Reports

Deep Circulation of the North Atlantic over the Last 200,000 Years: Geochemical Evidence

Abstract. *Variations in the cadmium/calcium ratio of North Atlantic Deep Water are recorded in the fossil shells of benthic foraminifera. The oceanic distribution of cadmium is similar to that of the nutrients, hence the cadmium/calcium ratio in shells records temporal variations in nutrient distributions. Data from a North Atlantic sediment core show that over the past 200,000 years there has been a continuous supply of nutrient-depleted waters into the deep North Atlantic. The intensity of this source relative to nutrient-enriched southern waters diminished by about a factor of 2 during severe glaciations. This evidence combined with carbon isotope data indicates that the continental carbon inventory may have been less variable than previously suggested.*

The CLIMAP program built upon advances in marine micropaleontology (*1*), isotopic geochemistry (*2, 3*), and Pleistocene chronology (*4, 5*) to map in considerable detail the variability of surface ocean temperature and circulation (*6, 7*). Less is known about the response of deep ocean circulation to the dramatic climate variations associated with the Pleistocene glacial advances and retreats, although Weyl (*8*) and Newell (*9*) have suggested on theoretical grounds that major changes could be expected. Duplessy *et al.* (*10*) have provided isotopic evidence suggesting that the Norwegian Sea was not a source of cold deepwater formation during the most recent glacial maximum. However, in a later study (*11*) they suggested that there nevertheless must have been some source of cold deep water to the North Atlantic. Streeter and Shackleton (*12*) used abundance variations in benthic foraminiferal species to suggest that the net production of North Atlantic Deep Water (NADW) was substantially diminished or that it had even ceased during the last glaciation. However, an unambiguous quantitative interpretation of population changes in benthic foraminifera has not been possible. Early reports suggesting simple correlations with deepwater parameters (*13*) have given way to more subtle and complex interpretations (*14*). Curry and Lohmann (*15*) used $\delta^{13}C$ data from benthic foraminifera in the Vema Channel to argue that NADW flow is reduced during glacial periods; the low sedimentation rates of their cores precluded detailed resolution of the change, however. Therefore, although it is plausible that deep ocean circulation varies during climatic change, the nature of this response is uncertain.

The oceanic distribution patterns of many chemical properties reflect deep ocean circulation patterns (*16, 17*). In particular, in the North Atlantic the formation of dense deep water from shallow nutrient-depleted waters of subtropical origin produces a low-nutrient "tongue" along the southerly spreading direction of NADW (*17*). If a record of past chemical distribution patterns in the deep ocean could be established, then deep ocean flow patterns in the past could be inferred. The Cd content of calcitic foraminiferal tests is one such indicator (*18*), since Cd is linearly related to phosphate in the deep ocean and the Cd content of certain benthic foraminifera reflects bottom-water chemistry (*19*). While at the surface, ocean water loses phosphate and Cd through biological activity and equilibrates with atmospheric O_2. The biogenic debris [with a Cd/phosphate ratio of 3.5×10^{-4} (*19*)] which settles out of the surface layers is remineralized at depth. "Older" deep water, therefore, contains less O_2 and more phosphate and Cd than water more recently at the surface. Because of the relation between O_2 consumption and the average chemical composition of organisms elaborated on by Redfield *et al.* (*20*), O_2-depleted water has high Cd concentrations, about 1.2 nmole/kg.

We have studied in detail the Cd/Ca, $\delta^{18}O$ (*15*), and $\delta^{13}C$ variability in benthic foraminifera from piston core *Chain* 82, station 31, core 11PC, to establish the timing and magnitude of chemical changes over the last 200,000 years. The core was raised from a water depth of 3209 m (near the NADW core density level) on the western flank of the Mid-Atlantic Ridge about 300 km northwest of the Azores (42°N, 32°W). It contains well-preserved planktonic and benthic foraminifera throughout its length. The average sedimentation rate over the interval studied is 2.6 cm per 1000 years. After pretreatment (*21*), we determined Cd at the Massachusetts Institute of Technology by using minor modifications of the graphite furnace atomic absorption spectrophotometric procedures described in (*18, 22*). Measurements of $\delta^{18}O$ and $\delta^{13}C$ were made at Woods Hole Oceanographic Institution on a Micromass 602E mass spectrometer (*23*). The isotopic and chemical data for this core are listed in Table 1.

The $\delta^{18}O$ record (Fig. 1) is readily correlated with the standard sequence of

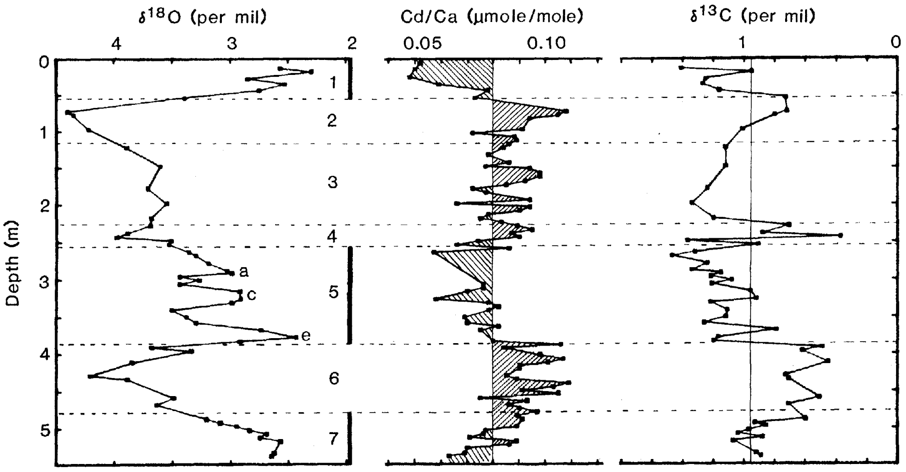

Fig. 1. Data from core *Chain* 82, station 31, core 11PC (42°N, 32°W; 3209 m). The values for $\delta^{18}O$ and $\delta^{13}C$ are relative to PDB.

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isotope stages and substages to stage 7 (*2, 24*). The range of values (from 2.3 per mil at 18 cm to 4.4 per mil at 73 cm) is comparable to that of other Atlantic sediment cores with high accumulation rates (*7, 12, 25*). The $\delta^{18}O$ record demonstrates that accumulation in the core has occurred without detectable hiatuses. Although benthic foraminiferal $\delta^{18}O$ is a function of three variables (the isotopic composition of the water, the temperature of the water, and the species analyzed), the major downcore variations are believed to result from isotopic changes in the world ocean due to fluctuations in the mass of isotopically light ice (*1, 3, 4*).

The benthic foraminifera from this core do not record Cd concentrations as high as the same genera in the deep Pacific [0.18 to 0.20 μmole of Cd per mole of Ca (*19*)]. Complete homogenization of the deep water of the world ocean would result in an average Cd/Ca ratio of 0.15 μmole per mole. Therefore, at no time has the flux of oxygenated, low-nutrient water to this location ceased. Significant Cd/Ca variations are correlated with the climatic record: the highest Cd/Ca ratio is observed in periods of extreme glacial climate (isotope stages 2 and 6), with lower values in periods of milder climate (isotope stages 1, 5, and 7). This relationship demonstrates that there have been climate-correlated changes in deep ocean circulation through time. There is not an exact synchroneity between $\delta^{18}O$ and Cd/Ca ratio. Although decoupling has been suggested on the basis of theory (*8, 9*), we believe that it is premature to elaborate on the significance of the observed differences until the phenomenon is confirmed in other cores.

The quantitative significance of the

Table 1. Oxygen and carbon isotope data (*Cibicidoides wuellerstorfi* except where noted) and Cd/Ca data for core *Chain* 82, station 31, core 11PC. The values for $\delta^{18}O$ and $\delta^{13}C$ are relative to PDB. The letters after the Cd/Ca data indicate the variety analyzed (u = *Uvigerina* spp. and c = *Cibicidoides kullenbergi*). The first number after each letter indicates the number of replicates included in the value reported and the second number specifies the number of analyses rejected from the mean. For Cd/Ca, points at 18, 28, 48, 61, 68, 88, 188, 346, 401, and 424 cm were also rejected; these were samples for which only a single analysis was possible and which appeared to be contaminated.

Depth	$\delta^{18}O$ (per mil)	$\delta^{13}C$ (per mil)	Cd/Ca (μmole/mole)	Depth	$\delta^{13}C$ (per mil)	$\delta^{18}O$ (per mil)	Cd/Ca (μmole/mole)
4–8			0.052 (c, 2, 1)	285–288	3.03	1.34	
11–15	2.57	1.41	0.050 (c, 1, 0)	288–293	2.99	1.15	
15–20	2.30	0.95		293–298	3.44	1.22	
23–27			0.048 (c, 1, 0)	298–303	3.27	1.08	
25–30	2.85	1.25		303–308	3.44	1.21	
33–37	2.53	1.27	0.059 (c, 1, 0)	308–313			0.076 (u, 1, 0)
41–46	2.75	1.16	0.078 (u, 1, 0)	313–318	2.92	0.96	0.070 (u, 2, 0)
51–56	3.54	0.64	0.082 (c, 1, 0)	323–328	2.91	0.92	0.058 (u, 2, 0)
51–56	3.23	0.82	0.063 (u, 1, 0)	328–331	2.99	1.22	0.078 (u, 2, 0)
70–75	4.40	0.72	0.108 (u, 1, 0)	328–331			0.096 (c, 1, 0)
75–80	4.35	0.80	0.105 (u, 1, 0)	333–338			0.082 (u, 2, 0)
80–85			0.094 (u, 2, 0)	338–343	3.51	1.11	0.078 (u, 1, 0)
95–100	4.22	1.01	0.091 (u, 1, 0)	348–350	3.38	1.12	0.069 (u, 2, 1)
100–105			0.072 (u, 2, 0)	355–360	3.30	1.26	0.070 (u, 2, 0)
105–110			0.088 (u, 3, 0)	360–365			0.082 (u, 2, 0)
110–115			0.089 (u, 3, 0)	365–370	2.74	0.79	0.075 (u, 1, 0)
115–120			0.086 (u, 7, 0)	375–380	2.44	1.17	
120–125	3.89	1.12	0.084 (u, 3, 0)	380–385	2.92	1.20	0.080 (c, 1, 0)
130–135			0.078 (u, 2, 1)	385–388			0.106 (u, 2, 0)
140–145			0.086 (u, 3, 0)	388–393	3.68	0.49	0.084 (u, 3, 0)
145–150	3.60	1.12	0.077 (u, 2, 0)	393–398	3.34	0.62	
150–152			0.094 (u, 2, 0)	398–403			0.098 (u, 2, 0)
154–159			0.098 (u, 2, 0)	403–408			0.107 (u, 5, 0)
159–164			0.098 (u, 4, 3)	408–413	3.85	0.45	0.101 (u, 3, 0)
165–170			0.092 (u, 1, 0)	413–417			0.090 (u, 2, 0)
170–175			0.085 (u, 2, 0)	417–422			0.090 (u, 2, 0)
175–180	3.71	1.24	0.072 (u, 3, 2)	426–431	4.21	0.73	0.085 (u, 3, 0)
181–185			0.077 (u, 2, 1)	431–436	3.89†	0.71†	0.089 (u, 2, 2)
190–195			0.094 (u, 2, 1)	436–441			0.109 (u, 2, 0)
195–200	3.55	1.34	0.066 (u, 2, 0)	441–446			0.103 (u, 1, 1)
200–205			0.094 (u, 2, 0)	446–451			0.091 (u, 1, 0)
205–210			0.090 (u, 7, 1)	451–455			0.105 (u, 1, 0)
210–215			0.078 (u, 1, 0)	457–460	3.49	0.51	0.075 (u, 1, 0)
215–220	3.68	1.20	0.075 (u, 2, 0)	460–465			0.093 (u, 2, 0)
220–225			0.083 (u, 3, 0)	465–470	3.64	0.71	0.086 (u, 1, 1)
225–230	3.53	0.66	0.089 (u, 2, 1)	470–475			0.090 (u, 2, 0)
225–230	3.85	0.77		475–480			0.097 (u, 2, 0)
231–235			0.095 (u, 3, 1)	480–485			0.089 (u, 2, 1)
235–240	3.89	0.88	0.087 (u, 2, 0)	485–490	3.21	0.60	0.091 (u, 2, 1)
240–245	3.98	0.37	0.090 (u, 2, 0)	490–495	3.09	0.93	
245–250	3.51	1.37	0.074 (u, 2, 0)	495–500	2.95	0.86	0.089 (u, 5, 0)
250–255	3.53	0.91	0.066 (u, 2, 0)	500–505	2.84	0.97	0.077 (u, 2, 0)
255–260			0.086 (u, 1, 0)	505–510	2.69	1.04	0.076 (u, 2, 0)
260–265	4.09*	1.21		510–515	2.75	0.88	0.071 (u, 2, 0)
260–265	3.44	1.42	0.057 (u, 1, 0)	515–520	2.57	1.07	0.089 (u, 3, 0)
260–265	3.28	1.30		520–525			0.086 (u, 2, 0)
265–270	3.45	1.50		525–530			0.070 (u, 1, 1)
265–270	3.15	1.43		532–537	2.63	0.92	0.069 (u, 1, 0)
275–280	3.19	1.24		537–541	2.65	0.89	0.063 (u, 1, 0)

*Suspect analysis. †*Cibicidoides* spp. This variety has been shown to exhibit responses to bottom-water conditions identical to that of *C. wuellerstorfi* (*23*).

variations recorded in this core can be examined with the use of a simple model of NADW chemical composition. The chemical composition of a variable component such as Cd in the NADW (C_D) is simply a flux-weighted average of the concentration in the waters that combine to form this water mass (*26*). These sources can be simplified to be represented by nutrient-depleted surface waters (C_S and F_S representing the surface-water concentration and flux, respectively) and entrained quantities of intermediate water (C_I and F_I) and Antarctic bottom water (C_B and F_B):

$$C_D = \frac{C_S F_S + C_I F_I + C_B F_B}{F_S + F_I + F_B}$$

where $F_D = F_S + F_I + F_B$. But since $C_S = 0$ and the phosphate concentrations in the "southern source" (SS) waters are nearly identical (*17*), this relation reduces to

$$C_D \simeq \frac{F_{SS}}{F_D} C_{SS}$$

In other words, the concentration of Cd in the NADW is inversely proportional to NADW flow rate. All other things being constant, the doubling of Cd recorded in this core would signify a halving of NADW flux. Even though it can be argued on the basis of the oceanic geochemical mass balance that C_{SS} cannot vary significantly (*27*), it is not clear that the flux of SS water has been constant. In the absence of other constraints on the flow of SS waters, we can only conclude that during glaciation the flux from nutrient-depleted NADW sources has diminished in importance relative to that of the southern deep and intermediate sources.

The Cd/Ca data also are relevant to the interpretation of $\delta^{13}C$ data in the benthic foraminifera of deep-sea cores. Shackleton (*25*) noted that $\delta^{13}C$ in core M12392 decreased by about 1 per mil during glacial periods. He attributed this change to a transfer of isotopically depleted continental reduced C (from tropical rain forests and northern hardwood forests) into the ocean during glaciation. Broecker (*28*) argued that this change was instead due to a transfer of organic C from the continental shelf into the deep sea. Our core also shows changes in $\delta^{13}C$ of similar magnitude during extreme glacial periods, but the record differs considerably from that of core M12392 in that our core has isotope stage 3 $\delta^{13}C$ values similar to those of stage 1. This record, however, is similar to that of other cores from the North Atlantic Ocean (*29*).

Regional benthic $\delta^{13}C$ differences imply that some oceanic basins may have contained different water masses in the past. Since the Cd signal reflects changes in the deep-ocean circulation patterns which also influence the distribution of $\delta^{13}C$ (*30*), Cd/Ca and $\delta^{13}C$ data from cores throughout the ocean can be used to evaluate the relative importance of C transfer and circulation changes on abyssal $\delta^{13}C$. Our evidence for changes in deep-ocean circulation patterns suggests that the magnitude of the C transfer signal must be less than previously suggested (*25, 28, 31*). Over the past 65,000 years (isotope stages 1 through 3), all paired analyses of $\delta^{13}C$ and Cd/Ca fall on the present-day relation between these variables. Shackleton's hypothesis (*25*) would require an oceanwide offset in the relationship between $\delta^{13}C$ and Cd/Ca. A quantitive evaluation of Broecker's hypothesis (*28*) requires more data and a more detailed analysis than is possible in this report. However, our interpretation of changing deep-ocean circulation patterns suggests that at least part of the $\delta^{13}C$ signal is not due to a C flux from the continental shelf.

From the Cd data presented here we conclude that the glacial NADW, while exhibiting a continuous flow with no evidence of cessation, is diminished relative to the flux of Antarctic water sources. In addition, benthic $\delta^{13}C$ data may reflect both circulation changes and transfers of reduced C between continents and ocean.

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References and Notes

1. J. Imbrie and N. Kipp, in *The Late Cenozoic Glacial Ages*, K. K. Turekian, Ed. (Yale Univ. Press, New Haven, Conn., 1971), p. 71.
2. C. Emiliani, *J. Geol.* **63**, 538 (1955).
3. N. Shackleton and N. Opdyke, *Quat. Res. (N.Y.)* **3**, 39 (1973).
4. W. Broecker and J. van Donk, *Rev. Geophys. Space Phys.* **8**, 169 (1970).
5. K. J. Mesolella, R. K. Matthews, W. S. Broecker, D. L. Thurber, *J. Geol.* **77**, 250 (1969).
6. R. M. Cline and J. D. Hays, *Geol. Soc. Am. Mem. 145* (1976); CLIMAP Project Members, *Science* **191**, 1131 (1976); J. D. Hays, J. Imbrie, N. J. Shackleton, *ibid.* **194**, 1121 (1976).
7. W. F. Ruddiman and A. McIntyre, *Science* **212**, 617 (1981).
8. P. K. Weyl, *Meteorol. Monogr.* **8**, 37 (1968).
9. R. Newell, *Quat. Res. (N.Y.)* **4**, 117 (1974).
10. J. C. Duplessy, L. Chenouard, F. Vila, *Science* **188**, 1208 (1975).
11. J. C. Duplessy, J. Moyes, C. Pujol, *Nature (London)* **283**, 479 (1980).
12. S. S. Streeter and N. J. Shackleton, *Science* **203**, 168 (1979).
13. S. S. Streeter, *Quat. Res. (N.Y.)* **3**, 131 (1973); G. P. Lohmann, *J. Foraminiferal Res.* **8**, 6 (1978); D. Schnitker, *Nature (London)* **248**, 385 (1974); B. H. Corliss, *Mar. Geol.* **31**, 115 (1979).
14. P. E. Belanger and S. S. Streeter, *Mar. Micropaleontol.* **5**, 401 (1980); M. Bremer and G. P. Lohmann, *Deep-Sea Res.*, in press; B. H. Corliss, in preparation.
15. W. Curry and G. P. Lohmann, *Quat. Res. (N.Y.)*, in press. The carbon isotopic ratio is defined as $\delta^{13}C$ =

$$\frac{R_{\text{sample}} - R_{\text{standard}}}{R_{\text{standard}}} \times 1000$$

where R = $^{13}C/^{12}C$ of the Pee Dee belemnite (PDB) standard. The oxygen isotopic ratio is defined as $\delta^{18}O$ =

$$\left[\frac{(^{18}O/^{16}O)_{\text{sample}}}{(^{18}O/^{16}O)_{\text{standard}}} - 1\right] \times 1000$$

16. J. L. Reid, *Limnol. Oceanogr.* **7**, 287 (1962).
17. A. Bainbridge, Ed., *GEOSECS Atlantic Expedition* (National Science Foundation, Washington, D.C., 1980), vol. 2.
18. E. A. Boyle, *Earth Planet. Sci. Lett.* **53**, 11 (1981).
19. K. Hester and E. Boyle, *Nature (London)* **298**, 260 (1982).
20. A. Redfield, B. Ketchum, F. Richards, in *The Sea*, M. N. Hill, Ed. (Interscience, New York, 1963), vol. 2, p. 26.
21. Samples 2 to 6 cm thick were taken continuously down the length of the core. The samples were shaken in dilute ammonium phosphate solution for about 2 hours, wet-sieved through a 150-μm nylon sieve, and dried. *Uvigerina* spp. and *Cibicidoides* spp. were handpicked for chemical and isotopic analysis.
22. The modifications were shown to produce data equivalent to that obtained in (*18*) but aided speed and reproducibility. The samples were first cleaned in 0.1*N* NaOH with a small amount of H_2O_2 added. The new reducing reagent was hydrazine. Samples were dissolved in 0.1*N* HNO_3. The sample size was approximately 1 mg (15 to 40 individuals). Replicates were analyzed when sample size permitted. The pooled standard deviation of the replicates indicates a 1σ precision of ± 0.005 μmole of Cd per mole of Ca. At the low absolute Cd concentrations encountered, contamination is a problem. Occasionally analyses indicated high concentrations (> 0.03 μmole of Cd per mole of Ca) off the trend of the data; these samples were reanalyzed where possible. If the new analysis confirmed that contamination had occurred, the high point was rejected. Low points were never rejected. About 15 percent of the total number of analyses were rejected on the basis of these criteria.
23. The analytical procedures are similar to those described by L. D. Keigwin, [*Earth Planet. Sci. Lett.* **45**, 361 (1979)], with the following exceptions. Both CO_2 and H_2O are condensed in a liquid nitrogen trap as the sample reacts and the National Bureau of Standards calcium carbonate standard NBS-20 (solenhojen limestone) is used as a working standard. Analyses of 34 NBS-20 samples during days when core 11PC samples were analyzed gave a reproducibility (1σ) of 0.08 and 0.06 percent for $\delta^{18}O$ and $\delta^{13}C$, respectively. Results were corrected to the PDB standard, where $\delta^{18}O$ NBS-20 = −4.18 per mil and $\delta^{13}C$ NBS-20 = −1.06 per mil. The species *Cibicidoides wuellerstorfi* was used in all but one sample because of its presence throughout the core and because it has been demonstrated to track bottom-water variations in temperature and $\delta^{13}C$ [P. E. Belanger, W. B. Curry, R. K. Matthews, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **33**, 205 (1981); D. W. Graham, B. H. Corliss, M. L. Bender, L. D. Keigwin, *Mar. Micropaleontol.* **6**, 483 (1981); F. Woodruff, S. M. Savin, R. G. Douglas, *Science* **212**, 665 (1981); L. D. Keigwin, Jr., *Init. Rep. Deep Sea Drilling Proj.*, in press; (*11*)].
24. N. J. Shackleton, *Philos. Trans. R. Soc. London Ser. B* **280**, 169 (1977).
25. N. J. Shackleton, in *The Fate of Fossil Fuel CO_2 in the Oceans*, N. Andersen and A. Malahoff Eds. (Plenum, New York, 1977), p. 401.
26. Biogeochemical cycling can redistribute elements vertically, but most of the transfer occurs in shallow waters, however, recycling nutrients between the surface and intermediate water masses with very little leakage into the deeper waters.
27. Over periods of time short relative to the residence times of an element (say, 2×10^4 years relative to a Cd residence time of 2×10^5 years), the steady-state mechanisms governing the oceanic inventory of the element cannot significantly change the total amount of the element in the ocean. The distribution of the element throughout the ocean can change dramatically, given an appropriate change in ocean circulation patterns. But, since the volume of the high-Cd

Pacific and Indian ocean waters (10^9 km^3) is much greater than that of the low-Cd Atlantic (0.3×10^9 km^3), redistribution of deep-ocean Cd can produce only slight changes in the SS Cd concentration. Transient nonsteady-state mechanisms could alter the oceanic inventory of an element. Broecker (*28*) argued that the δ^{13}C shift in benthic foraminifera could be caused by a transfer of continental-shelf C into the deep ocean and that this C was accompanied by "Redfield ratio" P, increasing the total P inventory by one-third. We argue that such a transfer is less likely for Cd because, during the weathering of organic sediments, much of the Cd originally contained in organic matter will be trapped into Fe and Mn oxides. This assumption is verifiable on the basis of analyses of Cd/Ca in benthic foraminifera from cores in the Pacific and Indian oceans (E. A. Boyle and L. D. Keigwin, in preparation).

28. W. Broecker, *Prog. Oceanogr.* **11**, 151 (1982).
29. This difference is also observed in core GPC 9 [W. Curry and G. P. Lohmann, *Geol. Soc. Am. Abstr. Programs* xx, 435 (1981)] and in core V30-97 (R. Fairbanks, personal communication).
30. P. Kroopnick, *Earth Planet. Sci. Lett.* **49**, 469 (1980).
31. More recent δ^{13}C data (N. J. Shackleton, personal communication) from the Pacific Ocean also demonstrate that the global δ^{13}C shift is less than observed in core M12392.
32. We thank M. Bender, W. Broecker, W. Curry, and B. Corliss for comments on the manuscript. J. Broda provided assistance in the selection and sampling of sediment cores. Core curation at Woods Hole Oceanographic Institution is supported by NSF grant OCE 20/25231. This research was sponsored by NSF grant OCE-8018665 (E.A.B.) and OCE 80-09015 (L.D.K.). Woods Hole Oceanographic Institution contribution 5100.

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Beryllium-10 in Australasian Tektites: Evidence for a Sedimentary Precursor

Abstract. *Each of seven Australasian tektites contains about 1×10^8 atoms of beryllium-10 (half-life, 1.53×10^6 years) per gram. Cosmic-ray bombardment of the australites cannot have produced the measured amounts of beryllium-10 either at the earth's surface or in space. The beryllium-10 contents of these australites are consistent with a sedimentary precursor that adsorbed from precipitation beryllium-10 produced in the atmosphere. The sediments must have spent several thousand years at the earth's surface within a few million years of the tektite-producing event.*

Accumulating geochemical and isotopic data have buttressed a terrestrial origin for tektites, probably in an impact event (*1*). Meanwhile, the search for cosmic-ray effects in tektites has been abandoned because of several early failures to detect such effects (*2*). Yet tektites could have acquired ^{10}Be, a long-lived product of cosmic-ray interactions, in any of several ways discussed below. Recent developments in accelerator mass spectrometry have lowered 1000-fold the detection limit for ^{10}Be [half-life, 1.53×10^6 years (*3*)]. We therefore decided to resume the search for ^{10}Be in tektites with the aim of constraining more narrowly hypotheses of tektite origin.

We concentrated on the youngest tektites, the 0.7×10^6-year-old australites, because they would have lost by decay the smallest fraction of any ^{10}Be present in their parent material at the time of formation. The 15×10^6-year-old moldavites, in contrast, would retain only about 10^{-3} of the original ^{10}Be. We analyzed a moldavite as one kind of blank. In addition, we measured the ^{10}Be in the two "bediasites" of uncertain provenance. One contained an appreciable amount of ^{10}Be, whereas the other did not. We have excluded these results from Table 1 because of the ease with which nominal bediasites can be confused with australites. Moniot *et al.* (*4*) describe the procedures used to measure ^{10}Be by accelerator mass spectrometry with the Rutgers tandem Van de Graaff and total Be by atomic absorption spectrometry.

The ^{10}Be contents of the australites (Table 1) are 2 to 20 times lower than those of most contemporary sediments, which range from 2×10^8 to 100×10^8 atom g^{-1} (*5–7*). Sediments from Mono Lake have ^{10}Be contents similar to those of the australites (*8*). The moldavite samples gave counting rates indistinguishable from those of the blanks. Various possible sources for the ^{10}Be present in australites are considered below; we favor a source in terrestrial, sedimentary material.

If a meteorite produced the australites, they could have recaptured some fraction of the projectile's ^{10}Be. A meteorite large enough to supply 1×10^8 atoms of ^{10}Be per gram to 10^8 tons of australites (*9*) would present a virtual 2π geometry to irradiation by cosmic rays in space. The surface production rate of ^{10}Be would be about 10 dpm kg^{-1} in stony material (*10*). The production rate would decrease toward the interior with a half thickness of about 40 cm. A projectile with a mass of at least 1×10^8 tons is required to supply this much ^{10}Be. For an iron meteorite the corresponding values are 3 dpm kg^{-1}, 10 cm, and 3×10^9 tons.

If the meteoritic component in tektites were indeed as large as 50 percent, its geochemical signature would be unmistakable; in fact, the trace element data indicate only a small meteoritic component in most samples (*11–13*). Moreover, a projectile of 10^8 tons would probably

Table 1. The Be and ^{10}Be contents of seven australites and one moldavite. Sources of samples: U samples, U.S. National Museum, R. Clarke; indochinite B, R. Fleischer; and SAM 433, V. Barnes.

Sample	Place of origin	Sample mass	Be (ppm)	^{10}Be (10^8 atom g^{-1})	Pre-treatment
		Australites			
U2172	Thailand	10.08 g	2.0 ± 0.2	0.79 ± 0.12	
U2349	Thailand	4.69 g	1.7 ± 0.2	0.74 ± 0.12	
U5424	Thailand*	10.06 g	2.2 ± 0.2	0.97 ± 0.19	
U1953	Santa Mesa, Philippines	8.02 g	2.2 ± 0.2	0.92 ± 0.21	
U1916	Luzon, Philippines	2.22 g	2.6 ± 0.2	1.28 ± 0.15	
Indochinite B (*21*)		1.65 g	2.7 ± 0.6	1.35 ± 0.29	
SAM 433 (*21*)	Kalgoorlie, Australia	<3 mg†		<30(<0.05)‡	24 hours in 0.1*M* HCl
SAM 433		<1 mg†		<50(<0.03)‡	24 hours in 1*M* HCl
SAM 433		0.35 g	1.9 ± 1.1	1.35 ± 0.27	0.3 hours in 49 percent HF
SAM 433		1.45 g	2.7 ± 0.2	2.04 ± 0.27	Residue
		Moldavite			
U2235	Lhenice, Bohemia, Moldavia	2.23 g	1.8 ± 0.2	<0.02	Bulk after 0.5 hour in 0.1*M* HCl
U2235		3.41 g	1.9 ± 0.2	<0.02	Bulk after mechanical abrasion of surface

*Muong Nong type. †Calculated as the difference in mass of the sample before and after leach. ‡Values referred to the total mass of tektite.

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