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Distinct year-to-year particle flux variations off Cape Blanc during 1988–1991: Relation to $\delta^{18}O$-deduced sea-surface temperatures and trade winds

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ABSTRACT

Particle fluxes measured from 1988 to 1991 adjacent to a coastal upwelling site off Cape Blanc showed significant interannual variability of fluxes and sea-surface temperatures (SST) deduced from stable oxygen isotope analysis of the planktonic foraminifera *Globigerinoides ruber* and, partly, of the pteropod *Limacina inflata.* For the duration of the study period, a decrease in the seasonality of SST's was observed, as well as a significant decrease in the average annual SST from 24.4° to 20.8°C. This cooling trend was mainly the effect of a drastic decrease in the summer to fall SST (from 27.2° to 21.8°C). In comparison, the winter-spring SST decreased only slightly from 20.3° in 1988 to 19.8°C in 1991. Concomitantly, we measured decreasing annual total, carbonate, biogenic opal and lithogenic fluxes and, in contrast, increasing marine organic carbon fluxes. During 1991, when cold SST's prevailed and the trade winds were rather high throughout, annual biogenic and lithogenic fluxes (except organic carbon) were lower by approximately a factor of two compared to the other years. Colder SST's, generally corresponding to stronger trade winds and upwelling intensity, did not result in increased biogenic opal and lithogenic matter sedimentation; but higher marine organic carbon fluxes were recorded. Decreasing summer-fall SST from 1988 to 1991 coincided with decreased carbonate sedimentation maxima which generally occurred during the warm summer season. In the summer of 1989, when SST's were the highest of the four-year sampling period and upwelling was less intense due to weak spring-summer trades, a large sedimentation pulse of pteropod shells was observed. Our data set does not yet provide conclusive evidence that the observed year-to-year flux and SST variations represent larger-scale, periodically occurring climatic variations in the eastern Atlantic but it offers insight into the prevailing large variability in biochemical cycles and processes in the eastern Atlantic.

1. Introduction

Year-round flux measurements using time-series sediment traps performed in various parts of the ocean show seasonality in particle fluxes, e.g. in the Sargasso Sea (Deuser and Ross, 1980; Lohrenz *et al.,* 1992), the Panama Basin (Honjo, 1982), the North Pacific (Honjo, 1984), the Arabian Sea (Ramaswamy *et al.,* 1991), the Bay of Bengal (Ittekkot *et al.,* 1991), the Southern Ocean (Wefer *et al.,* 1988) and in the

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North Atlantic (for a summary, see Ducklow and Harris, 1993). Other publications by Deuser *et al.* (1988, 1990) and Lyle *et al.* (1992), have shown significant year-to-year variability of fluxes. Due to the coupling of particle sedimentation with upper-ocean thermal structure and biology (Deuser, 1986), such long-term flux studies could provide useful information about larger-scale climatic variability. In combination with remote-sensing data they enable us to identify possible causes of periodically or episodically occurring sedimentation events which might not be detected in short-term studies. In addition, they facilitate the interpretation of paleoceanographic proxies and give a more reliable estimate of average fluxes to the seafloor.

As emphasized by Servain and Legler (1986), large-scale dynamics in the tropical Atlantic appear to be mostly wind-forced. In contrast to the Pacific, where interannual variations are strongest (ENSO = El Niño Southern Oscillation), seasonal signals should dominate in the Atlantic Ocean. However, year-to-year changes in SST and wind stress are also present in the Atlantic Ocean, especially in areas with low and highly seasonal SST, e.g. coastal and equatorial upwelling zones (Servain *et al.,* 1985; McClain and Firestone, 1993).

In this study, we present particle flux data from a coastal open-ocean transition system of the eastern tropical Atlantic, over a four-year period. We focus mainly on the interannual variation of fluxes rather than on seasonal events and composition of particles and we use the stable oxygen isotope composition of planktonic foraminifera to investigate the upper ocean SST's. The main questions addressed are: (1) how do SST's change over the four-year sampling period? (2) can we distinguish year-to-year variations in the magnitude of fluxes? (3) what is the connection between surface water properties (i.e. SST), trade wind velocities and fluxes of biogenic and lithogenic components?

2. Regional setting

The CB1-4 moorings were located in the Canary Current (Fig. 1a), which is strongly influenced by seasonally varying trade winds and entrainment of cold and nutrient-rich waters off northwest Africa (Hagen and Schemainda, 1989). Relatively cold waters from higher latitudes are advected there, having a strong meridional component. A 50–70 km wide coastal band of cold water which corresponds to a zone of "primary upwelling" is separated by a front from the Canary Current (Hagen, 1981) where a less intensive "secondary upwelling" may occur (Postel, 1990; Mittelstaedt, 1991). In the upper water column between 150 and 600 m depth, a transition zone between North Atlantic Central Water (NACW) and the less saline South Atlantic Central Water (SACW) can be observed near Cape Blanc (Hughes and Barton, 1974). This transition zone called the Central Water Boundary (Zenk *et al.,* 1991), undergoes spatial and temporal variations. This water boundary represents the southeastern side of the Canary/North Equatorial System (Zenk *et al.,* 1991).

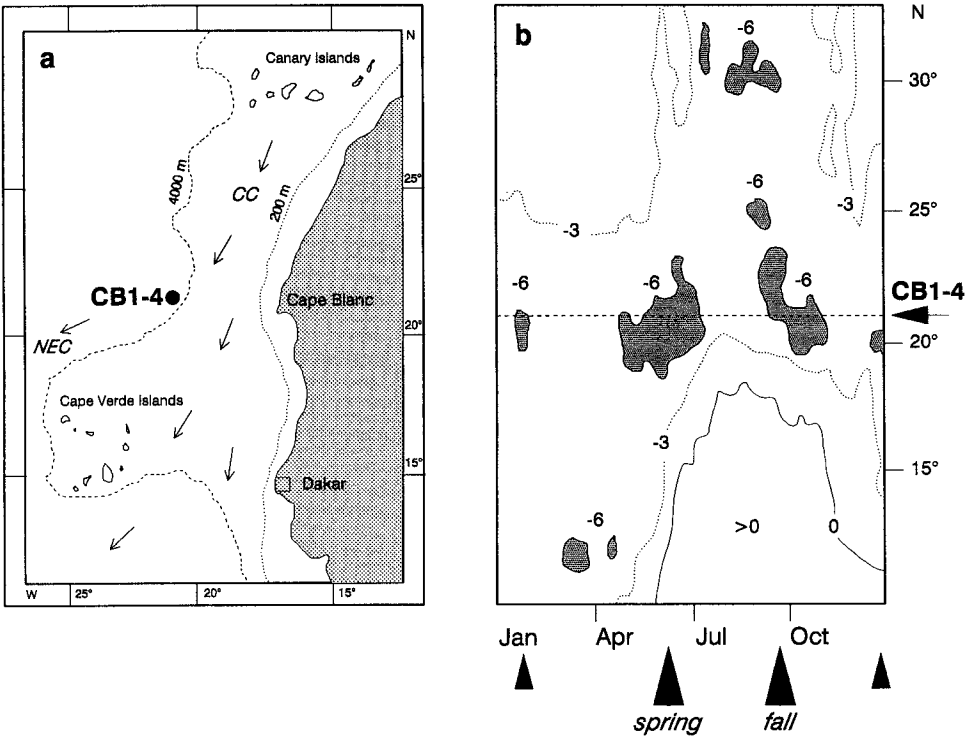

Figure 1. (a) Location of the mooring site off Cape Blanc (CC = Canary Current, NEC = North Equatorial Current). (b) Mean SST differences between coastal areas and open ocean for the period of 1969–1976 (modified, after Speth and Detlefsen, 1982). Large black arrows in spring and fall indicate strong seasonal cold-water intrusions from the coast off Cape Blanc; smaller arrows indicate lesser intensive upwelling periods.

South of Cape Blanc (south of 21N), the more nutrient-rich SACW is found, providing part of the source water for upwelling (Hughes and Barton, 1974). Furthermore, this southern offshore region is also influenced by tropical circulation in the surface layer (Mittelstaedt, 1991) with warm northward-moving counter currents prevailing, especially from July to October (Van Camp *et al.,* 1991).

Due to fairly persistent and strong northeastern trade winds off Cape Blanc, upwelling occurs throughout the year (Schemainda *et al.,* 1975; Van Camp *et al.,* 1991; Mittelstaedt, 1991) with periods of stronger intensity, mainly in spring (April–June) and fall (September–October) (Postel, 1990; Van Camp *et al.,* 1991). This seasonality is also documented in SST anomalies between the coastal zone and the central Atlantic averaged from 1969 to 1976 (Speth and Detlefsen, 1982; Fig. 1b). Phytoplankton biomass generally is high on the shelf along the entire coast and drops significantly beyond the shelf break, as indicated by chlorophyll values below 0.5 mg m^{-3}. However, off Cape Blanc (around 20N), "giant filaments" with chlorophyll concen-

trations of 1–2 mg m^{-3} are present several hundred kilometers offshore from the shelf break, persisting throughout the year (Van Camp *et al.,* 1991) with substantial seasonal and interannual variations (Bricaud *et al.,* 1987). These filaments are caused by the convergence of the southward flowing Canary Current and the poleward flowing counter current south of Cape Blanc (Mittelstaedt, 1991). McClain *et al.* (1990) and Gabric *et al.* (1993) have also reported the occurrence of significant advection of phytoplankton biomass from the coastal upwelling regime to the open ocean, along the coast of northwest Africa from 15N to 25N. Thus, the general region of enhanced pigments off northwest Africa is the result of the combined effects of offshore advection of coastal water and open ocean upwelling.

3. Material and methods

We deployed cone-shaped multisample sediment traps with 13, 20 and 22 cups and 0.5 (CB1,3,4) or 1.17 m^2 (CB2) collection areas (Table 1). All traps were fitted with a grid at the top. In 1988 and 1989, we used Mark V (CB1) and Mark VI (CB2) traps described in detail by Honjo and Doherty (1988). For the CB3 and CB4 deployments, we used time-series Kiel Aquatec sediment traps which are similar in shape and aspect ratio to the Mark V traps. The deployment data for all traps are given in Table 1.

The sampling cups were poisoned with mercuric chloride prior to and after deployment. The influence of various poisons and preservatives on the sediment trap material is discussed in Lee *et al.* (1992). Trap solution was prepared from filtrated seawater and suprapur NaCl was added to increase the density (around 40%). Samples were carefully wet-sieved through a 1 mm screen and "swimmers" (Knauer *et al.,* 1984) were removed by hand. However, "swimmers" did not account for much of the total material due to deep deployment positions of the traps (around 700 m and between 2195 and 3562 m). Significant material in the >1 mm fraction was only

Table 1. Locations of the Cape Blanc moorings (CB1–4), sampling durations and intervals.

Trap name	Trap type (opening)	Position	Water depth (m)	Trap depths (m)	Sampling duration	Samples × days
CB1	Mark VI (0.5 m^2)	20°45.3′N 19°44.5′W	3646	2195	22.03.88–08.03.89	13 × 27
CB2	Mark V (1.17 m^2)	21°08.7′N 20°41.2′w	4092	3502	15.03.89–24.03.90	22 × 17
CB3	Kiel SMT 230 (0.5 m^2)	21°08.3′N 20°40.3′W	4094	730 3557	08.04.90–30.04.91 29.04.90–08.04.91	18 × 21.5 16 × 21.5
CB4	Kiel SMT 230 (0.5 m^2)	21°08.7′N 20°41.2′W	4108	733 3562	05.03.91–19.11.91 05.03.91–19.11.91	20 × 10 20 × 10

found in the 3502 m sample from 1989. However, these were empty pteropod shells and not considered as "swimmers" (Kalberer *et al.,* 1993). The <1 mm fraction was split into aliquots, and the freeze-dried material was analyzed as described by Fischer and Wefer (1991). Lithogenic matter was estimated according to: Lith = Total flux - (opal flux + carbonate flux + 2 × C_{org} flux). Biogenic opal was determined by automated wet leaching (Müller and Schneider, 1993). Fluxes were not corrected for dissolution losses. Dissolved Si in the supernatants from the CB1 deployment, for instance, was used to check the seasonal and annual loss of biogenic opal in the cups. Opal loss due to dissolution was estimated between 1–7% of total biogenic opal and amounted to 2.5% on an annual basis (Fischer and Wefer, 1991). From our measurements, we conclude that dissolution in the cups is very rapid and is not related to the deployment time, and thus affects all samples to a similar extent. We therefore regard all fluxes as minimal values.

Stable carbon isotope measurements of total organic carbon were performed with a Finnigan MAT delta E mass spectrometer. Organic matter was oxidized with an Heraeus CHN Elemental Analyzer which was attached to a trapping box. The overall analytical precision deduced from an internal laboratory standard was better than 0.1% ($\pm 1\sigma$). Planktonic foraminifera were picked by hand from a 1/64 or 1/4 split (CB4) and dried at 60°C. Stable isotope analysis was performed on approximately 8–10 individuals of *Globigerinoides ruber* (white; 200–300 μm size fraction) per sample using a MAT 251 mass spectrometer and an automated preparation line (KIEL device). Calibration of the laboratory standard to PDB was achieved using NBS 18, 19 and 20 standards. Analytical precision is 0.07% for $\delta^{18}O$ ($\pm 1\sigma$). For the calculation of SST, we applied the paleotemperature equation of Craig and Gordon (1965) and used an equilibrium oxygen isotope offset of -0.35% for *G. ruber* white (Fairbanks *et al.,* 1980; Ravello and Fairbanks, 1992) which is an algal symbiont-bearing spinose species. According to Deuser *et al.* (1981), Deuser (1987), Deuser and Ross (1989) and Ravelo and Fairbanks (1992), spinose species such as *G. ruber* (white) calcify in the mixed layer or in the upper part of the seasonal thermocline (0–30 m), and thus record near-surface temperatures. $\delta^{18}O$ of modern seawater (SMOW) was computed from the relationship to salinity established by Craig and Gordon (1965) ($\delta^{18}O_{SW} = -21.1 + 0.61S$). Application of other paleotemperature equations or relationships between the stable oxygen composition of seawater and salinity (e.g. Ganssen and Sarnthein, 1983) did not result in significant SST differences.

4. Results

a. Currents. Aanderaa current meter measurements from approximately the same depths as the traps (20 m below) were available for CB3 and CB4 (upper traps) as well as for CB2 (lower trap, data uncomplete) and CB4 (lower trap). During the

collection period of CB3 (1990–1991), current speeds at that position ranged from 1 to almost 20 cm s^{-1} in 750 m water depth, which corresponds to the Antarctic Intermediate – Mediterranean Water masses (AAIW – MW; Zenk *et al.,* 1991). In spring and fall, the current direction was between 90 and 120°, in summer 90 to 270°, and in winter mostly southward. Similar current speeds were recorded for the following period in 1991, mostly between 5 and 10 cm s^{-1} with some spikes of 20 cm s^{-1}. Current directions varied significantly, but retained some similarities to the earlier sampling period. Near the deeper CB4 trap (3588 m) located in the NADW mass, current velocities were significantly lower with 3 cm s^{-1} on the average ranging from 0 to 8 cm s^{-1}. Flow direction in this water mass was generally southward.

b. Particle composition, seasonal and interannual flux variations. Total fluxes ranged from 8.2 to 361 mg m^{-2} day^{-1} and lithogenic fluxes from 2.3–157 mg m^{-2} day^{-1} (Fig. 2). From 1988 to 1991, total and lithogenic fluxes decreased significantly, but the relative contribution of lithogenic to total matter increased. This is also indicated in the decrease of the Lith./Corg ratios, especially from 1990 to 1991 (Fig. 2). The composition of the material, though changing during the four-year period, was dominated by carbonate (21–78%) and lithogenic components (upper traps: 12–43%; lower traps: 29–53%). Organic carbon comprised 4.3–21.4% of the upper and 2.4–13.6% of the lower trap samples. Biogenic opal constituted only 1.7 to 14.6%.

Background fluxes were relatively high in all years corresponding to rather continuous upwelling at the trapping site (Schemainda *et al.,* 1975; Van Camp *et al.,* 1991). Peak fluxes were generally measured in winter, spring, summer and fall. However, they differed in timing and magnitude from year to year (Fig. 2). The most prominent peaks at the deeper trap level were (1) in spring, summer and fall of 1988, (2) in summer 1989, (3) in winter/spring 1990 and (4) in winter, spring and summer 1991. Generally, the fall peak was weaker (except in 1988) and summer sedimentation events dominated during 1988, 1989 and 1991. Note, that the 1988 deep flux record (CB1 lower) represents a water depth of 2195 m, whereas the 1989–1991 deep trap series originate from 3502–3562 m water depth. This complicates the comparison of interannual fluxes, particularly organic carbon fluxes (Fig. 3), which are depth-dependent (Suess, 1980; Martin *et al.,* 1987). Organic carbon fluxes are shown in Figure 3. In the deeper waters they show a marked seasonality in 1988 and 1991 and much lower seasonality in 1989. Annually, highest carbon fluxes were measured in 1991, lowest values from 1988 to 1989 (Table 2). During 1991, we observed a strong seasonal organic carbon flux pattern with prominent peaks in winter, spring and summer appearing at both depth levels (Fig. 3). The summer maximum was significantly reduced in magnitude in the deeper part of the water column and was delayed by about 10 days (Fig. 3). This translates into relatively high sinking speeds for the organic matter fraction on the order of 280 m d^{-1}. Information about the

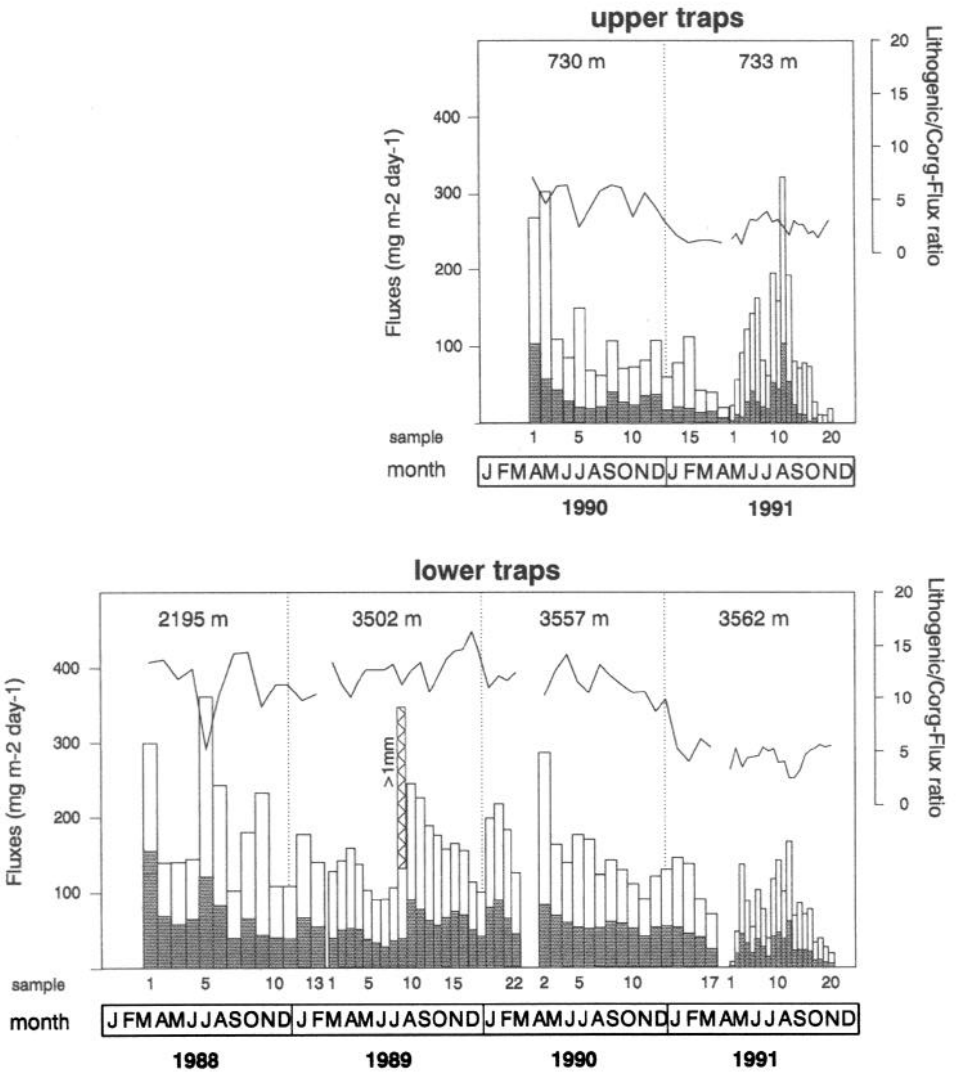

Figure 2. Seasonal and interannual total and lithogenic fluxes (shaded bars) (< 1 mm size fraction; upper traps: 1990–1991; lower traps: 1988–1991). In summer 1989, a distinct pteropod shell sedimentation pulse, with large-sized (> 1 mm) empty *L. inflata* and *C. pyramidata* occurred (Kalberer *et al.*, 1993). Lithogenic/C_{org} flux ratios are also shown.

source of the organic matter can be derived from the stable carbon isotope composition and the C/N-ratios. Both parameters are shown superimposed on the organic carbon fluxes in Figure 3. The $\delta^{13}C_{org}$-values varied between −19.3 and −22‰, the C/N-ratios (atom) between 5.4 and 11.7. During or just before high organic carbon sedimentation, e.g. in the summer of 1988 and spring of 1990 (upper trap) and summer of 1991 (upper and lower traps), respectively, the $\delta^{13}C_{org}$-values increased

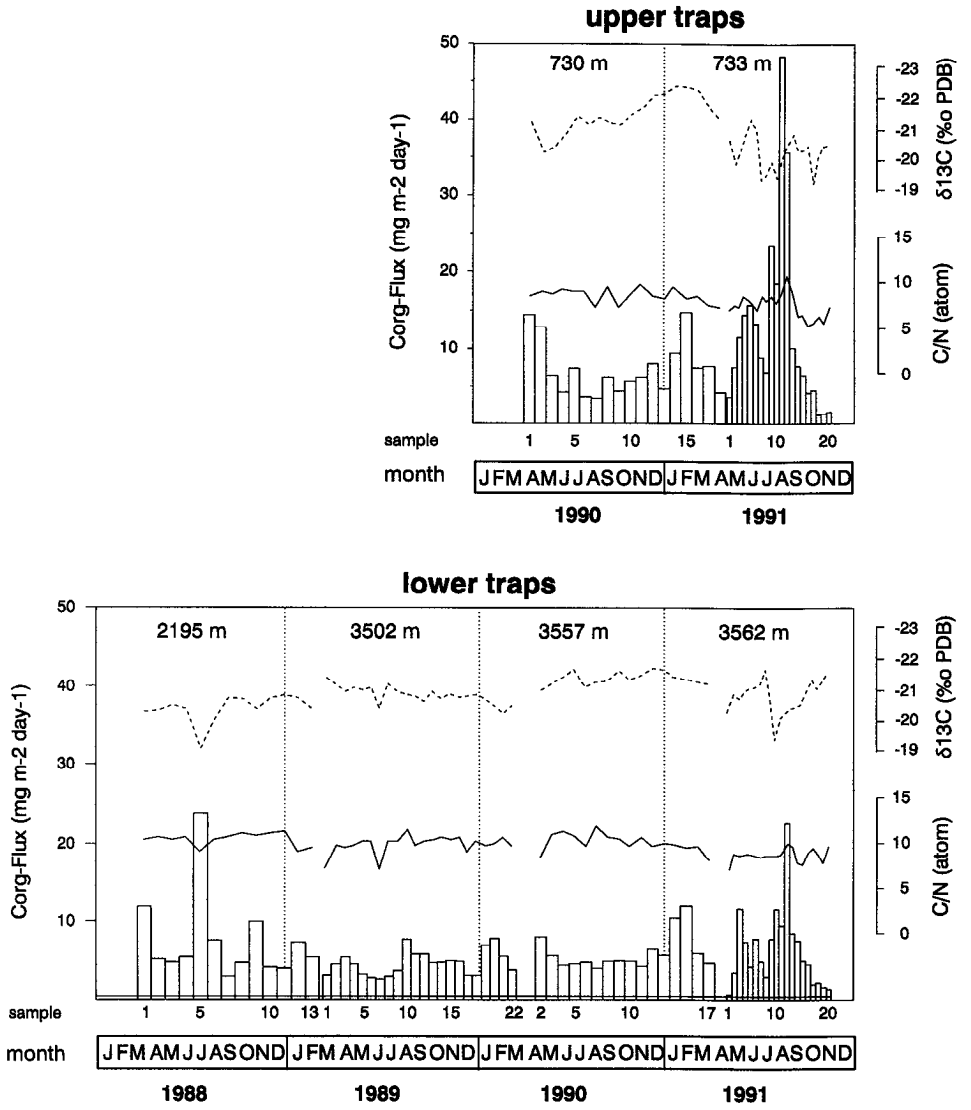

Figure 3. Seasonal fluxes of organic carbon (< 1 mm size fraction). C/N-ratios (atom) and the stable carbon isotope composition of the organic matter are superimposed.

sharply by 1.5–2‰. The stable isotope composition and C/N-ratios do not show any correlation. Rather, the isotope ratios are better correlated to the organic carbon fluxes although the correlation coefficient is not highly significant ($r^2 = 0.31, N = 72$).

Fluxes of skeletal material like biogenic opal and carbonate should not significantly change between 2200 (deployment 1988) and 3500 m (other deployments) water depth except the aragonite fluxes. Carbonate fluxes (Fig. 4a) parallel the total

Table 2. Annual fluxes, C/N, C_{org}/opal, carbon rain and stable carbon isotope ratios from 1988 to 1991 (< 1 mm size fraction).

			FLUXES in g m^{-2}							RATIOS		
Cape Blanc	Trap name	Trap depth m	Total	Opal	Corg	Nges	Carbo-nate	Litho-genics	C/N atom	Corg/opal wt/wt	Corg/ C(CaCO3)	Delta 13C ‰ PDB
Upper traps												
1988	no data	—	—	—	—		—	—	—	—	—	—
1989	no data	—	—	—	—		—	—	—	—	—	—
1990	CB3u	730	32,5	1,2	1,8	0,24	17,8	9,8	8,6	1,5	0,8	−21,3
1991	CB4u	733	26,8	1,4	3,4	0,49	11,3	8,1	8,1	2,4	2,5	−20,7
Lower traps												
1988	CB1l	2195	53,6	4,4	1,2	0,26	22,9	21,9	9,8	0,5	0,8	−20,1
1989	CB2l	3502	52,1	2,4	1,6	0,19	26,7	19,7	9,6	0,7	0,5	−20,9
1990	CB3l	3557	49,8	2,1	1,7	0,19	25,0	19,4	9,9	0,8	0,6	−21,1
1991	CB4l	3562	25,5	1,3	2,0	0,26	10,8	9,4	8,9	1,5	1,5	−18,5
Ratios lower/upper traps												
1990			1,5	1,8	0,9	0,8	1,4	2,0	1,2	0,5	0,7	0,99
1991			1,0	0,9	0,6	0,5	1,0	1,2	1,1	0,6	0,6	0,89

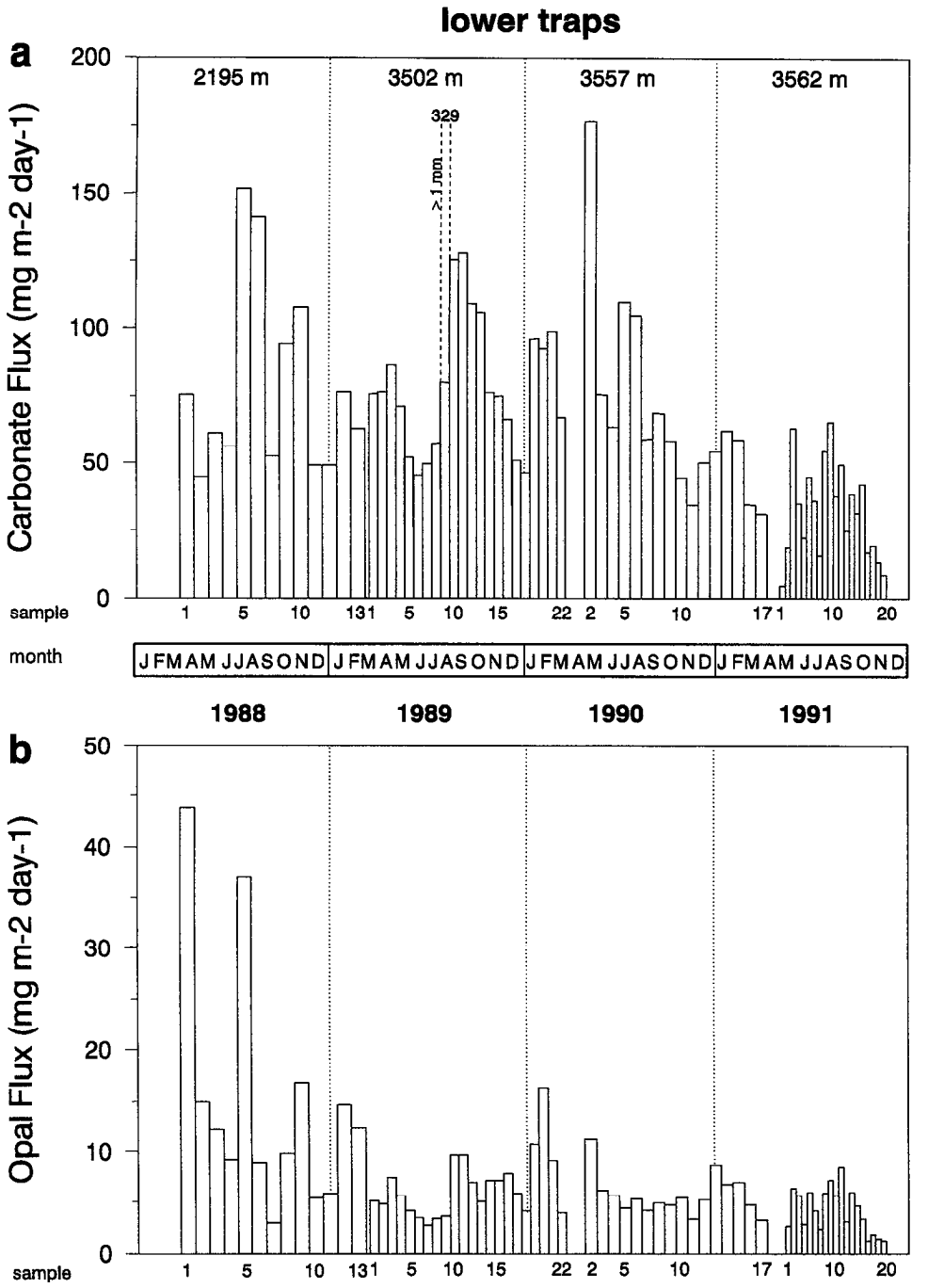

Figure 4. Carbonate (a) and biogenic opal fluxes (b) to the deeper traps (<1 mm size fraction). The broken line in the upper panel (1989) indicates a pteropod sedimentation spike in the >1 mm size fraction. Total carbonate (all sizes) then reached 329 mg m^{-2} day^{-1}.

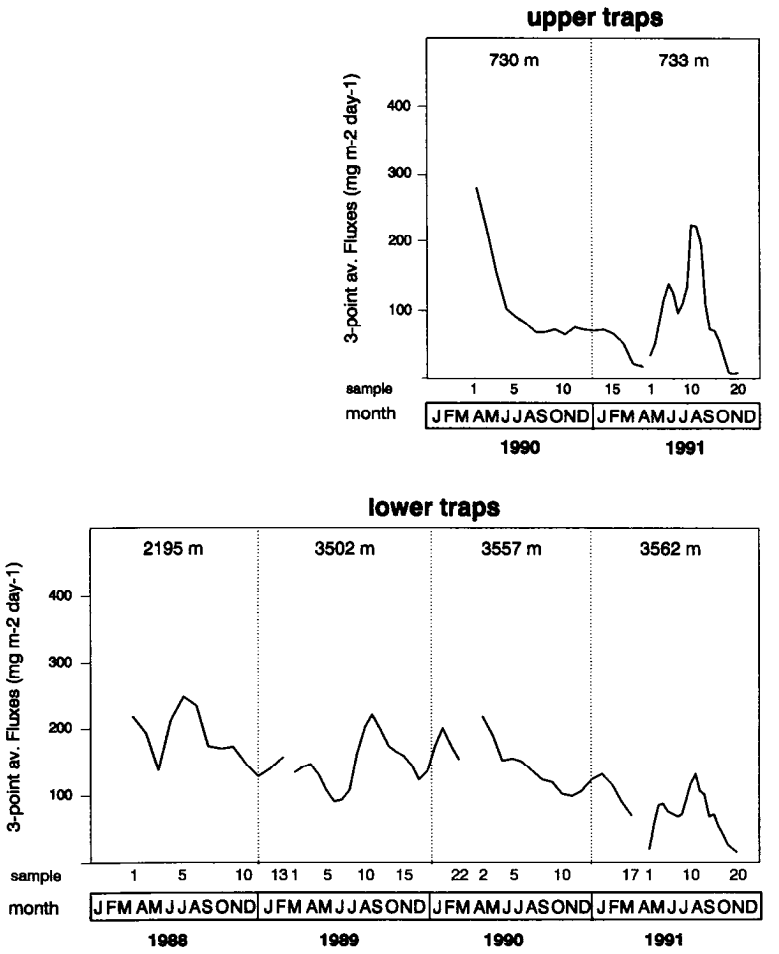

Figure 5. Three-point average total fluxes from 1988 to 1991. Note the general decrease in total fluxes during that period.

fluxes (Fig. 2). Highest opal fluxes with maxima around 44 mg m^{-2} day^{-1} were usually measured in winter and/or spring, decreasing in magnitude from 1988 to 1991 (Fig. 4b). In contrast, carbonate generally peaked in summer, also decreasing in magnitude during the four-year sampling period. Only in 1990 was a notable spring signal measured.

Longer-term changes of sedimentation are evident from three-point averages of total particle fluxes (Fig. 5). We observed a decrease in sedimentation from 1988 to 1991 for total mass and associated carbonate, biogenic opal and lithogenic matter (Table 2). Surprisingly, organic carbon revealed the opposite long-term trend (Fig. 3). The annual fluxes and various ratios for the entire sampling period are listed

in Table 2. The C_{org}/opal and the carbon rain ratio (C_{org}/C_{CaCO_3}) both increased continuously from 0.5 to 1.5, and 0.8 to 1.5, respectively, from 1988 to 1991.

c. Seasonality and interannual SST variability deduced from the $\delta^{18}O$-composition of G. ruber. Planktonic foraminifera are often used to monitor the surface hydrography in the oceans. The stable oxygen isotope ratios measured for these calcite skeletons can be translated to temperature values characteristic of the environment where these protozoans lived before falling into the trap. The white variety of *G. ruber,* for instance, lives in a near-surface environment (see Hemleben *et al.,* 1989 for a review), has a lifetime of less than one month, and persists in relatively high abundances throughout the year (Deuser *et al.,* 1981; Deuser, 1986; Hemleben *et al.,* 1989). Due to rapid sinking, these tests provide excellent records of the near-surface water temperature (Deuser, 1986).

Seasonal and interannual cycles of the $\delta^{18}O$ of *G. ruber* (white variety) are shown in Figure 6 for both trap levels. The white variety revealed a strong seasonality in $\delta^{18}O$, from -2.10 to $-0.36‰$. Higher values, corresponding to lower SST's, generally occurred in winter/spring, whereas lower values (higher SST's) were recorded in summer/fall. This is shown in Table 3 in conjunction with the minima and maxima of the isotope values. In addition, the attributed seasonal SST changes and the average SST's from 1988 to 1991 are listed. From 1989 to 1991, we observed a slight decrease in winter/spring SST from 20.3 to 19.8°C but a much stronger decrease in summer/fall SST from 27.2 to 21.8°C. Consequently, the average annual SST decreased from 23.7°C (1988) to 20.8°C (1991). Seasonality in SST, i.e. the maximum-minimum indicated by the $\delta^{18}O$ of *G. ruber,* was also drastically reduced from 6.9° in 1989 to 2.0°C in 1991. Specimens of the pink variety of *G. ruber* were collected only during warmer periods (see Deuser *et al.,* 1981), e.g. in fall 1988 (CB1) and summer/fall 1989 (CB2), respectively, but revealed values relatively close to the white variety. Therefore, these values are not shown here. In Figure 6, the seasonal $\delta^{18}O$-values of *G. ruber* (white) from the upper traps are also shown. As anticipated, the values recorded in the upper traps closely follow the pattern established from the deeper waters. Therefore, we were able to use the upper trap record of 1991 for the computations of seasonal SST variations. During 1991, only a few specimens of *G. ruber* (white) collected in the lower CB4 trap were available. As a consequence, the isotope data set is rather limited during this year and one might speculate that a warm period e.g. in fall 1991 (like in 1989) may have been missed. However, we obtained a four-year isotope curve of the pteropod *L. inflata* (Fischer and Kalberer, unpubl. data) which almost fills all gaps in winter-spring and fall 1991. These results confirm our interpretations concerning the lower seasonality in SST in 1991.

Table 3. Seasonality of the $\delta^{18}O$-values of *G. ruber* (white, deep trap collections) and estimated SST (equation after Craig and Gordon, 1965). Salinity values for the calculation of ocean water isotope composition (according to Craig and Gordon, 1965) were taken from Mittelstaedt (1991). A correction for a disequilibrium offset by *G. ruber* (white) of −0.35‰ was applied (Fairbanks *et al.*, 1980). The seasonal changes of SST during the four-year sampling period are also shown.

			$\delta^{18}O$ (‰ PDB)		Surface salinity (Mettelstaedt, 1991)		$\delta^{18}O$(‰ SMOW) (Craig & Gordon G., 1965)		Calculated SST in °C (Craig & Gordon G., 1965)		Seasonal SST variation	Annual SST average
Year	Warmest month	Coldest month	Min. month	Max. month	ppt	ppt	Min.	Max.	Max.	Min.	°C	°C
1988	Sep	Jan	−2,10 Sep	−0,76 Jan	35,90 Sep	36,30 Jan	0,70 Sep	0,94 Jan	26,9 Sep	21,8 Jan	5,1	24,4
1989	Aug–Oct	Mar–Jun	−1,97 Sep/Oct	−0,47 Apr	36,20 Sep/Oct	36,20 Apr	0,88 Sep/Oct	0,88 Apr	27,2 Sep/Oct	20,3 Apr	6,9	23,7
1990	Aug–Dec	Jan–Jun	−1,68 Nov	−0,53 Feb	36,45 Nov	36,30 Feb	1,03 Nov	0,94 Feb	26,5 Nov	20,8 Feb	5,7	23,6
1991	Aug	May–Jun	−0,99 Aug	−0,36 Jun	35,90 Aug	36,20 Jun	0,70 Aug	0,88 Jun	21,8 Aug	19,8 Jun	2,0	20,8

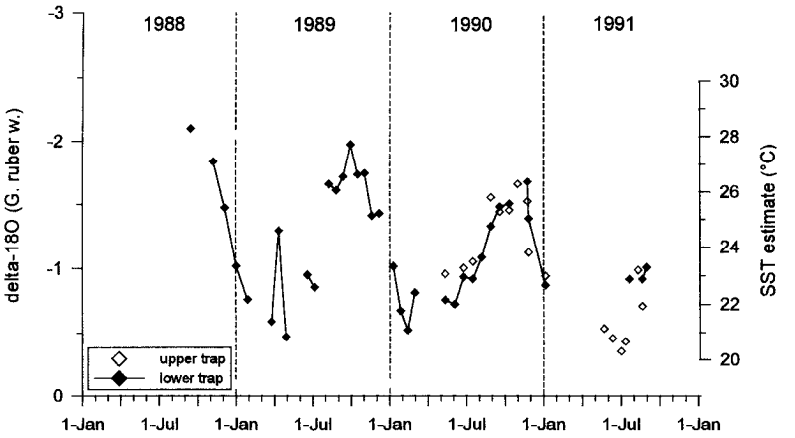

Figure 6. Seasonal and interannual cycles of the $\delta^{18}O$-ratios of *G. ruber* (white variety) from 1988 to 1991. The temperature scale is indicated at the left side. *G. ruber* white was collected rather continuously (although partly in very low numbers) throughout the four-year sampling period. Upper trap (CB3, 4) parallels deeper trap $\delta^{18}O$-values.

5. Discussion

a. Currents and trapping efficiency. Sediment traps do not sample downward fluxes of a certain limited area, but rather collect particles originating at various distances from the trap position. Horizontal current velocities and particle sinking speeds both influence advection and diffusion of particles in the ocean (Siegel *et al.,* 1990). Furthermore, currents in the vicinity of the traps influence trapping efficiency (Baker *et al.,* 1988; Gust *et al.,* 1992). At speeds > 12 cm s^{-1}, the relative trapping efficiency is reduced and the mean size and density of particles increases (Baker *et al.,* 1988). Such speeds were experienced by our shallow traps, but the deeper traps were affected by nearly constant currents and relatively low velocities as well as the consistent directions in the NADW during the four-year sampling period. Thus, these data can be reasonably interpreted and compared with respect to seasonality and interannual variability. In addition, a first approximation of particle sinking speeds occurring in this area resulted in relatively high values of 280 m s^{-1}, probably as a consequence of organic matrix loading by highly abundant lithogenic components, e.g. aeolian quartz grains (see Ramaswamy *et al.,* 1991).

b. Seasonality in SST, related trade wind velocities and fluxes. Coastal upwelling off northwest Africa is largely wind driven and intensified winds should lead to an enrichment in surface nutrients and, as a consequence, to enhanced flux rates. The plots of total mass fluxes from 1988 to 1991 (Figs. 2, 5) show that fluxes were commonly high during spring and (late) summer, and, to a lesser extent, in fall and winter (1988). These flux peaks correspond in general in timing to the occurrence of SST anomalies off Cape Blanc compiled by Speth and Detlefsen (1982; Fig. 1b),

which show highest temperature differences between upwelling and open ocean water in spring (May–June) and fall (September–October) and, to a lesser extent in winter (January). These anomalies give us indication of the amount of cold, nutrient-rich subsurface waters intruded into the photic zone off Cape Blanc, as well as the duration of upwelling which is largely trade wind driven (Wooster *et al.,* 1976; Van Camp *et al.,* 1991). However, SST anomalies from Speth and Detlefsen (1982) are values averaged over longer timescales (1969–1976).

In Figure 7 the wind speeds in the NE trade wind zone measured at Nouadhibou airport (Cape Blanc; data from Global-Telecommunications-System of the World Meteorological Center) are shown for the collection period from 1988 to 1991. Wind directions were generally from east to northeast. A strong seasonality can be recognized, with minima in winter-spring and maxima occurring generally in spring and summer. In summer, a distinct decline in wind speeds can be observed, generally coinciding with an increase in total and carbonate fluxes (Figs. 2, 4a), and low oxygen isotope values reflecting warm SST's (Fig. 6). However, this is not true for 1991 where lowest $\delta^{18}O$-values in summer are associated with high wind velocities. Low seasonality of wind speed, lowest winter and spring/summer wind velocity values and, consequently, lowest average annual wind speeds were measured in 1989. Higher winter values and highest average annual wind speed values were recorded in 1991.

Detailed inspection of Figures 2 and 5 reveals significant seasonal and interannual variability in the magnitude and timing of fluxes. The records indicate that the prominent total flux peaks occurred (1) during 1988 in spring and summer; (2) during 1989 in summer (3); during 1990 in winter and spring; (4) and during 1991 in winter, spring and summer. We observe no general correlation between fluxes and estimated SST's ($\delta^{18}O$ of *G. ruber*). This finding is in contrast to results obtained by Deuser (1986) for the Sargasso Sea where he found an inverse relationship between SST and the amount of particles leaving the euphotic zone. Total fluxes in this open ocean environment varied from 20 to 120 mg m^{-2} day^{-1} which is much lower than the 8.2 to 361 mg m^{-2} day^{-1} measured off Cape Blanc (Fig. 2).

Calcium carbonate in the traps was contributed by primary producers (mostly coccolithophorids) as well as planktonic foraminifera and pteropods. Shells of the aragonite-producing pteropods, which were mostly present in 1989, contributed between 5 and 36% of total carbonate, with summer values reaching 88% of the total carbonate (Kalberer *et al.,* 1993). Foraminifera constituted about one third of the total carbonate. Approximately one third was produced by coccoliths. Carbonate flux maxima (Fig. 4a) associated with higher SST's (Fig. 6) and diminished trade wind speeds (Fig. 7) generally occurred in summer and decreased from 1988 to 1991. This coupling was most evident during 1989 when a distinct carbonate sedimentation pulse (Figs. 2, 4a) coincided with a pteropod flux event in early August (see also Kalberer *et al.,* 1993). This pteropod sedimentation corresponded to a significant

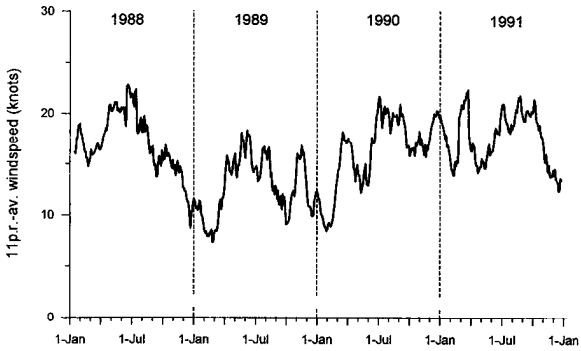

Figure 7. 11-point running average wind speeds (all directions) measured at Nouadhibou airport, Mauretania. Wind directions were mostly from the east and northeast. Data are from the Global-Telecommunication-System of the World Meteorological Center and were provided from the Deutsche Wetterdienst, Hamburg. Relatively low annual average values were measured in 1989. In contrast, almost continuously high values were obtained in 1991.

increase in SST, most probably in the middle of July as recorded by the $\delta^{18}O$-values of planktonic foraminifera (Fig. 6). Moreover, this year was characterized by the lowest trade wind velocities in winter and spring (Fig. 7) and lowest annual values resulting in relatively high offshore SST's.

Biogenic opal flux, contributed mostly by marine diatoms, generally peaked in winter and spring (Fig. 4b), commonly corresponding to high $\delta^{18}O$-values and low SST's (Van Camp *et al.,* 1991) (Fig. 6). The winter and spring opal peaks both decreased clearly, though not continuously, from 1988 to 1991 (Fig. 4b). This may indicate a tendency toward a decreasing supply of silicate-rich subsurface waters via upwelling in winter/spring from 1988 to 1991. On the other hand, we observed increased winter/spring wind speeds (Fig. 7) and a decrease in SST (Fig. 6) suggesting intensified upwelling (Van Camp *et al.,* 1991). Thus, the wind data from 1988–1991 presented in Figure 7 give no indication of a correlation between trade wind velocities in spring and biogenic opal fluxes. A possible explanation may be a change in the source waters for coastal upwelling. The area off Cape Blanc lies on the boundary of the northward-moving nutrient and silica rich SACW and the southward-flowing nutrient poorer NACW (Hughes and Barton, 1974; Zenk *et al.,* 1991). We speculate that this Central Water Boundary, which also separates the superficial Canary Current from warmer tropical waters flowing northwards, was located far north in 1988/1989 compared to 1991 and supplied larger amounts of nutrients (e.g. silica) into this upwelling area.

Organic carbon fluxes generally peaked in winter, spring and summer (Fig. 3). As already mentioned, we assume that the 1988 organic carbon fluxes measured at 2195 m should be higher than at about 3500 m, the depth range where the 1989 to 1991 traps sampled. Therefore, organic carbon sedimentation is assumed to have been highest in 1991; this holds true for the winter, spring and summer peaks, which

increased from 1989 to 1991. As discussed above, there is no correspondence to peak fluxes of carbonate and biogenic opal (Fig. 4a,b), both of which showed decreasing tendencies. During organic carbon sedimentation in summer, a decrease in the C/N ratios (atom) from about 10 to 8 was observed, indicating that rather fresh organic material was transported to the deeper water during those seasons. The stable carbon isotope composition ranged from about −19.1 to −22.5‰, too low for marine organic matter originating at water temperatures above 21°C (see Fontugne and Duplessy, 1981). Rather, these values indicate the influence of terrigenous organic matter, mostly wind transported pollen (S. Jahns, pers. comm., August 1994), which should have significantly lower isotope ratios around −26 to −28‰ (Jasper and Gagosian, 1990). We obtained a significant increase in the $\delta^{13}C_{org}$-ratios with values from about −21.5/−22‰ to −19/−19.5‰ during or immediately before the organic carbon flux maxima in summer (Fig. 3), i.e. during the warmer season, notably in 1988 and 1991. We did not measure significantly higher values during spring, when upwelling was most intense. In general, we found a correlation between the isotope ratios and the organic carbon fluxes from 1988 to 1991 but the correlation coefficient is not highly significant ($r^2 = 0.31$, N = 72). The most likely explanation for the coincidence of high carbon fluxes and high $\delta^{13}C_{org}$-values is that both reflect increased marine organic carbon production, where molecular dissolved $^{12}CO_2$ was rapidly reduced, and relatively more $^{13}CO_2$ was incorporated during photosynthesis (Deuser *et al.,* 1968; Degens *et al.,* 1968).

c. Relation between biogenic and lithogenic fluxes. Particle loading of biogenic materials by lithogenic components has an impact on sedimentation; for example, high amounts of relatively heavy mineral grains may increase sinking rates and reduce organic matter degradation during particle transport (Ramaswamy *et al.,* 1991). Indeed, lithogenic constituents off Cape Blanc, mostly quartz grains of aeolian origin (Wefer and Fischer, 1993) appear to be coupled to the organic carbon fluxes as shown in Figure 8. However, we obtained different regression lines. During 1991, lithogenic fluxes were lower by about a factor of two, but the organic carbon fluxes were significantly higher. This indicates a variable relationship between the fluxes of biogenic and lithogenic components from year-to-year. Ramaswamy *et al.* (1991) found Lith./C_{org}-ratios between 1 and 5 in the Arabian Sea. This range was only found in our upper trap records and in the deeper trap in 1991 (Fig. 2). The 1988–1990 deep water Lith./C_{org}-values oscillated between 5 and 15. The increase with depth during 1990–91 could reflect either organic matter degradation or increased scavenging of lithogenic components by marine particles, e.g. marine snow (Ramaswamy *et al.,* 1991). However, such an increase of lithogenic fluxes with depth which is often found in the ocean was not observed (Fig. 2). Instead, the C_{org}-flux decreased strongly by about 40% (on an annual basis; Table 2) between 733 and 3562 m (Fig. 3).

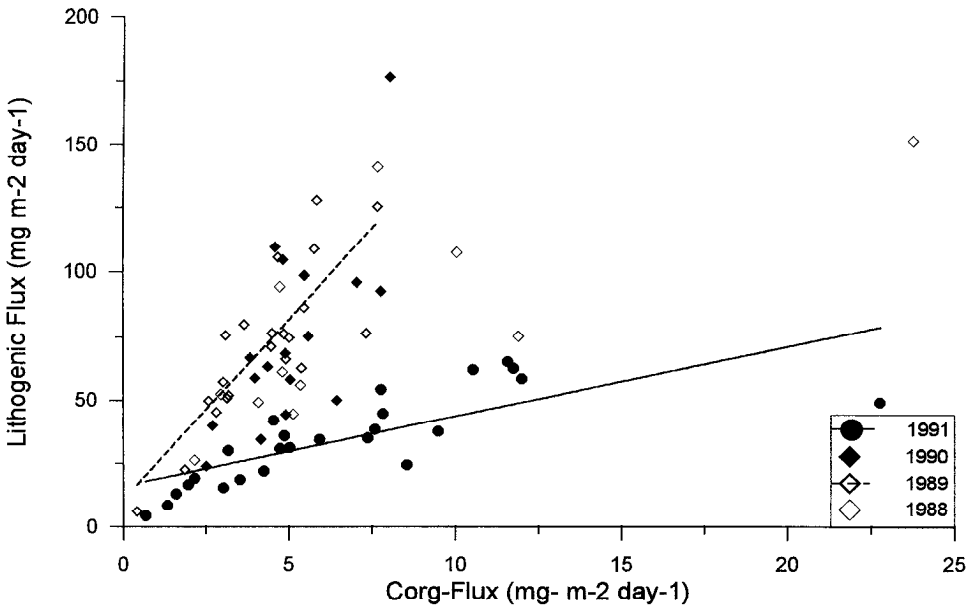

Figure 8. Correlation of lithogenic and organic carbon fluxes for the years 1988 to 1991. Note the different slopes for the regression line of samples collected in 1989 and 1991. The regression lines for 1988 and 1990 are not significantly different from 1989.

d. The long-term SST record. In this section we will discuss changes in SST seasonality based on the $\delta^{18}O$-values of *G. ruber* and *L. inflata* over the four-year sampling period and compare this record to measured values taken from the literature. Year-to-year SST changes and seasonality deduced from the $\delta^{18}O$-values of *G. ruber* are summarized in Table 3. Cold and warm seasons which usually occurred in winter/spring and summer/fall, respectively, differed both in magnitude and timing (Fig. 6). We found a cooling trend from 24.4 to 20.8°C from 1988 to 1991. In addition, a significant difference in SST-seasonality between the two "extreme" years 1989 (ΔT = 6.9°C) and 1991 (ΔT = 2°C) was recorded (Fig. 9). The low seasonality in 1991 is not an artifact of the limited data derived from *G. ruber.* A more continuous isotope curve has been obtained from the pteropod *L. inflata* (Fischer and Kalberer, unpubl. data) which confirms the strong difference in seasonality between 1989 and 1991. The SST distribution was also reflected in the annual total and carbonate flux pattern (Table 2). Annual total fluxes in the warmer years 1988–89 were higher compared to the colder year 1991. This was mainly due to increased carbonate sedimentation in boreal summer, particularly during 1989 when a pteropod-shell sedimentation pulse was detected (Kalberer *et al.,* 1993).

In Figure 10 a comparison of the estimated annual SST values from *G. ruber* to the annual trade wind velocities and the related annual fluxes is presented. Although the expected negative correlation between SST and wind speeds was not observed, the

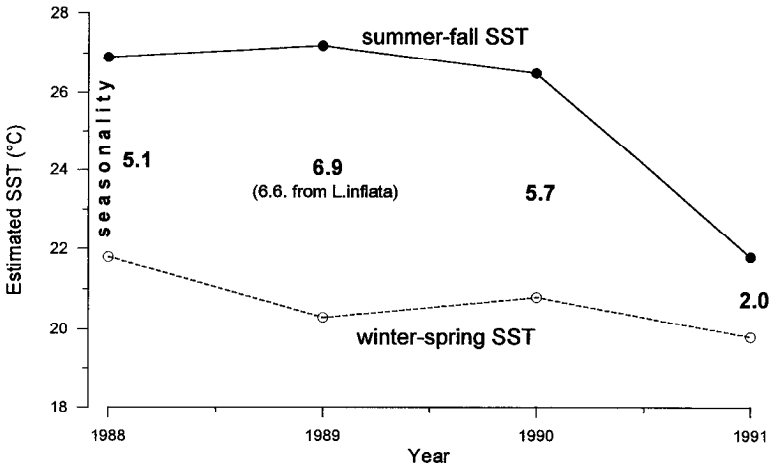

Figure 9. Seasonal and interannual SST changes estimated from the $\delta^{18}O$-record of *G. ruber* (white variety). A −0.35‰ disequilibrium offset for *G. ruber* was applied. Note the general trend in SST decrease with time and the change in seasonality of 6.9° to 2°C from 1989 to 1991 which is not an artifact of missing data points in 1991 (see Fig. 6 and text). The seasonal temperature difference recorded by the pteropod *L. inflata* in 1989 is 6.6°C (Kalberer *et al.*, 1993) and, thus, comparable to 6.9°C estimated from *G. ruber*.

two extreme years 1989 and 1991 are clearly distinguishable: 1989 was characterized by high annual SST's and low wind speeds whereas in 1991 cold SST's and high wind values were measured.

We compared measured SST values from the literature (Mittelstaedt, 1991: historical data from 1906–1977; Van Camp *et al.*, 1991: data from 1987) with SST data calculated from the $\delta^{18}O$-values of *G. ruber* (white; Table 3) and the pteropod *L. inflata* (Kalberer *et al.*, 1993). On average, SST's from the literature (Mittelstaedt, 1991; Van Camp *et al.*, 1991) varied between 19°C in winter/spring and 24° during summer/fall, a range which coincided only partly with our SST computations derived from $\delta^{18}O$ of *G. ruber* (19.8 to 27.2°C; Tab. 3). Specifically, our summer SST's were significantly higher by 2–3°C than measured values. Using higher equilibrium offsets for *G. ruber*, as reported in the literature (Ravelo and Fairbanks, 1992), would not change our results enough to make up for the discrepancy between estimated summer SST's from *G. ruber* and literature data.

Monthly seasonal SST's from 1964 to 1979 integrated over a wide area off Mauritania were compiled by Servain *et al.* (1985). They give a mean SST value for winter/spring (seasonal minimum) around 22°C and a mean value for late summer/fall (seasonal maximum) of 27.5°C. In this case, the minimum SST value obtained for winter/spring was significantly higher than our estimate of 19.8°C for June 1991. Considering these data, we assume that the situation off Cape Blanc in 1991 was characterized by (probably anomalously) cold conditions.

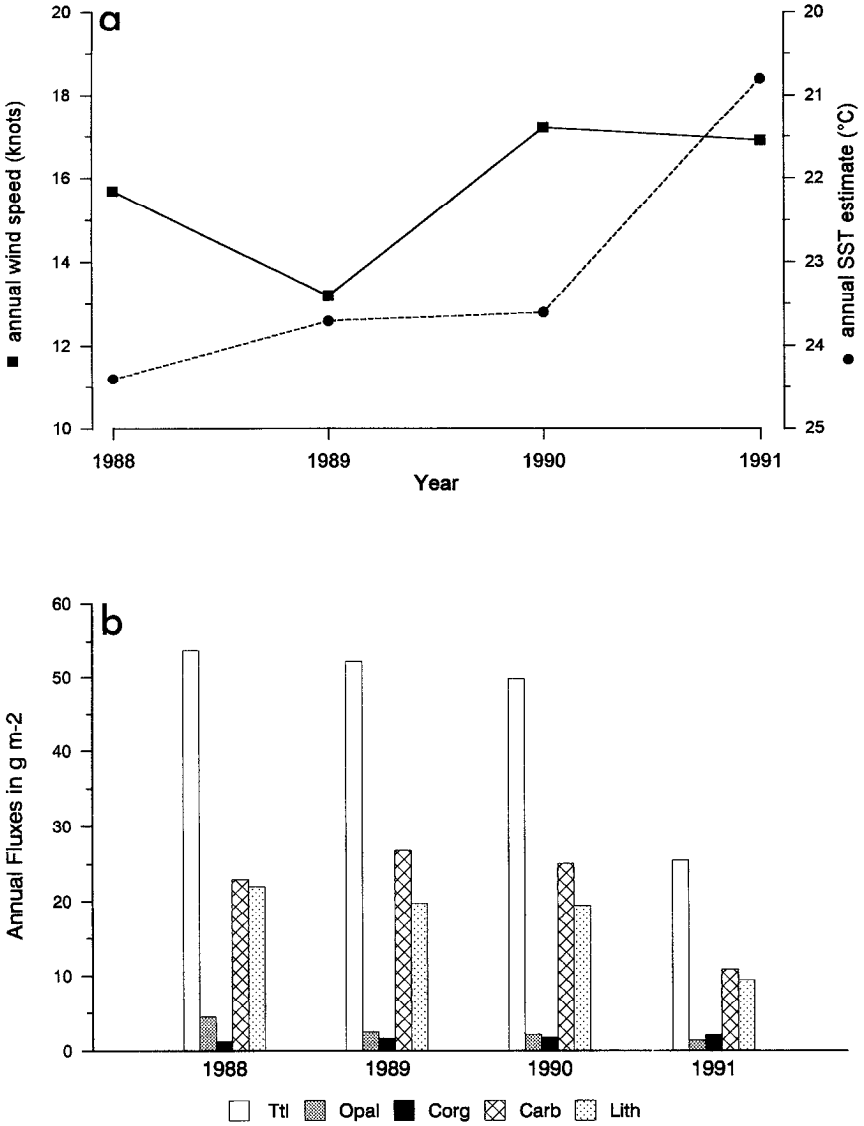

Figure 10. (a) Annual SST variation calculated from *G. ruber* compared to the annual wind velocity recorded at Nouadhibou airport (see Fig. 7). Note that high SST's and low wind values were determined for 1989 whereas low SST's and high wind speeds were obtained for 1991. (b) Annual total, opal, organic carbon and lithogenic fluxes recorded from 1988 to 1991. Note the continuous decrease in total, opal and lithogenic fluxes and the increase in organic carbon fluxes. Highest carbonate fluxes were measured in 1989 during lowest wind velocities/high SST's; highest organic fluxes were determined for 1991 when high wind speeds/low SST's prevailed.

An independent seasonal SST record was provided by young *L. inflata* of 250 μm size which were collected in unusually high numbers in 1989 (Kalberer *et al.,* 1993). This SST curve parallels the pattern exhibited by *G. ruber* during the same year. *L. inflata* migrates in a night-day cycle between 50 and 350 m water depth (Wormuth, 1981) but appears to incorporate carbonate in the shallowest depth range reached at night (Fabry and Deuser, 1991; Kalberer *et al.,* 1993). Therefore, this species should also monitor shallow water SST, probably slightly deeper (around 50 m) than *G. ruber* (0–30 m). Indeed, maximum SST in 1989 is similar, 6.9° for *G. ruber* and 6.6°C for *L. inflata* (Fig. 9) and absolute SST minima and maxima differ only slightly. From the *G. ruber* record we obtained 20.3 and 27.2°C in April and September/October (Tab. 3), respectively, compared to 19.5°C in April and 26.1°C in August derived from the $\delta^{18}O$ values of *L. inflata.*

Corresponding to the general decrease in annual mean SST estimates from 24.4 to 20.8°C (Tab. 3) total, carbonate, opal and lithogenic material fluxes decreased almost continuously from 1988 to 1991 (Fig. 10b; Table 2). Surprisingly, not only the annual opal fluxes decreased, but also the winter/spring peaks (Fig. 4b) which were attributed to increased upwelling of silicate-rich subsurface waters. Although SST's decreased, no indication of increased opal sedimentation was found. On the other hand, we observed an increase in marine organic carbon sedimentation (Figs. 3, 10b, Tab. 2). The organic carbon flux increased from 1989 to 1991 (the 1988 record is from shallower depth) in conjunction with a general SST decrease (Fig. 10) and, obviously, increased upwelling and advection of cold nutrient-rich water is what we might expect from other investigations (see Deuser, 1986, for a review). However, this tendency in the organic carbon sedimentation was not combined with an increasing opal sedimentation. This discrepancy might be due to increased production and sedimentation of non-opal bearing phytoplankters during 1991. A reduction in carbonate fluxes from 1989 to 1991 was mainly an effect of the continuous decrease in summer sedimentation, the period of highest SST's in the seasonal cycle. The interannual changes in carbonate and C_{org} sedimentation resulted in a drastic increase in the annual carbon rain ratio (C_{org}/C_{CaCO3}; Berger and Keir, 1984) from 0.5–0.8 in 1989–90 to 1.5 in 1991 (Table 2), and probably resulted in a more effective "biological pump" in the relatively cold year 1991.

We assume that the long-term $\delta^{13}C_{org}$-record which shows a slight tendency from higher to lower values from 1988 to 1990 (Fig. 3) reflects the increasing CO_2 availability due to decreasing SST's. In 1991, however, the $\delta^{13}C_{org}$-value (calculated on an annual basis with respect to the C_{org}-fluxes) and the C_{org}-fluxes are higher, and the carbon isotope values show a more pronounced seasonality compared to the other years. Organic carbon production and rapid sedimentation during that year may have significantly reduced the CO_2 availability in the surface layer (see Watson *et al.,* 1991; Robertson *et al.,* 1993). In conclusion, the long-term trend in the stable carbon isotope values appears to reflect an overall molecular CO_2 increase due to the

SST decrease or upwelling, whereas the rapid seasonal $\delta^{13}C_{org}$ increases occurring during the warmer season were probably due to biological uptake of CO_2.

On an annual basis, we did not observe a positive correlation between the biological fluxes (i.e. C_{org}) or the carbon rain ratio and the lithogenic fluxes as it was suggested by Ittekkot (1993). Lithogenic fluxes as well as the amount of pollen influx (S. Jahns., unpubl. data, August 1994) were reduced in the "cold" year 1991, whereas the sedimentation of organic carbon and the carbon rain ratios strongly increased. The amount of land-derived material transported to the open ocean is at least partly controlled by the strength of the trade wind system. Figure 7 shows that the annual trade winds were in fact stronger in 1991 and were blowing more continuously. This would lead us to expect a higher land-derived input into the open ocean. This was not the case, however, at least not for the Cape Blanc trap position. We offer two possible explanations: (1) during 1991 much less terrestrial dust was transported upwards into the Saharan Air Layer which later releases its load into the trade winds and, consequently into the ocean, (2) the material transported upward into the higher air layers and later downwards into the deeper trade wind layer was transported beyond the coast of northwest Africa and the Cape Blanc trap position.

6. Conclusions

As described by Servain and Legler (1986), relaxation of the trade winds in response to larger-scale climatic perturbations may cause anomalous warming in the tropical Atlantic in certain years. The best documented episode in the Atlantic with an unusually large amplitude in SST was observed in 1984. At that time, SST's in the tropical Atlantic were anomalously high, the trade winds were weak, and the Intertropical Convergence Zone (ITCZ) was displaced southward from its usual position (Weisberg and Colin, 1986). Off Mauritania (21N, 19W), SST's were also warmest during 1983/84 (McClain and Firestone, 1993).

From our flux and estimated SST data we conclude that 1989 was a relatively warm year off northwest Africa due to diminished NE trades and, probably, a northward-moving warmer counter current, especially from July to October as described by Van Camp *et al.* (1991). The year 1991, however, was characterized by much colder conditions, especially during summer and fall. However, we do not yet know if these year-to-year changes occur in certain periodicities as part of larger-scale climatic oscillations (e.g. El Niño effects) influencing the eastern Atlantic or are more a reflection of regional fluctuations in the extension of the filamental zone off Cape Blanc. Long-term observations of SST's and meteorological data performed off northwest Africa (Picault *et al.,* 1984; Michelchen, 1981) revealed periodicities of about 15, 5–6 and 2 years. These periodicities and unusual (episodic) sedimentation events can only be monitored with longer-term flux studies (e.g. Deuser, 1986). Such investigations will provide information about climatic oscillations and their impact

on upper ocean biology and particle sedimentation into the deep ocean and, the ultimate causes of episodic sedimentation events.

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REFERENCES

Baker, E. T., H. B. Milburn and D. A. Tennant. 1988. Field assessment of sediment trap efficiency under varying flow conditions. J. Mar. Res., *46,* 573–592.

Berger, W. H. and R. Keir. 1984. Glacial-holocene changes in atmospheric CO_2 and the deep-sea record, *in* Processes and Climate Sensitivity. J. E. Hansen and T. Takahashi, eds., Geophys. Monogr. Ser., 29, Washington, D.C., 337–351.

Bricaud, A., A. Morel and J. M. Adre. 1987. Spatial/temporal variability of algal biomass and potential productivity in the Mauritanian upwelling zone as estimated from CZCS data. Adv. Space Res., *7,* 53–62.

Craig, H. and L. I. Gordon. 1965. Deuterium and $\delta^{18}O$ in the ocean and the marine atmosphere, *in* Stable Isotopes in Oceanographic Studies and Paleotemperatures, E. Tongiori, ed., Consiglio Nazionale delle Richerche, Rome 1–122.

Degens, E. T., R. R. L. Guillard, W. M. Sackett and J. A. Hellebust. 1968. Metabolic fractionation of carbon isotopes in marine plankton.—I. Temperature and respiration experiments. Deep-Sea Res., *15,* 1–9.

Deuser, W. G. 1986. Seasonal and interannual variations in deep-water particle fluxes in the Sargasso Sea and their relation to surface hydrography. Deep-Sea Res., *33,* 225–246.

—— 1987. Seasonal variations in isotopic composition and deep-water fluxes of the tests of perennially abundant planktonic foraminifera of the Sargasso Sea: Results from sediment trap collections and their paleoceanographic significance. J. Foram. Res., *17,* 14–27.

Deuser, W. G., E. T. Degens and R. R. L. Guillard. 1968. Carbon isotope relationships between plankton and seawater. Geochim. Cosmochim. Acta, *32,* 657–660.

Deuser, W. G., F. E. Muller-Karger, R. H. Evans, O. B. Brown, W. E. Esias and G. C. Feldman. 1990. Surface ocean colour and deep-ocean carbon flux: how close is the connection? Deep-Sea Res., *37,* 1331–1343.

Deuser, W. G., F. E. Muller-Karger and C. Hemleben. 1988. Temporal variations of particle fluxes in the deep Subtropical North Atlantic: Eulerian versus Lagrangian effects. J. Geophys. Res., *93,* 6857–6862.

Deuser, W. G. and E. H. Ross. 1980. Seasonal changes in the flux of organic carbon to the deep Sargasso Sea. Nature, *283,* 364–365.

—— 1989. Seasonally abundant planktonic foraminifera of the Sargasso Sea: Succession, deep-water fluxes, isotopic compositions, and paleoceanographic implications. J. Foram. Res., *19,* 268–293.

Deuser, W. G., E. H. Ross, C. Hemleben and M. Spindler. 1981. Seasonal changes in species composition, numbers, mass, size and isotopic composition of planktonic foraminifera settling into the deep Sargasso Sea. Paleogeogr., Paleoclimatol., Paleoecol., *33,* 103–127.

Ducklow, H. W. and R. P. Harris. 1993. Introduction to the JGOFS North Atlantic Bloom Experiment. Deep-Sea Res., *40,* 1–8.

Fabry, V. J. and W. G. Deuser. 1991. Aragonite and magnesian calcite fluxes to the deep Sargasso Sea. Deep-Sea Res., *3,* 713–728.

Fairbanks, R. G., P. H. Wiebe and A. W. H. Be. 1980. Vertical distribution and isotopic composition of planktonic foraminifera in the western North Atlantic. Science, *207,* 61–63.

Fischer, G. and G. Wefer. 1991. Sampling, preparation and analysis of marine particulate matter, *in* The Analysis and Characterization of Marine Particles, D. C. Hurd and D. W. Spencer, eds., Geophys. Monogr., *63,* 391–397.

Fontugne, M. and J.-C. Duplessy. 1981. Organic carbon isotope fractionation by marine plankton in the temperature range −1 to 31°C. Earth Planet. Sci. Lett., *41,* 365–371.

Gabric, A. J., L. Garcia, L. Van Camp, L. Nykjaer, W. Eifler and W. Schrimpf. 1993. Offshore export of shelf production in the Cape Blanc (Mauretania) filament as described from coastal zone color scanner imagery. J. Geophys. Res., *98,* 4697–4712.

Ganssen, G. and M. Sarnthein. 1983. Stable isotope composition of foraminifers: the surface and bottom water record of coastal upwelling, *in* Coastal Upwelling, Its Sediment Record, Part A: Responses of the Sedimentary Regime to Present Coastal Upwelling, E. Suess and J. Thiede, eds., 99–121, Plenum Press, New York, London.

Gust, G., R. H. Byrne, R. E. Bernstein, P. R. Betzer and W. Bowles. 1992. Particle fluxes and moving fluids: experience from synchronous trap collections in the Sargasso Sea. Deep-Sea Res., *39,* 107–1083.

Hagen, E. 1981. Mesoscale upwelling variations off the West African coast, *in* Coastal and Estuarine Series. 1. Coastal Upwelling, F. A. Richards, ed., Am. Geophys. Union, Washington. 72–78.

Hagen, E. and R. Schemainda. 1989. Mittlere und jahreszeitliche Strukturen im Unterstrom (UUC) des Auftriebsgebietes vor Nordwestafrika. Beitr. Meeresk., *59,* 19–45.

Hemleben, C., M. Spindler and O. R. Anderson. 1989. Modern Planktonic Foraminifera, Springer Verlag, New York, 1–363.

Honjo, S. 1982. Seasonality and interaction of biogenic and lithogenic particulate flux at the Panama Basin. Science, *218,* 883–884.

—— 1984. Study of ocean fluxes in time and space by bottom-tethered sediment trap arrays: a recommendation, *in* Global Ocean Flux Study, Proceedings of a workshop, Nat. Acad. Press, Washington, D.C. 305–324.

Honjo, S. and K. W. Doherty. 1988. Large scale aperture time-series sediment traps; design, objectives, construction and application. Deep-Sea Res., *35,* 135–149.

Hughes, P. and E. D. Barton. 1974. Stratification and water mass structure in the upwelling area off northwest Africa in April/May 1969. Deep-Sea Res., *21,* 611–628.

Ittekkot, V. 1993. The abiotically driven biological pump in the ocean and short-term fluctuations in atmospheric CO_2 contents. Global Planet. Change, *8,* 17–25.

Ittekkot, V., R. R. Nair, S. Honjo, V. Ramaswami, M. Bartsch, S. Manganini and B. N. Desai. 1991. Enhanced particle fluxes in Bay of Bengal induced by injection of fresh water. Nature, *351,* 385–387.

Jasper, J. P. and R. B. Gagosian. 1990. The sources and deposition of organic matter in the Late Quarternary Pigmy Basin, Gulf of Mexico. Geochim. Cosmochim. Acta, *54,* 1117–1132.

Kalberer, M., G. Fischer, J. Pätzold, B. Donner, M. Segl and G. Wefer. 1993. Seasonal sedimentation and stable isotope records of pteropods off Cape Blanc. Mar. Geol., *113,* 305–320.

Knauer, G. A., D. M. Karl, J. H. Martin and C. N. Hunter. 1984. *In situ* effects of selected preservatives on total carbon, nitrogen, metals collected in sediment traps. J. Mar. Res., *42,* 445–462.

Lee, C., J. I. Hedges, S. G. Wakeham and N. Zhu. 1992. Effectiveness of various treatments in retarding microbial activity in sediment trap material and their effects on the collection of swimmers. Limnol. Oceanogr., *37,* 117–130.

Lohrenz, S. E., G. A. Knauer, V. L. Asper, M. Tuel, A. F. Michaels and A. H. Knap. 1992. Season variability in primary production and particle flux in the northwestern Sargasso Sea: U.S. JGOFS Bermuda Atlantic Time-series Study. Deep-Sea Res., *39,* 1373–1391.

Lyle, M., R. Zahn, F. Prahl, J. Dymond, R. Collier, N. Pisias and E. Suess. 1992. Paleoproductivity and carbon burial across the California Current: the multitracers transect. Paleogeogr., *7,* 251–272.

Martin, J. H., G. A. Knauer, D. M. Karl and W. W. Broenkow. 1987. VERTEX: carbon cycling in the northeast Pacific. Deep-Sea Res., *34,* 267–285.

McClain, C. R., W. E. Esaias, G. C. Feldman, J. Elrod and D. Endres. 1990. Physical and biological processes in the North Atlantic during the first GARP global experiment. J. Geophys. Res., *95,* 18,027–18,048.

McClain, C. R. and J. Firestone. 1993. An investigation of Ekman upwelling in the North Atlantic. J. Geophys. Res., *98,* 12,327–12,339.

Michelchen, N. 1981. Estimates of large-scale atmospheric pressure variations in the upwelling area off Northwest Africa. *in* Coastal and estuarine series 1. Coastal Upwelling, F. A. Richards, ed., Washington, American Geophysical Union, 17–20.

Mittelstaedt, E. 1991. The ocean boundary along the northwest African coast. Prog. Oceanogr., *26,* 307–355.

Müller, P. J. and R. Schneider. 1993. An automated leaching method for the determination of opal in sediments and particulate matter. Deep-Sea Res., *40,* 425–444.

Picault, J., J. Servain, A. J. Busalacchi and M. Seva. 1984. Interannual variability versus seasonal variability in the tropical Atlantic. Geophys. Res. Lett., *11,* 787–790.

Postel, L. 1985. The pelagic cross-shelf and alongshore boundaries of the North West African upwelling region and their annual variability in terms of zooplankton biomass, *in* Simp. Int. Afl. O. Afr. Inst. Pesq., Barcelona 1985, C. Bas, R. Margalef and P. Rubies, eds. 479–487.

Ramaswamy, V., R. R. Nair, S. Manganini, B. Haake and V. Ittekkot. 1991. Lithogenic fluxes to the deep Arabian Sea measured by sediment traps. Deep-Sea Res., *38,* 169–184.

Ravelo, A. C. and R. G. Fairbanks. 1992. Oxygen isotopic composition of multiple species of planktonic foraminifera: recorders of the modern photic zone temperature gradient. Paleoceanogr., *7,* 815–831.

Robertson, J. E., A. J. Watson, C. Langdon, R. D. Ling and J. W. Wood. 1993. Diurnal variation in surface pCO_2 and O_2 at 60N, 20W in the North Atlantic. Deep-Sea Res., *40,* 409–422.

Schemainda, R., Nehring, D. and S. Schulz. 1975. Ozeanologische Untersuchungen zum Produktionspotential der nordwestafrikanischen Wasserauftriebsregion 1970–1973. Geodätische und geophysikalische Veröffentlichungen, *4,* 1–88.

Servain, J. and D. M. Legler. 1986. Empirical orthogonal function analysis of tropical Atlantic sea-surface temperature and wind stress. J. Geophy. Res., *91,* 181–191.

Servain, J., J. Picaut and A. J. Busalacchi. 1985. Interannual and seasonal variability of the tropical Atlantic Ocean depicted by sixteen years of sea-surface temperature and wind stress, *in* Coupled Ocean-Atmosphere Model, J. C. J. Nihoul, ed., Elsevier Oceanography Service 40, Oxford, New York, Toronto, 211–237.

Siegel, D. A., T. C. Granata, A. F. Michaels and T. D. Dickey. 1990. Mesoscale eddy diffusion, particle sinking, and the interpretation of sediment trap data. J. Geophys. Res., *95,* 5305–5311.

Speth, P. and H. Detlefsen. 1982. Meteorological influences on upwelling off Northwest Africa. Rapports et Procès-Verbeaux des Reunions, conseille International pour l'Exploration de la Mer, *180,* 29–34.

Suess, E. 1980. Particulate organic carbon flux in the ocean—surface productivity and oxygen utilization. Nature, *288,* 260–263.

Van Camp, L., L. Nykjaer, E. Mittelstaedt and P. Schlittenhardt. 1991. Upwelling and boundary circulation off Northwest Africa as depicted by infrared and visible satellite observations. Prog. Oceanogr., *26,* 357–402.

Watson, A., C. Robinson, J. E. Robinson, P. E. le B. Williams and M. J. R. Fasham. 1991. Spatial variability in the sink for atmospheric carbon dioxide in the North Atlantic. Nature, *350,* 50–53.

Wefer, G. and G. Fischer. 1993. Seasonal patterns of vertical particle flux in equatorial and coastal upwelling areas of the eastern Atlantic. Deep-Sea Res., *40,* 1613–1645.

Wefer, G., G. Fischer, D. Fütterer and R. Gersonde. 1988. Seasonal particle flux in the Bransfield Strait. Deep-Sea Res., *35,* 891–898.

Weisberg, R. H. and C. Colin. 1986. Equatorial Atlantic Ocean temperature and current variations during 1983 and 1984. Nature, *322,* 240–243.

Wooster, W. S., A. Bakun and D. R. McLain. 1976. The seasonal upwelling cycle along the eastern boundary of the North Atlantic. J. Mar. Res., *34,* 131–141.

Wormuth, J. H. 1981. Vertical distribution and diel migrations of *Euthecosomata* in the northwest Sargasso Sea. Deep-Sea Res., *28A,* 1493–1515.

Zenk, W., B. Klein and M. Schröder. 1991. Cape Verde Frontal Zone. Deep-Sea Res., *38,* 505–530.

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