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The 8200 year B.P. event in the slope water system, western subpolar North Atlantic

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[1] Stable isotope, trace metal, alkenone paleothermometry, and radiocarbon methods have been applied to sediment cores in the western subpolar North Atlantic between Hudson Strait and Cape Hatteras to reveal the history of climate in that region over the past ~11 kyr. We focus on cores from the Laurentian Fan, which is known to have rapid and continuous accumulation of hemipelagic sediment. Although results among our various proxy data are not always in agreement, the weight of the evidence (alkenone sea surface temperature (SST), $\delta^{18}O$ and abundance of *Globigerinoides ruber*) indicates a continual cooling of surface waters over Laurentian Fan, from about 18°C in the early Holocene to about 8°C today. Alternatively, Mg/Ca data on planktonic foraminifera indicate no systematic change in Holocene SST. The inferred long-term decrease in SST was probably driven by decreasing seasonality of Northern Hemisphere insolation. Two series of proxy data show the gradual cooling was interrupted by a two-step cold pulse that began 8500 years ago, and lasted about 700 years. Although this event is associated with the final deglaciation of Hudson Bay, there is no $\delta^{18}O$ minimum anywhere in the Labrador Sea, yet there is some evidence for it as far south as Cape Hatteras. Finally, although the 8200 year B.P. event has been implicated in decreasing North Atlantic ventilation, and hence widespread temperature depression on land and at sea, we find inconsistent evidence for a change at that time in deep ocean nutrient content at ~4 km water depth.

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1. Introduction

[2] We discuss here stable isotope, alkenone, and trace metal proxy data from sediment cores along the eastern margin of North America from Cape Hatteras to Hudson Strait (Figure 1). These cores all have sedimentation rates that are sufficient to resolve paleoclimate and paleo-ocean change at the millennial to centennial timescale since the last glacial maximum (LGM). The very high sedimentation rates of our cores, as well as large changes in surface and deep ocean physical-chemical properties, provide an exceptionally high signal-to-noise ratio for paleoceanographic data. Our focus is on the Laurentian Fan where we have the most complete records and the best chronology. Large changes are expected in this region because the changing position and transports of the various currents north of the Gulf Stream give the slope waters between the Grand Banks and the Middle Atlantic Bight the most variable sea surface temperature (SST) of anywhere in the North Atlantic Ocean [*Petrie and Drinkwater*, 1993, and references therein]. Furthermore, it is widely thought that this part of the ocean should reflect discharge of meltwater and ice during the Younger Dryas (YD) interval and during the "8200 year" event as well as smaller climate changes such as the Little Ice Age and the Medieval Warm Period [*Keigwin and Pickart*, 1999]. The 8.2 ka event is clearly the most prominent of these events, in Greenland ice cores [*Alley et al.*, 1997] and in our data. In the highest resolution $\delta^{18}O$ data from Greenland, the two cold peaks of the event occurred between about 8250 and 8180 years B.P., and from start to finish the event lasted about 200 years. Thus it should be resolvable in only the highest quality marine sediment cores.

2. Oceanographic Setting

[3] The western boundary of the North Atlantic is marked by the confluence of two major current systems: the Gulf Stream and the Labrador Current (Figure 1). Waters of the equator-flowing Labrador Current have two main sources, each of Arctic origin (see review by *Loder et al.* [1998]). One flow enters the Nordic Sea from Fram Strait and follows the coast from east to west Greenland. In the northern Labrador Sea, the West Greenland Current is joined by a second Arctic flow through the Canadian archipelago and out the Davis Strait. Now the Labrador Current sensu stricto, this cold and fresh water follows the Labrador shelf break and is augmented by freshwater discharge from Hudson Bay and other estuaries. Although this is a surface current, recent work shows that strong heat loss every winter and weak stratification can lead to eddy shedding and convection as deep as 1000 m [*Pickart et al.*, 1997]. This process can occur to the northeast of the Grand Banks and can account for the upper (high chlorofluorocar-

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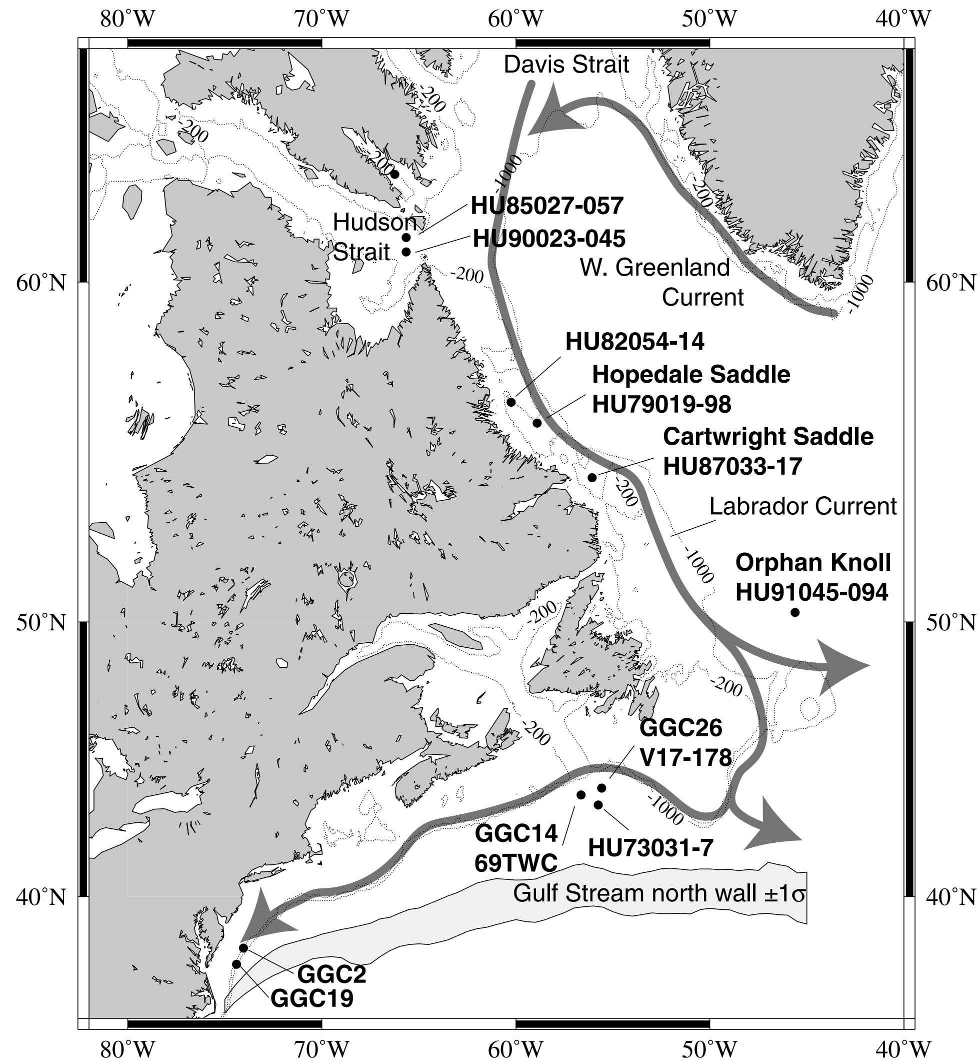

Figure 1. Base map of the slope water region equatorward from the Labrador Sea to Cape Hatteras with core locations discussed in this paper. The Labrador Current and its source waters are indicated by dark gray arrows. The north wall of the Gulf Stream is based on satellite observations [*Lee*, 1994].

bon, CFC) Labrador Sea Water. This high CFC water can be traced beyond Cape Hatteras far to the southwest [*Pickart and Smethie*, 1993].

[4] Because the Gulf Stream passes so close to the Tail of the Bank and may interfere with the Labrador Current, there is no clear view of how and when the Labrador Current rounds this corner to the Scotian Shelf. Nevertheless, it is known that the equatorward flow bifurcates in at least two places, with branches that retroflect eastward. The net result is that the equatorward flow on the Scotian shelf is on average about 1 Sv, compared to 3–8 Sv on the Labrador and Newfoundland shelves [*Loder et al.*, 1998]. *Rossby and Benway* [2000] proposed that the flux of cold and fresh shelf waters from Labrador may actually force 100 km excursions in the latitude of the Gulf Stream. To the south, *Spall* [1996a, 1996b] has already described a self-sustaining mechanism that couples Gulf Stream stability, eddy shedding, and zonal position with entrainment of Labrador Sea Water, and strength of the northern recirculating gyre. More recently, *Joyce et al.* [2000] used instrumental data to show that the slope water system could control Gulf Stream latitude through interactions at the separation point off Cape Hatteras. They link northward shifts of the Gulf Stream to the positive phase of the North Atlantic Oscillation (NAO) and southward shifts to the negative phase. Thus, not only are the slope waters north of the Gulf Stream thought to be a coupled system that oscillates on NAO timescales [*Pickart et al.*, 1999], but the Gulf Stream may be involved as well.

3. Methods

3.1. Sampling, Foraminifera, and Stable Isotopes

[5] One centimeter slices were removed from R/V Oceanus cruise 326 cores 14 and 26GGC (Table 1) at 2-cm spacing for studies involving foraminifera. This interval was doubled below 300 cm in 26GGC, where the rate of sedimentation is much higher. Particular intervals of interest were sampled as close as 1 cm. Samples for alkenone

Table 1. Location of Core Sites Discussed in This Paper

Core	Depth, m	Latitude, °N	Longitude, °W
CH07-98 2GGC	965	37 37.163	74 10.630
CH07-98 19GGC	1049	36 52.264	74 34.179
OCE3326 14GGC	3525	43 03.959	55 49.992
OCE3326 13MC	3440	43 03.97	55 50.03
OCE326 26GGC	3975	43 28.997	54 51.984
OCE326 25MC	3890	43 29.001	54 52.019
HU96029 69TWC	3506	43 03.972	55 50.030
HU85027-057PC	790	61 04.320	66 25.620
HU90023-045PC	845	60 56.822	66 08.280
HU82054-14PC	577	56 26.149	59 53.200
HU91045-94PC	3448	50 12.26	45 41.14
HU87033-17PC	514	54 36.99	56 10.6
HU79019-98	594	55 46.260	58 49.480
HU87033-17PC	514	54 36.99	56 10.6
V17-178	4006	43 29.0	54 52.0
HU73031-7PC	4055	42 58.700	55 14.900

geochemistry from 26GGC were paired with the foram samples at 4 cm intervals. Methods of sample preparation and analysis of foraminifera are similar to those of *Keigwin and Jones* [1995]. Stable isotope analyses on the benthic foram *Cibicidoides wuellerstorfi* were on the size fraction >150 μm, and the planktonic foram *Neogloboquadrina pachyderma* sinistral (s.) was analyzed at 150–250 μm. Cores 2 and 19GGC from R/V Cape Hatteras cruise 9807 were sampled at about 4 cm spacing for *N. pachyderma* dextral, and samples from CSS Hudson cores were provided by the curator at Bedford Institute of Oceanography.

3.2. Alkenones

[6] Numerous studies have demonstrated that sedimentary Uk'37 is a useful proxy for mean annual SST at 0 m water depth in the temperature range 0°–29°C [*Muller et al.*, 1998, and references therein]. Global core top calibrations of the alkenone paleothermometer [*Muller et al.*, 1998] are nearly identical to the original temperature calibration, which we use here, on the basis of cultures of *E. huxleyi* [*Prahl and Wakeham*, 1987]. The appropriateness of this calibration for surface waters over the Laurentian Fan is supported by the core-top (0–1 cm) alkenone SST of 8.8°C (see below), which is within the modern seasonal range of SST over that site, and is nearly identical to the 48-year average measured SST in Emerald Basin at 0 m, 8.7°C (data provided by K. Drinkwater). The details of our sampling and analytical procedures are given by *Sachs and Lehman* [1999]. Analytical precision is better than 0.20°C (1σ).

3.3. Trace Metals

[7] Cd/Ca was analyzed at MIT using the method of *Boyle and Keigwin* [1985] modified to utilize the reductive cleaning before the oxidative cleaning [*Boyle and Rosenthal*, 1996]. Mg/Ca samples were cleaned using the same method, excluding the reductive step, and were analyzed at Rutgers University using ICP-MS [*Rosenthal et al.*, 1999].

3.4. Radiocarbon Dating

[8] Samples for this project (Table 2) were radiocarbon dated at the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS) facility at Woods Hole Oceanographic Institution. In general, an attempt was made to date the same surface dwelling planktonic foraminifer in developing a chronology as were chosen for stable isotope analysis, but in many samples only mixed planktonic foraminifera could be analyzed. Foram samples of about 5 mg were cleaned ultrasonically in distilled water. Alkenones were purified from bulk sediment samples of early and late Holocene age using methods described by *Xu et al.* [2001] and *Ohkouchi et al.* [2002]. Because AMS ^{14}C analyses required ~0.5 mg of purified alkenones and their concentration was ~0.3 μg/g in early Holocene and ~1 μg/g in late Holocene sediment, and so as not to deplete an entire interval of core, channels of sediment 4 × 4 cm and 40 cm long (350–390 cm) and 30 cm long (20–50 cm) were removed from the split sediment core (26GGC) for early and late Holocene alkenone ^{14}C dates, respectively. In conversion from conventional ^{14}C years to calendar years B.P., no additional reservoir correction (R) was assumed than the ~400 years used by the CALIB program [*Stuiver et al.*, 1998].

4. Results and Discussion

4.1. Laurentian Fan

4.1.1. Sedimentation

[9] On the basis of numerous AMS ^{14}C dates (Table 2), sedimentation rates on the Laurentian Fan display regional variability that may be some function of normal hemipelagic sedimentation and overbank flow adjacent to the main valleys of the fan. If the dominant sedimentary processes are down channel flow of turbidites and deep western boundary current distribution of fine-grained overbank deposits [*Stow*, 1981], then we would expect the greatest sedimentation rates on the western side of the valley. *Skene and Piper* [2003] show instead that downstream and cross-stream variations in flow thickness of turbidites are the more important process. As shown in Figure 2, sedimentation rates were highest throughout the Holocene to the east. At the location of 25MC/26GGC the linear rate of sedimentation was 80 cm/kyr in the early Holocene, decreasing to ~30 cm/kyr younger than ~8 ka. In contrast, sedimentation rates were less than half that on the western levee of the eastern valley at 13MC/14GGC. This probably indicates that supply of sediment from the nearby continental slope of Newfoundland dominated sedimentation at the eastern core site during the Holocene, rather than overbank deposition from the eastern channel.

[10] Recent work of *Ohkouchi et al.* [2002] and *Mollenhauer et al.* [2003] has shown that alkenones, in addition to fine-grained (<63 μm) $CaCO_3$ and other organic components of sediment can have substantially older radiocarbon ages than coeval foraminifera, raising concerns about the reliability of alkenone-derived SST estimates at locations such as sediment drifts where substantial quantities of sediment are laterally advected. To evaluate this possibility on the Laurentian Fan, we radiocarbon-dated alkenones in early and late Holocene sediments from core 26GGC. Where our foram-based age model indicates sediments are about 11,000 yrs old, alkenones are less than 1000 yrs older, and they are about 1200 yrs older in the latest Holocene (Figure 2). If the assumption is made that

Table 2. Accelerator Mass Spectrometer Dates Used in This Paper, Including Those of *Keigwin and Pickart* [1999] on 13MC and 25MC

Cruise, Core	Depth, cm	Species	Conventional ^{14}C Age, years B.P.	Error, 1σ	NOSAMS No.	Calendar Age, years B.P.	Age Range
CH07-98, 19GGC	0.5	mixed planktonics	925	35	38860	521	541 to 504
	217	G. bulloides	8130	45	26433	8587	8641 to 8530
	305	G. inflata	9970	50	26434	10815	11128 to 10611
	369	mixed planktonics	10450	50	26435	11358	11899 to 10880
OCE326, 25MC	0	mixed planktonics	0[a]		16746	0	
	10.5	mixed planktonics	780	30	17732	416	436 to 385
	32.5	mixed planktonics	1680	50	17730	1221	1260 to 1164
	43.5	mixed planktonics	2040	25	17240	1568	1607 to 1537
OCE326, 26GGC	35	alkenones	2740	35	39019	2428	2539 to 2338
	66	N. pachyderma s.	2610	40	31452	2303	2328 to 2280
	96	N. pachyderma s.	3380	75	31453	3248	3337 to 3153
	120	G. inflata	3950	45	26436	3923	3981 to 3865
	161	mixed planktonics	5050	45	35599	5431	5458 to 5316
	195	N. pachyderma s.	6110	50	31454	6530	6610 to 6465
	204.5	N. pachyderma s.	6190	50	31455	6633	6674 to 6564
	234.5	G. menardii	7180	55	31456	7643	7676 to 7585
	240.5	N. pachyderma s.	7640	60	38435	8103	8158 to 8007
	248.5	N. pachyderma s.	8090	75	31476	8544	8628 to 8437
	270.5	mixed planktonics	8420	40	38436	8919	8964 to 8866
	310	G. inflata	8940	50	26437	9500	9778 to 9150
	370	alkenones	10050	65	41480	10982	11149 to 10651
	395	G. inflata	9720	55	26438	10327	10589 to 10297
	441	mixed planktonics	10050	50	38437	10982	11147 to 10655
OCE326, 13MC	0.5	mixed planktonics	0[a]		16737	0	
	1.5	mixed planktonics	165	40	16738	0	
	2.5	mixed planktonics	435	55	16739	0	
	3.5	mixed planktonics	570	30	16740	228	251 to 136
	5.5	mixed planktonics	735	30	16741	328	403 to 302
	7.5	mixed planktonics	1210	30	16742	720	752 to 686
	9.5	mixed planktonics	1390	35	17729	913	931 to 894
	11.5	mixed planktonics	930	45	16743	519	544 to 496
	14.5	mixed planktonics	1300	35	17731	827	884 to 774
	18.5	mixed planktonics	1380	30	16744	913	931 to 894
	30.5	mixed planktonics	2360	30	17047	1948	1979 to 1920
OCE326, 14GGC	128	G. inflata	9510	45	38545	10278	10303 to 10105
	203	N. pachyderma s.	11200	40	45401	12838	12902 to 12644
	236	mixed planktonics	12450	70	38547	13824	14087 to 13703
	256	mixed planktonics	12750	55	38548	14200	15082 to 14115
	285	mixed planktonics	12650	60	38431	14116	14315 to 14084
	362	mixed planktonics	13600	75	38687	15749	15988 to 15530
	404	N. pachyderma s.	15050	80	42493	17422	17162 to 17683
HU96029, 69TWC	0.5	mixed planktonics	670	40	14060	297	326 to 277
	4.5	mixed planktonics	1240	55	14061	770	857 to 715
	26.5	mixed planktonics	2880	25	17049	2695	2708 to 2674

[a]Fraction modern >1.

the greater ^{14}C ages of alkenones relative to planktonic foraminifera is due to mixing of "old" alkenones with those produced contemporaneously with the foraminifera, then a minimum of 15–20% of the alkenones in 26GGC are "pre-aged." (Alkenones that are 11,000 yrs old have one-fourth the ^{14}C activity as modern alkenones. Those that are 22,000 yrs old have one-sixteenth the ^{14}C activity.) While it would be preferable for all sedimentary components to have the same age, we note that age discrepancies are not unique to fine versus coarse, or organic versus inorganic fractions. The ^{14}C analyses of different species of planktonic foraminifera in the same sediment sample often yield a 1000–3000-year range of ages, or greater [*Duplessy et al.*, 1986; *Broecker et al.*, 1999, 2004; *Lowemark and Grootes*, 2004].

[11] Our radiocarbon results thus suggest that, while lateral advection of alkenones (and presumably fine grained material in general) to the Laurentian Fan may be significant and account in part for the high deposition rate there, we do not expect the process to incorporate quantities of "pre-aged" material sufficient to corrupt the SST_{alk} time series. While one might be tempted to subtract ~1000 yrs from the alkenone time series relative to the foram-based time series we have no reason to expect that the proportion of pre-aged alkenones remained constant through time. Furthermore, Holocene SST_{alk} records from throughout the NW Atlantic slope waters, from the North Carolina to the Nova Scotia shelves, all show near monotonic cooling of 3°–8°C through the Holocene (J. P. Sachs and L. D. Keigwin, unpublished data, 2004).

4.1.2. Oxygen Isotopes

[12] Results of oxygen isotope analysis and other analytical data are archived at the NGDC World Data Center for Paleoclimatology. The $\delta^{18}O$ of *N. pachyderma* s. at the site of 25MC/26GGC (this site is referred to hereafter as 26GGC) reveals a pattern similar to that reported earlier from the slope waters [*Keigwin and Jones*, 1995]. From the bottom of 26GGC at 11 ka to ~8.5 ka, $\delta^{18}O$ decreased by more than 1.5‰ to a minimum of about 1‰. After that

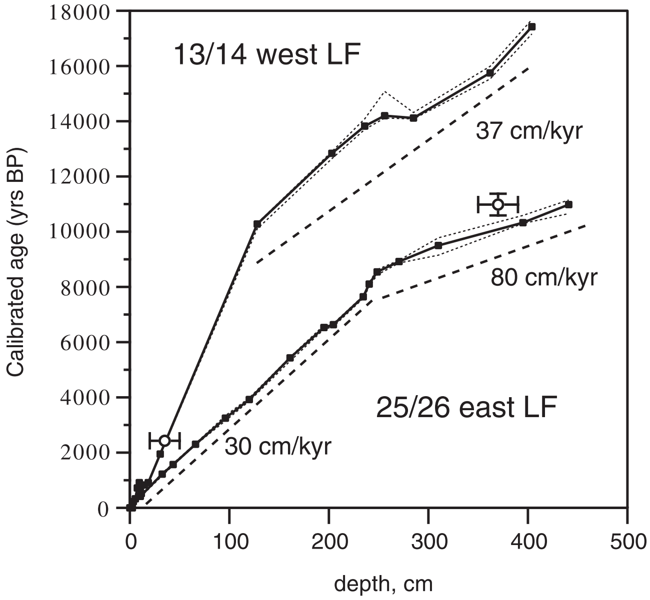

Figure 2. Age-depth relationship for Laurentian Fan (LF) cores to the east and west of the main (eastern) valley. Small solid squares are AMS dates on planktonic foraminifera calibrated to calendar years assuming a 400 year reservoir effect ($\Delta R = 0$) (Table 2), with $\pm 1\ \sigma$ errors indicated by the dotted lines. Calibrated AMS dates on alkenone samples are shown as open circles, with the *x* axis error bar showing the interval of core sampled. Dashed lines show linear sedimentation rates, for reference.

minimum, $\delta^{18}O$ increased gradually by about 1.25‰ to the sediment surface (Figure 3). When the data are detrended to account for the global ice volume effect [*Fairbanks*, 1989], it is clear that large changes in some combination of the $\delta^{18}O$ of local seawater and temperature dominate the small ice volume signal. New results at 26GGC differ from previous results in that the age of the low $\delta^{18}O$ is constrained by AMS dates of 8.1 and 8.5 calendar ka (Table 2); now we believe this event must be related to the final deglaciation of Hudson Bay that began as early as ~8.45 ka on the basis of the oldest date of marine incursion there [*Barber et al.*, 1999]. Previously, this event was defined as the interval where the *N. pachyderma* s. was ~1‰ [*Keigwin and Jones*, 1995]. Here, with much closer sampling, we define the event to be <1‰, and because of the scatter in the signal, we define its onset by the rapid decrease of $\delta^{18}O$ in the three point running average to <1‰ at 8.54 ka (Figure 3). We consider it to have ended at 7.80 ka, the second of two consecutive $\delta^{18}O$ minima, but we caution that these dates are subject to our assumption of no change in reservoir age.

[13] Although our Laurentian Fan gravity cores are relatively short, a longer perspective on the 8.2 ka event comes from patching the deglacial record of core 14GGC to the Holocene record of core 26GGC (Figure 4). Separated by only ~60 km, these sites should reflect the same surface water conditions (Figure 1). Core 14GGC contains several deglacial events on the century timescale. The $\delta^{18}O$ minimum at ~16 ka [*Keigwin and Jones*, 1995; *Piper and Skene*, 1998; *Skene and Piper*, 2003] is resolved here as two events, and several other minima are recognized

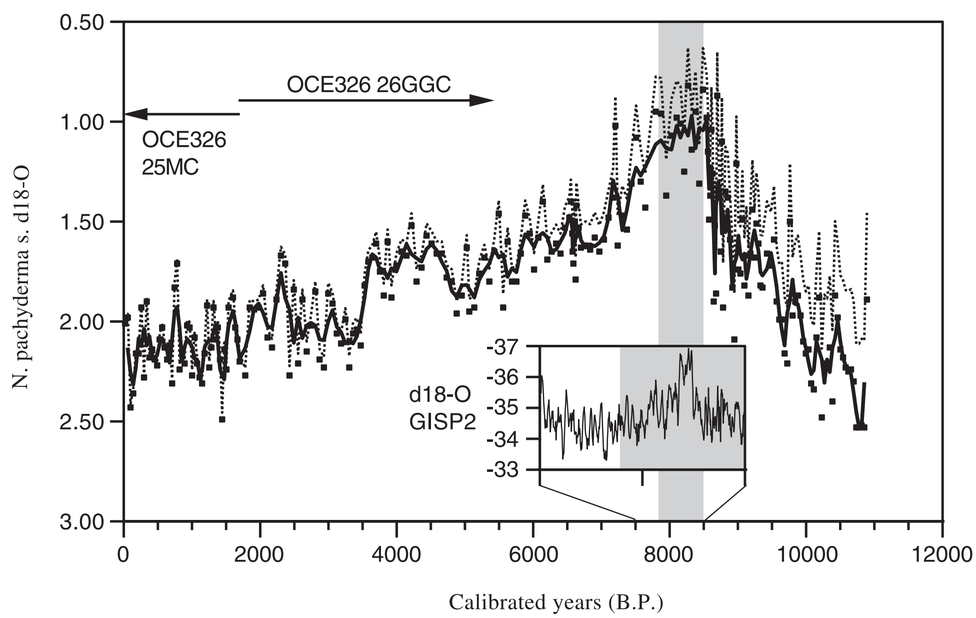

Figure 3. Average $\delta^{18}O$ results on N. *pachyderma* s. (squares) from the eastern levee of the eastern valley of Laurentian Fan (OCE326 26GGC, 25MC) plotted versus calendar age. The bold black line is the three point running average of those data, and the dotted line is the $\delta^{18}O$ of *N. pachyderma* s. corrected for the ice volume effect. Arrows indicate where the data from multicore (MC) 25 are patched to 26GGC, and the shaded column highlights the interval of minimum $\delta^{18}O$ of *N. pachyderma* between about 7800 and 8500 years B.P. The inset shows the high-resolution GISP2 $\delta^{18}O$ results from the interval 7500 to 8500 years B.P.

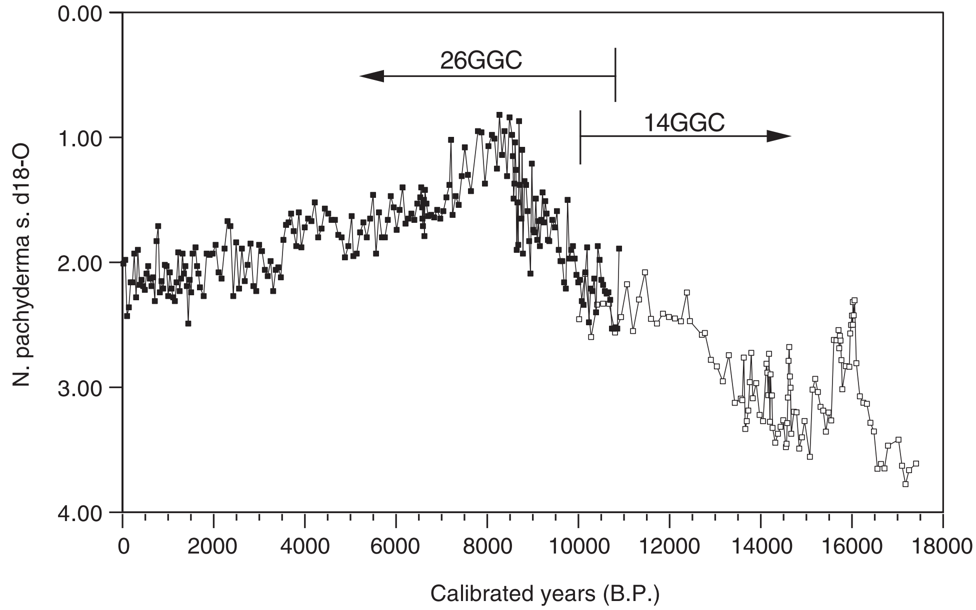

Figure 4. Composite time series of *N. pachyderma* s. δ^{18}O from the Laurentian Fan versus calendar age. Solid symbols are from 25MC and 26GGC to the east of the main (eastern valley), and open symbols are from 14GGC, to the west (Figure 1). These two sites are within about 60 km of each other.

between ~13 and 15 ka. The oldest event is probably equivalent to Heinrich event 1, and the similarity in shape and magnitude between that event and the one at 8.2 ka suggests that they may reflect a similar process. As reported before, there is no expression of Younger Dryas freshening in either δ^{18}O [*Keigwin and Jones*, 1995; *de Vernal et al.*, 1996] or dinoflagellate assemblages data [*de Vernal et al.*, 1996], even though that episode is clearly expressed in the % *N. pachyderma* s. and ice rafting (not shown) in 14GGC and other sites [*Skene and Piper*, 2003].

4.1.3. Sea Surface History

[14] When first identified, the Holocene minimum in δ^{18}O was interpreted as low salinity in near surface waters because elevated abundance of *N. pachyderma* s. in the same samples would indicate surface cooling [*Keigwin and Jones*, 1995]. Here we support that observation with *N. pachyderma* s. counts on the new core, as well as with Uk37 SST data (SST_{alk}) (Figure 5a) and Mg/Ca data (Figure 5b). The reader should beware that by SST, we actually mean calcification temperature for coccolithophores and planktonic foraminifera. At present we have no way of knowing if calcification for these microfossils occurred at the sea surface or within the surface mixed layer.

[15] Although there is no evidence for pronounced cooling 8200 years ago in the Mg/Ca derived SST (Figure 5b), we believe this is a robust cold/low SSS (sea surface salinity) event because the previously reported maximum in % *N. pachyderma* s. is present in 26GGC and also because it is a doublet (Figure 5d). The two-step character of the 8200 year event has been recognized previously in Greenland snow accumulation rates [*Alley et al.*, 1997], lake temperatures [*von Grafenstein et al.*, 1998], speleothem-based air temperatures in Ireland [*Baldini et al.*, 2002], and in release volume estimates for glacial Lake Agassiz [*Leverington et al.*, 2002]. Pronounced cooling over the Laurentian Fan is mirrored by dinoflagellate cyst (dinocyst) evidence for even larger cooling in the southern Labrador Sea near Orphan Knoll [*de Vernal and Hillaire-Marcel*, 2000] (Figure 5a). These authors showed a large decrease in SST from ~18°C at ~7.9 ka to ~5°C at 7.3 ka. We contend that this change, which marks an average Holocene stepwise cooling of about 4°C, is likely to be coincident with the changes we have identified at ~8.2 ka, within dating errors (see triangles along *x* axes in Figure 5a). The coincidence of the low δ^{18}O, the abundant *N. pachyderma* s. (in the context of the longer *Keigwin and Jones* [1995] record), and the SST_{alk} drop identifies the "8200" year event as the most prominent climate change in the slope water system during the past ~11 ka. It is notable that the beginning of the cooling/freshening event identified here (8.54 ka) is very close to the 8.45 ka of the oldest marine mollusk dated around Hudson Bay [*Barber et al.*, 1999], and that it may have lasted some 700 years. Because this interval in 26GGC is about 40 cm thick, it is not likely that bioturbation has smeared over 700 years an event that glaciologists think could have occurred in as little as a year [*Clarke et al.*, 2004]. Thus the oceanic expression of this event may have occurred over the course of centuries, even if the forcing was much briefer.

[16] For the remainder of the Holocene, SST_{alk} indicates continual cooling of 10°C with only minor variability. *Balsam* [1981] first noted that the slope waters over Laurentian Fan were warmer in the early Holocene. Declining SST is supported by both the higher percent abundance (Figure 5d) and lower δ^{18}O (Figure 5c) of the warm surface dwelling planktonic foram *Globigerinoides ruber* in the early Holocene. Long-term cooling through the Holocene has been recognized at many locations in the North Atlantic and is probably a response to changing seasonality driven by orbital change [*Marchal et al.*, 2002]. Typically, an early

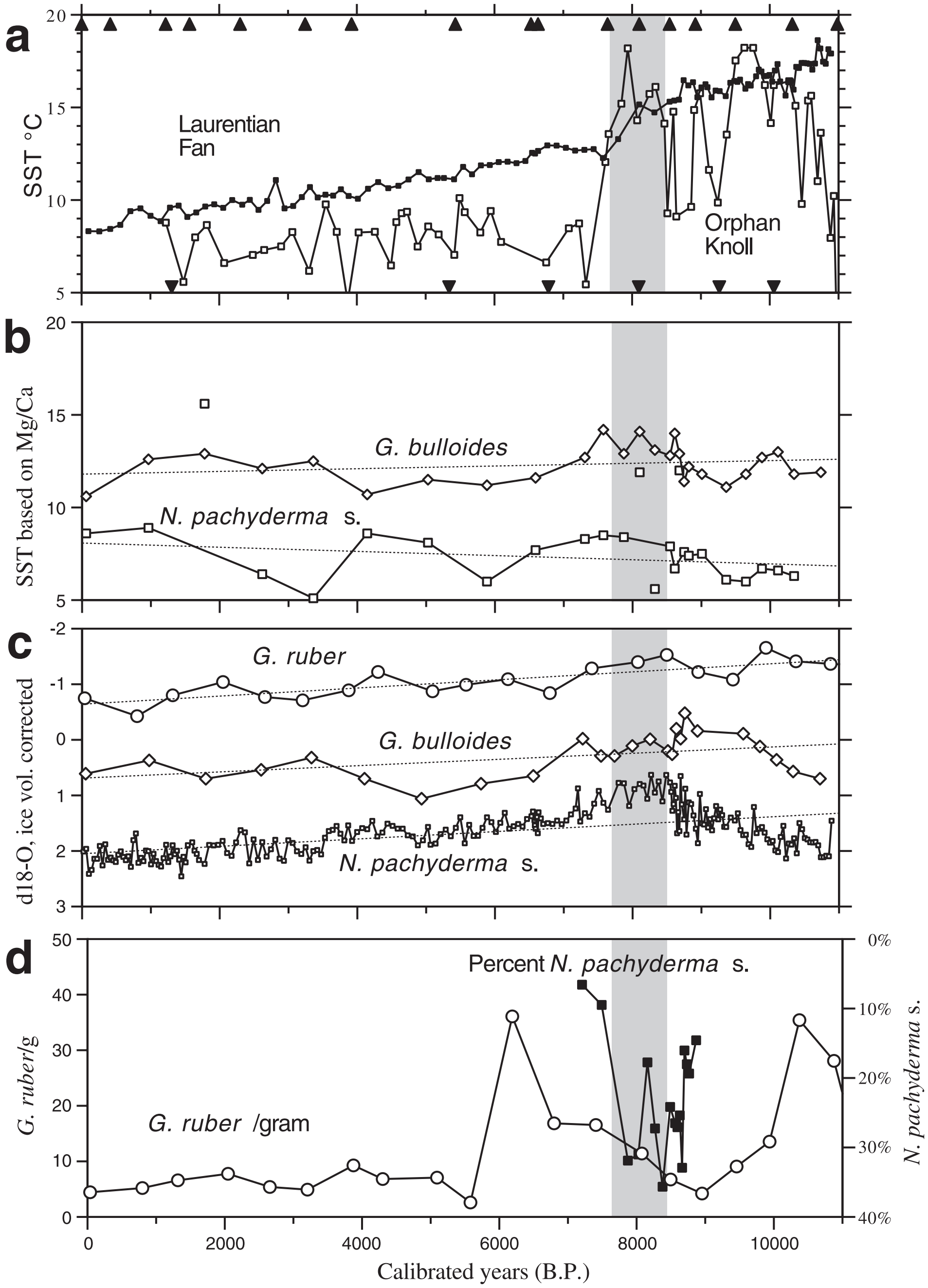

Figure 5. Holocene SST proxies in the slope water system off eastern Canada. (a) SST_{alk} on the Laurentian Fan compared to summer SST based on dinocysts over Orphan Knoll [after *de Vernal and Hillaire-Marcel*, 2000; *Solignac et al.*, 2004]. Chronological control is indicated by triangles for (top) Laurentian Fan and (bottom) Orphan Knoll. (b) Mg/Ca-based calcification temperature estimates for the planktonic foraminifera *G. bulloides* (diamonds; $Mg/Ca = 0.72e^{0.1T}$, where T is the temperature in °C; [*Barker and Elderfield*, 2002]) and *N. pachyderma* s. (squares; $Mg/Ca = 0.6e^{0.1T}$, where T is the temperature in °C [*Elderfield and Ganssen*, 2000]) from the Laurentian Fan. (c) The $\delta^{18}O$ results, corrected for the ice volume effect, on the planktonic foraminifer *G. ruber* from V17-178 (circles, after *Keigwin and Jones* [1995]), *G. bulloides* (diamonds), and *N. pachyderma* s. (squares) from the Laurentian Fan. In Figures 5b and 5c, the straight lines are regressions through the data. (d) Faunal proxies for SST. The cold water planktonic foram species *N. pachyderma* s. (solid squares) reaches maximum abundance around 8200 years ago on Laurentian Fan (note reversed *y* axis), whereas, in general, the warm water species *G. ruber* (open circles) indicates warm surface waters in the early Holocene (previously unpublished data from V17-178). In each panel the shaded column marks the interval of maximum cooling associated with the 8200 yr event.

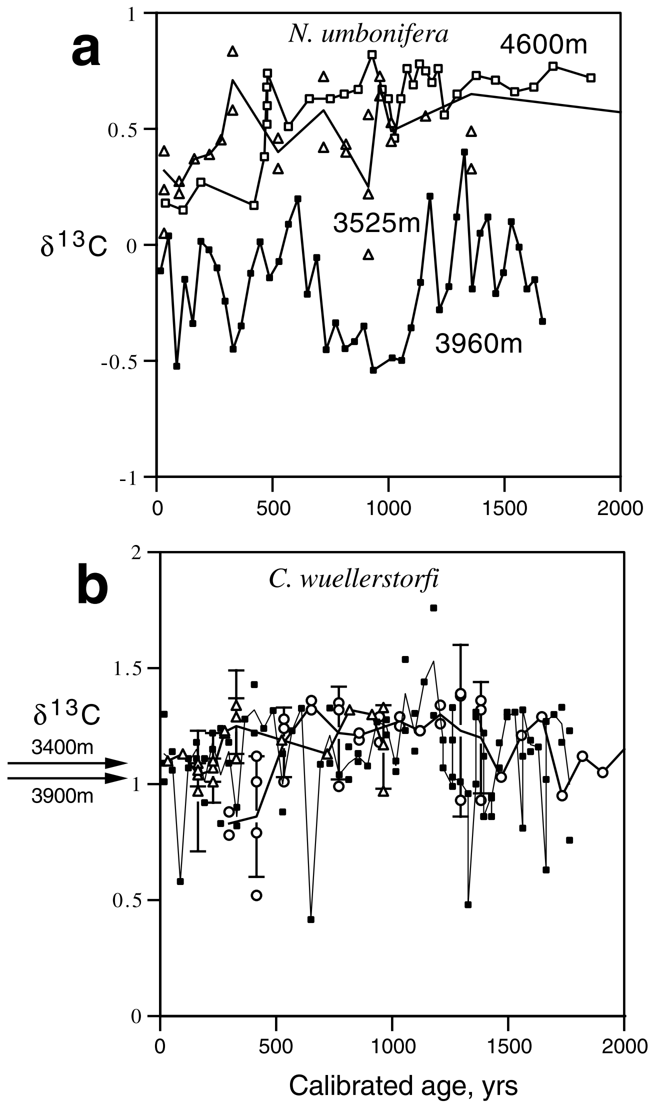

Figure 6. The $\delta^{13}C$ results on (a) *N. umbonifera* and (b) *C. wuellerstorfi* from sites on the Laurentian Fan and Bermuda Rise, western North Atlantic, using age models of *Keigwin and Pickart* [1999], *Keigwin and Boyle* [2000], and the chronology presented here for core HU96029 69TWC (Table 2). These results show a systematic and unexplained difference between the $\delta^{13}C$ of *N. umbonifera* at Laurentian Fan core 25MC at 3960 m (solid squares) and nearby Laurentian Fan core 13MC (triangles, 3525 m) and Bermuda Rise core BC-9 (open squares, 4600 m). On the other hand, $\delta^{13}C$ of *C. wuellerstorfi* from the same Laurentian Fan sites (same symbols as in Figure 6a) as well as at HU96029 69TWC (circles) are approximately the same, and insignificantly different from the $\delta^{13}C$ of ΣCO_2 measured over Laurentian Fan (arrows along *y* axis).

to mid-Holocene warm interval, or hypsithermal, is evident, and the data from many latitudes in the eastern North Atlantic show only modest post hypsithermal cooling of ~2°C [*Marchal et al.*, 2000]. In the Norwegian-Greenland Sea, there is geographic variability in long-term Holocene cooling of a few degrees, but the greatest change was 4°–5°C in the Irminger Sea [*Andersen et al.*, 2004]. From 11 ka to 8.5 ka SST_{alk} was high, but in reasonable agreement with SSTs of ~17°C measured at the site of 26GGC during July 1998 when the core was collected. Dinocyst assemblages also indicated high SST in the early Holocene on Laurentian Fan [*de Vernal et al.*, 1996] and Orphan Knoll (Figure 5a) [*de Vernal and Hillaire-Marcel*, 2000].

[17] Despite the evidence for early Holocene warmth and subsequent cooling, long-term Holocene cooling is inconsistent with SSTs on the basis of the Mg/Ca method (Figure 5b). SST estimates using *G. bulloides* [*Barker and Elderfield*, 2002] show about a 1°C cooling and estimates using *N. pachyderma* s. [*Elderfield and Ganssen*, 2000] show about a 1°C warming trend (Figure 6c). This inconsistency probably results from our poor understanding of the depth and season of growth of the various microfossils we have chosen for study. More attention needs to be paid to proxy calibration in a diversity of environments.

4.1.4. Additional Scenarios

[18] In the above section we argued that over the Laurentian Fan there may have been long-term decrease of SST through the Holocene. However, because there is no common thread among all the SST and $\delta^{18}O$ data, additional interpretations are possible. For example, if we take the SST_{alk} and the *N. pachyderma* s $\delta^{18}O$ at face value and assume that the nannofossils and *N. pachyderma* lived at the same time and place, then decreasing SST_{alk} and increasing $\delta^{18}O$ would signal a continual decrease in SSS from about 11 ka to 8 ka. This interpretation is not unreasonable because lowered SSS in the early Holocene has been documented around many margins of the North Atlantic by *de Vernal et al.* [2000], and is consistent with high surface runoff from melting glaciers. A direct record of this can be found in the history of summer melt layers in ice caps from the Canadian Archipelago [*Koerner and Fisher*, 1990], close to one of the source regions for the Labrador Current. A second source of freshening in the slope water system could be the numerous drawdowns of glacial Lake Agassiz that preceded the final event at ~8450 years B.P. However, according to the latest compilation of these events for the past 11,000 years, all were <10% of the volume of the final event [*Leverington et al.*, 2002]. If the foraminifera and coccolithophores grew at exactly the same time and place over the Laurentian Fan, we can solve a paleotemperature equation [*Shackleton*, 1974] for salinity by assuming the modern North Atlantic δ_{water}/salinity relationship [*Craig and Gordon*, 1965]. Using the data in Figure 5a results in a salinity estimate of 40 psu for 11 ka, and 37 for 8.5 ka. Clearly these estimates are too high, but the magnitude of change, nearly 3 psu, may be close to the true extent of salinity depression that set the stage for the 8.2 ka event.

[19] Alternatively, it is possible there was no significant long-term cooling through the Holocene, as indicated by Mg/Ca on both species of planktonic foraminifera (Figure 5b). For this scenario to be correct, coccolithophores and *G. ruber* must have changed their season or depth of calcification. For *G. ruber* this would be difficult to imagine because this is a well-known surface dweller and because it lives only in the summertime in slope waters

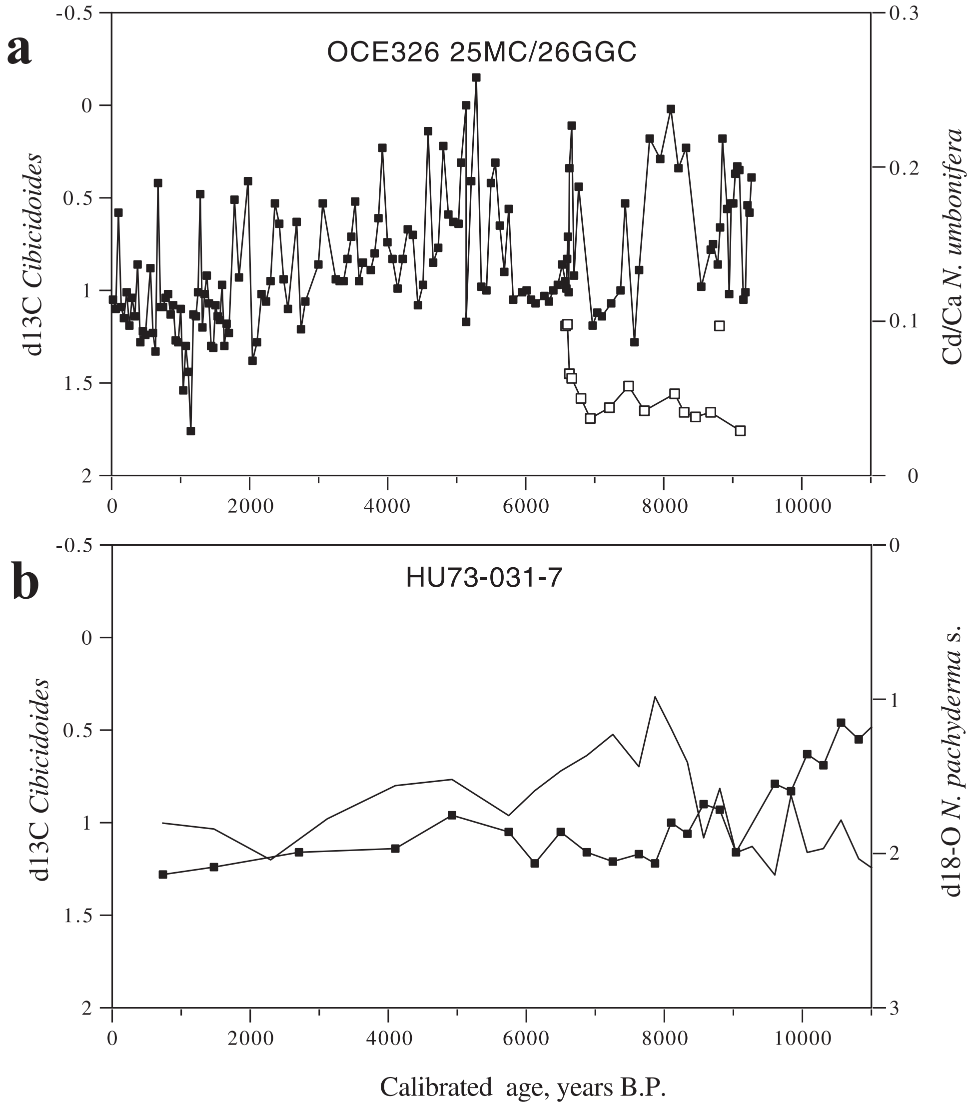

Figure 7. (a) Benthic $\delta^{13}C$ (*C. wuellerstorfi*, solid squares) and Cd/Ca (*N. umbonifera*, open squares) at the site of MC25 and GGC26 (3525m), on the Laurentian Fan. (b) The $\delta^{13}C$ of *C. wuellerstorfi* (solid squares) and $\delta^{18}O$ of *N. pachyderma* s. (line) from 4200m (HU73031-7) on the Laurentian Fan.

(Keigwin, unpublished plankton tow data). However, although coccolithophores favor high intensity sunlight, they live in a range of water temperatures and (for some unknown reason) could have changed their season of growth through the Holocene from summer to the early spring bloom. This speculation is consistent with the strong negative relationship between alkenone concentration in 26GGC sediments and SST_{alk} (not shown; r^2 = 0.75). Higher alkenone concentrations (and lower SST) would be expected with higher nutrient availability in the spring bloom. A strong argument against this interpretation is that satellite observations show coccolith production is overwhelmingly during summer at mid and high latitudes throughout the global ocean today [*Iglesias-Rodriguez et al.*, 2002], but we have no data on the integrated flux of coccoliths across the annual cycle. At present we lean toward the interpretation that there was indeed long-term cooling in the surface waters over Laurentian Fan.

4.1.5. Changes in Thermohaline Circulation 8200 Years Ago?

[20] Carbon isotope ratios in benthic foraminifera have a mixed history as proxies for nutrient content, hence deep ocean ventilation, in the western North Atlantic. This proxy received a boost when it was shown to reflect the same nutrient patterns as Cd/Ca in late Pleistocene benthics [*Boyle and Keigwin*, 1982]. Together these data made a compelling case for a reduction in the percentage of North Atlantic Deep Water (NADW) in the waters overlying the core site during glacial stages and substages. However, millennial scale changes in the $\delta^{13}C$ of *C. wuellerstorfi* in the late Holocene were not supported by Cd/Ca data [*Keigwin et al.*, 1991]; instead it appeared that *N. umbonifera* was a

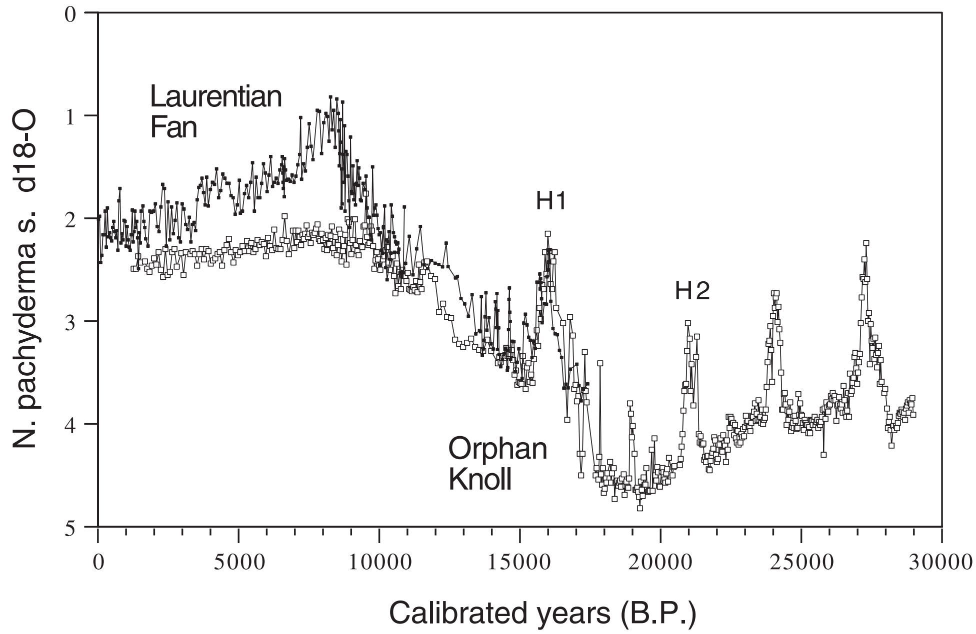

Figure 8. Oxygen isotope results on *N. pachyderma* s. from the Laurentian Fan (this study, solid points) compared to Orphan Knoll (open points) [*de Vernal and Hillaire-Marcel*, 2000] on independent age models. The most notable difference between δ^{18}O at these two locations occurred in the early Holocene when the surface waters over Laurentian Fan were anomalously low (fresh). Absence of the δ^{18}O minimum at ~8200 yrs B.P. on Orphan Knoll is curious, considering meltwater discharge at that time is thought to have been to the Labrador Sea [*Barber et al.*, 1999]. Two of the earlier low d18-o intervals on Orphan Knoll are identified as Heinrich (H) Events.

more reliable proxy for δ^{13}C of bottom water on the Bermuda Rise. More recently, it was found that δ^{13}C of *N. umbonifera* decreased abruptly during the Little Ice Age, but Cd/Ca did not increase [*Keigwin and Boyle*, 2000]. To our knowledge, there are no pairs of δ^{13}C and trace metal data that unambiguously show Holocene changes in NADW production.

[21] To develop a record of NADW history on Laurentian Fan, we measured late Holocene δ^{13}C in both *N. umbonifera* and *C. wuellerstorfi* from trigger weight cores and from multicores, and compared results to bottom water measurements there and results from the Bermuda Rise (Figure 6). In the Laurentian Fan region, δ^{13}C of *N. umbonifera* seems unreliable because it differs systematically between two cores that differ in water depth by only ~500 m (Figure 6a), where there is no bottom water δ^{13}C difference (arrows along *y* axis in Figure 6b). *N. umbonifera* data are always lower than the δ^{13}C of bottom water, but results at core 25MC (~4000 m) are significantly lower than at either core 13MC (3500 m) or on the Bermuda Rise (4600 m; data from *Keigwin and Boyle* [2000]). We do not know why the *N. umbonifera* δ^{13}C is anomalous at the 4000 m site on Laurentian Fan, including core top values that are more than 1‰ lower than modern bottom water, but similar δ^{13}C results on *C. wuellerstorfi* at three Laurentian Fan sites at different depths (Figure 6) led us to concentrate on that species for stable isotopes in the long Holocene section of 26GGC.

[22] Carbon isotope ratios of *C. wuellerstorfi* from 26GGC are low enough (<0.5‰) to indicate significantly decreased NADW production at several times in the Holocene (Figure 7a). For example, prominent events of low δ^{13}C are seen at ~5.0, 6.5, 8.2, and 9.0 ka. By far, the event associated with colder and fresher surface waters between ~8.5 and 7.8 ka is the most distinctive because of its duration and reproducibility. If taken literally, this would indicate near complete cessation of NADW and replacement by nutrient-rich water from the Southern Ocean. This scenario is supported by modeled freshening of the North Atlantic 8200 yrs ago [*Renssen et al.*, 2001], and is consistent with the interpretation of *Oppo et al.* [2003], in whose record the ~5 ka event is the most prominent (but the ~8.2 ka event is absent entirely). We interpret benthic δ^{13}C cautiously because of inconsistencies described above, and because other studies have suggested a link between benthic δ^{13}C and surface ocean fertility [*Zahn et al.*, 1986; *Mackensen et al.*, 1993]. Curiously, results from core HU73031-007 at ~4100 m on Laurentian Fan show none of the low δ^{13}C events (Figure 7b). Even though the sedimentation rate is low (~10 cm/kyr), this is one of the sites where low δ^{18}O around 8 ka was first noted [*Keigwin and Jones*, 1995], so at least some details of millennial scale climate change are preserved.

[23] Because of concerns about the reliability of δ^{13}C, we measured Cd/Ca on *N. umbonifera* as a second proxy for bottom water nutrient content around 8200 years ago at 26GGC (Figure 7a). Cd/Ca results at that time are typical of modern nutrient-poor waters, and do not support decreased NADW production. One sample at 8900 yrs B.P. is a single analysis with no replicate or similar adjacent sample.

Because of the potential for contamination during trace metal analyses, we do not believe that any confidence can be placed on one measurement with high Cd/Ca. Another sample at 6600 yrs B.P. had a nearby analysis that agreed (199.5 cm, 0.097 μmol/mol; 202 cm, 0.098 μmol/mol); it is possible that this high Cd event is real and bottom waters were nutrient-rich at that time. However, the major feature of the Cd/Ca data is the absence of high values in the 7600 to 8400 yrs B.P. interval with low δ^{13}C. Hence we cannot place any great confidence in the δ^{13}C record as indicating reduced NADW during the 8200 yr event. In our opinion, identifying Holocene changes in NADW production has such great societal relevance that paleoceanographers have a special obligation to bring more than one tracer to bear on the problem [*Keigwin and Boyle*, 2000].

4.2. Spatial Extent of Lowered Salinity 8200 Years Ago

[24] The low δ^{18}O that occurred 8200 years ago on the Laurentian Fan should provide a powerful tracer for mapping the spatial extent of the meltwater that may be responsible for widespread cooling at that time [*Alley et al.*, 1997; *Klitgaard-Kristensen et al.*, 1998; *von Grafenstein et al.*, 1998; *Barber et al.*, 1999; *Nesje and Dahl*, 2001]. Curiously, there is no δ^{18}O evidence for this event anywhere in the Labrador Sea, even though Hudson Strait is thought to have been the source. Within Hudson Bay itself, geophysical evidence for the flood (scour marks, channels) has been described by *Josenhans and Zevenhuizen* [1990], and hematite rich rocks from the northern part of the bay are thought to be the source of red clay marker beds within Hudson Strait [*Kerwin*, 1996].

[25] Of the many Labrador Sea cores that lack δ^{18}O evidence for the 8.2 ka event [*Hillaire-Marcel et al.*, 1994], core HU91045-94, from Orphan Knoll (Figure 1) has one of the best records of the latest Quaternary [*de Vernal and Hillaire-Marcel*, 2000]. It has a well-dated series of δ^{18}O on *N. pachyderma* s., and has a sufficient rate of sedimentation to resolve the 8.2 ka event if it occurred there. Figure 8 shows that Orphan Knoll and Laurentian Fan results are nearly identical early in the deglaciation, but diverge in the early Holocene. The absence of low δ^{18}O ~8200 years ago is notable, considering how well the event at 16 ka is represented at each location.

[26] Because Orphan Knoll is seaward of the Labrador Current (Figure 1), we reasoned that it might not record the low salinity influence of a major meltwater event if it was confined to coastal waters by the effect of its low density and high transport on the position of the shelf break front [*Chapman*, 2000]. However, on the southern Labrador shelf at Cartwright Saddle, there is no evidence of a major lowering of δ^{18}O at about 8200 yrs B.P. (Figure 9e) [*Andrews et al.*, 1999], nor is there any at sites farther to the north. Figures 9a and 9b show δ^{18}O on *N. pachyderma* s. from two cores in Hudson Strait that have ^{14}C dates on benthic foraminifera and mollusca [*Kerwin*, 1996; *MacLean et al.*, 2001; *Jennings et al.*, 2001]. Each has at least some indication of the two markers that are thought to represent the deglaciation of Hudson Bay (a low in magnetic susceptibility and a red layer [*Kerwin*, 1996; *Jennings et al.*,

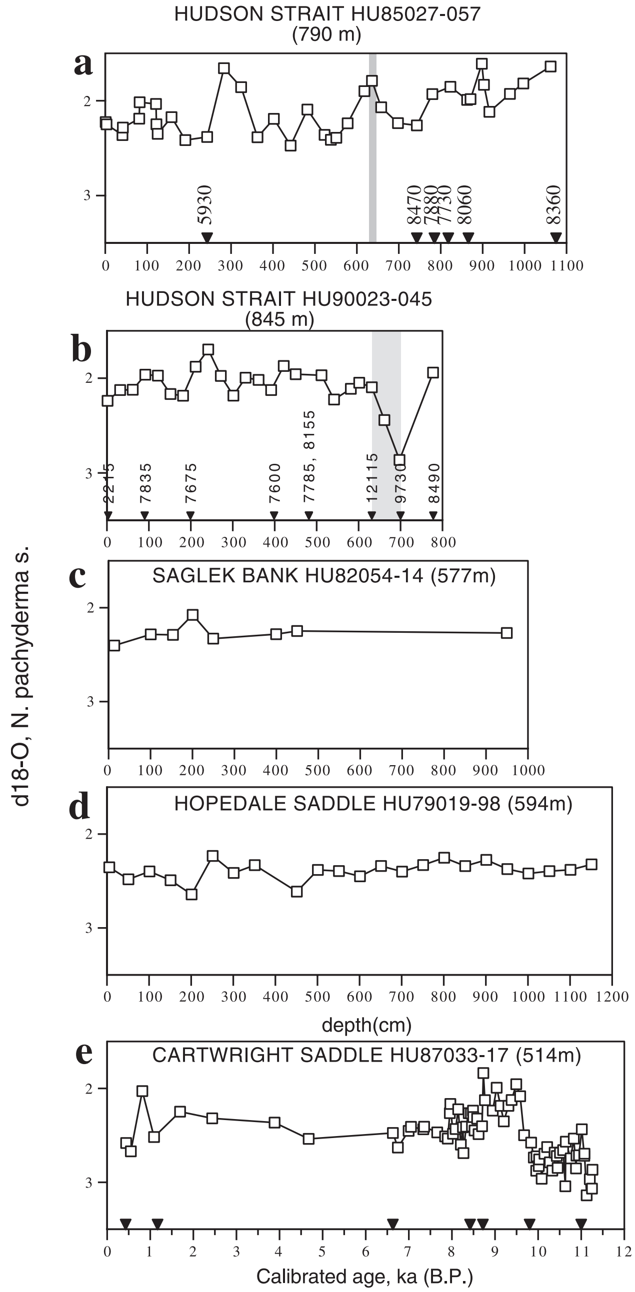

Figure 9. The δ^{18}O of *N. pachyderma* (s) from sites along the Labrador Shelf and Hudson Strait. (a and b) In Hudson Strait, cores have the red zone (shaded bar, Figure 9a) or the low magnetic susceptibility markers (shaded bar, Figure 9b) for the Hudson Bay deglaciation [*MacLean et al.*, 2001; *Jennings et al.*, 2001], but determining an accurate chronology is difficult when dating benthic foraminifera and mollusks in a dynamic environment that might have been affected by the discharge event. (c and d) The next sites farther south lack any evidence of low δ^{18}O, or sustained trends, and lack any chronology except to show they are Holocene [*Josenhans et al.*, 1986]. (e) At the southern end of this transect, Cartwright Saddle shows low δ^{18}O between ~9.5 and 8.5 ka. Note data from that site are plotted versus age, with dates indicated by triangles along *x* axis [*Andrews et al.*, 1999].

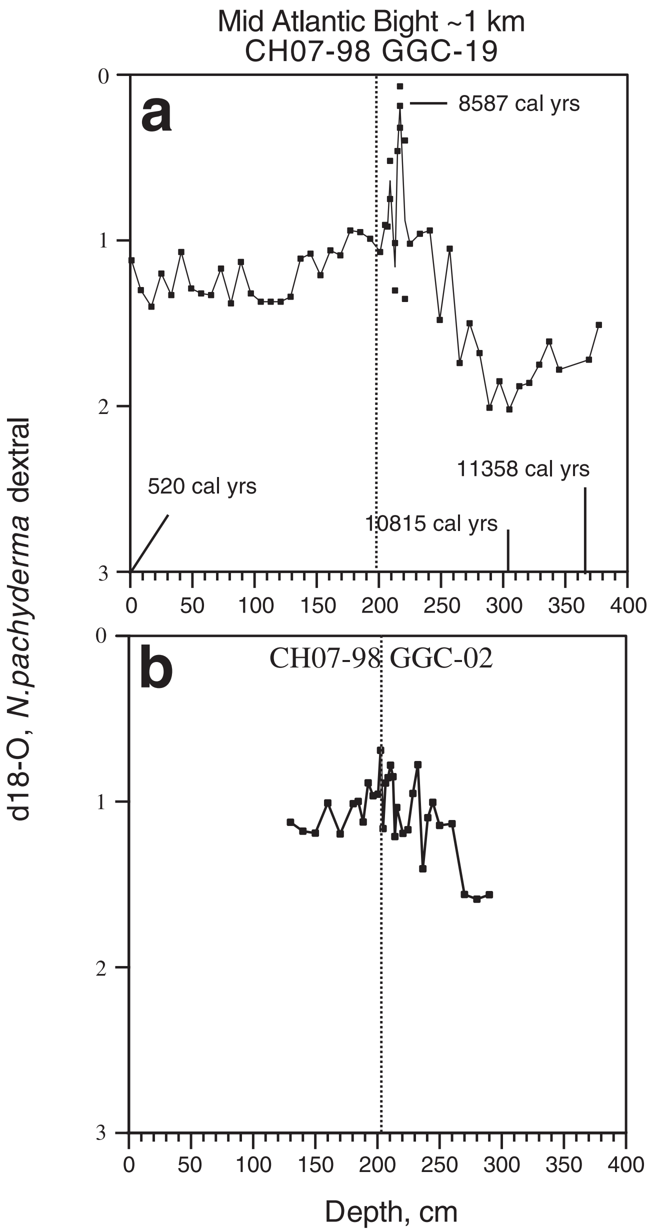

Figure 10. Oxygen isotope results on *N. pachyderma* (dextral) from two cores along the continental slope north of Cape Hatteras. Similar to Laurentian Fan cores, 19GGC has a pronounced $\delta^{18}O$ minimum that is coincident with the earliest marine incursion in Hudson Bay [*Barber et al.*, 1999]. Core 02GGC lacks this distinct minimum, but there is a suggestion of lower $\delta^{18}O$ in the correlative early Holocene interval. These cores are correlated at the first abundant appearance of *G. menardii* (dotted vertical line). This event is dated 7600 calendar years B.P. on the Laurentian Fan (Table 2) and elsewhere.

2001]). Although the ^{14}C dates do not make an especially reliable chronology, combined with the lithostratigraphy they suggest low $\delta^{18}O$ at ~8.0 to 8.5 ka should have been recorded in the strait, if fresh water passed through. This statement is true as well for cores from the Labrador Shelf basins (Figures 9c–9e) [*Josenhans et al.*, 1986; *Andrews et al.*, 1999]. Of course, there are many reasons why evidence for such an event might be missing. For example, benthic foram biostratigraphy and lack of a minimum in magnetic susceptibility suggests piston core 057 (Figure 9a) did not reach far enough to recover the true deglacial event [*Jennings et al.*, 2001], but for now the important point is the lack of low $\delta^{18}O$ observations for this event. Nevertheless, the lack of any trends in the data (as at Orphan Knoll), and the fact that $\delta^{18}O$ in Hudson Strait is lowest (as would be expected if salinity was lowest there) indicates that the *N. pachyderma* s. $\delta^{18}O$ is probably a reliable tracer for near surface temperature and salinity at these locations.

[27] Finally, we report $\delta^{18}O$ and ^{14}C results on two cores from along the continental slope north of Cape Hatteras (Figure 1). At CH07-98 19GGC we have found that $\delta^{18}O$ of *N. pachyderma* (dextral) decreased abruptly by nearly 1‰ about 8500 years ago (Figure 10a). The isotope stratigraphy at this site is similar to that seen on Laurentian Fan, far upstream in the slope waters, although the event to the south forms more of a spike. A nearby site (2GGC) confirms the generally lower $\delta^{18}O$ in the early Holocene, but lacks the sharp minimum found at 19GGC (Figure 10b). If further work supports the pattern of low salinity outlined here, then either meltwater found a previously unknown southern outlet ~8.5 kyr ago (e.g., St. Lawrence or Hudson Rivers) or it escaped unnoticed from the Labrador Sea along the continental shelf and mixed into the slope water system.

5. Conclusions

[28] The following points summarize our conclusions.

[29] 1. Alkenone thermometry indicates that SST in slope waters over the Laurentian Fan decreased continuously by about 10°C since the beginning of the Holocene. A similar cooling is recorded by dinoflagellate-based SST reconstructions by *de Vernal and Hillaire-Marcel* [2000] at Orphan Knoll in the southern Labrador Sea. Direct ^{14}C dating of alkenones on the Laurentian Fan shows that SST_{alk} is not affected as much by lateral advection of older sediment as other locations with focused sedimentation. The regular decline in SST_{alk} was interrupted by accelerated cooling that began ~8.5 ka and continued for about 700 yrs. Associated with this cold pulse was a large increase in the abundance of the polar planktonic foram *N. pachyderma* s. The maximum rate of cooling during this brief interval was nearly 5°C/kyr.

[30] 2. Maximum cooling around 8.2 ka is associated with minimum $\delta^{18}O$ in planktonic foraminifera only in the western subpolar gyre, between the Laurentian Fan and Cape Hatteras, far from the presumed source of discharge at Hudson Strait. At present there is no $\delta^{18}O$ indication of this event anywhere in the Labrador Sea or elsewhere in the subpolar North Atlantic. In the Norwegian Sea, where the event is found in other proxy data such as percent *N. pachyderma* s. [*Klitgaard-Kristensen et al.*, 1998; *Risebrobakken et al.*, 2003], the isotope effect of cooling probably outweighs the effect of low salinity [*Klitgaard-Kristensen et al.*, 2001], if present at all. This spatial

distribution of the $\delta^{18}O$ signal is inconsistent with geological and geophysical data on land that point to Hudson Strait as the source of 8.2 ka meltwater. We can reconcile the data on land with the paleoceanographic data only by calling on ad hoc explanations. For example, the event was too brief (as brief as one year) to be found in the Labrador Sea, or it was too fresh for foraminifera to survive. However, the data from Laurentian Fan suggest the discharge may have occurred over centuries (roughly 700 years) and if so, at least the beginning and end should have been recorded in the Labrador Sea. One could also argue that the event was trapped as a narrow coastal current in the Labrador Sea, but in the slope waters south of Newfoundland it is found well offshore.

[31] 3. The geographically restricted extent of low $\delta^{18}O$ around 8200 years ago indicates that low salinity surface water may have mixed with Gulf Stream water as it moved south in the slope water system and then returned eastward. If so, then salinity may never have become low enough to inhibit convection at sites in the open Labrador Sea and in the Norwegian-Greenland seas. This is consistent with lack of benthic Cd/Ca evidence of reduced NADW between 8.5 and 7.8 ka, despite low benthic $\delta^{13}C$.

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