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Deglacial Variability of Antarctic Intermediate Water Penetration into the North Atlantic from Authigenic Neodymium Isotope Ratios

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Deglacial variability of Antarctic Intermediate Water penetration into the North Atlantic from authigenic neodymium isotope ratios

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[1] Understanding intermediate water circulation across the last deglacial is critical in assessing the role of oceanic heat transport associated with Atlantic Meridional Overturning Circulation variability across abrupt climate events. However, the links between intermediate water circulation and abrupt climate events such as the Younger Dryas (YD) and Heinrich Event 1 (H1) are still poorly constrained. Here, we reconstruct changes in Antarctic Intermediate Water (AAIW) circulation in the subtropical North Atlantic over the past 25 kyr by measuring authigenic neodymium isotope ratios in sediments from two sites in the Florida Straits. Our authigenic Nd isotope records suggest that there was little to no penetration of AAIW into the subtropical North Atlantic during the YD and H1. Variations in the northward penetration of AAIW into the Florida Straits documented in our authigenic Nd isotope record are synchronous with multiple climatic archives, including the Greenland ice core $\delta^{18}O$ record, the Cariaco Basin atmosphere $\Delta^{14}C$ reconstruction, the Bermuda Rise sedimentary Pa/Th record, and nutrient and stable isotope data from the tropical North Atlantic. The synchroneity of our Nd records with multiple climatic archives suggests a tight connection between AAIW variability and high-latitude North Atlantic climate change.

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1. Introduction

[2] The role of Atlantic Meridional Overturning Circulation (AMOC) in abrupt climate change has been a subject of intense interest [*Broecker*, 1994; *Broecker and Denton*, 1989; *Gutjahr et al.*, 2010; *Lynch-Stieglitz et al.*, 2011; *Piotrowski et al.*, 2004]. Today, deep water forming in the high-latitude North Atlantic is compensated for by northward flowing intermediate and surface waters originating in the Southern Ocean. This northward flowing upper-limb of AMOC transports 1.22–1.3 $\times$ 10^{15} W of heat to the North Atlantic region and thus plays a crucial role in the global heat balance [*Larsen*, 1992; *Talley*, 2008]. During the Younger Dryas (YD) and Heinrich Event 1 (H1) cold periods of the last deglaciation, meltwater and iceberg discharges into the North Atlantic are thought to have been coincident with a significant reduction of AMOC [*Broecker*, 1994; *Broecker and Denton*, 1989] as well as a warming of the Southern Ocean [*Blunier and Brook*, 2001; *Blunier et al.*, 1998]. Geochemical evidence shows that radiocarbon-poor and nutrient-rich southern-sourced deep water replaced radiocarbon-rich and nutrient-poor North Atlantic Deep Water (NADW) below ~2 km during the YD [*Boyle and Keigwin*, 1987; *Keigwin*, 2004; *Sarnthein et al.*, 1994]. Similar ocean circulation conditions are also implied for the H1 cold period [*McManus et al.*, 1999, 2004]. Furthermore, nutrient-poor Glacial North Atlantic Intermediate Water filled the North Atlantic above ~2 km during both cold events [*Keigwin*, 2004; *Marchitto et al.*, 1998; *Sarnthein et al.*, 1994].

[3] Abrupt changes in the northward flow of AAIW associated with AMOC reduction have also been hypothesized [*Came et al.*, 2007; *Marchitto et al.*, 1998; *Pahnke et al.*, 2008; *Rickaby and Elderfield*, 2005; *Zahn and Stüber*, 2002], suggesting a potential connection between the Southern Ocean and high-latitude North Atlantic climate change. However, controversy persists as to whether the northward flow of AAIW is waxing or waning during abrupt cold events. One school maintains that there is an increase in the northward penetration of AAIW associated with weaker AMOC during both the YD and H1 cold events [*Pahnke et al.*, 2008; *Rickaby and Elderfield*, 2005; *Thornalley et al.*, 2011; *Zahn and Stüber*, 2002]. Others come to the opposite conclusion, namely that there is a weakening of AAIW at least during one of the deglacial events (YD) [*Came*

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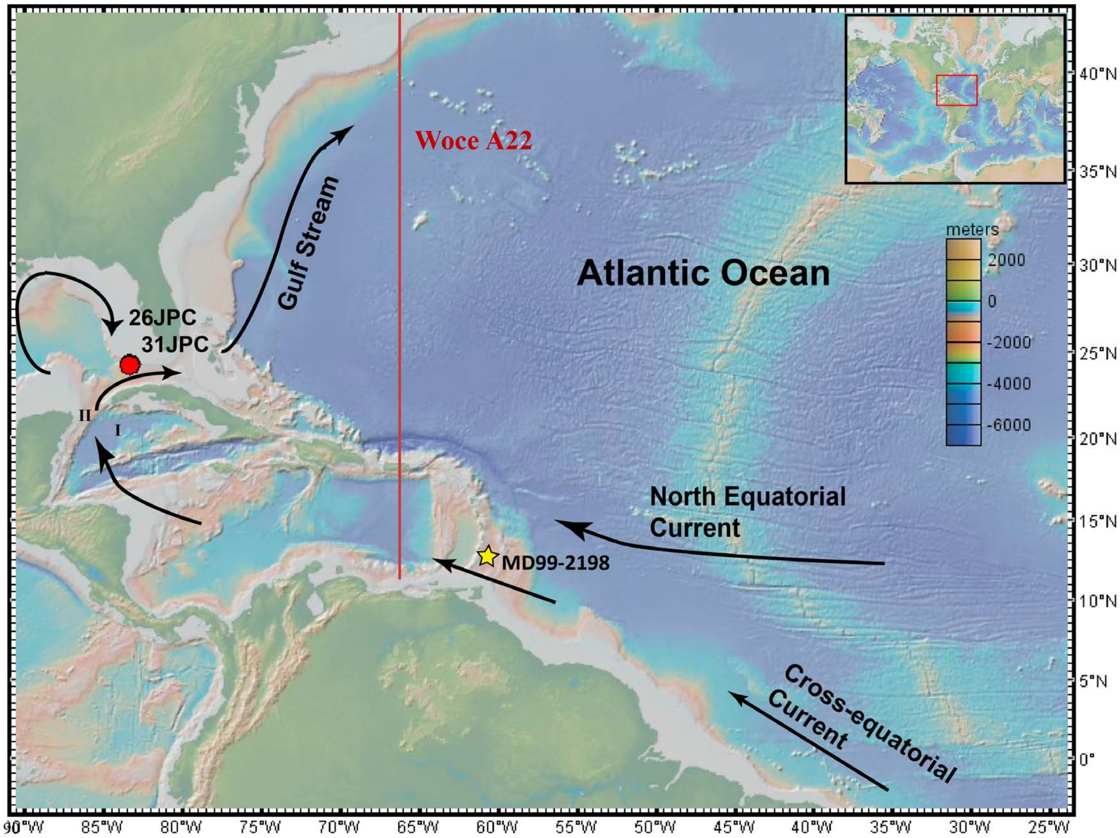

Figure 1. Map of the North Atlantic showing the locations of KNR166-2-26JPC (24°19.62′N, 83°15.14′W, 546 m) and KNR166-2-31JPC (24°13.18′N, 83°17.75′W, 751 m) in the Florida Straits (this study; http://www.geomapapp.org). Also shown is the location of MD99–2198 in the Tobago Basin [*Pahnke et al.*, 2008]. Superimposed is WOCE line A22 (red line, stations not shown) along 66°N. Symbols in this figure: I. Cayman Basin; II. Yucatan Channel.

et al., 2008]. In an attempt to shed light on the controversy, this study reconstructs water mass source changes across the last deglacial using authigenic Nd isotope records from Florida Straits sediments.

[4] Authigenic Nd isotope ratios (ε_{Nd}) are used as a reliable quasi-conservative water mass tracer [*Frank*, 2002]. Here, ε_{Nd} is defined as $\varepsilon_{Nd} = 10^4 \times [(^{143}Nd/^{144}Nd)_{sample}/(^{143}Nd/^{144}Nd)_{CHUR} - 1]$, where $(^{143}Nd/^{144}Nd)_{CHUR} = 0.512638$, and is the $^{143}Nd/^{144}Nd$ value of the bulk earth. These ratios are not fractionated by biological or chemical processes in the ocean. Using a proxy not prone to changes in biological processes or nutrient cycling enables us to unambiguously reconstruct the relationship between abrupt climate change and ocean circulation. Because Nd has a relatively short residence time (600–2000 years) [*Frank*, 2002] that is on the order of the global ocean circulation time, different oceanic water masses obtain their specific ε_{Nd} signatures from various continental weathering sources. For example, NADW preserves a distinct ε_{Nd} value of −13.5 [*Piepgras and Wasserburg*, 1980, 1987] from old continental materials surrounding the high-latitude North Atlantic, while Pacific deep and intermediate water ε_{Nd} values of −2 to −4 reflect young volcanic material supplied from the circum-Pacific region [*Piepgras and Wasserburg*, 1980; *Piepgras and Jacobsen*, 1988]. Water masses originating in the Southern Ocean (Antarctic Bottom water (AABW), Circumpolar Deep Water (CDW) and AAIW) are mixtures of both North Atlantic and Pacific waters, and thus have intermediate ε_{Nd} values of −6 to −9 [*Jeandel*, 1993]. Therefore, changes in the Nd isotope ratios in a water mass represent simple mixing between different water masses, unless additional Nd is added to the water transport path (e.g., scavenging or boundary exchange). In this study, we have reconstructed the deglacial variability of AAIW penetration into the subtropical North Atlantic by measuring authigenic Nd isotope ratios in sediments from two cores recovered from the Florida Straits.

2. Oceanographic Setting of the Study Area

[5] We have analyzed two cores, KNR166-2-26JPC (24°19.62′N, 83°15.14′W, 546 m) and KNR166-2-31JPC (24°13.18′N, 83°17.75′W, 751 m), recovered from the Florida Straits (Figure 1). Both cores 26JPC and 31JPC are located within intermediate-depth waters of the Florida Current, which under modern conditions represents a mixture of recirculated North Atlantic subtropical gyre water (17 Sv) and southern component waters (13 Sv) [*Schmitz and Richardson*, 1991; *Schmitz and McCartney*, 1993].

[6] Both North Atlantic subtropical gyre water and southern component water enter the Caribbean and flow northward toward the Florida Straits. According to *Wüst* [1964] and *Kameo et al.* [2004], AAIW in the southern Caribbean is

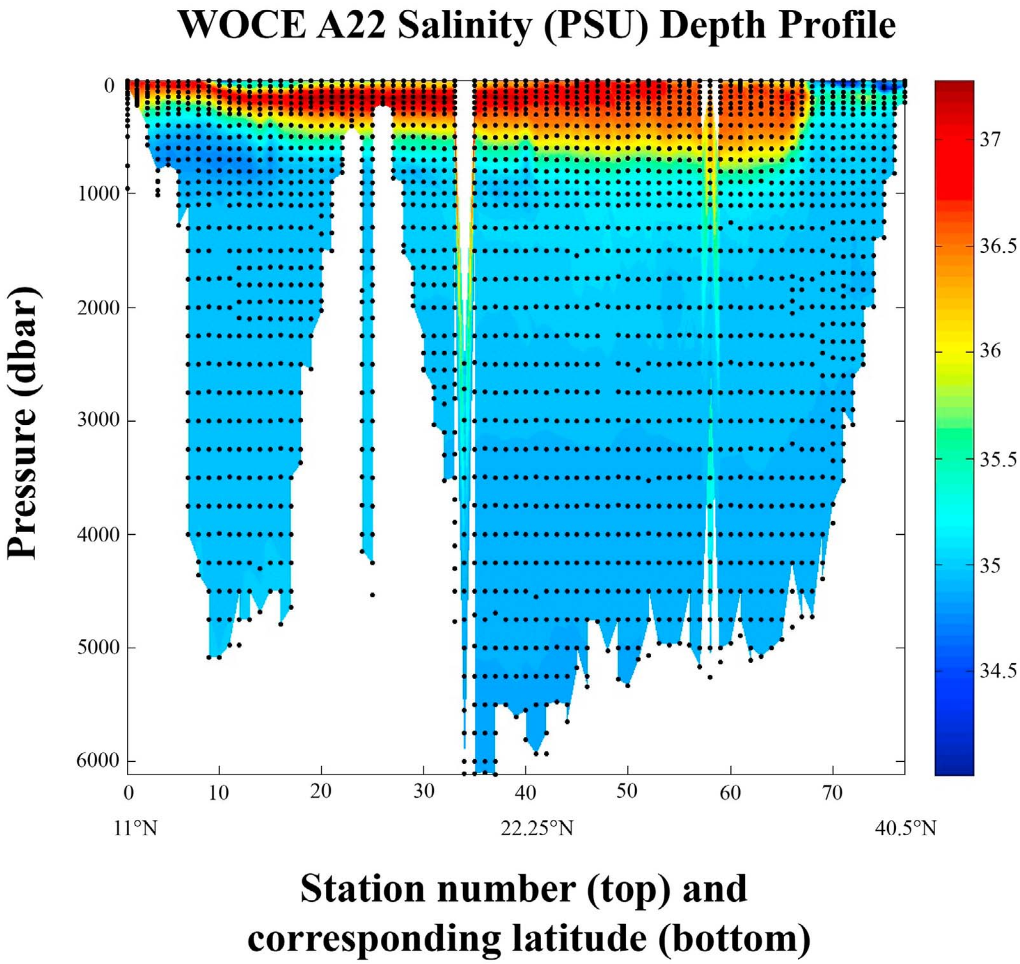

Figure 2. Salinity depth profile along WOCE line A22. Superimposed are CTD samples (dark circles).

identified by a salinity minimum at a depth ranging from 550 to 880 m. In the eastern Caribbean, this salinity minimum zone (salinities < 34.9) also exists (Figures 1 and 2, southern section of WOCE line A22) at intermediate depths, suggesting the penetration of AAIW into this region. The clearest manifestation of the low-salinity waters occurs at the southern end of WOCE A22 (∼11°N) between depths of ∼500 to 1100 m. The salinity minimum zone is thinner and its core is slightly deeper further north along WOCE A22. At ∼20°N, the core of AAIW occurs at a depth about 900 m, and ranges from 700 m to 1100 m. In the Cayman Basin and Yucatan Channel (Figure 1), the salinity minimum zone (i.e., AAIW) is found at depths between 585 m and 904 m [*Wüst*, 1964]. In the Florida Straits, the salinity minimum zone occurs below 400 m [*Schmitz and Richardson*, 1991] below which ∼80% of the Florida Current transport is of South Atlantic origin [*Schmitz and Richardson*, 1991]. Here, we interpret this water mass originating in the South Atlantic as AAIW.

3. Materials and Method

[7] The age model for 26JPC [*Lynch-Stieglitz et al.*, 2011] was recently updated using Calib 6.0 (M. Stuiver et al., Calib calibration program, version 6.0, 2011) and a standard marine reservoir correction of 400 years [*Schmidt and Lynch-Stieglitz*, 2011]. The same calibration technique was also applied to update the original age model for 31JPC [*Came et al.*, 2008] in this study. Sedimentation rates for both cores are much higher during the deglaciation than during either the last glacial maximum (LGM) or the Holocene. The shallower (26JPC) and deeper (31JPC) cores have average sedimentation rates of 68 cm·kyr^{-1} and 11 cm·kyr^{-1} during the last deglaciation (10–19 kyr), respectively. These high sedimentation rates allow high temporal resolution (sub-century scale) Nd isotope records. Sedimentation rates during the YD in 26JPC are extremely higher and reach an average of 190 cm·kyr^{-1}. This increase in sedimentation rate is not accompanied by changes in sediment composition or grain size. The effect that such changes in sedimentation rate have on ε_{Nd} leachate values is discussed further in section 4.1.

[8] Unfortunately, we have less confidence in the chronology of the abrupt events in core 31JPC. This unreliability stems from radiocarbon age reversals found in the core between 96 and 150 cm [*Came et al.*, 2008], corresponding to the period from 12.5 to 16.5 kyr (Figures 3 and 5). Specifically, within this interval for 31JPC, two age control points (one at 112 cm and the other at 128 cm) were excluded in our new age model (Figure 3) because of these radiocarbon age reversals. Although an isolated interval in 26JPC corresponding to the end of the YD from 11.5 to 11.95 kyr was also found to contain radiocarbon age reversals [*Lynch-Stieglitz et al.*, 2011], the remainder of the deglacial sediments within core 26JPC

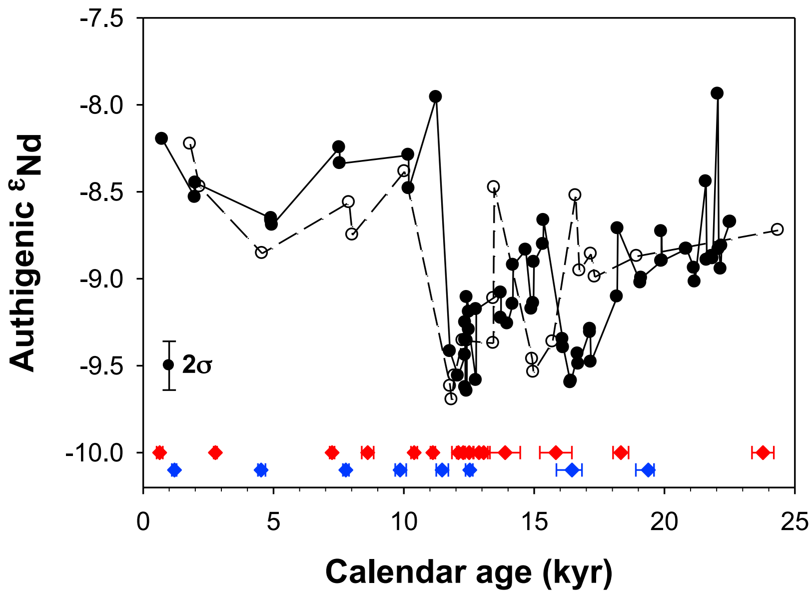

Figure 3. Authigenic ε_{Nd} record for 26JPC (solid circles) and 31 JPC (open circles). External error of ±0.13 (2σ) is indicated. The age model for both cores is based on AMS radiocarbon dates from previous studies [*Came et al.*, 2008; *Lynch-Stieglitz et al.*, 2011] and converted to calendar dates using Calib 6.0 and Marine09 for 26JPC [*Schmidt and Lynch-Stieglitz*, 2011] and 31JPC (this study). Age control points are indicated for both 26JPC (red diamonds) and 31JPC (blue diamonds). Offset of the timing of the YD and H1 between 26JPC and 31JPC may be a result of the uncertainty in the 31JPC age model. Refer to text for further discussion.

(spanning the YD through the LGM) are thought to be undisturbed and have been previously used to determine the timing of Florida Current transport and hydrologic cycle variability across the deglaciation [*Lynch-Stieglitz et al.*, 2011; *Schmidt and Lynch-Stieglitz*, 2011]. None of our ε_{Nd} data from 26JPC fall within this late deglacial age-reversal-interval of 450 years.

[9] Nd isotopes were analyzed in 63 sediment samples from core 26JPC and 22 sediment samples from core 31JPC, respectively. Seawater Nd is incorporated into authigenic Fe-Mn oxyhydroxide coatings of ocean sediments, whose isotopic composition reflects that of the ambient seawater within which the coatings form [*Goldstein and Hemming*, 2003]. Hence, authigenic Nd isotope ratios measured in sediments from intermediate water depths in the Florida Straits should vary between ε_{Nd} values for northern- and southern-sourced waters, reflecting varying inputs of recirculated North Atlantic gyre water and AAIW, respectively. Fe-Mn oxyhydroxides were extracted for Nd isotope analysis following a sequential extraction procedure modified from that used by others [*Chester and Hughes*, 1967; *Gutjahr et al.*, 2007; *Piotrowski et al.*, 2004]. Briefly, carbonates were removed by digesting bulk sediment samples multiple times with 10% acetic acid buffered with sodium acetate. Then 0.02M hydroxylamine hydrochloride with 25% acetic acid (v/v) was added to leach the authigenic Fe-Mn oxyhydroxides for two hours. This leachate was later converted to a nitrate matrix before column chemistry. For Nd isotope analysis of the detrital fraction of the sediments, samples were further leached with hydroxylamine hydrochloride overnight. 2M KOH was then added to remove biogenic opal, if any. Finally, the residual fraction of the sediments was digested in 2:5 (v/v) nitric acid and hydrofluoric acid, followed by aqua regia. The REEs in both the authigenic and detrital samples were separated using RE spec™ column chemistry and the Nd was separated from the REEs using Ln spec™ column chemistry. Nd isotope ratios were analyzed as Nd^+ by Thermal Ionization Mass Spectrometry (TIMS) at Texas A&M University. $^{143}Nd/^{144}Nd$ ratios were corrected for any Sm interference on mass 144 by monitoring mass 147. $^{143}Nd/^{144}Nd$ ratios were normalized to $^{146}Nd/^{144}Nd = 0.7219$. Reproducibility of Nd standard JNdi-1 during the course of this study gave a mean value of $^{143}Nd/^{144}Nd = 0.5121053 \pm 0.0000063$ (2σ, n = 25).

4. Results and Discussions

4.1. Timing of Changes in the Northward Penetration of AAIW

[10] Located approximately 13 km apart at different depths, 26JPC (546 m water depth) and 31JPC (751 m water depth) show similar ranges in authigenic ε_{Nd} values for the past 25 kyr (from -7.9 to -9.7 for 26JPC and from -8.2 to -9.7 for 31JPC; Figure 3 and Tables 1 and 2). Authigenic ε_{Nd} values determined for sediments deposited during the LGM are significantly lower than those determined for sediments deposited during the Holocene ($\sim$0.4 ε units lower for both cores). No discernible long-term trend is observed in either of these cores. Both downcore records show abrupt short-term excursions toward more negative values during the last deglaciation. The timing of each excursion is consistent with the timing of the YD and H1 cold events. Specifically, for 26JPC, more negative ε_{Nd} values are found between 16.1 and 18.2 kyr (Figure 3). Within the following 700 years and into the Bølling-Allerød (BA), ε_{Nd} values return to LGM levels. Similar to the first negative excursion, there is a second negative shift in ε_{Nd} values between 11.8

Table 1. Authigenic Fe-Mn Leachate and Detrital Fraction Nd Isotope Ratios of Core KNR166-2-26JPC

		Fe-Mn Oxyhydroxide Fraction		Detrital Fraction	
Depth (cm)	Age (ka)	$^{143}Nd/^{144}Nd$	ε_{Nd}	$^{143}Nd/^{144}Nd$	ε_{Nd}
2.5	0.71			0.512076 ± 2	−10.97 ± 0.03
3	0.74	0.512218 ± 2	−8.20 ± 0.04		
31	1.99	0.512201 ± 2	−8.53 ± 0.05		
31.5	2.01	0.512205 ± 2	−8.45 ± 0.04		
79	4.92	0.512194 ± 3	−8.65 ± 0.06	0.512170 ± 2	−9.13 ± 0.04
Duplicate		0.512194 ± 2	−8.67 ± 0.04		
79.5	4.95	0.512192 ± 3	−8.69 ± 0.06		
119	7.53	0.512215 ± 3	−8.25 ± 0.06	0.512146 ± 2	−9.60 ± 0.04
119.5	7.55	0.512211 ± 4	−8.34 ± 0.07		
207	10.18	0.512213 ± 2	−8.29 ± 0.04		
207.5	10.19	0.512203 ± 2	−8.48 ± 0.04		
300	11.25	0.512230 ± 2	−7.96 ± 0.04		
390.5	11.77	0.512155 ± 2	−9.42 ± 0.05		
443	12.08	0.512148 ± 2	−9.56 ± 0.05		
443.5	12.09			0.512093 ± 2	−10.63 ± 0.03
497	12.36	0.512154 ± 2	−9.44 ± 0.05		
497.5	12.36	0.512159 ± 5	−9.35 ± 0.11	0.512116 ± 2	−10.18 ± 0.04
Duplicate		0.512145 ± 2	−9.63 ± 0.04		
498	12.36	0.512164 ± 3	−9.25 ± 0.06		
515	12.42	0.512143 ± 2	−9.65 ± 0.04	0.512105 ± 2	−10.39 ± 0.03
515.5	12.42	0.512171 ± 2	−9.11 ± 0.05		
540.5	12.50	0.512167 ± 4	−9.19 ± 0.09		
541	12.50	0.512162 ± 6	−9.29 ± 0.12		
580.5	12.77	0.512147 ± 11	−9.58 ± 0.22	0.512095 ± 2	−10.59 ± 0.04
581	12.78	0.512168 ± 2	−9.18 ± 0.05		
645	13.73	0.512172 ± 5	−9.08 ± 0.09	0.512071 ± 2	−11.06 ± 0.05
Duplicate		0.512165 ± 2	−9.23 ± 0.05		
653	13.96			0.512011 ± 2	−12.23 ± 0.03
653.5	13.98	0.512163 ± 3	−9.26 ± 0.06		
659	14.18	0.512169 ± 4	−9.15 ± 0.07		
659.5	14.20	0.512181 ± 4	−8.92 ± 0.08		
672.5	14.68	0.512185 ± 4	−8.84 ± 0.07		
673	14.70			0.511996 ± 2	−12.53 ± 0.03
678.5	14.90	0.512168 ± 2	−9.17 ± 0.04		
680.5	14.98	0.512169 ± 2	−9.14 ± 0.04		
681	14.99	0.512181 ± 2	−8.91 ± 0.04		
690.5	15.35	0.512187 ± 5	−8.80 ± 0.11		
691	15.37	0.512194 ± 3	−8.67 ± 0.06		
709	16.10	0.512159 ± 2	−9.35 ± 0.04	0.512105 ± 3	−10.40 ± 0.06
709.5	16.12	0.512156 ± 2	−9.40 ± 0.04		
715	16.40	0.512146 ± 2	−9.60 ± 0.04	0.512062 ± 2	−11.23 ± 0.04
715.5	16.42	0.512146 ± 3	−9.59 ± 0.05		
720.5	16.67	0.512154 ± 3	−9.43 ± 0.06		
721	16.70	0.512151 ± 2	−9.49 ± 0.04		
730	17.15	0.512162 ± 4	−9.29 ± 0.07	0.512060 ± 2	−11.27 ± 0.03
Duplicate		0.512161 ± 1	−9.31 ± 0.03		
730.5	17.18	0.512152 ± 1	−9.48 ± 0.03		
750.5	18.18	0.512171 ± 2	−9.10 ± 0.04		
751	18.21	0.512191 ± 2	−8.71 ± 0.04	0.512053 ± 2	−11.40 ± 0.04
766.5	19.08	0.512175 ± 2	−9.02 ± 0.03		
767	19.11	0.512177 ± 2	−9.00 ± 0.04		
780.5	19.88	0.512182 ± 2	−8.90 ± 0.04	0.512064 ± 2	−11.19 ± 0.04
Duplicate		0.512191 ± 2	−8.73 ± 0.04		
781	19.91	0.512182 ± 3	−8.90 ± 0.05		
797	20.82	0.512185 ± 2	−8.83 ± 0.03		
797.5	20.85	0.512185 ± 2	−8.83 ± 0.04		
802.5	21.14	0.512180 ± 2	−8.94 ± 0.04	0.512059 ± 2	−11.30 ± 0.04
803	21.17	0.512176 ± 2	−9.02 ± 0.04		
810.5	21.59	0.512205 ± 2	−8.44 ± 0.03		
811	21.62	0.512182 ± 2	−8.89 ± 0.04		
814.5	21.82	0.512183 ± 4	−8.87 ± 0.07		
815	21.85	0.512183 ± 2	−8.89 ± 0.04		
818.5	22.05	0.512231 ± 4	−7.94 ± 0.08		
819	22.08	0.512185 ± 2	−8.83 ± 0.02		
820.5	22.17	0.512179 ± 2	−8.94 ± 0.05		
821	22.19	0.512186 ± 2	−8.81 ± 0.04		
826.5	22.51	0.512193 ± 2	−8.68 ± 0.04		
827	22.54	0.512193 ± 5	−8.67 ± 0.10		

Table 2. Authigenic Fe-Mn Leachate and Detrital Fraction Nd Isotope Ratios of Core KNR166-2-31JPC

		Fe-Mn Oxyhydroxide Fraction		Detrital Fraction	
Depth (cm)	Age (ka)	$^{143}Nd/^{144}Nd$	ε_{Nd}	$^{143}Nd/^{144}Nd$	ε_{Nd}
5.5	1.81	0.512216 ± 3	−8.23 ± 0.06		
8.5	2.17	0.512204 ± 2	−8.47 ± 0.04	0.512099 ± 3	−10.51 ± 0.05
28.5	4.57	0.512184 ± 2	−8.85 ± 0.05	0.512133 ± 2	−9.85 ± 0.05
64.5	7.90	0.512199 ± 2	−8.56 ± 0.05	0.512137 ± 2	−9.77 ± 0.04
65	8.03	0.512190 ± 1	−8.75 ± 0.03		
73.5	10.03	0.512208 ± 3	−8.38 ± 0.05	0.512059 ± 3	−11.30 ± 0.05
89	11.79	0.512145 ± 3	−9.62 ± 0.05		
89.5	11.84	0.512141 ± 2	−9.70 ± 0.04		
90.5	11.95	0.512148 ± 2	−9.56 ± 0.05	0.512105 ± 3	−10.40 ± 0.06
93.5	12.26	0.512158 ± 3	−9.36 ± 0.05		
95	12.42	0.512158 ± 2	−9.36 ± 0.03		
111	13.44	0.512158 ± 2	−9.37 ± 0.03		
Duplicate		0.512171 ± 3	−9.11 ± 0.07		
111.5	13.47	0.512204 ± 3	−8.48 ± 0.06	0.512056 ± 3	−11.35 ± 0.06
130.5	14.94	0.512153 ± 1	−9.46 ± 0.03		
131	14.97	0.512149 ± 2	−9.54 ± 0.03		
140.5	15.71	0.512158 ± 2	−9.36 ± 0.03	0.512073 ± 2	−11.03 ± 0.04
150.5	16.60	0.512201 ± 2	−8.52 ± 0.04		
151	16.74	0.512179 ± 2	−8.95 ± 0.04		
152.5	17.18	0.512184 ± 2	−8.86 ± 0.03		
153	17.33	0.512177 ± 3	−8.99 ± 0.06		
158.5	18.94	0.512183 ± 2	−8.87 ± 0.04		
177	24.36	0.512191 ± 2	−8.72 ± 0.03	0.512012 ± 3	−12.22 ± 0.05

and 12.8 kyr. This second abrupt decrease in ε_{Nd} values is followed by a rapid increase of ε_{Nd} values (less than 60 years) into the Holocene. A similar pattern with two abrupt negative spikes is also observed in 31JPC. The ε_{Nd} values in 31JPC are much lower between 14.9 and 15.7 kyr and between 11.8 and 12.3 kyr. As in 26JPC, there are rapid increases in ε_{Nd} values after each abrupt unradiogenic excursion.

[11] Concerns related to boundary-exchange processes are a potential caveat to tracing water mass ε_{Nd} signatures using authigenic Fe-Mn leachates from sediments along continental margins [*Lacan and Jeandel*, 2005b] such as the Florida Straits. Such processes may result in the overprinting of the true seawater Nd isotope composition with that derived from settling particles supplied by the continental margin. In order to test whether boundary exchange complicates our seawater Nd interpretations, we compared Nd isotope ratios in the detrital fraction to those in the authigenic fraction within the same sediment samples from both 26JPC and 31JPC. In both cores, the authigenic ε_{Nd} values are significantly higher than the detrital ε_{Nd} values (Figure 4a). If the Nd isotope values of marginal detrital particles were to impact ambient seawater Nd isotope signatures, there should be a good correlation between the leachate and detrital isotopic records. However, no such correlation exists for either core ($R^2 \leq 0.1$, Figures 4a and 4b). Indeed, during negative excursions in the authigenic ε_{Nd} record (YD and H1), there are positive excursions in the detrital ε_{Nd} record, the opposite of what one would expect if the detrital signature were to affect that of the authigenic component.

[12] Changes in sedimentation rate and/or grain size distribution may be important factors influencing variations in ε_{Nd} values of the Fe-Mn oxide leachate. Although there is an abrupt increase in sedimentation rate just before the start of the YD in core 26JPC, the increase does not correspond to a drastic change in sediment composition or grain size distribution. Throughout the deglaciation and Holocene, carbonate and organic C compositions vary minimally, and maintain values of about 85–90% and 0.5–1.0%, respectively (Table S1 in auxiliary material).[1] Coarse fraction (grain size greater than 63 μm) percentages in 26JPC drop from 22 to 30% during the early deglaciation and H1 to 5–10% during the rest of the deglaciation and the Holocene (Figure S1 in auxiliary material). One could argue that this change in grain size distribution may affect the sensitivity of our leaching technique. However, the timing of the changes in grain size distribution in 26JPC does not correspond to the timing of changes in sedimentation rate, or to the timing of the abrupt changes in the authigenic Nd isotope record. Furthermore, the sedimentation rate in core 31JPC is maintained at about 11 cm$\cdot$kyr^{-1} throughout the last deglaciation, and the magnitude and timing of authigenic ε_{Nd} leachate values are identical to those in 26JPC. Thus, we believe that changes in sedimentation rate or grain size distribution do not appear to affect the validity of our authigenic Nd isotope records, and we are confident that the leachate ε_{Nd} values represent true water mass Nd isotope compositions. Therefore, it is likely that the negative excursions of ε_{Nd} values in both cores reflect a reduction in the influence of a water mass that has higher ε_{Nd} values, i.e., AAIW. Most importantly, both negative excursions in both cores reach similar ε_{Nd} values, implying that the contribution of AAIW was reduced to the same extent during both the YD and H1 in the Florida Straits.

[13] During the last deglaciation, sea level rise is punctuated by episodic standstills and rapid rises. Specifically, in the Florida Straits, rapid sea level rise is found between 13.8 kyr and 14.5 kyr [*Locker et al.*, 1996]. Changes in sea level may affect the recirculation and/or depth range of AAIW in our study sites. However, the timing of rapid sea level rises

[1]Auxiliary materials are available in the HTML. doi:10.1029/2012PA002337.

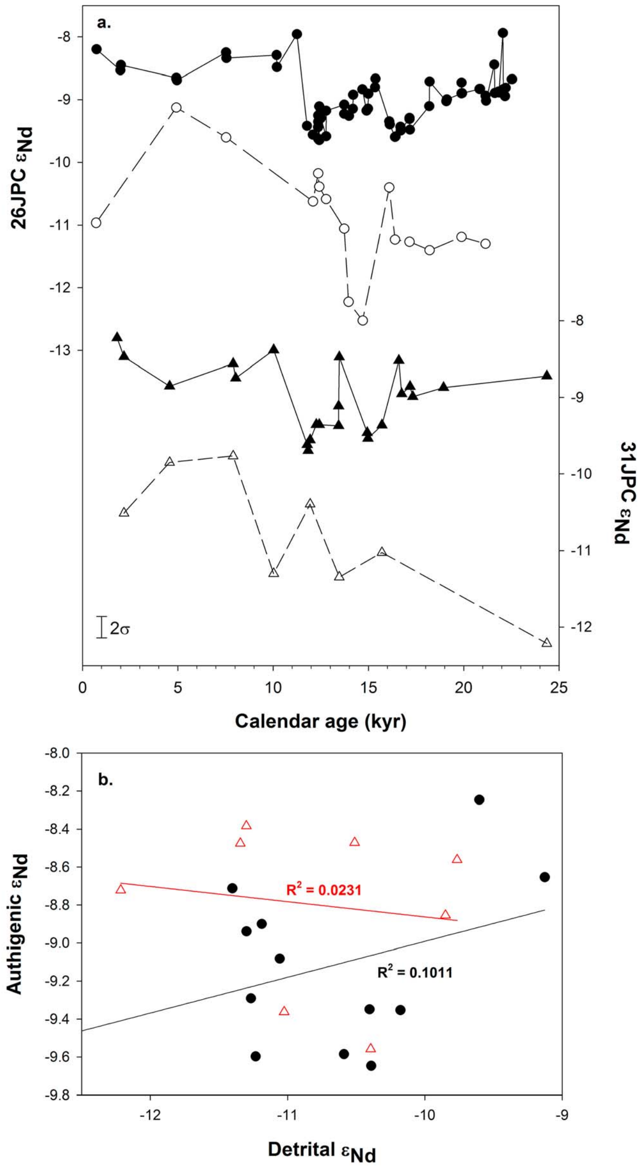

Figure 4. (a) The ε_{Nd} data for Fe-Mn leachate, i.e., authigenic, (filled symbols) and detrital (open symbols) fractions of sediment from cores 26JPC (top) and 31JPC (bottom). External error of ±0.l3 (2σ) is indicated. (b) Linear regression analysis of the ε_{Nd} values from both authigenic and detrital fractions of the sediment from 26JPC (black solid circles) and 31JPC (red open triangles). R^2 values are indicated for both 26JPC (black) and 31JPC (red).

(13.8 to 14.5 kyr) in the Florida Straits doesn't correspond to variations of our authigenic Nd isotope negative excursions (16.1 to 18.2 kyr and 11.8 to 12.8 kyr). Thus, changes in rapid sea level rises and/or recirculation may not have a significant influence on the variations in the authigenic Nd isotope record.

[14] Because we have more confidence in the deglacial age model for 26JPC, we focus on this ε_{Nd} record and its relationship to deglacial climate change (Figure 6). However, it is important to note that the difference in timing of the YD and H1 between 26JPC and 31JPC is likely due to a lack of age control points in the section of 31JPC between 12.5 to 16.5 kyr (Figure 3). Indeed, it might be possible, because of the identical patterns in authigenic ε_{Nd} between these two cores, to tie the excursions in Nd isotope ratios in 31JPC to those in 26JPC, whose age model is more tightly controlled by more radiocarbon dates.

4.2. Variations of AAIW in the Subtropical North Atlantic During the Last Deglaciation

[15] Today, NADW forms at high latitudes within the Labrador and Nordic Seas. NADW acquires its characteristic ε_{Nd} value of −13.5 by mixing of unradiogenic Labrador Seawater (ε_{Nd} = −13.9 ± 0.4) with more radiogenic overflow water masses in the Nordic Seas (Denmark Strait Overflow Water (ε_{Nd} = −8.2 ± 0.6) and Iceland-Scotland Overflow Water (ε_{Nd} = −8.4 ± 1.4)) [*Lacan and Jeandel*, 2004a, 2004b, 2005a]. Thus, changes in the location of deep water formation, or variations in terrigenous input to the high latitude North Atlantic could have a significant impact on ε_{Nd} values of the northern component water (NCW) end-member. During the YD and H1, production of NADW is believed to have been significantly reduced [*Broecker*, 1994; *Broecker and Denton*, 1989] and its glacial analog, Glacial North Atlantic Intermediate Water (GNAIW), is thought to have formed south of Iceland [*Duplessy et al.*, 1988; *Oppo and Lehman*, 1993; *Sarnthein et al.*, 1994]. The production location of GNAIW farther south in the North Atlantic likely imprinted this NCW end-member with a different ε_{Nd} value. Yet, a few studies suggest that NCW end-members (i.e., GNAIW and NADW) maintain similar ε_{Nd} values on glacial-interglacial timescales [*Abouchami et al.*, 1997; *Foster et al.*, 2007; *van de Flierdt et al.*, 2006]. However, only one of these studies focused on ε_{Nd} variations of the NCW end-members across the last deglaciation [*van de Flierdt et al.*, 2006]. The low sampling resolution in this study prevents the identification of abrupt climate events across the deglaciation, and any associated differences in NCW ε_{Nd} values. Furthermore, a recent study of sediments from the Blake Ridge [*Gutjahr et al.*, 2008] suggests that GNAIW has an ε_{Nd} value (−9.7 ± 0.4) during the LGM that is significantly different from that of modern NADW (−13.5). Unfortunately, a definitive determination of GNAIW ε_{Nd} values within the deglacial intervals of the Blake Ridge cores was not possible because of a potential problem with boundary exchange at this location [*Gutjahr et al.*, 2008].

[16] Our high-resolution Nd isotope record from 26JPC shows significantly negative excursions of ε_{Nd} values in the Florida Straits during the YD and H1 abrupt cooling events (Figure 6a). Both negative excursions have an average ε_{Nd} value of −9.4 ± 0.2, which agrees, within error, with the ε_{Nd} value of GNAIW suggested by *Gutjahr et al.* [2008]. If their interpretation of a distinct ε_{Nd} value for GNAIW (−9.7 ± 0.4) is correct, our authigenic ε_{Nd} record suggests that, during the YD and H1 events, the Florida Straits sites were bathed only by GNAIW, and that there is little to no penetration of AAIW into the subtropical North Atlantic. During the LGM, the northward penetration of AAIW is reduced, with an average ε_{Nd} value of −8.8 compared to that of −8.4 during the Holocene. This is in agreement with a previous study which showed that AAIW was still a significant contributor to the subtropical North Atlantic during the LGM [*Zahn and Stüber*, 2002].

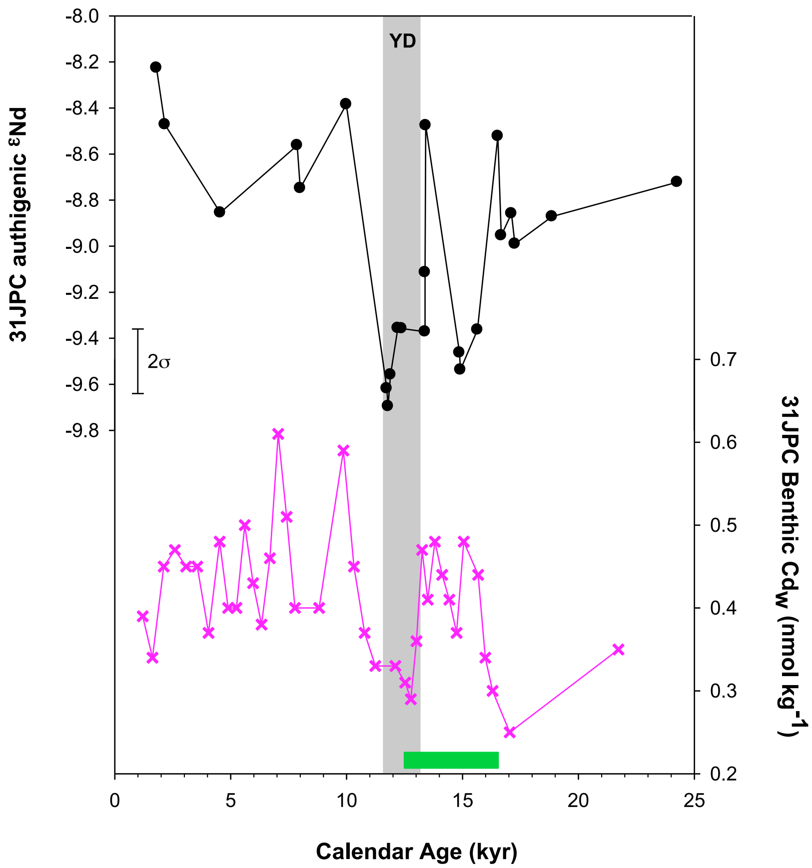

Figure 5. Comparison of authigenic ε_{Nd} record (top) and benthic Cd_w (bottom) [*Came et al.*, 2008] for 31JPC. External error for authigenic ε_{Nd} record of ± 0.13 (2σ) is indicated. Grey shading marks the YD as indicated in *Came et al.* [2008]. Green bar marks the interval containing the out of age sequence radiocarbon dates in 31JPC.

4.3. Connection Between AAIW Variability and North Atlantic Climate Change

[17] In the Florida Straits, estimates of seawater Cd values (Cd_w) were determined using benthic foraminifera from 31JPC (Figure 5) [*Came et al.*, 2008]. Significantly decreased Cd_w values during the YD were interpreted as indicating a reduction in the northward flow of nutrient-rich AAIW when AMOC was reduced [*Came et al.*, 2008]. Furthermore, higher Cd_w values in the Florida Straits during the BA warm period were interpreted as indicating an increase in the northward penetration of AAIW during this warm interval [*Came et al.*, 2008]. These conclusions are consistent with our authigenic ε_{Nd} records from both Florida Straits cores (Figures 3 and 5). However, care must be taken when interpreting nutrient proxies such as Cd in that they are sensitive to biological processes [*Boyle et al.*, 1995; *Marchitto and Broecker*, 2006] and thus are not unambiguous proxies of ocean circulation. Indeed the high-frequency variability in the Cd record [*Came et al.*, 2008] may be a reflection of this ambiguity.

[18] Changes in our authigenic ε_{Nd} record are also synchronous with other records of abrupt ocean circulation changes across the last deglaciation (Figures 6c, 6d, and 6e). A recent reconstruction of benthic foraminiferal oxygen isotope values from the margins of the Florida Straits was interpreted to record variations in Florida Current transport, most likely reflecting changes in the strength of AMOC across the deglaciation [*Lynch-Stieglitz et al.*, 2011]. Ice-volume-corrected benthic foraminiferal $\delta^{18}O_{calcite}$ values from 26JPC show an abrupt excursion toward lower values during the YD from 11.5 to 13.1 kyr (Figure 6c), which is not mirrored on the Bahama Margin of the Straits [*Lynch-Stieglitz et al.*, 2011], suggesting reduced Florida Current transport and a weaker AMOC state. Additionally, past atmospheric ^{14}C concentrations ($\Delta^{14}C$) reconstructed from three sediment cores in the Cariaco Basin recorded abrupt increases in atmospheric $\Delta^{14}C$ during the YD and H1 (Figure 6d), suggesting reduced ventilation of the deep ocean [*Hughen et al.*, 2000, 2004, 2006], consistent with a reduction in AMOC at these times. However, as pointed out by a recent study [*Southon et al.*, 2012], caution should be used when interpreting the increase in $\Delta^{14}C$ during the YD and H1 in the Cariaco Basin. Another record that is thought to represent changes in AMOC during the deglaciation is the sedimentary $^{231}Pa/^{230}Th$ record of *McManus et al.* [2004]. In sediments from the deep Bermuda Rise, elevated sedimentary $^{231}Pa/^{230}Th$ ratios (close to 0.093; Figure 6e) during H1 have been interpreted as recording a complete shutdown of

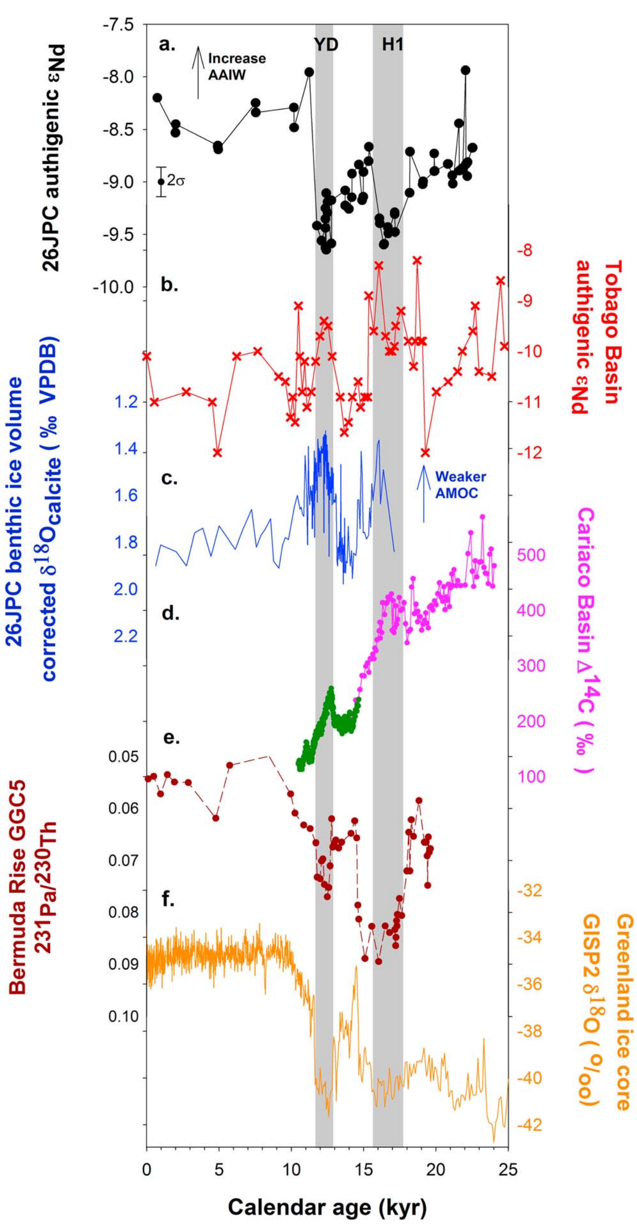

Figure 6. Comparison of the 26JPC authigenic Nd isotope record with multiple climatic records. (a) 26JPC authigenic ε_{Nd} (black circles) in this study. External error of ± 0.13 (2σ) is indicated. Arrow indicates the direction of increasing northward penetration of the AAIW. (b) Authigenic ε_{Nd} record from MD99–2198, Tobago Basin [*Pahnke et al.*, 2008] (red crosses). (c) Ice-volume-corrected benthic $\delta^{18}O$ based from 26JPC [*Lynch-Stieglitz et al.*, 2011] (blue). Arrow shows the direction of a weakening AMOC. (d) Atmospheric $\Delta^{14}C$ changes from Cariaco Basin 58PC [*Hughen et al.*, 2000] (green) and ODP Leg 165 holes 1002D and 1002E [*Hughen et al.*, 2004, 2006] (pink). (e) Sedimentary $^{231}Pa/^{230}Th$ variations (dark red solid circles) from Bermuda Rise site GGC5 [*McManus et al.*, 2004]. (f) Greenland GISP2 ice core $\delta^{18}O$ (orange). Grey bars mark the Younger Dryas (YD) and Heinrich 1 (H1), whose timings are based on the timings in GISP2 $\delta^{18}O$ record.

the AMOC [*McManus et al.*, 2004]. Another abrupt decrease in $^{231}Pa/^{230}Th$ ratios during the YD also suggests a significant reduction in the meridional overturning rate. Records of sedimentary $^{231}Pa/^{230}Th$ ratios can be problematic in that they may be susceptible to changes in opal flux [*Chase et al.*, 2002]. However, the timing of the changes in the Pa/Th record at the Bermuda Rise are coincident with the timing of our ε_{Nd} excursions in the Florida Straits, and suggest a complementary oceanic circulation explanation (Figures 6a and 6e). Specifically, our new authigenic ε_{Nd} record from 26JPC indicates that both the YD and H1 are characterized by little to no AAIW penetration into the subtropical North Atlantic. Therefore, the correlation between variations in 26JPC benthic ice-volume-corrected $\delta^{18}O_{calcite}$ values, Cariaco Basin atmospheric $\Delta^{14}C$, the Bermuda Rise sedimentary $^{231}Pa/^{230}Th$ ratios, and our authigenic ε_{Nd} record from 26JPC supports the idea that there is a direct positive relationship between the strength of the AMOC and the northward penetration of the AAIW into the North Atlantic. Our results suggest a strong connection between Southern Ocean circulation changes and high-latitude North Atlantic climate change.

4.4. Competing Hypotheses

[19] This interpretation of our authigenic Nd isotope record is at odds with an earlier interpretation of a Nd isotope record based on a deeper core (1330 m) from the Tobago Basin (MD99–2198) [*Pahnke et al.*, 2008] (Figures 1 and 6b). The Tobago Basin authigenic ε_{Nd} data (Figure 6b) display variations that are anti-correlated with those recorded at our Florida Straits sites. Indeed, the Tobago Basin site shows abrupt excursions of ε_{Nd} values toward more positive values during the YD and H1, which was interpreted as recording an increase in the northward penetration of AAIW into the tropical North Atlantic during these cold events [*Pahnke et al.*, 2008]. Furthermore, a recent study from the high-latitude North Atlantic (between 1200 m and 2300 m water depth) suggests that AAIW may be the source of extremely radiocarbon-depleted (negative $\Delta^{14}C$ excursion) water that penetrated as far north as ~60°N during the YD and H1 when AMOC was significantly reduced [*Thornalley et al.*, 2011]. Similarly, another study argued that high nutrient content (Cd/Ca and $\delta^{13}C$) reconstructed from benthic foraminifera at ~60°N (NEAP4K, 1600 m water depth) in the Atlantic suggested an extreme northward penetration of AAIW during the YD and H1 [*Rickaby and Elderfield*, 2005]. However, in each of these previous studies, the selected sediment cores are all *deeper* than the modern depth range of AAIW (500 m to 1100 m) [*Kameo et al.*, 2004; *Wüst*, 1964]. Considering that the Tobago Basin study site is located below the core of modern AAIW, it is possible that the authigenic ε_{Nd} values at this site are recording relative variations between NADW ($\varepsilon_{Nd} = -13.5$) and GNAIW ($\varepsilon_{Nd} = -9.7$), rather than those between NADW and AAIW ($\varepsilon_{Nd} = -6$ to -9). Perhaps the more radiogenic ε_{Nd} values reconstructed for the YD and H1 in the Tobago Basin [*Pahnke et al.*, 2008] recorded the shoaling of GNAIW instead of variations in the northward penetration of the AAIW. It is therefore also possible that the radiocarbon-depleted signatures [*Thornalley et al.*, 2011] and the high nutrient contents [*Rickaby and Elderfield*, 2005] recorded in intermediate depth cores from the high-latitude

North Atlantic during these events are a result of reduced intermediate-depth ventilation of GNAIW and a shoaling of this as well as underlying southern ocean water masses [*Came et al.*, 2008; *Roberts et al.*, 2010]. Reduced ventilation during the YD and H1 in the high-latitude North Atlantic at intermediate depths (1–2 km) would imprint this intermediate water mass with a radiocarbon-depleted signature as suggested by *Thornalley et al.* [2011]. Shoaling of the underlying southern sourced water could have brought in nutrient-rich signals at these sites located at a depth below AAIW. Furthermore, *Sortor and Lund* [2011] found no evidence for increased amounts of radiocarbon-depleted AAIW in the south Atlantic during the H1 or the YD, making it even more unlikely that the high-latitude north Atlantic records reflect changes in AAIW across the last deglaciation.

[20] One important hypothesis for ocean circulation during the YD and H1, as proposed by modeling studies [*Keeling and Stephens*, 2001; *Weaver et al.*, 2003] and geochemical evidence [*Rickaby and Elderfield*, 2005], is that meltwater fluxes to the high-latitude North Atlantic during these cold events reverse the density contrast between AAIW and NADW. Today, a stable conveyor circulation can be attributed to the density contrast between salty NADW and fresh AAIW: $\rho_{\mathrm{NADW}} > \rho_{\mathrm{AAIW}}$. During the LGM, even if AAIW became saltier, a stable meridional overturning circulation cell at shallower depths in the North Atlantic with $\rho_{\mathrm{GNAIW}} > \rho_{\mathrm{AAIW}}$ can still be maintained. However, during the YD and H1, the density contrast between northern- and southern-component water is proposed to be reversed, i.e., $\rho_{\mathrm{GNAIW}} < \rho_{\mathrm{AAIW}}$ [*Keeling and Stephens*, 2001; *Rickaby and Elderfield*, 2005; *Weaver et al.*, 2003], such that AAIW underlies GNAIW and penetrates further north at deeper depths. This ocean circulation scheme is termed a collapsed or an "off" mode of the overturning circulation. This collapsed mode of overturning circulation would result in a very different ocean circulation scheme where southern component waters are exported to the North Atlantic via glacial AABW and AAIW at deep and intermediate depths and are compensated by surface and subsurface return flow of the GNAIW. If this hypothesis is true, then the collapsed mode of the overturning circulation would allow AAIW (termed HAAIW in this case) during the YD and H1 to reach core sites that underlie modern AAIW. This could potentially explain why intermediate depth cores at low and high latitudes (MD99–2198, 1330 m, and NEAP4K, 1600 m) in the North Atlantic both record an increased influence of AAIW during the YD and H1. HAAIW may reach depths greater than that of our study sites in the Florida Straits, such that we are unable to monitor its variations across these cold events. This hypothesis is appealing in that it may be able to explain the contradictions between our authigenic Nd isotope records and Nd isotope and nutrient records from previous studies [*Pahnke et al.*, 2008; *Rickaby and Elderfield*, 2005]. However, a closer investigation shows that average authigenic $\varepsilon_{\mathrm{Nd}}$ values during the YD and H1 in our study ($\varepsilon_{\mathrm{Nd}} = -9.41 \pm 0.16$) are identical to those in *Pahnke et al.* [2008] ($\varepsilon_{\mathrm{Nd}} = -9.70 \pm 0.66$), suggesting that 26JPC and the Tobago Basin core are recording the same water mass during the YD and H1. It might be possible that the denser HAAIW dominates a much wider depth range (from 546 m, 26JPC, to 1600 m, NEAP4K) during the YD and H1 relative to Holocene. Even if this is possible, it cannot explain the contrasting variability in authigenic $\varepsilon_{\mathrm{Nd}}$ results in 26JPC and MD99–2198. In the Florida Straits, abrupt decreases in authigenic $\varepsilon_{\mathrm{Nd}}$ values during the YD and H1 suggest a reduced contribution of AAIW, while the same values in the Tobago Basin show abrupt increases during these periods. These increased $\varepsilon_{\mathrm{Nd}}$ values are identical to those in 26JPC during the same periods, yet were interpreted to record a strengthening of AAIW in the Tobago Basin. We feel that the simplest explanation to this contradiction is that AAIW does not reach either 26JPC or MD99–2198 during the YD and H1, and that both cores are recording the same northern component water.

5. Conclusions

[21] Our high-resolution authigenic $\varepsilon_{\mathrm{Nd}}$ records from cores 26JPC and 31JPC provide strong evidence that there was a significantly reduced presence of AAIW in the Florida Straits during both the YD and H1, when proxy records indicate that AMOC was greatly reduced. This interpretation directly contradicts a previous $\varepsilon_{\mathrm{Nd}}$ study from the Tobago Basin that argued for an increased northward penetration of AAIW during the YD and H1 [*Pahnke et al.*, 2008]. This disagreement most likely stems from the fact that the Tobago Basin record was generated using a core that lies beneath the modern AAIW depth range. Therefore, we suggest that the Tobago Basin study fails to record true AAIW signals across the last deglaciation. Results from our study suggest an important role for AAIW during abrupt climate events associated with the reorganization of AMOC, and a tight connection between Atlantic intermediate water circulation variability and high-latitude North Atlantic climate change.

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