

Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene

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Received 5 May 2003; revised 21 October 2003; accepted 7 November 2003; published 21 April 2004.

[1] Geochemical profiles from the North Atlantic Ocean suggest that the vertical $\delta^{13}\text{C}$ structure of the water column at intermediate depths did not change significantly between glacial and interglacial time over much of the Pleistocene, despite large changes in ice volume and iceberg delivery from nearby landmasses. The most anomalous $\delta^{13}\text{C}$ profiles are from the extreme interglaciations of the late Pleistocene. This compilation of data suggests that, unlike today (an extreme interglaciation), the two primary sources of northern deep water, Norwegian-Greenland Sea and Labrador Sea/subpolar North Atlantic, had different characteristic $\delta^{13}\text{C}$ values over most of the Pleistocene. We speculate that the current open sea ice conditions in the Norwegian-Greenland Sea are a relatively rare occurrence and that the high- $\delta^{13}\text{C}$ deep water that forms in this region today is geologically unusual. If northern source deep waters can have highly variable $\delta^{13}\text{C}$, then this likelihood must be considered when inferring past circulation changes from benthic $\delta^{13}\text{C}$ records. *INDEX TERMS:* 1050 Geochemistry: Marine geochemistry (4835, 4850); 1620 Global Change: Climate dynamics (3309); 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 4267 Oceanography: General: Paleoceanography; 4806 Oceanography: Biological and Chemical: Carbon cycling; *KEYWORDS:* paleoceanography, North Atlantic Deep Water, Pleistocene

Citation: Raymo, M. E., D. W. Oppo, B. P. Flower, D. A. Hodell, J. F. McManus, K. A. Venz, K. F. Kleiven, and K. McIntyre (2004), Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene, *Paleoceanography*, 19, PA2008, doi:10.1029/2003PA000921.

1. Introduction

[2] One of the major uncertainties concerning future climate change is the response of ocean thermohaline circulation to the redistribution of heat and moisture at Earth's surface. In particular, many climate models predict that enhanced temperatures and runoff in the Arctic-North Atlantic region could shut down, or severely weaken, deep water convection in the Norwegian-Greenland Sea (NGS) region and that changes here could have far-reaching effects on regional and global climate [Stocker and Wright, 1991; Manabe and Stouffer, 1993; Rahmstorf and Ganopolski, 1999; Ganopolski and Rahmstorf, 2001]. The belief that the NGS region is climatically sensitive and capable of propagating climate signals around the globe finds further support from geologic records of past climate and oceanographic

change. These studies describe the thermohaline circulation of the North Atlantic as a conveyor belt whose strength (or volume transport) is driven primarily by surface water property changes in the NGS [Broecker and Denton, 1989; Imbrie *et al.*, 1993]. Today, between 5 and 6 Sv of deep water flows over the sills east and west of Iceland and entrains a further 7–8 Sv of water just south of the sills [Dickson *et al.*, 1990]. These overflows thus form the greatest volumetric component of NADW (total volume flux estimated at 20 Sv). This water then flows as a deep western boundary current to the Southern Hemisphere and eventually to the other deep basins of the world. During glacial periods, convection in the NGS region is believed to have been greatly reduced by the freshening of surface water owing to glacial runoff [Broecker and Denton, 1989]. Correspondingly, the volume of NADW produced decreased, offset partially by intermediate water formation in the subpolar North Atlantic south of Iceland [e.g., Labeyrie *et al.*, 1987; Duplessy *et al.*, 1988; Oppo and Lehman, 1993; Sarnthein *et al.*, 1994; Keigwin and Lehman, 1994; Fronval *et al.*, 1995; Dokken and Jansen, 1999].

[3] Primarily on the basis of studies of the last glacial-interglacial (G-I) cycle, this view of an ocean circulation “conveyor” which is strong during interglaciations, weak during glaciations, and weaker still during episodes of dramatic ice sheet calving which occur on millennial time-scales, has dominated our view of the Pleistocene epoch [Boyle and Keigwin, 1987; Raymo *et al.*, 1990; de Menocal *et al.*, 1992; Oppo and Lehman, 1995; Oppo *et al.*, 1995; Vidal *et al.*, 1997; McIntyre *et al.*, 1999; Flower *et al.*, 2000;

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Table 1. Site Information

Site	Depth, m	Latitude/Longitude	Sedimentation Rate	Environment	Data Reference/Laboratory ^a
982	1145	58N, 16W	2.6	pelagic	<i>Venz et al.</i> [1999]; <i>Venz and Hodell</i> [2002]/UF
984	1650	61N, 24W	12.5	drift	this paper/WHOI, MIT
983	1983	60N, 24W	12.9	drift	<i>Raymo et al.</i> [1998]/MIT; <i>Kleiven et al.</i> [2003]/UF; <i>McIntyre et al.</i> [1999]/UCSC; this paper/UF, WHOI
980	2169	55N, 15W	10.3	drift	<i>Oppo et al.</i> [1998]/WHOI; <i>McManus et al.</i> [1999]/WHOI; <i>Flower et al.</i> [2000]/USF
981	2173	55N, 15W	5.2	drift	this paper/MIT
552	2301	56N, 23W	1.8	pelagic	<i>Shackleton and Hall</i> [1984]/CAMB
607	3427	41N, 33W	4.1	pelagic	<i>Raymo et al.</i> [1989]; <i>Ruddiman et al.</i> [1989]/LDEO
846	3307	3S, 91W	3.7	pelagic	<i>Mix et al.</i> [1995]/ORE

^aLaboratory codes: UF, University of Florida-Gainesville; WHOI, Woods Hole Oceanographic Institution; MIT, Massachusetts Institute of Technology; UCSC, University of California at Santa Cruz; USF, University of South Florida, St. Petersburg; CAMB, Cambridge University; LDEO, Lamont-Doherty Earth Observatory; ORE, University of Oregon, Corvallis.

Kleiven et al., 2003]. Both data studies and modeling efforts have led to an almost canonical belief that thermohaline convection in the North Atlantic/Norwegian-Greenland Sea is not only highly sensitive to the mean climate state of the region and the amount of glacial runoff but that the response of the “conveyor” then amplifies and propagates climate signals throughout the ocean-atmosphere system [e.g., *Stocker and Wright*, 1991; *Crowley*, 1992; *Imbrie et al.*, 1993; *Mikolajewicz et al.*, 1997; *Paillard and Cortijo*, 1999; *Hall and Stouffer*, 2001; *Ganopolski and Rahmstorf*, 2001].

[4] In this paper we present a compilation of both new and previously published data from the subpolar North Atlantic that allows us to examine the Pleistocene history of North Atlantic Deep Water formation. This compilation takes advantage of recent data from ODP Leg 162 which drilled closely spaced sites in the path of NGS overflows. By putting these records together on a common chronology, a fuller picture of the vertical and horizontal geochemical gradients of the region can be obtained. Likewise, since these records span the Pleistocene Epoch, we can examine whether the “Mid-Pleistocene Revolution” in ice volume and frequency of variation [e.g., *Berger and Jansen*, 1994] was associated with changes in the thermohaline circulation of the region as inferred from $\delta^{13}\text{C}$.

2. Methods and Stratigraphy

[5] Data spanning 27 G-I cycles were examined from a suite of deep-sea sites ranging in water depth from 1000 to 3500 m (Table 1). For each of 27 peak interglacial and 27 peak glacial intervals of the Pleistocene, spanning marine isotope stages (MIS) 1–64, we generate a vertical $\delta^{13}\text{C}$ profile from the subpolar North Atlantic on the basis of stable isotope records from seven DSDP/ODP sites (all data described in this study are available electronically from the NOAA World Data Center for Paleoclimatology at <http://www.ngdc.noaa.gov/paleo/paleo.html>; in addition, the data and supplementary figures can be found at http://www.raymo.com/maureen/climate_archives.php). The locations

of the sites (Figure 1/Table 1/Supplementary figure S1¹) are ideally suited to monitor the various components of NADW (indicated by arrows on Figure 1), including the NGS overflows and Labrador Seawater (LSW). Sites 980 and 981 are collocated in the path of the NGS Wyville Thomson Ridge Overflow (WTRO), while Sites 983 and 984 sit in the path of the Iceland Sea Overflow Water (ISOW). Each of these sites is located on an actively forming sediment drift where normal pelagic sedimentation rates are enhanced by the horizontal advection of fine (<63 μ) material suspended in a nepheloid layer associated with currents spilling over the Iceland-Faeroe-Scotland Ridge (sill depths given in Figure 1). Hence these four sites record physical and chemical changes in the downwelling limb of the global thermohaline “conveyor belt” which originates in the NGS.

[6] Three other sites, 552, 982, and 607, are located out of the direct path of the clay-laden NGS overflows in areas of normal pelagic sedimentation rates (Table 1). The two shallower sites (552 and 982) are influenced primarily by LSW [*Sy et al.*, 1997], while Site 607 at 3426 mbsl is in the path of the deep NADW western boundary current as it exits the northernmost Atlantic. Unlike the drift sites, millennial-scale variability is not easily resolved at these sites (see mean sedimentation rates in Table 1). We proceed on the assumption that, at any one time, each of these sites will volumetrically integrate the relative influence (at that site) of waters originating from the NGS, the open North Atlantic including the Labrador Sea, the Southern Hemisphere, and the Mediterranean Sea. By contrast, we do not assume that the chemical and physical characteristics of the various water masses have remained constant over the Pleistocene and will, in fact, argue that some source area characteristics have likely undergone dramatic shifts.

¹Auxiliary material is available at <ftp://ftp.agu.org/apend/pa/2003PA000921>

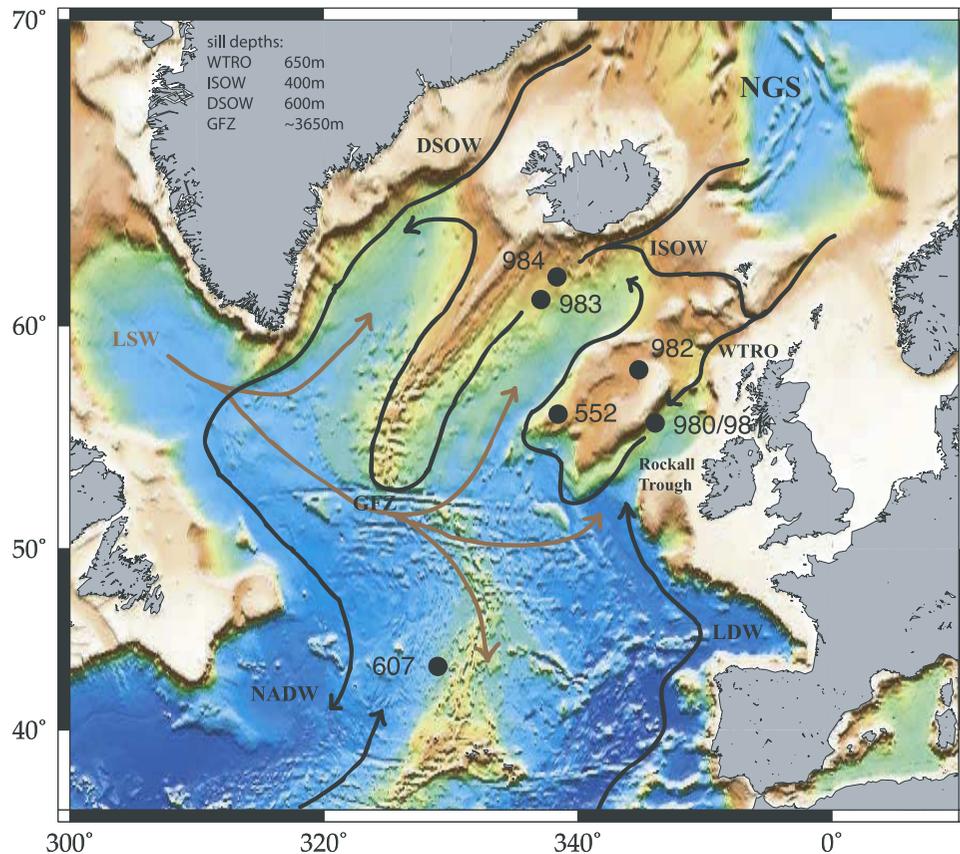


Figure 1. Locations of cores used in study. Paths of major deep water flows are indicated by arrows. Abbreviations not defined in text are: DSOW, Denmark Strait Overflow Water; LDW, Lower Deep Water; GFZ, Gibbs Fracture Zone.

[7] Initially, the $\delta^{18}\text{O}$ record from each site, measured on benthic foraminifera, was correlated to the chronology of Shackleton *et al.* [1990] (Figure 2). No systematic $\delta^{18}\text{O}$ offsets were observed between data sets run at different laboratories (Table 1), although such offsets can typically be up to 0.3‰ [Ostermann and Curry, 2000]. Because different sites and/or time intervals had widely varying sample spacing (ranging from a few hundred years to, for most intervals, a few thousand years), we linearly interpolated all the records to a uniform 3 kyr spacing (approximate average sampling interval of most records). After smoothing each record with a 3-point running mean (primarily for clarity of figures), each record was correlated to the target timescale (Figure 2) using Analyseries [Paillard *et al.*, 1996] (unsmoothed data can be viewed at http://www.raymo.com/maureen/climate_archives.php and in auxiliary material). Most records had a correlation coefficient of between 0.8 and 0.9 to the TARGET curve and the average number of age control points (tie lines) used for each 1.8 Myr long record was 80. Certain intervals of noticeably poorer correlation (e.g., the 983 and 984 records within the 1.4–1.6 Myr range) are due to the paucity of benthic foraminifera at those times. In addition, Site 552

has a drilling hiatus from ~ 1.3 to 1.6 Ma. All the other records are spliced high-resolution benthic isotope records with no obvious hiatuses.

[8] Correlation of $\delta^{18}\text{O}$ records inevitably involves a certain amount of subjectivity, and in this study we were relatively conservative in our use of tie points (one used every ~ 22 kyr, on average). This approach avoided rapid short-term sedimentation rate variations and was also least likely to bias the outcome of our analysis of vertical $\delta^{13}\text{C}$ gradients. To avoid bias, each record was also correlated independently to the target timescale; e.g., the North Atlantic sites were not correlated to each other in an iterative fashion. This also has the advantage of presenting a suite of data records plotted to a well-documented and widely accepted timescale (a simple algorithm could easily convert all these records to a different timescale if desired).

[9] After putting the records on the Shackleton *et al.* [1990] chronology, we identified peak glacial and peak interglacial intervals as indicated by $\delta^{18}\text{O}$, an ice volume/temperature proxy. These time slices (indicated by blue and yellow bars in Figures 2 and 3) are each 10,000 years long with the exception of the Holocene time slice, which

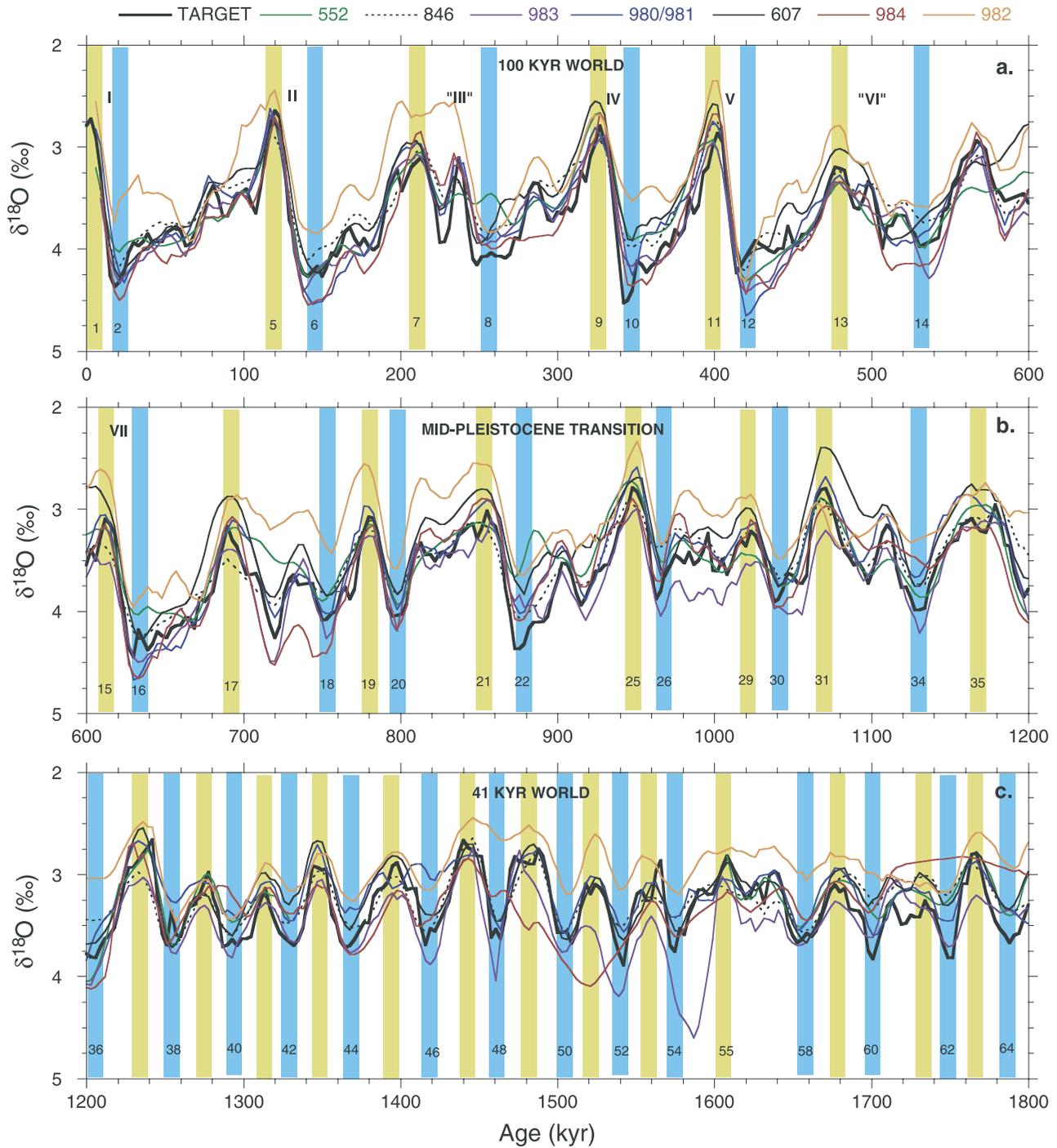


Figure 2. Benthic $\delta^{18}\text{O}$ records from seven Atlantic sites and one Pacific site, 846, correlated to the Shackleton *et al.* [1990] target record (<http://delphi.esc.cam.ac.uk/coredata/v677846.html>). Terminations and marine isotope stages are labeled. Terminations III and VI are in quotation marks to indicate that they do not meet the formal definition of a termination [see Raymo, 1997]. Blue and yellow bars defined as described in text.

includes only the last 8 kyr. While the $\delta^{18}\text{O}$ records include a small number of measurements made on genera other than *Cibicides*, the benthic $\delta^{13}\text{C}$ data (Figure 3) include only *Cibicides* data (with the vast majority being *wuellerstorfi* spp.), the species which has been shown to most accurately

record bottom water carbon chemistry [e.g., Curry *et al.*, 1988; McCorkle *et al.*, 1990; McCorkle and Keigwin, 1994; Tachikawa and Elderfield, 2002]. The possibility that this genus does not accurately record bottom water $\delta^{13}\text{C}$ within certain environments is discussed later in this paper.

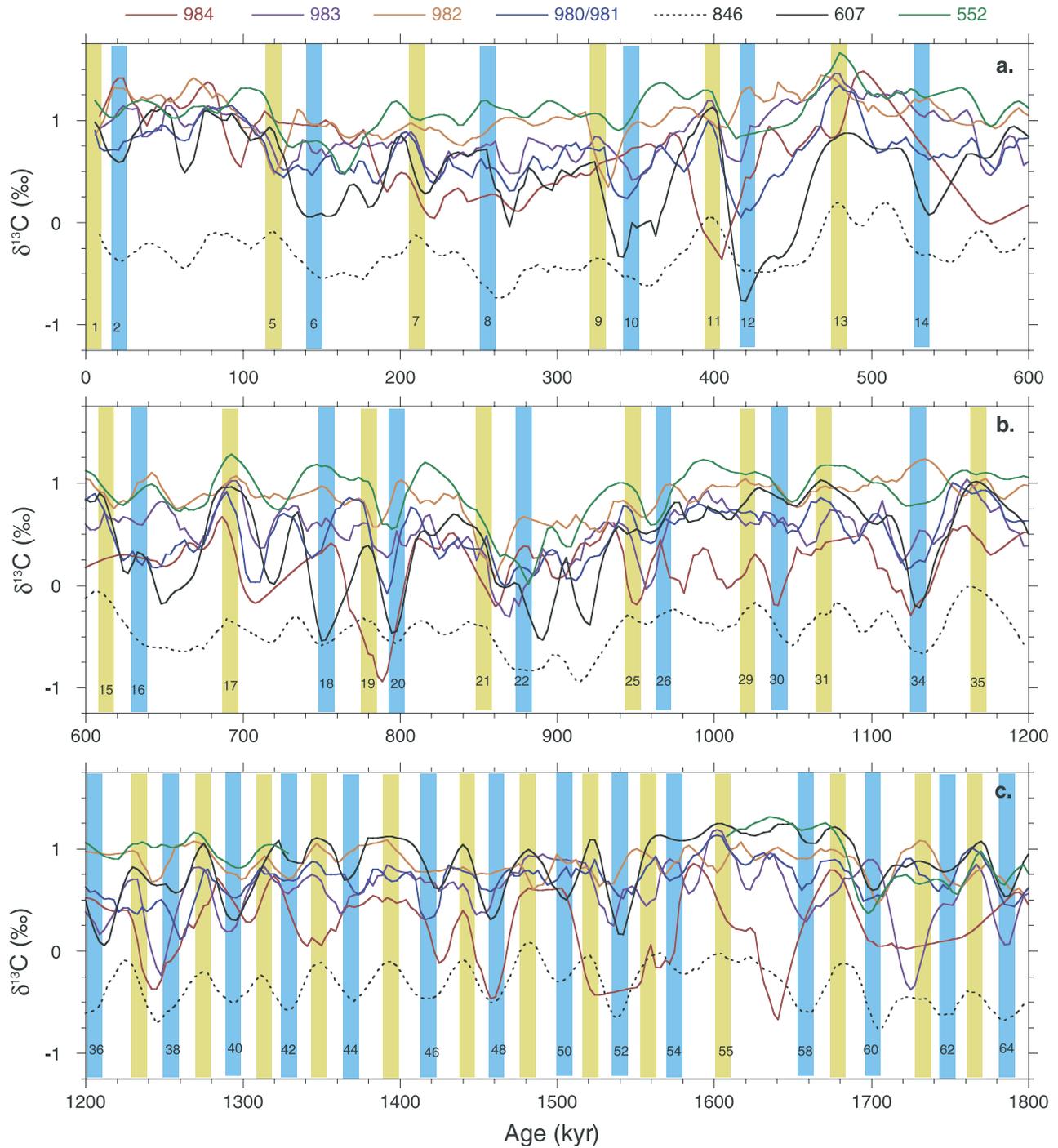


Figure 3. Benthic $\delta^{13}\text{C}$ data plotted to age on the basis of $\delta^{18}\text{O}$ correlations shown in Figure 2. All data shown on the basis of the genus *Cibicidoides*, with most analyses on *C. wuellerstorfi*. Blue and yellow bars indicate peak glacial and interglacial stages, respectively.

Clearly, and not unexpectedly, the $\delta^{13}\text{C}$ data show much greater spatial variability between sites than oxygen isotope data.

[10] To determine whether there are spatially or temporally coherent patterns in the $\delta^{13}\text{C}$ data, we reconstruct average vertical $\delta^{13}\text{C}$ profiles (depth transects) for each

glacial and interglacial maxima of the Pleistocene [e.g., after Curry and Lohmann, 1983]. We did this by taking the average value of the actual (not smoothed) *Cibicidoides* $\delta^{13}\text{C}$ data within each 10,000 time slice (as defined above). In some cases, no *Cibicidoides* data exist within a particular core within a particular 10 kyr time slice; in other cases, up

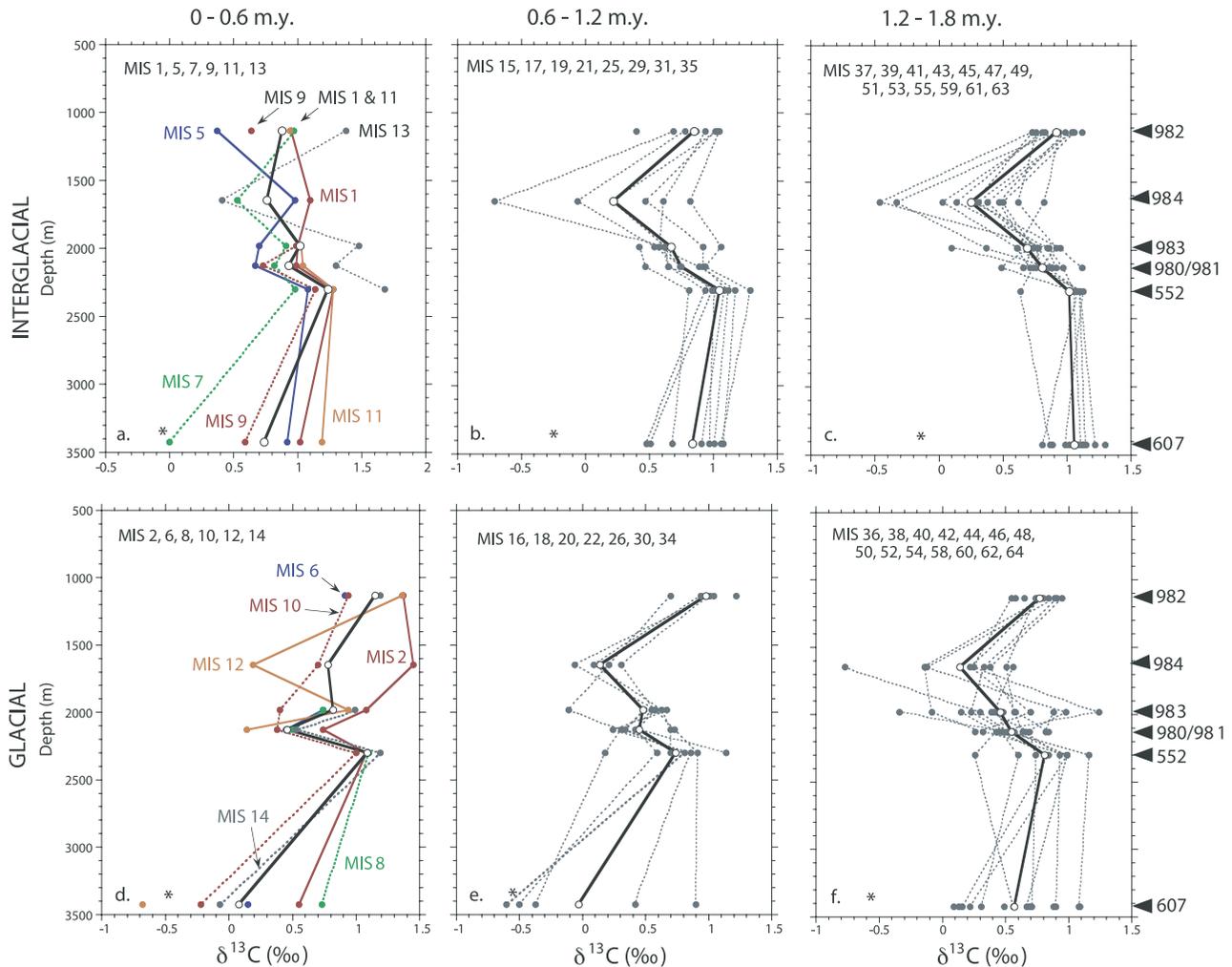


Figure 4. Average of $\delta^{13}\text{C}$ values within each 10,000 year time slice shown in Figure 3. Dashed and colored lines represent individual marine isotope stages named on panel. Solid line with open circles is mean vertical $\delta^{13}\text{C}$ gradient for time interval represented by that panel. Starred value is average $\delta^{13}\text{C}$ value for correlative Pleistocene time interval at Pacific Site 846.

to a few dozen analyses were available. Typically, 2–4 analyses of multiple *Cibicoides* specimens were present for each time slice.

3. Results

[11] Fifty-four vertical $\delta^{13}\text{C}$ reconstructions, each representing the average $\delta^{13}\text{C}$ composition of six sites within a correlative 10,000 year time slice, are shown in Figure 4. These data present 54 “snapshots” of the average spatial variability of bottom water $\delta^{13}\text{C}$ in the subpolar North Atlantic on orbitally forced G-I timescales. Each dashed (or colored) line represents an individual time slice (marine isotope stage), and the solid black line in each panel is the average vertical $\delta^{13}\text{C}$ gradient for the early, middle, or late Pleistocene interval shown in that panel. As discussed above, incomplete profiles are due to a lack

of data at some sites within some intervals (we did not connect dots across missing data). Much shorter piston core records near Site 980 (NA807022; Vidal *et al.* [1997]), Site 983 (EW9302-8JPC; Oppo *et al.* [1997, 2001]), and Site 984 (EW9302-14JPC; Oppo *et al.* [1997]) confirm the $\delta^{13}\text{C}$ patterns presented here over the intervals of overlap (back to MIS 6).

[12] At the nondrift sites (552, 982, and 607), normal pelagic sedimentation rates combined with bioturbation prevent clear resolution of millennial scale variability. However, at the higher accumulation rate sites (984, 983, 980, and 981), such variability is apparent [e.g., Raymo *et al.*, 1998; McIntyre *et al.*, 1999; Oppo *et al.*, 2001; Kleiven *et al.*, 2003]. By averaging the data within a 10,000 year time slice, we may be creating a “mean state” that does not really exist. For instance, high-resolution studies of the last glacial interval [Sarnthein *et al.*, 1994; Keigwin

and Lehman, 1994] suggest NADW production was suppressed to an even greater extent during the exceptionally cold Heinrich events. Similarly, modern data show that deep water characteristics vary in the North Atlantic on even decadal timescales (e.g., the great salinity anomaly) [Sy *et al.*, 1997; Dickson, 1997]. It may be that even millennial-scale data are averages of variability occurring at higher frequencies (centuries, for instance) and thus are also recording a “mean state” that does not really exist. For the purposes of this study we emphasize the obvious differences (or lack thereof) in the mean circulation state of the North Atlantic as it responds to orbital-scale glacial and interglacial modes of the climate system.

[13] From Figure 4 it is obvious that consistent temporal and spatial patterns are observed in the $\delta^{13}\text{C}$ data. During interglacial maxima of the early and middle Pleistocene (1.8–0.6 Myr, Figures 4b and 4c), the nondrift Sites 982, 552, and 607 show generally more positive $\delta^{13}\text{C}$ values, while the drift sites (983, 984, and to a lesser degree 980/981) show more negative values. Since each time slice represents the average $\delta^{13}\text{C}$ value over a 10,000 year period and because so many time slices show a similar pattern, we believe it is highly unlikely that this result is due to correlation errors. Although not presented in this paper, glacial onset and deglacial profiles also show a similar structure.

[14] The pattern of depleted $\delta^{13}\text{C}$ values between more enriched values at 1145 and 2301 m is also seen in the late Pleistocene interglacial profiles of MIS 7 and 13. It is probably not a coincidence that these particular interglaciations are also the only two of the last 0.6 Myr that do not attain the extremely depleted $\delta^{18}\text{O}$ values (warm climates?) characteristic of MIS 1, 5, 9, and 11. Of the two other complete late Pleistocene interglacial profiles (MIS 1 and 5), MIS 1 seems to be most in agreement with modern observations of relatively uniform $\delta^{13}\text{C}$ values throughout the water column of the subpolar North Atlantic.

[15] For early to mid-Pleistocene glacial intervals (Figures 4e and 4f), the $\delta^{13}\text{C}$ observed at middepths (between 1150 and 2300 m) is likewise more negative than that observed deeper (at 552) and shallower (at 982). In the latest Pleistocene (Figure 4d) the MIS2 profile appears to be the most anomalous of the entire 1.8 Myr, being the only glacial with intermediate depth $\delta^{13}\text{C}$ (at 1650 m) more enriched than sites deeper and shallower. None of the late Pleistocene glacial intervals conform neatly to the accepted glacial paradigm of two water mass layers with high- $\delta^{13}\text{C}$ GNAIW overlying low- $\delta^{13}\text{C}$ deep water (typically assumed to be of Southern Ocean origin). This is primarily due to the consistently enriched $\delta^{13}\text{C}$ values observed at Site 552 (2300 m).

[16] Taken as a whole, the two most significant differences between glacial and interglacial profiles seem to be: (1) the gradually more negative glacial $\delta^{13}\text{C}$ values observed at the deepest site, 607, over the course of the Pleistocene; and (2) the lack of a pronounced middepth $\delta^{13}\text{C}$ minima in the warmest late Pleistocene interglacial intervals for which a complete profile exists (e.g., MIS 1 and 5; Figure 4a). Most surprisingly, the early Pleistocene “41 kyr world,” here studied back to 1.8 Ma but which

extends beyond 2.5 Ma, shows a nearly identical pattern for the vertical water mass structure in the subpolar North Atlantic for both glacial and interglacial intervals (Figures 4, 5d, and 5e). Thus, despite the growth and decay of continent-spanning ice sheets and the pronounced influx of icebergs to the open North Atlantic during glacial intervals [e.g., Shackleton and Hall, 1984; Raymo *et al.*, 1986; Ruddiman *et al.*, 1986, 1989], the vertical $\delta^{13}\text{C}$ gradients at these sites appear to be relatively insensitive to this G-IG range in the mean climate state.

[17] Finally, note that mean ocean $\delta^{13}\text{C}$ changes occurring over the Pleistocene [e.g., Raymo *et al.*, 1997] would not affect the relative $\delta^{13}\text{C}$ differences between sites (e.g., the shape of the profiles). However, such whole ocean changes (caused by transfer of carbon into and out of the biosphere) could account for some of the absolute $\delta^{13}\text{C}$ variability observed over time. In Figures 5c–5e the mean glacial profiles are not corrected for G-I change in whole ocean $\delta^{13}\text{C}$ in each of the three intervals studied. However, the average mean ocean $\delta^{13}\text{C}$ change (MOC) for each interval, calculated using the average G-I $\delta^{13}\text{C}$ difference observed at Pacific Site 846, is noted on figures.

4. Origin of Low- $\delta^{13}\text{C}$ Water

[18] What is the origin of the low- $\delta^{13}\text{C}$ values observed at middepths? Do these profiles accurately depict bottom water chemistry and hence local hydrographic conditions over the Pleistocene? In particular, is it possible that the $\delta^{13}\text{C}$ record from Site 984, the site with the most pronounced negative excursions, is influenced by some analytical or geochemical artefact? While the removal of the 984 data would not eliminate the middepth “sandwich” layer of low- $\delta^{13}\text{C}$ water, it would certainly make it less pronounced.

[19] It has been shown that a high flux of organic matter to the sediment-water interface can create a microenvironment of atypically low $\delta^{13}\text{C}$ that can be incorporated into the test of epifaunal benthic foraminifera such as *C. wuellerstorfi* [Mackensen *et al.*, 1993]. Given the high sedimentation rates and fine-grained detrital flux associated with nepheloid layers, the possibility of such geochemical artefacts must be considered here. We offer three lines of evidence that lead us to believe that the $\delta^{13}\text{C}$ signals are true bottom water geochemical signals; first, the drift sites (which are generally lower in $\delta^{13}\text{C}$) are characterized by similar organic carbon contents as the nearby and more C^{13} -enriched nondrift sites (average value in drift sites is $\sim 0.2\%$; Jansen *et al.* [1996]). Thus no obviously larger flux of phytodetritus is observed, or inferred, for the drift sites relative to the nondrift sites.

[20] Second, analyses of living and core top *C. wuellerstorfi* recovered from a nearby Feni Drift site accurately reflect bottom water $\delta^{13}\text{C}$ [Tachikawa and Elderfield, 2002]. In that study the mean $\delta^{13}\text{C}$ value of *C. wuellerstorfi* measurements at the drift site (Site C) is slightly higher than (but well within 1σ of) the measured bottom water $\delta^{13}\text{C}$ value. Likewise, at a nearby nondrift site (Site A), *C. wuellerstorfi* values are again within 1σ of the measured bottom water $\delta^{13}\text{C}$ value. In a similar study undertaken on the eastern North American continental shelf, a limited

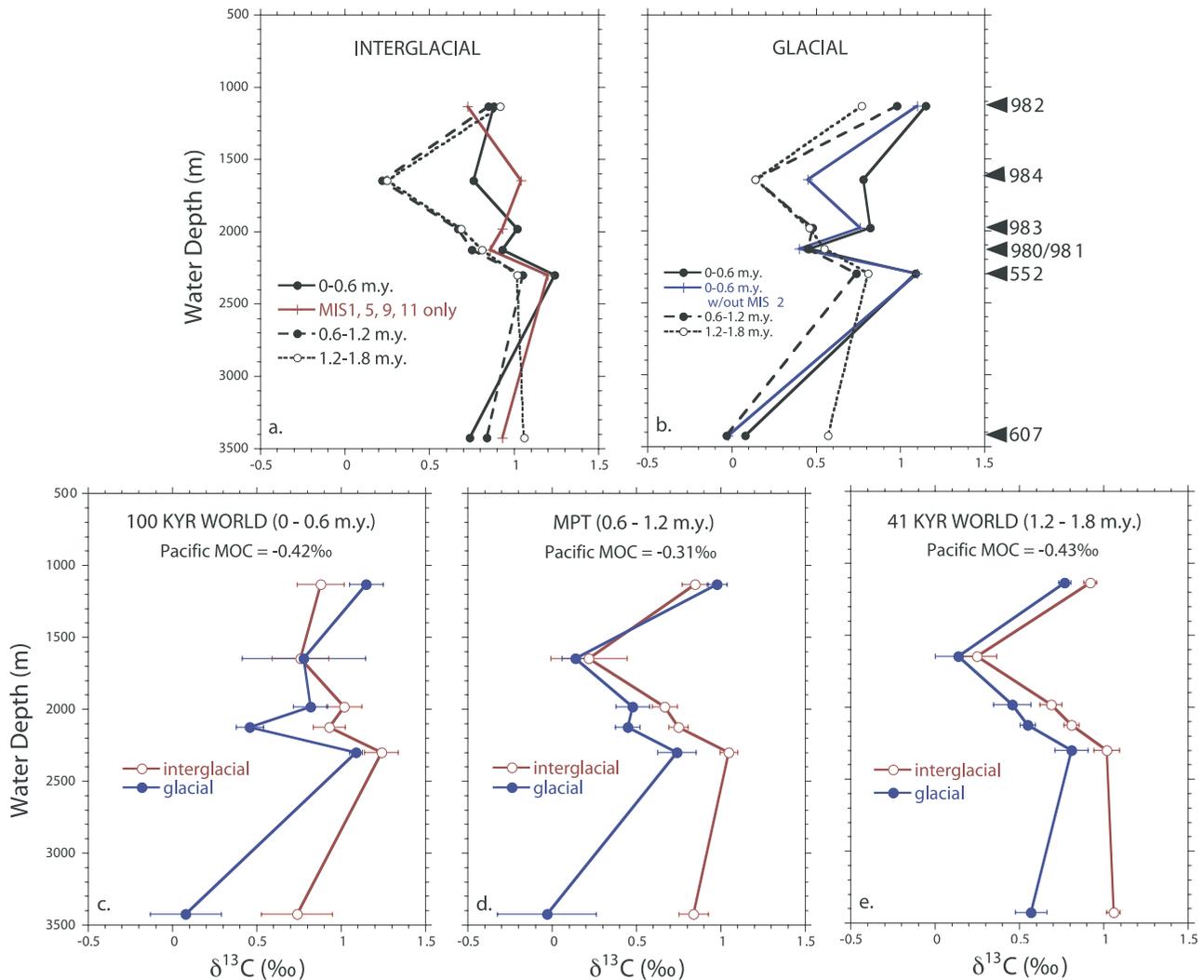


Figure 5. Comparison of mean vertical $\delta^{13}\text{C}$ gradients for early, middle, and late Pleistocene interglacial (a) and glacial (b) intervals. Lower panels (c)–(e) compare glacial with interglacial profiles within each of these three intervals. The error bars represent the standard error of the profiles shown in each panel of Figure 4. Glacial profiles are not corrected for a presumed G-I mean ocean $\delta^{13}\text{C}$ change (MOC). The magnitude of such changes are uncertain but are generally assumed to be a few tenths of a per mil. For each interval in Figures 5c–5e the average G-I mean ocean change (MOC) observed at the Pacific reference site (846) is noted.

number of *C. wuellerstorfi* measurements also suggest that this species is an excellent recorder of bottom water $\delta^{13}\text{C}$ when multiple specimens are analyzed [see *Corliss et al.*, 2002, Figure 6b, Table 4]. This study suggests that our biggest concern should be with any horizon with only a single specimen analysis since analytical and/or natural variance could bias our reconstructed gradients. In our study, multiple specimens were used in most analyses, and analyses were further averaged over a 10,000 year time slice. Overall, these studies do not support a systematic bias in *Cibicides* $\delta^{13}\text{C}$ data that would result in unusually depleted carbon isotope signals at drift sites.

[21] Our third line of evidence that the vertical $\delta^{13}\text{C}$ profiles are accurate bottom water signals comes from the

corresponding $\delta^{18}\text{O}$ profiles (generated by the same methodology and which could not be influenced by a “Mackensen effect”). These profiles also show a middepth layer of isotopically distinct water, suggesting that a unique water mass, slightly enriched in $\delta^{18}\text{O}$, is present at the drift sites (983 and 984 in particular; Figure 6). It seems unlikely that any analytical or sedimentary process could alter $\delta^{18}\text{O}$ values; we feel it is more likely that the $\delta^{18}\text{O}$ of the water is reflecting the temperature, salinity, and sea ice and brine rejection processes in the place of origin (discussed below).

[22] If one accepts that the stable isotope data accurately reflect bottom water chemistry, then what is the origin of the middepth low- $\delta^{13}\text{C}$ water that characterizes most of the Pleistocene? Can a sensible hydrographic model be found

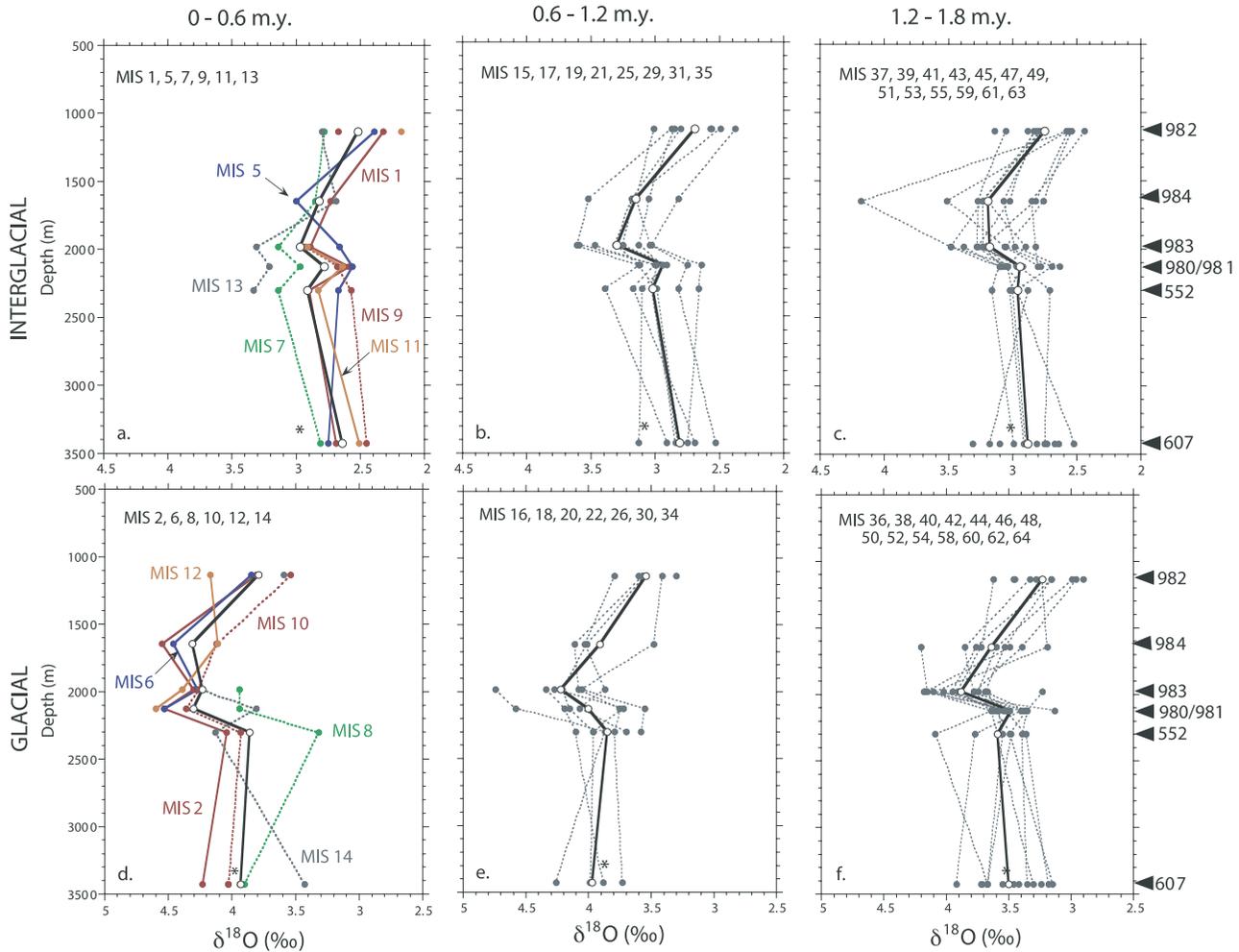


Figure 6. Average $\delta^{18}\text{O}$ values for time slices shown in Figure 3. Dashed and colored lines represent individual marine isotope stages named on panel. Solid line with open circles is mean vertical $\delta^{18}\text{O}$ gradient for time interval represented by that panel. Starred value is average $\delta^{18}\text{O}$ value for correlative time interval at Pacific Site 846. Note that interlaboratory $\delta^{18}\text{O}$ calibrations, which are more difficult than for $\delta^{13}\text{C}$, are typically not accurate to better than 0.1–0.2‰.

to explain the data profiles shown in Figures 4–6? We first considered whether a greatly enhanced oxygen minimum layer could have characterized the northeast subpolar Atlantic over most of the Pleistocene. The depth and magnitude of the middepth $\delta^{13}\text{C}$ reduction ($>0.5\text{‰}$) suggests it would need to be a layer over 500 m deeper and significantly more depleted in oxygen than is observed in this region today [*Geochemical Ocean Sections Study*, 1981] (<http://whpo.ucsd.edu/data/onetime/atlantic/a24/index.htm>). However, such a strong minimum zone would imply a dramatic increase in productivity over the region for which no sedimentary evidence is found. In addition, the depth of the oxygen minimum zone in the global ocean is controlled primarily by the depth of the photic zone and thus is typically between 500 and 1000 m depth. Given these observations and the very positive $\delta^{13}\text{C}$ values observed at 1100 m (Site 982), we do not believe that an oxygen minimum zone is responsible for the middepth $\delta^{13}\text{C}$ depletion. Indeed,

it seems unlikely that the drift sites with their vigorous contourite currents could maintain low oxygen conditions driven by a downward flux of organic matter.

[23] The second possibility we considered was that the low- $\delta^{13}\text{C}$ water originates from the Mediterranean Sea. We believe this unlikely because the shallowest site, 982, closest to the depth of Mediterranean Outflow Water (800–1200 m) always exhibits high- $\delta^{13}\text{C}$ values. In addition, studies of past MOW variability show this water mass to be characterized by enriched, rather than depleted, $\delta^{13}\text{C}$ values in the Pleistocene [e.g., *Vergnaud-Grazzini et al.*, 1989; *Schonfeld and Zahn*, 2000].

[24] A third possibility is that low- $\delta^{13}\text{C}$ Upper Circumpolar Deep Water (UCDW) could have extended farther north and been more volumetrically significant in the past. We rule this possibility unlikely given that, over most of the Pleistocene, the $\delta^{13}\text{C}$ observed in the Caribbean Sea (Site 502; *Oppo et al.* [1995]) is more positive than observed at

Site 984. This is a marginal basin ventilated over a sill (1800 m) at the approximate depth of Site 984 (1650 m) and the depth one would expect UCDW to occupy. Hence a Southern Hemisphere water mass would leave more, not less, of a geochemical signature in the Caribbean relative to sites further north. Similarly, shoaling of deep AABW, a conclusion one might come to if working with fewer records, seems unlikely given the more positive $\delta^{13}\text{C}$ signals observed at our deeper sites, 552 and 980/981.

[25] Finally, we considered whether the low- $\delta^{13}\text{C}$ water could have originated in the Norwegian-Greenland Sea and flowed into North Atlantic as WTRO and ISOW. In our study the most negative $\delta^{13}\text{C}$ water is typically observed at Site 984, the shallowest drift site closest to the NGS source, and therefore the site likely to have the least amount of mixing with water from within the subpolar North Atlantic. By contrast, Site 980/981 in the Rockall Trough is significantly influenced by Lower Deep Water today (Figure 1; McCartney [1992]) as well as by NGS overflows. Not only are the drift sites in the direct path of overflows today, active drift sedimentation has occurred throughout the Pleistocene at these sites as evidenced by elevated sedimentation rates (higher sedimentation rates are observed at Sites 983 and 984 than at 980/981, again consistent with these sites being closer to, and more influenced by, the NGS overflows; Table 1).

[26] On the basis of the above observations, we conclude that the most likely origin of the middepth $\delta^{13}\text{C}$ minimum is via overflows from the NGS. Benthic foraminiferal $\delta^{13}\text{C}$ data suggest that low- $\delta^{13}\text{C}$ values occurred in the NGS during the penultimate deglaciation and into the beginning of the last interglaciation (MIS 5e) [Duplessy and Shackleton, 1985; Oppo and Lehman, 1995; Fronval and Jansen, 1997; Bauch et al., 2000]. These transient low- $\delta^{13}\text{C}$ intervals have been attributed to the combined influence of a reduction in air-sea exchange at higher latitudes (owing to the presence of sea ice) as well as lower- $\delta^{13}\text{C}$ values in the tropical feed waters. This interpretation is further supported by a possible correlation between millennial $\delta^{13}\text{C}$ events in the NGS and middepth subpolar North Atlantic, implying constant exchange between the two basins [Oppo et al., 2001]. At these times, deep water may have formed by brine rejection under sea ice on the continental shelves by processes more analogous to those observed around Antarctica today [e.g., Dokken and Jansen, 1999].

[27] In addition to reduced air-sea isotopic exchange, two other possible sources/causes of low- $\delta^{13}\text{C}$ water in the NGS are (1) transport of Pacific water through the Arctic Ocean and/or (2) runoff from the continents surrounding the Arctic and NGS. Both of these sources are significant contributors to the water mass budget of the Arctic Ocean today, and both are typically depleted in $\delta^{13}\text{C}$ relative to modern NGS and North Atlantic surface water [Erlenkeuser et al., 1999]. Stable isotope studies also show that remineralization of organic matter on shallow continental shelves is an important contributor of depleted $\delta^{13}\text{C}$ to Arctic coastal waters [Polyak et al., 2003].

[28] The modern flux of Bering Strait-Pacific water to the Arctic is estimated to be ~ 0.8 Sv [Coachman and Aagaard,

1998], which is one half to equal the estimated runoff flux [Ekwurzel et al., 2001]. In addition, modern and Holocene studies have shown that the runoff and Pacific fluxes can vary in their relative influence in the Arctic on decadal to millennial timescales [e.g., Schlosser et al., 2002; Ekwurzel et al., 2001]. From these observations we speculate that the relative influence in the NGS of Arctic versus North Atlantic source waters could have changed in the past.

5. Role of the Norwegian-Greenland Sea in Pleistocene Climate Change

[29] Above we propose that the low- $\delta^{13}\text{C}$ water mass observed at middepths in the subpolar North Atlantic over much of the Pleistocene may have originated in the NGS. Lower- $\delta^{13}\text{C}$ values could have been due to enhanced sea ice cover in the region which would have reduced air-sea isotopic exchange as well as reduced exchange of surface waters with the open North Atlantic relative to the Arctic. Today, ~ 1 Sv enters the NGS via the Fram Strait relative to about 5 Sv from the open North Atlantic [Bonisch and Schlosser, 1995]. Could this ratio have been significantly different in the past?

[30] Many researchers have presented geologic evidence for more extensive, even permanent, sea ice cover in the NGS during the early to mid-Pleistocene. It has been proposed that the lack of carbonate preservation in this region from about 2.8 to 0.5 Myr ago may have been due to perennial sea ice cover suppressing surface water productivity and causing the reduction of convection and deep water ventilation [e.g., Kellogg, 1980; Baumann and Huber, 1999; Huber et al., 2000]. Only during the peak interglaciations of the latest Pleistocene is a strong inflow of Atlantic surface water documented in the NGS [Jansen et al., 1988; Henrich and Baumann, 1994; Baumann and Huber, 1999].

[31] Studies of siliceous microfossils come to similar conclusions. Examining ODP Sites 907 and 642 and other cores, Koç and Scherer, 1996; Stabell and Koç, 1996] concluded that no production of silica occurred in the surface waters of the NGS in the Pleistocene except for the Holocene and perhaps MIS 11. They inferred the presence of very harsh surface water conditions, most probably year-round sea ice cover. In contrast to the NGS, the diatom production at sites south of Iceland (Sites 919 and 983) follows a typical G/IG pattern with high productivity in interglaciations [Koç and Flower, 1998; Koç et al., 1999]. Thus these investigators conclude that a strong front must have existed near the Greenland-Scotland Ridge during most of the interglaciations of the Pleistocene.

[32] On the basis of their inferences of surface water conditions in the NGS, Henrich and Baumann [1994] proposed that brine formation during winter sea ice growth may have been the dominant mode of deep water production in the NGS. The very limited benthic and planktonic $\delta^{13}\text{C}$ records available from the Pleistocene NGS [e.g., Jansen et al., 1988; Flower, 1998] are typically more negative than $\delta^{13}\text{C}$ values observed at our drift sites,

consistent with water mass mixing and entrainment south of the sills. Likewise, heavy benthic $\delta^{18}\text{O}$ values in the NGS [Henrich and Baumann, 1994] are consistent with the more enriched $\delta^{18}\text{O}$ values observed in our drift sites (Figure 6).

[33] While no unambiguous proxies for sea ice extent have been discovered, we propose that the more uniform water column chemistry observed in the North Atlantic during late Pleistocene interglaciations (Figure 4) may be due to unusually open sea ice conditions in the NGS that allowed open ocean convection, better air-sea isotopic exchange, and more efficient utilization of surface nutrients. Under such conditions, deep water originating from the NGS would be indistinguishable in $\delta^{13}\text{C}$ from deep water originating from open ocean convection in the Labrador Sea or subpolar North Atlantic. By contrast, when the NGS was sea ice covered, dense water forming in this region, possibly through brine rejection, could have been characterized by more depleted $\delta^{13}\text{C}$ values typical of Arctic, rather than North Atlantic, feed waters. Such water, flowing over the southern sills, could explain the middepth $\delta^{13}\text{C}$ minima observed over much of the Pleistocene.

[34] Our finding of similar glacial and interglacial vertical $\delta^{13}\text{C}$ profiles (Figures 4 and 5) suggests that, over the first million years of the Pleistocene (1.8–0.6 Myr), the relative volumetric importance of water originating in the NGS versus the Labrador Sea/subpolar North Atlantic did not change significantly from glacial to interglacial times. Further, since we suggest above that the geochemical signature of NSOW may have changed in the latest Pleistocene (owing to changing surface water conditions in the NGS), it becomes impossible to evaluate the relative flux of the different NADW sources in the late Pleistocene. For example, during late Pleistocene interglaciations, it would be impossible to distinguish LSW from NSOW on the basis of $\delta^{13}\text{C}$ values alone. We can say for certain that the sediment accumulation rates remained elevated at the drift sites throughout the late Pleistocene, even during glacial periods, an indication of active contourite currents with nepheloid layers. Whether such horizontal sediment transport could be maintained by boundary currents circulating within the North Atlantic basins (e.g., in the absence of NGS overflows) poses an interesting physical oceanographic problem.

[35] Note that the $\delta^{13}\text{C}$ at the deepest NADW site, 607, gets progressively more negative, especially during glaciations, over the Pleistocene. Two possible sources for this $\delta^{13}\text{C}$ -depleted water are (1) the middepth water which we propose is originating in the NGS and (2) deep water of southern origin (e.g., AABW). While it may be possible that NSOW water (including Denmark Strait Overflow Water) could flow as density-driven current to depths as great as 3400 m (much like AABW spilling off continental shelves around Antarctica today), other geochemical data do not support a northern source origin. First, the $\delta^{18}\text{O}$ data suggest that the deepest water mass in the North Atlantic is more similar to Pacific deep water (the asterisk in Figure 6 as defined by mean values at Site 846) than to the drift site (middepth) water at times of lowered $\delta^{13}\text{C}$; and second, if an increased flux of $\delta^{13}\text{C}$ -depleted NSOW was responsible for

the deep $\delta^{13}\text{C}$ minima observed in Figure 4e relative to 4f, then we might expect this water mass to exert a relatively greater influence at the shallower depths as well (changing the appearance of the vertical profiles). It may be that the current 12–14 Sv of deep water originating from the NGS is a late Pleistocene “extreme interglacial” anomaly; this region may have been only a minor source of deep water over much of Pleistocene. Indeed, if NGS overflows had been a volumetrically significant source of deep water over much of the Pleistocene, then we would expect to find more evidence for this low- $\delta^{13}\text{C}$ /high- $\delta^{18}\text{O}$ water throughout the subpolar North Atlantic.

6. Conclusions

[36] In summary, our compilation suggests that for most of the Pleistocene, the components of NADW did not vary significantly on glacial-interglacial timescales. The similarity of the glacial and interglacial profiles at intermediate depths suggests that the relative volumetric production of LSW and NGS overflow waters did not change. We suggest the possibility that the Pleistocene glacial $\delta^{13}\text{C}$ signal at Site 607 (and many other deep Atlantic sites) may be more influenced by variations in the production rate of deep water around Antarctica rather than changes in NADW flux or the strength of the “conveyor belt” [e.g., Toggweiler, 1999; Keeling and Stephens, 2001].

[37] While new evidence continues to suggest that NADW production decreased during the LGM and was replaced at depth by SOW [Marchitto *et al.*, 2002], this may not be a useful paradigm for interpreting older records. For much of the Pleistocene the vertical water mass structure of the subpolar North Atlantic (and hence possibly deep water convection centers) appears to have changed little despite dramatic swings in the mean climate state. It may be that the large volumetric flux of NSOW water observed today is an “extreme interglacial” anomaly not typical of most of the Pleistocene. While speculative, the possibility arises that, rather than initiating and then transmitting climate changes around the globe as some models suggest, North Atlantic Deep Water variability played a relatively passive role in global climate changes over much of the Pleistocene. Northern Hemisphere high-latitude climate may be forced by climate variability originating at low latitudes and transmitted via the atmosphere [e.g., Cane, 1998; Cane and Clement, 1999; Wang *et al.*, 1999; Raymo and Nisancioglu, 2003]. Alternatively, changes in atmospheric CO_2 driven by restructuring of the Southern Ocean water column could also be forcing Northern Hemisphere climate via the atmosphere [e.g., Sigman and Boyle, 2000, and references therein; Keeling and Stephens, 2001].

[38] **Acknowledgments.** The authors would like to thank the scientists and crew of ODP Leg 162 for collecting a record-breaking amount of top quality core material. We also thank Chris Charles, Howie Spero, and two anonymous reviewers, whose comments and suggestions greatly improved this paper. MER would like to thank Susie Carter and Kristen Ganley for their help in the early stages of this project and the National Science Foundation for grants OCE-0118005 and OCE-0118001, which supported MER and DWO.

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