Q [*Adkins et al.*, 2002]. The slight decrease in core MD07-3076 *Cibicides* δ^{18}O from ~17 to ~15 ka could result from a slight bottom water warming and/or water δ^{18}O decrease that could be transferred from Antarctic surface waters to depth by deep water formation around the Antarctic continent in response to the deglacial warming in the high southern latitudes, as Antarctic ice and sub-Antarctic SST records indicate that surface warming began at ~18 ka in high southern latitudes and was already midway to its Holocene values by ~15 ka [*Barker et al.*, 2009; *Calvo et al.*, 2007; *Lamy et al.*, 2004; *Lemieux-Dudon et al.*, 2010; *Skinner et al.*, 2010].

[12] Although we lack well-dated continuous high-resolution records from North Atlantic cores located below 3200 m, the available records suggest that the onset of the glacial-interglacial decrease in benthic δ^{18}O occurred first at ~3000 m and later below (Figure 2), confirming the results of *Labeyrie et al.* [2005]. A later onset of the benthic δ^{18}O deglacial transition in deeper cores than in shallower cores is consistent with a reduction in advection below ~3000 m depth of the surface light δ^{18}O meltwater signal. Such a reduction is in turn consistent with the reduction in the rate of export of northern sourced deep waters deduced from North Atlantic Pa/Th records during Henrich Stadial 1 (HS1) [*Gherardi et al.*, 2009].

3.2. Deglacial Circulation Changes in the Atlantic Ocean Below 3000 m

[13] In this section, we examine the additional information contained in the dated *Cibicides* δ^{13}C records from Atlantic cores located below 3000 m depth. The occurrence of marked decreases in North Atlantic deep cores *Cibicides* δ^{13}C records indicates the presence of a poorly ventilated deep water mass in the North Atlantic during the HS1 and Younger Dryas (YD) cold intervals (Figure 3). These decreases in *Cibicides* δ^{13}C have been interpreted as the signature of the northward expansion of southern-sourced deep water (glacial version of AABW) in response to decreases in northern-sourced deep water (glacial version of NADW) formation [*Elliot et al.*, 2002; *Oppo and Lehman*, 1995; *Rickaby and Elderfield*, 2005; *Skinner and Shackleton*, 2006; *Vidal et al.*, 1997]. This interpretation is consistent with sediment Pa/Th measurements [*Gherardi et al.*, 2009; *McManus et al.*, 2004] and radiocarbon data [*Robinson et al.*, 2005; *Skinner and Shackleton*, 2004] showing that the rate of export of northern-sourced deep waters out of the North Atlantic basin decreased during HS1 and the YD.

[14] Benthic foraminifer Mg/Ca measurements conducted on Iberian Margin core MD99-2334K indicate that changes in deep water temperature paralleled changes in surface water temperature in the North Atlantic region, with colder water reaching site MD99-2334K during the HS1 and YD cold intervals than during the Bølling-Allerod (B-A) or Holocene periods [*Skinner et al.*, 2003; *Skinner and Shackleton*, 2006]. *Skinner et al.* [2003] and *Skinner and Shackleton* [2006] demonstrated that the water mass bathing site MD99-2334K during HS1 and the YD was characterized by low temperatures, δ^{18}O, DIC δ^{13}C and Δ^{14}C values, a signature they interpreted as reflecting the local dominance of southern-sourced deep water during these periods. They also showed that the 0.6‰ decrease in core MD99-2334K benthic foraminifer δ^{18}O from 17 to 15 ka is entirely explained by a decrease in bottom water δ^{18}O [*Skinner et al.*, 2003].

[15] A simple increase in the influence of southern-sourced deep water in the North Atlantic during HS1 is difficult to reconcile with our new South Atlantic record: over the time interval 17–15 ka, benthic δ^{18}O decreased by only ~0.2‰ in South Atlantic core MD07-3076, whereas it decreased by ~0.6‰ in core MD99-2334K. One might suggest that the weak decrease in South Atlantic benthic δ^{18}O could result from the counteracting impact of a ~0.6‰ decrease in deep water δ^{18}O and of a 1.5 to 2°C deep water cooling over the 17–15 ka interval. However, the temperature of Southern Ocean deep water was close to the freezing point at the LGM (18–24 ka) [*Adkins et al.*, 2002], so that it cannot have further decreased by 1.5 to 2°C over the 17–15 ka time interval. Therefore, the 0.2‰ decrease recorded in core MD07-3076 benthic δ^{18}O cannot be explained by a 0.6‰ decrease in bottom water δ^{18}O that would have been compensated by a large local cooling over that interval. We therefore suggest that cold, low-δ^{18}O, and low-δ^{13}C deep waters of northern origin have contributed to deep water composition at ~3100 m depth on the Iberian Margin over that time interval.

[16] South Atlantic core MD07-3076 *Cibicides* δ^{13}C data exhibit extremely low and relatively stable values centered on -1.06 ± 0.12‰ (1 σ, n = 64) over the 30–15 ka time interval which indicate that bottom water at this site was extremely poorly ventilated during the glacial and HS1 (Figure 3). Similar glacial *Cibicides* δ^{13}C values have been recorded in other South Atlantic cores [*Hodell et al.*, 2003; *Mackensen et al.*, 2001; *Ninnemann and Charles*, 2002], indicating that the water mass that bathed site MD07-3076 during the last glacial and HS1 occupied a large portion of the South Atlantic, extending vertically from the seafloor to ~2500 m depth south of 43°S, and equatorward to ~41°N below 4000 m depth [*Martínez-Méndez et al.*, 2009]. The presence of very poorly ventilated deep water in the South Atlantic at the end of the last glacial has recently been further supported by radiocarbon measurements on core MD07-3076 benthic foraminifera [*Skinner et al.*, 2010].

[17] The large decrease in *Cibicides* δ^{18}O recorded in core MD07-3076 site starting at ~14.9 ka is accompanied by a large increase in *Cibicides* δ^{13}C marking an abrupt increase in bottom water ventilation at site MD07-3076 (Figure 3). The onset of this *Cibicides* δ^{13}C increase is synchronous, within dating uncertainties, with the onset of the *Cibicides* δ^{13}C increase marking the HS1/B-A transition in North Atlantic cores (Figure 3). These observations suggest that

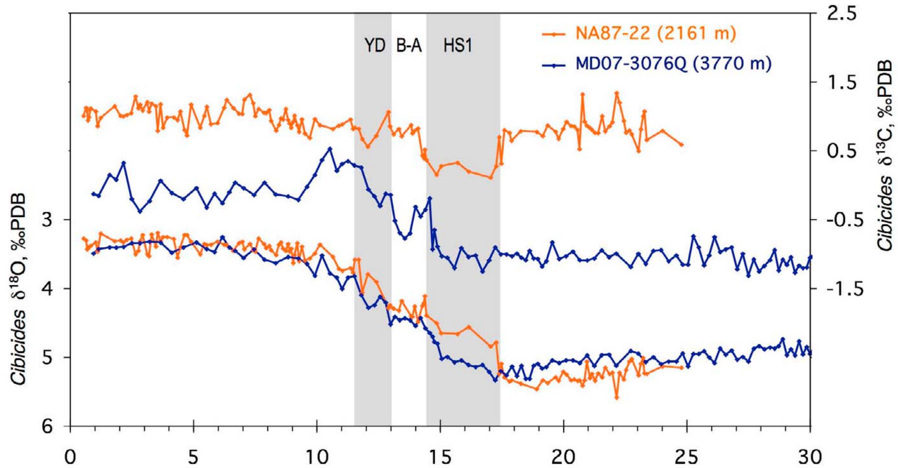

Figure 4. Comparison of South Atlantic core MD07-3076 and North Atlantic core NA87-22 *Cibicides* δ^{18}O and δ^{13}C records versus calendar age. Core NA87-22 *Cibicides* δ^{18}O has been corrected by +0.64‰ to account for its equilibrium offset with respect to *Uvigerina* [*Duplessy et al.*, 1984]. Complete reference for core NA87-22 isotopic measurements and age model are given in Table 1.

the increase in deep water formation in the North Atlantic that took place at the HS1/B-A transition led to an increase in the influence of well ventilated northern-sourced water at site MD07-3076 over a ~900 year interval, from ~14.7 to 13.8 ka. This transient increase in bottom water δ^{13}C at site MD07-3076 is coeval with a minimum in the benthic-atmosphere age offset [*Skinner et al.*, 2010], which is consistent with the arrival of younger waters to the South Atlantic site.

[18] From 13.6 ka on, bottom water δ^{13}C at site MD07-3076 progressively increased, reaching a second transient maximum between ~11.9 and ~9.9 ka. During the 13–10 ka interval, the difference between *Cibicides* δ^{13}C of northern and southern Atlantic cores was noticeably reduced, indicating that site MD07-3076 bottom waters were less isolated from ^{13}C rich water than prior to the HS1/B-A transition (Figure 3). In addition, radiocarbon ages measured on core MD07-3076 benthic foraminifera indicate that at that time, the very poorly ventilated glacial deep Southern Ocean water had substantially retreated from the Atlantic sector of the Southern Ocean at ~3800 m depth [*Skinner et al.*, 2010]. Therefore, the state of the ocean circulation that prevailed prior to the YD/Holocene transition was very different from the glacial state that prevailed prior to the HS1/B-A transition.

[19] The episode of high bottom water δ^{13}C at site MD07-3076Q over the YD and YD/Holocene transition could have multiple causes: an increase in southern-sourced deep water δ^{13}C in response to either, increased surface waters δ^{13}C due to sea ice retreat, increased deep water formation around Antarctica, a change in the location of deep water formation as suggested by [*Mackensen et al.*, 2001], or a combination of these three factors. However, part of the δ^{13}C signal could also result from an increased contribution of northern deep water masses to MD07-3076 bottom water composition, especially after the YD/Holocene transition (at ~11.7 ka) in response to increased deep water export out of the North Atlantic as evidence in Pa/Th [*Gherardi et al.*, 2009; *McManus et al.*, 2004] and other North Atlantic records.

[20] After 10 ka, core MD07-3076 *Cibicides* δ^{13}C decreased to reach Holocene values of ~−0.1‰, i.e., about 1‰ lower than North Atlantic deep cores *Cibicides* δ^{13}C values, indicating that the Atlantic Meridional Ocean Circulation (AMOC) adjusted to another circulation regime, characterized by lower ventilation in the South Atlantic at ~3800 m than in the North Atlantic at ~4500 m. This new circulation regime seems to have prevailed until ~5 ka (Figure 3). After 5 ka, the Bermuda Rise *Cibicides* δ^{13}C record indicates that ventilation progressively decreased at ~4500 m in the North Atlantic, suggesting an increasing contribution of AABW to bottom water composition at Bermuda Rise deep site.

4. Asynchrony Between Shallower and Deeper Atlantic Benthic δ^{18}O Signals

[21] Atlantic benthic δ^{18}O signals from the 1300–2200 m depth range lead Atlantic benthic δ^{18}O signals from cores located below 3000 m over the last glacial-interglacial transition (Figures 4 and 5). More specifically, a relatively large and rapid decrease in benthic δ^{18}O is observed at the beginning of HS1 (~17.5 ka), followed by a marked slow down or even a reversal in the trend of the benthic δ^{18}O signal for the shallowest cores (Figure 5). A second decrease in δ^{18}O, analogous to the one occurring at the beginning of HS1, takes place at the beginning of the YD (~13 ka) (Figure 5).

[22] The large light benthic δ^{18}O peaks seen in Norwegian Sea core ENAM93-21 during HS1 and the YD have been shown to result from brine formation in a context of iceberg discharges and meltwater input to the surface ocean [*Dokken and Jansen*, 1999]. Sea-ice formation causes an increase in surface water density by salt rejection with almost no change in water δ^{18}O. Therefore, during HS1 and the YD, brine formation transmitted the light δ^{18}O values resulting from surface water freshening to depth. In addition, because freshening of the surface ocean reduced or even stopped open-ocean convection during stadials, deep waters under-

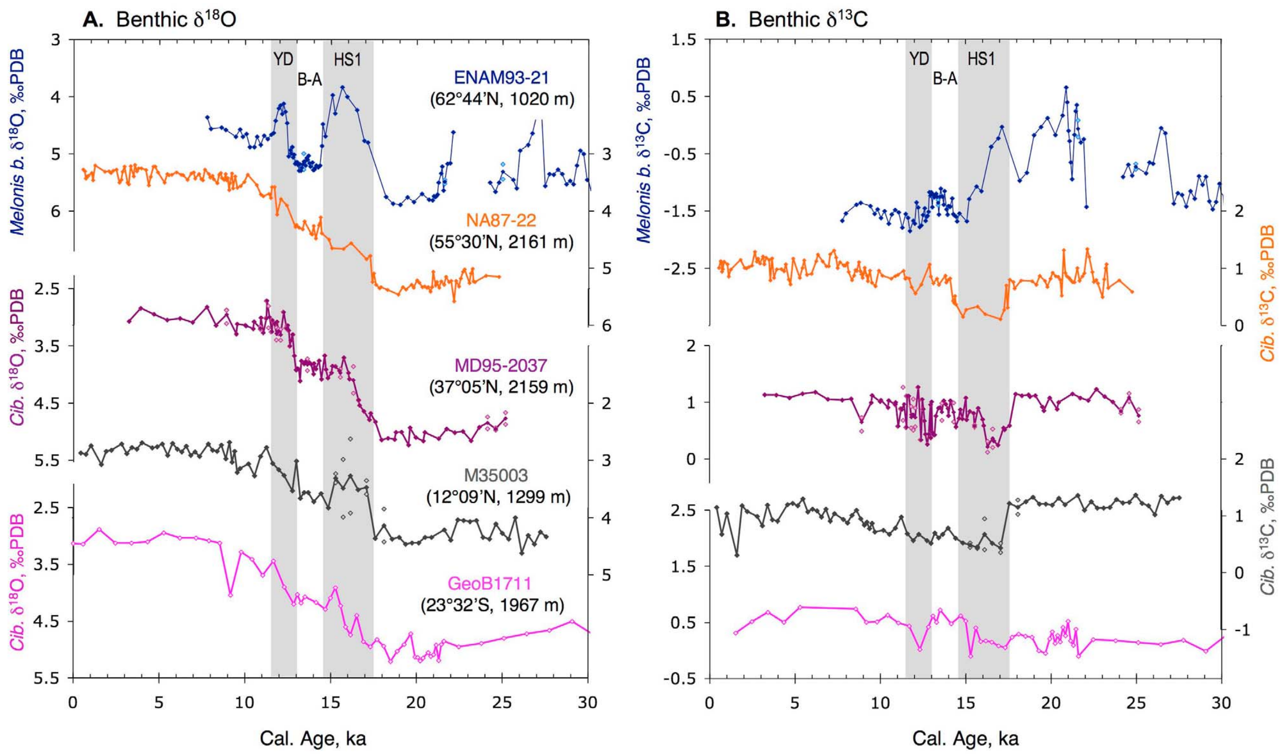

Figure 5. Norwegian Sea and Atlantic benthic isotopic records from the 1000–2200 m depth range versus calendar age. (a) Benthic δ^{18}O records. Cores NA87-22, MD95-2037, M35003, and GeoB1711 *Cibicides* δ^{18}O have been corrected by +0.64‰ to account for their equilibrium offset with respect to *Uvigerina* [*Duplessy et al.*, 1984]. Core ENAM93-21 benthic δ^{18}O signal is based on measurements on *Melonis barleanum* corrected by +0.36‰ for equilibrium offset [*Dokken and Jansen*, 1999; *Jansen et al.*, 1988]. (b) *Cibicides* δ^{13}C records for cores NA87-22, MD95-2037, M35003, and GeoB1711 and *M. barleanum* δ^{13}C record for Norwegian Sea core ENAM93-21. Light filled symbols indicate replicate measurements. Complete references for isotopic measurements and age models are given in Table 1.

neath the fresh water lid were characterized by low ^{13}C and ^{14}C concentrations [*Dokken and Jansen*, 1999; *Elliot et al.*, 2002], as shown by the large decreases in benthic δ^{13}C (Figure 5). The density of these brine-generated deep waters was high enough to create overflows across the sills [*Dokken and Jansen*, 1999] so that brine-generated deep waters entered the North Atlantic basin and formed part of North Atlantic intermediate and deep water masses during HS1 and the YD [*Meland et al.*, 2008; *Thornalley et al.*, 2010; *Waelbroeck et al.*, 2006].

[23] All North Atlantic benthic isotopic records of Figure 5 show the imprint of these brine-generated waters. These records reflect various levels of mixing of brine-generated waters with other water masses, depending on the water depth of the core and its proximal or distal location with respect to the Nordic Seas (Figure S1 in the auxiliary material).

[24] Benthic isotopic records of North Atlantic cores NA87-22 and MD95-2037, located downstream from the Norwegian deep water overflow, clearly exhibit the brine-generated water low δ^{18}O and low δ^{13}C signature during HS1 and the YD [*Labeyrie et al.*, 2005; *Waelbroeck et al.*, 2006]. Core M35003 *Cibicides* δ^{18}O decreases over HS1 and the YD have been interpreted as reflecting intermediate water warming [*Rühlemann et al.*, 2004]. According to modeling experiments, a strong reduction of the AMOC in response to freshwater input in the North Atlantic generates a temperature increase in middepth tropical Atlantic caused by a reduced inflow of cold intermediate and deep waters in conjunction with downward mixing of heat from the thermocline [*Rühlemann et al.*, 2004]. However, models used in this type of studies do not account for brine formation and cannot therefore simulate the impact of brine-generated dense waters on the salinity and temperature fields. Moreover, in the absence of additional data, it is not possible to establish whether the observed *Cibicides* δ^{18}O decreases result from warming, a decrease in water δ^{18}O, or a combination of both. The decreases observed in *Cibicides* δ^{13}C records of cores NA87-22, MD95-2037 and M35003 at 17.5 ka (Figure 5) are too large to be explained by changes in surface biological productivity alone. One could argue that the *Cibicides* δ^{13}C signal reflects a decrease in site M35003 bottom water ventilation resulting from decreased overturning circulation at ~1300 m depth during HS1. However, North Atlantic Pa/Th data suggest that there was on the contrary an increase in water export out of the North Atlantic basin above 1700 m depth during HS1 with respect to the Holocene [*Gherardi et al.*, 2009]. Therefore, available evidence supports the hypothesis that a mixture of brine-generated and overlying North Atlantic waters reached site M35003 over HS1.

[25] Because South Atlantic site GeoB1711 is located at ~1970 m depth, above the Walvis Ridge water depth, it is under the influence of the southward flowing water masses originating in the North Atlantic. Core GeoB1711 *Cibicides* δ^{18}O deglacial decrease starts around 17 ka, exhibits a plateau during the B-A period and resumes over the YD, similarly to what is observed in core MD95-2037 (Figure 5). However, core GeoB1711 *Cibicides* δ^{13}C signal is different from North Atlantic signals: comparing *Cibicides* δ^{13}C records from North Atlantic, South Atlantic and Southern Ocean cores from about 2000 m depth reveals that ventilation was higher in the North Atlantic than in the South Atlantic sites during the last glacial and Holocene. The relatively low *Cibicides* δ^{13}C values recorded in core GeoB1711 over the last glacial indicate mixing at site GeoB1711 of northern-sourced bottom waters with southern-sourced, less ventilated, CDW waters [*Waelbroeck et al.*, 2006]. Importantly, although core GeoB1711 *Cibicides* isotopic record is at lower resolution than the other records from the 1000–2200 m depth range examined in the present study, the timing of core GeoB1711 *Cibicides* δ^{18}O is nevertheless clearly similar to the latter records and distinct from that of the deeper benthic δ^{18}O records.

5. Discussion

[26] North Atlantic isotopic records from the 1300–2200 m depth range exhibit the imprint of low δ^{18}O and low δ^{13}C brine-generated waters during HS1, and during the YD to a lesser extent. The fact that there is no apparent brine-generated water signal at site M35003 during the YD could reflect the much smaller amplitude of the meltwater event and associated disruption of the AMOC during the YD than during HS1. Also, because sea level had already risen by 50% with respect to its LGM level at the beginning of the YD (whereas it had only risen by 15% at the beginning of HS1), shelf areas with water depths of less than 200 m and high potential for formation of large volumes of brine waters [*Meland et al.*, 2008] were reduced during the YD with respect to HS1.

[27] During HS1, the signature of depleted δ^{18}O water has been observed on the continental slope just south of the Iceland-Scotland Ridge, down to 3500 to 4050 m depth [*Barker et al.*, 2004]. Although the amount of brine-generated water dense enough to reach those water depths, was probably limited, the ~0.6‰ decrease in bottom water δ^{18}O observed over the course of HS1 at site MD99-2334K, located at ~3100 m on the Iberian Margin, argues for a contribution of this very dense brine-generated water to the composition of North Atlantic deep waters at ~3100 m, in addition to southern sourced deep waters during HS1.

[28] In section 3 and 4, we have described changes in water mass distribution that took place in the Atlantic over the last deglaciation, we will now briefly discuss what may have caused these changes. Available climate records versus calendar age over the last 30 kyr comprise U-Th dated sea level data, ^{14}C-dated SST records, Greenland ice core records dated by annual-layer counting [*Rasmussen et al.*, 2006; *Svensson et al.*, 2008] and Antarctic ice core records in the new glaciological chronology of *Lemieux-Dudon et al.* [2010], which is consistent for both Greenland and Antarctic ice cores. U-Th dated sea level data indicate that the onset of sea level rise over the last deglaciation took place at 19.01 ka ± 0.14 kyr [*Lambeck and Chappell*, 2001], which is earlier than any of the circulation changes described above (Figure 3). A recent study showed that Northern Hemisphere ice sheets and mountain glaciers started to retreat at 19–20 ka [*Clark et al.*, 2009], which is consistent with sea level data. Examining three of the most widely proposed mechanisms that induce changes in ice volume and feedback with the climate system (i.e., high northern latitude insolation, atmospheric CO_2, and tropical Pacific SSTs), these authors further showed that northern summer insolation was the primary mechanism for triggering the onset of Northern Hemisphere deglaciation. The earliest onset of deglacial surface warming observed in sub-Antarctic [*Barker et al.*, 2009; *Calvo et al.*, 2007; *Lamy et al.*, 2004] and some tropical [e.g., *Levi et al.*, 2007] SST records, took place at ~18 ka, in phase, within dating uncertainty, with the onset of the increase in air temperature over Antarctica and in atmospheric CO_2 (respectively at 17.9 ka ± 0.3 kyr and 17.6 ka ± 0.3 kyr [*Lemieux-Dudon et al.*, 2010]) and significantly later than the onset of sea level rise. The present study indicates that the first detectable changes in Atlantic Ocean circulation took place at the onset of the HS1 cold interval (~17.5 ka), also in phase, within dating uncertainty, with the onset of the increase in air temperature over Antarctica and atmospheric CO_2. These initial changes in Atlantic Ocean circulation occurred in the 1000–2200 m depth range, with marked decreases in North Atlantic benthic δ^{18}O, indicating the southward propagation of brine waters generated in the Nordic Seas in a context of iceberg discharges and meltwater input to the surface ocean. We suggest that these changes in Atlantic circulation occurred in response to northern ice sheet melting.

[29] In summary, benthic isotopic records from the present study indicate that the first detectable changes in Atlantic Ocean circulation took place at the onset of the HS1 cold interval and that the subsequent evolution of deep water circulation in the Atlantic Ocean was characterized by transient changes over the B-A, YD and early Holocene that lead to the Holocene ocean circulation regime starting at ~10 ka. Ocean circulation changes appear thus to have interacted with ice sheet melting, sea level rise, and changes in the global carbon cycle, to bring the climate system from a glacial to an interglacial state, in response to the initial triggering effect of the increase in northern summer insolation.

6. Conclusion

[30] We have shown that benthic isotopic records from different water depth ranges and locations within the Atlantic Ocean exhibit distinct patterns and significant differences in timing over the last deglaciation. This confirms that benthic δ^{18}O cannot be used as a global correlation tool with millennial-scale precision. However, the corollary of this finding is that independently dated benthic isotopic records provide a wealth of information on past circulation changes, where these are expressed in terms of temperature and water-δ^{18}O changes.

[31] Over the last deglaciation, such independent time scales can be established based on ^{14}C dating and correlation to absolutely dated ice core records. Using this approach, we have shown the following.

[32] 1. Changes in benthic $\delta^{18}O$ that are attributable to circulation changes first occurred in the 1000–2200 m range, with marked decreases in North Atlantic benthic $\delta^{18}O$ taking place at ~17.5 ka due to the southward propagation of brine waters generated in the Nordic Seas during HS1 (~17.5–14.5 ka).

[33] 2. The arrival of $\delta^{18}O$-depleted deglacial meltwater took place later at North Atlantic sites located below 3000 m depth than at sites located above 2200 m depth, as a result of the reduction in advection of northern-sourced deep waters during HS1.

[34] 3. The prevailing view that the low benthic and/or water $\delta^{18}O$ and DIC $\delta^{13}C$ values recorded in North Atlantic cores below 3000 m during HS1 is a signature of southern-sourced deep water alone is not compatible with our new South Atlantic deep isotopic record. In particular, northern sourced brine-generated deep water contributed to the lowering of North Atlantic water $\delta^{18}O$ at ~3100 m depth on the Iberian Margin over HS1.

[35] 4. South Atlantic waters at ~44°S and ~3800 m depth remained isolated from better ventilated northern-sourced water masses until after the HS1/B-A transition. At that time, the increase in deep water formation in the North Atlantic led to an increase in the influence of well ventilated northern-sourced water at site MD07-3076. This situation however only prevailed for ~900 years. After this episode of increased bottom water $\delta^{13}C$, site MD07-3076 fell back under the influence of low $\delta^{13}C$ Southern Ocean deep waters before its bottom water $\delta^{13}C$ progressively increased to reach a second, longer transient maximum over the YD and YD/Holocene transition. The timing of this second increase in site MD07-3076 bottom water $\delta^{13}C$ suggests that it was likely of southern origin, with a possible subsequent northern contribution attributable to the increase in NADW export that took place after the end of the YD.

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References

Adkins, J. F., et al. (2002), The salinity, temperature, and $\delta^{18}O$ of the glacial deep Ocean, *Science*, *298*, 1769–1773, doi:10.1126/science.1076252.

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.

Barker, S., et al. (2009), Interhemispheric Atlantic seesaw response during the last deglaciation, *Nature*, *457*, 1097–1102, doi:10.1038/nature07770.

Calvo, E., C. Pelejero, P. De Deckker, and G. A. Logan (2007), Antarctic deglacial pattern in a 30 kyr record of sea surface temperature offshore South Australia, *Geophys. Res. Lett.*, *34*, L13707, doi:10.1029/2007GL029937.

Charles, C. D., et al. (1996), Climate connections between the hemisphere revealed by deep sea sediment core/ice core correlations, *Earth Planet. Sci. Lett.*, *142*, 19–27, doi:10.1016/0012-821X(96)00083-0.

Clark, P. U., et al. (2009), The Last Glacial Maximum, *Science*, *325*, 710–714, doi:10.1126/science.1172873.

Coplen, T. B. (1988), Normalization of oxygen and hydrogen isotope data, *Chem. Geol.*, *72*, 293–297.

Dokken, T. M., and E. Jansen (1999), Rapid changes in the mechanism of ocean convection during the last glacial period, *Nature*, *401*, 458–461, doi:10.1038/46753.

Duplessy, J.-C., et al. (1984), ^{13}C record benthic foraminifera in the last interglacial ocean: Implications for the carbon cycle and the global deep water circulation, *Quat. Res.*, *21*, 225–243, doi:10.1016/0033-5894(84)90099-1.

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*, 343–360, doi:10.1029/PA003i003p00343.

Elliot, M., et al. (2002), Changes in North-Atlantic deep-water formation associated with the Dansgaard-Oeschger temperature oscillations (60–10 ka), *Quat. Sci. Rev.*, *21*, 1153–1165, doi:10.1016/S0277-3791(01)00137-8.

Gherardi, J.-M., L. Labeyrie, S. Nave, R. Francois, J. F. McManus, and E. Cortijo (2009), Glacial-interglacial circulation changes inferred from $^{231}Pa/^{230}Th$ sedimentary record in the North Atlantic region, *Paleoceanography*, *24*, PA2204, doi:10.1029/2008PA001696.

Govin, A. (2008), Instability of the ocean circulation: Response of northern and southern high-latitude climate to orbital variations during the Last Interglacial and last glacial inception, Ph.D. thesis, Univ. of Versailles Saint Quentin, Versailles, France.

Govin, A., E. Michel, L. Labeyrie, C. Waelbroeck, F. Dewilde, and E. Jansen (2009), Evidence for northward expansion of Antarctic Bottom Water mass in the Southern Ocean during the last glacial inception, *Paleoceanography*, *24*, PA1202, doi:10.1029/2008PA001603.

Hodell, D. A., K. A. Venz, C. D. Charles, and U. S. Ninnemann (2003), Pleistocene vertical carbon isotope and carbonate gradients in the South Atlantic sector of the Southern Ocean, *Geochem. Geophys. Geosyst.*, *4*(1), 1004, doi:10.1029/2002GC000367.

Imbrie, J., et al. (1984), The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}O$ record, in *Milankovitch and Climate*, edited by A. Berger et al., pp. 269–305, D. Reidel, Dordrecht, Netherlands.

Jansen, E., U. Bleil, R. Henrich, L. Kringstad, and B. Slettemark (1988), Paleoenvironmental changes in the Norwegian Sea and the northeast Atlantic during the last 2.8 m.y.: Deep Sea Drilling Project/Ocean Drilling Program Sites 610, 642, 643 and 644, *Paleoceanography*, *3*, 563–581, doi:10.1029/PA003i005p00563.

Keigwin, L. D., and G. A. Jones (1994), Western North Atlantic evidence for millenial-scale changes in ocean circulation and climate, *J. Geophys. Res.*, *99*, 12,397–12,410, doi:10.1029/94JC00525.

Keigwin, L. D., G. A. Jones, S. J. Lehman, and E. A. Boyle (1991), Deglacial Meltwater Discharge, North Atlantic Deep Circulation, and Abrupt Climate Change, *J. Geophys. Res.*, *96*(C9), 16,811–16,826, doi:10.1029/91JC01624.

Kroopnick, P. M. (1985), The distribution of ^{13}C of ΣCO_2 in the world oceans, *Deep Sea Res., Part A*, *32*, 57–84, doi:10.1016/0198-0149(85)90017-2.

Labeyrie, L., H. Leclaire, C. Waelbroeck, E. Cortijo, J.-C. Duplessy, L. Vidal, M. Elliot, B. Lecoat, and G. Auffret (1999), Insolation forcing and millennial scale variability of the North West Atlantic Ocean: Surface versus deep water changes, in *Mechanisms of Global Climate Change at Millennial Time Scales*, *Geophy. Monogr. Ser.*, vol. 112, edited by P. Clark, R. S. Webb, and L. D. Keigwin, pp. 77–98, AGU, Washington, D. C.

Labeyrie, L., et al. (2005), Changes in deep water hydrology during the Last Deglaciation, *C. R. Geosci.*, *337*, 919–927, doi:10.1016/j.crte.2005.05.010.

Lambeck, K., and J. Chappell (2001), Sea level change through the last glacial cycle, *Science*, *292*, 679–686, doi:10.1126/science.1059549.

Lamy, F., et al. (2004), Antarctic timing of surface water changes off Chile and Patagonian ice sheet response, *Science*, *304*, 1959–1962, doi:10.1126/science.1097863.

Lemieux-Dudon, B., et al. (2010), Consistent dating for Antarctic and Greenland ice cores, *Quat. Sci. Rev.*, *29*, 8–20, doi:10.1016/j.quascirev.2009.11.010.

Levi, C., L. Labeyrie, F. Bassinot, F. Guichard, E. Cortijo, C. Waelbroeck, N. Caillon, J. Duprat, T. de Garidel-Thoron, and H. Elderfield (2007), Low-latitude hydrological cycle and rapid climate changes during the last deglaciation, *Geochem. Geophys. Geosyst.*, *8*, Q05N12, doi:10.1029/2006GC001514.

Lisiecki, L. E., and M. E. Raymo (2005), A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}O$ records, *Paleoceanography*, *20*, PA1003, doi:10.1029/2004PA001071.

Lisiecki, L. E., and M. E. Raymo (2009), Diachronous benthic $\delta^{18}O$ responses during late Pleistocene terminations, *Paleoceanography*, *24*, PA3210, doi:10.1029/2009PA001732.

Little, M., et al. (1997), Rapid palaeoceanographic changes in the Benguela Upwelling System for the 160 000 y as indicated by abundances of planktonic foraminifer, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *130*, 135–161, doi:10.1016/S0031-0182(96)00136-8.

Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, and H. E. Garcia (2006), *World Ocean Atlas 2005*, vol. 1, *Temperature, NOAA Atlas NESDIS*, vol. 61, edited by S. Levitus, 182 pp., NOAA, Silver Spring, Md.

Lynch-Stieglitz, J., T. F. Stocker, W. S. Broecker, and R. G. Fairbanks (1995), The influence of air-sea exchange on the isotopic composition of oceanic carbon: Observations and modeling, *Global Biogeochem. Cycles*, *9*, 653–665, doi:10.1029/95GB02574.

Mackensen, A., et al. (1993), Benthic foraminiferal assemblages from the eastern South Atlantic Polar Front region between 35° and 57°S: Distribution, ecology and fossilization potential, *Mar. Micropaleontol.*, *22*, 33–69, doi:10.1016/0377-8398(93)90003-G.

Mackensen, A., et al. (2001), Late Pleistocene deep-water circulation in the subantarctic eastern Atlantic, *Global Planet. Change*, *30*, 197–229, doi:10.1016/S0921-8181(01)00102-3.

Martínez-Méndez, G., et al. (2009), Variable water column structure of the South Atlantic on glacial-interglacial time scales, *Quat. Sci. Rev.*, *28*, 3379–3387, doi:10.1016/j.quascirev.2009.09.022.

McManus, J. F., et al. (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, *428*, 834–837, doi:10.1038/nature02494.

Meland, M. Y., T. M. Dokken, E. Jansen, and K. Hevrøy (2008), Water mass properties and exchange between the Nordic seas and the northern North Atlantic during the period 23–6 ka: Benthic oxygen isotopic evidence, *Paleoceanography*, *23*, PA1210, doi:10.1029/2007PA001416.

Michel, E., L. D. Labeyrie, J.-C. Duplessy, N. Gorfti, M. Labracherie, and J.-L. Turon (1995), Could deep subantarctic convection feed the world deep basins during the Last Glacial Maximum?, *Paleoceanography*, *10*, 927–941, doi:10.1029/95PA00978.

Ninnemann, U. S., and C. D. Charles (2002), Changes in the mode of Southern Ocean circulation over the last glacial cycle revealed by foraminiferal stable isotopic variability, *Earth Planet. Sci. Lett.*, *201*, 383–396, doi:10.1016/S0012-821X(02)00708-2.

Oppo, D. W., and S. J. Lehman (1995), Suborbital timescale variability of North Atlantic deep water during the past 200,000 years, *Paleoceanography*, *10*, 901–910, doi:10.1029/95PA02089.

Orsi, A. H., et al. (1999), Circulation, mixing, and production of Antarctic Bottom Water, *Prog. Oceanogr.*, *43*, 55–109, doi:10.1016/S0079-6611(99)00004-X.

Rasmussen, S. O., et al. (2006), A new Greenland ice core chronology for the last glacial termination, *J. Geophys. Res.*, *111*, D06102, doi:10.1029/2005JD006079.

Rasmussen, T. L., et al. (1996), Circulations changes in the Faeroe-Shetland Channel correlating with cold events during the last glacial period (58–10 ky), *Geology*, *24*, 937–940, doi:10.1130/0091-7613(1996)024<0937:CCITFS>2.3.CO;2.

Rasmussen, T. L., et al. (1998), Cyclic sedimentation on the Faeroe Drift 53–10 ka BP related to climatic variations, in *Geological Processes on Continental Margins: Sedimentation, Mass-Wasting and Stability*, edited by M. S. Stoker, D. Evans, and A. Cramp, *Geol. Soc. Spec. Publ.*, 129, 255–267.

Reimer, P. J., et al. (2009), IntCal09 and Marine09 Radiocarbon Age Calibration Curves, 0–50,000 Years cal BP, *Radiocarbon*, *51*, 1111–1150.

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.

Robinson, L. F., et al. (2005), Radiocarbon variability in the western North Atlantic during the last deglaciation, *Science*, *310*, 1469–1473, doi:10.1126/science.1114832.

Rühlemann, C., et al. (1999), Warming of the tropical Atlantic Ocean and slowdown of the thermohaline circulation during the last deglaciation, *Nature*, *402*, 511–514, doi:10.1038/990069.

Rühlemann, C., S. Mulitza, G. Lohmann, A. Paul, M. Prange, and G. Wefer (2004), Intermediate depth warming in the tropical Atlantic related to weakened thermohaline circulation: Combining paleoclimate data and modeling results for the last deglaciation, *Paleoceanography*, *19*, PA1025, doi:10.1029/2003PA000948.

Schrag, D. P., et al. (1996), Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean, *Science*, *272*, 1930–1932, doi:10.1126/science.272.5270.1930.

Shackleton, N. J., and N. D. Opdyke (1973), Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperature and ice volume on a 105 year and 106 year scale, *Quat. Res.*, *3*, 39–55, doi:10.1016/0033-5894(73)90052-5.

Shackleton, N. J., M. A. Hall, and E. Vincent (2000), Phase relationships between millennial-scale events 64,000–24,000 years ago, *Paleoceanography*, *15*, 565–569, doi:10.1029/2000PA000513.

Siani, G., et al. (2000), Radiocarbon reservoir ages in the Mediterranean Sea and Black Sea, *Radiocarbon*, *42*, 271–280.

Sikes, E. L., et al. (2000), Old radiocarbon ages in the southwest Pacific Ocean during the last glacial period and deglaciation, *Nature*, *405*, 555–559, doi:10.1038/35014581.

Skinner, L. C., and N. J. Shackleton (2004), Rapid transient changes in northeast Atlantic deep water ventilation age across Termination I, *Paleoceanography*, *19*, PA2005, doi:10.1029/2003PA000983.

Skinner, L. C., and N. J. Shackleton (2005), An Atlantic lead over Pacific deep-water change across Termination I: Implications for the application of the marine isotope stage stratigraphy, *Quat. Sci. Rev.*, *24*, 571–580, doi:10.1016/j.quascirev.2004.11.008.

Skinner, L. C., and N. J. Shackleton (2006), Deconstructing Terminations I and II: Revisiting the glacioeustatic paradigm based on deep-water temperature estimates, *Quat. Sci. Rev.*, *25*, 3312–3321, doi:10.1016/j.quascirev.2006.07.005.

Skinner, L. C., N. J. Shackleton, and H. Elderfield (2003), Millennial-scale variability of deep-water temperature and $\delta^{18}O_{dw}$ indicating deep-water source variations in the Northeast Atlantic, 0–34 cal. ka BP, *Geochem. Geophys. Geosyst.*, *4*(12), 1098, doi:10.1029/2003GC000585.

Skinner, L. C., et al. (2010), Ventilation of the deep Southern Ocean and deglacial CO_2 rise, *Science*, *328*, 1147–1151, doi:10.1126/science.1183627.

Svensson, A., et al. (2008), A 60 000 year Greenland stratigraphic ice core chronology, *Clim. Past*, *4*, 47–57, doi:10.5194/cp-4-47-2008.

Talley, L. D. (2008), Freshwater transport estimates and the global overturning circulation: Shallow, deep and throughflow components, *Prog. Oceanogr.*, *78*, 257–303, doi:10.1016/j.pocean.2008.05.001.

Thornalley, D. J. R., H. Elderfield, and I. N. McCave (2010), Intermediate and deep water paleoceanography of the northern North Atlantic over the past 21,000 years, *Paleoceanography*, *25*, PA1211, doi:10.1029/2009PA001833.

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, doi:10.1016/S0967-0637(99)00092-8.

Vidal, L., et al. (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, doi:10.1016/S0012-821X(96)00192-6.

Vidal, L., et al. (1999), Link between the North and South Atlantic during the Heinrich events of the last glacial period, *Clim. Dyn.*, *15*, 909–919, doi:10.1007/s003820050321.

Waelbroeck, C., et al. (2001), The timing of the last deglaciation in North Atlantic climate records, *Nature*, *412*, 724–727, doi:10.1038/35089060.

Waelbroeck, C., et al. (2002), Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records, *Quat. Sci. Rev.*, *21*, 295–305, doi:10.1016/S0277-3791(01)00101-9.

Waelbroeck, C., et al. (2006), Distant origin of circulation changes in the Indian Ocean during the last deglaciation, *Earth Planet. Sci. Lett.*, *243*, 244–251, doi:10.1016/j.epsl.2005.12.031.

Waelbroeck, C., et al. (2008), Transferring radiometric dating of the Last Interglacial sea level high stand to marine and ice core records, *Earth Planet. Sci. Lett.*, *265*, 183–194, doi:10.1016/j.epsl.2007.10.006.

Zahn, R., and A. Stüber (2002), Suborbital intermediate water variability inferred from paired benthic foraminiferal Cd/Ca and $\delta^{13}C$ in the tropical West Atlantic and linking with North Atlantic climates, *Earth Planet. Sci. Lett.*, *200*, 191–205, doi:10.1016/S0012-821X(02)00613-1.

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