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The Atlantic Meridional Overturning Circulation and Abrupt Climate Change

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abrupt climate change, AMOC, Dansgaard-Oeschger, Heinrich events, Younger Dryas

Abstract

Abrupt changes in climate have occurred in many locations around the globe over the last glacial cycle, with pronounced temperature swings on timescales of decades or less in the North Atlantic. The global pattern of these changes suggests that they reflect variability in the Atlantic meridional overturning circulation (AMOC). This review examines the evidence from ocean sediments for ocean circulation change over these abrupt events. The evidence for changes in the strength and structure of the AMOC associated with the Younger Dryas and many of the Heinrich events is strong. Although it has been difficult to directly document changes in the AMOC over the relatively short Dansgaard-Oeschger events, there is recent evidence supporting AMOC changes over most of these oscillations as well. The lack of direct evidence for circulation changes over the shortest events leaves open the possibility of other driving mechanisms for millennial-scale climate variability.

INTRODUCTION

Abrupt Climate Change During the Last Glacial Cycle

Paleoclimate records recovered from land, the ocean, and glacial ice all point to a dynamic global climate over much of the last glacial cycle. In many locations in the Northern Hemisphere, abrupt changes in climate have occurred that span almost the full range of glacial to interglacial conditions, with the transition between climate states occurring in decades or less (Alley & Clark 1999, Voelker 2002). These abrupt climate changes are most clearly recorded in the climate records from glacial ice on Greenland (Andersen et al. 2004) and are referred to as Dansgaard-Oeschger (D-O) events. The warm intervals are referred to as interstadials, and the cold intervals are referred to as stadials. The climate records from the Greenland ice cores have often served as the template for abrupt climate change over the last 80,000 years (**Figure 1**). The abrupt changes observed in Greenland ice are most pronounced at times of intermediate climate and ice volume, with a relatively stable climate during the Holocene (0–10 ka) and the Last Glacial Maximum (LGM) (19–23 ka) (Buizert & Schmittner 2015, McManus et al. 1999, Schulz et al. 1999). Several mechanisms have been proposed to explain the origin of these abrupt climate changes (Clement & Peterson 2008), including sea-ice variability (Gildor & Tziperman 2003, Li et al. 2005), ice-shelf growth and decay (Petersen et al. 2013), shifts in preferred Northern Hemisphere planetary wave patterns (Seager & Battisti 2007, Wunsch 2006), and changes in the tropics (Clement et al. 2001). However, the prevailing paradigm is that the abrupt climate changes are a result of changes in the northward transport of heat by the Atlantic meridional overturning circulation (AMOC) (Broecker et al. 1985, Clark et al. 2002, Rahmstorf 2002).

The Atlantic Meridional Overturning Circulation Hypothesis for Abrupt Climate Change

The first abrupt climate change event linked to a collapse or weakening of the AMOC was the Younger Dryas (Rooth 1982). This abrupt change back to cooler conditions in the Northern Hemisphere, which occurred on the deglaciation (**Figure 1**), was first recognized in European pollen records and has since been linked to temperature and rainfall changes throughout the Northern Hemisphere (Carlson 2013). The link between the Younger Dryas cooling and the AMOC was supported by evidence for less extensive North Atlantic Deep Water (NADW) during the LGM and Younger Dryas (Boyle & Keigwin 1987), consistent with changes in the position of the polar front in the North Atlantic on the deglaciation (Ruddiman & McIntyre 1981). Providing further support for this idea were modeling and theoretical studies that suggested that there may be two stable states of the North Atlantic circulation (Manabe & Stouffer 1988, Stommel 1961). The idea that changes in the AMOC could lead to abrupt climate change was extended to explain the climate oscillations recorded in Greenland ice during marine isotope stage 3 (D-O events) (Broecker et al. 1985).

Different ideas have been put forward to explain what might cause the AMOC to change over the abrupt climate changes of the last glacial cycle. For the Younger Dryas, the rerouting of Laurentide Ice Sheet melt from the Mississippi River drainage to the St. Lawrence River drainage has been invoked (Carlson 2013, Carlson et al. 2007, Rooth 1982). This could have provided enough extra freshwater to the North Atlantic to interrupt the AMOC. Periodic purges of the continental ice directly into the North Atlantic are evident throughout the last glacial cycle. These Heinrich events were identified as layers of ice-rafted grains in North Atlantic sediments (Bond et al. 1992, Hemming 2004). Heinrich events occurred during some of the D-O stadials but not all of them. It is hypothesized that the melting of these iceberg armadas could have provided

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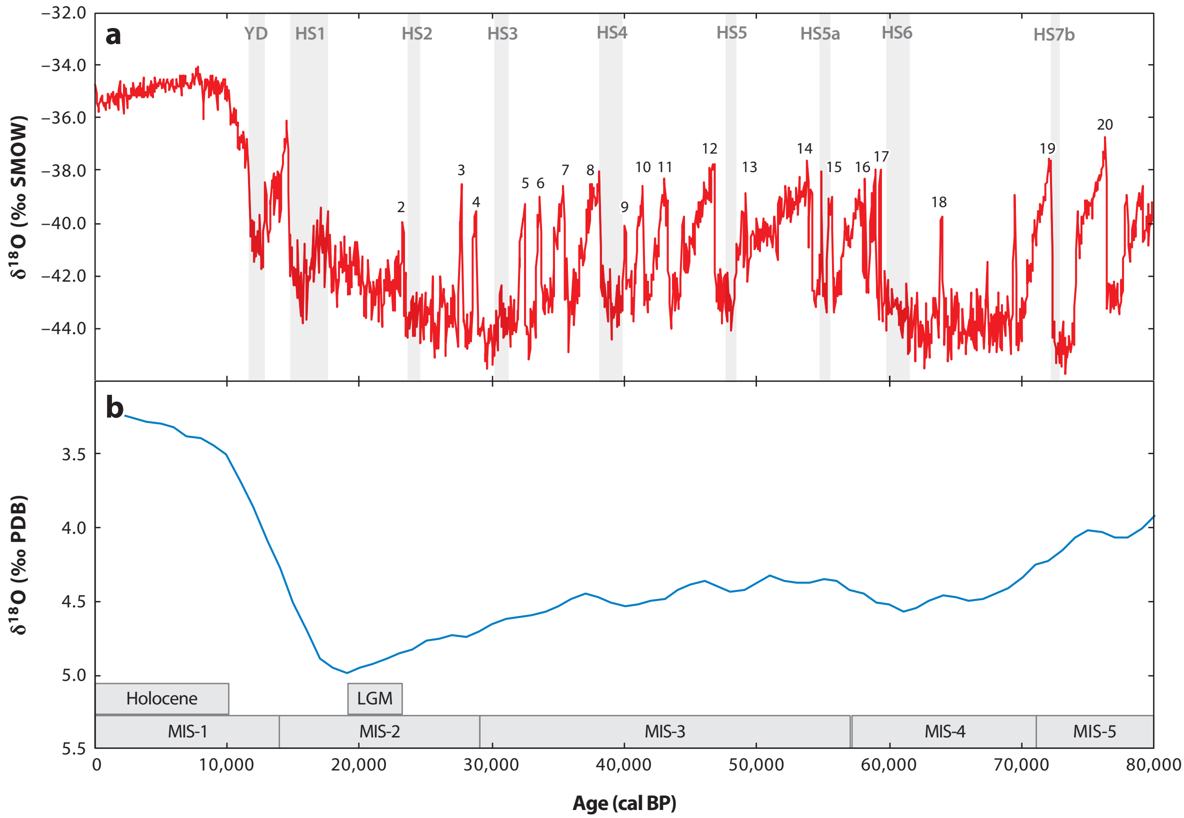

Figure 1

(*a*) Oxygen isotopic composition of Greenland ice, which serves as a proxy for temperature. Greenland interstadials (warm events) are numbered; the Younger Dryas (YD) cold event and Heinrich stadials (HSs) are indicated by the gray bars. The ages of the Heinrich stadials are taken from the U-Th-dated Chinese cave climate records (Wang et al. 2001). (*b*) Oxygen isotopic composition of deep-sea benthic foraminifera (Lisiecki & Raymo 2005), which reflects primarily changes in global ice volume, with some contribution from changes in deep-sea temperature. The time periods corresponding to the Holocene and Last Glacial Maximum (LGM) are marked, along with the marine isotope stage (MIS) boundaries (Lisiecki & Raymo 2005). Additional abbreviations: cal BP, calendar years before present; PDB, Pee Dee belemnite; SMOW, standard mean ocean water.

sufficient freshwater to interrupt the AMOC. Why the AMOC would be reduced during the D-O stadials that are not associated with Heinrich events is more difficult to explain. Modeling simulations are often driven by externally imposed changes in freshwater forcing, presumably representing contributions from melting land ice (Ganopolski & Rahmstorf 2001, Kageyama et al. 2010, Menviel et al. 2014), but self-sustained oscillations in the AMOC caused by variations in freshwater and heat transport through the ocean and atmosphere system have also been proposed (Broecker et al. 1990, Dokken et al. 2013, Zhang et al. 2014).

By the mid-1990s, a conceptual model had evolved in which the AMOC exhibited three different circulation states over the last glacial cycle (Alley et al. 1999, Sarnthein et al. 1994) (**Figure 2**): the warm circulation state of today, which is characterized by a strong and deep AMOC with deepwater formation in the Greenland-Iceland-Norwegian Seas; a cold circulation state typified by the LGM, with a shallower AMOC and deepwater formation shifted to the subpolar North Atlantic; and an off circulation state, which presumably prevailed after large inputs of freshwater,

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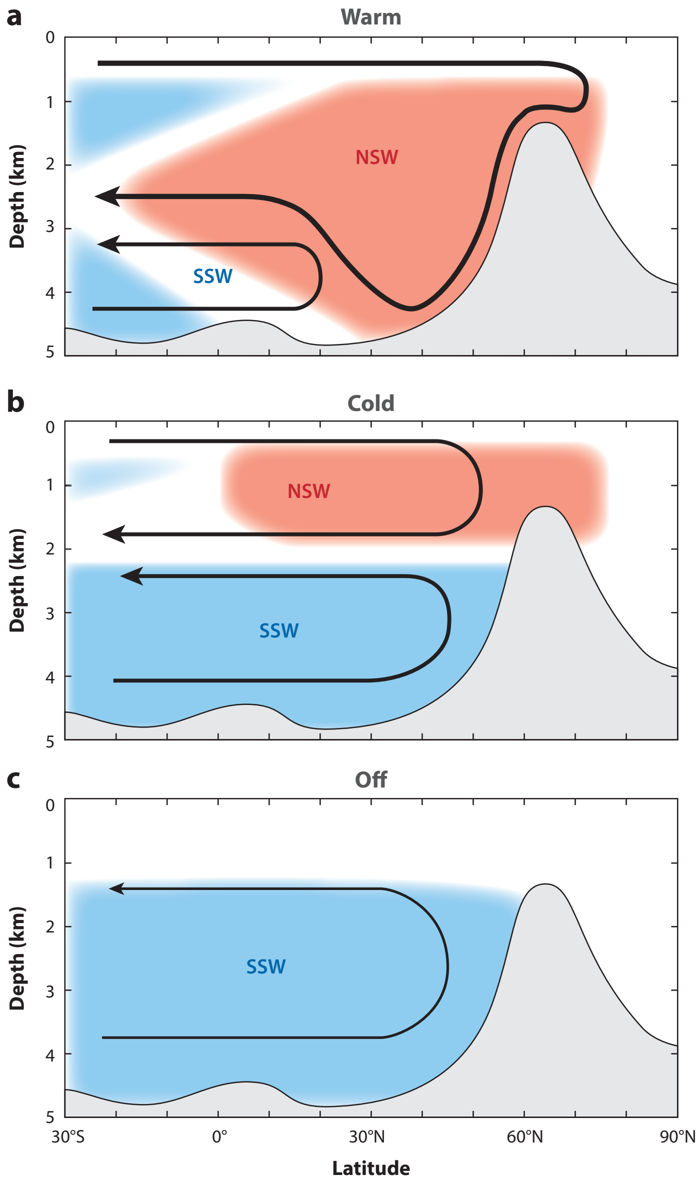

Figure 2

Conceptual model of circulation. The warm circulation state (panel *a*) holds for the Holocene and the marine isotope stage 3 interstadials, the cold circulation state (panel *b*) holds for the Last Glacial Maximum and marine isotope stage 3 stadials, and the off circulation state (panel *c*) results from the delivery of iceberg melt during Heinrich events. The approximate extent of waters with properties consistent with a northern source [Northern Source Water (NSW)] is shown in red, and the approximate extent of waters with properties consistent with a southern source [Southern Source Water (SSW)] is shown in blue. Meridional circulation cells are indicated with thicker arrows, denoting a more vigorous circulation. Modified by permission from Macmillan Publishers Ltd: *Nature* (Rahmstorf 2002), copyright 2002.

such as those that occurred during the Heinrich events (Rahmstorf 2002). The proposed cold and off circulation states were based on a combination of existing proxy data from ocean sediments and the response of ocean circulation models of varying degrees of complexity and freshwater inputs to the North Atlantic and other forcing changes.

In the years since this view of AMOC variability as the cause of glacial and deglacial abrupt climate change was put forward, considerable effort has been extended to document the changes in Atlantic circulation over these abrupt changes. In this article, I review the evidence from ocean sediments for changes in the AMOC associated with abrupt climate changes. The story is often far from clear. The records can be noisy and incomplete, the spatial coverage patchy, and the proxies subject to complicating factors. There is no one scenario that easily encompasses all of the proxy records, at least as they are currently interpreted. Although the true scenario is likely to be more nuanced and complex (or possibly completely different) than the conceptual model outlined above, this model provides a useful framework given the limited data available.

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PALEO-PROXY EVIDENCE FOR CHANGES IN THE ATLANTIC MERIDIONAL OVERTURNING CIRCULATION OVER THE LAST GLACIAL CYCLE

Proxies Relevant to Reconstruction of the Atlantic Meridional Overturning Circulation

The most direct evidence for AMOC changes comes from changes in subsurface water-mass properties (water-mass proxies) and indicators of the strength of subsurface flow (circulation proxies). Water-mass proxies include those that are related to biogeochemical cycling in the ocean as well as those that reflect abiotic exchanges with the ocean boundaries. However, in today's deep Atlantic, most water-mass properties can be explained by the relative mixtures of waters that left the surface in the North Atlantic (NADW) and those that left the surface in the Antarctic (Antarctic Bottom Water and Antarctic Intermediate Water). For times in the past, these waters are referred to more generally as Southern Source Water (SSW) and Northern Source Water (NSW). Cd/Ca ratios in benthic foraminifera serve as a proxy for PO_4 concentrations (high in SSW, low in NSW), stable carbon isotope ratios ($\delta^{13}C$) in foraminifera shells reflect seawater values (low in SSW, high in NSW), and radiocarbon ($\Delta^{14}C$) in benthic foraminifera and deepwater corals reflects age (old in SSW, young in NSW) (Lynch-Stieglitz & Marchitto 2014). Nd isotope ratios (εNd) are imprinted on waters by exchange with terrigenous sediments (high or more radiogenic in SSW, low or less radiogenic in NSW) and are recorded in precipitates on foraminifera and in sediments (Goldstein & Hemming 2014).

Circulation proxies include the terrigenous grain size in the sortable component of sediments (mean sortable silt grain size), which reflects the energy in near-bottom flows (large grains in energetic environments, and smaller grains where the near-bottom currents are weak) (McCave & Hall 2006). High-energy environments can reflect strong mean flow, strong variability and eddy activity, or both. The $^{231}Pa/^{230}Th$ ratio in sediments reflects larger-scale flows and the rain of particles through the water column. Both Pa and Th are produced from the decay of U in the water column and are scavenged by particles in the water column before being buried on the seafloor. In general, when circulation is strong, ^{231}Pa, which has a longer residence time in seawater, is preferentially exported from the basin. Once the in situ production of Pa and Th in sediments is accounted for, this results in low $^{231}Pa/^{230}Th$ in the deep Atlantic sediments today; if the circulation was weak in the past, the ratio in the open ocean should more closely reflect the ratio produced by the decay of U (high $^{231}Pa/^{230}Th$) (Yu et al. 1996). This basic picture is complicated by

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water-mass history, reversible scavenging, and different scavenging affinities for particles of different compositions (Deng et al. 2014). Considerable heterogeneity is expected and observed at different locations within the basin (Burckel et al. 2016). Oxygen isotope ratios in foraminifera reflect both the temperature of calcification and the oxygen isotopic composition of water, which is related to salinity in the upper ocean. This allows for the reconstruction of seawater density, which has been used to infer the strength of the large-scale overturning (Lynch-Stieglitz et al. 2006) and the strength of boundary currents (Lynch-Stieglitz et al. 1999a, 2014; Moreno-Chamarro et al. 2016). The mean flow in the ocean is largely geostrophic and thus is reflected in the mean density structure. This approach can provide a direct and straightforward measure of AMOC strength, but it relies on an accurate reconstruction of past seawater density at appropriate locations.

Each of these proxies for deepwater properties and flow has its strengths and weaknesses and instances when it may record other physical and biological processes and environmental conditions. Because an evaluation of the relative merits of the paleo-circulation proxies is beyond the scope of this article, I refer readers to the review by Lynch-Stieglitz & Marchitto (2014) for more detail on these proxies.

From the Last Glacial Maximum to the Present

There is now compelling and largely consistent evidence from several tracers for changes in deepwater circulation and properties over the deglaciation. During the LGM, deep water masses were arrayed differently than they are today. Multiple water-mass tracers suggest that SSW with properties similar to those of the Antarctic Bottom Water today dominated the glacial Atlantic below a depth of approximately 2 km, and that NSW with properties associated with the NADW today was found above 2 km (Curry & Oppo 2005, Howe et al. 2016, Keigwin 2004, Marchitto & Broecker 2006). This led to the interpretation of a shallower surface-to-deep overturning cell (the cold mode; **Figure 2**). In addition to the proxies for water-mass properties, there is evidence for a different LGM circulation from proxies that reflect the deep and intermediate water flow itself. The distribution of $^{231}Pa/^{230}Th$ in glacial-aged sediments from the deep Atlantic is consistent with a replacement of southward-flowing NADW with a more sluggish, northward-flowing SSW below 2 km (Jonkers et al. 2015, Negre et al. 2010). Both $^{231}Pa/^{230}Th$ and sedimentary grain size data support the idea of an active circulation in the upper water mass, consistent with a shoaling of the AMOC (Bradtmiller et al. 2014, Gherardi et al. 2009, Lippold et al. 2012). Evans & Hall (2008) used the sortable silt grain size proxy to show that the maximum in bottom current energy associated with southward-flowing NADW along the western boundary of the North Atlantic was reduced during the LGM, with a concomitant increase in bottom current speeds above 2.5 km, consistent with a shallower AMOC. Density reconstructions from ocean margin oxygen isotope data suggest both weaker flow through the Florida Straits and a reduced shear in the overturning circulation in the South Atlantic (Lynch-Stieglitz et al. 1999b, 2006). This would require that either the circulation cell was not associated with an interhemispheric surface-to-deep overturning, as is seen in today's AMOC, or the circulation was considerably weaker than today's AMOC in the Southern Hemisphere. However, regardless of whether the AMOC was stronger or weaker than it is today, there is broad consensus that there was some version of the conceptual model's cold circulation state during the LGM (Adkins 2013, Lynch-Stieglitz et al. 2007).

There is also now compelling evidence that the transition from the glacial to the modern AMOC was not unidirectional or smooth. The cold climate interval surrounding Heinrich stadial 1, although in many respects similar to the LGM, shows more extreme values in some circulation and water-mass tracers at some locations. As the deglaciation progressed, a more modern-like circulation set in, which was followed by a return to more glacial-like conditions during the Younger Dryas and the appearance of the modern strong AMOC in the early Holocene.

Properties that are more typical of SSW (low carbon isotope ratios and high Nd isotope ratios) are evident in the deep Atlantic during both Heinrich stadial 1 and the Younger Dryas (**Figure 3**). In addition to the water-mass tracers shown in **Figure 3**, deep Atlantic nutrient concentrations (reconstructed from seawater Cd) (Boyle & Keigwin 1987) and radiocarbon concentrations (Keigwin 2004, Robinson et al. 2005) show SSW values during these periods of cold Northern Hemisphere climate. In most cases, the deepwater properties during Heinrich stadial 1 and the Younger Dryas are quite similar to those observed during the LGM. In addition, intermediate waters along the western margin of the North Atlantic show water-mass properties more typical of North Atlantic origin during these intervals of cold climate, consistent with a decrease in the interhemispheric transport of Antarctic Intermediate Water (Came et al. 2008, Huang et al. 2014, Xie et al. 2012). Taken together, these data from the water-mass tracers suggest that the AMOC was shallow and weak during Heinrich stadial 1 and the Younger Dryas.

The shoaling of the southward deep western boundary flow associated with NADW that was inferred from the grain size data for the LGM persists through Heinrich stadial 1 (Hall et al. 2011) and reappears during the Younger Dryas (Evans & Hall 2008), consistent with a shallow AMOC during these cold events. Several other studies based on sediment grain size have also suggested reduced flow along the pathway of present-day NADW during both Heinrich stadial 1 and the Younger Dryas (Hall et al. 2011, McCave & Hall 2006, Praetorius et al. 2008). Oxygen isotopes in benthic foraminifera suggest a reduced density gradient across the Florida Straits during Heinrich stadial 1 and the Younger Dryas, consistent with a weakening of the upper branch of the AMOC (Lynch-Stieglitz et al. 2011, 2014). Subsurface and intermediate water warming in the western North Atlantic at several different locations has been inferred from proxies, with model studies suggesting that this subsurface warming is a direct consequence of weakened Atlantic overturning (Came et al. 2007, Liu et al. 2009, Lynch-Stieglitz et al. 2014, Ruhlemann et al. 2004, Schmidt et al. 2012). Perhaps the most often cited proxy record for changes in AMOC strength over the deglaciation is a record of $^{231}Pa/^{230}Th$ in sediments at a depth of 4.55 km on the Bermuda Rise (McManus et al. 2004). This record also suggests a longer residence time for deep waters in the Atlantic during Heinrich stadial 1 and the Younger Dryas relative to the time intervals before and after, with extreme values for Heinrich stadial 1. Overall, the proxy evidence seems to support a very weak AMOC during both Heinrich stadial 1 and the Younger Dryas (either some version of the off mode or a very weak cold mode), with perhaps greater weakening (the off mode) during Heinrich stadial 1.

However, even during the well-studied deglacial time period, data point to a more complex reality than a simple shift between two or three distinct circulation states. Upper deep waters and intermediate waters (above 2 km) in the open North Atlantic show a large shift from high to low $\delta^{13}C$ midway through Heinrich stadial 1 (Oppo et al. 2015) and a shift from lower to higher $\Delta^{14}C$ late in Heinrich stadial 1 (Chen et al. 2015). This large shift in the properties of intermediate and upper deep waters occurred at a time when the water-mass tracers in the deep waters and circulation proxies continue to suggest a weakened AMOC. It is not yet clear how these water-mass changes are related to circulation, but the timing of changes in the upper deep waters points to the possibility that, rather than a simple transition occurring between circulation states, the upper and lower deepwater changes may be decoupled in time.

Bradtmiller et al. (2014) compiled $^{231}Pa/^{230}Th$ data and examined evidence for differences in circulation between the Holocene, the LGM, and Heinrich stadial 1. Their interpretation suggests that during the LGM (the cold circulation in the conceptual model), the upper circulation cell was more vigorous than it is today, and the deeper circulation cell less so. However, they found the

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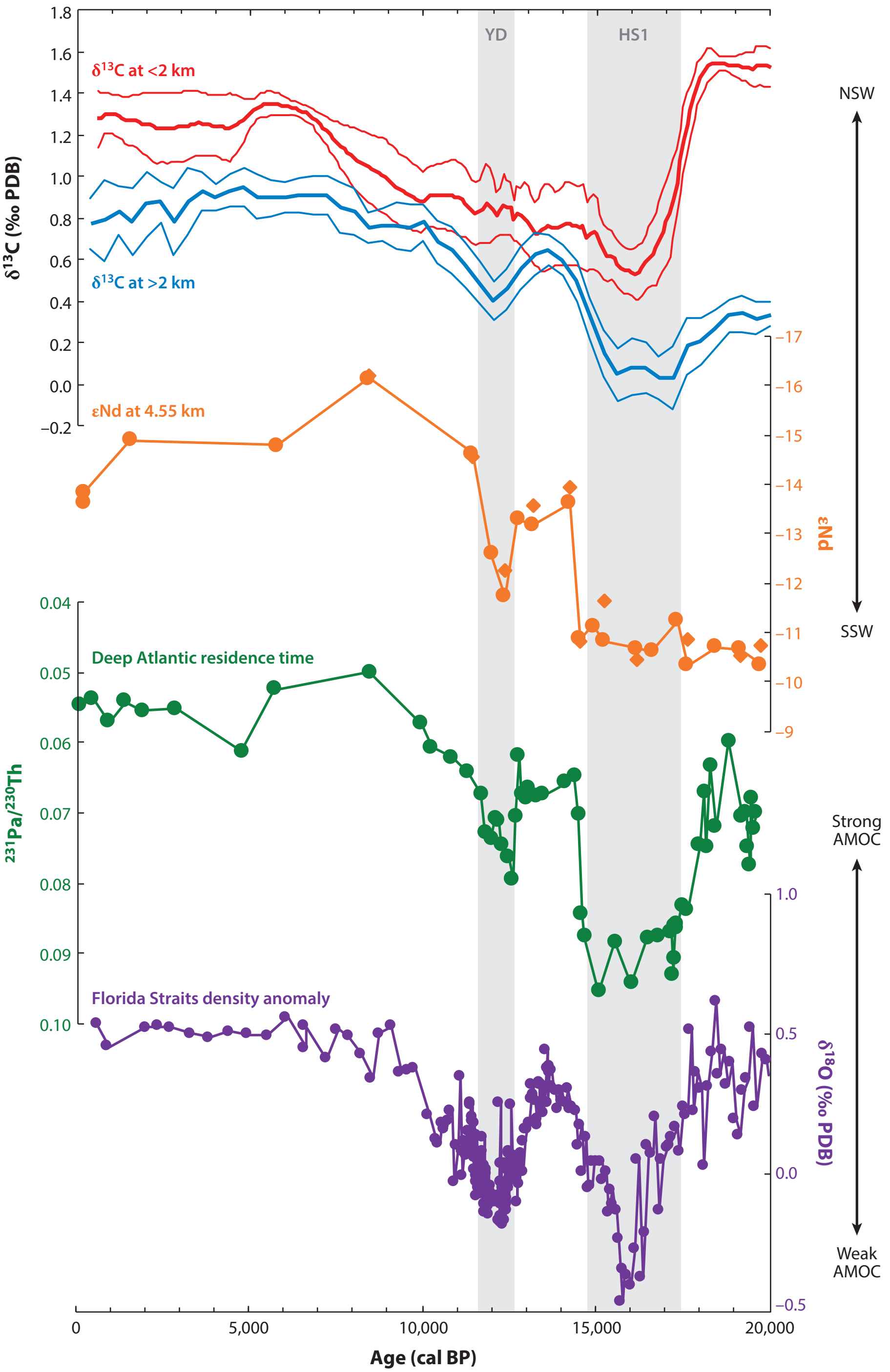

same general pattern for Heinrich stadial 1. A vigorous yet shallow overturning during Heinrich stadial 1 is difficult to reconcile with both the water-mass and circulation proxies from the upper ocean and the conceptual model for the off circulation.

Heinrich Events

The conceptual model described above would lead us to expect that during full glacial periods, the Heinrich events would be associated with a shift from the cold circulation mode to the off mode. In the upper 2 km of the North Atlantic, we would expect a transition from NSW to SSW during this time, or an aging in place of some of the NSW properties (lower $\Delta^{14}C$, higher nutrients, lower $\delta^{13}C$). In the lower 2 km of the North Atlantic, we would not necessarily expect a large change in these properties for a shift from the cold mode to the off mode because SSW would bathe the site in either mode. The conceptual model described above suggests that during marine isotope stage 3, the Heinrich events would be associated with a shift from the warm circulation mode to the off mode. In this case, we would expect excursions toward more SSW properties both above and below 2 km.

Many high-resolution benthic carbon isotope records have been generated in the search of this signature of deepwater change associated with the Heinrich events. Although some records above 2 km suggest the presence of nutrient-rich waters for some Heinrich events, the record is inconsistent among locations. This suggests that there may have been regional changes in circulation or, perhaps, productivity that can also affect the carbon isotope signature. This work has been hampered by poor time resolution and noisy data, so although good evidence for upper North Atlantic $\delta^{13}C$ change is not in hand at the present time, it is possible that it will emerge in the future.

Efforts to reconstruct deepwater $\delta^{13}C$ have been more successful. Lynch-Stieglitz et al. (2014) recently compiled well-resolved records from below 2 km in the North Atlantic, aligning these records based on radiocarbon and the distinctive benthic $\delta^{18}O$ pattern in marine isotope stage 3. Consistent with the expectations of the conceptual model, there is little change in deepwater $\delta^{13}C$ over the glacial-age Heinrich stadials (stadials 2 and 3) (**Figure 4**). Similarly, Nd isotopes at the Bermuda Rise (depth of 4.55 km) suggest that SSW persisted during the full glacial, with no indication of any additional changes in the deepwater source associated with the glacial-aged Heinrich stadials (stadials 1, 2, and 11) (Bohm et al. 2015). Also consistent with the expectations of the conceptual model, there is a clear shift in deepwater $\delta^{13}C$ from NSW to SSW values over the marine isotope stage 3 Heinrich stadials (stadials 4–6). However, contrary to expectations, during the Heinrich stadials that occurred during marine isotope stage 3, Nd isotope ratios do

Figure 3

North Atlantic water-mass and circulation proxy records over the deglaciation. Mean and 2-sigma standard-error $\delta^{13}C$ values (*thick* and *thin lines*, respectively) for the upper North Atlantic (<2 km) are shown in red, and those for the deep North Atlantic (>2 km) are shown in blue. The upper North Atlantic stack is the average of the records from above 2 km reported by Oppo et al. (2015); the deep North Atlantic stack was described by Lynch-Stieglitz et al. (2014). The deep (4.55 km) Nd isotope ratio (εNd) at the Bermuda Rise is shown in orange (Roberts et al. 2010). The $^{231}Pa/^{230}Th$ ratio, which can reflect changes in deepwater residence time, from the deep North Atlantic at the Bermuda Rise is shown in green (McManus et al. 2004). The ice-volume-corrected oxygen isotope ratio of benthic foraminifera on the Florida Margin, which can reflect changes in the density structure in the Florida Straits and the strength of the upper branch of the Atlantic meridional overturning circulation (AMOC), is shown in purple (Lynch-Stieglitz et al. 2011, 2014). Additional abbreviations: cal BP, calendar years before present; HS1, Heinrich stadial 1; NSW, Northern Source Water; PDB, Pee Dee belemnite; SSW, Southern Source Water; YD, Younger Dryas.

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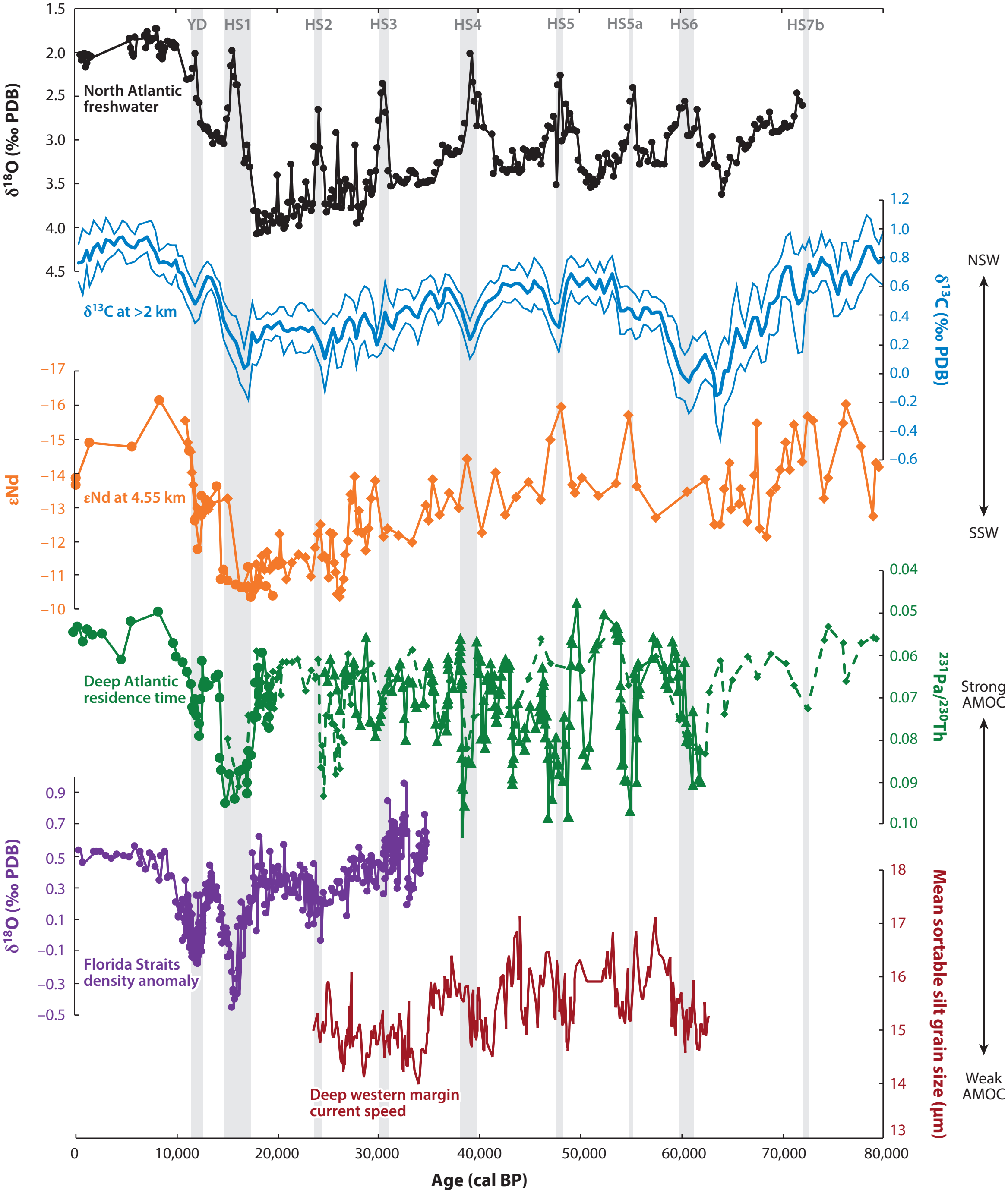

not shift toward more positive (SSW) values, indicating the continued presence of NADW at this site (**Figure 4**).

Although $^{231}Pa/^{230}Th$ ratios in sediments in deep waters on the Bermuda Rise approach the production ratio in association with Heinrich stadial 1, indicating long residence time for deep waters, such a clear signal is not seen for all of the other Heinrich stadials during full glacial conditions (stadials 2 and 3) (**Figure 4**). If these data are interpreted as a simple recorder of AMOC strength, then they do not support the conceptual model of an off circulation state during these Heinrich events. However, improved understanding of the controls on the cycling of Pa and Th and the incorporation of this knowledge into simple models warn against a straightforward interpretation of a record from a single site in terms of the strength of circulation over the entire water column (Burckel et al. 2016, Thomas et al. 2006). By contrast, the Pa/Th ratios during marine isotope stage 3 Heinrich stadials (stadials 4–6) are all associated with a clear excursion to the production ratio, consistent with the conceptual model of an off circulation.

Grain size data from the depth of the core of southward NADW flow along the western boundary suggest reduced southward flow of NADW throughout the glacial period between 35 ka and 15 ka (Hall et al. 2011, Hoogakker et al. 2007), consistent with a shoaling of this southward-flowing western boundary current during the LGM inferred from the depth transect of cores discussed above (Evans & Hall 2008). There is no indication of further flow speed reduction for the Heinrich stadials that occurred during this time period (stadials 2 and 3) (**Figure 4**). However, earlier in marine isotope stage 3, the grain size is similar to Holocene values at this site (Evans & Hall 2008) for much of the time, suggesting similar deepwater flow vigor. During this period, the grain size data suggest a return to low-flow glacial-like conditions during the earlier marine isotope stage 3 Heinrich stadials (stadials 4–6). Similarly, the excursion to low benthic oxygen isotope values on the Florida Margin for the Younger Dryas and Heinrich stadial 1 are not apparent for the glacial-aged Heinrich stadials 2 and 3. Such excursions can be interpreted as a reduction in the density gradient across the straits (as was the case for the Younger Dryas) or a more general warming of the upper North Atlantic in response to a weakened AMOC.

Taken together, these data suggest that the meltwater discharged during Heinrich stadials 2 and 3 did little to disrupt the glacial circulation state, which likely remained in the cold mode throughout. However, the grain size, Pa/Th, and carbon isotope data present a coherent story for a switch from the warm circulation state to an off mode for the earlier Heinrich stadials in marine isotope stage 3. Although the Nd isotope data at the Bermuda Rise seem to contradict

Figure 4

North Atlantic water-mass and circulation proxy records over the Heinrich events. The oxygen isotope ratio in the planktonic foraminifer *Neogloboquadrina pachyderma* from the western North Atlantic (Hillaire-Marcel & Bilodeau 2000) on an updated age model (Lynch-Stieglitz et al. 2014) is shown in black; low values reflect the presence of glacial meltwater. Mean and 2-sigma standard-error $\delta^{13}C$ values (*thick* and *thin lines*, respectively) for the deep North Atlantic (>2 km) are shown in blue (Lynch-Stieglitz et al. 2014). The deep (4.55 km) Nd isotope ratio (εNd) at the Bermuda Rise is shown in orange (Bohm et al. 2015, Roberts et al. 2010). The $^{231}Pa/^{230}Th$ ratio, which can reflect changes in deepwater residence time, from the deep North Atlantic at the Bermuda Rise is shown in green (Bohm et al. 2015 and Lippold et al. 2009, *dashed line with diamonds*; Henry et al. 2016, *solid line with triangles*; McManus et al. 2004, *solid line with circles*). The ice-volume-corrected oxygen isotope ratio of benthic foraminifera on the Florida Margin, which can reflect changes in the density structure in the Florida Straits and the strength of the upper branch of the Atlantic meridional overturning circulation (AMOC), is shown in purple (Lynch-Stieglitz et al. 2011, 2014). The mean sortable silt grain size along the western boundary of the North Atlantic at the depth of today's North Atlantic Deep Water, which can indicate the current speed on the deep western margin, is shown in dark red (Hoogakker et al. 2007); a large mean size indicates a vigorous flow. Additional abbreviations: cal BP, calendar years before present; HS, Heinrich stadial; NSW, Northern Source Water; PDB, Pee Dee belemnite; SSW, Southern Source Water; YD, Younger Dryas.

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this scenario, another record from the deep South Atlantic suggests a reduction in NSW export associated with the Heinrich events (Piotrowski et al. 2008).

Dansgaard-Oeschger Events

Although there are many different thoughts on the causes of the repeated abrupt warmings seen in the Greenland ice record (D-O events), the view that they represent the variability of the AMOC is widely held. In the conceptual model described above, stadials would be associated with the cold circulation state, and the interstadials would be associated with the warm circulation state. The Heinrich events occurred during some of the D-O stadials. However, there are often several D-O events between Heinrich events. This means that, in searching for evidence that the AMOC changed over the D-O events, one must be careful to look at the transitions between the non-Heinrich stadials and the interstadials, and not interpret evidence for AMOC changes related to the Heinrich events as evidence for D-O-related AMOC changes. The D-O events are often quite short, lasting only a few hundred to a few thousand years. This also makes it difficult to know whether the absence of D-O events in a water-mass or circulation proxy is due to the absence of change or the inability of that change to be recorded in ocean sediment records, which can be smoothed by bioturbation, the response time of the proxy, and other processes. Compounding the problem of limited time resolution is that abundances of foraminifera tend to be low when sedimentation rates are high: Even when sedimentation rates allow for good time resolution, limited numbers of individuals make it difficult to construct high-resolution records of foraminifera-based proxies. It is also difficult to date any events during marine isotope stage 3 precisely enough to know whether an excursion is related to a Greenland stadial or interstadial period. I therefore focus the discussion here on records from sediment cores with an independent (noncirculation) proxy that shows D-O variability (**Figure 5**). This both ensures that the core is capable of resolving D-O timescale variability if it is there and provides a way to align the timing of any events with the Greenland record.

In the conceptual model, a transition between cold and warm circulation states would be accompanied by a replacement of SSW (stadials) with NSW (interstadials) in the deep (>2 km) North Atlantic. Perhaps the best-resolved sedimentary records of surface ocean D-O variability in the North Atlantic come from the Portuguese Margin. Shackleton et al. (2000) reported from the same sediment core a planktonic isotope record that matches the ice-core record with high fidelity along with a benthic carbon and oxygen isotope record. Although it is clear that the marine isotope stage 3 Heinrich stadials are associated with an excursion in benthic $\delta^{13}C$ toward SSW values, similar excursions were not seen for all of the D-O stadials (**Figure 5*a***). Skinner & Elderfield (2007) reported similar results for a nearby location (**Figure 5*c***). Keigwin & Boyle (1999) presented planktonic oxygen isotope and benthic carbon isotope data over several D-O events from the deep North Atlantic at the Bermuda Rise. Although they found evidence for low $\delta^{13}C$ during non-Heinrich stadials, the record was noisy. A carbon isotope record from the South Atlantic suggests that there are some millennial-scale changes in deepwater properties in the deep South Atlantic (Charles et al. 1996), but a one-to-one correspondence with the D-O events is not seen. This same site in the South Atlantic has a record of Nd isotopes (Piotrowski et al. 2008), but as with $\delta^{13}C$, not every D-O event is resolved. Although these records showed the possibility of changes in NADW export associated with the D-O events, the resolution of the records was not sufficient to determine whether all of the D-O events were associated with water-mass property changes. However, these earlier attempts to document variability in deep water-mass properties were hampered by low abundances of benthic foraminifera, which led to lower temporal resolution and more noise than in the surface ocean records from the same cores.

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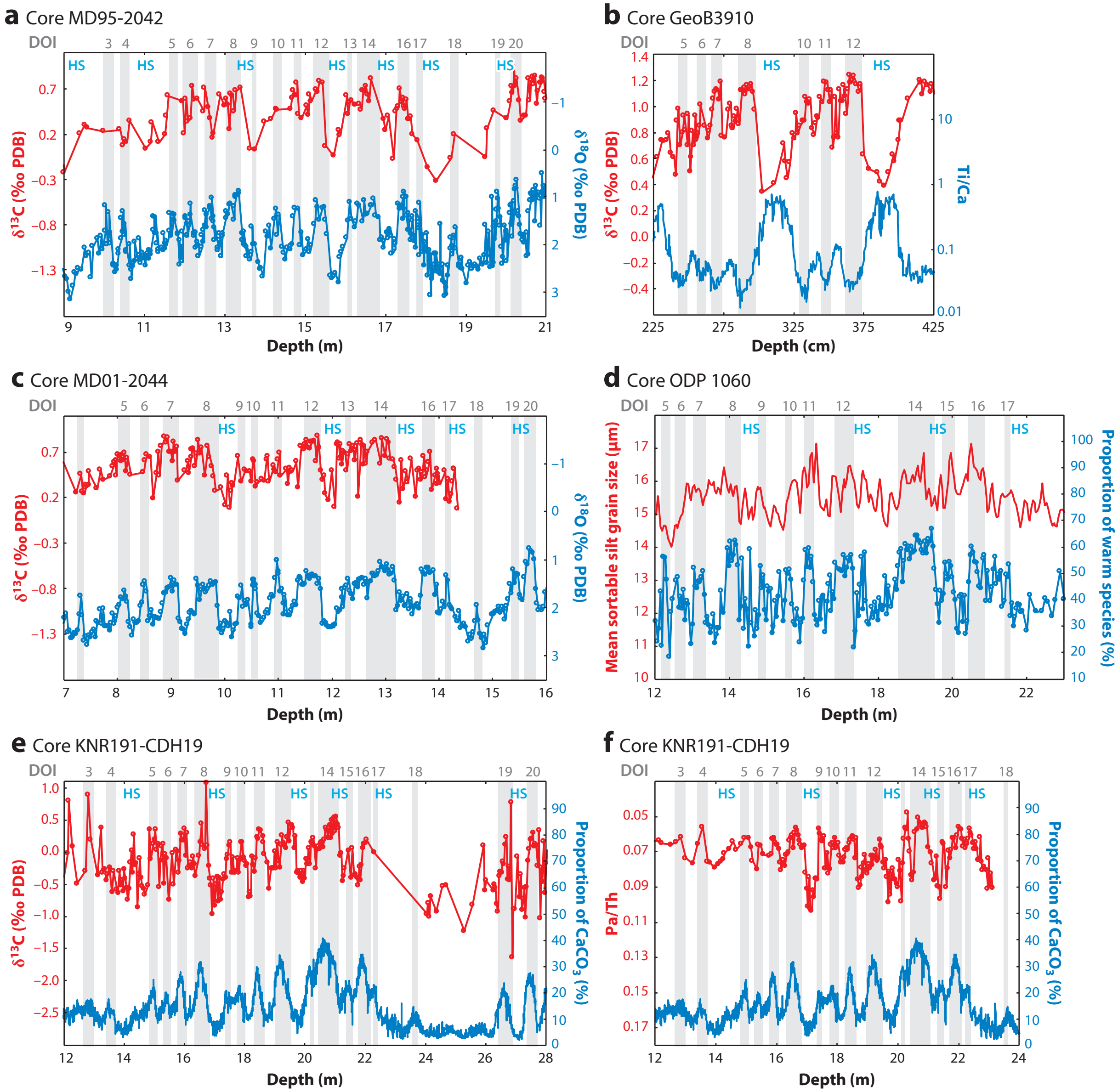

Figure 5

Water-mass and circulation proxies in sediment cores that resolve Dansgaard-Oeschger events in a surface proxy. (*a*) δ^{13}C of benthic foraminifera from the Iberian Margin at a depth of 3.15 km (Shackleton et al. 2000), plotted with the δ^{18}O of a planktonic foraminifer that resolves changes in surface conditions over the Dansgaard-Oeschger events. (*b*) δ^{13}C of benthic foraminifera from the Brazil Margin at a depth of 2.36 km, plotted with Ti/Ca ratios that record river runoff events (Burckel et al. 2015). (*c*) The same two proxies as in panel *a* from another sediment core on the Iberian Margin at a depth of 2.66 km (Hodell et al. 2013, Skinner & Elderfield 2007, Skinner et al. 2007). (*d*) Mean sortable silt grain size on the deep western boundary (Hoogakker et al. 2007), plotted with the percentage of warm species present (Vautravers et al. 2004). (*e*) δ^{13}C of benthic foraminifera from the Bermuda Rise at a depth of 4.55 km, plotted with the proportion of $CaCO_3$, which can reflect terrigenous sediment input and dissolution of carbonate on the seafloor (Henry et al. 2016). (*f*) Pa/Th from the Bermuda Rise at a depth of 4.55 km, plotted with the proportion of $CaCO_3$ (Henry et al. 2016). Abbreviations: DOI, Dansgaard-Oeschger interstadial; HS, Heinrich stadial; PDB, Pee Dee belemnite.

However, more recently, a sediment core from the tropical Atlantic off Brazil showed excursions toward low $\delta^{13}C$ for all but the shortest D-O stadials (Burckel et al. 2015) (**Figure 5*b***). The core is shallower (2.3 km) and farther south than the Iberian Margin cores, and so perhaps is relatively more sensitive to changes in water-mass boundaries associated with the marine isotope stage 3 AMOC changes. But a highly resolved benthic $\delta^{13}C$ record from the Bermuda Rise (depth of 4.55 km) has recently been published (Henry et al. 2016), and this core also clearly shows changes for all but the very shortest of the D-O events (**Figure 5*e***), consistent with the conceptual model of a switch from the cold mode (D-O stadials) to the warm mode (D-O interstadials). However, additional records from a variety of locations that can fully resolve $\delta^{13}C$ changes over the D-O events are needed to fully evaluate the nature of the deep water-mass and circulation changes over the D-O events.

One property of deep Atlantic sediments that appears to record D-O variability in a number of locations is the proportion of $CaCO_3$ (%$CaCO_3$) or other measures of carbonate versus terrigenous sedimentation. Keigwin & Jones (1994) noted the correlation between the %$CaCO_3$ in deep western North Atlantic sediments and the Greenland climate, and they ascribed the lower values during stadials to a combination of increased dilution by terrigenous sediments and enhanced dissolution resulting from the replacement of NSW with SSW. Changes in the overlying production of carbonate and the delivery of organic matter to the sediments (which can also promote dissolution) are additional mechanisms that can change the %$CaCO_3$ in the sediment. Owing to the multiple influences on this proxy, it has not been traditionally viewed as a water-mass tracer. However, it is possible that the D-O variability in %$CaCO_3$ and related measures (Ti/Ca) are, in fact, primarily records of the relative proportion of NSW and SSW at the deep Bermuda Rise site. In a recent study using a deep South Atlantic core, Gottschalk et al. (2015) constructed records from several related proxies, including those that measure dissolution intensity in sediments and reconstruct carbonate ion concentrations in deep water. Unlike %$CaCO_3$, these proxies are not affected directly by increased dilution by terrestrial material. Gottschalk et al. (2015) made a strong case that the changes in dissolution intensity at this location are driven by changes in the input of the less corrosive NSW relative to SSW and found a good correspondence to many of the D-O cycles from the Greenland ice core.

Evidence for or against AMOC changes during D-O events based on the circulation proxies has also, until recently, been limited and contradictory. Density reconstructions from the Florida Straits extend only as far back as D-O interstadial 6. Although there is no evidence of circulation change over the very short events, there are $\delta^{18}O$ excursions consistent with a strong upper branch of the AMOC for D-O interstadials 5 and 6 (Them et al. 2015). Thornalley et al. (2013) presented a thorough and convincing analysis of the flow speed of deep western boundary currents off North America over the transition between late marine isotope stage 5 and stage 4 based on the sortable silt grain size proxy. They found that the flow speeds at all depths from 2 to 4.5 km are similar to those in the Holocene for all of the D-O interstadials in this time interval. They also found that, although the flow speed above 2.5 km clearly increased during cold intervals, this increase was accompanied by a decrease in flow speed at the depth of today's NADW maximum only for the cold intervals during full glacial conditions. This would suggest that, at least for some of these D-O events (D-O interstadials 19 and 20), the Greenland warmings do not necessarily correspond to a strengthening in the AMOC. If anything, the evidence appears to show that the circulation during the stadials that immediately preceded the warmings was stronger (although shallower) than it was during the D-O interstadials. Although there are no similarly well-resolved records from above 2.5 km for marine isotope stage 3, Hoogakker et al. (2007) reported grain size data from a site near the depth of the present-day maximum flow speed associated with NADW (**Figure 5*d***). This record does not show consistent variability associated with the marine isotope stage 3 D-O

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events, supporting the idea that, like the stage 5 events, the D-O warmings may not correspond to a stronger AMOC. Although the grain size data do not appear to reflect changes in circulation over the D-O events, the changes in the magnetic mineral content in a series of cores in the North Atlantic do show D-O variability and have been interpreted as reflecting increased deep flow from the Nordic Seas during D-O interstadials (Kissel et al. 1999). Also, in the same Bermuda Rise sediment core in which they documented δ^{13}C changes over most of the D-O events, Henry et al. (2016) found changes in Pa/Th over most of the D-O events as well, further bolstering the case for D-O variability in the AMOC.

If D-O variability is caused by changes in the strength and/or pattern of the AMOC, as is widely assumed, then do the currently available data support this view? With many years of effort, I think that we are finally coming to the point where there is good evidence to support this view. Some records of deepwater properties do show changes over most of the D-O events, most notably the newer δ^{13}C record from the Bermuda Rise (Henry et al. 2016). Other relatively high-resolution records do not show these changes, but at least in some cases, this is likely due to the failure of the sedimentary record to record D-O variability (because of low sedimentation rates, noisy proxies, etc.). In addition, there is now good evidence from the Pa/Th circulation tracer for changes over most of the D-O events. Do the existing data support the conceptual model that D-O variability represents switches from the cold to warm circulation state? The water-mass proxy data do support this view, with a switch from SSW to NSW in the deep Atlantic. The Pa/Th data from the Bermuda Rise also seem to support a switch from a more vigorous flow (interstadial warm state) to a less vigorous but still active flow (stadial cold state) during the non-Heinrich stadials. However, inferences about the depth of NSW export from grain size measurements along the western boundary of the North Atlantic suggest that the flow remained vigorous at the depth of today's NADW during the non-Heinrich stadials, which is not consistent with the simple conceptual model. It also remains possible that the shortest D-O events (such as D-O interstadials 2, 3, 4, and 18) were not accompanied by ocean circulation changes, as records that are clearly capable of resolving these changes have yet to emerge.

INDIRECT EVIDENCE FOR CHANGES IN THE ATLANTIC MERIDIONAL OVERTURNING CIRCULATION

Despite the difficulties in finding direct evidence for changes in AMOC strength or structure over many of the abrupt climate changes seen in the Greenland ice record, such changes have been widely thought to be integral to the mechanism of abrupt climate change for at least 30 years. This is likely due to the strong links between the AMOC and patterns of surface ocean and climate change. Ruddiman & McIntyre (1981) documented that the polar front in the North Atlantic, which today trends from the southwest to the northeast, assumed a more zonal position during the LGM and the Younger Dryas. Bond et al. (1993) provided evidence that the frontal position also changes during the abrupt climate changes of marine isotope stage 3. General circulation model studies in which the AMOC was disrupted by the introduction of excess freshwater to the North Atlantic predicted a pattern of temperature change that was similar to observations, with large cooling over the North Atlantic and Greenland and more modest temperature changes elsewhere (Kageyama et al. 2010, Manabe & Stouffer 1995). More recently, Guillevic et al. (2013) showed that the spatial gradients in temperature and accumulation rates over the D-O events on the Greenland Ice Sheet are consistent with the patterns expected for changes in the amount of sea ice versus open water in the Nordic Seas. Although the changes in the patterns of sea surface temperatures in the North Atlantic, sea-ice cover, and climate over Greenland can have multiple causes, all are consistent with the changes that are expected for switches between periods where there is NADW formation in the far north Atlantic (the warm state of the conceptual model) and where there is not (the cold and off states of the conceptual model).

Because the AMOC transports heat northward across the equator, a reduced AMOC should lead to warming in the Southern Hemisphere (Crowley 1992), and such an antiphasing was discovered between the Greenland and Antarctic ice-core records (Barker et al. 2011, Blunier & Brook 2001). A compilation of temperature records suggests a timing for this antiphased climate change between the hemispheres on the deglaciation that is quite similar to that observed in the circulation proxies (Clark et al. 2012, Shakun & Carlson 2010). In models, the decrease in northward heat transport leads to a more southerly position of the intertropical convergence zone (e.g., Zhang & Delworth 2005), and many data have been generated that support such a change at the time of the Heinrich stadials (Arz et al. 1998, Mulitza et al. 2008, Peterson et al. 2000, Wang et al. 2004). It may turn out that this fingerprint of AMOC changes on the surface ocean and atmosphere is as good a proxy for AMOC variability as any of the deep-sea-based approaches outlined above. Ritz et al. (2013) and Zhang et al. (2015) have taken this approach and estimated the magnitude of AMOC change for the deglaciation and marine isotope stage 3 using model-based spatial fingerprints and records of sea surface temperature change. When this approach is used, however, care must be taken to exclude other mechanisms of climate change that could result in similar fingerprints.

DISCUSSION AND OUTLOOK

The evidence is abundant and clear that there were changes in the AMOC over the deglaciation that can explain the climate oscillations during this period. Similarly, it is clear that at least some of the Heinrich events were associated with significant changes in the AMOC. Although it has been difficult to directly document changes in the AMOC over the shorter D-O events, there is now evidence supporting AMOC changes over most of these oscillations.

Here, I have reviewed only the evidence for changes in ocean circulation associated with the abrupt climate changes of the last glacial cycle. I have not attempted to discern whether AMOC changes were the ultimate cause of the abrupt climate changes, a strong feedback on Northern Hemisphere climate change, or simply a response to strong climate changes in the Northern Hemisphere that do not require changes in the AMOC. Although most of the evidence is consistent with the long-standing hypothesis that the Younger Dryas cold event was caused by the routing of glacial meltwater into the North Atlantic (Carlson et al. 2007), the relationships among glacial melt, Heinrich events, Heinrich stadials, and Heinrich-related AMOC changes appear to be more complex. Evidence is strengthening that the weaker AMOC during Heinrich stadials may have triggered ice-sheet instability rather than being caused by it (Álvarez-Solas et al. 2011, 2013; Gutjahr & Lippold 2011; Marcott et al. 2011; Stanford et al. 2011).

It is even more challenging to generate records using proxies and sediments that can resolve ocean circulation changes with the same temporal resolution necessary to establish the sequence of events and understand the ultimate cause of the D-O events. On these very short timescales, one needs to carefully establish whether there may be time lags in the recording of processes by proxies and consider how sedimentary processes can cause offsets in signals from different phases, even in the same sediment core. As with the Heinrich events, the existing data point to complex relationships among meltwater fluxes, AMOC changes, and climate. Barker et al. (2015) showed that the more modest iceberg discharge into the North Atlantic during the stadials systematically lagged behind the stadial coolings, suggesting that the iceberg discharge was not necessary to cause the cooling. Henry et al. (2016) showed a significant lead (200 years) of ocean circulation changes before sea surface temperature changes in the North Atlantic subtropical gyre, which

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they presumed are synchronous with Greenland temperature changes. Although such a lead might support the idea of AMOC changes driving the Greenland temperature cycles, it is unclear why there should be a lag between Greenland temperature and AMOC-related heat transport changes.

Finally, no clear evidence has emerged for ocean circulation changes associated with the shortest D-O events. This could be because the ocean records are incapable of recording the change or, alternatively, because the AMOC did not change during these events and did so only for longer D-O events. If we had the quality records to know that the AMOC did change for all D-O events, that would bolster the idea that the AMOC is a necessary component of the cause of the D-O events, rather than a response or positive feedback to changes that have their origin elsewhere.

SUMMARY POINTS

1. There is evidence for a shallower and weakened Atlantic meridional overturning circulation (AMOC) during the Last Glacial Maximum (LGM), and perhaps a further weakening during Heinrich stadial 1 that was followed by a transition to a modern-like AMOC and a return to a circulation state similar to that of the LGM during the Younger Dryas.
2. There is evidence that during periods when the AMOC is strong, it weakens around the time of Heinrich events (thought to be discharges of glacial ice into the North Atlantic). However, there is also evidence that a weaker AMOC may trigger ice-sheet instability and input of glacial melt to the North Atlantic rather than the other way around.
3. There is now evidence for AMOC changes for most of the Dansgaard-Oeschger events. However, more high-resolution records are needed to fully document the nature and extent of the circulation changes. As with the Heinrich events, the relationships between freshwater input, AMOC, and climate changes are not yet fully understood.
4. Surface temperature and precipitation patterns are consistent with the modeled climate fingerprint for AMOC variability over the last glacial cycle.

DISCLOSURE STATEMENT

The author is not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

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