

^{10}Be distribution in sediments of two cores from the north Atlantic: An approach for evaluating environmental change

A. A. ALDAHAN, G. POSSNERT and E. OLAUSSON

A.A. AlDahan, G. Possnert and E. Olausson. 1992 04 15: ^{10}Be distribution in sediments of two cores from the north Atlantic: An approach for evaluating environmental change. *Bulletin of the Geological Institutions of The University of Uppsala*, N.S. Vol 16, pp. 69–80. Uppsala. ISSN 0302-2749.

Distribution of the cosmic-ray produced isotope ^{10}Be was measured in two sediment cores from the north Atlantic ocean. Dissolution with 8M HCl indicates ^{10}Be concentrations of $2.7\text{--}5.7 \times 10^8$ and $4.1\text{--}8.0 \times 10^8$ at/g in sediment of cores 290 and 299 respectively. Dissolution by $\text{HF-H}_2\text{SO}_4$ reveals that about up to 40 % of the ^{10}Be content is retained in the HCl-insoluble fraction. Variations in ^{10}Be content of the two cores are compared to age estimates made by ^{14}C -dating, $\delta^{18}\text{O}$ (in *Neogloboquadrina dutertrei*, *Globorotalia inflata* and *Cibicides wuellerstorfi*), other microfossils and carbonate content. Sediments of core 299 extends to about 150 ka at a depth of 5.5 m thus giving an accumulation rate of about 4 cm/ka, whereas a sediment accumulation rate of about 1 cm/ka emerges from the data of core 290 (about 600 ka at a depth of 6.17 m). Although not fully resolved by the number of measurements made in this study, there appear to be some enhanced ^{10}Be values which correspond to about 100–130 ka repeated intervals. These enhanced values may relate to orbital forcing periods and thus could serve as stratigraphic correlation markers of marine sediments in a way similar to signals retained by the $\delta^{18}\text{O}$.

A.A. AlDahan, *The Svedberg Laboratory and Department of Quaternary Geology, Uppsala University, S-75121 Uppsala, Sweden*; G. Possnert, *The Svedberg Laboratory, Uppsala University, S-75121 Uppsala, Sweden*; E. Olausson, *Hultvågen 9, S-44006, Gråbo, Sweden; 4th February, 1992.*

Introduction

During recent decades extensive efforts have been launched into investigating natural variation in marine sediments with the help of oxygen, carbon and beryllium isotopes, among others (e.g. Emiliani, 1955, 1978; Shackleton & Opdyke, 1973; Tanaka & Inoue, 1980; Mix & Fairbanks, 1985; Berger, 1989; Bourles et al., 1989; Zahn et al., 1991). One important outcome was the possibility of utilizing changes in the oxygen isotopic ratio as indicator of both age and glaciation periods. Furthermore, correlation between the oxygen and carbon isotopic data proved to be a successful approach in detecting environmental variations (e.g. Labeyrie & Duplessy, 1985; Jansen, 1989). The correlation between radionuclides produced by cosmic-rays, such as ^{10}Be , and $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ provided additional assessment of chronological events in marine sediments (e.g. Ku et al., 1985; Southon et al., 1987; Eisenhauer et al., 1987; Henken-Mellies et al., 1990). The ^{10}Be isotope, with its half-life of 1.5 Ma, has several features which make it useful for dating of marine sediments, and in particular

when other dating procedure cannot be achieved properly. However, the data on ^{10}Be in marine sediments indicate non exponential decrease with time, particularly during the last 1 Ma, and accordingly efforts are concentrated on obtaining a ^{10}Be time scale with time markers in a way similar to the $\delta^{18}\text{O}$ time scale. In most such studies (e.g. Southon et al., 1987; Henken-Mellies et al., 1990) it was difficult to substantiate a model that satisfies a systematic time-dependent variation in ^{10}Be , because the available data apparently support a lithological contribution to the variation observed.

In the study, we present data of two sediment cores which illustrate some of the problems addressed with respect to the ^{10}Be dating. We have compared the results of ^{10}Be with dating by oxygen isotopes, ^{14}C , microfossils and calcite content. Furthermore microscopic (light and scanning electron microscopy) and X-ray diffraction examinations were carried out in order to evaluate the effects of sediment composition and secondary (diagenetic) environmental changes on the dating parameters used.

Location and description of the cores

The sediments used in this study are the about 6.6 m and 5.5 m long piston cores referred to as cores 290 and 299 respectively. Core 290 was collected from a water depth of 1995 m at latitude 41° 35'N and longitude 29°37'W and core 299 from a water depth of 4288 m at latitude 47°41'N and longitude 12°43'W by the Albatross Swedish Deep Sea Expedition during 1947–48 (Fig. 1). The foraminifera in core 290 were studied by Phleger et al. (1953), the lithology and chemistry of both cores by Olausson (1960) who also proposed a tentative stratigraphy (Olausson, 1971). The color of the sediments varies between whitish-grey and pale-yellowish without a distinct banding or lamination. Lithification increases downwards and is accompanied by a formation of friable chalky character. The grain size is generally ≤ 0.5 mm in both cores with observable variation in the calcite content (Fig. 2). Micro- and nanno-fossil shells dominate the materials of core 290 with contents generally exceeding 80 vol. %. The planktonic population comprises ≥ 90 vol. % of the microfossils and the rest is benthic (Phleger et al., 1953). In core 299, most of the calcite occurs as micritic matrix and the content of microfossils is below 50 vol. %. Quartz, feldspar and clay minerals are the most common minerals besides calcite, whereas minor amounts of Fe-oxides, Mn-oxides and Ti-oxides also occur.

X-ray diffraction analysis of total samples and separated $\geq 63 \mu\text{m}$, $\leq 20 \mu\text{m}$ and $\leq 4 \mu\text{m}$ fractions (made by wet sieving and sedimentation) indicate that calcite is the major mineral with some variable proportions of quartz, feldspar, illite, chlorite and smectite in core 290. The proportion of calcite varies in core 299 from 75 to 20 vol. % with occurrence of minor dolomite, gypsum and halite.

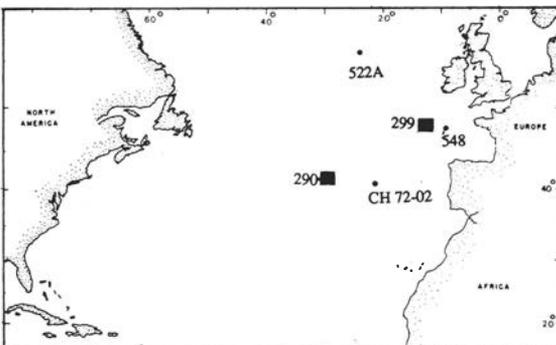


Fig. 1. Location map of the cores studied (black rectangles) and sites 522A, 548 and CH72-02.

Dissolution of some samples with 5–10 % acetic acid left insoluble residue which is mainly composed of quartz, feldspar, clay minerals aggregates, siliceous fossil shells (radiolarians), organic matter remains and some subordinate amounts of Fe-, Ti- and Mn-oxides.

Transmitted light microscopic examination of thin sections revealed occurrence of coarse ($\sim 10 \mu\text{m}$ large), loose, micritic aggregates and intact fossil shells at the upper parts of the cores (Fig. 2A and B). The micritic matrix becomes more compacted with increasing depth (below about 3 m) and the aggregation feature disappears (Fig. 2C and D). Dissolution and breakage of some fossil shells as well as occurrence of authigenic clay minerals (mostly a chlorite-smectite phase) are observed at depths below about 5 m in core 290 (Fig. 2E) and in most of core 299 (Fig. 2F). Dissolution of the nanofossil shells is observed in the depth interval below 5 m of core 290, as revealed by the scanning electron microscope. Finecrystalline (up to $10 \mu\text{m}$ large) calcite, apparently resembling microspar, is found in samples showing dissolution of the fossil shells.

Analytical methods

Determination of ^{10}Be

The determination of ^{10}Be concentration was carried out on 14 samples from core 290 and 8 samples from core 299. The selection of these samples was based on choosing both closely (25–50 cm apart) and widely (~ 1 m apart) spaced intervals to account for short and long range variations in the activity of ^{10}Be (this topic is further elaborated in the discussion section). Furthermore, ^{10}Be was determined in microfossil-rich $\geq 63 \mu\text{m}$ fraction of four samples as well as the fractions $\leq 20 \mu\text{m}$ and $\leq 4 \mu\text{m}$ of one sample from core 290 and on four HCl-insoluble residues from core 299. A briefing of the chemical extraction procedure of ^{10}Be from the sediments is given below and more details are given in AlDahan & Possnert (1990).

To an ignited (at 900°C) amount of 0.15–0.5 g of each sample were added 0.25–1 mg Be carrier and 10 ml 8M HCl and the mixture was left (with continuous stirring) for 1–2 hours on a water bath at 95°C . Generally small amounts, not exceeding 8 %, of colorless residue remains after filtration of samples from core 290 whereas the percentage of insoluble residue varied from about 20–60 % in samples of core 299. Hydroxides were precipitated from the filtered solution by adding 25 % NH_3 and

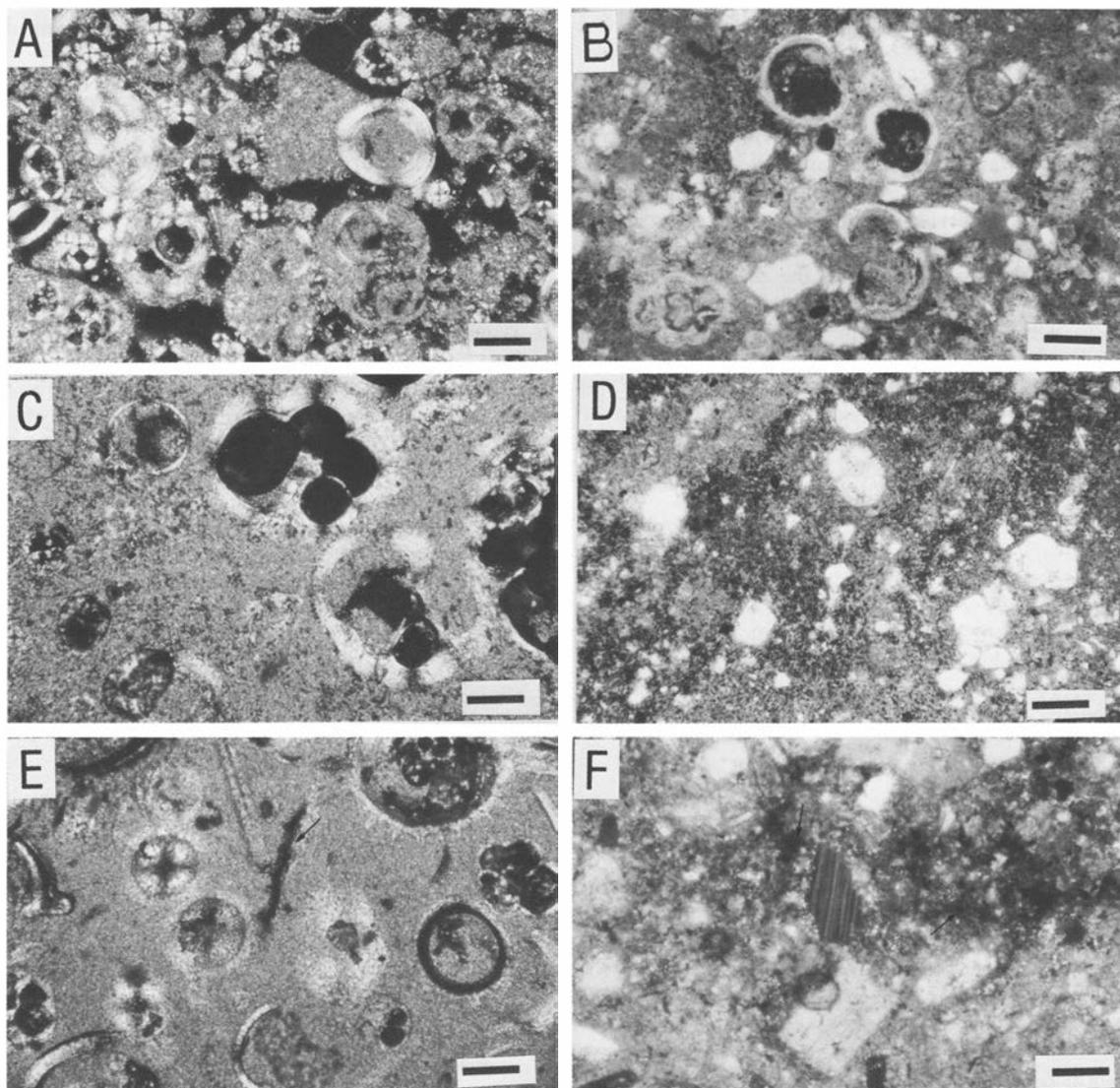


Fig. 2. Thin section of samples from the cores studied photographed under polarized light (bar scale=50 μ m) showing: A and B - The occurrence of loose micritic aggregates and intact fossil shells at the upper parts of the cores (A-core 290 and B-core 299). C and D - Increasing compaction at the middle parts of the cores which is associated with penetration of the matrix in the fossil (Fig. 2C, core 290). The high content of detritus is well illustrated in Fig. 2D, core 299. E and F - Dissolution of some fossil shells (Fig. 2E, core 290). Detrital grains (Fig. 2F, core 299) and occurrence of authigenic clay mineral mass (arrows) at the lower parts of the cores.

the mixture was again filtered. The filtrate was dissolved in 6 M HCl and the pH of the solution was adjusted to 3.5 with NaOH and EDTA-2Na. This solution was passed through an ion exchange column (packed with 10 ml of Dowex AG 50 w-X8 50-100 resin washed with EDTA-2Na based solution) after addition of 0.5 ml 30 % H₂O₂. Beryl-

lium was eluted from the column with 50 ml 4M HCl and separated as Be-hydroxide with 25 % NH₃. This Be-hydroxide was converted to BeO at 600°C for two hours and mixed with silver powder for the accelerator mass spectrometer analysis.

The analytical determinations of the ¹⁰Be content were performed with the Uppsala EN-tandem Van

de Graaff accelerator used as an ultra sensitive mass spectrometer. $^9\text{BeO}^-$ ion currents in the order 100–500 nA, produced in a SIMS Cesium sputter ion source, were accelerated to 18 MeV. The break-up of $^{17}\text{O}^{+5}$ beam from the injected $^{10}\text{Be}^{17}\text{O}^-$ molecule was charge-integrated in a side Faraday cup after the analysing magnet in order to normalize the output current from each sample. Simultaneously $^{10}\text{Be}^{+3}$ was measured in a detector telescope located at $+30^\circ$ after the switching magnet (Possnert & Åström, 1986). An absolute $^{10}\text{Be}/^9\text{Be}$ ratio was achieved by the use of standards with known concentrations; we used a standard from the National Institute of Standards and Technology and a standard prepared at our laboratory which cover the $^{10}\text{Be}/^9\text{Be}$ ratio at about $1-3 \times 10^{-11}$. Most isobaric $^{10}\text{B}^{+3}$ ions were eliminated by a 10 μm thick Cu-foil acting as an entrance window for the $^{10}\text{Be}^{+3}$ beam to the detector telescope. A background level in the order of 5×10^{-14} was obtained for a blank sample prepared from a beryllium metal. This background is far below the $^{10}\text{Be}/^9\text{Be}$ ratio of the samples which was around 10^{-11} . Errors in the ^{10}Be data (including machine stability and accuracy) are generally below 8 %.

Determination of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$

Because of rarity and/or extensive dissolution and fracturing of fossils in parts of core 299, the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were performed only on samples from core 290. For this purpose two planktonic (*Neogloboquadrina dutertrei* and *Globorotalia inflata*) and one benthic (*Cibicides wuellerstorfi*) foraminifers were picked for determination of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. These fossil shells were separated from samples spaced about 10 cm apart, thus covering the whole of core 290 with 65 samples. The preparation and analytical procedures using a VG 903 Micro-mass mass spectrometer are described by Shackleton (1974). The analytical error in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ determinations is ± 0.1 ‰.

Determination of ^{14}C

Carbon 14 measurements were made on 3 bulk samples from core 299 and 3 > 63 μm fossil-rich samples from core 290. The samples were ultrasonically cleaned and leached with 0.5 HCl at several stages, of which gas from the first leaching was disregarded. The rest of the released CO_2 was reduced to graphite by iron powder loaded in 0.25 mm in diameter Cu-holder for the accelerator mass spectrometer. The Uppsala EN-tandem Van de

Graaff accelerator was used for the measurement of ^{14}C as described by Possnert (1990).

Results and interpretations

The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data

The choice of the three types of microfossils used for the determination of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ was aimed to detect changes in near surface water masses as well as the bottom water mass. The planktonic species *Neogloboquadrina dutertrei* and *Globorotalia inflata* were selected to represent variations in the ocean water layers classed as intermediate and deep by Bé (1977). The benthic species *Cibicides wuellerstorfi* is one of the most frequently analysed benthic fossils in marine sediments and its data can be used for regional correlation. These three fossils show deviation in their isotopic equilibrium with sea water. The magnitude of deviation for *C. wueller-*

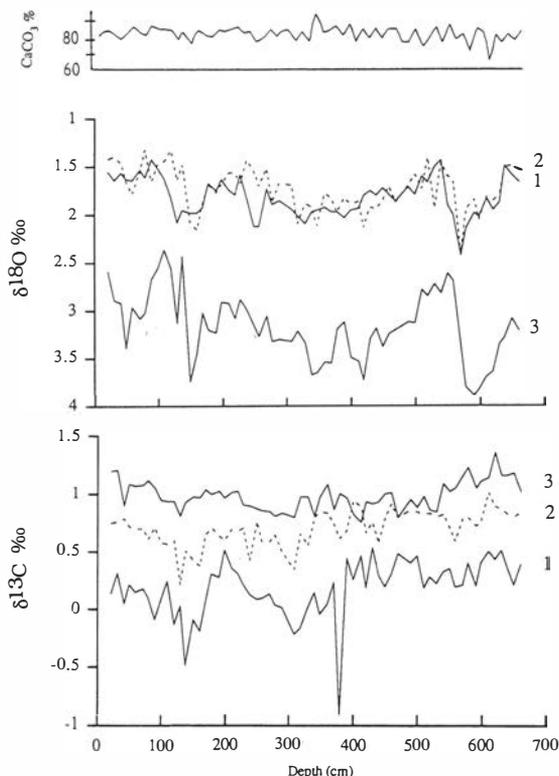


Fig. 3. Variation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ with Depth in the separated fossils from core 290. *Neogloboquadrina dutertrei* (1), *Globorotalia inflata* (2) and *Cibicides wuellerstorfi* (3). Numbers above $\delta^{18}\text{O}$ peaks refer to oxygen isotopic stages. The calcite content is from Olausson (1960).

torfi is -1.01‰ (PDB) $\delta^{18}\text{O}$ and -1.04‰ (PDB) $\delta^{13}\text{C}$ (Graham et al., 1981). The disequilibrium offset for the planktonic species is affected by several factors, including, e. g. seasonal temperature variation, location within the water column, species difference and maturation habit (e. g. Jansen, 1989). In this study, however, we have not carried out any disequilibrium correction for the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data of the fossils used. The results of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ presented in this paper are given in PDB notation.

Although the purpose of oxygen and carbon isotopes was to provide an independent chronology for core 290, some remarks are also made with respect to the scope of variations in the data. The $\delta^{18}\text{O}$ data shown in Table 1 and Figure 3 indicate a range of variation in the planktonic species between 1.3 and 2.4‰. Slight enrichment with values ranging between 2.4 and 3.9 is found in the benthic species. This enrichment on the benthic species reflects mainly the lower temperature of the bottom compared to the surface ocean water (e. g. Emiliani, 1971; Erez & Honjo, 1981).

In all $\delta^{18}\text{O}$ trends (Fig. 3 and in particular Fig. 4) there is a slight but rather consistent enrichment in the heavier isotope with increasing depth in the core. The estimate from the fit of the linear regression line (Fig. 4) indicates a value of about 0.4‰ increase at sample depth of 6.17 m. This feature is interesting because similar distribution of $\delta^{18}\text{O}$ in benthic foraminifera can be observed in other published records (Shackleton & Opdyke, 1973; Prell 1982; Zimmerman et al., 1985; Jansen et al., 1988) down to isotopic stage 16. Beyond this stage the $\delta^{18}\text{O}$ records show generally a trend of continuous depletion in ¹⁸O. Olausson (1990), has explained this depletion, starting from stage 12, as

Table 1. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (in ‰ versus PDB) results of separated fossils from core 290.

Depth (cm)	C. wuellerstorfi		G. inflata		N. dutertrei	
	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
21.7	1.2	2.6	0.8	1.4	0.1	1.5
31.7	1.2	2.9	0.8	1.4	0.3	1.6
42.2	0.9	2.9	0.8	1.5	0.0	1.6
49.7	1.1	3.4	0.7	1.6	0.2	1.6
59.2	1.1	3.0	0.7	1.8	0.2	1.6
71.7	1.1	3.1	0.7	1.5	0.2	1.5
79.2	1.1	3.0	0.6	1.3	0.1	1.6
89.8	1.1	2.7	0.7	1.6	-0.1	1.4
99.5	0.9	2.5	0.6	1.5	0.1	1.5
109.0	0.0	2.4	0.6	1.4	0.3	1.6
119.7	0.9	2.6	0.6	1.3	-0.1	1.8
129.2	0.8	3.1	0.2	1.6	0.0	2.1
137.2	0.9	2.4	0.5	1.5	-0.5	2.0
149.3	1.0	3.7	0.4	2.1	-0.1	2.0
159.2	1.0	3.5	0.4	2.2	-0.2	2.0
168.2	1.0	3.0	0.6	1.9	0.0	1.9
177.3	1.0	3.2	0.7	1.7	0.3	1.7
189.2	1.0	3.2	0.6	1.8	0.3	1.8
197.7	1.0	2.9	0.6	1.7	0.5	1.6
209.3	1.0	2.9	0.7	1.6	0.4	1.7
219.3	1.0	3.1	0.7	1.6	0.3	1.8
227.8	0.9	2.9	0.7	1.7	0.2	1.6
238.3	0.9	3.0	0.4	1.4	0.1	1.8
249.8			0.8	1.5	0.1	2.1
257.3	0.9	3.3	0.6	1.7	0.1	2.1
269.3	0.9	3.0	0.6	1.5	0.1	1.7
277.8	0.8	3.3	0.6	1.9	0.0	1.9
289.3	0.8	3.3	0.5	1.7	0.0	1.9
308.3	0.8	3.3	0.3	1.7	-0.2	1.9
317.3	1.0	3.2	0.7	2.1	-0.2	2.0
329.2	1.0	3.3	0.6	1.9	0.0	2.1
339.2	0.8	3.7	0.8	1.9	0.2	2.0
347.3	1.0	3.7	0.9	2.1	-0.0	2.0
359.2	1.1	3.5	0.8	1.8	0.0	1.9
369.3	0.9	3.5	0.7	1.8	0.2	2.0
378.3	1.0	3.2	0.6	2.0	-0.9	2.0
389.2	1.0	3.1	0.7	1.8	0.4	2.0
399.2	0.8	3.5	0.9	1.9	0.3	2.0
411.7	0.8	3.5	0.9	1.9	0.5	1.9
419.2	0.9	3.7	0.7	2.1	0.2	1.8
429.2	0.9	3.3	0.8	1.9	0.5	1.8
439.2	0.9	3.2	0.6	1.9	0.3	1.8
449.2	1.0	3.4	0.8	1.9	0.2	1.7
459.2	1.0	3.2	0.9	1.7	0.3	1.8
469.2	0.8	3.2	0.8	1.9	0.5	1.9
489.2	1.0	3.1	0.9	1.7	0.4	1.7
499.2	0.9	3.1	0.9	1.6	0.5	1.8
509.2	1.0	2.8	0.8	1.7	0.2	1.6
518.7	0.9	2.8	0.8	1.4	0.3	1.7
529.2	0.9	2.7	0.8	1.9	0.2	1.5
539.2	1.1	2.8	0.8	1.5	0.3	1.4
549.2	1.0	2.6	0.7	1.6	0.4	1.9
557.2	1.1	2.7	0.6	1.6	0.2	2.0
569.2			0.8	2.4	0.2	2.4
577.2	1.2	3.8	0.8	1.9	0.4	2.1
589.7	1.1	3.9	0.7	1.8	0.2	2.0
597.2	1.1	3.8	0.8	2.0	0.4	2.0
609.2	1.1	3.7	1.0	1.8	0.5	1.8
619.2	1.4	3.6	0.9	1.9	0.4	2.0
629.2	1.2	3.3	0.9	1.8	0.5	1.9
637.3	1.2	3.3	0.8	1.5	0.4	1.5
649.2	1.2	3.1	0.8	1.5	0.2	1.6
660.7	1.0	3.2	0.9	1.5	0.4	1.7

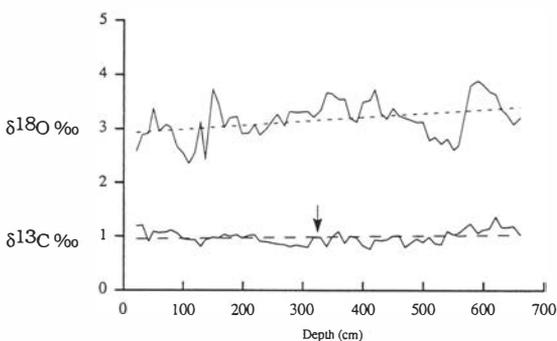


Fig. 4. Correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of *Cibicides wuellerstorfi* in core 290; The dashed lines represent a linear fit regression to the data. The arrow shows the approximate depth at which the pattern of $\delta^{13}\text{C}$ variation changes.

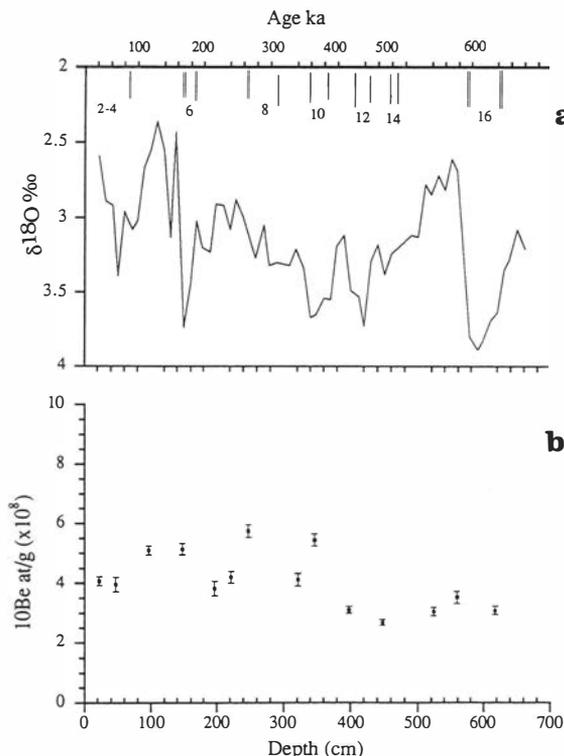


Fig. 5. The $\delta^{18}\text{O}$ data of *Cibicides wuellerstorfi* (a) and ^{10}Be concentrations of bulk samples (b) of core 290. Age scale is based on $\delta^{18}\text{O}$ stages as numbered in (a).

to a larger melt of the Antarctic ice sheet, reducing its volume to ca. 2/3 relative to its pre-Elsterian size. This melting occurred during the Elsterian-Saalian ice age, times of which were much milder than the earlier ice age, resulting in a reduced ice accumulation in the Antarctic.

Spectral correlation of $\delta^{18}\text{O}$ isotopic data of the *C. wuellerstorfi* with the isotopic stages given by Shackleton & Opdyke (1973), Imberia et al. (1984) and Zimmerman et al. (1985), indicate that core 290 extends to stage 16 (Fig. 5). Furthermore compared with the faunal stratigraphy (Phleger et al., 1953), the $\delta^{18}\text{O}$ stages correlate with cold (glacial) periods identified. These correlations give an age of about 0.65 Ma for the bottom of the core.

The $\delta^{13}\text{C}$ results of the three fossil types analysed are shown in Table 1 and Figure 3. The maximum range of variation in the $\delta^{13}\text{C}$, being between -0.9 and 0.5 ‰, occurs in *N. dutertrei*. The lowest range, between 0.8 and 1.4 ‰, is exhibited by *C. wuellerstorfi*, which shows heavier isotopic signature than the planktonic fossils. Besides the two

major peaks in Figure 3 at 1.4 m and 3.7 m, there are also two major patterns in the distribution of $\delta^{13}\text{C}$ data of the fossils. The former feature is well illustrated by the planktonic fossils and is highly reduced in the benthic, whereas the latter feature is best illustrated by the benthic fossils. The two major patterns of the $\delta^{13}\text{C}$ are separated at a depth of about 3.1 m where a slight depletion of ^{13}C ($\delta^{13}\text{C}$ changes from 1.2 to 0.8) occurs between the upper 0.21 m and 3.1 m and a slight enrichment ($\delta^{13}\text{C}$ changes from 0.8 to 1.2) from 3.1 m and downwards (Fig. 4).

As also shown by Figure 3, the calcite content profile of core 290 failed to indicate age markers that can be clearly correlated with the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data.

The ^{14}C dating

Table 2 shows ^{14}C ages of three bulk samples from core 299 and of the $> 63 \mu\text{m}$ fossil-rich fraction of three samples from core 290. The data indicate about 10 ka for the accumulation of about 40 cm of sediment in core 299 but only about 20 cm in core 290. Estimation of accumulation rate in core 290 for the first 0.5 m based on ^{14}C data gives a rate at about 2 cm/ka compared to 1 cm/ka when using $\delta^{18}\text{O}$ stages. In this interval the $\delta^{18}\text{O}$ stages are not resolved completely and thus they do not offer exact estimation of accumulation rate. However, the rest of the $\delta^{18}\text{O}$ profile (Fig. 5) indicates a variable accumulation rate in this core which may relate to a hiatus and/or erosion or to a real change in sedimentation rate.

The ^{14}C results of core 299 give an accumulation rate of about 4 cm/ka for the first 0.5 m of the core. Based on carbonate content, Olausson (1965) estimated an age of 120–140 ka for the bottom of core 299 (Fig. 6). This means an accumulation rate of about 4 cm/ka which resembles that obtained by the ^{14}C data.

Table 2. ^{14}C ages of bulk samples from core 299 and $>63\mu\text{m}$ fossil fraction from core 290.

Sample	depth (cm)	^{14}C age BP (years)
Core 299		
1	1.0	10465 ± 175
2	13.5	14930 ± 380
3	39.5	20890 ± 495
Core 290		
1	22.5	9070 ± 180
2	31.75	13630 ± 205
3	42.5	20215 ± 270

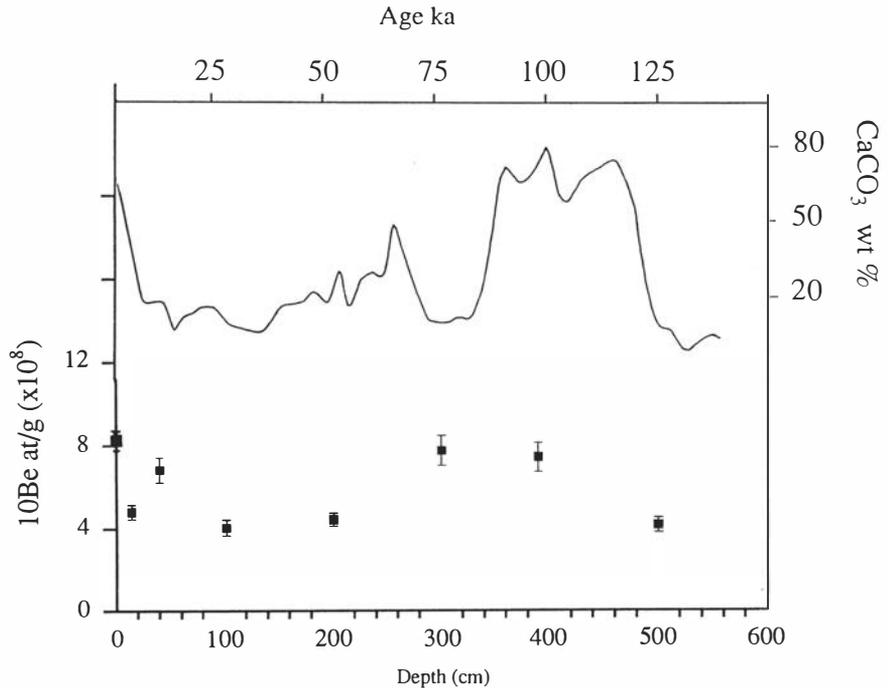


Fig. 6. ^{10}Be concentrations of bulk samples (rectangles) from core 299. The calcite content (Olausson, 1960) was used for age calibration.

The estimated accumulation rates in core 290 and 299 are comparable to other data from the eastern north Atlantic (Fig. 1) for comparative intervals of sites 552A (Zimmerman et al., 1985), 548 (Vergnaud Grazzini et al., 1985) and CH72-02 (Southon et al., 1987) which show values of about 1.7, 1.7 and 4.5 cm/ka respectively. The relatively slow rate obtained for core 290 corresponds also to the correlation based on microfossils proposed for a six site (including the studied one) profile (Phleger et al., 1953).

The ^{10}Be data

The concentrations of ^{10}Be in the cores studied are shown in Table 3 and Figures 5 and 6. The ^{10}Be concentration is normalized to gram dried sample (at 110°C). In core 290 the highest value is of samples from a depth of 247.5 cm (5.7×10^8 atoms/g) and the lowest value (2.7×10^8 atoms/g) is found in a sample from a depth of 447.5 cm. ^{10}Be concentrations in the $\geq 63 \mu\text{m}$ microfossil-rich fraction of four samples (Table 4) and of the $\leq 4 \mu\text{m}$ fraction of the sample at a depth 147.5 cm (Table 5) show reduced values when compared to their bulk sample counterparts (Table 3). Furthermore, most of the ^{10}Be con-

Table 3. ^{10}Be concentration and chemical parameters of the investigated samples from cores 290 and 299.

Sample No.	Sub-bottom depth (cm)	Ignited/dried	wt. % HCl insoluble residue	$^{10}\text{Be}^1$ ($\times 10^8$ atoms/g)
Core 290				
1	22.5	0.572	8	4.05 ± 0.15
2	47.5	0.560	7	3.94 ± 0.24
3	96.5	0.560	5	5.09 ± 0.15
4	147.5	0.588	8	5.11 ± 0.19
5	196.5	0.572	5	3.81 ± 0.24
6	221.5	0.585	4	4.20 ± 0.19
7	247.5	0.600	8	5.74 ± 0.21
8	321.5	0.576	6	4.11 ± 0.21
9	346	0.580	2	5.44 ± 0.21
10	397.5	0.568	2	3.10 ± 0.12
11	447.5	0.580	4	2.68 ± 0.11
12	523.5	0.572	3	3.04 ± 0.14
13	559	0.564	2	3.52 ± 0.20
14	617.5	0.564	6	3.07 ± 0.14
Core 299				
1	1	0.732	44	8.01 ± 0.60
2	13.5	0.850	47	4.80 ± 0.35
3	39.5	0.834	61	6.82 ± 0.61
4	101.5	0.843	51	4.05 ± 0.37
5	200.5	0.824	32	4.40 ± 0.32
6	301	0.640	24	7.73 ± 0.73
7	390.5	0.642	23	7.40 ± 0.71
8	500.5	0.832	50	4.16 ± 0.34

¹⁾ Normalized to gram dried (at 110°C) sample.

Table 4. ^{10}Be concentrations in the fossil-rich $\geq 63 \mu\text{m}$ fraction of four samples from core 290.

Sample	Depth (cm)	Wt.% ($\geq 63 \mu\text{m}$)	^{10}Be at/g ¹⁾ ($\times 10^8$)
3	96.0	36	0.75 ± 0.07
4	147.5	57	1.19 ± 0.11
8	397.5	44	1.03 ± 0.09
11	617.5	24	0.43 ± 0.04

¹⁾ corrected for the wt.% $\geq 63 \mu\text{m}$ fraction in each sample

centration in the sample from 147.5 depth occurs within the $\leq 20 \mu\text{m}$ fraction (Table 5). Microscopic examinations indicate that nannofossils (having calcite composition) are the dominant constituent of the $\leq 4 \mu\text{m}$ fraction whereas most clay minerals, feldspar, quartz, siliceous fossils and Fe and Mn oxides particles occur with the $\leq 20 \mu\text{m}$ fraction. This feature suggests that most ^{10}Be is located with the noncalcitic materials which supports similar conclusions by other authors (e.g. Bourles et al., 1989; Henken-Mellies et al., 1990). However, the noncalcitic materials in the sediments examined revealed variable mineralogy and origin (detrital and authigenic components), as is well demonstrated by samples from core 299. In this core, the ^{10}Be concentrations show a dependence on the content of HCl-insoluble fraction, as shown by Figure 7 and Tables 3 and 6. The highest ^{10}Be concentrations in HCl soluble fraction is of sample 1 at a depth of 1 cm (8.0×10^8 atoms/g) and the lowest is of sample 4 (4.1×10^8 atoms/g) at a depth of 101.5 cm (Table 3). The values of ^{10}Be in the insoluble fraction vary from 0.7 to 3.3×10^8 atoms/g (Table 6) and indicate that appreciable amounts can be retained by the insoluble fraction. This together with observed trends between the content of HCl-insoluble residue and ^{10}Be (Fig. 8a) and the ratio of ignited to dry samples and ^{10}Be (Fig. 8b) in cores 290 and 299 suggest that total dissolution is necessary to recover all ^{10}Be from sediments. However, there is no clear correlation between these parameters (Fig. 8); in particular for samples containing more than 10–15

Table 5. ^{10}Be concentrations in the size fraction of sample 4 (depth=147.5 cm) of core 290.

Fraction (μm)	Wt.%	^{10}Be at/g ¹⁾ ($\times 10^8$)
≤ 4	8.4	1.09 ± 0.06
≤ 20	34.6	2.78 ± 0.12
≥ 63	57.0	1.19 ± 0.11

¹⁾ corrected to wt.% of each fraction in the sample

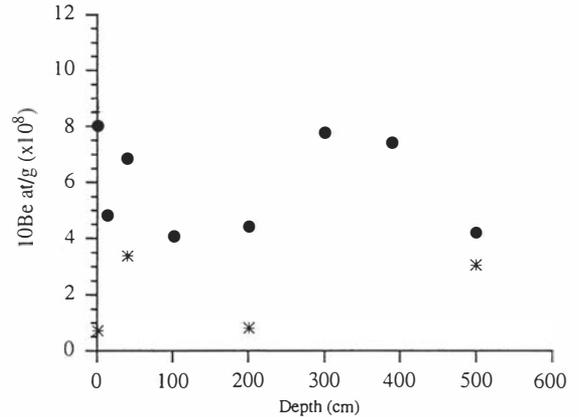


Fig. 7. ^{10}Be concentrations in bulk samples (filled circles) and HCl-insoluble fraction (stars) from core 299.

wt % HCl-insoluble residue. This variable relationship between ^{10}Be and the noncalcitic content of the samples may indicate that not only the content but most likely the type (mineralogical composition) of

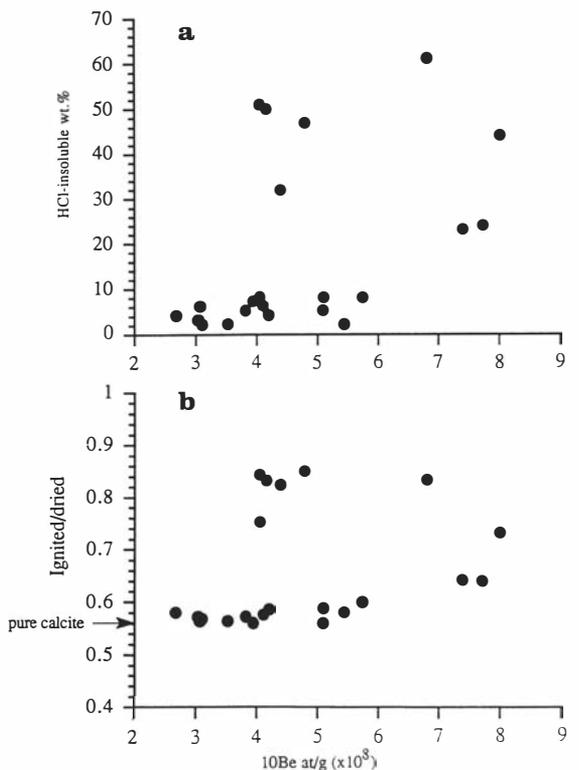


Fig. 8. Correlation between ^{10}Be concentrations and HCl-insoluble content (a) and ratio of ignited to dried samples (b) from cores 290 and 299.

Table 6. ^{10}Be concentration in the HCl-insoluble fraction of four samples from core 299.

Sample no	Sub-bottom depth (cm)	^{10}Be atoms/g ($\times 10^8$)
1	1	0.72 ± 0.07
3	39.5	3.37 ± 0.23
5	200.5	0.81 ± 0.07
8	500.5	3.04 ± 0.10

noncalcitic material partly controls the content of ^{10}Be . This phenomenon does not offer an explicit main location site(s) of ^{10}Be in the sediment study. The incompatible ionic radii of Be (0.35\AA) and Ca (0.99) may be the main reason for rejection of ^{10}Be by the calcite lattice and the preference of ^{10}Be for minerals containing elements with smaller ionic radii like clay minerals and Fe and Mn oxides and hydroxides.

As is shown by figure 8 the data on ignition weight loss can be used to decide whether to use total dissolution or HCl-dissolution procedure for extraction of ^{10}Be from a sediment sample. Theoretically, ignition of pure calcite gives a ratio of 0.56, higher values (in the samples studied) are mainly due to the occurrence of relatively volatile free components (e. g. quartz, feldspar, siliceous fossils) whereas lower values reflect volatile-rich components (e. g., organic matter, clay minerals, Fe and Mn hydroxides). In samples studied the organic matter content is rather low (organic carbon content in the cores studied reaches up to 0.4 %; Olausson, 1965) and accordingly ignition loss is mainly related to calcite, clay minerals and metal-hydroxides.

Microscopic examinations of the studied samples reveal evidence of diagenetic alteration such as, for example, dissolution of fossil shell and formation of authigenic calcite, dolomite and clay minerals. At the present time we have not investigated the effects of this alteration on variation in ^{10}Be concentrations. The few available data from the Pacific Ocean sediment (down to 1 m) indicate possible mobilization of Be during diagenesis (Sharma et al., 1989; Bourles et al., 1989). But conclusions emanating from these studies suggest a comparable decay trend of ^{10}Be in pore water, diagenetic phillipsite and enclosing sediments. This may hopefully mean a homogenized diagenetic fractionation which does not overprint the primary signal of the sediment.

In spite of the complication involved in relating main site(s) of ^{10}Be , a clear difference is observed between the ^{10}Be content of the comparatively chronological equivalent (^{14}C dated) surface sample of cores 290 and 299. ^{10}Be content of the surface sample in core 290 is 4.1×10^8 at/g (at 22.5 cm depth)

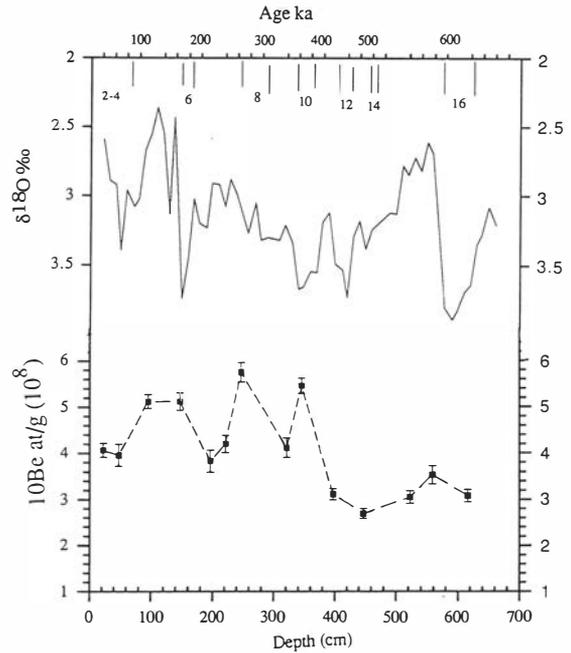


Fig. 9. Matching between ^{10}Be content of bulk samples and $\delta^{18}\text{O}$ of the *Cibicides wuellerstorfi* from core 290. The age scale is based on $\delta^{18}\text{O}$ stages as numbered below the scale.

compared to 8.0×10^8 at/g (at 1 cm depth) of core 299 which partly relates to the lower calcite content of the latter sample. Results of two surface samples collected from water depths of 2900 and 2909 m in the Atlantic (Bourles et al., 1989) and a sample from depth 3485 m (Southon et al., 1987), show concentrations that vary between 3.6 – 5.4×10^8 atoms/g. Although it is rather risky to compare these results because of bioturbation, sampling disturbance and exact depth of the surface layer, it is clear that ^{10}Be concentrations in surface sediments of the Pacific ocean are 4–10 times higher than those of the Atlantic (cf. Eisenhauer et al., 1987; Sharma et al., 1989; Bourles et al., 1989).

Discussion and conclusions

Although not yet well established, the distribution of ^{10}Be concentration is apparently not constant in the surface sediments of the ocean floor. This in addition to the non linear variation of ^{10}Be content with depth in the studied samples restricts the possibility of determining ^{10}Be age from a simple decay equation. This seems to be a problem in ^{10}Be dating

of the last million years. Therefore in this study we tried to compare the ^{10}Be data with the dating and correlation obtained from ^{14}C , $\delta^{18}\text{O}$, microfossils and carbonate content. The interesting feature in the ^{10}Be data is occurrence of enhanced concentrations (Figs. 5 and 6) which could be used as time markers. Data on ^{10}Be distribution in carbonate-rich sediments, such as the cores studied, indicate also enhanced concentrations at some levels (Southon et al., 1987; Eisenhauer et al., 1987). These enhanced signals are related to larger amounts of noncalcitic components in the sediments. Although results of this study partly support this conclusion, it is enigmatic to find a similar pattern of enhanced ^{10}Be concentration in the sediments of the Pacific ocean which are dominated (to more than 95 %) by noncalcitic components (e. g. Eisenhauer et al., 1987). Furthermore, enhanced ^{10}Be signals occur in samples from core 290 studied in spite of their rather constant noncalcitic content.

Data on ^{10}Be from ice cores (Raisbeck et al., 1987; Beer et al., 1988) also indicate enhanced ^{10}Be signals at several levels. Judging from our data and those from other sediment cores (cf. Southon et al., 1987; Eisenhauer et al., 1987; Henken-Mellies et al., 1990), the variation in content of the noncalcitic components cannot alone explain the enhanced ^{10}Be signals of carbonate-rich sediments. Furthermore, there is some periodicity of the signals which, in core 290 studied, fall around 100–130 ka, 250–270 ka, 360–370 ka and 570–580 ka (Fig. 9). Maxima with higher amplitude and lower ages are observed in ice cores; the ages fall around 14 ka, 35 ka and 60 ka. Irrespective of cause, these signals are expected to be of global extent and may appear in marine sediments. However, as suggested by Krishnaswami & Lal (1982), the enhanced signals of ^{10}Be with periodicities < 100 ka are appreciably attenuated in marine sediments due mainly to particle mixing even for a fast sedimentation rate. Thus the enhanced ^{10}Be activity in core 290 may result from Milankovitch insolation effects (mostly eccentricity related orbital forcing), for details see e. g. Berger (1989). It is difficult to match exactly the enhanced ^{10}Be signals observed in this core with $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. But a rough correlation indicates occurrence of enhanced periods at maximum temperature increase, deglaciation peaks (Fig. 9). A similar phenomenon also characterizes data on ice cores (Raisbeck et al., 1987). Evidence obtained from ice cores supports a real increase (2–3 fold) in ^{10}Be at atmospheric precipitation. This relationship is not clear but obviously reflects addition of ^{10}Be during deglaciation (warm) periods to ocean waters. Estimation of this effect by Henken-Mellies et al. (1990) suggests an increase in the order of 2–20 % in ^{10}Be

of sediment during a deglaciation period which is expected to be homogenized by bioturbation.

Another possible source of enhanced ^{10}Be signals is variation in the chemical composition of noncalcitic components (both the detrital and authigenic). Warm and cold periods around the oceans would supply mineralogically different detritus; weathering under warm climate produces more Fe- and Mn-rich particles (like clay minerals) than under cold climate. Adsorption and/or incorporation of ^{10}Be into these materials (cf. Li, 1981; Kusakabe & Ku, 1984) as well as by organic particulates (cf. Raisbeck et al., 1980; Measures & Edmond, 1983) are major transport mechanisms to the sediment. This could result in variation of ^{10}Be concentration in deep waters of the Atlantic during glaciation-deglaciation periods and may thus partly explain a development of enhanced ^{10}Be concentrations. For example, the presentday concentration of ^{10}Be in deep water of the Atlantic is less than that of the Pacific by about 25–50 %.

Although the main cause of glaciation-deglaciation episodes is still disputable, it is clear that these episodes were associated with climatic changes. Astronomical tuning related to mainly cyclic events induced by the influence of the earth's orbital geometry has been proposed as the main cause of climatic variability (e. g. Emiliani, 1978; Berger, 1989; Jouzel et al. 1989; and references therein). Also effects of solar cycle modulation and the geomagnetic field excursions on cosmic ray intensity and a variable precipitation intensity were suggested as the most favourable cause for the enhanced ^{10}Be activity in ice cores (Raisbeck et al., 1987; Beer et al., 1988). With the available data of ^{10}Be in this study, it is premature to establish the main cause(s) of the enhanced signals. But there are promising indications that such enhanced signals can be used as time markers in marine sediments. A combination of these signals with other dating markers means an interesting correlation possibility for evaluation of extraterrestrial effects on the earth climatic changes. This task is, however, still faced with several complications, among them being the effects of terrestrial processes. Several studies indicate that climatic changes may have been brought about or triggered by terrestrial catastrophic events such as intense volcanism and/or immense forest fires (cf. Porter, 1986; Jouzel et al., 1989; Scuderi, 1990). These events induce climatic noise through the input of CO_2 and acid (sulphate) aerosols to the atmosphere. The effects of these events should be understood for better evaluation of extraterrestrial effects.

Acknowledgements. – We would like to thank Cliff Robinson and Jonas Åström for their active help with the AMS measurements. Åke Nyberg helped with analysis of core 299. Kenneth Inn kindly provided us with the NIST ¹⁰Be standard. Funds for this project were provided by the Swedish Natural Science Research Council (NFR) and the Knut and Alice Wallenberg Foundation.

REFERENCES

- AlDahan, A.A. & Possnert G. 1990: Advances in the chemical separation of ¹⁰Be from sediments. *Publication of the Svedberg Laboratory and Department of Radiation Science, Uppsala*, 18 p.
- Bé, A.W.H. 1977: An ecological, zoogeographical and taxonomic review of recent planktonic foraminifera. In Ramsay A.T.S. (ed.): *Oceanic Micropaleontology*, Academic Press, New York, p. 1–88.
- Beer, J., Siegenthaler U., Bonani, G., Finkel, R.C. Oeschger, H., Suter, M. & Wölfli, W. 1988: Information on past solar activity and geomagnetism from ¹⁰Be in the Camp Century ice core. *Nature* 331: 675–679.
- Berger, A. 1989: Pleistocene climatic variability at astronomical frequencies. *Quaternary International* 2, 1–14.
- Bourles, D., Raisbeck, G.M. & Yiou, F. 1989: ¹⁰Be and ⁹Be in marine sediments and their potential for dating. *Geochimica et Cosmochimica Acta* 53: 443–452.
- Eisenhauer, A., Mangini, A., Segl, M., Beer, J., Bonani, G., Suter, M. & Wölfli, W. 1987: High resolution ¹⁰Be and ²³⁰Th profiles in DSDP site 580. *Nuclear Instruments and Methods in Physical Research B* 29: 326–331.
- Emiliani C. 1955: Pleistocene temperatures. *Journal of Geology* 63: 538–578.
- Emiliani C. 1971: Depth habitats of growth stages of pelagic foraminifera. *Science* 173: 1122–1124.
- Emiliani C. 1978: The cause of the ice ages. *Earth and Planetary Science Letters* 37: 349–352.
- Erez, J. & Honjo S. 1981: Comparison of isotopic composition of planktonic foraminifera in plankton tows, sediment traps and sediments. *Paleogeography, Paleoclimatology, Paleoecology* 33: 129–156.
- Graham, D.W., Corliss, B.H., Bender, M.L. & Keigwin, L.D. Jr 1981: Carbon and oxygen isotopic disequilibria of recent deep-sea benthic foraminifera. *Marine Micropaleontology* 6: 483–497.
- Henken-Mellies, W.U., Beer, J., Haller, F., Hsu, K.J., Shen C., Bonani, G., Hofman, H.J. Suter, M. & Wölfli, W. 1990: ¹⁰Be and ⁹Be in south Atlantic DSDP site 519: relation to geomagnetic reversals and to sediment composition. *Earth and Planetary Science Letters* 98: 267–276.
- Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W. & Shackleton, N.J. 1984: The orbital theory of Pleistocene climate: support from a revised chronology of the marine ¹⁸O record. In Berger, A. (ed.), *Milankovitch and Climate*, D. Reidel, Dordrecht, p. 269–305.
- Jansen, E., Bliel, U., Henrich, R., Kringstad, L. & Slettmark B. 1988: Paleoenvironmental changes in the Norwegian Sea and the northwest Atlantic during the last 2.8 Ma: DSDP/ODP sites 610, 642, 643, and 644. *Paleoceanography* 3: 563–581.
- Jansen E. 1989: The use of stable oxygen and carbon isotope stratigraphy as a dating tool. *Quaternary International 1*: 151–166.
- Jouzel, J. Barkov, N.I., Barnola, J.M., Genthon, C., Korotkevitch, Y.S., Kotlyakov, V.M., Legrand, M., Lorius, C., Petit, J.P., Petrov, V.N., Raisbeck, G., Raynaud D., Ritz, C. & Yiou, F. 1989: Global changes over the last climatic cycle from the Vostok ice core record (Antarctica). *Quaternary International* 2: 15–24.
- Krishnaswami, S. & Lal, D. 1982: Deep Sea Sedimentation: Processes and chronology. In Currie, L.A. (ed.): *Nuclear and chemical dating techniques*, pp. 363–388. *American Chemical Society, Ser. No. 176*, Washington, DC.
- Ku, T.L., Southon, J.R., Vogel, J.S., Liang, Z.C. Kusakabe, M. & Nelson, D.E. 1985: ¹⁰Be distribution in deep sea drilling project site 576 and site 578 sediments studied by accelerator mass spectrometry. *Initial Reports of the Deep Sea Drilling Project vol. LXXXVI*, p. 539–546.
- Kusakabe, M. & Ku, T.L. 1984: Incorporation of Be isotopes and other trace metals into marine ferromanganese deposits. *Geochimica et Cosmochimica Acta* 48: 2187–2193.
- Laberyie, L.D. & Duplessy, J.-C. 1985: Changes in the oceanic ¹³C/¹²C ratio during the last 140000 years: high latitude surface water records. *Palaeogeography, Palaeoclimatology, Palaeoecology* 50: 217–240.
- Li, Y.H. 1981: Ultimate removal mechanisms of elements from the ocean. *Geochimica et Cosmochimica Acta* 45: 1659–1664.
- Measures, C.I. & Edmond, J.M. 1983: The geochemical cycle of ⁹Be: a reconnaissance. *Earth and Planetary Science Letters* 66: 101–110.
- Mix, A.C. & Fairbanks R. 1985: North Atlantic surface-ocean control of Pleistocene deep-ocean circulation. *Earth and Planetary Science Letters* 73: 231–243.
- Olausson, E. 1960: Description of sediment cores from the north Atlantic ocean. *Reports of the Swedish Deep-Sea Expedition, 1947–1948*, 7 (5), p. 229–286.
- Olausson, E. 1965: Evidence of climate changes in north Atlantic Deep-Sea cores. In Sears M. (ed.), *Progress in Oceanography*, Pergamon Press, New York, p. 221–254.
- Olausson, E. 1971: Quaternary correlations and the geochemistry of oozes. In Funnell B.M. (ed.), *The Micropaleontology of Oceans*, Cambridge Univ. Press, Cambridge p. 375–398.
- Olausson, E. 1990: A post-Cromerian rise in sea-level. *Proceedings of the International Conference on the Role of the Polar Regims in Global Change, Fairbanks, Alaska* (in print).
- Phleger, F.B., Parker, F.L. Peirson, J.F. 1953: North Atlantic Foraminifera. Sediment cores from the North Atlantic Ocean. *Reports of the Swedish Deep-Sea Expedition, 1947: 1948*, 7 (1), p. 1–22.
- Porter, S.C. 1986: Pattern and forcing of northern hemisphere glacier variations during the last millennium. *Quaternary Research* 26: 27–48.
- Possnert, G. 1990: Radiocarbon dating by the accelerator technique. *Norwegian Archaeological Review* 23: 30–37.
- Possnert, G. & Åström, J. 1986: ¹⁴C measurements with the Uppsala tandem. In Hedges, R.E.M. & Hall, E.T. (eds.), *Workshop on techniques in accelerator mass spectrometry*, Oxford, p. 167–170.
- Prell, W.L. 1982: Oxygen and carbon isotope stratigraphy for the Quaternary of hole 502B: Evidence for two

- modes of isotopic variability. *Initial Reports of the Deep Sea Drilling Project vol. LXVIII*, p. 455–464.
- Raisbeck, G.M., Yiou, F., Fruneau, M., Loiseaux, J.M., Lieuvin, M., Ravel, J.C., Reys, J.M. & Guichard, F. 1980: ^{10}Be concentration and residence time in the deep ocean. *Earth and Planetary Science Letters* 51: 275–278.
- Raisbeck, G.M., Yiou, F., Bourles, D., Lorius, C., Jouzel, J. & Barkov, N.I. 1987: Evidence for two intervals of enhanced ^{10}Be deposition in Antarctic ice during the last glacial period. *Nature* 326: 273–277.
- Scuderi, L.A. 1990: Tree-ring evidence for climatically effective volcanic eruptions. *Quaternary Research* 34: 67–85.
- Shackleton, N.J. 1974: Attainment of isotopic equilibrium between ocean water and benthonic foraminifera genus *Uvigerina*: isotopic changes in the ocean during the last glacial. *Colloques Internationaux du CNRS* 219: 203–209.
- Shackleton, N.J. & Opdyke, N.D. 1973: Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28–238: Oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale. *Quaternary Research* 3: 39–55.
- Sharma, P., Church, T.M. & Bernat, M. 1989: Use of cosmogenic ^{10}Be and ^{26}Al in phillipsite for the dating of marine sediments in the south Pacific ocean. *Chemical Geology (Isotope Geosci. Sec.)* 73: 279–288.
- Southon, J.R., Ku, T.L., Nelsin, D.E., Reys, J.L., Duplessy, J.C. & Vogel, J.S. 1987: ^{10}Be in a deep-sea core: implication regrading ^{10}Be production changes over the past 420 ka. *Earth and Planetary Science Letters* 85: 356–364.
- Tanaka, S. and Inoue, T. 1980: ^{10}Be evidence for geochemical events in the north Pacific during the Pliocene. *Earth and Planetary Science Letters* 49: 34–38.
- Vergnaud Grazzini, C. & Saliège, J.F. 1985: Oxygen and carbon isotopic composition of benthic foraminifera at deep sea drilling project sites 548 and 549 (hole 549A): Pleistocene climatic changes and circulation in the northeastern Atlantic. *Initial Reports of the Deep Sea Drilling Project vol. LXXX, part 2*, p. 805–815.
- Zahn, R., Rushdi, A., Pisias, N.G., Bornhold, B.D., Blaise, B. & Karlin R. 1991: Carbonate deposition and benthic $\delta^{13}\text{C}$ in the subarctic Pacific: implication for changes of the oceanic carbonate system during the past 750000 years. *Earth and Planetary Science Letters* 103: 116–132.
- Zimmerman, H.B., Shackleton, N.J., Backman, J., Kent, D.V., Baldauf, J.G. Kaltenback, A.J. & Morton, A.C. 1985: History of Plio-Pleistocene climate in the northeastern Atlantic, deep sea drilling project hole 552A. *Initial Reports of the Deep Sea Drilling Project vol. 81*, p. 861–875.