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STABLE ISOTOPE RECORD AND LATE QUATERNARY SEDIMENTATION RATES AT THE ANTARCTIC CONTINENTAL MARGIN

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ABSTRACT. Four cores from the Antarctic continental margin located between 50 and 200 km from the present-day ice-shelf edge, were selected for sedimentological and mass spectrometer analyses. The first stable isotope records of the Southern Polar Ocean can be correlated in detail with global isotope stratigraphy. Together with magnetostratigraphic, sedimentological and micropaleontological data, the record provides stratigraphic and paleoceanographic information back to the Jaramillo Subchron (910 ka).

Although the isotope values have been altered by diagenetic processes in the sediments, which are poor in carbonate, an interpretation is possible using correlation with the sedimentological parameters. Oxygen isotope data give indications for a meltwater spike at the beginning of interglacials, when large scale melting of parts of the ice-shelves took place. The synchronous record of the benthic and planktic $\delta^{13}C$ signals reflect continuous bottom water formation also during glacials.

Primary productivity was strictly reduced during glacials due to continuous ice coverage in the Weddell Sea. The climatic improvement at the beginning of an interglacial is associated with peak values in biologic activity lasting for about 15 kyr.

During one climatic cycle, mean sedimentation rates at the continental margin decrease with increasing distance from the continent from 5.2 to 1.3 cm/kyr. Maximum sedimentation rates of 25 cm/kyr at the beginning of an interglacial and minimum sedimentation rates of 0.6 cm/kyr during glacial periods have been calculated. These rates are mainly controlled by movements of the ice-shelf edge and ice-rafting.

Introduction

The Antarctic circumpolar ocean with its extensive ice-shelves in the Weddell and Ross Seas, is one of the Earth's principal sources of oceanic bottom waters and strongly influences the global climate (Foster & Middleton, 1980). The quantitatively most important part of the Antarctic Bottom Water (AABW) originates in the Weddell Sea and its influence on sedimentation has been detected as far north as 40°. The downcore distribution of Quaternary sediments from the Weddell Sea is, via complex interactions between the ice-

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shelf edge, the sea ice distribution, and oceanographic and biologic processes, related to global climatic changes (Pudsey *et al.*, 1988; Fütterer *et al.*, 1988). On the eastern continental margin of the Weddell Sea, distinct cycles in the sedimentary environment mainly reflect changes in the hydrographic regime in response to the retreat and advance of ice-shelves and sea ice (Grobe, 1986).

Prior to this study, sediments from south of the present-day Antarctic Polar Front have not been included in the construction of Quaternary isotope stratigraphy, because of the paucity of biogenic carbonate (Hays *et al.*, 1976; Shackleton, 1977; Imbrie *et al.*, 1984; Pisias *et al.*, 1984; Prell *et al.*, 1986; Martinson *et al.*, 1987; Williams *et al.*, 1988). All interpretations of southern high latitude glacial and interglacial paleoenvironments to date (*e.g.* Grobe, 1986; Ledbetter & Ciesielski, 1986; Burckle & Abrams, 1987; Pudsey *et al.*, 1988; Fütterer *et al.*, 1988) have therefore suffered from the lack of detailed, worldwide correlatable, stratigraphic data. Only a few isotope records from sub-Antarctic cores are published (Hays *et al.*, 1976; Labeyrie & Duplessy, 1985; Labeyrie *et al.*, 1987; Oppo & Fairbanks, 1987; Curry *et al.*, 1988). Recently, a stable isotope stratigraphy based on both benthic and planktic foraminifers from 69°S has been presented and correlated in detail with the global isotope stratigraphy (Mackensen *et al.*, 1989).

In this paper we use the new stratigraphy scheme as a reference both for correlation within a suite of cores perpendicular to the eastern Weddell Sea continental margin, and also for correlation with the global climate cycles as documented in sediments and in oxygen and carbon isotopes of planktic (*N. pachyderma*) and benthic (*E. exigua*) foraminifers. The high resolution stratigraphy permits calculation of sedimentation rates in the various cores during individual climatic cycles.

AREA OF INVESTIGATION

The continental margin in the eastern part of the Weddell Sea can be divided into a continental shelf (>500/600 m water depth), an upper slope (500 - 1800 m), a slightly dipping flat terrace, about 200 km wide, (1800 - 3000 m), and a lower slope down to the abyssal sea floor (4500 m, Fig. 2). This subdivision of the slope morphology is responsible for the characteristic sedimentary conditions and the accumulation of mostly undisturbed sedimentary sequences, which are merely affected by gravitational transport and turbidites.

Today, the sedimentary environment on the shelf is controlled by the Antarctic Coastal Current transporting very cold (-1.8 °C) water masses of low salinity (34.3 - 34.4‰) southwestwards (Eastern Shelf Water, ESW; Foldvik *et al.*, 1985). On the upper slope, down to about 2000 m water depth, the relatively warm Weddell Deep Water (WDW) and the modified Weddell Deep Water reside. In the area of investigation the continental terrace and upper parts of the lower slope between 2000 m and 4000 m are overlain by the Antarctic Bottom Water (AABW), whereas the colder Weddell Sea Bottom Water (WSBW) is found below 4000 m (Carmack & Foster, 1977; Weiss *et al.*, 1979; Foldvik *et al.*, 1985; Østerhus, 1987). Analysis of 45 box core samples revealed virtually carbonate-free surface sediments under the main part of the WDW flow as well as at abyssal depths below 4000 m where the WSBW is present (Mackensen *et al.*, in press). On the continental terrace and upper parts of the lower slope, over which the AABW flows, surface sediments contain between 1 and 5% $CaCO_3$. Sedimentological investigations from this area have revealed that Quaternary sediments from the continental terrace between 1800 and 3000 m water depth contain sufficient calcareous foraminifers for stable isotope analysis (Fig. 1).

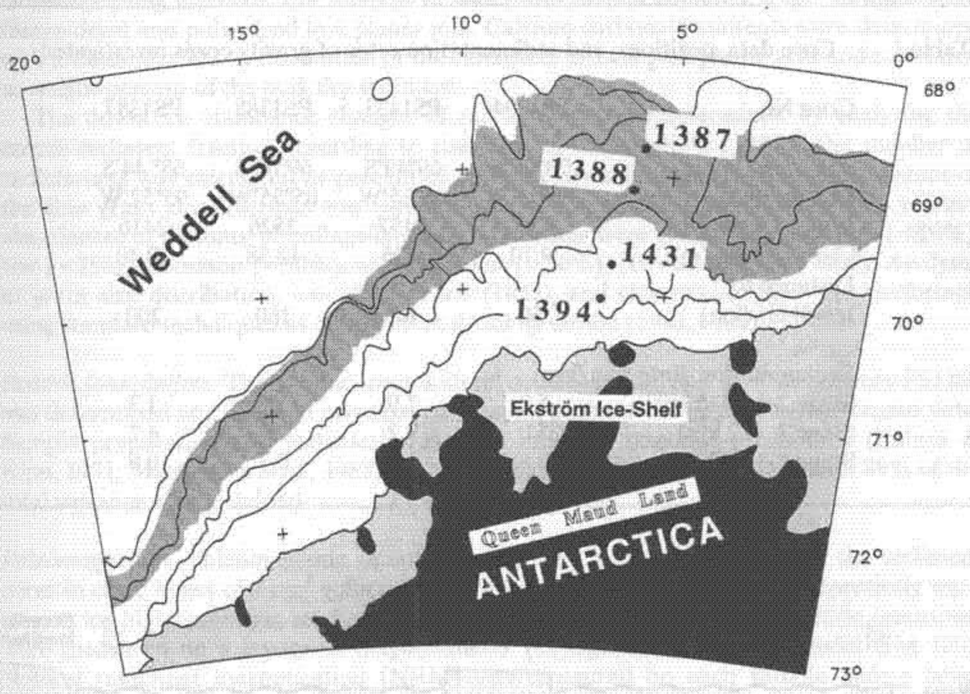

Figure 1 Locations of gravity cores from the eastern Weddell Sea continental margin off Ekström Ice-Shelf. Shaded area indicates surface sediments with significant carbonate content (1-10%) mainly consisting of the planktic foraminifer *Neogloboquadrina pachyderma*.

MATERIAL AND METHODS

Sediment cores were recovered during *Polarstern* cruises ANT I/2 (1983) and ANT IV/3 (1985/86), (Hempel, 1983; Fütterer, 1987) on four profiles perpendicular to the continental margin between 68°S and 72°S. The selection of core locations was based on 3.5 kHz sediment echosoundings. Preliminary sedimentological investigations in this area were carried out along a profile off Kapp Norvegia (Grobe, 1986). Four gravity cores (diameter 12 cm) from the continental terrace off Atka Bay have been chosen as representative of the sedimentary conditions at water depths between 1800 and 3000 m and investigated in detail (Table 1, Fig. 1). PS1394 is located on a sediment fan at the foot of the upper steep slope, PS1431 was taken from the upper slope, PS1388 from the lower central slope, and PS1387 from the northernmost part of the terrace (Fig. 2).

Sedimentologic Parameters. All cores were opened and described on board. Samples for micropaleontological, sedimentological, and paleomagnetic investigations were taken every 10 cm downcore unless obvious changes in color and grain size of sediments necessitated

Table 1 Core data, positions, and sedimentation rates of gravity cores investigated.

Core No.	PS1394	PS1431	PS1388	PS1387
Latitude	70°06'S	69°49'S	69°02'S	68° 44'S
Longitude	06°51'W	06°35'W	05°55'W	05°52'W
Water Depth (m)	1700	2457	2526	2416
Core Length (m)	9.10	9.35	12.38	10.40
Distance to Ice-Shelf (km)	50	80	160	200
Sedimentation Rate (cm/kyr)				
Stage 2-5 (115 kyr)	5.2	2.9	1.9	1.3
Stage 1 (12 kyr)	25.0	9.2	5.0	4.2
Stage 2-4 (62 kyr)	3.6	1.0	0.7	0.6

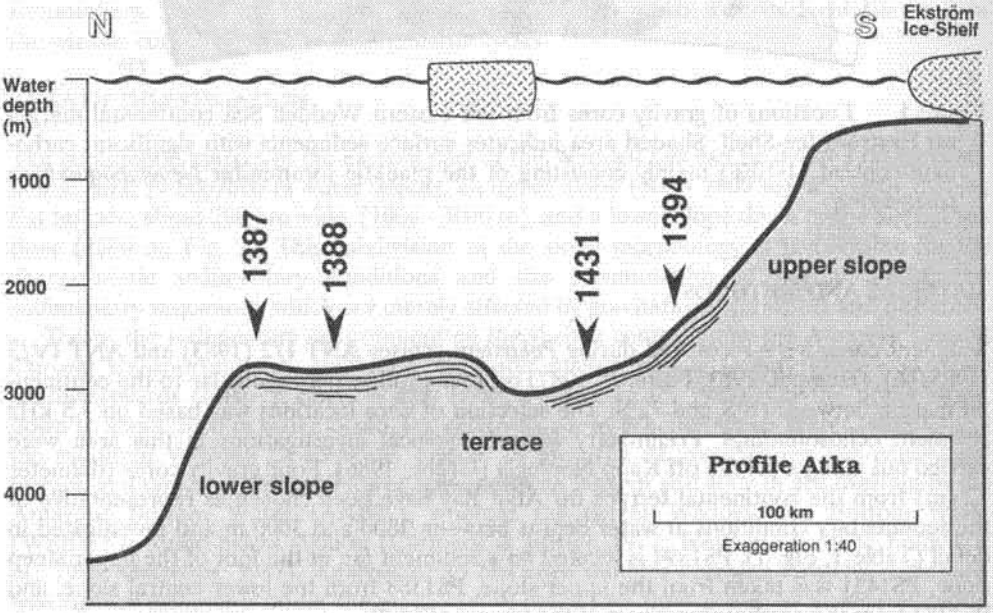

Figure 2 Morphology of the investigated profile off Ekström Ice-Shelf. The continental margin in this part of the Weddell Sea can be divided into the shelf, an upper slope, a flat terrace and a lower slope down to the abyssal sea floor. Core locations are indicated by arrows.

closer sampling intervals. For analysis of water and carbon contents, 2 cm^3 samples were freeze-dried and pulverized in a planet mill. Calcium carbonate contents were determined with a Coulomat 702 by dissolution of the carbonate in 14% phosphoric acid and calculated as weight-percent of the bulk dry sediment.

The downcore abundance changes of radiolarians were determined by analyzing the coarse sediment fraction according to the method of Sarnthein (1971). The number of radiolarians was calculated as percent of the bulk dry sediment. The biogenic content of the fine grain size fractions was estimated from smear-slide investigation. The relative abundances of diatoms, silicoflagellates, and coccoliths were given as barren, rare (<10%), few (<25%), common (<50%), and abundant (>50%), (Grobe & Kuhn, 1987). Analyses of grain size distribution, ice-rafted debris (IRD), and clay mineralogy were performed using standard techniques as described in detail by Grobe (1986, 1987).

Benthic foraminifers. The benthic foraminiferal content (>125 mm fraction) of core PS1388 was determined and Q-mode principal component analysis was applied on the census data. Sample preparation and statistical treatment followed standard procedures (Imbrie & Kipp, 1971; Mackensen *et al.*, 1985). Three principal components which explain 89% of the total variance were retained.

Paleomagnetism. Paleomagnetic samples were taken from the split half of the sediment cores in cubic boxes of 7 cm^3 volume. Sampling intervals of 5 and 10 cm, respectively were chosen for high resolution studies. Paleomagnetic directions and magnetization intensities were measured on a cryogenic magnetometer (Cryogenic Consultans, Model GM 400). Natural remanent magnetization (NRM) was measured on each sample before being subjected to a systematic demagnetization treatment (Schoenstedt GSD-1 single axis demagnetizer) involving approximately 10 steps on each sample with a maximum alternating field intensity of at least 80 mT.

Stable isotopes. The stable isotope measurements were made on planktic foraminiferal tests of sinistrally coiled *Neogloboquadrina pachyderma* with samples containing on average 6 specimens of the 125-250 mm size fraction. In addition, in Core PS1388 the tests of the benthic foraminiferal species *Epistominella exigua* were analysed. This species was found to occur persistently throughout the core, although in very low numbers in some horizons (Mackensen *et al.*, 1989). *Epistominella exigua* is an epibenthic abyssal species which feeds on phytodetritus (Mackensen *et al.*, 1985; Gooday, 1988). Tests of *E. exigua* collected live are precipitated close to the $\delta^{18}O$ isotopic equilibrium of ambient bottom water (Barrera & Mackensen, unpubl. data).

Wherever possible, only foraminifers without secondary calcification or signs of partial dissolution were selected for isotope measurements. The foraminifers were dissolved in 100% orthophosphoric acid at 70 °C using an automatic carbonate preparation device with 46 positions (Finnigan MAT) which is connected on-line to a Finnigan MAT 251 mass spectrometer. Unlike other preparation techniques, by this method each sample is reacted separately by adding a defined quantity of acid. This avoids a possible memory effect which may occur if numerous samples react in a common acid bath. The amount of carbonate used for one measurement varies from 40 to 60 μg, but if necessary, even samples weighing as little as 20 μg can be measured successfully. This corresponds to either 2 - 6 specimens of *N. pachyderma* or about 20 - 30 tests of *E. exigua*.

In areas and periods with low sedimentation rates, the analysis of only a few specimens may reveal an isotope signal severely biased because of bioturbation (Schiffelbein, 1986; Zahn *et al.*, 1986). The measurement of paired samples from the same stratigraphic level, however, did not show significant differences between subsamples. We therefore assume that bioturbation has had no drastic influence on the reliability of our data.

All measurements were carried out against a laboratory standard (natural well gas) which has been calibrated to the VPDB scale (Vienna Pee Dee Belemnite) using the IAEA (International Atomic Energy Agency) distributed reference sample NBS 19 (Hut, 1987). The linearity and proportionality of the mass spectrometer with respect to the VPDB scale was checked using the NBS 18 and NBS 20 standards. Standard deviations of measurements, as determined by 500 measurements of our laboratory carbonate standard (Jurassic limestone) which was measured routinely on 8 positions within each sample carrousel, are $<0.06^{o}/_{oo}$ for $^{12}C/^{13}C$ and $<0.09^{o}/_{oo}$ for $^{16}O/^{18}O$. Measurements have been corrected for mass spectra effects according to the method of Craig (1957) and isotope ratios are given in δ-notation versus VPDB.

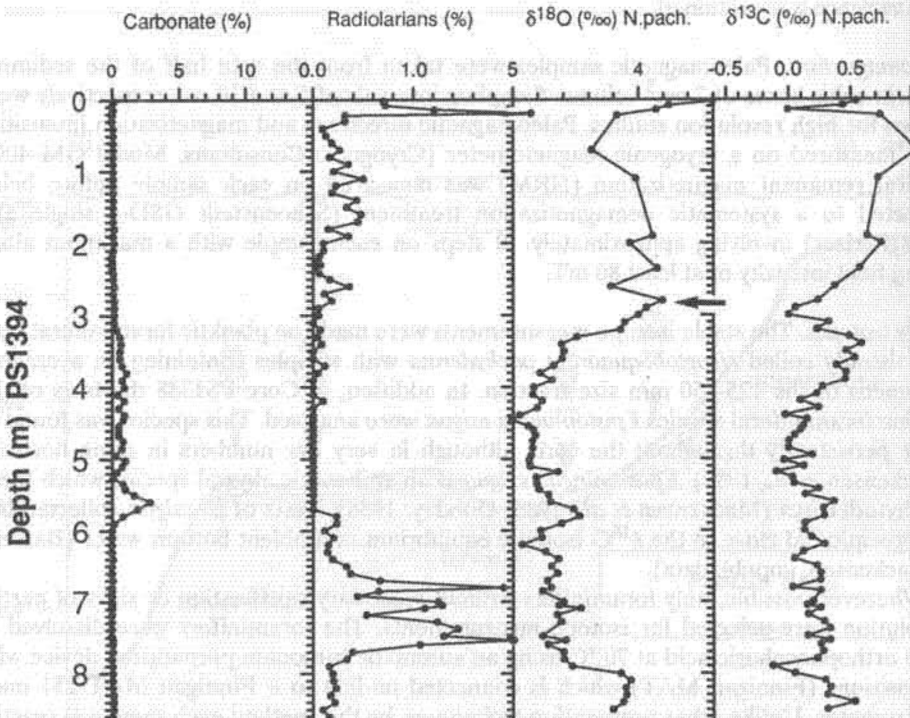

Figure 3 Stable isotope record and contents of biogenic components in Core PS1394. Note high sedimentation rate of 25 cm/kyr during the Holocene Epoch. Arrow indicates Antarctic meltwater spike during deglaciation.

Results

Distribution of Biogenic Components. On the eastern Weddell Sea continental margin the typical glaciomarine sediments consist of silty clay or clayey silt with varying amounts of sand and gravel. In all cores horizons with a significant carbonate content of 1 - 15% alternate with horizons lacking calcareous particles (Figs. 3 - 6, Appendix I - IV). The carbonate content originates from planktic foraminifers (*N. pachyderma*) and a few benthic foraminiferal species. Rare coccoliths only occur in horizons with high carbonate content. The thickness of the uppermost calcareous sediment unit decreases from 3 to 1.5 m with increasing distance from the continent, whereas the carbonate content increases from 1 to 20%. Below 7 m and 4.5 m in Cores PS1388 and PS1387 respectively, carbonate is present throughout (Figs. 5, 6).

Siliceous microfossils are concentrated in distinct horizons similar to the carbonate-rich layers. The thickness of the opal-bearing sediment sequence decreases from 3 to 0.5 m with increasing distance from the continent (Figs. 3 - 6). Opaline particles of the sand size frac-

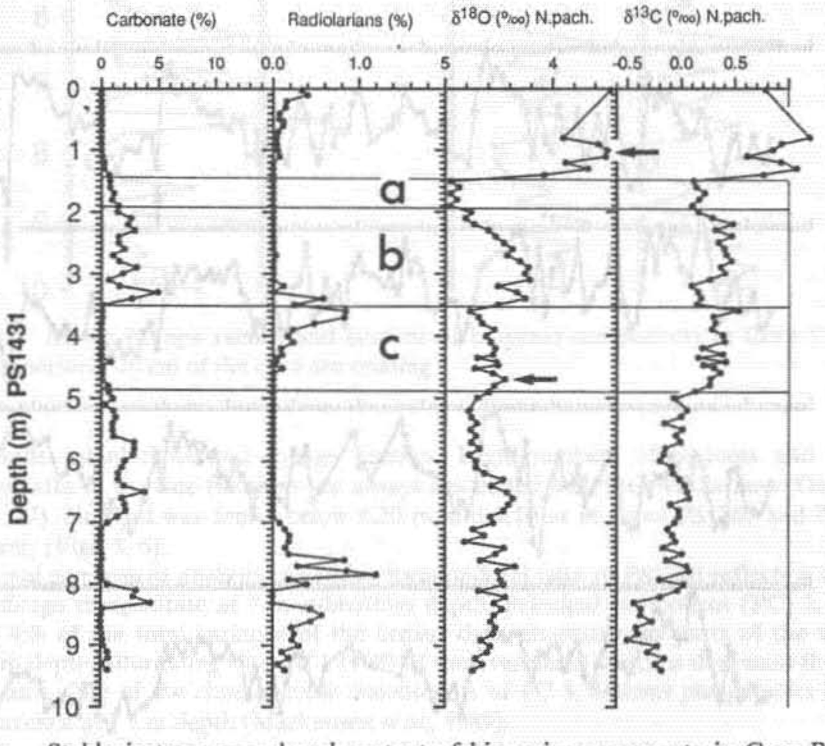

Figure 4 Stable isotope record and content of biogenic components in Core PS1431. Letters a, b and c indicate different 'diagenetic regimes'. In regime c isotopic data are strongly influenced by diagenetic processes due to the low carbonate content, oxygen isotope ratios are changed to heavier, carbon isotope ratios to lighter values. Arrows indicate Antarctic meltwater spike during deglaciation.

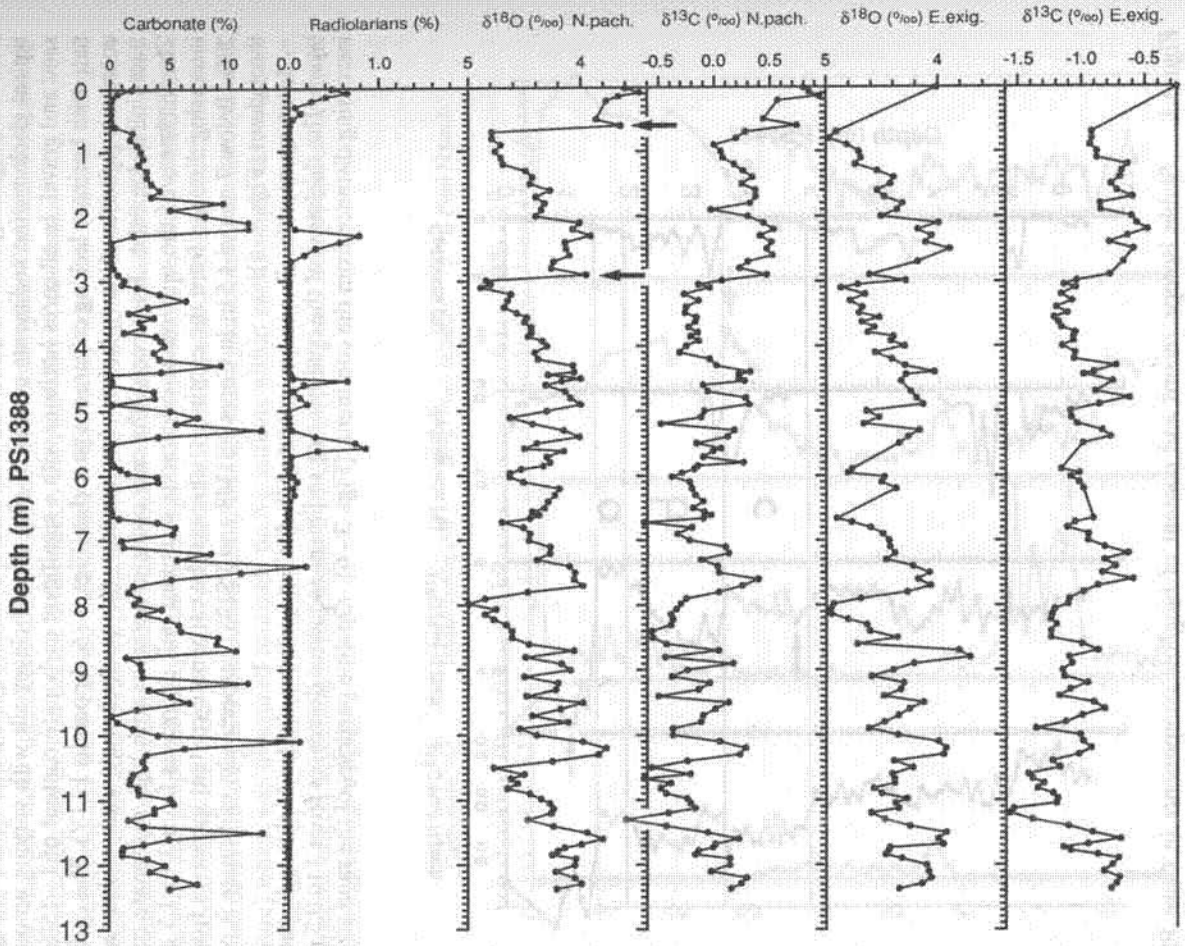

Figure 5 Stable isotope record and content of biogenic components in Core PS1388. Planktic (*N. pachyderma*) as well as benthic (*E. exigua*) foraminifers were measured. Arrows indicate Antarctic meltwater spike during deglaciation.

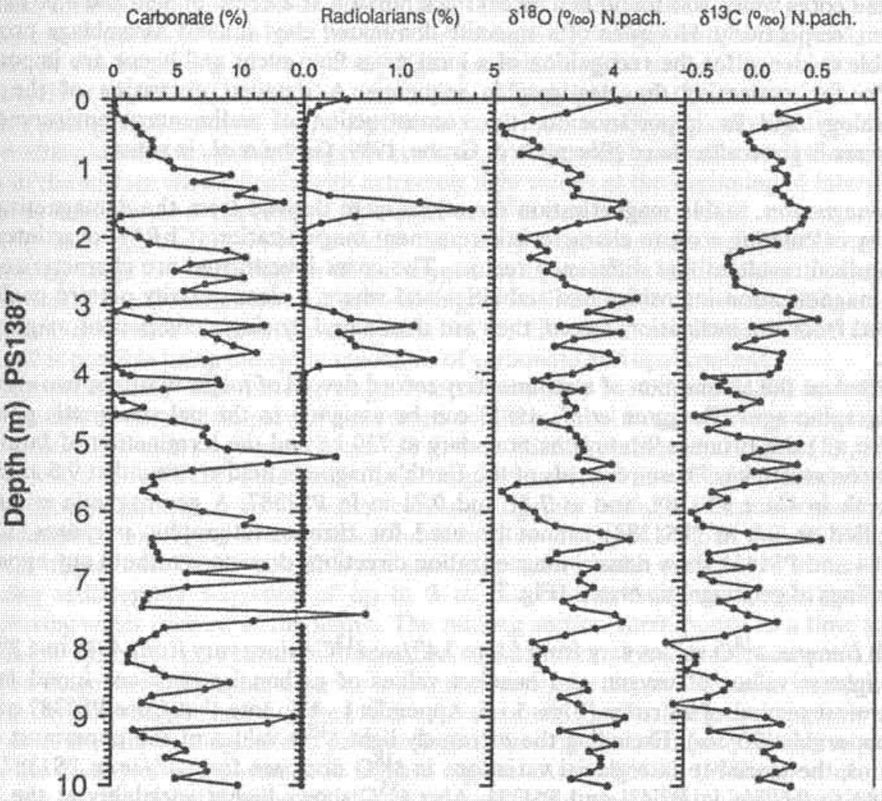

Figure 6 Stable isotope record and content of biogenic components in Core PS1387. The uppermost 20 cm of the core are missing.

tion include radiolarians and sponge needles. High numbers of diatoms and a few silicoflagellates in the fine fractions are always associated with the radiolarians (Grobe & Kuhn, 1987). No opal was found below 6.20 m and 4.10 m in cores PS1388 and PS1387 respectively (Figs. 5, 6).

Principal component analysis of benthic foraminiferal data of PS1388 reflects a change in assemblage composition at 7 m subbottom depth. Principal component (PC) 3, which explains 4% of the total variance of the census data, characterizes parts of the section below 7 m depth, alternating with PC 1 (79% of total variance) which is dominant throughout the core. One of the characteristic constituents of PC 3, *Melonis pompilioides* has its last occurrence at 7.1 m depth (Mackensen *et al.*, 1989).

Sedimentology. Water content, calculated in percentage of dry weight, decreases downcore from 90% in the uppermost centimeters to 50% at a depth of 10 m. A pronounced decrease in water content of 10% at 6.10 m in Core PS1388 parallels a decrease of about 5% at 4.20 m depth in Core PS1387. This corresponds to the maximum amount of smectite in these cores which was found in a 40 cm thick horizon at a depth of 5.80 and 4.00 m sub-bottom, respectively. Horizons of a smectite-dominated clay mineral assemblage provide valuable evidence for the recognition of a local mass flow event and hence are important criteria for evaluating the stratigraphic sequence. A detailed discussion of the clay mineralogy and its importance for the reconstruction of sedimentary environmental processes is given elsewhere (Ehrmann & Grobe, 1989; Grobe *et al.*, in press).

Paleomagnetism. Stable magnetization directions were derived from the demagnetization data by calculating a mean characteristic remanent magnetization (ChRM) over intervals of identical resultant and difference vectors. The cores investigated are characterized by high magnetization intensities and stabilities and where a clear polarity pattern could be defined from the inclination record, they are dominated by single component magnetizations.

Based on the assumption of a sedimentary record devoid of major hiatuses, two chronostratigraphic ages (Berggren *et al.*, 1985) can be assigned to the paleomagnetic polarity pattern, (1) the Brunhes/Matuyama boundary at 730 ka and the termination of Jaramillo Subchron at 910 ka. These reversals of the Earth's magnetic field are found at 9.5 and 12.1 m depth in Core PS1388, and at 7.11 and 9.71 m in PS1387. A geomagnetic excursion identified at 9.8 m (PS1388) cannot be used for chronostratigraphic purposes. Cores PS1394 and PS1431 show normal magnetization directions downcore without any apparent recordings of geomagnetic events (Fig. 7).

Stable Isotopes. $\delta^{18}O$ values vary from 5.0 to 3.4°/oo; $\delta^{13}C$ values vary from -0.75 to 1.2°/oo. The lightest values of oxygen and heaviest values of carbon isotopes are found in the uppermost section of all cores (Figs. 3 - 6, Appendix I - IV; note that Core PS1387 misses the uppermost 20 cm). Excluding the extremely light $\delta^{18}O$ values of the uppermost core sections, the glacial to interglacial variations in $\delta^{18}O$ decrease from 1°/oo in PS1387 and PS1388 to 0.6°/oo in PS1431 and PS1394. Also $\delta^{13}C$ shows higher variability at the core locations furthest from the continent. Heaviest oxygen values are found in Core PS1394 at 4.50 m, PS1431 at 1.70 m, PS1388 at 0.80 m and in PS1387 at 0.40 m depth. A further maximum of $\delta^{18}O$ values occurs at 8.00 m in PS1388 and at 5.70 m in PS1387.

Discussion

Oxygen Isotope Record. The oxygen isotope record in Cores PS1387 and PS1388 is complete and typical of the deep sea record elsewhere back to early stage 7 (Fig. 7). In the benthic record of PS1388, stage 5 shows the characteristic three less positive $\delta^{18}O$ peaks (Emiliani, 1955; Imbrie *et al.*, 1984; Pisias *et al.*, 1984; Prell *et al.*, 1986) and stage 7 is characterized by the marked positive $\delta^{18}O$ excursion (Shackleton & Opdyke, 1973; Hays *et al.*, 1976). X-radiographs of core slabs show a weakly laminated horizon at 5.57 m in PS1388 which may represent a small distal turbidite in event 7.5 sediments.

The most positive isotopic events at 0.7 to 0.8 m depth in Core PS1388 were chosen as stratigraphic fixed points for a time-scale because benthic and planktic records coincide at these levels (Mackensen *et al.*, 1989). Although $\delta^{18}O$ fluctuations in the benthic and planktic records are almost parallel, they differ from the position of a few of the less positive peaks during interglacials. This may be due to the fact that in some of the horizons benthic foraminifers are rare or absent, and the few planktic individuals picked out for isotope analyses could have been reworked. Alternatively, a slight lead of the planktic $\delta^{18}O$ values at the beginning of isotopically light periods, may be caused by a meltwater lid (Labeyrie *et al.*, 1986). During the destruction of large parts of the Antarctic ice-shelves during post-glacial sea-level rise, a significant contribution of isotopically light water from the melting ice-shelves and icebergs (Weiss *et al.*, 1979) could have influenced the composition of the surface water. Peaks with extremely light values at the beginning of interglacials point to a meltwater event (stage 1, 5; arrows on $\delta^{18}O$ curve; Figs. 3 - 5).

The basal age of PS1394 is not older than early stage 5 and that of PS1431 not older than early stage 7. Interpretation of the oxygen isotopic data is difficult due to diagenetic alteration of the foraminifers in areas with higher sedimentation rates and sediments with minimum carbonate content (see discussion below). However, correlation with PS1388 and PS1387 is possible using the cyclic variations of carbonate and opal content.

Sediments older than isotope stage 7 are found only in Cores PS1387 and PS1388. Correlation of the middle part of our record with the global isotope stratigraphy becomes ambiguous. Between stage boundaries 7/8 (6 m in PS1388, 4 m in PS1387) and 14/15 (7 m in PS1388, 5 m in PS1387) the record is incomplete or biased due to missing and displaced sediments. A drastic drop in water content, together with a sharp increase in the abundance of smectite and magnetization intensities in Cores PS1388 and PS1387, indicates a major mass flow event. This can also be correlated with other cores in the area. A missing sedimentary sequence of up to 3 m can be calculated from the gradient of decreasing water content at the hiatus. The missing section corresponds to a time span of about 200 kyr based on a mean sedimentation rate of 1.6 cm/kyr (Mackensen *et al.*, 1989).

The magnetostratigraphic data serve as a framework for the stratigraphic interpretation of the oxygen isotope record of the lower core sections (Fig. 7). The paleomagnetic record and several sedimentological and paleontological variations, which are probably associated with the mid-Pleistocene change in climate regime around 600 ka (Williams *et al.*, 1988), corroborate the identification of the characteristically steep and highly positive $\delta^{18}O$ excursion of event 16.2 (Prell *et al.*, 1986; Williams *et al.*, 1988) at 8 m in PS1388 and 5.70 m in PS1387 (Fig. 7) on the basis of the following observations: (1) foraminiferal carbonate is present continuously below core depths of approximately 7 m (PS1388) and 5 m (PS1387), (2) biogenic opal is absent below this level, (3) cycles and amplitudes of the curve, obtained by plotting the number of benthic foraminiferal specimens per sample versus depth in core, change at 7 m in PS1388, and (4) principal component analysis of benthic foraminiferal data reflect a change in assemblage composition (Mackensen *et al.*, 1989).

Between isotopic event 16.2 and stage 19, stage 17 can be recognized in the benthic record. Stages 15 and 13 following stage 16 up to the hiatus can be identified in the benthic as well as in the planktic record. With the base of the hiatus coinciding with stage boundary 12/13 (486 ka, Williams *et al.*, 1988) and the upper limit with the transition from stage 7 to 8 (285 ka), the hiatus includes stages 9 to 12 with a time span of about 200 kyr. This corresponds well with the value calculated from the abrupt decrease in water content at the corresponding core depth.

Between stage 19, the top of which corresponds to the Brunhes/Matuyama boundary, and stage 27 at the top of the Jaramillo Event, isotopic stages 21, 23 and 25 can be identified (Fig. 7). Also in this part of the record the isotopic as well as the carbonate curves from both cores can easily be correlated. Core catcher ages of both cores are around 920

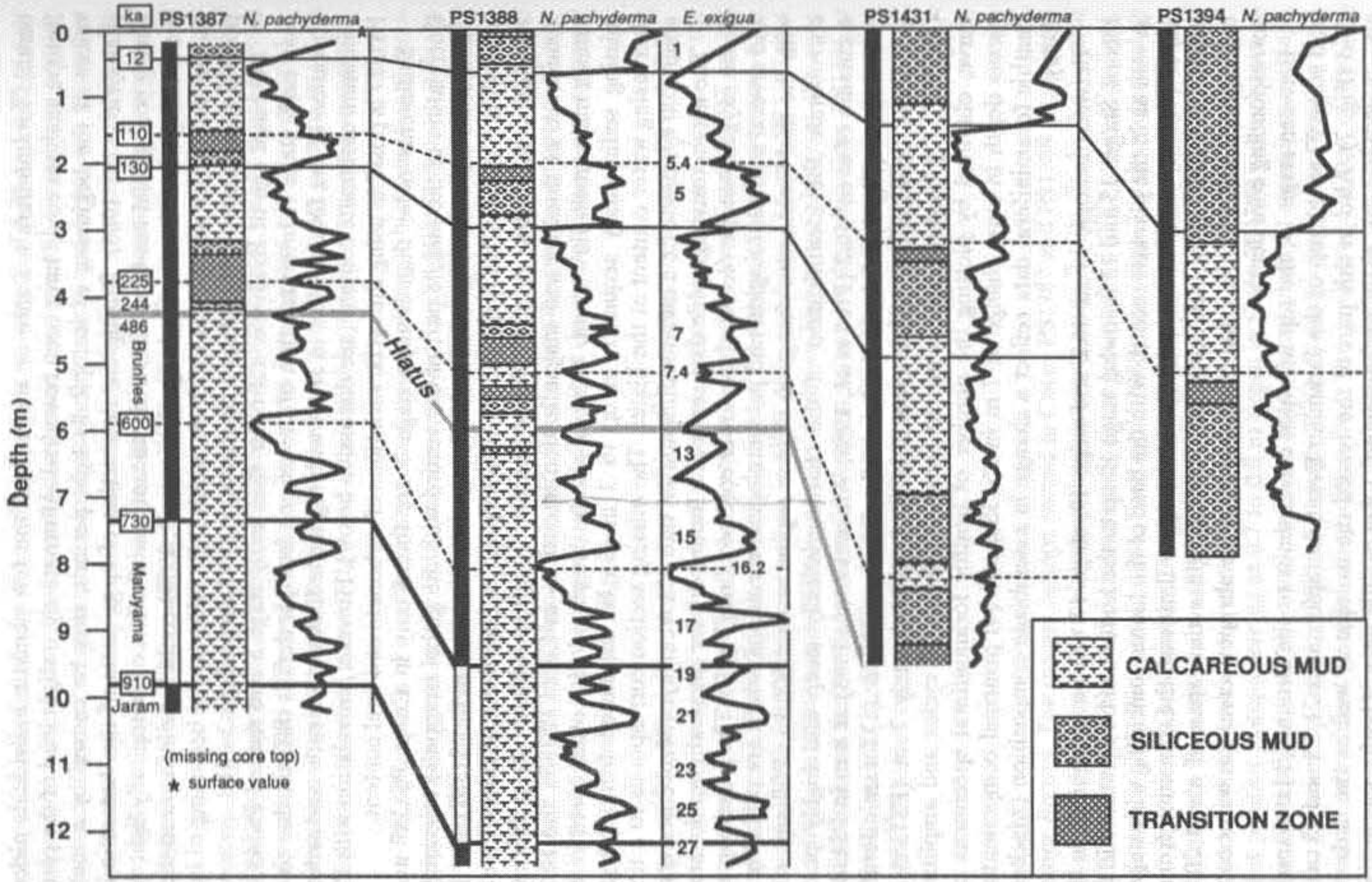

Figure 7 Inter-core correlation by means of oxygen isotopes, paleomagnetic data, and lithology. The isotope curves coincide well throughout. The hiatus shown in PS1387 and PS1388, caused by a mass flow event, represents a time span of about 200 kyr, corresponding to the isotope stages 8 to 12. Ages of stage boundaries are from Martinson *et al.* (1987).

ka. The base of PS1387 might be slightly older.

Carbon Isotope Record. The downcore carbon isotope distribution of both *N. pachyderma* and *E. exigua* correlates with the oxygen isotope curves (Figs. 3 - 6). The carbon isotopes follow the glacial/interglacial cycles by shifting to ^{13}C-enriched values during most interglacials and to ^{13}C-depleted values during glacial periods. This fluctuation, which amounts to 0.9°/oo, can partly be explained by global changes in the isotopic composition of the total carbon dioxide in the ocean which is dependent on the higher amount of ^{13}C-depleted organic carbon stored in the terrestrial biomass (Shackleton, 1977). Since these changes are responsible for only a minor part of the shift, the glacial/interglacial $\delta^{13}C$-cycles are probably a function of mainly global changes in ocean productivity, nutrient and bottom water mass distribution (Curry *et al.*, 1988; Duplessy *et al.*, 1988; Sarnthein *et al.*, 1988, Sarnthein & Winn, in press).

The difference in $\delta^{13}C$ values of the benthic foraminifer *E. exigua* of about 0.9°/oo between the last glacial and late Holocene time falls in the same range as calculated by Curry *et al.* (1988) for *C. wuellerstorfi* from South Atlantic cores (26°- 44° S). Compared with *C. wuellerstorfi*, the peak values of *E. exigua* are about 0.5°/oo lower, indicating that this species does not incorporate the carbon isotopes in equilibrium with the dissolved bicarbonate of the ambient seawater (Woodruff *et al.*, 1980).

As *E. exigua* prefers an epibenthic microhabitat, as does *C. wuellerstorfi* (Gooday, 1988), the difference in equilibrium is assumed to be caused by a constant vital effect, rather than by varying amounts of decaying organic carbon in the uppermost layers of the sediment (Zahn *et al.*, 1986). An almost parallel variation of carbon isotope composition for benthic and planktic foraminifers is typical for regions where bottom water formation occurs (Duplessy *et al.*, 1988). A reconstruction of the geographic distribution of $\delta^{13}C$ in benthic foraminifers in the Atlantic Ocean indicates that Antarctic bottom water (AABW) has been continually forming during the last 150 kyr (Duplessy *et al.*, 1988). Following their arguments, the parallelism of the benthic and planktic $\delta^{13}C$ changes, which exists throughout the sequence of all cores, indicates that the AABW formation never ceased during the last 1 million yr.

Diagenetic Alteration of the Isotope Signal. Minimum values of 3.6 to 3.7°/oo for $\delta^{18}O$ are found in the surface sediments of all cores. In general, values lighter than 4°/oo are typical of the Holocene sedimentary sequence. Samples with pre-Holocene ages are in the range 4 to 5°/oo, and do not attain lighter values even in those sediments formed during the peak warm times of stages 5 or 7. For example, the difference between the $\delta^{18}O$ values of event 5.5 and the 'nominal value' for the Holocene Epoch increases from about 0.3°/oo in PS1387 to 0.9°/oo in PS1431 and PS1394 as the continent is approached. The 150 cm thick Holocene sequence with typical $\delta^{18}O$ values in PS1431, the sharp gradient of the oxygen isotope curve at stage boundary 1/2, and the clearly defined stage 2, excludes bioturbation as a mechanism for bias of the isotope signal. The continously decreasing $\delta^{13}C$ values with sediment depth (*e.g.* PS1431), and the increasing difference in the $\delta^{18}O$ values between stages 5.5 and 1 from Core PS1387 to PS1431 in association with increasing sedimentation rates and decreasing carbonate content, calls for diagenetic alteration of the foraminiferal tests.

In general, each climatic cycle of the oxygen isotopic curve can be linked with three sedimentary facies types (*e.g.* Fig. 4). Facies (a) represents glacial deposits with heavy oxygen isotope ratios (4.8 - 5.0°/oo) and a low carbonate content, sedimentary Facies (b) has a higher foraminiferal carbonate content with light isotopic values (4.2 - 4.4°/oo), and the opal bearing Facies (c) is characterized by heavier isotope ratios (4.6 - 4.8°/oo) and an almost total absence of carbonate.

The correlation of the oxygen isotope ratios of Core PS1431 with Cores PS1387 and PS1388 shows that in sections with a significant amount of carbonate the variations are similar (Fig. 7; Fig. 4a, b). Comparison with the Holocene sedimentary facies shows that Facies (c) represents the peak warm interval of about 20 kyr during an interglacial. The carbonate content in Facies (c) is limited to only a few planktic foraminifers - some of the samples only contained 3 to 6 tests per 100 cm^3. Diagenetic dissolution within the sedimentary column may have altered the isotopic composition of the tests towards heavier values by selectively removing the isotopically lighter carbonate (Savin & Douglas, 1973; Berger & Killingley, 1977). This effect increases close to the continent because of a higher dilution by terrigeneous material from the shelf. Core PS1394 shows an increase in the oxygen isotopic composition from 3.5 to 4.0°/oo, already in 20 cm depth within sediment Facies (c), where sedimentation rates are highest during Holocene time.

The concurrent shift of the carbon isotope data to lighter values (Fig. 4c) may be explained by sulphate reduction. Low $\delta^{13}C$ values are characteristic of carbonate, formed under reducing conditions during bacterial consumption of organic matter (Claypool & Kaplan, 1974). Sulfate reduction processes could have occurred more intensively in the biosiliceous Facies (c) where high organic carbon contents coincided with low carbonate content. Additionally, bicarbonate which is heavy with respect to $\delta^{18}O$ may have been formed by this process (Belanger *et al.*, 1981). However, this hypothesis requires diagenetic overgrowth on the foraminiferal tests or exchange by diffusion.

Cyclicity of Primary Production. In the discussion about variations of CO_2 in the atmosphere during the Pleistocene climatic cycles, the Southern Ocean is supposed to have been a sink for CO_2 during glacial times due to high productivity (Keir, 1988), processes which were caused by higher solar radiation and reduced circulation of surface waters (Sundquist & Broecker, 1985). On the other hand, a reduced consumption of nutrients is discussed because of the extensive sea ice coverage during glaciations (Mix & Fairbanks, 1985). Estimates of the sea ice distribution during the last glacial maximum show continuous ice coverage south of 60° S (Hays, 1978; Cooke & Hays, 1982). During these times light was the limiting factor for productivity in the Southern Ocean. Thus, in terms of global glacial productivity, the ocean around Antarctica has been called the 'loser' compared with low and mid latitudes (Sarnthein & Winn, in press). The glacial sediments with little or no biogenic components, reduced bioturbation, low IRD-content and mainly consisting of mud, support the hypotheses of strongly reduced productivity and sedimentation rates during glaciations (Grobe, 1986).

On evaluating the sedimentological data concerning biogenic productivity, particularly in the uppermost part of the cores of Holocene age, and during isotopic events 5.5 and 7.5, we find that there is additional evidence to suggest higher primary productivity during peak warm times which began with the transitions from glacial to interglacial periods and lasted for about 15 to 20 kyr. This time span corresponds to the duration of the peak warm times of stages 7, 5 and the Holocene Epoch. The preservation of opaline skeletal elements from planktic (radiolarians, diatoms) as well as of benthic organisms (sponges), provides evidence of a higher productivity compared to the remaining interglacial and glacial times.

Sediments with a significant amount of silica will be passed on only if production exceeds solution (Broecker & Peng, 1982). Normally, the biogenic opal, formed in the euphotic zone, will be remineralized during the settling process through the water column (Calvert, 1968; Lisitzin, 1972).

In the siliceous sediment facies carbonate is rare, because the increasing flux of organic carbon to the ocean floor during peak warm times raises the CCD. In this facies the organic carbon content is up to 0.4% higher than in the remaining parts of the cores. The siliceous facies alternates with sediments with a significant carbonate content. The amount of $CaCO_3$ depends on the productivity, the level of the CCD and the supply of the more alkaline NADW during interglacials. The carbonate content increases when the opal content decreases and vice versa. Analyses of recent undisturbed surface samples, containing planktic foraminifers as well as radiolarians, show that the Holocene change from a predominantly siliceous facies to a calcareous facies has just begun.

During periods of high productivity, a large amount of Si is buried in the sediment and taken out of the silica cycle. Consequently, the decreasing silica content of the ocean may also lead to a decrease in the production of siliceous organisms after peak warm times. Of course, siliceous organisms will remain part of the plankton, but the reduced flux to the ocean floor is not sufficient to preserve silica in the sediment.

Opal-bearing sequences only occurred during the peak warm phases of the last interglacial stages 1, 5, 7 and possibly 11. We suggest that, only after long and pronounced glacial periods with strongly reduced primary productivity in high latitudes, Antarctic ocean waters have been provided with sufficient dissolved nutrients to trigger extensive blooms of siliceous plankton. Therefore, the lack of silica below stage 7 (11) may be associated with the change in frequencies of the dominant climatic cycles of 40 to 100 kyr at around 600 ka (Prell *et al.*, 1986; Williams *et al.*, 1988).

Additional evidence for the cyclical variations in productivity is also given from sediment cores in the southern part of the Weddell Sea off the Filchner Ice-Shelf (Melles, pers. comm.) as well as in the northwestern Weddell Sea off the South Orkney Islands (Brehme, pers.comm.). The sedimentological parameters there can be correlated in detail with those of the investigated area on the basis of their biogenic content. These results show that the variations in productivity are not only a local effect, but occur at least throughout the Weddell Sea.

Mechanisms controlling the cyclical variations in opal production can be explained by the interactions of water masses, sea ice coverage and deep convection during the climatic cycles. The North Atlantic Deep Water (NADW) is supposed to be one source for nutrients, which is conveyed during peak warm times from the northern high latitudes to the Southern Ocean. The Norwegian Sea has been the dominant source area for NADW only during full interglacial conditions as during the Holocene and isotopic event 5.5 (Duplessy *et al.*, 1988). Additionally, the varying extent of the sea ice during warm and cold periods may have controlled the cyclical variations in production and preservation of opal. During the transition from a glacial to an interglacial, when deep convection restarts in response to katabatic winds and the decrease of sea ice, more nutrients are available in surface waters which are then slowly used up.

Sedimentation Rates. The high productivity during the first 15 kyr of interglacial stages, documented in the sediments by a significant amount of opal, provides, in addition to the isotopic data, an important tool for inter-core correlation and calculation of sedimentation

Table 2 Stratigraphic fixed points used for calculating sedimentation rates in Cores PS1431 and PS1388. Ages of isotopic events are from Martinson *et al.* (1987).

Isotopic Event	Age (ka)	PS1431 Depth (m)	PS1388 Depth (m)
2.0	12.05		0.65
2.2	17.85	1.50	0.80
5.0	73.91	2.25	1.25
5.1	79.25		1.40
5.2	90.95		1.60
5.3	99.38		1.80
5.4	110.79	3.20	2.00
6.2	135.10	5.00	3.10
7.0	189.61	6.20	3.80
7.4	224.89	8.10	5.10
8.0	244.18	9.40	5.85

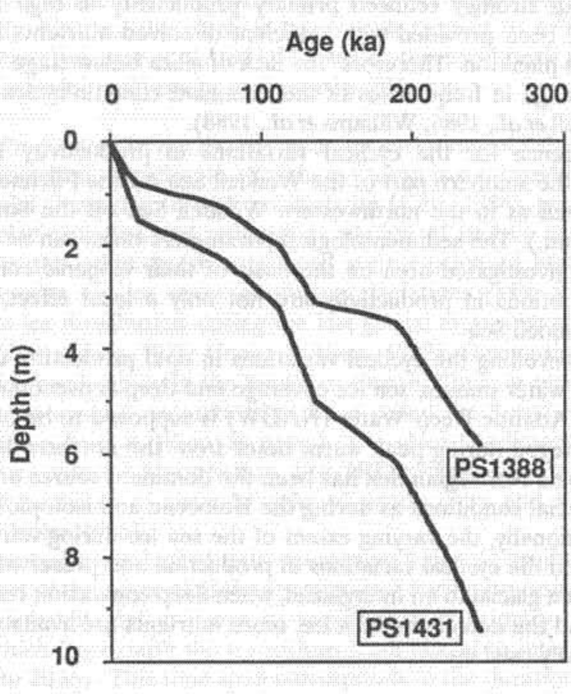

Figure 8 Age - depth plot of Cores PS1431 and PS1388. High sedimentation rates at peak warm times decrease with the climatic deterioration during the transition to glacial times. The mean sedimentation rate increases approaching the continent.

rates (Table 1, Fig. 8). The 'silica event' in the sediment is short, its boundaries are well-defined and supposedly synchronous at all locations along the continental margin. The first significant occurrence of opal in each sedimentological cycle can be defined as the beginning of an interglacial stage. The $\delta^{18}O$ values lead the distribution of opal. After reduction of the ice-shelf (meltwater spike) and the pack-ice coverage, the primary production increases (silica event). The decreasing lag between the meltwater spike and the silica event with increasing distance from the shelf, as documented in the cores, can be explained by the decreasing sedimentation rates.

Stratigraphic fixed points (Table 2, Martinson *et al.*, 1987) have been used for age - depth plots (Fig. 8) and to calculate sedimentation rates during the last 300 kyr. The mean sedimentation rates during one climatic cycle decrease from 5.2 to 1.3 cm/kyr down the continental slope (Fig. 9). The very high sedimentation rate during the last 12 kyr (approximately 4 - 25 cm/kyr depending on core location) is due to the increased input of terrigenous material at the beginning of interglacials, when the calving of glaciers and ice-shelves leads to a drastically increased supply of ice-rafted debris. This indicates principally a continuous change from a high sedimentation rate during deglaciation and interglacial maxima, to moderate and low rates during glaciations (0.6 - 3.6 cm/kyr) and agrees with earlier published data (Fütterer *et al.*, 1988; Zahn *et al.*, 1985). Sedimentation rates are highest at PS1394 because of the high input of terrigeneous material close to the shelf.

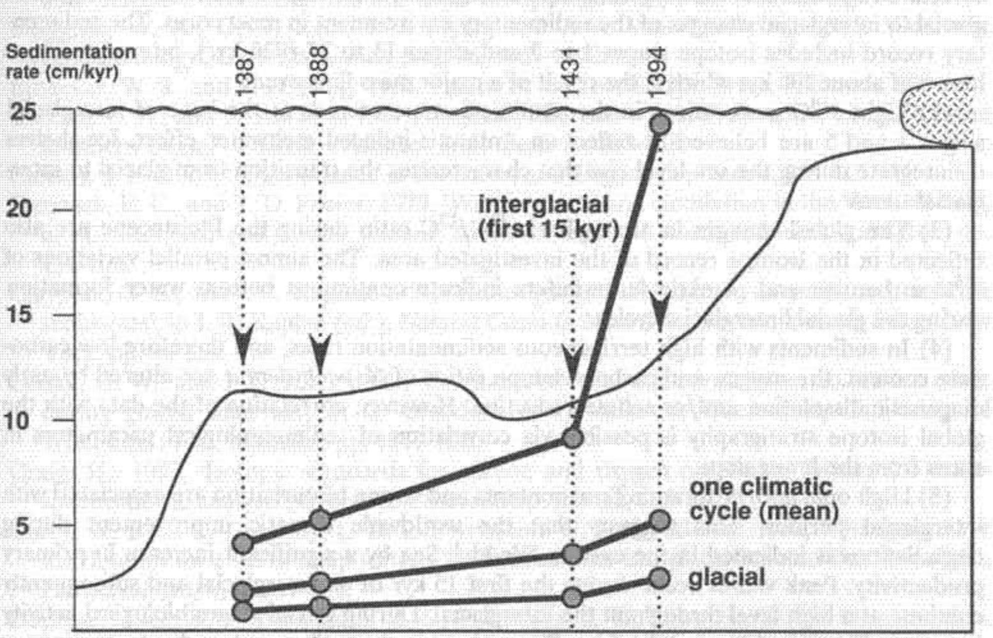

Figure 9 Sedimentation rates plotted according to core locations. Sedimentation rates generally decrease with increasing distance from the continent and during a climatic cycle from interglacial to glacial times (Table 1). They are highest close to the shelf at the beginning of interglacials due to a high input of ice-rafted debris.

During glaciations, which coincide with low sea-level, the grounding line of the ice-shelf occurs at the continental shelf edge. Consequently, small advances of the ice-shelf will have enabled transport of reworked and unsorted sediments across the continental shelf edge and down the slope. During the glacial maximum, gravity flow transport has been the dominant sedimentation process on the upper terrace, because of the morphology of the continental margin in this area with its steep flank (Fig. 9).

There is evidence from grain size analysis that during interglacial stages sedimentation rates on the lower part of the continental slope have been mainly controlled by the production and solution of planktic foraminifers, which reach 20% of the bulk dry sediment, rather than by current transportation and the degree of ice-rafting (Grobe, 1986). The high interglacial sedimentation rates are reduced to 1 cm/kyr during glacials due to decreasing biogenic productivity and iceberg transportation. Sedimentation rates generally decrease both with increasing distance from the continent and with decreasing temperature as a climatic cycle proceeds.

Conclusions

The most important results and main implications can be summarized as follows:

(1) Pleistocene sediments from the Antarctic continental margin have been dated and correlated by means of an oxygen isotope stratigraphy allowing detailed reconstruction of glacial to interglacial changes of the sedimentary environment in most cores. The sedimentary record includes isotope stages 1 to 7 and stages 13 to 27 (920 kyr), interrupted by a hiatus of about 200 kyr which is the result of a major mass flow event.

(2) Light $\delta^{18}O$ peak values in the planktic isotope records at the base of interglacial stages 1 and 5 are believed to reflect an Antarctic-induced meltwater effect. Ice-shelves disintegrate during the sea-level rise that characterizes the transition from glacial to interglacial times.

(3) The global changes in atmospheric $^{12}C/^{13}C$ ratio during the Pleistocene are also reflected in the isotope record of the investigated area. The almost parallel variations of $\delta^{13}C$ in benthic and planktic foraminifers indicate continuous bottom water formation during the glacial/interglacial cycles.

(4) In sediments with high terrigeneous sedimentation rates, and therefore low carbonate content, the oxygen and carbon isotope ratios of *N. pachyderma* are altered by early diagenetic dissolution and/or sulfate reduction. However, correlation of the data with the global isotope stratigraphy is possible via correlation of sedimentological parameters in cores from the lower slope.

(5) High opal and/or foraminiferal contents and strong bioturbation are associated with interglacial periods. This suggests that the worldwide climatic improvement during deglaciations is indicated in the eastern Weddell Sea by a significant increase in primary productivity. Peak values occur during the first 15 kyr of an interglacial and subsequently continue at a high level throughout the interglacial. During glacial stages biological activity was strongly reduced as a result of insufficient light in the surface waters due to continuous sea ice coverage.

(6) Sedimentation rates at the Antarctic continental margin are extremely high close to the shelf at the beginning of an interglacial (25 cm/kyr). This is supposedly an effect of the destruction of large parts of the ice-shelves, when the sea-level rises during deglaciation.

Sedimentation rates decrease with increasing distance from the continent and from interglacial to glacial stages down to a minimum of 0.6 cm/kyr. Mean sedimentation rates during a climatic cycle decrease from 5.2 to 1.3 cm/kyr down the continental slope.

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APPENDIX

Table 3 Stable isotope data of *N. pachyderma*, carbonate content and % radiolarians in Core PS1394

Depth (m)	$\delta^{18}O$ *pachy.* (°/oo) VPDB	$\delta^{13}C$ *pachy.* (°/oo) VPDB	Carbonate (%)	Radiolarians (%)
0.02	3.40	0.54	0.17	0.70
0.08	3.76	0.44	0.00	0.72
0.14	3.86	0.00	0.08	1.21
0.19	-	-	0.00	2.18
0.22	4.17	0.74	0.00	0.32
0.30	-	-	0.00	0.32
0.40	-	-	0.00	0.10
0.50	-	-	0.00	0.13
0.60	-	-	0.00	0.19
0.70	4.38	1.04	0.00	0.26
0.80	-	-	0.00	0.16