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Variability in the Benguela Current upwelling system over the past 70,000 years

C.P. Summerhayes[1], D. Kroon[2], A. Rosell-Melé[3], R.W. Jordan[1,4], H-J. Schrader[5,6], R. Hearn[1,7], J. Villanueva[8], J.O. Grimalt[8] and G. Eglinton[3,9]

[1]*Institute of Oceanographic Sciences Deacon Laboratory, Brook Road, Wormley, Surrey GU8 5UB, UK*
[2]*Department of Geology and Geophysics, University of Edinburgh, West Mains Road, Edinburgh EH9 3JW, UK*
[3]*Organic Geochemistry Unit, School of Chemistry, University of Bristol, Cantocks Close, Bristol BS8 1TS, UK*
[4]*Department of Earth Sciences, Yamagata University, Yamagata 990, Japan*
[5]*Ocean Drilling Program, College Station, Texas 77845-9547, USA*
[6]*Geologisk Institut, Universitetet i Bergen, Allegaten 41, N-5007, Bergen, Norway*
[7]*AEA Technology, Analytical Sciences Centre, Harwell, Didcot, Oxfordshire OX11 0RA, UK*
[8]*Department of Environmental Chemistry (CID-CSIC), Jordi Girona 18-26, 08034, Barcelona, Spain*
[9]*Biogeochemistry Centre, Geology Department, University of Bristol, BS8 1TS, UK*

Abstract – This study was designed to see if the intensity and location of upwelling in the Benguela Current Upwelling System off Namibia changed significantly during the last 70,000 years. Most of the analytical work focused on geochemical, micropalaeontological and stable isotopic analyses of a 6.5m long combined pilot and piston core, PGPC12, from 1017m on the continental slope close to Walvis Bay. The slope sediments are rich in organic matter. Most of it is thought to represent deposition beneath a productive shelf edge upwelling system, but some is supplied by downslope near-bottom flow of material probably resuspended on the outer continental shelf. Temporal changes in upwelling intensity as represented by fluctuations in the accumulation of organic matter do not show the simple 'classical' pattern of less upwelling and lower productivity in interglacials and more upwelling and higher productivity in glacials, but instead show a pattern of higher frequency fluctuations. The broad changes in organic carbon accumulation reach maxima at times when the earth-sun distance was greatest, indicating that this accumulation responded to changes in the precession index; at these times monsoons would have been weakest and Trade Winds strongest. Maximum accumulation of organic matter on the slope occurred in the last interstadial (isotope stage 3), and coincided with coldest sea surface temperatures as recorded by alkenone data (U^{k}_{37}), and by nannofossil assemblages. It is attributed largely to increased productivity *in situ*, rather than the lateral supply of material eroded from older organic rich deposits exposed by the lowering of sealevel at that time. The enhanced productivity is attributed to a strengthening of upwelling-favourable winds in this area in response to the minimal solar insolation typical of this period. Diatoms generally are not abundant in these sediments, so appear to be unreliable indicators of productivity over the continental slope. When sealevel was lowest (isotope stages 2 and 4) organic matter previously deposited on the continental shelf was eroded and dumped on the continental slope; this reworked material constitutes up to 43% of the flux of organic matter to the slope at these times. This process did not affect the slope in stage 3, when sealevel fell by only 50m. The accumulation of terrigenous material was highest in stages 2 and 4. The available data suggest that the terrigenous influx at those times was primarily aeolian. We interpret this to mean that more of the winds then came from the east ('Berg' winds), bringing an influx of aeolian dust from the hinterland; these easterlies were less favourable for upwelling than were the more southerly Trade Winds that dominated during stage 3. Carbonate accumulation was least in stages 2 and 4, largely in response to dissolution induced by CO_2-rich bottom waters.

CONTENTS

1. INTRODUCTION

In this paper we use geochemical, stable isotopic and micropalaeontological studies of a 6.5m long core from the continental slope (Core PGPC12, Fig.1) in an attempt to ascertain the history of upwelling in the Benguela Current Upwelling System off Namibia over the past 70,000 years. The Benguela Current Upwelling System is one of the five great continental margin upwelling systems, with analogues off Peru, California, Northwest Africa and in the northeastern Arabian Sea. While a great deal has been learned about these analogues and their history in recent years, thanks to extensive programmes of piston coring and deep ocean drilling (e.g. RUDDIMAN, SARNTHEIN and SHIPBOARD PARTY, 1989; SUESS, VON HUENE and SHIPBOARD PARTY, 1990), our knowledge about the history of upwelling off Namibia is poor in comparison (SUMMERHAYES, PRELL and EMEIS, 1992). Only the history of the northernmost end of the Benguela Current Upwelling System has been examined in any detail, by drilling and piston coring on the easternmost end of the Walvis Ridge (HAY, SIBUET and SHIPBOARD PARTY, 1984; DIESTER-HAASS, 1987; DIESTER-HAASS, MEYERS and ROTHE, 1986, 1992; OBERHÄNSLI, 1991; SCHMIDT, 1992), and in the southern Angola Basin (BORNHOLD, 1973; JANSEN, 1985; SCHNEIDER, 1991; SCHNEIDER, DAHMKE, KOLLING, MÜLLER, SCHULZ and WEFER, 1992; MEYERS, 1992). South of the Walvis Ridge, the distribution of surface sediments beneath the upwelling system is well known (Fig.2; BREMNER, 1981; CALVERT and PRICE, 1983; SUMMERHAYES, 1983; ROGERS and BREMNER, 1991), but all that is known of the history of upwelling is derived from a single poorly dated piston core from the continental slope at 24°S (PC16; Fig.1) (DIESTER-HAASS, ROTHE and SCHRADER, 1988). Our work on PGPC12 adds significantly to the database, and enables the previous work to be re-evaluated.

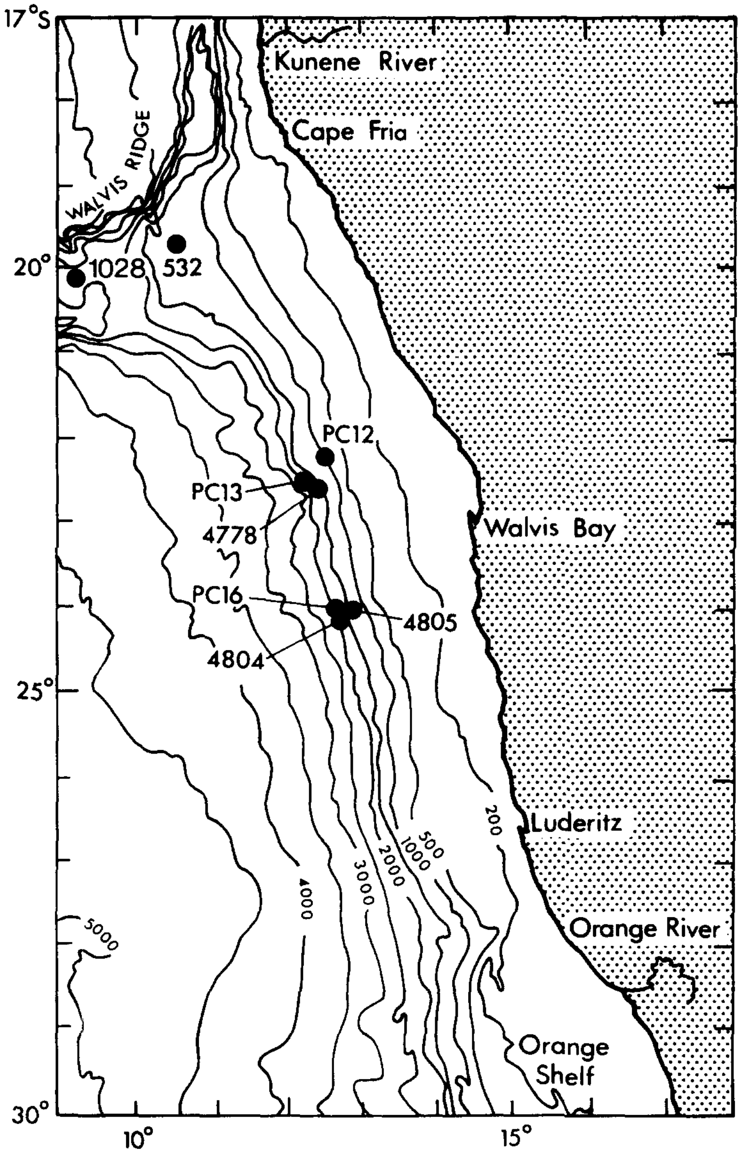

FIG.1. Location map showing sites of cores used in text: PGPC12 (as PC12), PC13, PC16, 1028, DSDP532, and UCT cores 4804, 4805 and 4778. Bathymetry in metres from DINGLE and NELSON (1993).

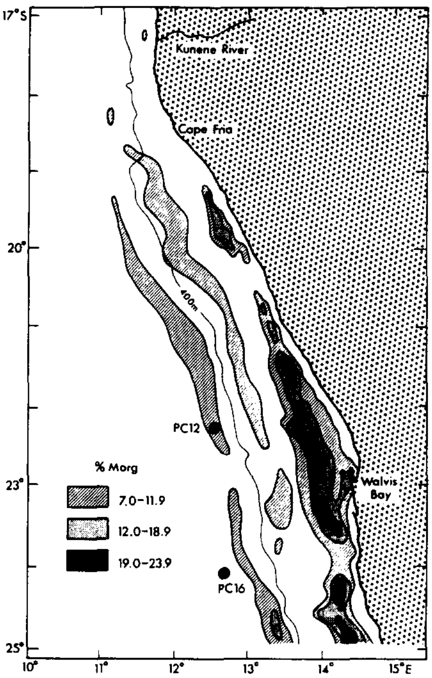

FIG.2. Distribution of organic-rich sediment off Namibia north of Lüderitz, from BREMNER (1981). Morg = total non-skeletal organic matter = C_{org} x 1.8.

The need for a comprehensive study is underlined by the continuing controversy regarding how the data available from DSDP Site 532 on the Walvis Ridge off northern Namibia (Fig.1) should be interpreted. For instance, Diester-Haass and her colleagues (DIESTER-HAASS, 1985; DIESTER-HAASS *et al* 1986, 1988, 1992) interpreted their data to suggest that upwelling was stronger during interglacials whereas DEAN, HAY and SIBUET (1984) and OBERHÄNSLI (1991) suggested that upwelling was stronger during glacials, as it was further north in the Angola Basin (e.g. SCHNEIDER, 1991). The choice of model profoundly affects broader attempts at regional palaeoceanographic reconstructions (e.g. HAY and BROCK, 1992).

Our study of PGPC12 offers a new opportunity to test the competing hypotheses of climate controlled sedimentation in the Benguela Current Upwelling System, and to ascertain how this major system has been affected by fluctuations in climate and sea-level during the Late Quaternary. Our aims were to find out:

(1) if the intensity of upwelling had changed through time over the past glacial-interglacial cycle, and if so when, how, why, and by how much?

(2) if the centre of upwelling had migrated with time, and if so when, where, how and why?

Studies of the history of intensity and distribution of upwelling are important in a general sense in that they may tell us about the location and intensity of the mid-latitude high pressure cells over the ocean. They may also tell us about the biogeochemical history of the oceans, which is dictated largely by coastal processes during times when upwelling is enhanced (BERGER, SMETACEK and WEFER, 1989); low latitude upwelling and its associated pulsating export productivity must be considered a strong additional factor, if not the major agent, in changing ocean and atmospheric chemistry during glacial-interglacial changes (SARNTHEIN, WINN, DUPLESSY and FONTUGNE, 1988).

2. METHODS

In this study we have used piston core PC12 and its accompanying pilot gravity core PG12, both collected on R/V *Chain* cruise 115 on 29 December 1973 from 1017m on the continental slope just north of Walvis Bay (22°16.0'S; 12°32.3'E) (Fig.1; SUMMERHAYES, 1974). PC12 was 620cm long. It consisted mostly of dark olive grey (5Y 3/2) calcareous ooze that was faintly banded and mottled and gave off a slight smell of H_2S; the colour lightened slightly to olive grey (5Y 4/2) at 308-456cm; the bottom 82cm consisted of 'flow in' so was discarded. PG12 was 82cm long. It contained similar but slightly lighter sediment (olive: 5Y 4/4). As explained below there is no overlap between the surface sediments sampled by PG12 and the older sediments sampled by PC12; combined they represent at least 620cm of sediment record. In this paper the combined core is designated PGPC12.

After their collection, PC12 and PG12 were cut into 1.5m long sections, then split lengthways into sampling and archive halves, sampled at 20cm intervals (SUMMERHAYES, BORNHOLD and EMBLEY, 1979), and stored in the core repository at the Woods Hole Oceanographic Institution. In August 1991 the storage tubes were opened for re-sampling; the sediment was still quite damp, but the 1.5m long sections had shrunk by about 8% in length, so that gaps had appeared between adjacent segments of each core section. Each section was reconstituted by eliminating shrinkage cracks, re-measured, and sampled at intervals equivalent to 1cm of the original core so as to enable us if necessary to resolve cyclic signals of high frequency. Most of the samples were obtained from the working half of the core, but where the previous sampling had depleted the working half, the archive half was sampled, using stratigraphic markers to align the sampling and archive halves of the core as precisely as possible. Great care was taken to avoid contaminating the samples with material from the smear zone around the edges of the core.

Different properties of the sediment were analysed at different centres. To establish an accurate stratigraphy, samples 5cm apart were analysed for oxygen and carbon isotopes by Kroon at East Kilbride; where higher resolution was needed the sample spacing was reduced to 1cm. On the basis of the age model derived from these analyses eight samples were chosen for ^{14}C dating at the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at Woods Hole Oceanographic Institution. The most detailed data set, for organic carbon (C_{org}) and $CaCO_3$ at one cm intervals, was obtained at IOS (HEARN, 1992). As an independent check on the stratigraphy, and to provide palaeoceanographic information, samples 10cm apart were analysed for coccoliths by Jordan at IOS, and samples 5cm apart were analysed for diatoms by Schrader at Bergen. Organic geochemical data were used as proxies for past sea surface temperatures, through the $U^{k}_{37'}$ ratio; samples mostly 10cm apart were analysed for this and other biomarkers by Rosell and Eglinton at Bristol and by Villanueva and Grimalt at Barcelona. Clay minerals in selected samples were analysed by X-Ray Diffraction at Woods Hole by Summerhayes.

2.1 Oxygen and carbon isotope stratigraphy

The samples were washed on a sieve with a mesh size of 125μm. Right coiling tests of the planktonic foraminifer *Neogloboquadrina pachyderma* (Ehrenberg) were hand-picked under a binocular microscope from the 125-250μm fraction. Foraminiferal sample weights were typically between 0.05 and 0.1mg, and composed of between 8 and 10 specimens. The picked tests were soaked in methyl alcohol for several minutes, then cleaned in an ultrasonic bath to remove contaminants. After cleaning, excess methyl alcohol was removed with tissue paper and residual alcohol was allowed to evaporate. Tests were then dissolved in orthophosphoric acid (specific gravity 1.9) at 90°C, and the resulting CO_2 gas was analysed using a VG Isogas Prism mass spectrometer, with reference to a standard marble (SM1); the values thus obtained were converted to a PDB standard following the method of CRAIG (1957). Precision for oxygen isotope analyses was 0.08‰, and for carbon isotopes 0.06‰ (using SM1 sample weights of between 0.05 and 1.0mg). Samples were analysed at 5cm intervals except at the top of the piston core where the interval was 1cm.

2.2 Age determination

Radiocarbon ages were measured separately on the organic fraction and on foraminifera extracted from selected samples to provide pairs of ages down core at key depths (Table 1). Most of the picked foraminifera were right or left-coiling *N. pachyderma*. The measurements were made by Dr Ann McNichol. Carbon dioxide generated from the samples by acid hydrolysis or combustion was reacted to form graphite using an Fe/H_2 catalytic reduction. Graphite was pressed into targets which were analysed along with standards and process blanks on an accelerator mass spectrometer. Radiocarbon ages were calculated using 5568 (yrs) as the half-life of radiocarbon.

2.3 Carbon and carbonate stratigraphy

Samples were dried at 100°C for 24 hours, ground by hand, and kept dry in a dessicator until analysed. A Coulometer was used to measure their C_{org} and $CaCO_3$ contents (see HEARN, 1992, for details). Briefly, after splitting each sample into two, one subsample was combusted to convert all of its organic matter and carbonate to CO_2, which was measured as total carbon (TC). The other subsample was treated with acid to convert its $CaCO_3$ fraction to CO_2, which was measured as inorganic carbon (IC): C_{org} was derived by difference (TC-IC). Bulk standards were used to calibrate the instrument.

Every tenth sample was analysed in duplicate, giving analytical precisions of ±1.04% for $CaCO_3$, and ±0.79% for C_{org}. Replicate analyses of bulk standards shows that the method was accurate to within ±1.21%. Table 1A shows inter-laboratory comparisons of $CaCO_3$ data in several samples from PGPC12; agreement is good (within ±10%), but Hearn's data are slightly on the low side. Table 2B shows inter-laboratory comparisons of C_{org} data in samples mostly from PGPC12 or provided by BP. Agreement between Hearn, Bristol and BP is good (on average within ±7%), but Hearn's data are slightly on the high side. There is poorer agreement between the C_{org} data of Hearn and Bremner (Table 2B). This may be explained as follows: the Morgans Wet Chemical Technique (MORGANS, 1956) used by Bremner (personal communication) to analyse for C_{org} involves adding to the sample excess oxidising agent (potassium dichromate) in sulphuric acid, then back-titrating to null point with ferrous sulphate. It is assumed that the sediment contains nothing other than organic matter that could be oxidised by the potassium dichromate, and that the nature and state of the organic matter is constant. In fact, these organic-rich sediments contain pyrite as well as organic matter; also, organic matter may be lost during effervescence when acid is added to the sample. MORGANS (1956) quotes an organic carbon recovery of 75% for the method, so methodological artefacts may explain why Bremner's data are so low compared with Hearn's (Table 2B; Set III).

In assessing these interlaboratory comparisons it needs to be borne in mind that only sets IV, V and VI used subsamples of the same homogenised samples. Sets I, II and III represent analyses carried out at different times on samples from the same intervals, which may explain some of the discrepancy.

TABLE 1. AMS^{14}C dates on core PGPC12, provided by National Ocean Sciences AMS Facility, Woods Hole Oceanographic Institution (Reports 94-021, 94-064 and 94-097). See text for details.

Number[*6]	**Material**	**Depth (cm)**	**Analytical Age[*1]**	**Error (+/-)**	**Calendar Age[*5]**	**Age Difference[*4]**	**Sedimentation Rate (cm 1000y^{-1})**	
							Organic	**Foram**
OS-2327	Organic Matter	10-11	3690	35	3690	+2619		
OS-3456	Foram[*2]		5990	75	6309**			
							9.4	14.2
OS-2756	Organic Matter	45-46	6820	40	7401	+1382		-
OS-3454	Foram[*2]		7880	85	8783**			
							6.0	8.9
OS-2328	Organic Matter	75-76	10700	55	12396	-253		-
OS-3455	Foram[*2]		10500	55	12143**			
							?	?
OS2367	Organic Matter	113-114	16750	100	19834	-3205		
OS2853	Foram[*2]		14100	55	16629**			
							9.8	8.7
OS2757	Organic Matter	142-143	19250	110	22783			
OS2858	Foram[*3]		16850	80	19954**	-2829		
							53.7	8.8
OS-2329	Organic Matter	167-168	19650	85	23248			
OS-2856	Foram[*3]		19250	85	22783**	-465		
							11.7	9.7
OS-2330	Organic Matter	217-218	23400	110	27516			
OS-2857	Foram[*3]		23800	120	27961**	+445		
							15.7	9.8
OS-2331	Organic Matter	267-268	26300	140	30704			
OS-2855	Foram[*3]		28500	190	33056**	+2200		

*1 no reservoir corrections or calibration to calendar years
*2 *N. pachyderma* right coiling
*3 *N. pachyderma* mainly left coiling
*4 foram age younger (-) or older (+) than organic matter age
*5 ages converted to calendar years using the method of BARD (1990)
*6 National Ocean Sciences AMS Facility accession number
** ages used for age model

TABLE 2. Interlaboratory comparison of $CaCO_3$ and C_{org} analyses made at the same sample depths (sets I, II, III) or on the same samples (sets IV, V, VI).

2A. Calcium carbonate analyses ($CaCO_3$) (see text for discussion of analytical methods). Samples in Set I were from 20cm intervals down PG12 (4 samples) and PC12 (19 samples); samples in Set II were from 20cm intervals down PC12, and were not the same as those in Set I. Dr J.M. Bremner's analyses (personal communication) were carried out in South Africa using the Karbonate-Bombe method; Summerhayes' analyses were carried out at Woods Hole by LECO (SUMMERHAYES, 1983, his Fig.13).

Set I Analyst	Number of samples	Av.% $CaCO_3$	**Set II** Analyst	Number of samples	Av.% $CaCO_3$
HEARN (1992)	23	45.6	HEARN (1992)	27	42.4
BREMNER (pers.comm.)	23	48.3	SUMMERHAYES (1983)	27	46.2
difference		-2.7%	difference		-3.8%
% difference		5.9	% difference		9.0

2B. Organic carbon analyses (C_{org}) (see text for discussion of analytical methods). Samples in Set III were the same as in Set I; samples in Set IV were also from PGPC12; Set V consisted of 4 samples from PGPC12 and one standard rock sample; Set VI comprised 5 samples provided by BP Research. Dr J.M. BREMNER's analyses (personal communication) were carried out in South Africa using the Morgans Wet Chemical Technique (MORGANS, 1956).

Set III Analyst	Number of samples	Av.% C_{org}	**Set IV** Analyst	Number of samples	Av.% C_{org}
HEARN (1992)	23	6.76	HEARN (1992)	3	6.87
BREMNER	23	4.49	Bristol Univ.	3	6.68
difference		+2.27%	difference		+0.19%
% difference		33.6	% difference		2.7
Set V Analyst	**Number of samples**	**Av.% C_{org}**	**Set VI** Analyst	**Number of samples**	**Av.% C_{org}**
HEARN (1992)	5	5.77	HEARN (1992)	5	5.17
BP Research	5	5.12	BP Research	5	4.86
difference		+0.65%	difference		+0.13%
% difference		11.3	% difference		6.0

2.4 Coccolith stratigraphy

The relative abundance of coccolith species in PC12 was assessed by counting 300 individuals per smear slide at selected intervals downcore. Samples were prepared by using a toothpick to extract a small amount of dried sediment which was then mixed with a drop of distilled water and spread in a series of streaks on the slide. After drying the slide on a hot plate, a drop of Norland Optical Adhesive 61 was added, followed by a cover slip. When the slide had cured in sunlight, coccoliths were identified under polarised light by using the oil immersion lens of a Leitz Orthoplan microscope at a magnification of x1250; they were counted in continuous transects until 300 specimens had been reached.

2.5 Siliceous organisms

'Rice-grain' sized dried samples taken every 5cm down core in PC12 were boiled in a 50:50 solution of concentrated hydrogen peroxide and concentrated HCl for *ca* 30 minutes to remove organic matter and $CaCO_3$ and to disperse the material. Residues were washed 6 times with distilled water using a centrifuge at 3000rpm, then transferred to 5ml sample vials to which one drop of 40% formaldehyde was added to prevent fungal growth. Using an automatic pipette, 200μl of well-homogenised sample was placed on a 20x40mm cover slip coated with a dilute solution of 98% distilled water and 2% Kodak photo-flow. The cover slips were dried at room temperature and mounted in Merck Mountex on a heating plate at 120°C. Slides were scanned at 500x magnification using a Zeiss phase contrast microscope. Detailed floral analysis was done using a Zeiss phase contrast oil immersion objective.

Abundances of diatoms (all marine planktonic diatoms, *Azpeitia nodulifer, Chaetoceros* resting spores, and *Delphineis karstenii*), silicoflagellates, radiolaria, sponge spicules, phytoliths, and freshwater diatoms were assessed by scanning on microscopical traverses representing *ca* 1/40th of the slide area and counting all opaline particles as one unit if they represented more than 1/2 of the original intact particle.

2.6 Molecular stratigraphy

Aliquots (*ca* 100mg) of each sample were analysed independently in Bristol and Barcelona. About 100mg of each sample was freeze dried overnight, spiked with an internal standard (hexatriacontane) and treated three times with 1.5ml of (3:1) dichloromethane:methanol to extract the solvent soluble organic matter. The total extract was vacuum-evaporated to dryness. In Barcelona, these extracts were redissolved in 2ml of acetone and total pigments were measured at 666nm by absorption spectrophotometry. An extinction coefficient equal to phaeophorbide-a (E_{666}nm = 70.2cm^{-1}M^{-1}) was used. These extracts were again vacuum concentrated to dryness. Both in Bristol and in Barcelona, all extracts were redissolved with 0.5ml of n-hexane and cleaned through a Pasteur pipette filled with 1g of deactivated silica which was eluted with 6ml of (9:1) hexane:dichloromethane. The unretained fraction was vacuum concentrated to dryness and silylated with bis(trimethylsilyl)tri-fluoroacetamide, redissolved in n-tetradecane and analysed by gas chromatography (GC).

Two different GC instruments were used for analysis: at Bristol, a Varian 3600 equipped with a septum programmable injector and a flame ionization detector (FID), and at Barcelona, a Carlo-Erba Model 4160 equipped with a split/splitless injector and an FID. The first instrument was fitted with a Chrompack CP Sil-5 column (50m long; 0.25mm internal diameter; 0.15μm film thickness). Hydrogen was used as carrier gas. The GC oven was temperature programmed from 150 to 300°C at 6°C min^{-1}, followed by an isothermal period of 35min. The GC injector was temperature programmed from 150 to 300°C at 200°C min^{-1} followed by an isothermal period of 35min. The detector was set at 300°C. The second instrument was fitted with a J&W DB-5 column (30m; 0.25mm; 0.2μm). Hydrogen was used as the carrier gas. Oven temperatures were programmed from 180 to 320°C at 20°C min^{-1}, followed by an isothermal period of 20min. Injection was in the splitless mode (split valve closed for 35secs). Injector and detector temperatures were 300 and 330°C respectively.

The peak areas of the target compounds (long chain alkenones and alkenoates) were integrated, quantified using the internal standard, and identified by coelution with external standards (RECHKA and MAXWELL, 1988) and by mass spectrometric analysis. The $U^{k}_{37'}$ index was calculated using the method of BRASSELL, EGLINTON, MARLOWE, PFLAUMANN and SARNTHEIN (1986), and the sea surface temperature for the euphotic zone derived using the equation $U^{k}_{37'}=0.034T+0.039$ (PRAHL and WAKEHAM, 1987; PRAHL, MUEHLHAUSEN and ZAHNLE, 1988). The error of the analysis was estimated to be $\pm 0.003 U^{k}_{37'}$ units.

The sum of the abundances of the long chain alkenones and alkenoates has been calculated for the samples studied. It appears expressed as AA in the plots and the text. These compounds are known to be indicators of the input of the prymnesiophyte algae to the environment (MARLOWE, BRASSELL, EGLINTON and GREEN, 1984a; MARLOWE, GREEN, NEAL, BRASSELL, EGLINTON and COURSE, 1984b). AA should mainly reflect the relative changes through time in the productivity of the Prymnesiophyceae, as well as changes in the environmental conditions that prevailed during deposition and burial. The error in the measurement does not exceed 4% of the total signal.

The measurement of the absorbance of the organic extract at 660nm is a reflection of the total abundance of chlorophyllic pigments in the samples. As for the AA, the pigment abundances may reflect not only productivity but also conditions of deposition and subsequent diagenesis. The two series of $U^{k}_{37'}$ values obtained in Bristol and Barcelona were compared and the whole analytical procedure was re-examined when deviations lager than 5% were observed. The error of the various measurements does not exceed ±10% of the total signal.

3. RESULTS

For the most part the following presentation of results focuses on those from PGPC12. However, because our study has a regional as well as a local dimension, we also record below selected comparative data from three nearby cores: PC16; 1018; and DSDP532 (see Fig.1 for locations).

3.1 Stratigraphy

AMS^{14}C dates (Table 1) and oxygen isotopes provide the stratigraphic framework for PGPC12 and enabled us to define the age model shown in Fig.3. From Table 1 we used foraminiferal ages as the basis for our age model of PGPC12. We then compared the oxygen isotope record of PGPC12 with the orbitally-tuned, stacked, standard isotope record of MARTINSON, PISIAS, HAYS, IMBRIE, MOORE and SHACKLETON (1987). Precise correlation proved difficult, but the comparison enabled us to pick the boundary between isotope stages 2 and 3 at about 120cm in PC12. In addition, we were able to compare the down-core oxygen isotope profile of PGPC12 with that of a nearby slope core (1711) in which stage 5 was reached (N.B. PRICE, personal communication), which enabled us to identify the boundary between isotope stages 3 and 4 at 450.5cm in PC12. Access to unpublished Sr geochemical profiles from both cores (N.B. PRICE, personal communication) helped us to fix the position of this boundary with confidence. We interpret an abrupt transition from light $\delta^{18}O$ values, averaging 0.7‰, in the pilot core PG12, to heavier $\delta^{18}O$ values of about 2‰ at the top of the piston core PC12 (Fig.3A), as indicating that there is a small amount of section missing between the pilot core and the piston core at the transition from the Holocene (isotope stage 1) to the last glaciation (isotope stage 2). Age and isotope data suggest a gap of some 4000 years (caused by over penetration by a poorly rigged piston core).

Assuming that the rate of sedimentation during that missing period was much the same as it was before and after (*ca* 8cm $1000yrs^{-1}$ based on the ^{14}C data in Table 1), then some 30cm of section may be missing and the top of the piston core PC12 represents a depth in the combined PGPC12 section of 112cm (82cm of pilot core PG12 + 30cm of missing section). Accordingly, to create a realistic depth profile for the combined core, PGPC12, we have inserted a 30cm gap into the depth section at the base of the pilot core (PG12), such that the piston core (PC12) part of PGPC12 begins at 112cm.

An abrupt shift in the carbonate values also suggested that there was some section missing between the base of PG12 (60.9% $CaCO_3$) and the top of PC12 (41.3% $CaCO_3$). In order to see if our choice of 112cm for the top of the low carbonate sequence in PGPC12 was reasonable, we checked the downcore profiles of carbonate in four other cores from the slope near PGPC12 (PC16, 4778, 4804 and 4805 in Fig.1). They also have high carbonate values at the surface dropping to values of less than 60% $CaCO_3$ at depths of between 70 and 280cm (average 125cm) suggesting that our choice of 112cm for the top of PC12 is reasonable.

The coccolith data confirm that the core is young. Figure 3B shows the down-core profile of the two main coccolith species *Emiliania huxleyi* and *Gephyrocapsa muellerae*. As there is no sign near the base of the core of the *Gephyrocapsa aperta* zone, the cored section is likely to be younger than 150ka (cf WEAVER and HINE, in press). Until recently it had been thought that the *E.huxleyi/ G.muellerae* transition occurred at about 73-85ka (THIERSTEIN, GEITZENAUER, MOLFINO and SHACKLETON, 1977; WEAVER and HINE, in press). While this may have been true for open oceanic environments, the transition appears to have occurred later and over a longer time span (25-50ka) in upwelling regions, as at ODP 658, off northwest Africa (JORDAN, ZHAO, EGLINTON and WEAVER, in press), as well as at PGPC12.

3.2 Sedimentation rate

From the age model (Fig.3) we can estimate changes in the rate of sedimentation (S_R) with time (Fig.3D). S_R was highest in the last interstadial (isotope stage 3), dropped in the last glaciation (stage 2), was lowest at the beginning of the Holocene (stage 1) then rose to a level higher than stage 2 in the upper part of stage 1. We have no useful information on stage 4 and have assumed a rate similar to that in stage 3.

Knowing S_R enables us to calculate the rate of accumulation of different components (such as $CaCO_3$) using the following equation:

$$S_{CO3} = S_R.P_s(1\text{-}P).CaCO_3.10^{-4}mg.cm^{-2}y^{1} \qquad (1)$$

where S_{CO3} = rate of accumulation of carbonate; S_R = sedimentation rate; P_s = dry sediment density; P = porosity.

To calculate porosity we used measurements of the water content, made by Summerhayes in 1974 on fresh samples at 20cm intervals down core (Table 3). Since water content was more or less constant within narrow limits over 100cm long sections of the core, the average water content for each section was used to derive the porosity for our 1cm sample intervals. Knowing the $CaCO_3$ content of each sample, its density was calculated by assuming that it was a mixture of pure carbonate (with a density of $2160mg.cm^{-3}$), and pure clay (with a density of $1520mg.cm^{-3}$) (TUCHOLKE and VOGT, 1979).

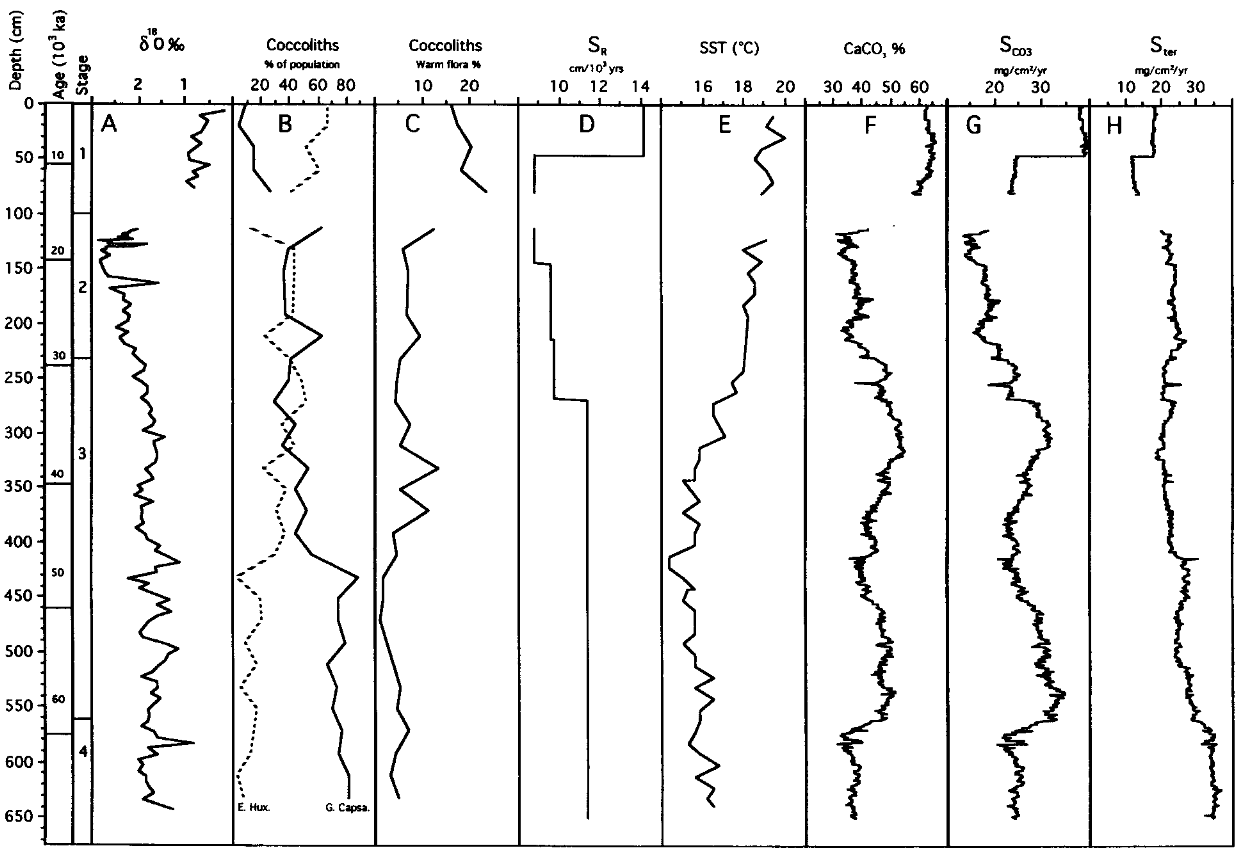

FIG.3. Down-core distribution of selected sediment properties, plotted against depth, age, and oxygen isotope stage. Ages were calculated in calendar years using the method of BARD, HAMELIN, FAIRBANKS and ZINDLER (1990). A: oxygen isotopes; B: the main coccolith species *Emiliania huxleyi* (dashed line), and *Gephyrocapsa muellerae* (solid line); C: warm floral assemblage *Gephyrocapsa oceanica, Florisphaera profunda* and *Umbilicosphaera sibogae*; D: Sedimentation rate in cm.$1000y^{-1}$; E: sea surface temperature (SST) calculated from U^k_{37} ratio; F: % $CaCO_3$; G: rate of $CaCO_3$ accumulation (S_{CO3}); H: rate of accumulation of terrigenous material (S_{ter}), calculating terrigenous material as [100%-($CaCO_3$%+C_{org}%)];

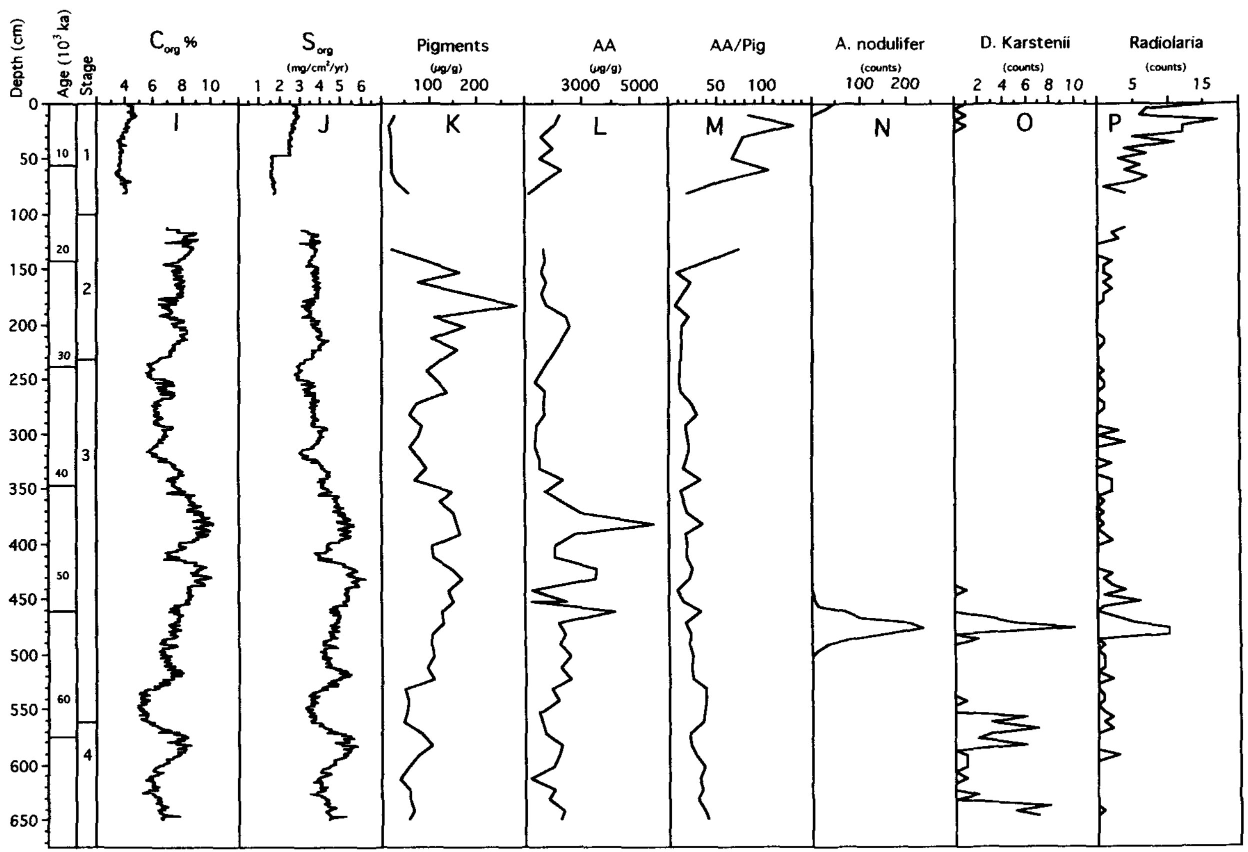

FIG.3. Cont. I: C_{org}; J: rate of accumulation of C_{org}(S_{org}); K: pigment abundance; L: alkenone-alkenoate (AA) abundance; M: AA/pigment ratio; N: *Azpeitia nodulifer* (open ocean) which is almost identical to that of *Chaetoceros* spores (upwelling assemblage) (counts per unit area); O: *Delphineis karstenii* (diatoms of the upwelling assemblage) (counts per unit area); P: radiolaria (counts per unit area).

TABLE 3. Water contents down core PGPC12. Differences between wet and dry weights of samples analysed at 20cm intervals down core, and averaged for selected depth spans.

Depth (cm)	Water Content %	Range %
0-89	65	
112-211	61	60-63
212-311	57	51-61
312-411	62	58-64
412-511	59	55-62
512-611	54	52-57
612-647	54	52-56

3.3 Indications of bioturbation and mixing

Differences within the same samples between the ages obtained on organic matter, representing fine-grained material, and planktonic foraminifera, representing coarse-grained material (see Table 1 for examples) are caused in some instances by bioturbation and in others by the admixture of older materials reworked from the shelf and transported to the site across the shelf edge when sealevel was low. Differential bioturbation, causing a lag in the rate of burial of coarser components (a common phenomenon documented by THOMSON, COOK, ANDERSON, MACKENZIE, HARKNESS and MCCAVE, in press, among others), is evident from the slightly greater age for planktonic foraminifera than organic matter at 10, 45, 217 and 267cm (Table 1). We interpret the reversal of this trend in samples at 75, 113, 142 and 167cm (Table 1) to indicate a supply of older organic matter to the slope in response to reworking on the shelf when sealevel was low in isotope stage 2 and rising at the beginning of stage 1.

Using the differences in age between the AMS^{14}C dates from the organic matter and associated foraminifera (Table 1) we can use the method of WEAVER and THOMSON (1993) to estimate the amount of mixing of *in situ* and reworked organic material in stage 2. We take the organic matter in stage 2 (113cm) to be a mixture of organic matter produced *in situ* (i.e. in local surface waters) (m_1) and transported (m_2) material. The measured (organic matter) age is 19,834 years (t_2), while the expected age (what it would have been had we been dealing only with bioturbation and not reworking) is (16,629-1300) years (t_1). The 1300 is the average by which organic matter age is younger than associated foraminiferan age in the two deepest dated samples in Table 1. We assume that the age of reworked material (t_3) is no older than 39,000 years, based on the observation that given the mixture of *E. huxleyi* and *G. muellerae* at 113cm the sediments reworked into stage 2 are unlikely to be older than those at 322cm (see Fig.3B). We also assume that the coccolith flora over the site and at the source of reworking on the shelf were similar, which accords reasonably well with what is known of the modern distribution of coccoliths on this margin (cf GIRAUDEAU, 1992). Given these conditions then:

$$(m_1 \times e^{-\lambda t1}) + (m_2 \times e^{-\lambda t3}) = (m_1 + m_2) \times e^{-\lambda t2} \qquad (2)$$

Because t_3 (39,000 years) is large, ($m_2 x e^{-\lambda t3}$) tends to zero. Therefore:

$$\frac{m_1}{(m_1 + m_2)} = \frac{e^{-\lambda t2}}{e^{-\lambda t1}} \tag{3}$$

given that $\lambda = \ln 2/5568$ (Libby half life), then:

$$\frac{m_1}{(m_1 + m_2)} = \frac{0.08466}{0.14834} = 57\%$$

Given these constraints, 57% of the material could be *in situ*, and 43% reworked. If the reworked material were younger than 39,000 years (t_3) then its proportion would increase.

3.4 Proxy temperature signals

The alkenone unsaturation index, $U^k_{37'}$, is believed to be a proxy for the local mean annual sea surface temperature (SST) in the euphotic zone (Fig.3E) (BRASSELL *et al*, 1986; EGLINTON, BRADSHAW, ROSELL, SARNTHEIN, PFLAUMANN and TIEDEMANN, 1992). There is good evidence that the index is unaffected by diagenesis, even where most of the sedimentary alkenone content is reduced by oxic decomposition (PRAHL *et al*, 1988). From the base of the core to 422cm (*ca* 50ka) SST shows a gradual cooling from about 17°C to about 15°C, with coldest temperatures late in isotope stage 4 and early stage 3. From 422cm upwards, SSTs rise to about 20°C at the top of the core (Fig.3E). These top core temperatures are similar to present summer SSTs over the site (SHANNON, 1985). The large scale fluctuation of about 5.5°C between the present day and the maximum cooling event is similar to that recorded between glacial and interglacial stages in other upwelling areas (EGLINTON *et al*, 1992). However, it was a surprise to find that maximum cooling occurred in the interstadial (stage 3), rather than during the two most recent glacial maxima (stages 2 and 4); the $U^k_{37'}$ SST profile is much the same further north in the Angola Basin (MÜLLER, SCHNEIDER and RUHLAND, 1994). A similar cold event has been recorded from stage 3 in the Arabian Sea (TEN HAVEN and KROON, 1991).

Independent confirmation that surface waters over the site were coldest in stage 3 comes from micropalaeontological data. For instance, the coccolith flora tends to follow the $U^k_{37'}$ quite closely. This is especially true for the 'warm' water flora (*Gephyrocapsa oceanica, Florisphaera profunda* and *Umbilicosphaera sibogae*) (see Figs 3C and 3E), and for *E. huxleyi* ('warm') and *G. muellerae* ('cold') (see Figs 3B and 3E). *E. huxleyi* tends to be cosmopolitan rather than temperature dependent, but HINE (1990) shows that *G. muellerae* is abundant in glacials, and JORDAN *et al* (in press) show that in the upwelling area off Northwest Africa (ODP Site 658) the ratio of *E. huxleyi* to *G. muellerae* increases sharply from stage 2 to stage 1. Further support for maximum cooling in stage 3 is suggested by the increase in abundance of left-coiling *Neogloboquadrina pachyderma* at that time at sites on the Walvis Ridge (SCHMIDT, 1992; OBERHÄNSLI, 1991), and in the Angola Basin (JANSEN, personal communication).

How may the admixture of reworked organic matter in stage 2 have affected the $U^k_{37'}$ profile? Since the $U^k_{37'}$ data show that SSTs were warm in stage 2, while those in stage 3 (at 322cm) were cool, we might deduce from the mixing data that the actual SST in stage 2 was even warmer than shown, and has been lowered a little by mixing with older (colder SST) material.

Although oxygen isotopes may also give some information about local water temperature, the oxygen isotope signal (Fig.3A) is also an ice-volume signal and probably reflects the isotopic character of subsurface water brought into the area by advection, rather than the SST over site PGPC12.

3.5 Carbonate signal

$CaCO_3$ (Fig.3F) is most abundant in the global 'warm' period (stage 1), least abundant in global 'cool' periods (stages 2 and 4), and intermediate in the interstadial (stage 3), so it is not surprising that its distribution is crudely correlated with that of the oxygen isotopes (Fig.3A); the correlation breaks down below 572cm. The distributions of $CaCO_3$ and rates of carbonate accumulation (S_{CO3}) are somewhat similar (see Figs 3F and 3G). $CaCO_3$ was also more abundant and S_{CO3} was more rapid in stage 1 than in glacial stages nearby on the eastern Walvis Ridge at DSDP Site 532 (DIESTER-HAASS, 1985; SCHMIDT, 1992) and core 1028 (SCHMIDT, 1992) (see Fig.1 for core locations), although there the S_{CO3} peak in stage 3 was missing.

For DSDP532, DIESTER-HAASS (1985) found that more foraminifera were fragmented (i.e. dissolution was more common) where $CaCO_3$ was low (in glacials); she attributed an increase in dissolution at these times to an increase in pore water acidity caused by decomposition of organic material, which, she inferred, accumulated at faster rates in glacials. SCHMIDT (1992) considered it more likely that some of the dissolution in cold stages on the Walvis Ridge could have been caused by the global increase in the CO_2 content of bottom waters in glacials. Although we did not examine PGPC12 samples for indices of dissolution, SUMMERHAYES *et al* (1979) showed that the $CaCO_3$ peaks centred near 297 and 522cm in PGPC12 are associated with abundant sand-sized foraminifera, suggesting to us that dissolution was minimal at those times. An examination of the extent of dissolution of foraminiferal remains had been carried out on core PC16 from the slope nearby (Fig.1) by DIESTER-HAASS *et al* (1988).

Before we could interpret their data we had to revise their stratigraphy of PC16, which had been based on comparison with DSDP Site 532, the stratigraphy of which had been subsequently revised by OBERHÄNSLI (1991). Because the stratigraphic data for stages 1-4 at Site 532 is sparse, we used instead as our stratigraphic control for PC16 the high resolution record from core 1028 on the Walvis Ridge (SCHMIDT, 1992) (Fig.1). Our resulting reinterpretation of the stratigraphy of PC16 is given in Fig.4; it suggests that this 8m core spans oxygen isotope stages 1 through 4 rather than 1 through 3. Our criteria for the revised stratigraphy of PC16 are as follows:

Stage 1 is characterised by high $CaCO_3$, low C_{org}, a low planktonic foraminiferal fragmentation index, and a low ratio of benthic to planktonic foraminifera;

Stage 2 is characterised by the lowest $CaCO_3$, highest C_{org}, abundant fragmentation, and a high benthic/planktonic ratio;

Stage 3 is characterised by moderate $CaCO_3$, moderate C_{org}, moderate fragmentation, and some very low benthic/planktonic ratios;

Stage 4 is characterised by a significant drop in $CaCO_3$, moderate C_{org}, abundant fragmentation, and a high benthic/planktonic ratio.

This new interpretation of PC16 (Fig.4) shows that there is a crude correlation between (i) increased dissolution (as represented by fragmentation of planktonic foraminifera and by the ratio of benthonic to planktonic foraminifera) and (ii) the low carbonate of the main glacial periods (stages 2 and 4).

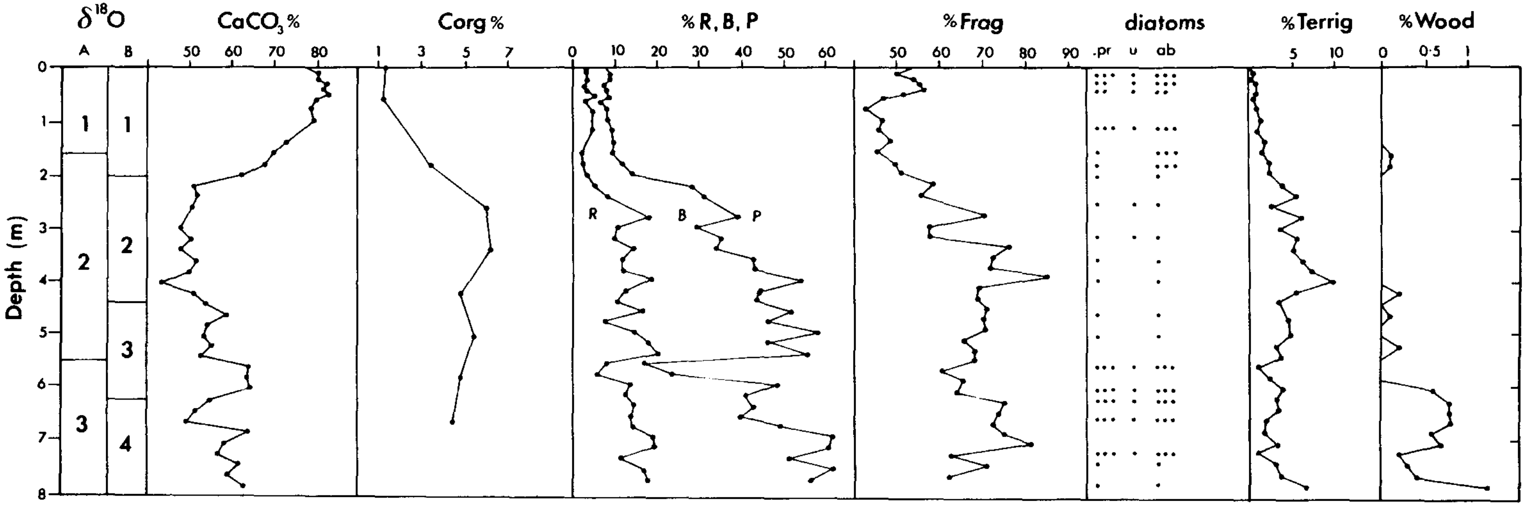

FIG.4. A revised stratigraphy for PC16, showing data from DIESTER-HAASS *et al* (1988). Oxygen isotope stratigraphy at far left is from DIESTER-HAASS *et al* (1988); that on the right is our revision based on a comparison of the stratigraphic features with those at 1028 on the Walvis Ridge (given in SCHMIDT, 1992). C_{org} data from SUMMERHAYES (1983). R = radiolaria; B = benthic foraminifera; P = planktonic foraminifera in the sand fraction. Frag = fragmentation index for planktonic foraminifera calculated as [fragments/(fragments + whole tests)] x 100. Diatom pr = preservation: (•) = poor; (••) = moderate; (•••) = good. Diatom ab = abundance: (•) = few; (••) = common; (•••) = abundant. Diatom u = species indicative of upwelling. Terrig = percent terrigenous material in the coarse silt fraction (40-63μm). Wood = percent wood fibres in the coarse silt fraction.

These various lines of evidence suggest to us that dissolution played an important part in controlling carbonate levels in PGPC12, and was strongest in stages 2 and 4.

Comparing the $CaCO_3$ and $U^k_{37'}$(SST) data (Figs 3F and 3E) we find that the high carbonate of the top 90cm correlates with the warmest SSTs; elsewhere the correlation is poor or non-existent, perhaps reflecting a change in the structure of the water mass between the global cold stages and the others.

3.6 Non-skeletal organic matter signals

Organic carbon (C_{org}) is abundant in PGPC12, ranging from about 4% near the top of the core to as much as 10% at 372-422cm (Fig.3I). These values are equivalent to a range of about 7% to 18% in non-skeletal organic matter. The core-top values compare well with those of modern slope sediments, while the maxima down-core approach the values found in muds from the inner shelf upwelling area (see Figs 2 and 3I).

C_{org} content is more or less inverse to $CaCO_3$, because the more abundant carbonate component dilutes the less abundant organic matter (Figs 3F and 3I). Organic matter is also diluted by terrigenous material. The effects of both diluents (carbonate and terrigenous materials) can be removed by considering the rate of accumulation of organic carbon (S_{org}) (Fig.3J), which proves to have been highest when surface waters were at their coldest in stages 3 and 4, and lowest when surface waters were warmest, in stage 1 (Figs 3E and 3J). Whether all of the fine scale structure of the S_{org} signal reflects changes in surface water temperature remains to be seen; certainly the cold peaks in the SST curve at about 582cm, 412cm, 372cm and 292cm do seem linked with S_{org} peaks that may represent short-lived upwelling events (Figs 3E and 3J). This link suggests that much of the S_{org} profile reflects variations in upwelling and surface water productivity.

Nevertheless, not all of the variations in the S_{org} profile reflect surface water productivity. We have already established above that some 43% of the organic matter in stage 2 consists of older material probably reworked from the continental shelf during low stands of sealevel. Thus the rate of accumulation (S_{org}) attributable to contemporary production then is likely to have been nearer $2mg.cm^{-2}y^{-1}$ than $4mg.cm^{-2}y^{-1}$. What can we tell about the S_{org} and mixing in stage 4, at a time too old to be dated by ^{14}C, but when sealevel was deeper than in stages 1 and 3? To answer this question we examined the relation between S_{org} (Fig.3J) and the rate of accumulation of terrigenous material (S_{ter}; Fig.3H), which is also likely to have increased at low stands of sealevel. Overlaying the S_{ter} and S_{org} curves between 322 and 522cm we find that the ratio of S_{ter}/S_{org} increases in a similar way down core into stage 4 and up core into stage 2. This suggests that the reworking of organic material which characterised stage 2 may also apply to stage 4, and that production was about half that suggested by the S_{org} data during both stages.

Some of the variation in C_{org} down-core could be a function of preservation, since more organic matter is preserved when the rate of sedimentation is high (MÜLLER and SUESS, 1979). Even so, the lack of any clear relationship between either S_{org} (Fig.3J), or C_{org} (Fig.3I), and S_R (Fig.3D) suggests that changes in S_R have not influenced the abundance of organic matter to any great extent within PGPC12. Excepting the mixing concentrated in stages 2 and 4, then, it seems reasonable to suppose that the S_{org} profile reflects mostly surface water productivity. Mixing continued after stage 2 into the base of stage 1, but steadily diminished as sealevel rose (Table 1; see age differences).

The C_{org} signal is similar in several respects to that of other nearby cores. PC16 confirms the pattern of C_{org} increasing from stage 1 to stage 2 (Fig.4). At DSDP site 532, C_{org} rises from 1.6% in stage 1 to 4-5% in stages 2-4 (GARDNER, DEAN and WILSON, 1984; SCHMIDT, 1992). As at

PGPC12, the S_{org} is higher in stage 3 than in stage 2 at DSDP532 (SCHMIDT, 1992). Although C_{org} values are much lower in core 1028 (max. 1.2%; SCHMIDT, 1992), the patterns of distribution of C_{org} and S_{org} in core 1028 are remarkably similar to those in PGPC12, with prominent peaks in stage 4, early stage 3, and stage 2. Particularly intriguing is the rise in C_{org} towards the top of stage 1 at both sites PGPC12 (Figs 3I, 3J) and 1028 (SCHMIDT, 1982).

Previous work on the organic geochemistry of the pilot core part of PGPC12 shows that its organic matter is predominantly marine (GAGOSIAN and FARRINGTON, 1978). These stage 1 sediments have a C/N ratio of 8.3 (GAGOSIAN and FARRINGTON, 1978), which is typical of marine sedimentary organic matter. The ratio is slightly higher than found in particulates over the slope (C/N=7; BISHOP, KETTEN and EDMOND, 1978). The deeper sediments of PGPC12 have a similarly low C/N ratio averaging 9.8 (data from SUMMERHAYES, 1983, his Fig.13). These values are much like those found at similar depths on the slope off the NW African upwelling area (C/N=8.7) by MÜLLER, ERLENKEUSER and VON GRAFENSTEIN (1983). It is difficult to reconcile these ratios with any appreciable contribution from terrestrial sources. The progression of C/N ratios from 7 in surface waters to 9.8 down-core is probably a result of bacterially-mediated diagenesis.

Pigments and alkenone-alkenoate (AA) abundances closely follow C_{org} for most of the core (Figs 3I, 3K and 3L), confirming a marine origin for most of the non-skeletal organic matter. The marked rise in AA abundance at depths <90cm reflects the increasing abundance of *E. huxleyi* (Figs 3B and 3L). The pigment-C_{org} and AA-C_{org} relationships seem to break down between 197cm and 90cm during the last glacial maximum (Figs 3I and 3K and 3L). If this reflected a change in the type of organic mtter we would expect to see a concomitant change in the AA/pigment ratio, but it remains constant throughout most of this interval (Fig.3M).

3.7 Biogenic silica signal

Throughout most of PGPC12 diatoms are either absent, or present in only trace amounts (<10 counted) (Figs 3N and 3O). The exceptions are at four depth zones: 0-21cm at the top of stage 1; 448-498cm (basal stage 3); and 558-583cm and 628-648cm, both in stage 4; diatoms are most abundant at 448-498cm. Almost all of the diatoms are marine, and most are well-preserved. In samples containing abundant *Azpeitia nodulifer* there were some signs of mechanical breakage and chemical dissolution. The lack of sorting and the good preservation suggests that the marine diatoms were not transported to the site by bottom currents.

Three distinct floral assemblages were identified:

1. A marine oceanic, warm water flora, dominated by *Azpeitia nodulifer;*
2. A neritic assemblage dominted by *A. ehrenbergii*;
3. An upwelling assemblage dominated by *Chaetoceros* resting spores and *Delphineis* spp.

In some samples these assemblages were mixed, as seen previously in this area by SCHUETTE and SCHRADER (1981). The oceanic assemblage consisted of *Azpeitia nodulifer, Pseudoeunotia doliolus* and *Thalassiothrix longissima.* It was most prominent at the base of stage 3 (458-498cm), and subordinate in stage 1 (0-21cm). The neritic assemblage consisted of *Actinocyclus ehrenbergii, Thalassionema nitzschiodes, Actinoptychus undulatus* and *A. splendens*, and was most prominent in stages 1 (0-21cm) and 3 (458-498cm). The upwelling assemblage contained abundant *Chaetoceros* resting stages, *Delphineis karstenii, Chaetoceros* setae fragments, and *A. ehrenbergii.* It made a significant minor contribution to the large diatom peak in stage 3 (458-498cm) and dominated the small population of diatoms in stage 4 (558-583cm and 638-648cm). The peak in abundance of the warm water oceanic flora at the base of stage 3 (Fig.3N) is difficult to explain, as it occurred within the period of major cooling (Fig.3E).

Radiolaria (which are a minor constituent in modern slope sediments here; BREMNER, 1981) occurred in all of the samples containing diatoms, and in a few more besides (Fig.3P). They were most abundant at the top of the core. Two of the samples that contained diatoms also contained some silicoflagellates (528 and 578cm). Sponge spicules, mostly monaxic and well-preserved, were present in almost all samples.

Radiolaria, diatoms and organic carbon all increase upwards at the top of the core (Figs 3I, 3N, 3O and 3P). We suggest that the close relationship between these three components reflects trends in productivity. Nevertheless, there is active dissolution of diatom tests today in this area, which may explain the difference in composition between the diatom populations of surface waters and bottom sediments (BREMNER, 1981). Diagenetic dissolution of diatoms may explain the absence of diatoms and radiolaria from much of the core, as has been documented nearby in the Angola Basin by JANSEN and VAN DER GAAST (1988). SCHRADER (1972) concludes that commonly 80-90% of the diatom record is dissolved in the upper part of the sediment column by diagenesis.

The association between diatoms and organic matter at the very top of the core is not maintained down-core, suggesting that there is a decoupling of the production and/or the preservation of organic matter and siliceous organisms in sediments older than stage 1. This decoupling occurs at present within the shelf system, accounting for the decrease in abundance of diatoms seawards across the shelf both in surface waters and in bottom sediments while non-skeletal organic constituents stay abundant (HART and CURRIE, 1960; BREMNER, 1981; ROGERS and BREMNER, 1991).

The diatom abundance picture for PGPC12 is consistent with that at PC16, where well-preserved diatoms are also abundant in stage 1; at the base of stage 3; and in stage 4 (Fig.4). Diatoms also increase slightly in abundance in stage 1 at DSDP532, but as at PGPC12 they are less abundant than radiolaria (DIESTER-HAASS, 1985). Despite these similar abundance patterns, there are important differences between the diatom floras at PGPC12 and PC16. The constituent species of PC16 are more typical of coastal upwelling, suggesting that it was closer to, or more influenced by, coastal upwelling. Even today there is more organic matter in the fine fraction of slope muds at 25°S near PC16 than at 22°S near PGPC12 (BREMNER, 1981). Taken together these data suggest that in the Holocene and during the last glacial, upwelling was more intense nearer to Lüderitz and PC16 than to Walvis Bay and PGPC12 (Fig.1).

Phytoliths derived from terrestrial plants are common trace components in most samples, but freshwater diatoms are rare, with only a single specimen being found in each of three scattered samples, suggesting that there have been no major changes in the strength of offshore winds during the past 70,000 years.

3.8 Terrigenous signal

The rate of accumulation of terrigenous material, S_{ter} (Fig.3H), can be estimated as the reciprocal of the sum of C_{org} and $CaCO_3$, because siliceous skeletal remains are not abundant in this core. S_{ter} was most rapid in stage 4, and was high in stage 2 and for a short time in mid-stage 3. The lowest S_{ter} was in the early Holocene.

The terrigenous material consists mostly of clay, the mineralogy of which is likely to reflect climate change. We had access to analyses of clay minerals made (i) on PC12 (i.e. deeper than 112cm in PGPC12) and (ii) on equivalents of the carbonate-rich stage 1 sediments of PGPC12 from nearby gravity cores 4804, 4805, and 4778 (Fig.1); these data are presented in Table 4. The stage 1 sediments contain on average about 50% illite, 34% montmorillonite, and 16% kaolinite + chlorite. On average their illite content increases downwards from 42% to 56%, while their montmorillonite decreases from about 45% to 31%. In the older sediments (represented by PGPC12 in Table 4) this down-core trend continues: illite increases to *ca* 65%; montmorillonite decreases to *ca* 25%; kaolinite + chlorite drop to *ca* 10%, as against 16% in the gravity core samples.

TABLE 4. Clay mineralogy of slope cores off Walvis Bay; PGPC12 depths are those of the combined core. Determined by X-ray diffraction

Core	depth (cm)	Montmorillonite %	Illite %	Kaolinite + Chlorite %
4804	0-2	47.8	38.8	13.4
	80-82	37.9	49.7	12.4
	100-102	34.0	51.9	14.1
4805	0-2	43.9	42.1	14.0
	20-22	36.8	47.3	15.9
	40-42	27.9	52.9	19.2
	60-62	30.1	54.7	15.2
	80-82	30.1	57.8	12.1
	100-102	33.5	49.4	17.1
4778	0-2	42.7	44.0	13.3
	20-22	33.4	42.4	24.2
	40-42	24.2	54.8	21.0
	60-63	25.9	55.4	18.7
	120-122	30.8	50.5	18.7
	150-152	28.8	59.2	12.0
PGPC12	112-114	22.8	64.3	12.9
	212-214	29.0	60.2	10.8
	292-294	21.4	67.9	10.7
	532-534	27.3	65.2	7.5
	592-594	24.1	66.8	9.1

4. DISCUSSION

4.1 Oceanographic and sedimentary setting

A brief description of the modern oceanographic and sedimentary regime is needed to set the scene for the interpretation of results.

4.1.1. Upwelling associated with the Benguela Current. The Benguela Current is the eastern boundary current of the South Atlantic Subtropical Gyre (Fig.5; PETERSON and STRAMMA, 1991). It bends northwest away from the coast at about 30°S, crossing the Walvis Ridge at about 22-25°S and 5-7.5°E (STRAMMA and PETERSON, 1989; NELSON and HUTCHINGS, 1983). Shorewards of the main stream of the current there is a wide sluggish flow along the continental margin into the Angola Basin (PETERSON and STRAMMA, 1991). Within this sluggish flow the prevailing S and SE winds drive coastal upwelling of cold, nutrient-rich South Atlantic Central Water (SACW) from depths of generally 200-300m (SHANNON, 1985), ranging down to 350-500m (SCHELL, 1970; DINGLE and NELSON, 1993). SACW originates by sinking at the Subtropical Front (Fig.5), otherwise known as the Subtropical Convergence (SCHELL, 1970; NELSON, 1989). Beneath the SACW lies cold, nutrient-rich Antarctic Intermediate Water (AAIW) which originates by sinking at the Polar Front (Fig.5). Along the eastern margin of the SE Atlantic the AAIW occurs at depths of 450-900m, with its core at about 600m (STRAMMA and PETERSON, 1989).

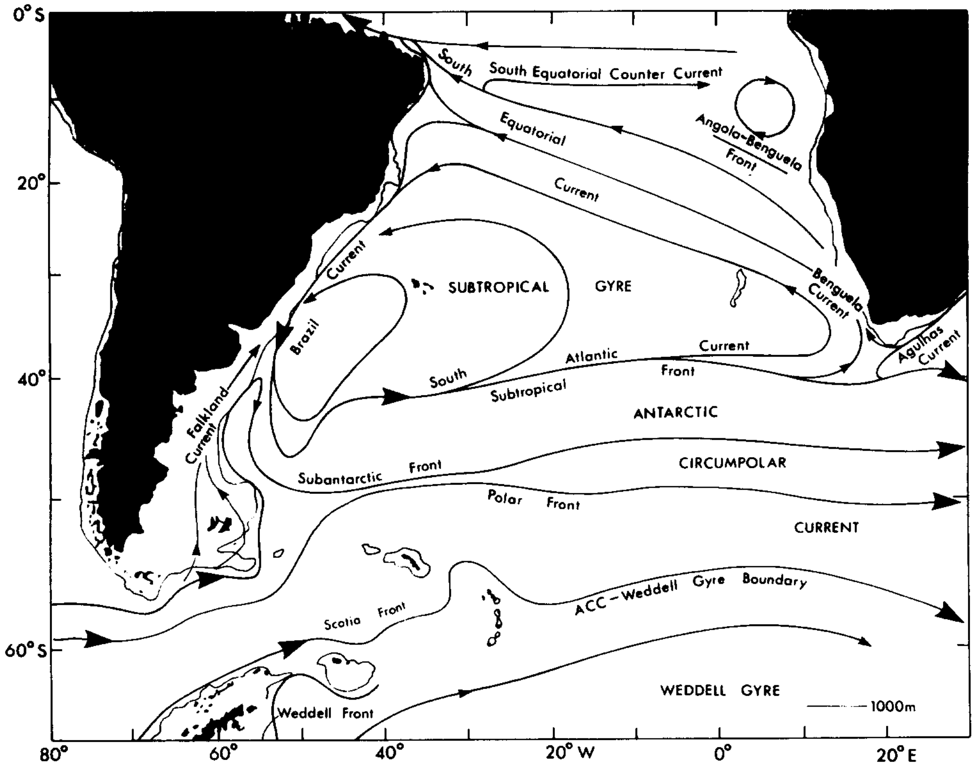

FIG.5. Regional oceanography showing main features like Benguela Current and Angola Front, from PETERSON and STRAMMA (1991).

For much of the year cool upwelled waters cover most of the continental shelf and much of the continental slope (SHANNON, 1985), forming what we refer to here as the Benguela Current Upwelling System. These cool waters move north over the shelf in the Benguela Coastal Current (HAY and BROCK, 1992). Offshore they are separated from the warmer waters of the subtropical Atlantic by a well developed thermal front (SHANNON, 1985). This quasi-stationary boundary is associated with the shelf edge, although it is distorted intermittently by eddies, plumes, filaments and jets which extend west into the SE Atlantic (LUTJEHARMS and STOCKTON, 1987; WALDRON and PROBYN, 1992; SHILLINGTON, HUTCHINGS, PROBYN, WALDRON and PETERSON, 1992). Some of the filaments can be 1000km long and extend up to 1300km west from the coast (LUTJEHARMS, SHILLINGTON and DUNCOMBE RAE, 1991). These plumes and filaments greatly extend the area of high productivity associated with upwelling by rapidly advecting cold, nutrient-rich water from the coastal upwelling sites to the open ocean (LUTJEHARMS *et al*, 1991). However, their impact on productivity seawards of the continental margin is thought likely to be small, given their wide spacing, their slow rate of progression, and their low concentrations of nutrients and organisms (low relative to those over the shelf, though high relative to the surrounding ocean) (SHILLINGTON *et al*, 1992).

To the north the Benguela Current Upwelling System is bounded, at about 16°S, by the Angola-Benguela Front (Fig.5; SHANNON, 1985), which is the convergence between the cold Benguela Current and the Benguela Coastal Current and the warm and saline waters of the Angola Current (PETERSON and STRAMMA, 1991). At the front, both the Benguela Coastal Current and the easternmost branch of the Benguela Current turn west (STRAMMA and PETERSON, 1989. The front migrates seasonally between about 14°S and 16.5°S (SHANNON, AGENBAG and BUYS, 1987). In the late austral summer the front weakens, and warm, saline, Angolan water penetrates south along the coast, occasionally reaching 20°S, or even 22°S in 'Benguela El Niño' events (SHANNON, *et al*, 1987; BOYD, SALAT and MASO, 1987; MANN, 1992).

To the south the coastal upwelling system is bounded by the Agulhas Bank, south of which lies the Subtropical Front (Fig.5). At times this front is perturbed, resulting in an equatorward flow of cold filaments of Subantarctic Surface Water (SHANNON, LUTJEHARMS and AGENBAG, 1989). South of the African coast and north of the Subtropical Front the warm waters of the Agulhas Current flow in from the Indian Ocean and return east (Fig.5), generating eddies of warm water that spin off to the northwest in the Benguela Current (DUNCOMBE RAE, 1991).

The coastal upwelling varies seasonally (SHANNON, 1985). During the austral winter and spring, water cooler than 16°C extends along the entire coast, but in summer and autumn its northward extent is reduced. The seasonal signal is more pronounced off Namibia (i.e. north of the Orange River) than further south, not least because of the penetration of Angolan water south along the north coast of Namibia in late austral summer and early autumn (BOYD, *et al*, 1987). Despite this penetration, there is still upwelling year round at the coast in the northern region (SHANNON and PILLAR, 1986).

Although upwelling is more or less continuous along the Namibian coast, it is particularly strong at 25°S (centre of the Lüderitz upwelling cell), somewhat less so at 22°S (Walvis cell) and at 19°S (Namibia cell), and weakest at 17°S (Cunene cell) (LUTJEHARMS and MEEUWIS, 1987). The Lüderitz cell is the coldest, the most persistent, and extends the furthest offshore. LUTJEHARMS and MEEUWIS (1987) found a strong correlation between intensity of upwelling and the direction and strength of coastal winds. There is also a loose association between the location of upwelling cells and the shape of the seabed, upwelling being more intense where deep water is closest to the coast (SHANNON, 1985).

Along the margin of southwestern Africa most primary production takes place over the shelf (PITCHER and MITCHELL-INNES, 1992). Production in the shelf edge system reaches 2.1-2.9gC $m^{-2}d^{-1}$; although higher than that in surrounding surface water offshore, this is much less than the high values typical of the coastal zone (up to 11g C $m^{-2}d^{-1}$; HUTCHINGS, ARMSTRONG and MITCHELL-INNES, 1986). Off the Namibian coast, north of the Orange River, average annual production in the coastal upwelling region is 2g C $m^{-2}d^{-1}$ (SHANNON and PILLAR, 1986). Within the coastal upwelling system there is a change from high biomass diatom blooms in turbulent, nutrient-rich water near the coast, to a flagellate community (including nannoplankton) at much lower biomass levels in stratified, nutrient-depleted water over the middle and outer shelf (e.g. PITCHER, WALKER, MITCHELL-INNES and MOLONEY, 1991). Phytoplankton, like diatoms and dinoflagellates, tend to be concentrated 10-25km from shore near the most recently upwelled water (SHANNON and PILLAR, 1986), where blooms form quickly in the 7-10 day cycle of upwelling (BRANCH, BARKAI, HOCKEY and HUTCHINGS, 1985). Generally, diatoms bloom at SSTs of 12-15°C, while flagellates dominate the phytoplankton at higher temperatures (MITCHELL-INNES and PITCHER, 1992). Zooplankton populations take longer to grow, so tend to be more abundant further offshore (SHANNON and PILLAR, 1986; BRANCH *et al*, 1985), for instance over the middle shelf (TIMONIN, ARASHKEVICH, PRITS and SEMENOVA, 1992).

4.1.2. Shelf edge upwelling. The occurrence of shelf edge upwelling was first proposed by HART and CURRIE (1960). In the south, near Cape Town, BANG (1971, 1973) and BANG and ANDREWS (1974) showed that there is a well-defined front associated with the shelf edge; this front is separate from the inshore frontal system bounding the coastal upwelling centres over the inner shelf. It extends to some depth over the slope, overlying the 1000m line, so is dynamically more important than the inshore front, and persists whether or not there is any wind-driven coastal upwelling. The front forms the seaward side of a divergence zone some 8-10 miles wide in which upwelling and/ or vertical mixing takes place, and is associated with an increase in the pelagic population, as is evident from high acoustic back-scattering (BANG, 1973), and abundant copepods (VERHEYE, HUTCHINGS, HUGGETT and PAINTING, 1992). Along the shelf edge front is a strong, equatorward frontal jet with velocities of up to 120cm s^{-1} at depths of 100-180m (BANG and ANDREWS, 1974). At times high chlorophyll values are associated with the jet, and this enhanced availability of food (BANG and ANDREWS, 1974; SHANNON, WALTERS and MOSTERT, 1985; HUTCHINGS *et al*, 1986) may support the spawning anchovies whose eggs and larvae are transported north in the jet (SHANNON, HUTCHINGS, BAILEY and SHELTON, 1984).

Recently, BARANGE and PILLAR (1992) and BARANGE, PILLAR and HUTCHINGS (1992) have confirmed that upwelling associated with an equatorward current that may reach velocities of 40cm s^{-1} also takes place at the shelf edge off Namibia (Fig.6) (see also NELSON, 1985, 1989). High productivity for the Namibian shelf break region is confirmed by the high concentrations of organisms (SUMMERHAYES, HOFMEYR and RIOUX, 1974; BARANGE *et al*, 1992; PAGÈS, 1992; OLIVAR, RUBIÉS and SALAT, 1992; TIMONIN *et al*, 1992), associated with high concentrations of nutrients (BAILEY, 1979 in NELSON, 1985) and organic particles (BISHOP *et al*, 1978).

We assume that upwelling at the shelf edge is perennial, since it seems to be associated with the shelf edge and largely independent of coastal wind stress. Average surface water temperatures at sites PGPC12 and PC16 range from 15.5 to 16°C in winter and spring, to about 19°C in summer and about 18°C in autumn (SHANNON, 1985).

Enhanced productivity appears to be a common feature along the shelf edge even outside upwelling areas (e.g. PINGREE and MARDELL, 1981; HOLLIGAN and GROOM, 1986; NEW and PINGREE, 1990). Vertical mixing induced by the interaction of tides with the change in topography at the shelf edge may be one cause (e.g. NELSON, 1985; BANG, 1973). Another factor may be the wind stress curl, the location of the axis of maximum wind speed along the coast (BAKUN and NELSON, 1990; HAY and BROCK, 1992). The elevation of the hinterland just inland from the Namibian coast makes the winds strongest along a narrow corridor parallel to the coastline and 200-300km offshore (KAMSTRA, 1985; BAKUN and NELSON, 1991), more or less over the shelf edge and upper slope. In this offshore zone Ekman transport driven by the wind produces divergence and upwelling at and immediately landward of the corridor's axis, which is balanced by convergence and downwelling on the seaward side of the corridor. This same upwelling process (known as Ekman Pumping) produces open ocean upwelling in the Arabian Sea beneath the axis of the SW Monsoonal jet well offshore from the Arabian coastal upwelling zone (BROCK, 1991).

Diatoms do not dominate the phytoplankton populations in the upper waters in the shelf edge upwelling, which does not bring as cold water to the surface as the upwelling in the coastal zone does. Nevertheless, diatoms abound at the thermocline over the slope (BISHOP *et al*, 1978), and radiolaria also flourish in these shelf edge waters (BISHOP *et al*, 1978). Previously unpublished information shows that while whole diatoms are common in surface waters, fine grained siliceous debris, most probably fragments of diatoms, is common near the bottom over the continental slope (Table 5). The abundance of siliceous fragments in bottom waters (Table 5) suggests that diatoms dissolve on sinking in this environment.

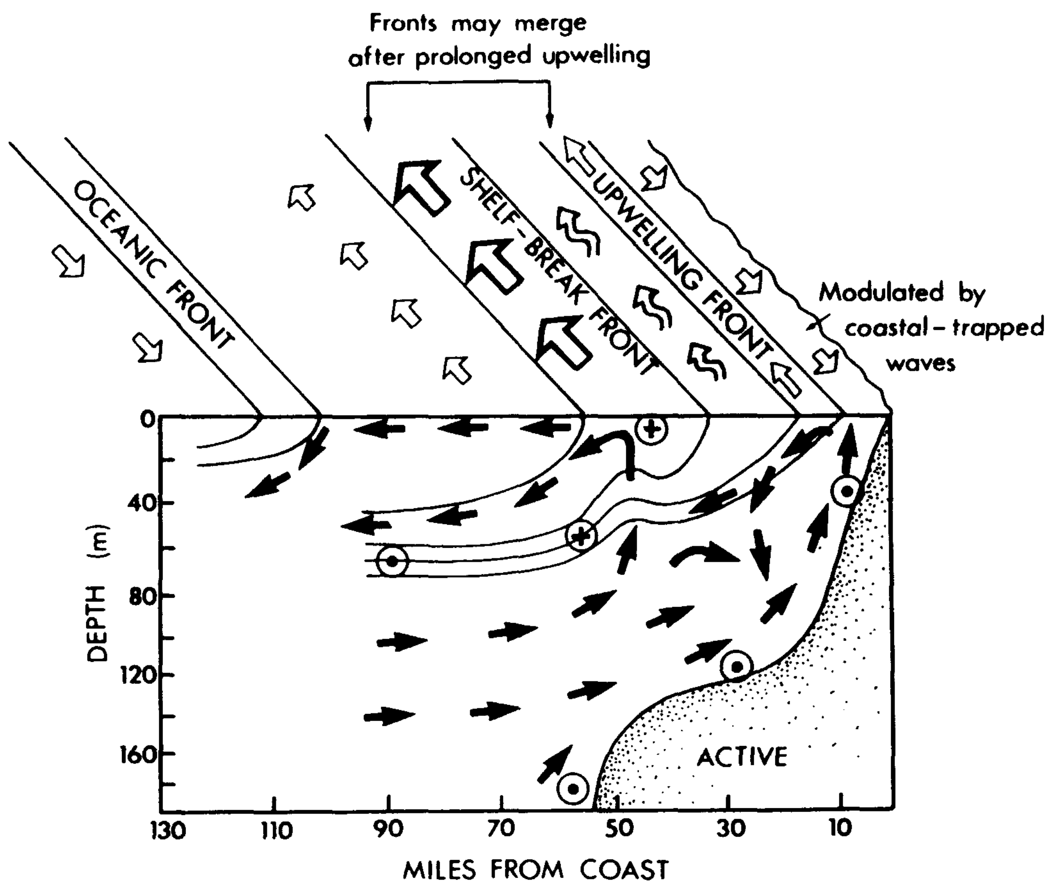

FIG.6. Conceptual three-dimensional model of upwelling system off Namibia (from BARANGE and PILLAR, 1992), showing isotherms, cross shelf flow, surface flow, and directions and locations of subsurface currents (crosses in circles indicate northward motion; dots in circles indicate poleward motion). The upwelling and shelf break fronts are major and persistent; the oceanic front is not.

TABLE 5. Suspended matter concentrations in surface and subsurface water over continental slope off Walvis Bay. Samples collected by 30l Niskin bottle on CHAIN cruise 115 by C.P. SUMMERHAYES and B.D. BORNHOLD and filtered aboard ship. Station 22 collected on December 30, 1973; Station 23 on January 1, 1974.

Station #	Nearest Piston Core	Position	Water Depth (m)	Height of Sample Above Seabed	Suspensate Concentration mg/l	% Combustible (ie organic)	Character of Suspensate
22	PC13	22°36'S 12°08.4'E	2429	Surface Water	0.22	75	common diatoms including large *Coscinodiscus*; mostly whole. Some plant-like fibres. Aggregates rare. Fine, red-brown mineral dust, less abundant than diatoms.
				100m	0.07	100	organic flocs.
				50m	0.04	100	organic flocs.
23	PC16	24°03.1'S 12°41'E	2078	Surface Water	0.19	42	Fairly common diatoms; no large *Coscinodiscus*. Common fine, red-brown mineral dust. Aggregates uncommon.
				100m	0.16	68	Mostly organic fragments; aggregates common.
				50m	0.19	68	Mostly fine grained siliceous skeletal debris; few unbroken diatoms - mostly fragments; fairly common irregular aggregates; few mineral grains.
				15m	0.45	71	As at 50m.

4.1.3. The poleward undercurrent. Beneath the thermocline over the continental shelf and slope there is a poleward undercurrent (HART and CURRIE, 1960; SHANNON, 1985; CHAPMAN and SHANNON, 1985). Poleward flow at about 5km d^{-1} is more or less ubiquitous in near-bottom water across the shelf and slope (DINGLE and NELSON, 1993). DINGLE and NELSON (1993) argue that the bottom waters well up from 350-500m and mvoe landward across the shelf. However, SMITH (in press) argues on theoretical grounds that in upwelling areas the poleward undercurrent drives an offshore Ekman transport near the bottom (see also SMITH, 1983). North of Lüderitz the water in the undercurrent is low in oxygen, which drops below 1ml l^{-1} at the current's core, and which probably originates from an oxygen poor source in the Angola Basin (SHANNON, 1985).

4.1.4. Sediments beneath the Benguela Current System. Beneath the Benguela Current Upwelling System there are three along-shore bands of organic-rich sediments (Fig.2): (a) a near-shore diatomaceous mud belt; (b) a middle shelf zone dominated by fecal pellets; and (c) a slope zone at about 1000m (BREMNER, 1981; ROGERS and BREMNER, 1991). The inshore and middle shelf muds are separated by a zone of coarse sediment centred at depths of about 150m. Seaward of the diatom belt the muds are carbonate rich (ROGERS and BREMNER, 1991; BREMNER, 1981) and dominated by coccoliths and foraminifera (GIRAUDEAU, 1992, 1993). Nowhere is terrigenous material abundant in the sediments except in the coastal strip (BREMNER, 1981). The sources of terrigenous material are aeolian transport of dust from the Namib Desert by easterly 'Berg' winds (SHANNON, 1985; BREMNER, 1981), and a smaller riverine input. Muds from the Kunene and Orange Rivers are moved primarily south by the prevailing bottom currents (ROGERS and

BREMNER, 1991; BREMNER and WILLIS, 1993). As a result, most of the clay fraction of slope sediments off Namibia comes from the Kunene (BREMNER and WILLIS, 1993); closer to shore the clays have a progressively more Namib desert provenance. Between the Orange and the Kunene the rivers are small and only contain enough water to reach the coast about once every 3 years. However, through rare flash floods they can transport large amounts of sediment to coastal waters (BREMNER and WILLIS, 1993).

The narrow belt of organic-rich diatomaceous mud nearshore contains up to about 15% C_{org} (CALVERT and PRICE, 1971, 1983; SUMMERHAYES, 1972, 1983; BREMNER, 1981; ROGERS and BREMNER, 1991). Seawards of these muds the shelf is covered in places with older pelletal phosphorite deposits (SUMMERHAYES, BIRCH, ROGERS and DINGLE, 1973; BREMNER, 1981; ROGERS and BREMNER, 1991). Cores show that there are older diatomaceous muds on the middle shelf, buried beneath a sediment lag deposit (SUMMERHAYES, 1972; DIESTER-HAASS, 1978; ROGERS and BREMNER, 1991), suggesting a fluctuating history of deposition of organic rich sediment in response to changes in climate and/or sea-level on the shelf (ROGERS and BREMNER, 1991).

4.1.5. Transport of organic matter to the continental slope. Despite the potential barrier of the shelf break front (Fig.6) (BARANGE and PILLAR, 1992; BARANGE *et al*, 1992; BANG and ANDREWS, 1974; HUTCHINGS *et al*, 1986), there is a mechanism for the seawards export of organic matter from the shelf in surface waters in the form of the filaments, plumes and eddies that intermittently breach the front (LUTJEHARMS and STOCKTON, 1987; WALDRON and PROBYN, 1992 - their Fig.2; WALDRON, PROBYN, LUTJEHARMS and SHILLINGTON, 1992; SHILLINGTON *et al*, 1992). Nevertheless, SHILLINGTON *et al* (1992) suggest that these processes do not represent a large scale loss of productive coastal water to the open ocean.

Little work has been done on sediment dynamics and particle flux off Namibia. Along a transect off Walvis Bay, BISHOP *et al* (1978) showed that the highest concentrations of particulates were (i) in the water column at the base of the mixed layer (near 50m) over the continental slope, and (ii) close to the seabed over the shelf and shelf edge. Suspensates near the bottom contained less C_{org} than those in the water column, so were assumed to be resuspended bottom sediments. Although BISHOP *et al* (1978) took no samples deeper than 400m (the shelf break), previously unpublished information confirms that suspended matter is abundant near the bottom over the slope (Table 5), making it likely that resuspended material is cascading down the slope. If the near-bottom suspensate over the slope was being transported by bottom currents we would expect to see this reflected in the topography in the form of sediment drifts or other current features, which are not present (SUMMERHAYES *et al*, 1979). The difference in suspensate concentration between stations 22 and 23 (sampled 2 days apart) at 100m and 50m off the bottom over the slope (Table 5) may reflect either (1) the inception of a downslope flow over station 23 after station 22 was sampled, or (2) irregularities in downslope flow, with less at station 22 and more at station 23; the two are around 180km apart, so might not be expected to be identical in any case.

Red, wind-blown mineral dust is common in surface waters over the slope, confirming the role of wind in supplying terrigenous material to slope waters (Table 5). This dust was not obvious near the bottom, probably because these fine particles were bound up in the aggregates common in the near-bottom waters (Table 5). Surface waters over the slope at 50m are rich in particulates comprising organisms and fine-grained fecal pellets, probably of copepods (BISHOP *et al*, 1978). Copepods are abundant here (TIMONIN *et al*, 1992) and provide the means of stripping organic matter from the water column and depositing it rapidly onto the slope below.

Organic enrichment of the muds of the continental slope at depths of about 1000m (Fig.2) is a consistent pattern seen on many of the world's continental margins (PREMUZIC, BENKOVITZ, GAFFNEY and WALSH, 1982). How does the organic matter get there? Walsh argues in favour of significant export of organic matter to the slope from the continental shelf (WALSH, 1989, 1991; WALSH, BISCAYE and CSANADY, 1988); others present evidence against it (HICKELLS, BLACKBURN, BLANTON, EISMA, FOWLER, MANTOURA, MARTENS, MOLL, SCHAREK, SUZUKI and VAULOT, 1991; ROWE, THEROUX, PHOEL, QUINBY, WILKE, KOSCHORECK, WHITLEDGE, FALKOWSKI and FRAY, 1988; FALKOWSKI, FLAGG, ROWE, SMITH, WHITLEDGE and WIRICK, 1988; BRULAND, BIENFANG, BISHOP, EGLINTON, ITTEKKOT, LAMPITT, SARNTHEIN, THIEDE, WALSH and WEFER, 1989). Along the US Atlantic margin 80% of the flux to the seabed is from surface production over the slope, and just 20% is from downslope near-bottom flow of material resuspended from the shelf by tidal action or internal waves or storms according to BISCAYE, ANDERSON and DECK (1988), and PALANQUES and BISCAYE (1992). Calculations suggest that about 10% of shelf production is exported from the shelf as a result of near-bottom suspension (BISCAYE *et al*, 1988). A similar picture emerged from a study of the Rhone shelf (DURRIEU DE MADRAN, NYFFELER and GODET, 1990) suggesting that most seaward transport was focused in canyons and not on the adjacent slope (MONACO, BISCAYE, SOYER, POCKLINGTON and HEUSSNER, 1990). In contrast, JAHNKE and SHIMMIELD (in press) suggest that more than half of the organic matter on the US Atlantic slope may have been supplied by lateral transport. Resuspension of sediments on the shelf is likely to be a result of large tidal phase current surges produced by the internal tide generated at the shelf break, which can increase bottom currents there by as much as 10% (HEATHERSHAW, NEW and EDWARDS, 1987). One effect is to transport fine-grained bottom sediments offshore (down-slope) (HEATHERSHAW *et al*, 1987).

NELSON (1985) presents evidence for mixing near the seabed on the shelf off southern Africa that is attributable to the instability of internal tide waves as they reach the margin, providing a possible mechanism for resuspension. Such studies, and the evidence in Table 5, suggest that the processes of resuspension and cross shelf transport followed by down-slope cascading, which are believed to be common along continental margins (MCCAVE, 1972), probably provide a significant minor amount of the organic matter on the slope off Namibia.

The zone enriched in organic matter on the slope is also enriched in sand-sized fecal pellets of polychaete worms (BREMNER, 1981; 1983); its organic matter is about evenly split between the silt and clay fractions (BREMNER, 1983). In places (e.g. near 25°S; BREMNER, 1981, 1983), the organic content of the silt and clay fractions increases from the outer shelf down the slope to near the 1000m isobath. So even if one source of organic matter on the slope is the export of organic matter from the shelf, there must be another source of organic matter over the slope, probably production resulting from shelf edge upwelling.

4.2 Interpretation of PGPC12 data

Comparing PGPC12, PC16, 1028, and Site 532 we find certain basic similarities. Stage 1 was warm. The sediments that accumulated then are rich in carbonate but poor in organic matter; they contain noticeable minor amounts of radiolaria, but only minor or trace amounts of diatoms. In contrast, taking stages 2, 3 and 4 together as representing glacial conditions, surface waters were colder than in stage 1. The sediments that accumulated then are poorer in carbonate, radiolaria and diatoms, but richer in organic matter.

The patterns of accumulation at the three sites (PGPC12, 1028, and 532) differ to varying degrees, the most obvious difference being that sedimentation rates at the northern two sites averaged about $3cm^{-1}1000y^{-1}$, while those at PGPC12 averaged about $10cm^{-1}1000y^{-1}$

The raw abundance data have been affected by dilution (e.g. of C_{org} by $CaCO_3$ and terrigenous material, and of $CaCO_3$ by terrigenous material) so we cannot interpret them directly in terms of upwelling history and sediment supply without first converting them to rates of accumulation (Figs 3G, 3H and 3J). The most reliable comparative data on rates of accumulation come from SCHMIDT (1992) for 1028 and for stages 2-4 of site 532 (NB: recovery of stage 1 at Site 532 was inadequate). There are no data on accumulation rates at PC16, because it has not been analysed for oxygen isotopes or ^{14}C age.

4.2.1. Organic matter and biogenic silica signals. Given the modern patterns of upwelling (SHANNON, 1985) and of organic enrichment of sediments (Fig.2) it is not surprising that the stage 1 sediments of PGPC12 are richer in C_{org}, accumulated faster, and have a higher S_{org} than their equivalents at 1028 and DSDP532 on the Walvis Ridge abutment. This differentiation extends back into isotope stages 2, 3 and 4, indicating that the upwelling associated with the Lüderitz cell has been consistently stronger than that off northern Namibia throughout the period studied.

The decreases in C_{org} and S_{org} from stages 2-3 to stage 1 at PGPC12 suggest that there has been a weakening of upwelling and the supply of organic matter since the last glaciation, especially since stage 3 when surface waters were at their coldest (according to $U^{k}_{37'}$ data). Much the same trend appears at Site 532 where the S_{org} in stage 3 is double that of stages 1 or 2 (SCHMIDT, 1992, her Fig.15b). The increase in supply of organic matter in stage 3 at site 532 coincides broadly with the influx of left-coiling *Neogloboquadrina pachyderma* noted by OBERHÄNSLI (1991). A somewhat different picture emerges at 1028 on the Walvis Ridge, which is well seawards of the shelf edge upwelling zone (see Fig.1), where there was little difference in S_{org} between Holocene and glacial stages (SCHMIDT, 1992, her Fig.15a). Nevertheless, as at site 532 and PGPC12, there is a peak in S_{org} in stage 3 that is also associated with abundant left-coiling *N. pachyderma* (SCHMIDT, 1992).

Although *N. pachyderma* has been taken as an indicator of cooling (e.g. SCHMIDT, 1992), recent work suggests that a preponderance of the left-coiling variety is a response to food supply rather than temperature (UFKES and ZACHARIASSE, 1993); this supports the notion that upwelling (i.e. nutrient supply and productivity) increased in stage 3. Indeed, left-coiling *N. pachyderma* is abundant today nearshore under the areas of coastal upwelling from Cape Town to the Kunene River (GIRAUDEAU, 1993), so is unlikely to indicate introduction of polar surface water. We take it as another crude indicator of upwelling intensity, confirming the picture from $U^{k}_{37'}$ and S_{org}. Presumably the lack of enrichment of diatoms in association with S_{org} in stage 3 is because diatom production is highest where upwelling is most intense - at the coast, not in the shelf edge upwelling system; the two systems are ecologically different.

Superimposed on these regional differences in S_{org} at 1028, DSDP532, and PGPC12 is a cyclic pattern. Ignoring high frequency variability, such as that expressed by the local excursions in S_{org} at 60ka (582cm) and 50ka (517cm), we find broad maxima in S_{org} at the top of stage 1, in each of stages 2 and 3, and approaching the base of stage 4 (Fig.3J). The peaks of these broad maxima coincide with maxima in the precession index, times when the earth-sun distance was greatest in June, thereby decreasing insolation, and which occurred at around 72ka, 47ka, 22ka, and 1ka (BERGER and LOUTRE, 1991); at these times the monsoons were relatively weak and the trade winds (hence upwelling) and seasonality were relatively strong (SCHNEIDER, 1991). In contrast, the broad intervening minima in S_{org} coincide with minima in the precession index, times when the earth-sun distance was least in June, thereby increasing insolation, and which occurred at around 60ka, 34ka and 11ka; at these times the monsoons were stronger and the trade winds (hence upwelling) and seasonality were weaker (SCHNEIDER, 1991). A close association between precessional 'cooling' and high S_{org} has been documented nearby, at 1028 on the Walvis Ridge by SCHMIDT (1992), and in the Angola Basin, by SCHNEIDER (1991).

A linear response to variations in the precession index cannot account for the development of a maximum in S_{org} and a minimum in SST in stage 3 between 40-55ka ages (Figs 3E and 3J). It must be recalled that much of what appears as palaeoceanographic variability in the southern hemisphere is a consequence of orbital forcing at high latitudes in the northern hemisphere being transmitted to the southern hemisphere by North Atlantic Deep Water (IMBRIE, MCINTYRE and MIX, 1989). Variations in the obliquity of the earth's axis combined with variations in precession to produce a period of consistently low insolation at 65°N between 35-60ka. Transmission of this signal to the southern hemisphere would have further strengthened the trade winds and weakened the monsoons. Upwelling of cold productive water would have been intensified as a consequence, helping to explain both the high S_{org} and the low SST of stage 3. Stronger Trade Winds would also have enhanced mixing in surface waters, thereby deepening the thermocline and increasing productivity seawards of the upwelling zone (SCHNEIDER, 1991). External forcing of the S_{org} signal explains the coeval enrichment of organic matter in stage 3 off Namibia (this paper and SCHMIDT, 1992), in the Angola Basin (MÜLLER *et al*, 1994), in the Arabian sea (TEN HAVEN and KROON, 1991) and off northwest Africa (MÜLLER *et al*, 1983).

Other S_{org} patterns down-core may also be attributable to the pattern of insolation at 65°N, for instance in stage 1 at around 10ka insolation was very high (BERGER and LOUTRE, 1991) and S_{org} was lowest (Fig.3J); at this point the monsoons were strongest and the trade winds (hence upwelling) weakest (SCHNEIDER, 1991).

In order to explain the enrichment of glacial sediments (stage 2 through 4) in organic matter at Site 532, DIESTER HAASS (1985), believing the lack of diatoms to indicate an absence of upwelling influence, assumed that the organic matter might have been supplied to the slope by erosion of shallower water organic rich muds exposed when sea-level dropped in glacial times (this same assumption underlies the interpretation of PC16 by DIESTER-HAASS *et al*, 1988). There is now hard evidence to support her idea, in the form of (1) AMS^{14}C dates indicating substantial mixing of reworked organic matter in stage 2 (Table 1), and (2) the increases in S_{ter} in stages 2 and 4 (Fig.3H). However, Diester-Haass was only partly right, as the AMS data (Table 1) show that reworked material was not mixed into the slope sediments during stage 3.

Although the difference in S_{org} between stages 2, 3 and 4 can be partly explained by the changes in external forcing, changes in sealevel, which fell to -50 to -60m in stage 3, to -80m in stage 4, and to -130m in stage 2 (SHACKLETON, 1987) also played an important role. The fall to -130m would have pushed the coastline seawards by 40km, narrowing the shelf from its present 115km average (Fig.1) to about 75km; the fall to -50m in stage 3 would have exposed much less of the shelf, which would still have been about 100km wide; the fall to -80m in stage 4 would have had an intermediate effect. During both stages 2 and 4 there would have been more exposure and erosion of the shelf than in stage 3, hence more supply of terrigenous material, possibly muddying the water and reducing productivity; there would also have been more reworking of organic matter.

What organic and sedimentary processes would have prevailed during stage 3, when upwelling seems to have been most intense? The main sources of organic matter on the slope then would have been deposition from (i) the shelf edge upwelling system; (ii) occasional seaward excursions of the inner shelf upwelling front, bringing zooplankton production from the seaward side of the front if not diatoms from its landward side; (iii) periodic excursions of jets and filaments of inner shelf water across the shelf edge front and out into the SE Atlantic; and (iv) a slightly increased rate of supply of contemporary organic rich sediment resuspended from the outer shelf by tidal currents impinging on or near the new shelf break (formerly near 400m deep, now near 350m deep). A net increase both in production in the surface waters over the slope, and in the supply of resuspended material from the shelf, would have increased the rate of accumulation of organic matter on the slope. The fall in sealevel of -50m in stage 3 might have caused some erosion of organic rich diatomaceous muds previously deposited nearshore, but there is no evidence to suggest that reworked organic material or diatomaceous muds from this zone reached the slope in bulk at site PGPC12 at this time.

How would oceanographic conditions have changed between stage 3 and stage 2, when sealevel fell a further 70-80m? There would still have been both a shelf break front and a coastal upwelling front, as in the BARANGE and PILLAR (1992) model (Fig.6), but instead of being 75km apart as they are today they would have been only about 35km apart, similar to the situation prevailing today off the Cape Peninsula (SHANNON, 1985), where the two fronts may even coalesce when coastal upwelling is particularly intense (HUTCHINGS *et al*, 1986; PITCHER, BROWN and MITCHELL-INNES, 1992). Such a coalescence would have increased production over the slope, but there is no sedimentary evidence for that having occurred - in fact the reverse. There is evidence for formation of a diatomaceous mud in glacial times just seaward of the present inner shelf break (presumably the old coastline) at 150m (BREMNER, 1981; SUMMERHAYES, 1972; DIESTER-HAASS, 1978). This may represent the coastal upwelling centre during stage 2. The shallowing of the outer shelf edge from 400m to 270m during stage 2 will have encouraged reworking by tidal currents of the organic rich mud that had accumulated on the present middle and outer shelf in stage 3. From the AA/pigment ratio of stage 2 sediments at PGPC12 we know that the eroded and reworked material was of much the same composition as the material deposited on the slope (and presumably the outer shelf) during stage 3 (Fig.3M). Since diatoms are absent from stage 2 in PGPC12, the nearshore diatomaceous muds exposed by a fall in sealevel are unlikely to have been the source of reworked material.

What would have happened to the organic rich diatom muds deposited on the inner shelf in stages 3 and 5 and exposed to erosion as sealevel fell during stages 2 and 4? Given the prevailing equatorward alongshore current patterns (BREMNER, 1981; ROGERS and BREMNER, 1991), it is possible that as sealevel fell these muds were carried in suspension north along the coast, perhaps as far as the Angola Basin, just as, today, the suspended load of the Amazon is swept north along the Brazilian coast to the Caribbean rather than out into the equatorial Atlantic (MILLIMAN, SUMMERHAYES and BARRETTO, 1975). Given the 25cm s^{-1} northward movement of surface water (SHANNON, 1985; NELSON, 1989), progress would have been slow. Suspended material moving seawards and sinking over the middle shelf would eventually have been entrained in the poleward undercurrent, which moves south with speeds of about 5-10cm s^{-1} (NELSON, 1989).

For her interpretation of the history of sedimentation in the late Quaternary of Site 532, DIESTER-HAASS (1985) had assumed that diatoms were the primary indicators of upwelling throughout the section. This followed from the observations of GARDNER *et al* (1984) and DEAN *et al* (1984) that diatom abundance declined markedly between the Pliocene and early Pleistocene towards a diatom-poor late Quaternary section described by GARDNER *et al* (1984) as 'post-upwelling'. Today, diatoms certainly are a primary indicator of the strong upwelling near the coast, but they are not as abundant over the slope, where upwelling is less strong. Over the slope then (or at the shelf edge) diatoms may not be primary indicators of the upwelling that takes place there. Even today there is only a tenuous relationship between diatoms and upwelling off southwestern Africa. Upwelling takes place over the entire region at some time in the year, but although the southern shelf is the more productive (SHANNON, 1985), there are no significant accumulations of diatoms south of the Orange River - the diatom muds occur north of Lüderitz (ROGERS and BREMNER, 1991).

This ecological differentiation between coastal and shelf edge upwelling systems provides one possible explanation for the apparently conflicting signals down-core, with the organic matter telling one story, and the diatoms another. Another possible answer to the problem could lie in the content of silica in Atlantic waters, which BERGER and HERGUERA (1992) argue renders diatoms unsatisfactory as indicators of productivity. In the Atlantic, the lower concentration of silica in subsurface waters leads diatom tests to be more weakly silicified than in the Pacific, making the relationship between the opal flux and supply strongly non-linear and so making the opal flux unreliable as a quantitative index of production. They go on to suggest that preservation of diatoms may well have deteriorated in glacials when there was even less silica in Atlantic subsurface waters. Hence the scarcity of diatoms in stages 2-4 at PGPC12 and Site 532 may be more a function of increased dissolution than of reduced productivity. Silica dissolution is important locally. Even today off this margin, diatoms are more abundant than Radiolaria in surface waters over the slope (BISHOP *et al*, 1978) but very uncommon in slope sediments, so that their small opal fraction is dominated by Radiolaria (BREMNER, 1981). The prevalence of dissolution makes it difficult to relate sedimentary opal to surface productivity (JAHNKE and SHIMMIELD, in press). Our organic matter data, and the available evidence on diatoms and their preservation (e.g. Table 5) indicates that the assumption of DIESTER-HAASS (1985) and DIESTER-HAASS *et al*, (1986, 1992) that diatoms are the primary indicators of productivity over the shelf edge and slope is invalid for the Late Quaternary along this margin.

We suggest that sediments rich in organic matter but containing few if any siliceous organisms were deposited from 'old' upwelled water. Coccolithophorids and other non-siliceous phytoplankton tend to flourish in this 'older' upwelled water, which has become well-stratified, whereas diatoms tend to flourish in 'newer' upwelled water close to shore where the water column is still poorly stratified (see oceanographic review, above).

Radiolaria are common albeit minor constituents of present slope sediments in this area (BREMNER, 1981; 1983). They were also more common in stage 1 than they were during glacials at both sites PGPC12 and DSDP532 (DIESTER-HAASS, 1985), probably in response to a reversion from an 'enhanced' shelf edge upwelling ecology (stage 2) to the 'normal' shelf edge upwelling ecology typical of today and much more influenced by the open ocean.

Conditions favoured diatom accumulation intermittently during stage 4 and early in stage 3 at both PGPC12 (Figs 3N and 3O) and PC16 (Fig.4). However, since the diatom populations differ between stages 3 and 4 at PGPC12, we cannot interpret their occurrence as simple responses to upwelling. The prominent peak in diatoms in stage 3 at PGPC12 has a distinctly oceanic flavour; it is a mix of almost equal amounts of *Azpeitia nodulifer* (oceanic) (Fig.3N) and *Chaetoceros* spores (upwelling), with subordinate *Delphineis Karstenii* (upwelling) (Fig.3O). As it occurs at the boundary between stages 3 and 4 it may reflect a short-lived response of the system to a change in the climate. The least ambivalent association of diatoms with upwelling is that in stage 4 at PGPC12, where *D. Karstenii* (upwelling indicator) (Fig.3O) is associated with a peak in S_{org} and a distinct short-lived cold water event (Figs 3E and 3J). Just how much upwelling took place in stages 2 and 3 nearby at site PC16 is difficult to establish without hard data and rates of accumulation. The high C_{org} values of stage 2 there (Fig.4) may, as at PGPC12, reflect reworking and mixing when sealevel was at its lowest, rather than increased upwelling.

Can we use our C_{org} data to estimate palaeoproductivity? BERGER and HERGUERA (1992) and JAHNKE and SHIMMIELD (in press) argue that given the complex routes by which organic matter arrives on the continental slope, either from the overlying surface waters or laterally via resuspended material in bottom water cascading down the slope, accurate estimates of palaeoproductivity in that environment are impossible. In addition, MIX (1989) shows that given the sensitivity of palaeoproductivity estimates to sedimentation rates, the application of C_{org} data to palaeoproductivity estimates demands extremely accurate chronologies, which we lack - given our few ^{14}C dates, and the difficulties we found in matching our $\delta^{18}O$ curve precisely to the SPECMAP standard. Productivity calculations would be pointless for stage 2, for which about 43% of the organic matter has probably been reworked. A rough qualitative measure of past productivity is given by the S_{org} curve (Fig.3J), bearing in mind that some 40% of the peak in stage 2 is reworked material, and that there was probably extensive reworking during stage 4 (though to a lesser extent since sealevel was higher). Removing the reworked element does not eliminate the increases in S_{org} in stages 2 and 4 compared with stage 1, it merely reduces the signal. Our conclusion that productivity increased somewhat in stages 2 and 4 and was at a maximum in stage 3 is consistent with palaeoproductivity estimates based on cores further seawards in this area (MIX, 1989) suggesting higher production in stages 2 through 4 than at present. The resolution in those cores was not enough to identify stage 3 as a discrete entity.

In summary, our new data from PGPC12 support the notion that upwelling was more intense in glacials than in interglacials off Namibia, as proposed by DEAN *et al* (1984) and OBERHÄNSLI (1992) on the basis of data from DSDP532, but contrary to the models of DIESTER-HAASS (1985) and SCHMIDT (1992). This finding is consistent with glacial increases in productivity associated with the Benguela Current Upwelling System in the Angola Basin north of the Walvis Ridge (e.g. SCHNEIDER, 1991).

4.2.2. Carbonate signals. The abundant supply of nutrients along the Namibian margin leads to a steady offshore increase in the abundance of $CaCO_3$, coccoliths and foraminifera in bottom sediments away from the coastal diatomaceous mud belt (BREMNER, 1981; GIRAUDEAU, 1992, 1993), and has helped to maintain a moderately high rate of accumulation of carbonate skeletal remains (S_{CO3}) through time at PGPC12 (Fig.3G) and elsewhere on the continental slope (SCHMIDT, 1992; DIESTER-HAASS, 1985). Like C_{org}, $CaCO_3$ was most abundant when productivity (i.e. nutrient supply) was highest, in stage 3 (Figs 3F and 3I). At PGPC12, 1028, and DSDP532, S_{CO3} increases towards the top of stage 1 (Fig.3G and SCHMIDT, 1992). This increase parallels that seen in biogenic silica (Fig.3P) and S_{org} (Fig.3J) at PGPC12, and most likely reflects the precessionally-driven strengthening of the trade winds over the past 10ky.

There is abundant evidence for enhanced dissolution of foraminiferal remains during glacials at PC16 (Fig.4), 1028 (SCHMIDT, 1992), and DSDP532 (SCHMIDT, 1992; DIESTER-HAASS, 1985) and by implication at PGPC12, where sand-sized foraminifera are abundant only where $CaCO_3$ is high (SUMMERHAYES *et al*, 1979). Dissolution may explain much of the reduction in $CaCO_3$ in the sediments of the glacial isotope stages 2 and 4 (Fig.3F). During the highly productive interstadial, stage 3, when both S_{CO3} and S_{org} were high, there is an inverse relationship between S_{CO3} and S_{org} centred on 422cm (48ka). This suggests that decomposition of the organic matter in the sediments also contributed to the dissolution of carbonate (Figs 3G and 3J). During stages 2 and 4 dissolution could have been enhanced either by a fall in pore water pH caused by the decomposition of organic matter within the sediments, or by the impingement on the slope of bottom water richer in CO_2 in the poleward undercurrent. The striking relationship between S_{CO3} (Fig.3G) and $\delta^{18}O$ (Fig.3A), and the general independence of S_{CO3} (Fig.3G) from SST (Fig.3E) suggests there has been a link between the carbonate content of local bottom waters and the global events controlling the isotopic composition of the water mass that was independent of the local upwelling process. We suggest that the linking process was the introduction of cold bottom water rich in CO_2 in the poleward undercurrent, which would have kept sediment carbonates especially low during the global cold stages 2 and 4.

In stage 1 the general global warming, with an abrupt transition from one climatic state to another, was accompanied by a major oceanographic change at PGPC12. Productivity, represented by S_{org} (Fig.3J), dropped yet again as surface water warmed further. Warm oceanic conditions favoured the production of nannoplankton ooze which remained carbonate rich because it was no longer subject to significant dissolution. This transition was accompanied by a profound change in ecology from a system dominated by *Gephyrocapsa muellerae* to one dominated by *Emiliana* *huxleyi* (Fig.3B); the accompanying large increase in the AA/pigment ratio (Fig.3M) probably reflects an increased abundance of coccolithophorids in Recent times, as indicated by the increased abundance of $CaCO_3$.

4.2.3. Non-carbonate (terrigenous) signals. The terrigenous accumulation curve (S_{ter}) (Fig.3H) is crudely related to the precession index, with maxima near 22ka, 47ka and 72ka when the earth-sun distance was greatest and Trade Winds were strongest. The crude association between S_{ter} and S_{org} (Figs 3H and 3J) may reflect the increased seasonality reported by SCHNEIDER (1991) at these times, giving rise to increased productivity in one season and large dust storms in another. However, not all of the S_{ter} signal is necessarily aeolian; some maxima may have been caused by (1) erosion caused by the lowering of sealevel; or (2) increased riverine runoff (more pluvial climate in the hinterland).

Although S_{ter} tended to be higher in glacial stages 2 and 4, when sealevel was lower, than in stage 3, the relation between S_{ter} and sealevel is neither linear nor simple. The lowering of sealevel might have increased terrigenous supply to the slope by several means, for instance: (1) increased erosion of the hinterland by rivers down cutting to a new base level; (2) erosion of the newly exposed coastal plain by wave action; or (3) increased erosion of the outer shelf (now only 230-270m deep at its outer edge, instead of 360-400m) by enhanced bottom currents as implied by VAN ANDEL and CALVERT (1971). Options (1) and (2) are possible, but (3) seems unlikely given the water depth and what is known of the speed of the poleward undercurrent.

Stronger, or more frequent, offshore winds bringing dust from the desert hinterland may have contributed to increasing S_{ter} during the glacials. In PC16, DIESTER-HAASS *et al* (1988) observed a marked increase in coarse-grained aeolian dust particles (quartz and mica) during stage 2 and a smaller increase during stage 4 (Fig.4). An increase in the incidence of easterly 'Berg' winds in stage 2 (and perhaps 4) may explain the observed patterns. Easterly winds are less conducive to upwelling, which could help to explain why S_{ter} and S_{org} diverge in stages 2 and 4 (Figs 3H and 3J).

Past rainfall patterns in southern and southwestern Africa lend support to the idea that during the glacials, as the thermal gradient between the equator and the poles steepened, so the S. Atlantic high pressure system moved north (VAN ZINDEREN BAKKER, 1984), allowing cool wet polar air to penetrate southwestern Africa to about 24°S between Lüderitz and Walvis Bay (COETZEE and SCHOLTZ, 1983; TYSON, 1986). Such an increase in rainfall in the hinterland may explain why at PC16 stage 4 contains abundant wood fibres (Fig.4). If stage 4 was more humid than stage 2, this would explain its higher S_{ter}.

Down-core at PGPC12 the increase in illitic clay suggests the influence of the Namib Desert area increased while that of the Kunene River waned. Increased illite content could reflect: (1) an increased supply of illitic material from the local hinterland by rivers or winds; (2) a weakening of the supply of less illitic material carried south along the slope from the Kunene River by the poleward undercurrent; (3) weakening of the poleward undercurrent; or simply (4) the greater proximity of the coring site to the coast at low sealevel times. When sealevel was low, during isotope stages 2 and 4, the discharge from the Kunene, to the north of the Walvis Ridge, may have gone straight across the very narrow shelf into the Angola Basin rather than being diverted south; thus the change in clay mineralogy may have had a physical rather than a climatic cause.

4.2.4. Regional climatic and oceanographic change. During the Last Glacial Maximum the 7° of latitude northward displacement of the Polar Front towards Cape Town (MACKENSEN, GROBE, HUBBERTEN and KUHN, 1994), the 2° northward displacement of the Subtropical Front in the southwest Indian Ocean (PRELL, HUTSON, WILLIAMS, BÉ, GEITZENAUER and MOLFINO, 1980), and the 2-5°C cooling of Subantarctic Surface Waters (MORLEY and HAYS, 1979), all suggest that the thermal gradient south of Africa steepened in glacials, displacing the S. Atlantic mid-latitude high pressure cell north by 2-5° of latitude (TYSON, 1986).

During the last glaciation (stages 2-4) there was a 2-4°C cooling of the surface waters in the northern Cape Basin (MORLEY and HAYS, 1979; EMBLEY and MORLEY, 1980; CLIMAP PROJECT MEMBERS, 1981; CHARLES and MORLEY, 1988). This cooling may have been caused in part by reduced advection of heat around the Cape of Good Hope from the Indian Ocean in eddies spun off from the Agulhas Current (GORDON, 1986). Also the Agulhas Current itself was cooler during glacials (PRELL *et al*, 1980; WINTER and MARTIN, 1990). Another source of cooling, an increased leakage of cold Subantarctic Surface Water across the Subtropical Front in jets and eddies, as described today by SHANNON *et al* (1989), would have cooled the Benguela Current near Cape Town, which appears not to have happened (CLIMAP PROJECT MEMBERS, 1981). Increased flow of warm Agulhas water into the southeast Atlantic following the Last Glacial Maximum is invoked by CHARLES and MORLEY (1988) to explain their radiolarian data. SCHNEIDER (1991) suggested that during glacials Agulhas Current inflow to the Cape Basin would have increased at times when the precession index was smallest (earth-sun distance in June was least) and decreased when the precession index was largest, thereby accentuating changes in seasonality, the highest seasonality occurring when the precession index was largest and the Trade Winds were strongest.

The equatorwards movement of the pressure system forced a similar shift in the Trade Winds, and the steeper thermal gradient will have strengthened them (TYSON, 1986). The increased temperature differential between the ocean and the landmass in glacials would also have contributed to the strengthening of the coastal wind field (HAY and BROCK, 1992). These increases in the wind strength would have increased the wind stress curl near the shelf edge, thereby enhancing upwelling there (HAY and BROCK, 1992). The northward shift in the Trade Winds probably weakened the upwelling intensity south of Lüderitz, but the increased productivity as represented by S_{org} data from PGPC12 implies that the Trade Winds strengthened in the vicinity of Lüderitz, and the evidence from DSDP532 suggests that the winds were stronger there too.

Unfortunately there is no evidence from onshore of what happened to the winds or climate northwards of Lüderitz during the last glaciation (stages 2-4) (VAN ZINDEREN BAKKER, 1984). The S_{org}, SST and left-coiling *N. pachyderma* data imply that upwelling-favourable southerly winds predominated at and to the north of Lüderitz in the interstadial, stage 3, creating throughout the study area a cold event which was intensified by the insolation minimum. The coeval flux of terrigenous material (S_{ter}) suggests that, just as today, easterly 'Berg' winds seasonally blew dust offshore. Earlier, in glacial stage 4, winds off Lüderitz were probably less upwelling favourable (i.e. less southerly) and more offshore and dust-laden (i.e. more easterly). The less intensive upwelling led to the surface waters being slightly warmer in stage 4 than in stage 3 (Fig.3E). The winds appear to have been even less upwelling-favourable in stage 2, leading to less upwelling and even warmer surface waters (though still not quite as warm as today's). As in stage 4, 'Berg' winds in stage 2 were better developed than those in stage 3, giving rise to a higher S_{ter}. At the end of stage 2 there was a substantial reduction in the strength of both the Trade Winds and the 'Berg' winds, thereby lowering both S_{org} and S_{ter}. DIESTER-HAASS *et al* (1988) had suggested that the change in the Trade Winds in glacials shifted the centre of upwelling northwards away from the vicinity of PGPC12 and PC16, but we find no evidence for that northward shift; indeed, our evidence suggests that upwelling intensified there in stage 3, and to a lesser extent also in stages 2 and 4, compared with today.

The shift in the Trade Winds also had an important effect at the northern end of the Benguela Current Upwelling System near the Walvis Ridge. Today, the Angola-Benguela Front marking the northern boundary of the system (Fig.5) is a perennial feature extending up to 1000km offshore, mostly between 14-16°S, but ranging from 12-17°S (MEEUWIS and LUTJEHARMS (1990). It lies furthest north in the austral winter and early spring, the main period of upwelling off northern Namibia, when the S. Atlantic high is centred between 20-23°S. It is furthest south in the late austral summer and autumn, when warm water penetrates south and the S. Atlantic high is centred near 28-35°S. In glacials, when the S. Atlantic high pressure cell migrated north and the mid latitude high pressure belt was compressed, the front might have been close to 12-14°S year-round (JANSEN, 1990; JANSEN and VAN IPEREN, 1991). Thus, year-round during glacials much of the southern Angola Basin would have lain beneath the productive waters of the Benguela Current Upwelling System, whether the Trade Winds were stronger or not, enriching bottom sediments in organic matter as shown by BORNHOLD (1973).

This model does not necessitate a shift in upwelling away from the northern Walvis Ridge during glacials as implied in the model of DIESTER-HAASS (1985). She assumed that today the Benguela Current turns west at about 20°S, bringing nutrient-rich water over DSDP site 532 (Fig.1), and during glacial times its turning point was further north, taking nutrient rich water northwards along the shelf past DSDP site 532 rather than over it. Our re-evaluation of the data suggests (a) that at present the Benguela Current turns west at 17-18°S rather than 20°S (MEEUWIS and LUTJEHARMS, 1990); (b) that during glacials it turned west even further north thereby influencing the Angola Basin more then than at present; (c) that DSDP site 532 is supplied by upwelling from the same shelf edge system that supplies site PGPC12, and was not by-passed by nutrient-rich shelf waters in glacials; (d) that upwelling actually increased in glacial stage 3 at both sites, consistent with a strengthening of the Trade Winds and an intensification of the Benguela Current Upwelling System; (e) the resulting increase in productivity was felt seawards in the Cape Basin (MIX, 1989), but appears not to have been experienced seawards of the shelf edge upwelling system at 1028, where organic accumulation apparently remained unchanged from glacial to Holocene (SCHMIDT, 1992).

On many other continental margins, where the shelf edge is close to the worldwide average depth of 130m, it is not clear if enrichment of slope sediments in organic matter in glacials reflects a real increase in upwelling and productivity, or merely the seawards displacement of the zone of upwelling and productivity that characterised the continental shelf when sealevel was high. Off Namibia, where the shelf is anomalously deep at 400m, the coastal upwelling system on the shelf was never displaced as far as the shelf edge in glacials, so what we see in PGPC12 (apart from the reworked component in stages 2 and 4) does indicate a real increase in the intensity of upwelling in glacials. Further seawards, foraminiferal and C_{org} data show that this wind-driven increase in productivity affected much of the SE Atlantic off southwestern Africa during the Last Glacial Maximum (MIX, 1989). Presumably these offshore increases in productivity reflect increased rates of seaward transport of nutrients and export production in filaments and plumes like those that breach the shelf edge upwelling system today.

We do not find any evidence to support the suggestion of HAY and BROCK (1992) that the subsurface waters providing the source of upwelling off Namibia were less nutrient-rich (i.e. less productive) in glacials than today. They assumed that less generation of NADW in glacials might have caused a concomitant reduction in the production of AAIW, thereby lowering productivity in the upwelling cells. We can find no evidence in the Namibian margin cores to support that contention.

5. CONCLUSIONS

1. Organic matter is enriched on the continental slope off Namibia in response to upwelling and/ or mixing processes taking place in a shelf edge upwelling system. That system seems to be quasi-permanent and physically independent of the much better known coastal upwelling system. Nervertheless, the two systems may be linked in terms of productivity, as seaward advection of

nutrient-rich 'old' upwelled water from the coast may enhance the nutrient content of upwelling water at the shelf edge. While Table 5 suggests that some organic matter probably reaches the slope in suspension in near-bottom flows originating on the outer shelf, if the SEEP experiment is anything to go by (BISCAYE *et al*, 1988) only 20% of the flux reaching the slope will have been exported from the shelf in this way; the rest represents production *in situ* in overlying surface waters. Field experiments are needed off Namibia to test these ideas.

2. A northward shift in the S. Atlantic high pressure cell during the last glaciation (isotope stages 2-4), perhaps by as much as 5°, intensified upwelling-favourable southerly Trade Winds, thereby intensifying upwelling and increasing productivity along the Namibian margin.

3. Cyclicity in the precession index, with a periodicity of about 20ky, appears to control periodicity in organic carbon production and accumulation at PGPC12 on the slope off Walvis Bay, as well as on the Walvis Ridge where this link was recognised by SCHMIDT (1992). Productivity increased in those periods when the earth-sun distance was greatest and insolation least. The largest productivity signal occurred in stage 3, when the precession index and obliquity signal both contributed to minimising insolation. At that time the southerly, upwelling-favourable Trade Winds were at their strongest, increasing upwelling and productivity; surface waters were at their coldest as a result. The external forcing led to similar cooling and increased production in stage 3 in the Angola Bain, off northwest Africa, and in the Arabian Sea.

4. During the period of strongest upwelling and productivity, in stage 3, strong seasonality is suggested by a moderately high, coeval rate of accumulation of both organic matter and terrigenous material. In the absence of significant signs of aeolian sedimentation, some of the terrigenous influx may be attributed to runoff resulting from a more humid climate in the hinterland.

5. In stage 3 (the interstadial of the last glaciation) not only was sealevel only 50m or so lower than at present, but also the shoreline was still about 100km from site PGPC12. This is not far from the present shoreline, so it seems likely that the overall oceanographic conditions were similar to those of today, with widely separate coastal and shelf edge upwelling systems. However, because the water was shallower there may have been more resuspension of contemporaneous organic rich sediment from the outer shelf than there is today. This enhanced resuspension, together with the intensified upwelling, supplied more organic matter to the slope that we see today. The mixing data (Table 1) show no evidence for a supply of organic matter reworked from older deposits.

6. SST increased and accumulation of organic matter diminished in stage 2 (full glacial conditions) compared with stage 3 (the interstadial). Thus it seems likely that climatic conditions favoured less upwelling in stage 2. The increase in aeolian components in stage 2 at PC16 confirms a climate change, suggesting there had been an increase in the strength or frequency of seasonal easterly 'Berg' winds off the desert.

7. Although sealevel was at its lowest (-130m) in stage 2, the coast was still about 75km from the shelf edge so it is likely that there would still have been both a coastal and a shelf edge upwelling system. The nearshore diatomaceous mud belt appears to have been displaced seawards to about the 160m line on the modern middle shelf some 50-60km from the shelf edge. We found no evidence to suggest that this nearshore deposit spread in such a way as to influence sedimentation on the slope in stage 2.

8. The S_{org} and ^{14}C data suggest that productivity at the shelf edge was only slightly greater in stage 2 than it has been on average in stage 1, the increase in S_{org} in stage 2 compared with stage 1 resulting

mostly from reworking of older material (of late stage 3 age) from the outer shelf.

9. Conditions in stage 4 seem to have been broadly similar to those in stage 2.

10. Diatoms are unreliable indicators of productivity in these Late Quaternary slope sediments, probably because the ecology of the shelf edge system is not favourable to diatom production, which reaches a maximum in the freshly upwelled waters of the coastal upwelling system, and the water depths are not favourable to diatom preservation.

11. Compared with stage 1, organic carbon accumulation rates (S_{org}) in stage 3 increased fourfold at PGPC12, and doubled at DSDP site 532 (calculated from data in SCHMIDT, 1992), suggesting that although upwelling increased in intensity during the last glaciation between Lüderitz and the Walvis Ridge, the increase was greater off southern Namibia.

12. During glacials the northern margin of the Benguela Current Upwelling System (the Angola-Benguela Front) probably stayed close to its present northernmost (winter) position between 12-14°S during both summer and winter, enabling the Benguela Coastal Current to bring more nutrient-rich water into the southern Angola Basin then. That enrichment did not take place at the expense of upwelling along the Namibian margin, from Lüderitz northwards. However, upwelling in glacials off the South African margin south of Lüderitz probably did diminish in response to northward movement of the pressure system and concomitant reduction in upwelling-favourable coastal winds.

13. According to MIX (1989) productivity also increased in the wider S. Atlantic seawards of the shelf edge frontal zone in the last glaciation (stages 2-4), though this increase is not apparent in the data from site 1028 on the Walvis Ridge seaward of site 532 (SCHMIDT, 1992) for reasons that are not clear.

14. The patterns of carbonate accumulation off Namibia show strong dissolution maxima especially at times of lowest sealevel (stages 2 and 4) and at times of maximum accumulation of organic matter (mid stage 3). Global enrichment of bottom water in dissolved CO_2 in glacial maxima is a preferred mechanism for creating widespread dissolution of carbonate in cold stages 2 and 4. Dissolution in mid stage 3 is probably a response to the enhanced decomposition of organic matter in bottom sediments.

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