Strong middepth warming and weak radiocarbon imprints in the equatorial Atlantic during Heinrich 1 and Younger Dryas

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Abstract

We present a benthic foraminiferal multiproxy record of eastern equatorial Atlantic (EEA) middepth water (1295 m) covering the last deglacial. We show that EEA middepth water temperatures were elevated by 3.9 ± 0.5°C and 5.2 ± 1.2°C during Heinrich event 1 (H1) and Younger Dryas (YD), respectively. The radiocarbon content of the EEA middepth during H1 and YD is relatively low and comparable to the values of the pre-H1 episode and Bølling-Allerød, respectively. A transient Earth system model simulation, which mimics the observed deglacial Atlantic Meridional Overturning Circulation (AMOC) history, qualitatively reproduces the major features of the EEA proxy records. The simulation results suggest that fresh water-induced weakening of the AMOC leads to a vertical shift of the horizon of Southern Ocean-sourced water and a stronger influence of EEA sea surface temperatures via mixing. Our findings affirm the lack of a distinctive signature of radiocarbon depletion and therefore do not support the notion of interhemispheric exchanges of strongly radiocarbon-depleted middepth water across the tropical Atlantic during H1 and YD. Our temperature reconstruction presents a critical zonal and water depth extension of existing tropical Atlantic data and documents a large-scale and basin-wide warming across the thermocline and middepth of the tropical Atlantic during H1 and YD. Significant difference in the timing and pace of H1 middepth warming between tropical Atlantic and North Atlantic likely points to a limited role of the tropical Atlantic middepth warming in the rapid heat buildup in the North Atlantic middepth.

1. Introduction

The deglacial climate transition from the Last Glacial Maximum to the Holocene was interrupted by two major returns to cold conditions associated with the Heinrich 1 (H1) and Younger Dryas (YD) events (\cite{hemming2004, liu2009, marcott2014, ngrip-members2004, shakun2012}). Though some studies suggest that ice sheet instabilities are not the trigger of the onset of H1 and YD (\cite{alvarez-solas2013, barker2015, thornalley2010}), ice sheet instabilities and meltwater-induced weakening of the Atlantic Meridional Overturning Circulation (AMOC) are considered by others as one of the major forcings of the climate perturbations during H1 and YD (\cite{liu2009, mcmanus2004, menviel2011, renussen2015, timmermann2005}). Subsurface warming in and northward heat transport from the tropical Atlantic may have contributed to North Atlantic middepth warming which has been invoked as a terminating process for the H1 and YD events (\cite{dokken2013, marcott2011, petersen2013, shaffer2004, thiagarajan2014}). The magnitude and spatial extent of tropical Atlantic middepth warming and its temporal and causal relationship to the middepth warming in the North Atlantic is, however, not well established. Most of the middepth oceanographic reconstructions in the tropical Atlantic are limited to the western part of the basin and quantitative reconstruction of middepth temperature is very limited (\cite{camer2007, camer2008, freeman2015, huang2014, lund2015, lynche2014, mangini2010, oppo2015, ruhlmann2004, schmidt2012, sortor2011, xie2012, xie2014}). Here we focus on the middepth of the eastern equatorial Atlantic (EEA) and examine last deglacial oceanographic changes using sensitive Mg/Ca paleothermometry and benthic foraminiferal Ba/Ca, $\delta^{13}C$, and $\Delta^{14}C$ reconstructions. Our proxy records along with the results of a transient model simulation reveal a substantial difference in the temporal evolution of middepth warming in the tropical and North Atlantic consistent with the idea of a limited thermal exchange between the tropical and North Atlantic during H1.
We focus on the MD03-2707 marine core sediment that was recovered from the eastern equatorial Atlantic (02°30.11′ N, 09°23.68′ E, water depth 1295 m) (Figure 1a). Water column profiles of salinity, salinity-temperature, and temperature-\(\Delta\text{CO}_2^2\)/\(C_0^3\) relationships show that the core site is currently located in the transition zone between North Atlantic Deep Water (NADW) and Antarctic Intermediate Water horizon (AAIW) (Figure 1). At the location of MD03-2707 the bottom water is characterized by salinity, temperature, and \(\Delta\text{CO}_2^2\)/of 34.8 practical salinity unit (psu), 4.2°C, and 50 \(\mu\text{mol/kg}\), respectively (Figures 1b–1d).

MD03-2707 consists of a 37.4 m long core sediment that continuously covers the last 155 thousand years before present (kyr B.P.) [Weldeab et al., 2007]. In this study, we focus on the sediment depth between 497.5 and 802 cm that covers the time window between 10 and 20.5 kyr B.P. (Figure 2). Radiocarbon measurements in tests of benthic and planktonic foraminifers (see below) were carried out at the Leibniz Labor for Radiometric Dating, University of Kiel, Germany. We added seven (7) new radiocarbon datings of Globigerinoides ruber (pink) tests to the existing nine (9) radiocarbon data of the same species in the target sediment section [Weldeab et al., 2007]. In this study, we focus on the sediment depth between 497.5 and 802 cm that covers the time window between 10 and 20.5 kyr B.P. (Figure 2). Radiocarbon measurements in tests of benthic and planktonic foraminifers (see below) were carried out at the Leibniz Labor for Radiometric Dating, University of Kiel, Germany.

We added seven (7) new radiocarbon datings of Globigerinoides ruber (pink) tests to the existing nine (9) radiocarbon data of the same species in the target sediment section [Weldeab et al., 2007]. The age model and age model uncertainty of the target sediment segment are established using 23 radiocarbon datings of Globigerinoides ruber pink, the Marine13 data set [Reimer et al., 2013], a constant reservoir age correction of 400 years, and the Bayesian statistical software “Bacon” [Blaauw and Christen, 2011] (Figure 2).

We carried out 13 radiocarbon datings of mixed epibenthic foraminifers (Cibicides spp.). We calculated the \(\Delta^{14}\text{C}\) (‰) of bottom water using the equation formulated by Adkins and Boyle [1997]: \(\Delta^{14}\text{C}\) (‰) = ([e\(^{-14}\text{C}\) age/8033]/(e\(^{-\text{cal age}/8266}\) – 1)\(\times 1000\), with cal age and \(^{14}\text{C}\) being calendar age of the sediment (Figure 2) and radiocarbon
elements were analyzed by the isotope dilution/internal standard method [Martin and Lea, 2002] using a Thermo Finnigan Element2 sector inductively coupled plasma mass spectrometer (ICP-MS). Analytical reproducibility, assessed by analyzing consistency standards matched in Mg/Ca and Ba/Ca ratios to dissolved for
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**Figure 2.** Age model of MD03-2707 target segment (497.5–802 cm) based on 16 radiocarbon data of *G. ruber* pink tests, constant reservoir age correction of 400 years and calculated using Bacon, a Bayesian statistics software [Blaauw and Christen, 2011]. Calibrate age model control points (blue) and the age-depth model with gray stippled lines showing 95% confidence intervals and red curve showing a single “best” model based on the weighted mean age for each depth.

2.2. Model Simulation

We use the three-dimensional atmosphere-vegetation-ocean sea ice-carbon cycle model LOVECLIM [Goosse et al., 2010]. The ocean-sea ice component of LOVECLIM (CLIO) consists of a primitive equation ocean general circulation model with a resolution of 3 × 3 [Goosse et al., 1999]. CLIO uses a free surface and is coupled to a dynamic-thermodynamic sea ice model [Fichefet et al., 1994; Fichefet and Maqueda, 1997]. There are 20 unevenly spaced levels in the vertical with thicknesses ranging from 10 m near the surface to ~700 m below 3000 m. The strength of vertical mixing is coupled to the roughness of bottom topography [Decloedt and Luther, 2010] as presented in Friedrich et al. [2011]. Mixing along isopycnals, the effect of mesoscale eddies on transports and mixing as well as downsloping currents over continental shelves is parameterized [Goosse et al., 2010]. The atmospheric component of LOVECLIM is the ECBilt model, a spectral T21 model based on quasi-geostrophic equations extended by estimates of ageostrophic forcing terms [Opsteegh et al., 1998]. It consists of three levels in the vertical and contains a full hydrological cycle that is closed over land by a bucket model for soil moisture and a runoff scheme. The atmosphere, ocean, and sea ice components of the LOVECLIM model are coupled by exchange of momentum, heat, and freshwater. The marine carbon cycle component of LOVECLIM is the Liège Ocean Carbon Heteronomous (LOCH) model [Mouchet, 2011]. LOCH is a three-dimensional ocean carbon cycle model with prognostic equations for dissolved inorganic carbon, total alkalinity, phosphate, dissolved and particulate organic matter, silicate, and radiocarbon. LOCH is coupled to CLIO using the same time step. Biogeochemical tracers that are subject to advection and mixing are advected and mixed using the same circulation fields and mixing parameters, respectively, as in CLIO. The surface partial pressure of CO₂ is computed from total alkalinity, dissolved inorganic carbon,
temperature, and salinity. The parameterization of CO₂ exchange with the atmosphere follows Wanninkhof [1992]. More details on the LOVECLIM model can be found in Goosse et al. [2010].

Experimental Design: The LOVECLIM model is used to perform a transient simulation covering the period of 22–10 ka B.P. The transient simulation is forced by the time evolution values of atmospheric CO₂ concentrations [Lüthi et al., 2008]. Orbital parameters vary set according to Berger [1978] and Northern Hemispheric ice sheet extent and height are obtained from a recent ice sheet simulation [Ganopolski and Calov, 2011]. Orbital parameters and greenhouse gas concentrations are updated every model year. Ice sheet extent and height are updated every 100 model years. In order to avoid internally generated, self-sustained oscillations of the AMOC [Friedrich et al., 2010], a constant Last Glacial Maximum (LGM) bathymetry [Roche et al., 2007] is used.

The model is spun up for 10,000 model years using constant climate boundary conditions of 22 kyr B.P. to account for the long equilibration timescale of marine radiocarbon. Atmospheric radiocarbon is not transient but set to 0‰ throughout the entire simulation. The rationale behind this simplified approach is the focus on the effects of ocean circulation changes on marine radiocarbon. The latter become better identifiable when large changes in atmospheric radiocarbon [Reimer et al., 2013] are omitted in the simulation.

Freshwater (FW) forcing is applied to the North Atlantic region ~50°N–70°N to mimic AMOC changes during the H1-Bølling-Allerød (BA)-YD transition (Figures 3a and 3b). The FW forcing is applied without the use of salt.
compensation as the latter leads to artifacts in the response of the ocean circulation [Stocker et al., 2007]. For Heinrich Event 1, a substantial weakening of the AMOC around 18 kyr B.P. is generated by FW hosing over the North Atlantic with a strength of 0.2 sverdrup (Sv) (1 Sv = 10^6 m^3/s) and a duration of 1000 model years (Figure 3a). After ~3300 model years a pulse of 0.2 Sv FW extraction is applied for 500 model years resulting in the recovery of the AMOC (Figure 3b). The YD period is simulated by slightly weaker and shorter FW forcing. It should be noted that the FW forcing is idealized, and it is not intended to represent recorded changes in sea level or ice-rafted debris. Strength and duration of the forcing has been chosen according to the model’s AMOC sensitivity as well as to simulate a temporal evolution of the AMOC that is in qualitative agreement with the Bermuda Rise 231Pa/230Th record [McManus et al., 2004] (Figure 3b). The latter is largely corroborated by a spatially comprehensive study [Bradtmiller et al., 2014], although the spatial pattern is complex, indicating that 231Pa/230Th may harbor imprints other than that of deglacial AMOC changes. An additional passive tracer was added to the ocean model to elucidate the response of the Atlantic water mass composition to the large-scale ocean circulation changes generated in the wake of the recurring AMOC weakening-recovery cycle during the H1-BA-YD transition. The tracer is designed to distinguish between southern sourced and northern sourced water masses. To represent a Southern Hemisphere end-member, the passive tracer is restored to a value of +1 at the surface of the Southern Ocean south of 60°S using a relaxation timescale of 10 days. For the North Atlantic the tracer is designed to represent NADW and is restored to a value of +1 at the surface of the North Atlantic between ~55°N and 70°N. Outside the restoring areas the tracer can evolve freely and is advected and mixed as all other tracers of the ocean model. The specific design of the water mass tracer allows identifying a local water mass composition by comparing the local value to the end-member values of ±1.

The zonal mean of the water mass tracer for the Atlantic after the model spin-up is shown in Figure 3c. The two end-member regions can be clearly discerned. The southward propagation of NADW can be identified by a tongue of high values at depths of around 2500 m. Due to northward transport of near-surface waters from the Southern Hemisphere to the North Atlantic, the surface is characterized by lower values and thus a stronger impact of Southern Hemisphere waters. Below the main depth of NADW, the increasing impact of Antarctic Bottom Water (AABW) can be seen from lower values of the water mass tracer. Overall it must be noted, however, that the tracer can only identify the respective fraction of one water mass with respect to its antagonistic water mass based on their geographical origin. Despite this caveat, the water mass tracers will be important assets for our analysis of the change in tropical Atlantic water mass composition during the last glacial termination.

3. Results and Discussion

The temporal focus of this study is centered on the time interval between 9.5 and 20.5 kyr B.P. (Figure 2) and aims to shed light on the oceanographic conditions of the EEA during the last deglaciation. The sediment accumulation in the investigated MD2707 section varies between 16 and 50 cm/kyr (Figure 2). With a sampling interval of 2 cm, the resolution of the proxy time series varies between centennial and multidecadal. Between 20.5 ± 0.4 and 18 ± 0.4 kyr B.P. the δ18O, Ba/Ca, and Mg/Ca time series show little variance with average values of 4.12 ± 0.16‰, 1.88 ± 0.35 μmol/mol, and 3.5 ± 0.22 mmol/mol (3.9 ± 0.6°C), respectively (Figure 4). The δ13C exhibits an average value of 0.38 ± 0.13‰ with large fluctuations (19–20.5) that do not show a clear trend. Over the same time window, the Δ14C of bottom water varies between 262 ± 27 and 169 ± 27‰, respectively (Figure 4e). Starting at 17.5 ± 0.4 kyr B.P., Ba/Ca and Mg/Ca rise and δ18O and δ13C decline synchronously, reaching plateau values between 16.6 ± 0.5 and 14.0 ± 0.2 kyr B.P. (Figure 4). Concomitant with these changes, benthic foraminiferal δ13C drops from 247 ± 32‰ to 46 ± 6‰. Between 14.5 ± 0.2 and 13 kyr B.P. Ba/Ca and Mg/Ca drop significantly and δ13C increases, on average, by 0.3‰. Starting at 12.9 ± 0.17 kyr B.P. Mg/Ca and δ18O continue to rise and peak at 12 ± 0.34 kyr B.P. These changes are, however, not accompanied by significant and systematic changes in δ13C and Δ14C. A brief and significant Ba/Ca decline is centered at 12.3 ± 0.1 kyr B.P. Starting at 11.6 ± 0.36 kyr B.P., Mg/Ca declines rapidly and reaches at 9.6 kyr B.P. a value of 3.07 mmol/mol (Figure 4d). We removed the temperature effect from benthic foraminiferal δ18O and obtained an estimate δ18O of seawater that is a composite local salinity and global level changes. The low resolution and relatively large error allow only insights into an orbital-scale trend that, as expected, appear to be dominated by the size of the ice sheets. We note that the Mg/Ca series was presented in Weldeab et al. [2016] without discussing the temporal and paleoclimatic context of the record.
3.1. Middepth Warming and Water Mass Changes

The most prominent features of our multiproxy record are centered between 17.5 ± 0.3 and 14.5 ± 0.2 kyr B.P. and between 12.9 ± 0.2 and 11.6 ± 0.2 kyr B.P. (Figures 4 and 5). Taking into consideration age model uncertainties, these two episodes are synchronous with the timing of H1 and YD. During H1, $\delta^{13}C$ and Ba/Ca in *Cibicides pachyderma*, an epibenthic foraminifera, show a negative and positive trends, respectively (Figures 4d and 4e). $\delta^{13}C$ in benthic foraminifers is used as a proxy for the $\delta^{13}C$ signature of dissolved inorganic carbon (DIC) that, in turn, is used as a qualitative indicator for source and age of the water mass [Oppo et al., 2015, and reference therein]. Ba/Ca in benthic foraminifers is used as a proxy for refractory nutrients [Hall and Chan, 2004; Lea and Boyle, 1989]. Several processes can, however, complicate the interpretation of these proxies. The lowest benthic foraminiferal $\delta^{13}C$ value (0.01 ± 0.06‰) is centered between 16.5 and 14.5 kyr B.P. and falls within the timing of a rapid drop of $\delta^{13}C$ of atmospheric CO2 ($\delta^{13}C_{\text{atmosphere}}$) from $-6.5 \pm 0.05$‰ to $-6.8 \pm 0.05$‰ [Schmitt et al., 2012]. Estimates of past air-sea gas exchange rates are highly uncertain [Gebbie, 2014; Schmittner et al., 2013]. However, as shown in a model simulation [Menviel et al., 2012], without invoking changes in ocean circulation or export productivity a temperature-corrected $\delta^{13}C_{\text{atmosphere}}$ drop of 0.3 ± 0.01‰ can explain only a negligible fraction of a 0.3‰ decrease in the EEA $\delta^{13}C$ record.

The $\delta^{13}C$ record of *G. ruber* pink (a mixed layer dweller) in MD03-2707 shows relatively low values (mean value: 1.23 ± 0.1‰, n = 21) during H1 relative to the pre-H1 episode (mean value: 1.49 ± 0.13‰, n = 61) (Figure 5). Whereas low $\delta^{13}C$ value of upwelled DIC is reflected by low $\delta^{13}C$ values in *G. ruber* tests, low CO2$_2$ concentration and low temperature, which characterize upwelled waters, have the opposite effect (i.e., high $\delta^{13}C$) and a significantly stronger impact on $\delta^{13}C$ of *G. ruber* tests [Peeters et al., 2002]. The relatively low $\delta^{13}C$ of *G. ruber* pink and continuous sea surface temperature (SST) increase (Figures 5j and 5k) during H1, therefore, argue against wind-induced upwelling. Similarly, low Ba/Ca values in *G. ruber* pink (Figure 5l), a proxy for runoff [Weldeab et al., 2007, 2014a, 2014b], do not support enhanced runoff and an elevated riverine nutrient supply to the surface water. We conclude, therefore, that proxies for sea surface conditions indicate a relatively low surface productivity and, by inference, most likely reduced export productivity. Furthermore, the increase of benthic foraminiferal Ba/Ca and its decoupling from the magnitude and trend of Ba/Ca in *G. ruber* (Figures 5e and 5l) supports our argument that the bottom water enrichment in refractory nutrients is unrelated to changes in surface water processes. We thus infer that the
increase and decrease of the benthic foraminiferal Ba/Ca and δ^{13}C, respectively, reflect changes in residence time or source of the bottom water. An increase of ventilation age or a flow of radiocarbon-depleted water during the H1 episode is indicated by slightly reduced ΔΔ14C values (ΔΔ14C = Δ14C_{water} - Δ14C_{atmosphere}) relative to the values of the Bølling-Allerød (Figure 5c). We note, however, that prior to H1 the ΔΔ14C record already shows relatively low values that considering the uncertainty in ΔΔ14C estimates are essentially indistinguishable from those of the H1 episode. Therefore, in contrast to the enrichment of refractory nutrients (Figures 5e and 5g) and consistent with the coral-based ΔΔ14C record from the EEA [Chen et al., 2015], our ΔΔ14C record does not reveal a distinctive feature that is exclusively and unambiguously associated with the H1 episode.

The Mg/Ca-based BWT reconstruction reveals episodes of significant bottom warming (Figure 4d). It is also known that changes in carbonate ion concentration, expressed as ΔCO2/C0^{3}, affect benthic foraminiferal Mg/Ca [Elderfield et al., 2006; Marchitto et al., 2007a; Rosenthal et al., 2006; Weldeab et al., 2016]. Sensitivity of Mg/Ca in Globobulimina spp. to changes in ΔCO2/C0^{3} pore water is estimated to be 0.009 ± 0.0044 mmol/mol per unit ΔCO2/C0^{3} (μmol/kg) [Weldeab et al., 2016]. While, to our knowledge, there is no ΔCO2/C0^{3} reconstruction available for the water depth of our core location, a low resolution record from the EEA (water depth: 2945 m) indicates a ΔCO2/C0^{3} bottom water change of 9 μmol/kg from the last glacial (36.4 ± 7.4 μmol/kg) to the early Holocene (27.1 ± 6.4 μmol/kg).

According to Weldeab et al. [2016], the ΔCO2/C0^{3} bottom water change of 9 μmol/kg would correspond to a ΔCO2/C0^{3} pore water change of 22 μmol/kg and a potential bias of 0.6 ± 0.3°C in the BWT estimate, which is with the uncertainty of the Mg/Ca-BWT calibration for Globobulimina spp. [Weldeab et al., 2016].

It stands out that the H1 episode was marked by strong bottom water warming with an average value of 3.9 ± 0.5°C above that of the pre-H1 episode (Figures 4d and 6d). The onset of the warming is synchronous with the timing of changes in Ba/Ca and δ^{13}C records. The termination of bottom water warming occurred at 14 ± 0.2 kyr B.P., approximately 500 ± 200 years after the onset of AMOC intensification, as inferred from 231Pa/230Th records [Bradtmiller et al., 2014; McManus et al., 2004] and Bølling-Allerød warming in northern high latitude [NGRIP-members, 2004]. The BWT decline during the Bølling-Allerød was followed by bottom water warming of 5.2 ± 1.2°C whose timing and duration coincides with that of the YD. In contrast to the H1 record, the temperature rise during YD is not accompanied by an increase of refractory nutrients and

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**Figure 5.** Results of proxy and model simulation analyses. (a) 231Pa/230Th time series [McManus et al., 2004], a proxy for changes in AMOC strength, (b) simulated AMOC response to freshwater forcing, (c) Δ^{13}C (Δ^{13}C_{benthic foraminiferal} - Δ^{13}C_{atmosphere}), (d) Δ^{14}C: simulated changes in 14C at the water depth of MD2707 site relative to the 14C values at the start of the experiment, (e) Ba/Ca in tests of Cibicides pachyderma, (f) simulated changes in phosphate concentration, (g) δ^{13}C analyzed in tests of Cibicides pachyderma, and (i) simulated changes in water mass influence of the Southern Ocean (-1) and North Atlantic (+1) end-members at MD2707 site. (j) δ^{13}C of G. ruber pink, (k) Mg/Ca-based SST estimates [Weldeab et al., 2007], and (l) Ba/Ca in G. ruber pink, a proxy for a relative runoff estimates [Weldeab et al., 2007].
In summary, the MD03-2707 record reveals that the EEA middepth experienced a large-scale warming and changes in source or residence time of intermediate level water during the H1. The YD episode differs from H1 by showing the most prominent middepth warming and the absence of an enrichment of refractory nutrient, suggesting a relatively reduced impact of a weak freshwater forcing on the middepth of the eastern equatorial Atlantic.

3.2. Results of Transient Model Simulation

To explore the driving mechanisms of the bottom water warming and changes in EEA bottom water tracers, we carried out an idealized transient model simulation forced with freshwater input into the North Atlantic (Figure 3). In response to the freshwater forcing, the AMOC weakens significantly during H1 and YD (Figures 3b and 5b). In qualitative agreement with the proxy data (Figure 5d), the simulation results (Figure 5c) suggest a decrease of $\Delta^{14}C_{\text{bottom water}}$, an increase of bottom water nutrient concentration (Figure 5f), bottom water warming (Figure 6e), and a relative increase of Southern Ocean influence (Figure 5i).

While freshwater forcing during YD results in a moderate weakening of the AMOC (Figure 5b), consistent with the proxy records, the simulation results do not show an increase of refractory nutrients (Figure 5f) and only a small increase of Southern Ocean-sourced water (Figure 5i). According to the results of the model simulation, the enhanced bottom water residence time and increased refractory nutrients over our core site during H1 are a result of an increase of Southern Ocean-sourced water relative to the decreasing volume of North Atlantic Deep Water (Figures 3d and 5i). The reduction of the cold North Atlantic water inflow and the overall sluggish circulation during H1 enhances the influence of SST on the middepth temperature via downward diffusive heat exchanges, leading to significant intermediate water warming (Figure 6). Even though model and proxy results are in good qualitative agreement with respect to a bottom water temperature increase in response to an AMOC weakening, the simulated timing and amplitude of the warming differs from the Mg/Ca record.
One possible reason for this discrepancy is the abruptness of the simulated AMOC change (Figure 5b) vis-à-vis the slower reconstructed AMOC reorganization (Figure 5a).

Qualitatively consistent with the proxy record, the modeling results for YD indicate a BWT rise (Figure 6e) and a lack of nutrient enrichment (Figure 5). The difference in the strength and duration of the freshwater forcing during H1 and YD explains the deviant tracer imprints in the intermediate water depth of the equatorial Atlantic. The stronger bottom water warming during the YD relative to that of H1 can be partly explained by the elevated greenhouse gas and solar insolation forcings and their impact on the EEA SST (Figures 6a–6d). Taken together, the results of the model simulation qualitatively reproduce the prominent features found in the EEA deglacial proxy records. Both the model and proxy data indicate that the intermediate water of the eastern equatorial Atlantic responded sensitively to meltwater-induced AMOC weakening. Deviating from the proxy BWT reconstruction, the result of the model run does not show persistent bottom water warming after the termination of H1. Because the major BWT drop occurred 500 ± 200 years after the reinvigoration of the AMOC (Figure 6e), it is likely that the sustained warming in the tropical Atlantic middepth reflects an equatorward transport of relatively warm middepth North Atlantic water (see below).

### 3.3. Timing and Spatial Extent of Middepth Warming

A quantitative reconstruction of subsurface and middepth temperature changes in the tropical Atlantic during the last deglacial was so far limited to the western part of the basin [Came et al., 2007; Schmidt et al., 2012]. The middepth and thermocline warming of western tropical Atlantic during H1 and YD is estimated ~3°C and 2–4°C, respectively [Came et al., 2007; Lynch-Stieglitz et al., 2014; Rühlemann et al., 2004; Schmidt et al., 2012]. Taking uncertainties of the age models and temperature estimates into account, the magnitude and timing of western Atlantic middepth warming is comparable to and synchronous with these of the eastern equatorial Atlantic (Figures 6d and 6f). While age model uncertainties put constraints on the confidence of centennial-scale comparison, the persistent EEA middepth warming 500 ± 200 years after the end of H1 (Figures 4c and 6d) is also evident in the subsurface temperature record of the western tropical Atlantic [Schmidt et al., 2012]. Overall, thermocline and middepth warming in the western and eastern tropical Atlantic indicate a basin-wide and, therefore, large heat accumulation in response to an AMOC perturbation during H1 and YD, consistent an Atlantic thermocline deepening [Cessi et al., 2004; Timmermann et al., 2005].

Available BWT records from the North Atlantic middepth reveal a varying timing and pace of BWT rise during H1 (Figures 6) [Marcott et al., 2011; Thiagarajan et al., 2014]. Though the low resolution, a potential effect of carbonate ion concentration on the Mg/Ca and the multispecies nature of the Mg/Ca record do not allow a robust comparison of the pace and magnitude of thermal changes across the transition of last glacial and H1 episodes, a BWT reconstruction from the northwestern Atlantic middepth suggests H1 warming [Marcott et al., 2011]. While discontinuous, another record of western North Atlantic temperature estimates across the water depth between 1176 and 2600 m indicates an abrupt temperature rise of 6°C between 15.1 and 15.8 kyr B.P. [Thiagarajan et al., 2014] that coincides with the timing of the height of the AMOC decline, as inferred from 231Pa/230Th records [Bradtmiller et al., 2014] and the peak of ice-rafted debris (IRD) across the north Atlantic (Figure 6) [Gherardi et al., 2005; McManus et al., 2004; Stern and Lisiecki, 2013]. The gradual middepth warming in the tropical Atlantic (Figures 6d and 6f) preceded the abrupt warming of the North Atlantic middepth by about 3000 years (Figures 6b and 6c). Because of its spatial coverage and the robustness of its age model, we compare the temperature record from the North Atlantic middepth [Thiagarajan et al., 2014] with the thermocline and middepth temperature records from tropical Atlantic (Figures 6d–6g). While the discontinuity of the North Atlantic middepth record does not allow a detailed comparison of temperature trends throughout the H1 episode, the relatively abrupt and large (~6°C) warming at 15.1–15.8 kyr B.P. is unmatched by a comparable magnitude and pace of warming in the tropical Atlantic records that reveal a gradual BWT rise starting from 19 kyr B.P. (Figure 6). Conversely, the large tropical Atlantic middepth warming during the YD episode contrasts strongly with the concomitant cooling (relative to the Bølling-Allerøds) of North Atlantic middepth (Figure 6). These findings indicate a thermal decoupling between the tropical and North Atlantic middepths and challenge the notion of rapid heat exchanges between the middepths of the tropical and North Atlantic during H1 and YD [Dokken et al., 2013; Petersen et al., 2013; Shaffer, 2004]. We hypothesize that heat exchange was weak and the abrupt and
delayed North Atlantic middepth warming most likely occurred in response to an outflow of warm Arctic and Nordic Sea subsurface water that has been recently identified [Cronin et al., 2012; Thornalley et al., 2015].

3.4. Young-to-Moderate Age of Middepth Atlantic Water During H1

The abrupt H1 warming in the middepth of North Atlantic [Thiagarajan et al., 2014] was accompanied by the presence of very old water that has been initially interpreted to originate from the Southern Ocean [Robertson et al., 2005; Thiagarajan et al., 2014; Thornalley et al., 2011]. Linking the very old middepth North Atlantic water to advection of Southern Ocean-sourced middepth water is at odds with our results from the middepth of the tropical Atlantic that serves as a conduit for the AAIW. The upper part of northern tropical Atlantic middepth (≤1000 m water depth) appears to be marked by a strong influence of relatively young North Atlantic water [Came et al., 2008; Huang et al., 2014; Xie et al., 2012, 2014]. Consistent with previous studies [Chen et al., 2015; Cléroux et al., 2011; Freeman et al., 2015; Lund et al., 2015; Mangini et al., 2010; Sortor and Lund, 2011] our results reveal the absence of a radiocarbon-depleted middepth water in the tropical Atlantic during H1 (Figure 7). The implication of this observation is twofold. First, given the relatively young-to-moderate age of middepth water in the tropical Atlantic (Figure 7), it is difficult to reconcile that the strong radiocarbon depletion of middepth North Atlantic water can be solely attributed to entrainment of extremely old Southern Ocean water into the AAIW. An alternative explanation, as suggested in recent studies [Thornalley et al., 2015; Wilson et al., 2014], is that an outflow of radiocarbon-depleted Nordic Sea subsurface waters substantially contributed to the radiocarbon depletion of middepth North Atlantic water. The existence of strongly radiocarbon-depleted water in the middepth of the eastern tropical Pacific and Indian Ocean during H1 and YD has been interpreted to reflect a northward spread of upwelled, extremely old glacial water from the Pacific and Indian Ocean segments of the deep Southern Ocean [Bryan et al., 2010; Marchitto et al., 2007b]. In analogy to the interpretation of the tropical Pacific data [Marchitto et al., 2007b], Mangini et al. [2010] suggested an invasion of radiocarbon-depleted Southern Ocean-sourced water into the tropical Atlantic middepth. In agreement with the growing lines of evidence from the tropical Atlantic [Chen et al., 2015; Cléroux et al., 2011; Lund et al., 2015; Sortor and Lund, 2011], our $\Delta^{14}$C record does not support the hypothesized spread of an extremely old middepth water from the Southern Ocean into the middepth of the tropical Atlantic [Mangini et al., 2010]. One scenario is that large-scale upwellings of extremely old middepth water from the Southern Ocean during H1 and YD did not occur. Alternatively, if a northward spread of radiocarbon-depleted water occurred then air-sea exchanges of CO$_2$ and advective and diffusive mixing must have significantly diluted the radiocarbon-depleted water as it entrained into the AAIW and advected into the tropical Atlantic. Another scenario arises from proxy data that have been interpreted to suggest that during the early deglacial the deep South Atlantic was isolated from the middepth South Atlantic and that the latter shows no sign of ventilation events until 15 kyr B.P. [Lund et al., 2015; Rickaby et al., 2010; Yu et al., 2014]. Depending which of the two scenarios is more likely, it has implication for our understanding of the mechanism that led to the early deglacial atmospheric CO$_2$ rise.

Figure 7. $\Delta^{14}$C ($\Delta^{14}$C benthic foraminifera/cold water coral minus $\Delta^{14}$C atmosphere) reconstructed in tropical Atlantic [Cléroux et al., 2011; Freeman et al., 2015; Lund et al., 2015; Mangini et al., 2010], North Atlantic [Thornalley et al., 2011], and tropical Pacfic [Marchitto et al., 2007b] cores.
4. Summary and Conclusion

This study demonstrates that during H1 and YD the middepth of EEA was marked by a strong warming, with an average temperature rise of 3.9 ± 0.5°C and 5.2 ± 1.2°C, respectively. Together with the records from the western tropical Atlantic, the EEA record reveals a basin-wide and large-scale heat accumulation across the thermocline and middepth at times of a weakened AMOC. Our idealized transient model simulation qualitatively reproduces the major features of oceanographic changes and relates the middepth warming to a relative increase of downward heat diffusion due to the sluggish replenishment of cold middepth North Atlantic [Cessi et al., 2004]. The results of our thermal and water mass reconstructions during H1 and YD have an implication for our understanding of interhemispheric heat and water mass exchanges. A northward transport of extremely old Southern Ocean-sourced middepth water, as hypothesized in previous studies, is not consistent with the relatively young age of middepth water across the tropical Atlantic. Therefore, the hypothesized spread of old upwelled Southern Ocean-sourced water into the tropical Atlantic middepth may have not occurred or was rejuvenated by air-sea gas exchanges and advective/diffusive mixing as it was entrained into the AAIW and transported northward Atlantic. Contrasting the timing and pace of middepth warming during H1 and YD, a heat exchange between the tropical and North Atlantic middepth is not supported by the currently available data. Consequently, the rapid heat buildup in the middepth of North Atlantic must be related to changes within the subsurface and middepth of the northern high latitude.

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