Detecting Holocene changes in thermohaline circulation

L. D. Keigwin* and E. A. Boyle

Woods Hole Oceanographic Institution, Woods Hole, MA 02543; and Massachusetts Institute of Technology, Cambridge, MA 02139

Throughout the last glacial cycle, reorganizations of deep ocean water masses were coincident with rapid millennial-scale changes in climate. Climate changes have been less severe during the present interglacial, but evidence for concurrent deep ocean circulation change is ambiguous.

Evidence for abrupt millennial-scale changes in climate has held the attention of paleoclimatologists for more than a decade. The motivations for this attentiveness derive as much from what we know as what we don't know. We know that changes are periodic or quasiperiodic, that at least some of the events are global in scale, that they are often linked to thermohaline circulation changes, that they seem to occur throughout the Pleistocene, and that the changes during the last ≈8,000 yr have been modest. On the other hand, we don't know whether there is a physical forcing mechanism or whether there is some long-term resonance in the climate system, and we don't know the phasing of millennial events among geographic regions and among different parts of the climate system. Most importantly, we don't know whether human activities could worsen the effects of natural oscillations in the climate system, or even trigger independent events.

Study of the present interglacial epoch (the Holocene) at most open sea locations is complicated by the generally lower rates of sediment accumulation compared with colder climates. This is because when the climate is warmer and sea level is higher, shelves and estuaries trap sediments that would otherwise flow down the continental margin to the abyss. Accordingly, the Holocene climate signal is often attenuated by bioturbation. However, in cores of glacial ice, rates of ice accumulation are higher than during cold episodes because warm air carries more moisture. Yet despite their higher ice accumulation rates, ice cores have lower amplitude climate signals in the Holocene than in the Pleistocene (1). Thus, it appears that climate during the past 11,000 yr is probably more stable than at any other time during the past 100,000 yr.

Much of our work has focused on the paleoclimate record of the Bermuda Rise, in the northern Sargasso Sea (Fig. 1). It was there that we found production of North Atlantic Deep Water (NADW) decreased during one of the best known millennial-scale climate events, the Younger Dryas (YD) cooling that interrupted deglacial warming (2). Contrary to some skepticism (3), this finding has recently been corroborated by Bond *et al.* (4). A later study of the Bermuda Rise showed that deglacial oscillations about every 2,000 yr in climate proxy data continued right through the Holocene, although with reduced amplitude (5). This observation was taken as support that the kind of glacial to Holocene climate variability well known from land (6) was also preserved in the best deep sea sediment cores from the North Atlantic. Bond *et al.* (4) later reached the same conclusion by documenting lithologic and planktonic faunal changes with a quasiperiod of ≈1,500 yr from sediment cores in the subpolar North Atlantic.

One of the most striking characteristics of millennial scale climate events during the Pleistocene is their association with changes in thermohaline circulation (THC) in the deep North Atlantic Ocean (7–12). When NADW flow is strong, a strong northward flux of warm near surface waters in the North Atlantic maintains mass balance and warms northern Europe. Thus, an interruption of NADW flow could plunge higher latitudes of the North Atlantic region into frigid conditions. Although Holocene climate events are relatively minor on a glacial/interglacial perspective, the small Holocene changes in the polar vortex and atmospheric storminess documented by O'Brien *et al.* (1) would probably cause widespread disruption to human society if they were to occur in the future. However, at present, only the sediment grainsize study of Bianchi and McCave (13) indicates systematic THC changes in the Holocene. These authors argued that, for every millennial-scale cold episode since ≈8,000 yr ago, there was a decrease in the flux of Iceland-Scotland Overflow Water, one of the components of NADW.

Promising Candidates for Study. Of the various Holocene events, only two candidates are well enough known that we might search for associated THC changes. The first is known as the 8.2 ka event (≈8,000–8,400 yr ago) and is the most prominent decrease in $\delta^{18}O_{ice}$ (cooling) in the Holocene section of Greenland ice cores. Alley *et al.* (14) summarize the evidence that the 8.2 ka event was about half the intensity of the YD event and was globally distributed, but their estimate of ≈6°C temperature depression applies only to the atmosphere over Greenland. In the Norwegian Sea, planktonic foraminiferal assemblage changes suggest sea surface temperature decreased by ≈2°C (15). Because the response of various indicators of the climate system 8,200 yr ago was so similar to the YD, Alley *et al.* (14) speculated that it might be similarly associated with a fresh water-induced shutdown of THC, but the source of fresh water was unknown. Just recently, Barber *et al.* (16) revised the chronology for the final deglaciation of Hudson Bay, making that discharge of meltwater nearly synchronous with the 8.2 ka event. Modeling the climatic effects of fresh water input to the North Atlantic indicates that ocean convection is most sensitive to high-latitude inputs, near the regions of present-day deep water formation (17–19). Nevertheless, there are presently no paleochemical data that suggest the production of NADW was actually curtailed 8,200 yr ago. Most likely, this event has not been found on the Bermuda Rise because sedimentation rates are lowest in the early Holocene (5).

The second event is the Little Ice Age (LIA), the most recent in the long series of millennial scale climate oscillations (20). It is generally thought to have been expressed

Abbreviations: NADW, North Atlantic Deep Water; THC, thermohaline circulation; LIA, Little Ice Age; BC, box core; YD, Younger Dryas.

*To whom reprint requests should be addressed. E-mail: lkeigwin@whoi.edu.

PERSPECTIVE

SPECIAL FEATURE

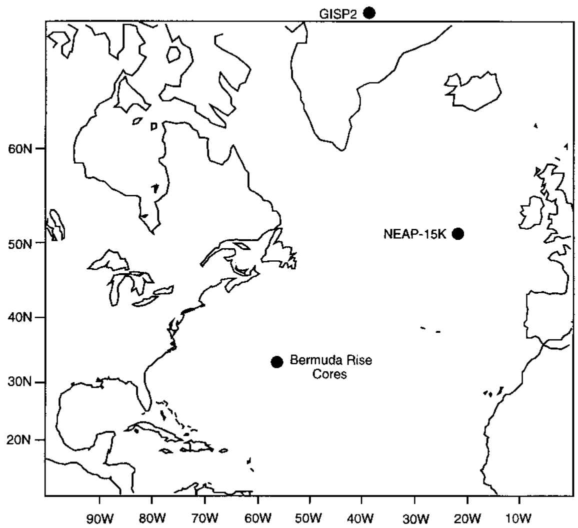

Fig. 1. Site locations discussed in text. Bermuda Rise core sites, at ≈4,600 m in the northern Sargasso Sea, have been studied for many years by the authors (2, 5, 12, 20, 27). Sedimentological evidence for thermohaline circulation change in the Holocene at core NEAP-15K (2,848-m water depth) is discussed in ref. 13, and the Holocene paleoclimate record of Greenland ice core GISP2 is discussed in ref. 1.

as a 1°C cooling in the Northern Hemisphere between ≈1500 and 1900 AD (21, 22), with slightly greater cooling (≈1.7°C) indicated by $\delta^{18}O_{ice}$ in Greenland (23). The LIA, which is well known from historical records in the North Atlantic region, had profound effects on fishing and on agriculture at high altitudes and high latitudes (24). Mounting evidence indicates that the LIA was a global event, and that its onset was synchronous within a few years in both Greenland and Antarctica (25, 26). It is also known that the last interglacial epoch (the Eemian) was climatically stable the way the Holocene is, but that the Eemian ended with an abrupt millennial-scale circulation change (27, 28). Could the LIA be analogous to the terminal Eemian event?

The Deep Ocean Record of the Little Ice Age. Because the LIA was a time of minimum carbonate content in the western North Atlantic, and because millennial-scale carbonate minima throughout the last glacial cycle were also times of suppressed deep convection in the northern North Atlantic region, we developed proxy geochemical data for deep ocean ventilation change for the LIA interval in the deep Sargasso Sea (Fig. 2). Carbon isotope ratios and Cd/Ca were measured from subcores of Box Core (BC)–004, which was described previously (20). A box core provides a large volume sample of the upper ≈50 cm of the seafloor, but it must be subsampled for detailed stratigraphic study. Benthic foraminifera are generally rare in BC-004 (Fig. 2*f*); it was

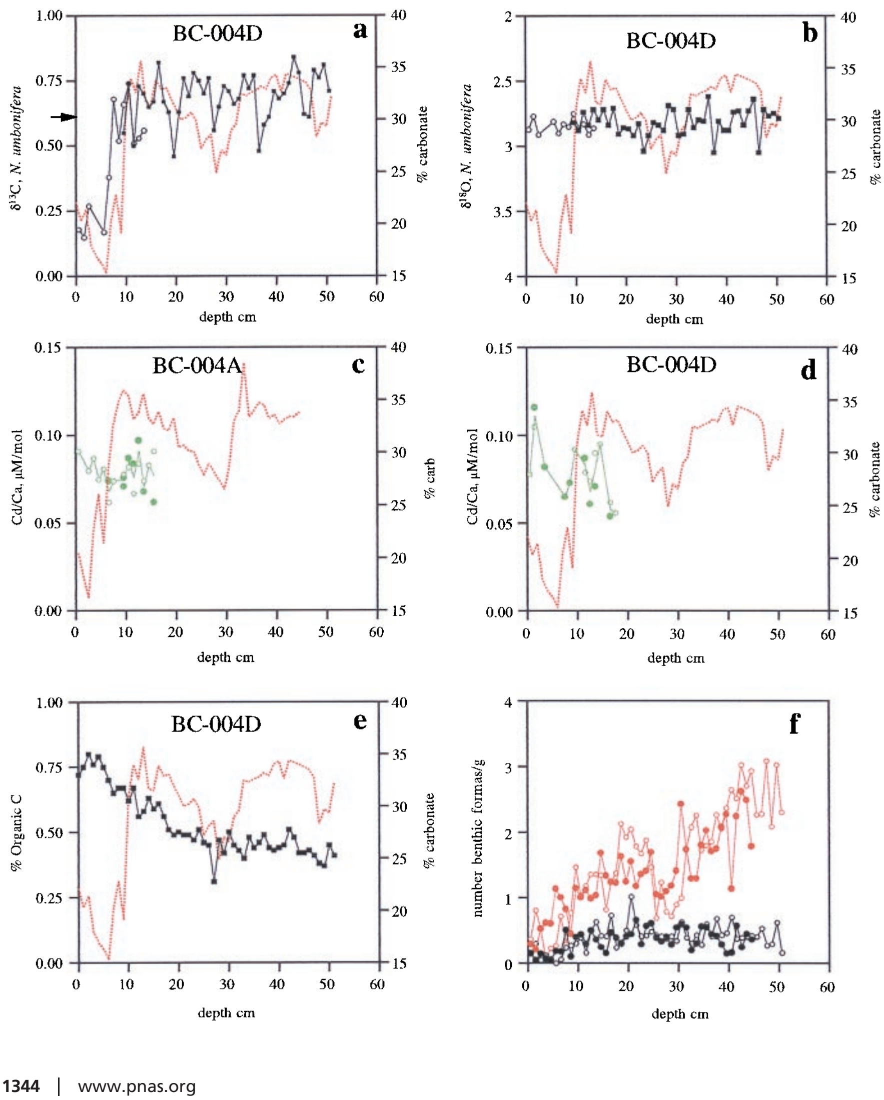

Fig. 2. New paleochemical and sedimentological data for the past ≈3 ka at Bermuda Rise core HU89038 BC-004. For *a*–*e*, the dashed red curve is the calcium carbonate content at the respective sub cores of BC-004 (after 20). In *a* and *b*, $\delta^{13}C$ and $\delta^{18}O$ (respectively), filled data points are analyses on a VG 903 mass spectrometer and open points were measured on a Finnigan 251. It was necessary to switch to the more sensitive Finnigan 251 for the stable isotope analyses in the LIA samples where the benthic foram abundance was minimal (because of higher clay and silt fluxes). In *a*, the measured bottom water value for $\delta^{13}C$ of total CO_2 on the Bermuda Rise is shown by an arrow on the *y* axis (0.66 ‰). Cd/Ca data on *N. umbonifera* (open symbols) and *Cibicidoides* (filled symbols) are shown for the subcores in *c* and *d*, with the average value for each depth shown as a solid line. *e* shows the percent organic carbon in BC-004D, and *f* presents the abundance of *Cibicidoides* (filled black circles, BC-004A; open black circles, BC-004D) and the abundance of *N. umbonifera* (filled red circles, BC-004A; open red circles, BC-004D).

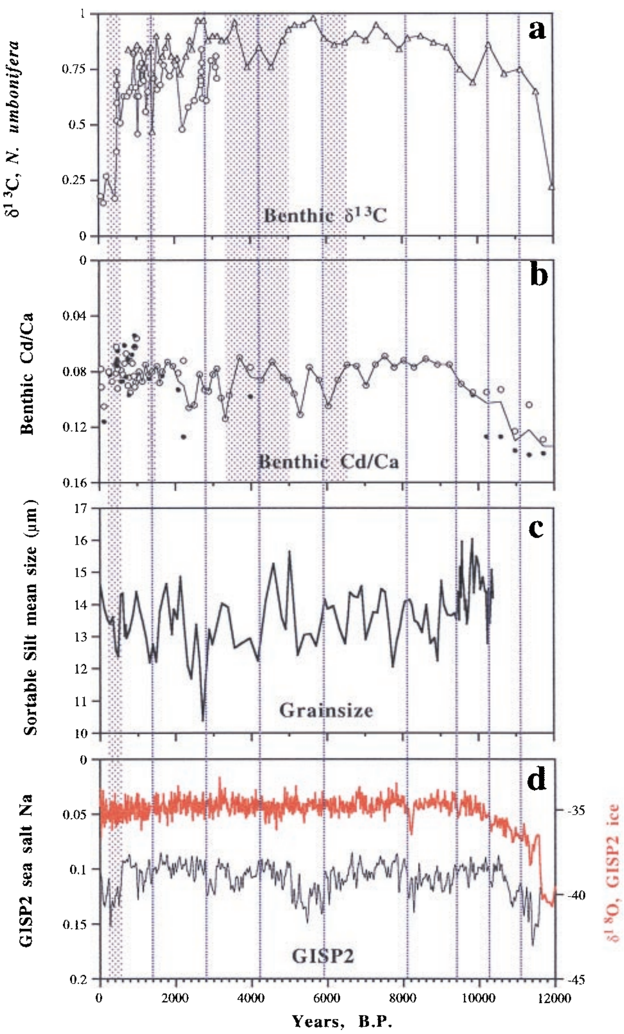

Fig. 3. Time series [(calendar years before present (B.P.)] of paleoclimatic and paleoceanographic data from the North Atlantic region. (*a*) The $\delta^{13}C$ data from Fig. 2*a* plotted with previously unpublished $\delta^{13}C$ data on *N. umbonifera* from Bermuda Rise core KNR31 GPC-5. (*b*) The Cd/Ca data from Fig. 2 *c* and *d* plotted with the data from ref. 2 (filled points, *Cibicidoides*; open points, *N. umbonifera*). Mean size of sortable silt on Gardar Drift in the NE Atlantic (13) are in *c*. A decrease in grainsize would indicate decreased flow (less winnowing) of Iceland-Scotland Overflow Water. Holocene data from Greenland ice core GISP2 (*d*) show little variability in air temperature over Greenland ($\delta^{18}O_{ice}$ data in red), except during the 8.2 ka event, and greater variability in sea salt Na concentration (black) at that and other times, such as the Little Ice Age (after ref. 1). Dashed lines running through all four panels mark the age of Holocene cold events identified by Bond *et al.* (4), and stippled bands mark the late Holocene cold events identified on the Bermuda Rise (5, 20).

necessary to analyze one species (*Nutallides umbonifera*) for $\delta^{13}C$ in one subcore (BC-004D), and Cd/Ca on both *Cibicidoides spp.* (mostly *Cibicidoides wuellerstorfi*) and *N. umbonifera* in subcores A and D.

Isotopic results show good agreement between the two instruments (see Fig. 2 caption) with typical late Holocene values for $\delta^{18}O$ (Fig. 2*b*), and similar $\delta^{13}C$ values where data from the two laboratories overlap (Fig. 2*a*). However, whereas most of the $\delta^{13}C$ data from the 12- to 50-cm interval in the core are typical of the expected late Holocene seawater value at this location, the LIA samples are marked by substantially lower values (Fig. 2*a*). If these data accurately reflect bottom water nutrient levels, then during the LIA NADW was nearly completely replaced by nutrient-rich circumpolar water in the deep western North Atlantic.

In contrast, Cd/Ca appears to increase only slightly in the same samples (Fig. 2 *c* and *d*). This increase is driven by samples in only the upper 2 cm of the cores, and is markedly less than what would be expected from a 0.5 ‰ drop in $\delta^{13}C$. The two tracers cannot both be reflecting deep ocean nutrient levels accurately. Although neither proxy is perfect, we do not presently have enough information to determine which is more reliable. If the isotopic data are taken literally, we must question why $\delta^{13}C$ does not return to modern (NADW) values at the top of the core. Box core tops are sometimes disturbed in the recovery process, and the core is bioturbated, but those processes should influence the trace metal and isotopic data equally. Carbon isotope ratios are also known to be influenced by the rain rate of organic carbon (29, 30), but those problems are thought to be of greatest concern in upwelling regions, not the centers of subtropical circulation like the Sargasso Sea. Organic carbon content at our core site does indeed decrease from the core top to background values of ≈0.4% (Fig. 2*e*), but that is probably typical of the diagenetic burndown of labile organic material at most deep sea locations. If we were to calculate the mass accumulation rate of organic carbon, there would be a large spike during the LIA because the rate of sedimentation spikes (20), but that artifact would not be preserved in the sedimentary record as the percent organic carbon decreases to a steady state value.

On the other hand, Cd/Ca in *C. wuellerstorfi* is thought to be lowered by carbonate dissolution in corrosive bottom waters (31). Our unpublished observations indicate poorer preservation of the foraminiferal fauna during the LIA in these samples, so it is possible that carbonate dissolution has lowered Cd/Ca in LIA samples of *C. wuellerstorfi*. Work in progress shows that Cd/Ca in *N. umbonifera* is not affected by bottom water undersaturation (E.A.B., unpublished work). Hence, we caution that it is premature to conclude that there was actually a change in THC associated with the LIA.

Other North Atlantic Indicators of Holocene Climate. One test of our Bermuda Rise observations is to study shallower locations where the potential for carbonate dissolution is not so great. Unfortunately, at present only a few open marine locations are known to resolve the LIA, and shallower sites might not be as sensitive to changes in the densest components of NADW. Another way to evaluate LIA results is in the context of other time series from the Holocene (Fig. 3). When our $\delta^{13}C$ data are plotted with results from other Bermuda Rise cores, it is clear that the LIA decrease is unprecedented in the Holocene. The only previous time when $\delta^{13}C$ was so low was during the YD event of ≈12 ka (Fig. 3*a*). Earlier Holocene events could be muted in this series because of lower rates of sedimentation, but the fact that LIA $\delta^{13}C$ was as low as the YD is surprising in light of the Alley *et al.* (14) observation that the 8.2 ka event was only about one-half the intensity of the YD. In contrast, the smaller LIA cooling is not noticeable in $\delta^{18}O_{ice}$ series in Fig. 3*d*, but both the LIA and the 8.2 ka event are marked by increased sea salt Na that reflects atmospheric storminess (1). Compared with earlier Holocene data, the LIA maximum in Cd/Ca ≈100 yr ago does not appear to be substantial (Fig. 3*b*). Similar Cd/Ca maxima were achieved several other times during the Holocene, but none of those maxima approached the level of NADW decrease found during the YD, and those maxima are not clearly associated with the Holocene cold episodes noted previously (4, 5, 20). As noted by Bianchi and McCave (13), grainsize evidence for changes in the Iceland-Scotland Overflow Water indicates variability close to the ≈1,500-yr quasiperiod in ice rafting reported by Bond *et al.* (4), but the finer grained sediment deposited during the LIA does not particularly stand out in the record (Fig. 3*c*). Furthermore, although both the ice-rafted debris episodes and the grainsize changes may have similar periodicities in their respective cores, there is not an exact match between every ice-rafted debris event and every grainsize event. Likewise, the ice-rafted debris events are not clearly correlated with excursions in benthic $\delta^{13}C$ or Cd/Ca. In fact, there is surprisingly little agreement among the proxy data in Fig. 3, except possibly during the LIA.

Where To Go from Here? Our view is that the role of possible THC changes during the LIA is so important that it deserves special attention. As Broecker *et al.* (26) point out, if deep watermass changes have occurred in the past several hundred years, then the ocean is not in steady state and that will change many of our assumptions. Of the many millennial-scale climate changes known from the last glacial cycle, the LIA is the only one documented by contemporary humans (24), and it occurred right at the beginning of modern instrumental observations. Thus, as climatologists attempt to extend instrumental observations with climate proxy data, paleoceanographers will bear a special burden in determining whether the LIA climate change was amplified by THC changes. Considering the importance of identifying THC changes during the most recent of the climate system's millennial-scale oscillations, we recommend that at least two proxies for paleocirculation be brought to bear at critical core sites. That we have failed to generate conclusive results at a site as well known as the Bermuda Rise highlights the challenges ahead.

We thank E. Franks and E. Roosen for assistance in the laboratory. This research was funded by National Science Foundation Grants OCE-9709686 and 9402198.

1. O'Brien, S. R., Mayewski, P. A., Meeker, L. D., Meese, D. A., Twickler, M. S. & Whitlow, S. I. (1995) *Science* **270,** 1962–1964.
2. Boyle, E. A. & Keigwin, L. D. (1987) *Nature (London)* **330,** 35–40.
3. Sarnthein, M., Winn, K., Jung, S. J. A., Duplessy, J.-C., Labeyrie, L., Erlenkeuser, H. & Ganssen, G. (1994) *Paleoceanography* **9,** 209–267.
4. Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I. & Bonani, G. (1997) *Science* **278,** 1257–1266.
5. Keigwin, L. D. & Jones, G. A. (1989) *Deep-Sea Res.* **36,** 845–867.
6. Denton, G. H. & Karlen, W. (1973) *Quat. Res.* **3,** 155–205.
7. Oppo, D. W. & Lehman, S. J. (1995) *Paleoceanography* **10,** 901–910.
8. Charles, C. D., Lynch-Stieglitz, J., Ninnemann, U. S. & Fairbanks, R. G. (1996) *Earth Planet. Sci. Lett.* **142,** 19–27.
9. Curry, W. B. & Oppo, D. W. (1997) *Paleoceanography* **12,** 1–14.
10. Vidal, L., Labeyrie, L., Cortijo, E., Arnold, M., Duplessy, J. C., Michel, E., Becque, S. & van Weering, T. C. E. (1997) *Earth Planet. Sci. Lett.* **146,** 13–27.
11. Zahn, R., Schonfeld, J., Kudras, H. R., Park, M.-H., Erlenkeuser, H. & Grootes, P. (1997) *Paleoceanography* **12,** 696–710.
12. Keigwin, L. D. & Boyle, E. A. (1999) *Paleoceanography* **14,** 164–170.
13. Bianchi, G. G. & McCave, I. N. (1999) *Nature (London)* **397,** 515–517.
14. Alley, R. B., Mayewski, P. A., Sowers, T., Stuiver, M., Taylor, K. C. & Clark, P. U. (1997) *Geology* **25,** 483–486.
15. Klitgaard-Kristensen, D., Sejrup, H. P., Haflidason, H., Johnsen, S. & Spurk, M. (1998) *J. Quat. Sci.* **13,** 165–169.
16. Barber, D. C., Dyke, A., Hillaire-Marcel, C., Jennings, A. E., Andrews, J. T., Kerwin, M. W., Bilodeau, G., McNeely, R., Southon, J., Morehead, M. D. & Gagnon, J.-M. (1999) *Nature (London)* **400,** 344–348.
17. Manabe, S. & Stouffer, R. J. (1993) *Nature (London)* **364,** 215–218.
18. Delworth, T. L., Manabe, S. & Stouffer, R. J. (1997) *Geophys. Res. Lett.* **24,** 257–260.
19. Mauritzen, C. & Hakkinen, S. (1997) *Geophys. Res. Lett.* **24,** 3257- 3260.
20. Keigwin, L. D. (1996) *Science* **274,** 1504–1508.
21. Bradley, R. S. & Jones, P. D. (1993) *The Holocene* **3,** 367–376.
22. Dahl-Jensen, D., Mosegard, K., Gundestrup, N., Clow, G. D., Johnsen, S. J., Hansen, A. W. & Balling, N. (1998) *Science* **282,** 268–271.
23. Stuiver, M., Grootes, P. M. & Braziunas, T. F. (1995) *Quat. Res.* **44,** 341–354.
24. Grove, J. (1988) *The Little Ice Age* (Methuen, New York).
25. Kreutz, K., Majewski, P. A., Meeker, L. D., Twickler, M. S., Whitlow, S. I. & Pittalwala, I. I. (1997) *Science* **277,** 1294–1296.
26. Broecker, W. S., Sutherland, S. & Peng, T.-H. (1999) *Science* **286,** 1132–1135.
27. Adkins, J. F., Boyle, E. A., Keigwin, L. & Cortijo, E. (1997) *Nature (London)* **390,** 154–156.
28. Hall, I. R., McCave, I. N., Chapman, M. R. & Shackleton, N. J. (1998) *Earth Planet. Sci. Lett.* **164,** 15–21.
29. Zahn, R., Winn, K. & Sarnthein, M. (1986) *Paleoceanography* **1,** 27–42.
30. Mackensen, A., Hubberton, H. W., Bickert, T., Fischer, G. & Futterer, D. K. (1993) *Paleoceanography* **8,** 587–610.
31. McCorkle, D. C., Martin, P. A., Lea, D. W. & Klinkhammer, G. P. (1995) *Paleoceanography* **10,** 699–714.