

# UC Irvine

## UC Irvine Previously Published Works

### Title

Increased production of cosmogenic  $^{10}\text{Be}$  during the Last Glacial Maximum

### Permalink

<https://escholarship.org/uc/item/4vd276dn>

### Journal

Nature, 357(6379)

### ISSN

0028-0836

### Authors

Lao, Yong  
Anderson, RF  
Broecker, WS  
[et al.](#)

### Publication Date

1992-06-01

### DOI

10.1038/357576a0

### Copyright Information

This work is made available under the terms of a Creative Commons Attribution License, available at <https://creativecommons.org/licenses/by/4.0/>

Peer reviewed

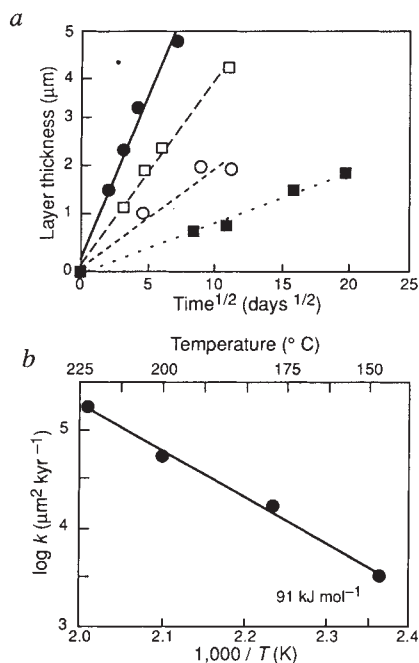


FIG. 3 a, Graph of measured hydration rim thickness against the square root of time at (■) 150, (○) 175, (□) 200 and (●) 225 °C for Indochinite tektite reacted in hydration experiments at relative humidity 100%. b, Arrhenius plot demonstrating the temperature dependence of the hydration rate constants for tektite hydration at 100% relative humidity. The activation energy is 91 kJ mol<sup>-1</sup>. Linear regression fits to these data give hydration rate constants of 0.096, 0.208, 0.390 and 0.697 μm day<sup>-1</sup>, respectively, where the correlation coefficients are greater than 0.98. The errors for the measurements of hydration rate constants fall within the plotted points.

rate for etching exceeds that for diffusion, etching is the dominant reaction process if the reaction conditions are suitable. Our results also suggest that, by examining the structure of reacted natural glass, it may be possible to infer details of the alteration environment. □

## Increased production of cosmogenic <sup>10</sup>Be during the Last Glacial Maximum

Yong Lao\*, R. F. Anderson\*, W. S. Broecker\*, S. E. Trumbore†‡, H. J. Hofmann† & W. Wolfli†

\* Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964, USA

† Institut für Mittelenergiephysik, ETH-Honggerberg, CH-8095 Zurich, Switzerland

**BERYLLIUM-10** (half-life 1.5 Myr) is produced by spallation of nitrogen and oxygen atoms by cosmic rays in the upper atmosphere. Its production rate is proportional to the flux of cosmic rays, which is modulated by solar activity and the strength of the Earth's magnetic field<sup>1,2</sup>. Weakening of the magnetic field allows more cosmic rays to impinge on the Earth's atmosphere, thereby increasing <sup>10</sup>Be production. Here we report that the ocean-wide average accumulation rate of <sup>10</sup>Be in Pacific sediments, which reflects the global average production rate of <sup>10</sup>Be (ref. 3), was at least 25% greater during the height of the most recent ice age (~24,000–16,000 yr ago) than during the Holocene (the past 10,000 yr). The higher production rate of <sup>10</sup>Be records the lower intensity of the geomagnetic field during that period and is consistent with the hypothesis developed to explain the younger <sup>14</sup>C ages of fossil corals compared with ages obtained by U–Th dating<sup>4</sup>. These results also point to a more general need to consider variations in production rate in geochronological studies using other cosmogenic nuclides.

The intensity of the Earth's magnetic field varies with time; for example, during the period from ~23,000 to ~15,000 yr BP (before present) its average value may have been ~50% lower than during the past 10,000 yr (ref. 5). Such variations greatly influence the production rates of cosmogenic nuclides such as <sup>14</sup>C and <sup>10</sup>Be. In a study of Barbados fossil corals, Bard *et al.*<sup>4</sup> found that the <sup>14</sup>C ages of the corals grown during the Last Glacial Maximum (LGM; ~20,000 yr BP) were ~3,500 yr younger than their U/Th ages. They attributed this discrepancy to a <sup>14</sup>C production rate ~40% higher during the LGM, caused by the lower intensity of the geomagnetic field. Palaeointensity data, however, are scarce, and methodological limitations can make them unreliable<sup>6</sup>. The postulated higher production rate of cosmogenic nuclides during lower geomagnetic field intensity must be verified through an independent record such as <sup>10</sup>Be.

The different geochemical cycles of carbon and beryllium require that different approaches be used to evaluate past production rates of <sup>14</sup>C and of <sup>10</sup>Be. After production, <sup>14</sup>C enters a large reservoir of atmospheric CO<sub>2</sub>, which is then well mixed, providing a homogeneous atmospheric <sup>14</sup>CO<sub>2</sub> source<sup>7</sup>. In contrast, <sup>10</sup>Be is a particle-reactive nuclide; it becomes associated with aerosols shortly after production and is stripped from the atmosphere on a timescale of months<sup>8</sup>. The deposition of <sup>10</sup>Be onto the Earth's surface varies considerably with time and place and is strongly influenced by local weather conditions<sup>2,8</sup>. Consequently, the deposition rate of <sup>10</sup>Be at any single site on land (for example in ice cores<sup>9,10</sup>) may not reliably record changes in its global average production rate. A different approach that integrates <sup>10</sup>Be deposition over a large geographic area must be used. Most of the <sup>10</sup>Be produced in the atmosphere enters the oceans, simply because the oceans cover two-thirds of the Earth's surface. The residence time of dissolved <sup>10</sup>Be in the oceans is long enough (~500–1,000 yr; refs 3, 11) for ocean mixing to smooth out latitudinal gradients in <sup>10</sup>Be deposition from the atmosphere, producing a nearly uniform concentration of <sup>10</sup>Be in the deep waters throughout an ocean basin such as in the

Received 3 July 1991; accepted 14 April 1992.

- Montanari, A. *et al. Geology* **11**, 668–671 (1983).
- Sigurdsson, H. *et al. Nature* **349**, 482–487 (1991).
- Margolis, S., Claeys, P. & Kyte, F. *Science* **251**, 1594–1597 (1991).
- Kastner, M., Asaro, F., Alvarez, W. & Alvarez, L. *J. non-cryst. Solids* **67**, 463–464 (1984).
- Izett, G., Dalrymple, G. & Snee, L. *Science* **252**, 1539–1542 (1991).
- Barkatt, A. A. *et al. Geochim. cosmochim. Acta* **48**, 361–371 (1984).
- O'Keefe, J. *Advances in Ceramics*, Vol. 20, *Nuclear Waste Management II*, 689–692 (American Ceramic Society Inc., Westerville, 1986).
- Glass, B. P. *J. non-cryst. Solids* **67**, 333–344 (1984).
- Lutze, W. & Ewing, R. (eds) *Radioactive Waste Forms of the Future* (North Holland, New York, 1988).
- Ewing, R. C. *Scientific Basis for Nuclear Waste Management I*, 57–68 (Plenum, New York, 1979).
- Berger, R. & Ericson, J. in *Recent Advances in Science and Technology of Materials* (ed. Bishay, A.) (Plenum, New York, 1974).
- LaMarche, P. H., Rauch, F. & Lanford, W. A. *J. non-cryst. Solids* **67**, 361–369 (1984).
- Friedman, I. & Long, W. D. *Science* **191**, 347–352 (1976).
- Mazer, J. J., Stevenson, C. M., Ebert, W. L. & Bates, J. K. *Am. Antiq.* **56**, 504–513 (1991).
- Abrajo, T. A., Bates, J. K., Woodland, A. B., Bradley, J. P. & Bourcier, W. L. *Clays Clay Miner.* **38**, 537–548 (1990).
- Advocat, T., Crovisier, J., Vernaz, E., Ehret, G. & Charpentier, H. *Mat. Res. Soc. Symp. Proc.* Vol. 212, 57–64 (Pittsburgh, 1991).
- Kring, D. A. & Boynton, V. V. *Geochim. cosmochim. Acta* **55**, 1737–1742 (1991).
- Jiang, X. & Peacor, D. *Clays Clay Miner.* **39**, 1–13 (1991).
- Brindley, G. W. & Brown, G. in *Crystal Structures of Clay Minerals and their X-Ray Identification* (Mineralogical Society, London, 1980).
- Bornhold, B. D. & Giresse, P. *J. sedim. Petrol.* **55**, 653–664 (1985).
- Montanari, A. *J. sedim. Petrol.* **61**, 315–339 (1991).
- Bursill, L. A. *J. Solid State Chem.* **10**, 72–94 (1974).
- Banfield, J. F., Veblen, D. R. & Smith, D. *J. Am. Miner.* **76**, 343–353 (1991).
- Bates, J. K., Bradley, J. P. & Bourcier, W. L. *High-Level Radioactive Waste Management: Proc. 2nd ann. int. Conf.*, Vol. 1, 720–727 (American Nuclear Society, La Grange Park, 1991).
- Ebert, W. & Bates, J. K. *Mat. Res. Soc. Symp. Proc.* Vol. 176, 339–346 (Pittsburgh, 1990).
- Friedman, I. & Smith, R. *Am. Antiq.* **25**, 476–522 (1960).
- Marshall, R. R. *Geol. Soc. Am. Bull.* **72**, 1493–1520 (1961).
- Lee, R. R., Leich, D. A., Tombrello, T. A., Ericson, J. E. & Friedman, I. *Nature* **250**, 44–47 (1974).

ACKNOWLEDGEMENTS. We thank D. Veblen, G. Ulmer and F. F. Rietmeijer for reviews. This work was supported by the U.S. Department of Energy, Office of Environmental Restoration and Waste Management.

‡ Present address: Department of Geosciences, University of California at Irvine, California 92717, USA.

TABLE 1 Average ratios of  $[Pa]_{G/H}$ ,  $[Be]_{G/H}$  and  $[Al]_{G/H}$  for cores in the Pacific Ocean

Core	Latitude	Longitude	Water depth (m)	Sedimentation rate (cm kyr <sup>-1</sup> )	Holocene* sample age (kyr BP)	Glacial* sample age (kyr BP)	$[Pa]_{G/H}$	$[Be]_{G/H}$	$[Al]_{G/H}$
RC14-121	54° 51' N	170° 41' W	2,532	20.0	1.2-2.1	23.3-28.3	0.38 ± 0.20	1.07 ± 0.39	2.18
V20-122	46° 34' N	161° 41' E	5,563	4.5	2.0-6.0	17.9-23.9	0.95 ± 0.10	1.20 ± 0.10	2.33
RC14-105	39° 41' N	157° 33' E	5,630	6.0	1.8-5.3	18.8-24.8	1.13 ± 0.11	1.67 ± 0.11	1.62
V21-146	37° 41' N	163° 02' E	3,968	3.8	2.8-6.4	19.0-24.0	1.05 ± 0.09	1.64 ± 0.11	1.74
V32-126	35° 19' N	174° 54' E	3,870	2.3	3.3-7.3	17.6-23.8	0.99 ± 0.08	1.43 ± 0.08	1.40
V32-128	36° 27' N	177° 09' E	3,623	1.7	5.2-7.3	17.2-23.9	1.20 ± 0.12	1.95 ± 0.11	2.27
W8709A-1	41° 33' N	131° 57' W	3,680	1.3	5.0-5.6	18.9-21.0	0.64 ± 0.05	1.40 ± 0.07	1.36
W8709A-8	42° 16' N	127° 41' W	3,111	10.0	1.0-4.1	17.4-19.4	0.72 ± 0.09	0.89 ± 0.10	1.44
W8709A-13	42° 07' N	125° 45' W	2,712	20.0	4.7-5.4	19.8-21.6	0.63 ± 0.20	0.81 ± 0.24	1.16
V21-59	20° 55' N	158° 06' W	2,992	0.8	7.1-9.6	14.8-24.3	1.04 ± 0.07	1.33 ± 0.06	1.47
V28-238	01° 01' N	160° 29' E	3,120	1.7	4.8-8.5	21.0-25.3	0.98 ± 0.07	1.50 ± 0.09	ND
RC11-210	01° 49' N	140° 03' W	4,420	1.2	6.0	16.7-19.6	1.02 ± 0.06	1.06 ± 0.06	1.06
V19-28	02° 22' S	84° 39' W	2,720	5.0	5.3	20.6-24.5	0.69 ± 0.06	1.39 ± 0.10	1.13
V19-29	03° 35' S	83° 56' W	3,157	10.0	1.8-4.5	19.8	0.82 ± 0.08	1.18 ± 0.10	1.10
TT154-10	10° 17' S	111° 20' W	3,225	2.0	5.9-6.8	20.1-22.6	0.69 ± 0.04	1.13 ± 0.06	1.47
V19-55	17° 00' S	114° 11' W	3,177	1.2	5.1-7.7	18.8-23.3	0.86 ± 0.05	1.08 ± 0.05	1.48
E17-9	63° 05' S	135° 07' W	4,848	3.6	2.4	18.0	0.65 ± 0.04	0.83 ± 0.05	1.78
RC15-61	40° 37' S	77° 12' W	3,771	3.8	2.0-7.1	18.2-25.4	1.12 ± 0.11	1.02 ± 0.07	1.42
Average†							0.91 ± 0.05	1.25 ± 0.07	1.52

Beryllium-10 was analysed by accelerator mass spectrometer, <sup>230</sup>Th and <sup>231</sup>Pa by radioactivity measurement, and Al by direct source plasma measurement.  $[Pa]_{G/H}$ ,  $[Be]_{G/H}$  and  $[Al]_{G/H}$  are the average Pa/Th, Be/Th and Al/Th ratios in the glacial sediments divided by the average Pa/Th, Be/Th and Al/Th ratios in the Holocene sediments, respectively, using equation (2) in the text. The quoted errors for  $[Pa]_{G/H}$  and  $[Be]_{G/H}$  include all sources of errors propagated throughout the calculation. The approximate uncertainty for analysis of Al in individual samples was ~5%. See ref. 21 for details of methods and sedimentation rates. ND, not determined.

\* Two samples from each time period were analysed for <sup>10</sup>Be, <sup>230</sup>Th and <sup>231</sup>Pa nuclides and Al contents for most of the cores. The ages for each core bracket the time interval during which the sediments were deposited. For some cores, only one sample was taken, so only one age is given. The age information is from literature based on <sup>18</sup>O, <sup>13</sup>C and <sup>14</sup>C isotope stratigraphy.

† The uncertainties for the averages of  $[Pa]_{G/H}$  and  $[Be]_{G/H}$  ratios are the standard error of the mean: for  $[Be]_{G/H}$ ,  $n=18$ ; for  $[Pa]_{G/H}$ ,  $n=17$ , excluding RC14-121 because the composition of the glacial sediment is very different from that in the Holocene sediment<sup>27</sup>. For the average of  $[Al]_{G/H}$  ratios,  $n=17$  because no Al was measured for V28-238.

Pacific<sup>12</sup>. If the Pacific Ocean serves as a large homogeneous reservoir of cosmogenic <sup>10</sup>Be, just as the atmosphere does for <sup>14</sup>C, then a record of the concentration of <sup>10</sup>Be in the deep Pacific over the period of interest should indicate the extent to which the global average production rate of <sup>10</sup>Be has varied.

No reliable monitor of the concentration of <sup>10</sup>Be in sea water has been identified, so the supply of <sup>10</sup>Be to the ocean must be evaluated from its rate of accumulation in deep-sea sediments. Evaluating the average accumulation rate of <sup>10</sup>Be in marine sediments is, however, complicated by the fact that the intensity of the scavenging processes that remove <sup>10</sup>Be from the water column to the sediments varies spatially, leading to a heterogeneous pattern of <sup>10</sup>Be accumulation. For example, fluxes of <sup>10</sup>Be to sediments near ocean margins, where relatively high biological productivity combined with the supply of eroded continental material greatly enhances scavenging, are more than an order of magnitude greater than fluxes of <sup>10</sup>Be to pelagic sediments underlying oligotrophic mid-ocean gyres<sup>13-16</sup>. Given this variability, and the possibility that the regional patterns of scavenging intensity have changed over time, it seems unlikely that a 40% change in <sup>10</sup>Be production rate (as is suggested by <sup>14</sup>C production rate) can be reliably identified in a marine record of <sup>10</sup>Be at a single site, even though the spatial concentration of <sup>10</sup>Be in the deep sea is relatively uniform. Two steps are needed to detect a change of this magnitude. First, the <sup>10</sup>Be record must be examined at many sites representing a variety of marine environments. Second, to constrain accumulation rates of <sup>10</sup>Be more reliably, they should be normalized to a tracer whose source is accurately known.

Naturally occurring uranium-series radionuclides provide the needed tracers. Thorium-230 (half-life 75,200 yr) and <sup>231</sup>Pa (half-life 32,500 yr) are produced in the ocean by decay of dissolved <sup>234</sup>U and <sup>235</sup>U, respectively. Thorium-230 is an extremely particle-reactive nuclide, such that it is removed from sea water on a timescale (decades<sup>17</sup>) much less than its radioactive half-life. Consequently, the flux of <sup>230</sup>Th to the sediments accumulated

at the time of interest is equal to its well known production rate. Normalizing <sup>231</sup>Pa and <sup>10</sup>Be to <sup>230</sup>Th therefore enables us to evaluate the accumulation rates of <sup>231</sup>Pa and <sup>10</sup>Be in the ocean. Evaluating such fluxes by normalizing to <sup>230</sup>Th has been discussed extensively<sup>16,18-20</sup>. Briefly, the flux ( $F$ ) of a sedimentary component  $N$  is estimated from the ratio of the concentration of  $N$ ,  $[N]$ , to that of initial unsupported <sup>230</sup>Th (that produced by decay of <sup>234</sup>U in the water column),  $[^{230}\text{Th}]$ , in sediment as

$$F(N) = P_{\text{Th}}^* [N] / [^{230}\text{Th}] \quad (1)$$

where  $P_{\text{Th}}$ , the production rate of <sup>230</sup>Th, is directly proportional to water depth. We selected 18 independently dated deep-sea cores<sup>21</sup> (Table 1) to which we applied this technique.

Like <sup>10</sup>Be, <sup>231</sup>Pa is preferentially removed from sea water in some ocean-margin regions of high particle flux even though the production rate of <sup>231</sup>Pa is uniform throughout the ocean<sup>16</sup>. An extensive study of Holocene sediments throughout the Pacific has shown that the patterns of enhanced <sup>231</sup>Pa deposition are much like those of enhanced <sup>10</sup>Be deposition (relative to <sup>230</sup>Th) at ocean-margins<sup>13</sup>. The <sup>231</sup>Pa to <sup>230</sup>Th production ratio has not changed for the last glacial cycle because of the fixed uranium isotopic composition and constant uranium concentration in the ocean due to the very long residence time of uranium (~200,000 yr; ref. 22). Therefore, the average of <sup>231</sup>Pa/<sup>230</sup>Th ratios of the glacial sediments in a representative set of cores should be the same as that of the Holocene sediments. We test whether the selection of the 18 cores provides an unbiased representation of the average oceanic flux of <sup>10</sup>Be using <sup>231</sup>Pa/<sup>230</sup>Th ratios in the sediments. For this purpose, we introduce a new parameter.

$$[N]_{G/H} = [N/\text{Th}]_G / [N/\text{Th}]_H \quad (2)$$

where  $N$  is the concentration of either initial unsupported <sup>231</sup>Pa (that produced by decay of <sup>235</sup>U in the water column), or <sup>10</sup>Be; Th refers to the concentration of initial unsupported <sup>230</sup>Th; G and H indicate the average results for samples from the glacial

(~24,000 to ~16,000 yr BP) and the Holocene (the past 10,000 yr), respectively. Because the production rate of  $^{230}\text{Th}$  ( $P_{\text{Th}}$ ) is constant in equation (1), the  $[N]_{\text{G/H}}$  ratio is indicative of a change in the flux of  $^{231}\text{Pa}$  or  $^{10}\text{Be}$  from the glacial to the Holocene.

At any individual site, the deviation of  $[\text{Pa}]_{\text{G/H}}$  from 1.0 could reflect changes in patterns of scavenging intensity which may result from changes in the pattern of particle flux or in sediment particle composition. But the ocean-wide average  $[\text{Pa}]_{\text{G/H}}$  ratio must, as noted above, be equal to 1.0. The average of the  $[\text{Pa}]_{\text{G/H}}$  ratios is  $0.91 \pm 0.05$  (Table 1), close to 1.0, indicating that the cores selected here provide a reasonable representation of the average sediment in the Pacific (with respect to deposition of the tracer nuclides). In contrast to  $[\text{Pa}]_{\text{G/H}}$ , the average  $[\text{Be}]_{\text{G/H}}$  ratio is  $1.25 \pm 0.07$  (Table 1), suggesting that the spatially averaged flux of  $^{10}\text{Be}$  to the deep Pacific sediments during the glacial was ~25% greater than during the Holocene. For reasons discussed below, this value probably represents a lower limit for an increased global average production rate of  $^{10}\text{Be}$  during the glacial.

Increased erosion of continents cannot have supplied the extra  $^{10}\text{Be}$ . Flux of  $^{10}\text{Be}$  associated with continental dust has been estimated<sup>21</sup> to be ~5% of the total supply of  $^{10}\text{Be}$  to the deep Pacific during the Holocene, assuming the  $^{10}\text{Be}$  content of Chinese loess<sup>23</sup> to be representative of that of continental dust deposited over the Pacific. We measured the aluminum content of the sediments and calculated from this the fluxes of aluminosilicate dust to the Pacific during the glacial relative to the Holocene, using equation (2). The  $[\text{Al}]_{\text{G/H}}$  ratios indicate that the average flux of continental material to the deep Pacific was ~50% larger during the glacial (Table 1). If the  $^{10}\text{Be}$  content of the continental dust did not vary, then this represents a negligible (~2%) increase of the continent-derived  $^{10}\text{Be}$  accumulated in the deep Pacific during the glacial compared with that during the Holocene.

Virtually all of the  $^{10}\text{Be}$  carried by rivers is removed in estuaries<sup>10</sup>, so changes in the riverine flux of  $^{10}\text{Be}$  cannot have influenced  $^{10}\text{Be}$  accumulation in deep Pacific sediments. Beryllium-10 supplied to the oceans with meltwaters during the period of glacial retreat (from ~14,000 to ~8,000 yr ago; ref. 24) may have contributed ~7% of the total supply of  $^{10}\text{Be}$  to the world oceans, if we assume the  $^{10}\text{Be}$  concentration in ice sheets on the North American and European continents to be the same as that in today's Greenland ice<sup>25</sup>. Most of meltwater  $^{10}\text{Be}$  would have been supplied to the Atlantic (through confined drainage basin systems) and removed in the estuaries there. Changes in the pattern of ocean circulation are unlikely to have significantly increased the partitioning of  $^{10}\text{Be}$  from other oceans into the Pacific<sup>13</sup>. Therefore, the deep Pacific acts as a relatively closed basin with respect to  $^{10}\text{Be}$  after its deposition from the atmosphere. We have thus estimated the global average production rate of  $^{10}\text{Be}$  during the Holocene ( $\sim 1.5 \times 10^6$  atoms  $\text{cm}^{-2}$   $\text{yr}^{-1}$ ) from the average accumulation rate of  $^{10}\text{Be}$  in the Holocene sediments in the Pacific<sup>13</sup>.

Because bioturbation tends to smooth out signal differences, the average  $[\text{Be}]_{\text{G/H}}$  ratio of 1.25 may be a lower limit because sediment accumulation rates in some of the open-ocean cores are sufficiently low that some component of glacial sediments continues to be retained in the surface mixed layer. Furthermore, as the average  $[\text{Pa}]_{\text{G/H}}$  ratio is slightly less than 1.0 (Table 1), the ocean-wide average accumulation rate of  $^{231}\text{Pa}$  during the glacial may be slightly underestimated using the cores in this study. The average accumulation rate of  $^{10}\text{Be}$  during the glacial may similarly be underestimated because the pattern of scavenging of  $^{10}\text{Be}$  is similar to that of  $^{231}\text{Pa}$ .

In summary, our measured average  $[\text{Be}]_{\text{G/H}}$  ratio of 1.25 is a lower limit for the change in the average accumulation rate of  $^{10}\text{Be}$  in the Pacific Ocean; supply of  $^{10}\text{Be}$  to the Pacific (therefore the production rate of  $^{10}\text{Be}$ ) must have been at least 25% larger during the glacial than during the Holocene. The

result of this study is clearly consistent with the finding that  $^{14}\text{C}$  production rate may be ~40% higher during the LGM<sup>4</sup>, and supports the hypothesis relating the increased production of  $^{14}\text{C}$  to a lower geomagnetic field strength. Implications for  $^{14}\text{C}$  dating of climate records, archaeological artefacts and so on are obvious. Exposure age dating of surface rocks and landforms based on *in situ* production rates of cosmogenic nuclides such as  $^{10}\text{Be}$ ,  $^{26}\text{Al}$  and  $^{36}\text{Cl}$  (ref. 26) will be affected similarly by fluctuations of the geomagnetic field intensity. Variability of the geomagnetic field intensity further back in time, to ~300,000 yr BP, can be examined using the approach described here, ultimately limited by the precision with which unsupported  $^{230}\text{Th}$  can be measured in marine sediments. □

Received 9 March; accepted 1 May 1992.

- Lal, D. & Peters, B. *Encyclopedia of Physics* 46/2, 551–612 (Springer, New York, 1967).
- Lal, D. *Geophys. Res. Lett.* **14**, 785–788 (1987).
- Lao, Y. *et al. Earth planet Sci. Lett.* (in the press).
- Bard, E., Hamelin, B., Fairbanks, R. G. & Zindler, A. *Nature* **345**, 405–410 (1990).
- Mazaud, A., Laj, C., Bard, E., Arnold, M. & Tric, E. *Geophys. Res. Lett.* **18**, 1885–1888 (1991).
- Banerjee, S. K. *From Mantle to Meteorites: a Gsrland of Perspective* (eds Gopalan K. *et al.*, 229–238 (Indian Academy of Sciences, Bangalore, 1989).
- Libby, W. F. *Radiocarbon Dating* (University of Chicago Press, 1952).
- Brown, L., Stenslend, G. J., Klein, J. & Middleton, R. *Geochim. cosmochim. Acta* **53**, 135–142 (1989).
- Yiou, F., Raisbeck, G. M., Bourles, D., Lorius, C. & Barkov, N. I. *Nature* **316**, 616–617 (1985).
- Raisbeck, G. M. *et al. Nature* **326**, 273–277 (1987).
- Kusakabe, M. *et al. Earth planet. Sci. Lett.* **102**, 265–276 (1991).
- Kusakabe, M. *et al. Earth planet. Sci. Lett.* **82**, 231–240 (1987).
- Mangini, A. *et al. Nucl. Instr. Meth. Phys. Res. B5*, 353–358 (1984).
- Brown, L., Klein, J. & Middleton, R. *Geochim. cosmochim. Acta* **49**, 153–157 (1985).
- Eisenhauer, A. *et al. Geological History of the Polar Oceans: Arctic versus Antarctic* (eds Bliel, U. & Thiede, J.) 475–487 (Kluwer, Dordrecht, 1990).
- Anderson, R. F. *et al. Earth planet. Sci. Lett.* **96**, 287–304 (1990).
- Anderson, R. F., Bacon, M. P. & Brewer, P. G. *Earth planet. Sci. Lett.* **62**, 7–23 (1983); **66**, 73–90 (1983).
- Bacon, M. P. & Rosholt, J. N. *Geochim. cosmochim. Acta* **46**, 651–666 (1982).
- Suman, D. O. & Bacon, M. P. *Deep-Sea Res.* **36**, 869–878 (1989).
- Francois, R., Bacon, M. P. & Suman, D. O. *Paleoceanography* **5**, 861–877 (1990).
- Lao, Y. thesis, Columbia Univ. (1991).
- Cochran, J. K. *Uranium Series Disequilibrium: Applications to Environmental Problems* (eds Ivanovich, M. & Harmon, R. S.) 384–430 (Clarendon, Oxford, 1982).
- Shen, C. thesis, Univ. of Bern (1986).
- Fairbanks, R. G. *Nature* **342**, 637–642.
- Beer, J. *et al. Nature* **331**, 675–679 (1988).
- Lal, D. A. *Rev. Earth planet. Sci.* **16**, 355–388 (1988).
- Sancetta, C., Heusser, L., Labeyrie, L., Naidu, A. S. & Robinson, S. W. *Mar. Geol.* **62**, 55–68 (1985).

ACKNOWLEDGEMENTS. We thank R. G. Fairbanks for reviewing an early draft of this manuscript. This research was supported by the US NSF.

## Enhanced Cenozoic chemical weathering and the subduction of pelagic carbonate

Ken Caldeira

Earth System Science Center and Department of Geosciences, Pennsylvania State University, 248 Deike Building, University Park, Pennsylvania 16802, USA

THE observed trend of increasing oceanic  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios during the late Cenozoic led Raymo *et al.*<sup>1</sup> to propose that chemical weathering rates increased at this time as a result of enhanced weatherability of silicate rocks. They suggested that this was due in turn to continental uplift, primarily in the Himalayas and the Andes. Because weathering involves the reaction of silicates with atmospheric carbon dioxide, considerations of changes in weathering rates must take into account the need to balance the global carbon cycle. To maintain this balance on timescales greater than  $10^6$  yr, enhanced weathering requires an increased flux of  $\text{CO}_2$  into the atmosphere<sup>2</sup>. Without such an increased flux, weathering rates could not increase, and one is then forced to search for other explanations for the observed  $^{87}\text{Sr}/^{86}\text{Sr}$  trend, such as a changing riverine strontium isotope composition<sup>3–6</sup>. Here I assume that the strontium isotope record does indeed reflect enhanced weathering in the late Cenozoic, but propose that its cause may have been an