More free publications from Archimer

OCEANOLOGICA ACTA 1978 - VOL. 1 - N° 2

ART. N° 399
Contribution COB N° 566

Late Quaternary climatic changes in Western Tropical Africa deduced from deep-sea sedimentation off the Niger delta

Pleistocene
Niger
Paleoclimate
Oceanic environment
Pléistocène
Niger
Paléoclimat
Milieu océanique

L. Pastouret [a], H. Chamley [b], G. Delibrias [c], J. C. Duplessy [c]
J. Thiede [d]
[a] Centre Océanologique de Bretagne, BP 337, 29273 Brest.
[b] Laboratoire de Géologie Marine, Centre Universitaire de Marseille-Luminy, 13288 Marseille.
[c] Centre des Faibles Radioactivités, Cnrs, 91190 Gif-sur-Yvette.
[d] Institutt for Geologi, Universitetet i Oslo, Postboks 1047 Blindern, Oslo 3, Norway.

Received 1/12/77, in revised form 22/12/77, accepted 12/1/78.

ABSTRACT

— The oxygen isotopes ratios of benthic foraminifera and detailed radiocarbon ages of the organic matter of an over 15 m long sediment core from the outer Niger delta allow us to date the oxygen isotope stage boundaries 1/2 to 11 500 (± 650) years BP, 2/3 to approximately 23 000 (± 2 000) years BP. The composition of the predominantly terrigenous clays and accessory pelagic fossils reflects the evolution of the climate over the southwestern Sahel zone and the response of the Eastern Tropical Atlantic to these climatic fluctuations during the Late Quaternary.

The dilution of the pelagic fossil concentrations by the terrigenous material and the oxygen isotopes ratios of planktonic foraminifera indicate large fluctuations in the freshwater discharge from the Niger, with high precipitations over the drainage area of this river from 4 500 (± 300) to 11 500 (± 650) years BP and from 11 800 (± 600) to 13 000 (± 600) years BP while the time intervals in between were as dry as today. Relative increase of kaolinite during wet phases and the association of smectite, chlorite and attapulgite during dry ones characterize the response of the weathering in the Niger drainage basins to the climatic fluctuations. The occurrence of 10-14 Å mixed-layers prior to 26 000 years BP is correlated with moderate alteration of the crystalline substratum outcrops from the middle-lower part of the Niger Basin. High quartz concentrations are particularly typical for the transition between oxygen isotope stages 1 and 2 at the inception of heavy precipitations in the southern Sahel zone. Sedimentation rates were quite constant, 30-35 cm/1 000 years; they became unusually large at the beginning of the Holocene from 10 900 (± 650) to 11 500 (± 650) years BP where they reached more than 600 cm/1 000 years.

Bottom waters around 1 100 m depth in the Gulf of Guinea responded to changes in paleo-oceanography of the entire Atlantic Ocean as well as to local influences. Abnormal carbon isotopes ratios and the drastic changes from a highly diversified fauna (during stages 2 and 3, and during the last part of stage 1 after approx. 7 000 years BP) to a

Contribution n° 566 du Département Scientifique du Centre Océanologique de Bretagne.

poorly diversified fauna in the intervening time span point to the development of a local benthic environment which cannot easily be compared with the corresponding continental and slope environments of the entire Atlantic Ocean.

Oceanol. Acta, 1978, **1**, 2, 217-232.

RÉSUMÉ

Variations climatiques du Quaternaire récent en Afrique de l'Ouest tropicale interprétées à partir des sédiments profonds du delta du Niger

Nous avons suivi sur 30 000 ans environ l'alternance phase sèche-phase humide caractéristique du Pléistocène des basses latitudes, ainsi que son empreinte dans l'histoire sédimentaire, à travers l'étude d'une carotte prélevée dans la partie externe du delta du Niger. La courbe des variations des teneurs en isotopes de l'oxygène des tests de foraminifères benthiques a été calibrée par datations au ^{14}C ; les zones de transition entre les stades isotopiques 1, 2 et 2,3 ont pu être datées respectivement de 11 500 ($\pm$ 650) ans BP et de 23000 ($\pm$ 2 000) ans BP environ.

Des précipitations très importantes ont eu lieu entre 4 500 ($\pm$ 300) ans et 11 500 ($\pm$ 650) ans et à un degré moindre entre 11 800 ($\pm$ 600) ans et 13 000 ($\pm$ 600) ans, les périodes intermédiaires étant aussi sèches que la période actuelle. L'augmentation relative de la kaolinite pendant les phases humides et celle de la smectite, la chlorite et l'attapulgite pendant les phases sèches caractérisent la réponse du bassin du Niger aux variations de l'intensité de l'érosion. La présence d'interstratifiés de comportement intermédiaire entre l'illite, la vermiculite et la smectite souligne l'existence avant 26 000 ans d'une période d'érosion intense du socle affleurant dans la région du bassin moyen-inférieur. Le maximum du stade 2 daté de 14 000-16 000 ans se traduit par une période de sécheresse soulignée par l'arrivée de smectite et une réponse bien marquée du milieu océanique (maximum de carbonate de calcium). Les concentrations élevées de quartz localisées dans la zone de transition entre les stades 1 et 2 (11 800-13 000 ans) sont le témoin du début des grandes précipitations survenues dans la région du Sahel. Le taux de sédimentation, relativement constant, de l'ordre de 30-35 cm/1 000 ans, devient exceptionnellement élevé au début de l'Holocène puisqu'il dépasse en moyenne entre 10 900 ($\pm$ 650) ans et 11 500 ($\pm$ 650) ans 600 cm/1 000 ans.

Dans les eaux de surface, les apports d'eau douce originaire du Niger ont influencé la productivité primaire, la concentration des organismes pélagiques et les teneurs en isotopes de l'oxygène des tests de foraminifères planctoniques; la salinité s'est abaissée au cours de l'Holocène de 8 ‰ environ. Les conditions de milieu régnant au niveau du fond ont été fortement influencées par la sédimentation détritique et les apports de matières organiques. Un changement radical s'est produit dans la faune benthique qui, de très diversifiée pendant les stades 3 et 2 et la partie supérieure du stade 1 (postérieurement à 7 000 ans environ), devient peu diversifiée voire pauvre dans les intervalles de temps, et notamment à l'Holocène inférieur. Des rapports isotopiques anormaux du carbone mettent l'accent sur les conditions tout à fait particulières de l'environnement benthique qui ont régné, à certaines époques, dans ce secteur du golfe de Guinée.

Oceanol. Acta, 1978, **1**, 2, 217-232.

INTRODUCTION

The outer Niger delta in the Gulf of Guinea is situated in a climatically very sensitive region of the world close to the boundary between the African tropical rain forest belt and the semi-arid to arid desert belts. While it is known today that the surface water temperatures of large parts of the tropical and subtropical oceans have changed relatively little between glacial and interglacial episodes of the Late Quaternary (CLIMAP Project Members, 1976), a wide range of various data such as lake level (Delibrias *et al.*, 1973; Servant, 1973; Street, Grove, 1976; Grove, Street, 1977) or the distribution of dune systems (Sarnthein, Diester-Haass, 1977) during the same time seem to indicate rapid spatial and temporal changes in the tropical vegetational belts.

This study discusses the response of the adjacent Niger drainage basins and of the tropical East Atlantic to the climatic changes prevailing from the last Glacial to the present Interglacial. We therefore chose a deep-sea core from the outer Niger delta with a very high sedimentation rate. This core was collected during the Walda cruise of R.V "Jean-Charcot" (CH 22 KW 31, 03°31'1N, 05°34'1E, 1 515 cm length, 1 181 m water depth). The sediment consists of dark olive gray (5 Y 3/2) to very dark gray (5 Y 3/1) homogenous mud. This core had been studied in previous papers (Thiede *et al.*, 1974; Pastouret *et al.*, 1974); ^{14}C dates lead us to change stratigraphic and sedimentation rate interpretations as we discuss below.

Since we have to decipher the temporal relationship of any changes in such a core, stratigraphy is of prime concern. We used the oxygen isotopes record of calcareous benthic foraminifera and radiometric ages obtained from the organic carbon content of the sediments to correlate this core with the Late Quaternary chronostratigraphy.

Surimposed on the classical oxygen isotopes record in planktonic foraminifera, variations are observed. They reflect the influence of the Niger freshwater discharges which finally controls the deposition of fine-grained terrigenous material; the large freshwater discharge produces also variations in the composition of the planktonic foraminiferal faunas. The quartz concentrations of the sediment and the composition of the clay mineral assemblages respond to climatic changes through erosional processes over the continental hinterland.

The opal concentrations of the sediments are sensitive indicators of the ocean productivity, while the stable carbon isotopes ratios of planktonic foraminifera relate to salinity changes of the Gulf of Guinea surface waters. The stable carbon isotopes ratios of benthic foraminifera, the preservation of planktonic foraminiferal tests and of other calcareous components, and compositional variations of the foraminiferal faunas are used to characterize the environment close to the sea floor.

STRATIGRAPHY

As a consequence of the vicinity of the continent, core KW 31 is rich in terrigenous components. The calcium carbonate content is generally lower than 10% and the coarse fraction ($\varnothing > 63\ \mu m$) lower than 2% (Thiede *et al.*, 1974). Therefore, classical tools like carbonate content variations or even faunal analysis cannot be applied as no good correlations with cores from the open ocean are evident.

Radiocarbon dates have been obtained on the total organic matter. As the organic carbon content is about 1% all along the core, a special counter requiring only 100 cm^3 of CO_2 was used. The limit of detection corresponds to an age of 26 000 years. Results are reported in Table 1 and Figure 1. They indicate a very high and rapidly changing sedimentation rate as follows:

- 30 cm/1 000 years from the top of the core to 300 cm;
- 640 cm/1 000 years between 300 cm to 700 cm;
- 35 cm/1 000 years between 700 cm to 1 200 cm;
- below 1 200 cm radiocarbon activities cannot be measured, indicating an age greater than 26 000 years BP.

As the sedimentation rate shows large fluctuations, the stratigraphy of this core has been refined through the measurement of $^{18}O/^{16}O$ ratios in benthic foraminifera along the whole section. These ratios are widely used as stratigraphic tools (Shackleton, Opdyke, 1973); during the glacial period isotopically light ice accumulated on the continents leaving the oceans slightly enriched in ^{18}O. All of these variations of ocean water isotopic composition are entirely reflected in benthic foraminifera. Moreover when climate varies, deep water temperatures are unlikely to change as they are controlled by the temperature of Antarctic Bottom Water, which is assumed to have remained close to the freezing point. Therefore isotopic variations of calcareous benthic foraminifera from distant deep sea sediment cores are very similar as they reflect directly the volume of ice stored on the continents. $\delta\ ^{18}O$ maxima and minima can be correlated and have been defined as stages by Emiliani (1955). Tests of two benthic species, *Cibicides pseudoungerianus* (Cushman) and *Osangularia pacifica* (Cushman) have been analyzed according to the procedure used in the Cnrs Gif-sur-Yvette laboratory (Duplessy, 1978). The standard deviation of one analysis, deduced from replicate measurements of the same carbonate, is 0.07‰. Between 100 and 700 cm, *Cibicides pseudoungerianus* was not measured because this species occurred only rarely; *Osangularia pacifica* was therefore analysed.

As already observed for other benthic foraminifera species (Duplessy *et al.*, 1970; 1975) *Osangularia* is 0.33‰ isotopically heavier than *Cibicides*. We therefore added −0.33‰ to *Osangularia* values in order to

Table 1
Results of ^{14}C Ages.

Depth (cm)	Sample number	^{14}C ages years BP
0- 6	GIF-3950	2 660 ± 250
27- 35	GIF-3951	3 670 ± 300
120- 125	GIF-3949	6 750 ± 410
312- 316	GIF-3959	10 900 ± 650
397- 400	GIF-3813	11 200 ± 650
705- 708	GIF-3752	11 500 ± 650
817- 820	GIF-3815	13 600 ± 600
877- 880	GIF-3816	16 100 + 1 300 − 1 100
1 192-1 195	GIF-3930	24 600 + 2 400 − 1 800
1 407-1 410	GIF-3952	⩾ 26 000

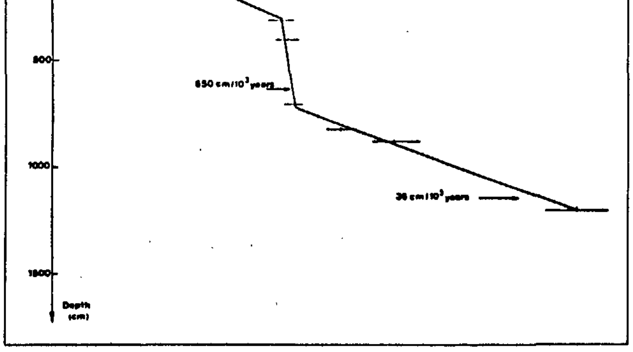


Figure 1
Sedimentation rates in core CH 22 KW 31. ^{14}C ages are plotted against sample depth.

obtain a continuous and comparable record. Oxygen isotopes measurements are reported in Table 2 using the δ notation defined by the relationship:

$$\delta^{18}O = \left[\frac{(^{18}O/^{16}O)\text{ sample}}{(^{18}O/^{16}O)\text{ reference}} - 1\right] \times 1\,000.$$

The standard used is the Chicago PDB standard.

The curve for benthic foraminifera (Fig. 2) shows well defined isotopic stages. They are highly increased due to the unusual high sedimentation rates. From the top to 700 cm, Emiliani's stage 1 characterized by $\delta^{18}O$ values close to the modern ones is correlated with the Holocene. The transition between stages 1 and 2 occurs from 700 to 850 cm and the $\delta^{18}O$ shift is $1.6‰$. This value probably represents a good measurement of the amplitude of the sea water $\delta^{18}O$ change between the Last Glacial and the Holocene as the effects of bioturbation, which homogenize the sediments, are reduced in this core. A similar change in the $^{18}O/^{16}O$ ratio of benthic foraminifera between stages 1 and 2 has been measured in two cores with sedimentation rates higher than 5 cm/1 000 years (Ninkovich, Shackleton, 1975; Duplessy, 1978). A flat, well defined $\delta^{18}O$ maximum is observed between 860 and 900 cm; it corresponds to the glacial conditions of stage 2. The transition between stages 2 and 3 occurs from 1 100 to 1 150 cm. The lower end of the core belongs to isotope stage 3, which is characterized by $\delta^{18}O$ values $0.8‰$ smaller than those of the maximum of stage 2.

A comparison between the benthic record and the ^{14}C dates shows that the sedimentation rate changed in phases with the climatic fluctuations. The major event occurred at the beginning of the Holocene, when the sedimentation rates increased to 640 cm/1 000 years (level 300 cm-700 cm). As the sediments in this core are predominantly of terrigenous origin, the beginning of the post-glacial event might be characterized by a strong increase of the flux of particulate matter brought by the Niger river to the ocean. The tentative stratigraphy originally proposed by Thiede *et al.* (1974) for core KW 31 has therefore to be revised. The bottom of the core falls within isotopic stage 3. If we assume that the sedimentation rate calculated below 700 cm, remained constant before 25 000 years BP, the deepest deposits of this core are 33 000 years old.

The boundary between isotope stages 2 and 3 is located at 1 150 cm; using sedimentation rate its age estimated to 23 000 years BP (± 2 000) is in good correlation with the major ice readvance described as the Catfish Creek Drift in Illinois and Wisconsin by Goldthwait *et al.* (1965), as Iowan or Pinedale glaciation in Midwestern and North-Central United States by Lemke *et al.* (1965), and in Britain (Coope *et al.*, 1971; Coope, 1975). The age of 16 100 years (± 1 300) for the maximum of stage 2 is the age of the highest volume of ice stored on the continents and therefore of the lowest sea level. This figure fits fairly well with field observations, which according to a worldwide compilation made by Emery *et al.* (1971), date the minimum sea level between 15 000 and 18 000 years BP.

PALEOCLIMATIC FLUCTUATIONS IN THE NIGER DRAINAGE BASIN

The temporal relationship between waxing and waning of northern hemisphere ice sheets in the high latitude continental areas and alternation of pluvial and dry periods in low latitude has been the subject of controversy (see Flint, 1971; Grove, Street, 1977; and others). As arid and humid climatic belts have changes when climate varied, a dry period at one place could be synchronous with a wet period at another place and there is *a priori* no general worldwide solution. Moreover this problem is complicated by the fact that the mean climatic evolution (measured by the continental ice volume variations) is only well recorded in the oceans but the consequences of dry and wet periods are essentially well developed on the continents. Accurate correlations between continental ice volume and rainfall intensity variations can therefore be obtained only by the study of deep-sea cores close enough to the continents to record the response of the ocean to climatic variations over the adjacent land mass. Pleistocene sediments from the outer Niger delta offer an opportunity to study glacial-pluvial relationship in the drainage basin of this river by correlating $\delta^{18}O$ stratigraphy, the freshwater discharge and the composition of terrigenous detritus which reflect the climatic regime above this nearby part of the West Africa.

Variations of the Niger freshwater discharge

Continental rainwater even at low latitude are relatively poor in ^{18}O. Thus, marine waters diluted by freshwaters are isotopically more negative than undiluted oceanic waters. As core KW 31 is located on the lee of the Niger river plume, it was expected that oxygen isotopic analysis might produce information more relevant to continental rainfall intensity than to temperature changes which are already known to have been small in this area during the late Quaternary (Duplessy *et al.*, 1974; CLIMAP Project members, 1976). Oxygen isotopic analyses have been performed on the planktonic foraminiferal species *Globigerinoides ruber* pink variety, which was sufficiently abundant in most samples. This species is well known to have a surficial depth habitat and also to tolerate relatively large changes in water salinity (Bé, Tolderlund, 1971). Results are reported in Table 2. The salient feature of the oxygen isotopic record in core KW 31 (Fig. 2) is a major anomaly during most of the Holocene and near the end of the Last Glacial in the transitional zone between stages 1 and 2. This anomaly is superimposed onto an isotopic record (compared with the benthic record) which is characteristic of the open ocean. A peak with $\delta^{18}O$ reaching $-3‰$ extends from 40 to 650 cm and covers the period 4 000-11 500 years BP. This event is interpreted as a consequence of the large freshwater discharge which ended abruptly around 4 000 years. The amplitude of $\delta^{18}O$ variations of *G. ruber* between the core top and stage 2 maximum is close to $2‰$. This value is very similar to that measured in the nearby core CH 7107 (Duplessy *et al.*, 1974), but away from the influence of the Niger river. Thus the modern and the glacial

Table 2

Results of oxygen and carbon isotopic analysis in the tests of Cibicides pseudoungerianus. Osangularia pacifica *and* Globigerinoides ruber. *KR 30, surface sample taken with Reinecke corer.*

Samples depth (cm)	*Cibicides pseudoungerianus*		*Osangularia pacifica*	
	Oxygen	Carbon	Oxygen	Carbon
KR 30 top	+ 2.23	+ 0.57		
22- 25	+ 2.31	+ 0.64		
43- 45	+ 2.32	+ 0.73		
57- 60	+ 2.22	+ 0.62		
62- 65	+ 2.27	+ 0.38		
67- 70	+ 2.30	+ 0.46		
102- 105	+ 2.40	− 0.03		
107- 120	+ 2.72	+ 0.08		
120- 122	+ 2.71	− 0.15		
122- 125	+ 2.72	− 0.11		
82- 125			+ 2.82	− 1.40
142- 180			+ 3.19	− 1.03
202- 220			+ 3.12	− 0.99
207- 260			+ 3.09	− 1.23
287- 340			+ 3.08	− 1.24
347- 360			+ 2.91	− 1.40
422- 440			+ 3.14	− 0.91
480- 485			+ 3.09	− 1.21
517- 520			+ 3.05	− 1.43
520- 525			+ 2.98	− 1.37
540- 545	+ 2.64	− 0.07	+ 2.97	− 1.28
560- 565			+ 2.91	− 1.31
580- 585			+ 3.10	− 1.16
590- 625			+ 3.02	− 1.65
640- 645			+ 2.90	− 1.33
645- 648	+ 2.64	− 0.41		
675- 678	+ 2.56	− 0.28		
680- 685	+ 2.61	− 0.08	+ 2.95	− 1.32
700- 705	+ 2.79	+ 0.36		
720- 725	+ 3.10	+ 0.19		
740- 745	+ 3.11	+ 0.26		
760- 765	+ 2.85	+ 0.16		
780- 785	+ 3.05	+ 0.23		
800- 805	+ 3.20	+ 0.23		
807- 810	+ 3.01	+ 0.23		
820- 825	+ 3.08	+ 0.27		
827- 830	+ 2.93	+ 0.15		
840- 845	+ 3.00	+ 0.22		
847- 850	+ 3.27	+ 0.40		
860- 865	+ 3.85	+ 0.62		
880- 885	+ 3.86	+ 0.61		
900- 905	+ 3.87	+ 0.57		
920- 925	+ 3.62	+ 0.64		
940- 945	+ 3.60	+ 0.49		
960- 965	+ 3.33	+ 0.53		
980- 985	+ 3.22	+ 0.43		
1 000-1 005	+ 3.40	+ 0.49		
1 010-1 015	+ 3.22	+ 0.42		
1 020-1 025	+ 3.60	+ 0.29		
1 030-1 035	+ 3.58	+ 0.34		
1 040-1 045	+ 3.60	+ 0.05		
1 060-1 065	+ 3.23	+ 0.45		
1 080-1 085	+ 3.52	+ 0.41		
1 100-1 105	+ 3.48	+ 0.30		
1 120-1 125	+ 3.16	+ 0.54		
1 140-1 145	+ 3.27	+ 0.32		
1 160-1 165	+ 2.89	+ 0.05		
1 180-1 185	+ 3.06	+ 0.49		
1 200-1 205	+ 3.09	+ 0.43		
1 220-1 225	+ 2.93	+ 0.45		
1 240-1 245	+ 2.97	+ 0.43		
1 260-1 265	+ 3.11	+ 0.50		
1 280-1 285	+ 3.08	+ 0.51		
1 300-1 305	+ 3.03	+ 0.41		
1 320-1 325	+ 3.13	+ 0.38		
1 340-1 345	+ 3.04	+ 0.42		
1 360-1 365	+ 3.12	+ 0.43		
1 380-1 385	+ 3.11	+ 0.41		
1 400-1 405	+ 3.09	+ 0.31		
1 420-1 425	+ 3.13	+ 0.38		
1 442-1 445	+ 3.19	+ 0.39		
1 460-1 465	+ 3.05	+ 0.45		
1 480-1 485	+ 3.14	+ 0.46		
1 502-1 505	+ 3.11	+ 0.38		
1 510-1 515	+ 3.05	+ 0.38		

Samples depth (cm)	*Globigerinoides ruber*		Samples depth (cm)	*Globigerinoides ruber*	
	Oxygen	Carbon		Oxygen	Carbon
KR 30 top	−2.49	+ 1.82	840- 845	−0.60	+ 1.67
bottom	−2.33	+ 1.92	847- 850	−0.69	+ 1.67
20- 25	−2.27	+ 1.61	860- 865	−0.46	+ 1.66
25- 27	−2.31	+ 1.83	880- 885	−0.40	
30- 37	−2.49	+ 1.85	900- 905	−0.60	+ 1.74
37- 40	−2.85	+ 1.76	920- 925	−0.39	+ 1.75
43- 45	−2.77	+ 1.84	940- 945	−0.73	+ 1.63
57- 60	−2.88	+ 1.84	960- 965	−0.39	+ 1.77
62- 70	−3.05	+ 1.45	980- 985	−0.26	+ 1.68
82- 85	−2.73	+ 1.61	1 000-1 005	−0.60	+ 1.76
100-102	−2.99	+ 1.35	1 020-1 025	−0.95	+ 2.01
120-155	−2.90	+ 0.02	1 040-1 045	−0.81	+ 2.07
180-222	−2.83	+ 0.05	1 060-1 065	−0.76	+ 1.67
307-335	−2.98	+ 0.40	1 080-1 085	−1.30	+ 1.67
372-420	−2.55	+ 0.61	1 100-1 105	−1.52	+ 1.85
432-455	−2.34	+ 0.47	1 120-1 125	−0.82	+ 1.79
467-475	−2.77	+ 0.74	1 140-1 145	−0.90	+ 1.89
487-495	−2.52	+ 0.99	1 160-1 165	−1.38	+ 1.92
520-525	−2.35	+ 1.14	1 180-1 185	−1.14	+ 1.87
532-555	−2.54	+ 1.09	1 200-1 205	−0.98	+ 1.82
560-565	−2.30	+ 1.18	1 220-1 225	−1.02	+ 1.85
569-582	−2.90	+ 0.77	1 240-1 245	−1.00	+ 1.82
595-618	−2.69	+ 0.72	1 260-1 265	−0.91	+ 1.88
625-638	−3.02	+ 0.34	1 280-1 285	−0.61	+ 1.77
665-678	−1.92	+ 0.31	1 300-1 305	−0.92	+ 1.90
680-685	−1.79	+ 1.36	1 320-1 325	−0.91	+ 1.91
700-705	−1.12	+ 1.64	1 360-1 365	−0.87	+ 1.89
720-725	−1.32	+ 1.62	1 380-1 385	−0.82	+ 1.90
740-745	−1.79	+ 1.66	1 400-1 405	−0.90	+ 1.76
760-765	−1.70	+ 1.56	1 440-1 445	−0.94	
780-785	−1.86	+ 1.48	1 460-1 465	−1.33	+ 1.85
800-805	−1.63	+ 1.71	1 480-1 485	−1.00	+ 1.88
807-810	−1.53	+ 1.69	1 500-1 505	−0.73	+ 1.78
820-825	−1.33	+ 1.62	1 510-1 515	−0.98	+ 1.93
827-830	−1.46	+ 1.63			

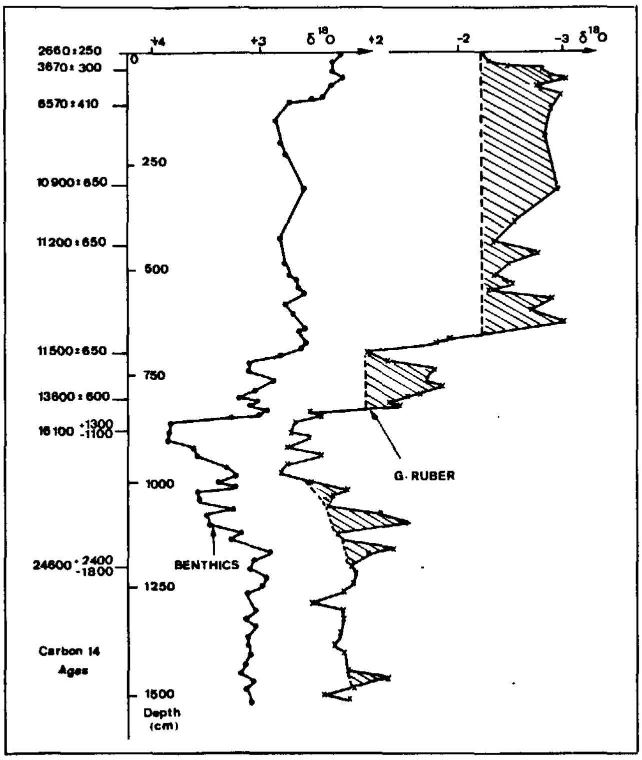


Figure 2

Oxygen isotopic composition of benthic foraminifera (Cibicides pseudoungerianus *and* Osangularia pacifica) *and* Globigerinoides ruber *(pink variety) in core CH 22 KW 31. The composition of* Cibicides pseudoungerianus *was measured in most samples except between 100 and 700 cm in which* Osangularia pacifica *was determined. Values obtained were readjusted, see text p. 219.* ^{14}C *ages are mentionned in the left column. Shaded areas represent fresh-water discharges.*

outflow of the Niger river are very similar. This means the Late Glacial and Holocene pluvial period in West Africa is now finished. A negative isotopic peak is also observed between 740 cm and 780 cm and is dated to about 11 500 to 13 000 years BP. It is synchronous with a humid phase occurring shortly before 12 000 years BP in the tropical Africa (Street, Grove, 1976). It has been correlated to interstadial periods observed in the European climatic records (Alleröd and Bölling). During the period 16 000-30 000 BP, wet episodes of short duration are clearly recorded, mainly at 1 100 and 1 160 cm. The impact of these events which occurred around 22 000 (+ 2 000) years BP is confirmed by an increase in the sediments of terrigenous detrital materials and plant remains (Thiede *et al.*, 1974).

It appears therefore that most of the rainfall over the drainage basin of the Niger river occurred at the beginning of the Holocene and finished abruptly 4 000 years ago. This observation can be correlated with those made in the Chad area (Maley, 1977) and others continental basins where the Holocene lacustral phase ended around 4 800 years BP (Street, Grove, 1976). During the rest of the time recorded in core KW 31, the pluvial regime of the equatorial and tropical western African continent and the Niger discharge appear to have been from the isotopic point of view, close to those observed today. During glacial periods, wet episodes occurred which lasted up to 1 000 years, but at present it does not seem possible to relate them to any important change in the general climatic evolution of the western Africa during stage 2 and the late stage 3.

Significance of the detrital clay supply

Nature and origin of the clay fraction

X-ray diffraction measurements on less than 2 µm non calcareous particles, supplemented by transmission electron microscopy, show a large preponderance of well-shaped but small kaolinite. Numerous other minerals are present: smectite of variable crystallinity, badly crystallized illite, illite-smectite irregular mixed-layers, a 7-10 Å complex including disordered halloysite and kaolinite with probable kaolinite-smectite mixed-layers as well as quartz, occur throughout the core. Chlorite, illite-vermiculite irregular mixed-layers, attapulgite (palygorskite) in short and broken fibres, feldspars, goethite and subamorphous iron oxides occur sporadically (Fig. 3).

Kaolinite reflects the erosion of the wide-outcropping pedologic formations of the north-equatorial and humid tropical African climatic belts. It originates from well-drained lateritic soils (Millot, 1964), chiefly in upstream parts of river basins, together with goethite and sub-amorphous iron oxides which are responsible for the reddish color of clay fractions. These minerals are supplied by the Niger river (Millot, 1953) and are widely distributed in the river delta (Porrenga, 1966; Klingebiel *et al.*, 1975) and in the intertropical marine Atlantic area (Griffin *et al.*, 1968).

Smectite chiefly comes from the lower part of rivers and badly-drained soils of the tropical and sub-tropical climatic belts (Paquet, 1969); this agrees with the blackish color of marine clays (Fig. 3) and continental soils rich in smectite (Paquet, 1969) as well as increasing smectite content in the downstream parts of the Niger river basin (Porrenga, 1966). The irregular 10-14 Å mixed-layers express the incomplete degradation of mica and/or smectite minerals at the bottom of alteration profiles or in zones sparsely covered by vegetation. A pedological origin is also probable for the 7-10 Å complex. This would represent either the onset of continental alteration of volcanic rocks (disordered kaolinite, halloysite: Sieffermann *et al.*, 1968), or the beginning degradation of detrital kaolinite that accumulated downstream in organic-rich zones (kaolinite-smectite mixed-layers: Chamley *et al.*, 1976). It should be noted that the chamosite/berthierine granules formed on the continental margin (Porrenga, 1966; Giresse, Odin, 1973) are not present in the deep-sea clay assemblage.

The other minerals occurring in KW 31 can be directly related to the rocks exposed in the Niger drainage basin. This is the case of the typical primary minerals (illite, chlorite, quartz, feldspars), as well as for attapulgite and a part of smectites which are abundant in Cretaceous and Paleogene deposits (in Millot, 1964; Chamley, 1971). These rare minerals are preserved from complete pedologic destruction, their abundances in sediments increase in the dry zones of northwest Africa (Chamley *et al.*, 1977).

In summary, the clay fraction of KW 31 sediments is marked by the abundance of kaolinite, and by a high mineralogical diversity. These minerals come mainly from soils eroded in the upper and lower parts of the Niger drainage basin and also from crystalline and sedimentary rocks.

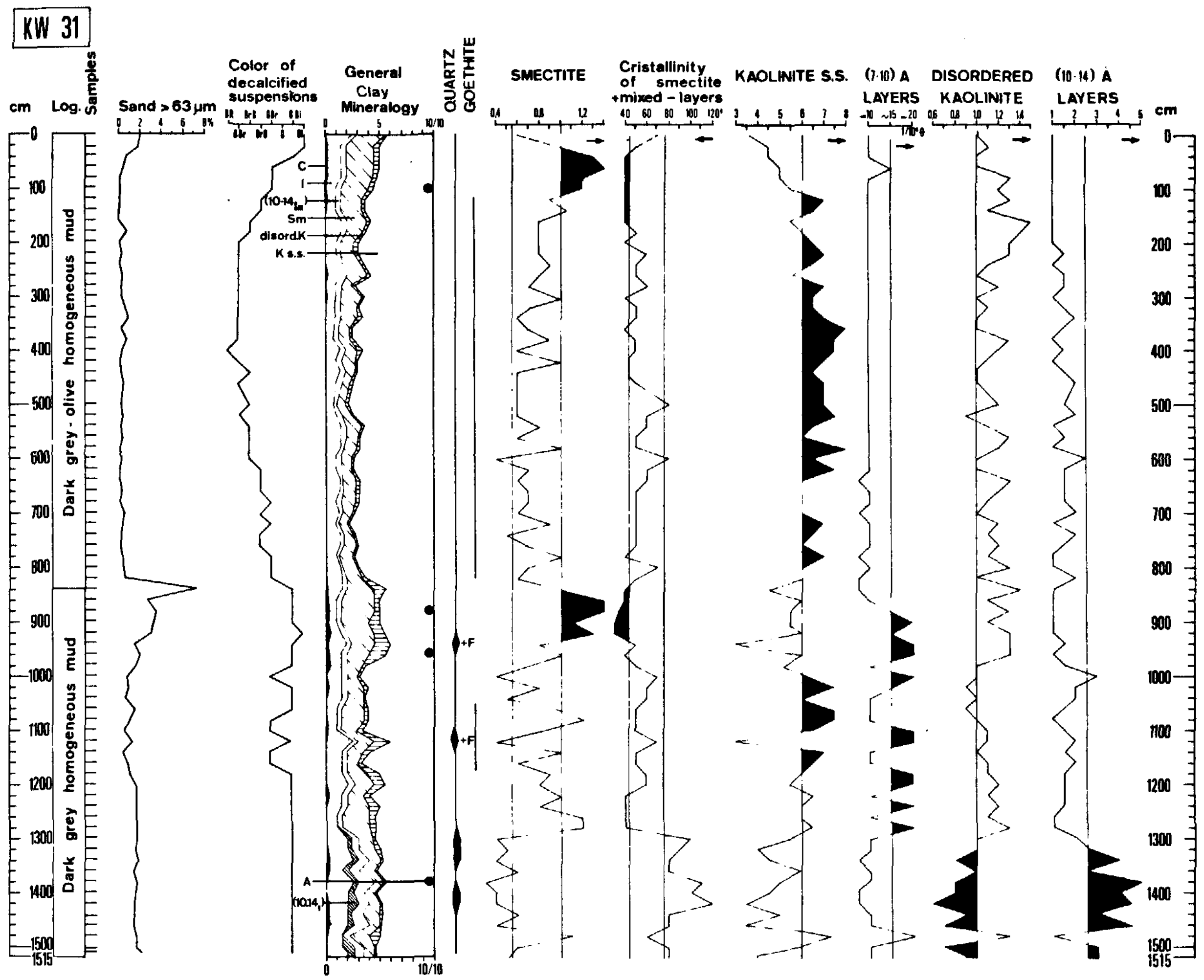


Figure 3

Description, coarse fraction frequency and mineralogy of the clay fraction:

– *Color: R, red; B, beige; Br, brown; G, gray; Bl, black.*

– *Clay minerals: C, chlorite, I, illite;* (10-14_{Sm}), (10-14_{V}), *illite-smectite, illite-vermiculite irregular mixed-layers; Sm, smectite; K, kaolinite; A, attapulgite; F, feldspars.*

– *Smectite abundance:* 18/10 Å *peak height ratio, glycolated sample.*

– *Smectite crystallinity: angle of* 18 Å *peak, glycolated sample.*

– *Kaolinite s.s.:* 7.1/10 Å *peak height ratio, glycolated sample.*

– (7-10) Å *layers:* 8.4 Å *peak height ratio on glycolated and natural sample.*

– *Disordered kaolinite to halloysite: breadth of* ~ 10 Å *reflection in* 1/10 θ, *glycolated sample.*

– (10-14) Å *layers:*

$$\frac{12\text{ Å } \textit{reflections height, natural sample}}{17\text{ Å } \textit{reflections height, glycolated sample}}.$$

→ *Direction of increase.*

Clay mineral variations along the core

Noticeable mineralogical changes occur along the core KW 31 (Fig. 3). The main part of the Holocene sequence, from the transition of Emiliani's stages 1 and 2 at about 840 cm, is marked by a relative increase in kaolinite and goethite contents, shown by the red-brownish or gray-brownish color of the clayey sediments. These minerals and colors chiefly characterize the middle and upper zones of the Niger basin. Their increase is probably due to an augmentation of rainfalls on the river basin, occurring in three ways: *a*) the formation of kaolinite in well-drained soils is favoured by greater wetness and slightly higher temperature during interglacial intervals; *b*) rainfalls produce a more important erosion of the soils which is attested by a sharp increase of the marine sedimentation rate (Fig. 1) and the dilution of marine organisms by detrital clays (Thiede *et al.*, 1974); *c*) the transport energy becomes greater and allows the clay minerals to be transported over the long distance from the distant part of the Niger river drainage basin.

The uppermost part of the Holocene sequence, above 120 cm, shows an increase in smectite content and the presence of chlorite and attapulgite. As smectite chiefly originates from parts of the river basin close to the delta, its increase in marine sediments indicates a decrease of the distance to the source area of the terrigenous detritus. This change in the mineral assemblages was probably caused by a decrease in the continental rainfalls during the Late Holocene. Arguments for such a diminution of the freshwater discharge consist in an augmentation of the concentration of pelagic fossils above 120 cm (Fig. 4 and Thiede *et al.*, 1974) and the augmentation of oxygen isotopic ratio of *Globigerinoides ruber* (Fig. 2).

The isotopic stage 2 between about 840 and 1 180 cm includes two mineral zones. The upper zone between

840 and 940 cm has the same characteristics as the uppermost levels of the core, with a better crystallized smectite and some chlorite and attapulgite in dark gray clayey sediments. The interpretation is the same: a relatively dry climate over the continent led to the diminution of hydrolysis processes and favoured the erosion and then the transport of coastal soils and rock formations. This is supported by plankton isotopic data (part 3, Fig. 2), indicating a minimum of freshwater discharge. As shown by benthic foraminiferal isotope data, the smectite-rich levels correspond to the full glacial conditions: the increase of smectite could be enhanced by a lowering of the sea level at the end of Last Glacial period, causing increased erosion of the coastal formations rich in smectite.

The lower part of isotopic stage 2 shows an alternation of levels rich in kaolinite and levels rich in a 7-10 A irregular complex. This could be interpreted as an augmentation in continental humidity, but not as great as during the Holocene. This interpretation is supported by the isotopic data. The kaolinite increases indicate short periods of heavy rainfalls. As stated above the 7-10 A complex increases are not yet well explained. Nevertheless their occurrence seems to express an alternation of wet and dry periods, allowing the existence of strong rainfalls in the volcanic areas of the Niger basin (eastern part?) and/or the temporary accumulation of kaolinite in downstream badly drained zones (delta?).

In the lower part of the core below 1 180 cm corresponding to the rather constant isotopic stage 3 two zones have been identified. The upper zone between 1 180 and 1 300 cm looks like the one directly above, but kaolinite-rich levels are less marked while 7-10 A complex levels are more marked: the mean continental humidity appears a little lower than above.

The base of the core below 1 300 cm shows an increase in irregular 10-14 A mixed-layers, illite and quartz, the local occurrence of chlorite, and a decrease in smectite and kaolinite crystallinity and abundance. This mineral assemblage is not usually described in subequatorial zones, but a distant origin via marine currents is unlikely because southwards the zone influenced by the Congo river is even more equatorial and marked by high contents of kaolinite (Bornhold, 1973). Northwards the mixed-layer minerals coming from mid-latitudinal zones must pass round West Africa, which is not supported by large-scale current data (*in* Emery *et al.*, 1974). Moreover no argument exists for an autochtonous mineralogical alteration: sediments are not rich in organic matter and the open-sea quaternary detritic environments do not favour such phenomena. The 10-14 A mixed-layers and associate minerals (illite, chlorite, quartz) probably originate directly from the african continent. Their assemblage evokes the moderate alteration of the crystalline substratum. Outcrops of this substratum chiefly exist in the middle-lower part of the Niger basin (Benin, Nigeria). Maybe the mineralogical association at the bottom of core KW 31 reflects an increase in rainfalls and erosion in this part of the river basin.

In summary, the clay mineral assemblages from the core KW 31 reflect large shifts in the continental climatic zones over the Niger basin. The Holocene is very humid except during its most Recent period. The upper part of Last Glacial, during isotopic stages 3 and 2, successively suggests moderate rainfalls on middle-lower areas of the basin, a rapid alternation of wet and rather dry periods, and at least a dry and rather cool period.

Quartz distribution

Quartz is one of the most common components of hemipelagic sediments. The source rock geology and the climatic regime above the source area control the amount and grain size of the clastic quartz which can be induced into the water masses covering the continental margins either in aqueous or in aerial suspension (Heath *et al.*, 1974; Molina Cruz, Price, 1977). Quartz concentrations of the sediments of core KW 31 have been determined after removing the calcium carbonate contents of the bulk sediment with acetic acid. The quartz content has been measured by X-ray diffraction applying the technique described by Till and Spears (1969) and now widely used to analyse marine sediments (Heath *et al.*, 1974; Heath *et al.*, 1976; Ellis, 1972).

The quartz content of the KW 31 sediments is approximately 8% in average (weight percent of the carbonate and opal free sediment) throughout the core (Fig. 4, Table 3). Its distribution however is quite irregular with an increase in the 100-400 cm section of the core and a more pronounced maximum in the 700-900 cm section. The two maxima correspond to time spans approximately from 6 000 to 11 000 years BP and from 11 500 to 16 000 years BP. Both intervals are separated by a section which was deposited during a very short time at sedimentation rates as high as > 600 cm/1 000 years. Since both the calcareous and opal remains of the oceanic plankton are at a minimum during this time, and the core was taken from the top of a small morphologic elevation, and since the very fine grained texture of the sediments is virtually unchanged, it can be concluded that this section of the core has recorded the effect of an important paleoclimatic event affecting the availability of suspended fine grained terrigenous detritus. These important maxima of the quartz concentrations are found in core sections which were deposited approximately 7 000-9 000 years BP and around 12 000 years BP. The earlier maximum coincides with a relatively humid climatic regime with probably increased runoff of freshwater due to relative high lake levels in the Chad region (Servant *et al.*, 1969; Street, Grove, 1976). The older maximum is much more difficult to interpret since such a clear relationship does not exist. Nevertheless, it seems important to notice that these high quartz concentrations have been deposited following the peak of the Last Glacial when an arid climate prevailed over large parts of Africa and when this climate had changed wide regions into deserts with sand dunes (Sarnthein, 1977). This wet period of the stage 1/2 transition can be correlated with a strong decrease of the oxygen isotopes ratios of planktonic foraminifera and has been interpreted to correspond to the interstadial periods of the european realm (Alleröd and

Bölling). The precipitations must have washed large amounts of terrigenous materials with relatively high concentrations of quartz into the Gulf of Guinea. The quartz concentrations decrease drastically at 11 500 years BP when during a time of heavy rainfalls (see discussion above) and humid climate the supplies of the KW 31 terrigenous components were dominated by clay minerals.

RESPONSE OF THE OCEANIC ENVIRONMENT

Climatic and environmental conditions seem to have been quite stable during the last 125 000 years in the eastern tropical Atlantic (Duplessy *et al.*, 1974; CLIMAP Project Members, 1976; Gardner, Hays, 1976). Changes in composition detected in core KW 31 thus reflect the response of the local oceanic environment to the changes of the Niger discharge, which reflect the variations of the intensity of rainfalls on the continents.

Surface waters

Salinity reduction

$\delta^{18}O$ variations of *Globigerinoides ruber* during the period 4 000-11 500 years BP have been interpreted as reflecting a large influx of freshwater. This phenomenon implies a high reduction of the salinity of the surface water where *Globigerinoides ruber* is living. An independant check can be obtained by measuring the $\delta^{13}C$ variations of this species. The following mechanism is implied: $\delta^{13}C$ values of total dissolved CO_2 in freshwaters are much smaller than the marine ones. This depletion in heavy carbon isotopes reflects the fact that about 65% of groundwater CO_2 is produced by the oxidation of organic matter (Labeyrie *et al.*, 1967)

Table 3

Quartz and opal concentrations in core CH 22/KW 31. Quartz, weight percent of the carbonate and opal free sediment. Opal, weight percent of the carbonate free sediment.

Sediment depth from top (cm)	Opal (%)	Quartz (%)	Sediment depth from top (cm)	Opal (%)	Quartz (%)
17	1.98	6.33	727	2.39	
32	6.81	6.19	747	3.50	13.92
47	4.24	5.60	767	2.74	11.92
72	9.70	6.29	787	3.37	14.01
87	10.48	6.65	807	2.39	12.47
107	10.34	6.44	827	3.50	10.08
127	9.95	7.07	837	2.86	14.17
147	6.54	6.87	857	5.29	10.52
167	7.02	8.28	877	5.67	8.43
187	3.19	7.71	897	2.54	8.76
197	3.02	8.08	917	3.67	7.75
227	6.24	13.90	937	2.15	7.02
247	1.54	8.10	957	6.51	6.59
267	2.51	7.96	977	5.67	6.75
287	1.69	7.46	997	4.26	5.75
307	3.64	7.29	1 017	5.97	7.34
325	3.61	7.72	1 030	6.07	6.80
347	4.03	8.02	1 080	5.80	7.31
367	4.71	8.33	1 125	6.17	6.78
387	4.34	6.92	1 150	3.25	7.35
407	5.16	6.89	1 177	2.71	15.54
427	2.41	7.93	1 197	2.12	6.61
437	1.48	7.66	1 217	2.98	7.01
457	2.63	7.31	1 237	3.37	8.12
477	3.17	7.00	1 257	2.51	7.57
497	2.52	6.81	1 277	1.67	7.44
517	3.03	6.64	1 297	2.20	7.12
537	5.28	6.84	1 317	6.06	7.89
547	6.81	6.10	1 337	5.69	7.72
557	6.08	6.75	1 357	4.79	7.87
585	2.67	7.15	1 377	6.89	7.93
605	2.37	7.66	1 397	7.95	7.55
611	2.71	6.17	1 417	6.48	7.37
625	1.08	6.98	1 437	4.76	7.72
645	1.06	9.86	1 457	5.09	7.21
665	1.02	9.18	1 477	4.50	7.82
685	3.25	5.97	1 497	1.53	7.52
705	2.78	10.83	1 507	2.33	7.74

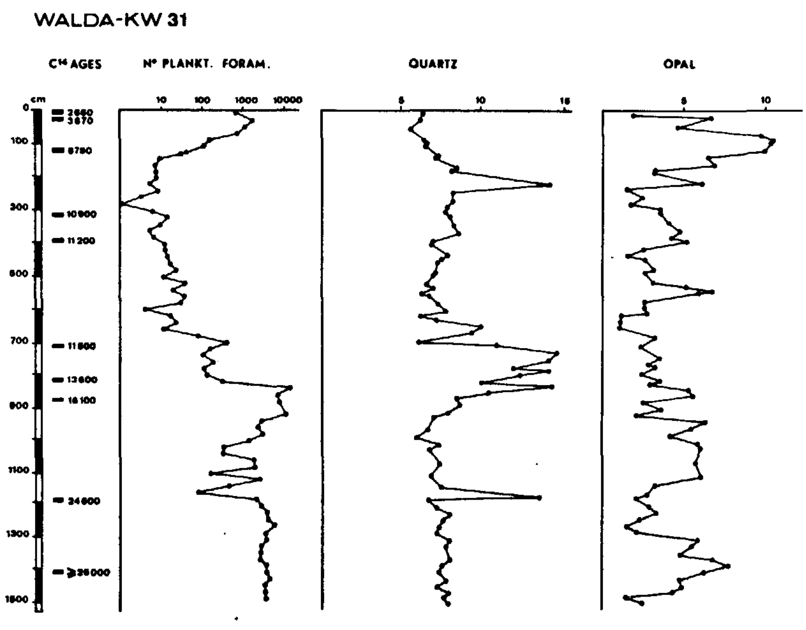


Figure 4

Abundance of planktonic foraminifera in number of specimens per gramme of dry sediment, quartz and opal distributions. Quartz: weight percent of the carbonate and opal free sediment. Opal: weight percent of the carbonate free sediment.

which has, on the continent, a $\delta^{13}C$ close to $-25‰$. Total CO_2 dissolved in sea water has a $\delta^{13}C$ of about $+2‰$, due to the isotopic exchange with atmospheric CO_2 (Duplessy, 1972). Therefore, in estuary and coastal environment, the mixing of fresh and marine waters is reflected in the isotopic composition of the total dissolved CO_2: $\delta^{13}C$ increases with salinity according to an hyperbolic relationship which is fully defined when total CO_2 concentrations and isotopic ratios are known for both sea and freshwaters. This relationship has been determined for the present day Niger delta (Duplessy, 1972).

$\delta^{13}C$ values of *G. ruber* along core KW 31 remain close to the modern ones during most of isotopic stages 2 and 3 (Fig. 5). This confirms that the Niger discharge during the Last Glacial was similar to the modern one. During the Late Pleistocene/Holocene pluvial period, a strong shift towards the smaller $\delta^{13}C$ is observed: the mean $\delta^{13}C$ between 60 and 685 cm is $+0.81$ per mil lighter than the mean modern value. Assuming that the present $\delta^{13}C$-salinity relationship can be applied to the pluvial period, we can estimate a mean salinity reduction of $8‰$ during the pluvial period. During paroxismal events, such as that recorded around 200 cm, the salinity reduction could have been higher than $10‰$, which is also supported by high quartz concentrations (Fig. 4).

Variations in the planktonic foraminiferal fauna

The distribution of the planktonic foraminiferal assemblages has been already discussed in previous papers (Thiede *et al.*, 1974; Pastouret *et al.*, 1974). In the inner part of the Gulf of Guinea freshwater discharge and turbidity have greatly influenced the zooplankton distributions. Correlations of the foraminifer fauna with temperature fluctuations from 30 000 years BP are not obvious, on the other hand species frequencies during stage 2 are similar to those occurring during stage 3 (Fig. 6). Faunal composition changed during Holocene age; frequencies of *Globigerinoides ruber* and *G. sacculifer*, *Globoquadrina dutertrei* and *Globorotalia scitula* increased whereas *Globigerina quinqueloba* became less abundant.

The greatest abundance of *Globigerinoides ruber* (pink and white subspecies) in core CH 22 KW 31 occurs between 100 cm and 700 cm, where averages change from 10 to 20% with a maximum of 30% between 200 and 400 cm. This sediment section is correlated with the lowest salinity value measured along the core. *G. ruber* is a warm water euryhaline epipelagic species (Bé, Tolderlund, 1971). Its distribution in the oceans shows maximum abundance in waters over $36‰$ or below $34.5‰$ salinity (Bé, Tolderlund, 1971; Blanc *et al.*, 1975).

Globoquadrina dutertrei shows a maximum abundance between 80 and 500 cm; its frequency increases from 4 to 8% with a peak to 12% at 270 cm level. On the other hand, a higher frequency located between 620 cm and 820 cm should be noted corresponding approximately to a dilution episode centered on 12 000 years BP. *G. dutertrei* is a subtropical-tropical species indicative of low salinity conditions (Ruddiman, 1971; Cita *et al.*, 1977); it is abundant in Gulf of Guinea sediments (Schott, 1966).

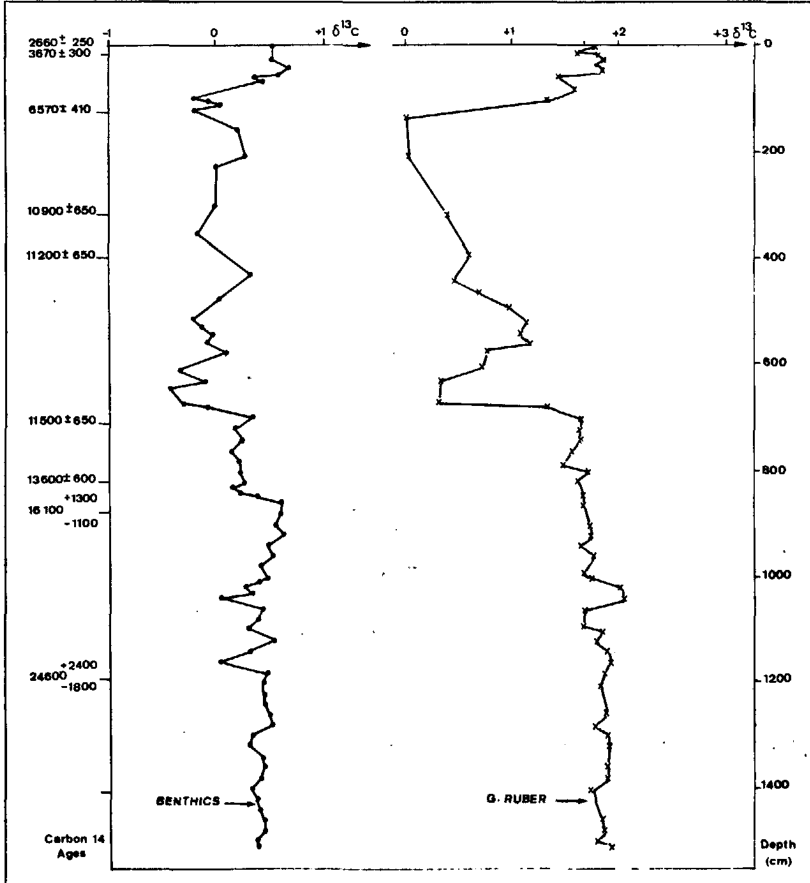


Figure 5

Variations of carbon isotopes ratios in benthic foraminifera (Cibicides pseudoungerianus *and* Osangularia pacifica) *and* Globigerinoides ruber *(pink variety) in core CH 22 KW 31.* Osangularia pacifica *was determined only between 100 and 700 cm (see text). In order to obtain a continuous benthic record we have added + 1.29 to* Osangularia $\delta^{13}C$ *values to correct the specific fractionation of* Osangularia *versus* Cibicides.

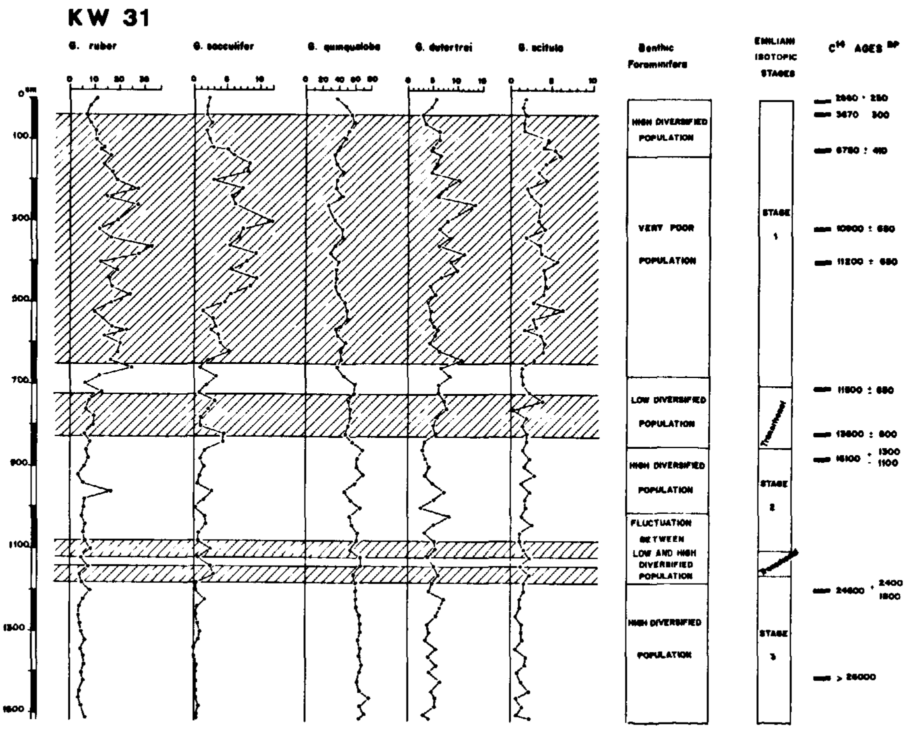


Figure 6
Frequency variations of Globigerinoides ruber, G. sacculifer, Globigerina quinqueloba, Globoquadrina dutertrei, Globorotalia scitula *in core CH 22 KW 31 (frequencies are in percent). The other species are not plotted. Variations of benthic foraminiferal faunas (after Thiede* et al., 1974*). Emiliani's isotopic stages and* ^{14}C *ages are plotted on the right hand side of the figure. The main freshwater discharges are shown by shaded areas.*

Globorotalia scitula is most abundant in the upper part (100 cm to 650 cm) of core KW 31. Its frequency grows from 2 to 4% with a peak to 7% around 150 cm. It is a bathypelagic species usually found in cold water masses: however only few data on its distribution in sediments are available (Schott, 1966). Our results seem to indicate capability to live in low-salinity waters.

The abundance of *Globigerinoides sacculifer* increases during Holocene time and particularly in horizon 120-500 cm in which its frequency reaches 10%. It should be noted however that at the bottom of the core (1 200-1 515 cm) during isotopic stage 3 this species is rare. The increasing abundance of *G. sacculifer* around a 1 160 cm depth, seems to be correlated to an increase of the freshwater discharge. These results would favor the capability for *G. sacculifer* also to tolerate a low-salinity environment. This is not in agreement with the current knowledge about the ecology of this species. *G. sacculifer* is known to be a warm water, epipelagic species in competition with *G. ruber* for a similar ecologic niche (Bé, Tolderlund, 1971; Hecht, 1976) and salinity seems to be the hydrographic variable responsible for the difference observed in the distributions of peak abundances of these species. *G. sacculifer* is believed to be a stenohaline species which occurs more abundantly in water masses with salinity between 34.5 to 36‰. Bé, Tolderlund (1971) have shown that this species is rare in the central water mass of the South Atlantic Ocean and in areas off the coasts of Brazil and Angola. On the other hand *G. sacculifer* is abundant in the Gulf of Guinea where surface water masses with low salinities occur.

Globorotalia quinqueloba frequency is lower between 60 and 650 cm than in the above or below sections of core KW 31. This change could be correlated with low salinity and/or high turbidity waters inputs instead of temperature variations as *G. quinqueloba* frequency did not vary during isotopic stages 2 and 3. This epipelagic species shows a distribution range in cold waters from 1 to 21°C with a maximum abundance at below 12°C (Bé, Tolderlund, 1971). The response of this species to low salinity of Holocene age in core KW 31 is difficult to interpret as no data have been published concerning the behaviour of *G. quinqueloba* toward salinity.

Variations of the productivity

Biogenic opal is found enriched in zones under the productive surface water masses along almost all continental margins (Berger, 1976), although this pattern can be considerably disturbed due to the diluting effect of the high input of terrigenous clastics in such regions (Molina Cruz, Price, 1977). A certain amount of opal silica, however, might be derived from the continents as well, since displaced freshwater diatoms and phytoliths are a common component of hemipelagic sediments (Diester-Haass *et al.*, 1973). From the results of the coarse fraction and smear slide analysis (Thiede *et al.*, 1974) we know that siliceous sponge spicules, radiolarians and diatoms are present in the fine grained, clayey deposits recovered in core KW 31.

The opal content of the sediments has been estimated using the same sample as for quartz measurements after they have been heated to 1 000°C for conversion of the opal silica to cristobalite (Goldberg, 1958; Calvert, 1966; Moore *et al.*, 1973). The samples of core KW 31 have been analyzed twice to assure a good quality of the data. The results of the analyses of the sample pairs were always close together. The average of these pairs of measurements are also presented in Table 2 and Figure 4.

The opal content is approximately 4% in average (weight percent of the carbonate-free sediment) throughout the core (Fig. 4). The best developed opal maximum can be observed at 50-150 cm below the core top in approximately 4 000-7 000 years old sediments. It should be noted that the quartz concentrations drop considerably below their average value during this time span. While opal concentrations are relatively low in the 600-850 cm core section which has been deposited approximately 11 400 to 14 000 years BP and which is characterized by a high input of terrigenous material, the 950-1 200 cm (approximately 17 000-22 000 years BP) and 1 300-1 450 cm ($\geqq$ 26 000 years BP) core sections have revealed two clear though less pronounced maxima of opal silica concentrations. Since the appearance of the sediments recovered in core KW 31 is remarkably homogeneous and in view of the core location, it is clear that compositional variations of the marine fossil assemblages are the result of the changes of the hydrography of the bottom as well as the surface water masses in the inner Gulf of Guinea. The opal contents are probably dominantly composed of diatoms and radiolarians, other remains such as sponge spicules are scarce. Though it is very difficult to quantify the postdepositional dissolution of the opaline fossils (Schrader, 1971; Berger, 1976), the most prominent opal concentrations in this core are assumed to reflect time spans of relatively fertile surface water masses in the Gulf of Guinea. It is especially interesting to note that the most prominent maximum of the opal concentration at the top of the core coincides with a large decrease of the $\delta^{18}O$ ratio of planktonic foraminifera, an increase of this ratio in benthic foraminifera, and an increase of the $\delta^{13}C$ ratios in both benthic and planktonic foraminifera. This zone is also characterized by great changes in the planktonic and benthic foraminiferal faunas. These observations can be correlated with paleoclimatic changes that occurred during this time when the prominent humid phase of the late Holocene gave way to an arid phase approximately 4 000 years ago. It should also be noted that the calcium carbonate contents of the bulk sediment increases from < 10 to > 24% in the 850-1 000 cm section of the core which corresponds to the peak of the last Glacial with a dry climate and low sea level (Faure, Elouard, 1967). Planktonic foraminifera (Fig. 4) as well as pelagic Gastropods occur more frequently, while benthic foraminiferal faunas are highly diversified (Fig. 6).

Variations of bottom environmental conditions

Correlations between $^{13}C/^{12}C$ curves are given on Figure 5 both for benthic *(Cibicides pseudoungerianus, Osangularia pacifica)* and planktonic *(Globigerinoides ruber)* foraminifera. Particularly the sudden decrease of $^{13}C/^{12}C$ ratios at 670 cm is remarquable. Inputs of dissolved CO_2 of continental origin cannot only explain such a variation in the deep environment. The well established thermic stratification of water masses in equatorial zones prevents convection movements and mixing of deep and surface water masses. These variations of bottom environmental conditions could be explained in three ways:

1) a decrease in the global ocean circulation; in this case deep waters would stay longer in contact at the bottom with the sea floor. Chemical reactions in the benthic boundary layers as well as the destruction of organic matter produces CO_2 poor in ^{13}C, but these reactions consume O^2. In this way the depositional environment became impoverished in dissolved oxygen, but rich in CO_2 with a decreasing $^{13}C/^{12}C$ ratio. This kind of partial stagnation of oceanic deep water masses at the bottom was suggested by Weyl (1968) in conjunction with the termination of a glacial stage. Nevertheless it should be noticed that in core KW 31 this stagnation phase did not occur at the end of a glacial period but after the deglacial period, which is not consistent with Weyl's theory. Moreover results obtained on cores from North Atlantic or Indian Ocean (J. C. Duplessy, unpublished data) have shown that $^{13}C/^{12}C$ ratio of benthic foraminifera did not vary greatly during the Holocene period. The phenomenon we observed in core KW 31 is therefore not a general one and must be interpreted in terms of local variations of the marine environment;

2) a local deep-water stagnation. This sort of phenomenon generating sapropel deposits has been observed in Black Sea-type basins (Degens, Hecky, 1974) or in the Eastern Mediterranean Sea (Pastouret, 1970; Ryan, 1972; Ryan, Hsü *et al.*, 1973; Cita *et al.*, 1977). But the observations we made on the site location of core KW 31 and the absence of sapropelic layers in the recovered section led us to eliminate this hypothesis;

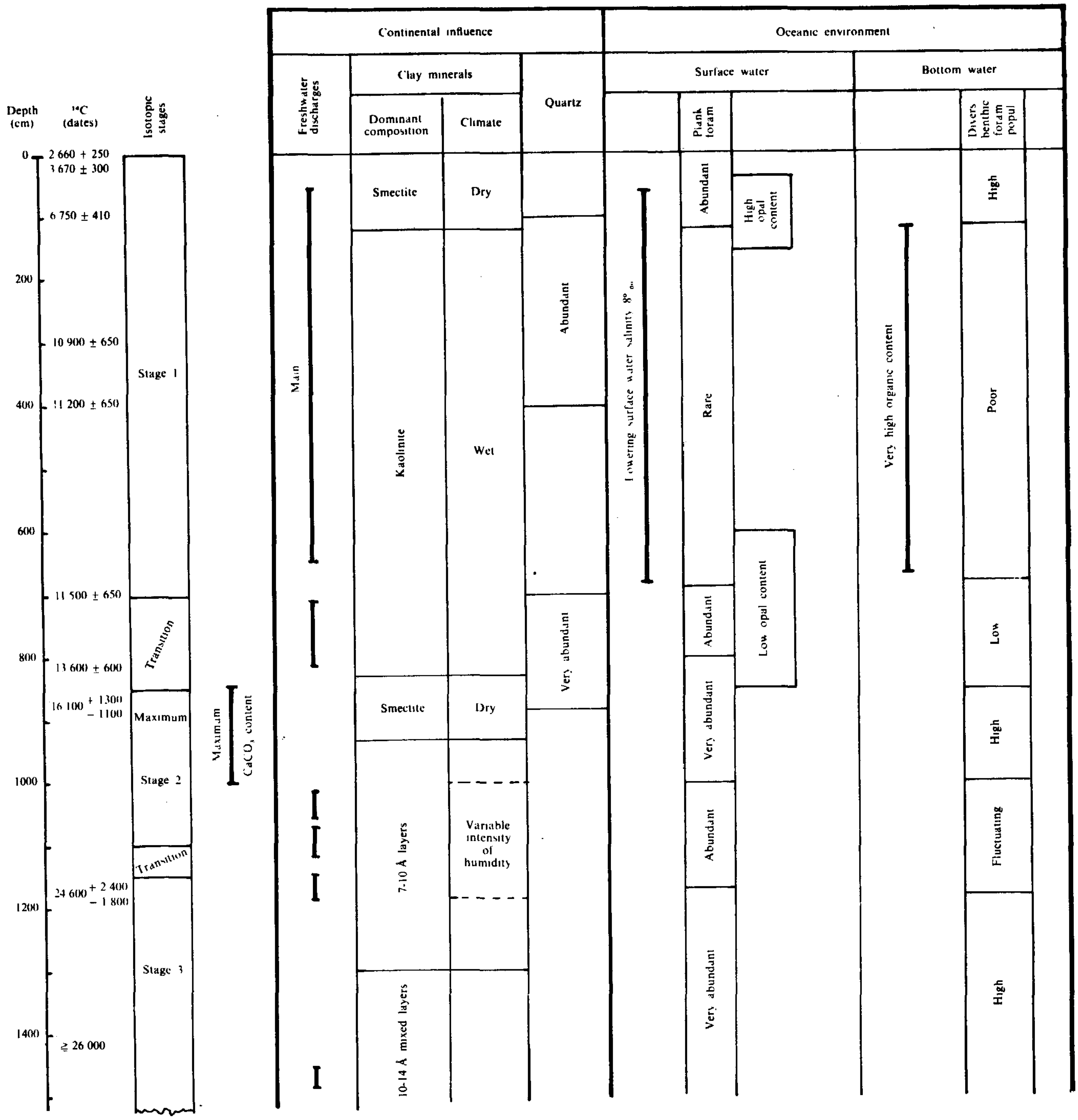
3) a local increase of the benthic CO_2 production generated by oxydation of the organic matter supplied to the ocean by the Niger river. Similar observations have been already described in estuarine environment (Gameson, Barrett, 1958). Richards (1965) pointed out that in marine waters the increases of the organic matter inputs were accompanied by an increase of the dissolved oxygen consumption. This hypothesis could only explain the variations of carbone isotopes ratio we observed.

The depositional environment close to the sea-floor has greatly influenced the benthic foraminiferal faunas; those variations have been previously discussed (Thiede *et al.*, 1974). The most important feature observed in this fauna (Fig. 6) consists in a change from highly diversified assemblage in the lower part of core KW 31 during stages 2 and 3 prior 14 000 years BP, to a low diversified to poor assemblage in the upper part (140-670 cm) of the core, except from 140 cm to the top, that is to say prior to 6 700 years BP. The zone with particularly low diversities between 680 and 850 cm can be correlated with the transitional zone between isotopic stages 1 and 2. Following this impoverishment and at the same time as the sedimentation rates increase as a consequence of the incipient pluvial period on the continental area, the benthic population became very poor in the lowest part of the Holocene (6 750 to 11 500 years BP).

The highly diversified fauna is characterized by the occurrence of the following species: *Euuvigerina peregrina, Bulimina costata, Cassidulina carinata, Cibicides*

Table 4

Sedimentological events recorded in CH 22KW31core



pseudoungerianus, Planulina wuellerstorfi, Gaudryina atlantica, Gyroidina soldanii, Nonion barleanum. Agglutinated species are most abundant between 0 and 140 cm. The poor benthic population comprises mostly *Globobulimina affinis, Bulimina exilis* and *Osangularia pacifica.* The disappearance of *Cibicides pseudoungerianus* between 140 and 680 cm should be noted.

DISCUSSION (Table 4)

The radiocarbon dates and the isotopic record we have obtained on the core KW 31 led us to establish a detailed stratigraphy of the sediment section which permits us to discuss on one hand the succession of dry and wet periods of the late Pleistocene in the Niger drainage basin and their imprint in the marine sedimentation in the inner part of gulf of Guinea and on the other hand, several geological, climatological, paleoenvironmental events occurring as well on the continental area as in the nearby ocean.

The boundaries between Emiliani isotopes stages 1 and 2, 2 and 3 previously dated respectively to 13 000 and 32 000 years BP (Shackleton, Opdyke, 1973) have been precised. The first one was measured to 11 500 (± 650) years BP and the second one estimated to 23 000 years BP (± 2 000). The isotopic stage 2 corresponding to the latest maximum ice sheet extension and the lowest sea-level regression was dated to 16 100 (± 1 300) years. This date is slightly earlier than that proposed by Faure and Elouard (1967) for the western African Coast but in good agreement with the world wide compilation made by Emery *et al.* (1971).

According to the atmospheric circulation pattern proposed by Manabe and Hahn (1977) the last ice age period is correlated in the intertropical Africa with dry climate. The interglacial climatic episodes are marked by increasing runoffs and freshwater discharges in the inner part of Gulf of Guinea as shown by the comparison between $\delta\ ^{18}O$ variations of planktonic and benthic Foraminifera. Freshwater discharges were low between 30 000 and 16 000 years except during short intervals around 22 000 years BP; they increased from 13 600 years and became very large in the Holocene between 11 500 and 4 000 years; then they decreased rapidly. This evolution can be correlated with late Quaternary lake-level fluctuations in central Africa (Street, Grove, 1976) and particularly from the Chad basin (Servant, 1973; Maley, 1977) except for the incipient wet period dated to 13 600 years and which we interpret as being equivalent to the european Bölling-Alleröd stages. Aridity was intense and widespread between 22 000 and 12 500 years in the intertropical zone (Burke *et al.*, 1971), lakes levels were very low, semi-desertic conditions extended and eolian dune systems were well developped. Since 12 000 years, humidity became greater, lakes levels increased. The maximum of the lacustrine phase is dated to 8 000-9 000 years. The dry period still persisting today began around 4 000 years and was interrupted only in areas such as northwestern Sahara, Mauritania, Niger, Chad by short-lived wet episodes between 3 500 and 3 000 years and around 1 800 years. Since 3 000 years desertic conditions persist in the central part of Sahara.

In core KW 31 during isotopic stage 3, 30 000 to 23 000 years, $\delta\ ^{13}C$ was rather constant. This is interpreted as the consequence of low or moderate inputs of organic matter to the ocean (Thiede *et al.*, 1974). This observation can be correlated with the sparse plant cover occurring at this period in the tropical Africa area (Street, Grove, 1976). As shown by our results, increasing plant cover started around 12 500 years and extended probably around 11 500 years.

Dry and wet phases alternations are marked on the continent by differential soils erosion in various geographic areas of the wide drainage basin, such as different composition of the clay mineral assemblages. During humid periods, kaolinite associated with goethite give evidence of erosion in the upstream parts of river basin and in the delta region, whereas smectite associated with chlorite and attapulgite reflect low erosion conditions during dry periods of downstream badly-drained soils. On the other hand the unusual occurrence between 33 000 and 25 500 years of irregular illite-smectite and illite-vermiculite mixed-layers shows alteration of crystalline substratum outcropping in the middle-lower part of the Niger basin (Benin, Nigeria).

The sedimentation rate is generally high ($\geq$ 30 cm/1 000 years) except at the beginning of the Holocene when during 600 years at least it was exceptional ($\geq$ 600 cm/1 000 years). This event appears to be related to increasing rainfalls and could be explained by overflowing of lakes in previously separated drainage basins or by a sudden break of natural dams such as eolian dunes built during the arid period in the northern part of the Niger drainage basin. Nevertheless it seems more satisfying to suggest the occurrence of intense washing of soil formations at the onset of the Holocene pluvial period. The solid transport of a river can greatly vary between dry and wet seasons. For the Rhone river (France) for example the ratio of solid transport is 200 times higher during the pluvial season (Chamley, 1971). Moreover the highest quartz concentrations occurred just after the maximum dry period between 11 500 and 14 000 years before the increasing sedimentation rate.

In the inner part of gulf of Guinea freshwaters inputs rich in organic and particulate matters during humid periods have greatly influenced the oceanic conditions such as productivity and benthic environment. In the surface waters, the fertility decreased, in response to high turbidity and decreasing salinity which was in average $8‰$ lower during Holocene period. Good correlations have been shown between scarcity of planktonic foraminifera and dilution peaks occurring at 23 000 years, 12 000 years and lower part of Holocene. Frequencies variations of some species such as *Globigerinoides ruber*, *G. sacculifer*, *Globoquadrina dutertrei* et *Globorotalia scitula* can be interpreted only in terms of adaptation to environmental conditions. Test fragmentation interpreted as dissolution related to high organic matter content, is present in the entire section (Pastouret *et al.*, 1974). If dissolution had played an important role (disappearance of low resistant species) it would be difficult to explain the quite constant frequency of *Globorotalia quinqueloba* during isotopic stages 3 and 2 and on the other hand, its decreasing abundance during the Holocene since both low resistant species and organic matter inputs became more abundant. Dissolution has probably in that case a minor influence only.

At the bottom, environment conditions are greatly influenced by organic and detrital discharges originating from Niger. Oxygen isotopes data on benthic foraminifera reflect the deep ocean conditions mainly related to ice sheet fluctuations. Carbon isotopes data have put in evidence very high organic carbon and benthic CO_2 production at the base of the Holocene (6 000-11 500 years). This event appears to be related to massive organic matter discharges and has been interpreted as a local effect interesting the outer Niger delta zone. These discharges have influenced the benthic foraminifera populations distributions.

Acknowledgements

We wish to express our grateful thanks to R. Kerbrat who was in charge of the extensive technical laboratory procedures involved and N. Guillo-Uchard for typing the manuscript. The clay mineralogical study was made possible by the financial support provided by the Centre National pour l'Exploitation des Océans and the technical assistance of M. Acquaviva and C. H. Froget. The quartz and opal analyses were carried out by P. Price at the School of Oceanography of Oregon State University, Corvallis, USA. These analyses and J. Thiede's work were effected with the support of CLIMAP (US-NSF grant IDO 75-22133). J. Thiede, in particular, wishes to acknowledge the hospitality of the Centre Océanologique de Bretagne (Cnexo) where this joint work was begun in 1973. A. Beauregard, J. Antignac and B. Le Coat were in charge of the isotopes analytical procedures. G. A. Auffret and G. Pautot read the manuscript and made helpful comments.

We greatly acknowledge Professor Richard Reyment for his remarks and comments.

We became acquainted after the manuscript was submitted with the existence of Nedeco (Netherlands Engineering Consultants) reports concerning investigations on the Niger delta area.

REFERENCES

Bé A. W. H., Tolderlund D. S., 1971. Distribution and ecology of living planktonic foraminifera in surface waters of the Atlantic and Indian Oceans, in *The Micropaleontology of the Oceans*, edited by B. M. Funnell and W. R. Riedel, Cambridge University Press, 105-149.

Berger W. H., 1976. Biogenous deep-sea sediments: Production, preservation and interpretation, in *Chemical Oceanography*, **5** (2nd Ed.), edited by J. P. Riley and R. Chester, Academic Press, London, 265-388.

Blanc F., Blanc-Vernet L., Laurec A., Le Campion J., Pastouret L., 1975. Application paléoécologique de la méthode d'analyse factorielle en composantes principales : interprétation des microfaunes de foraminifères planctoniques quaternaires en Méditerranée. II. Étude des espèces de Méditerranée orientale, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, **18**, 293-312.

Bornhold B. D., 1973. Late Quaternary sedimentation in the Eastern Angola basin, *Techn. Rep.*, Woods Hole Ocean. Inst., 213 p.

Burke K., Durotoye A. B., Whiteman A. J., 1971. A dry phase south of the Sahara 20 000 years ago, *West Afr. J. Archeol.*, Ibadan, **1**, 1-8.

Calvert S. E., 1966. Accumulation of diatomaceous silica in the sediments of the Gulf of California, *Geol. Soc. Amer. Bull.*, **77**, 589-596.

Chamley H., 1971. Recherches sur la sédimentation argileuse en Méditerranée, *Sc. Géol. Strasbourg*, mem. 35, 225 p.

Chamley H., Diester-Haass L., Lange H., 1977. Terrigenous material in East Atlantic sediment cores as an indicator of NW African climates, *"Meteor" Forsch.-Ergebn.*, R. C. (in press).

Chamley H., Durand J. P., Trauth N., 1976. Interstratifiés kaolinite-smectite dans le Valdo-Fuvélien au toit de la bauxite des Alpilles (Provence), *C.R. Acad. Sc., Paris*, **283** série D, 439-442.

Cita M. B., Vergnaud-Grazzini C., Robert C., Chamley H., Ciaranfi N., D'Onofrio S., 1977. Paleoclimatic record of a long deep-sea core from the Eastern Mediterranean, *Quaternary Res.*, **8**, 205-235.

CLIMAP Project Members, 1976. The surface of the ice-age earth, *Science*, **191**, 1131-1137.

Coope G. R., 1975. Climatic fluctuations in northwest Europe since the Last Interglacial, indicated by fossil assemblages of Coleoptera, In Ice Ages: Ancient and Modern, *Geol. J., sp. issue*, **6**, 153-168.

Coope G. R., Morgan A., Osborn P. J., 1971. Fossil Coleoptera as indicators of climatic fluctuations during the last glaciation in Britain, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, **10**, 87-101.

Degens E. G., Hecky R. E., 1974. Paleoclimatic reconstruction of late Pleistocene and Holocene based on biogenic sediments from the Black Sea and a tropical African lake, in *Les méthodes quantitatives d'étude des variations du climat au cours du Pléistocène*, Cnrs, 13-24.

Delibrias G., Gasse F., Rognon P., 1973. Existence de lacs importants au Pléistocène Supérieur (34 000, 23 000 ans BP) dans l'Afar méridional (Éthiopie), *C.R. Acad. Sc. Paris*, **277**, série D, 2633-2636.

Diester-Haass L., Schrader H. J., Thiede J., 1973. Sedimentological and paleoclimatological investigations of two pelagic ooze cores off Cape Barbas, North-West Africa, *"Meteor" Forsch.-Ergebn.*, R. C., **16**, 19-66.

Duplessy J. C., 1972. La géochimie des isotopes stables du carbone dans la mer, *thèse doct. ès Sci. Phys.*, Univ. Paris-VI, 196 p.

Duplessy J. C., 1978. Isotope studies, in *Climatic changes*, Cambridge University Press (in press).

Duplessy J. C., Chenouard L., Reyss J. L., 1974. Paléotempératures isotopiques de l'Atlantique Equatorial, in *Les méthodes quantitatives d'étude des variations du climat au cours du Pléistocène*, Cnrs, 251-258.

Duplessy J. C., Chenouard L., Vila F., 1975. Weyl's theory of glaciation supported by isotopic study of Norwegian core K 11, *Science*, **188**, 1208-1209.

Duplessy J. C., Lalou C., Vinot A. C., 1970. Differential isotopic fractionation in benthic foraminifera and paleotemperatures reassessed, *Science*, **168**, 250-251.

Ellis D. B., 1972. Holocene sediments of the South Atlantic Ocean: the calcite compensation depth and concentrations of calcite, opal, and quartz, *MS Thesis*, 77 p., School of Oceanography, Oregon State University, Corvallis, Oregon.

Emery K. O., Niino H., Sullivan B., 1971. Post-pleistocene levels of the East China Sea, in *Late Cenozoic Glacial ages*, edited by K. K. Turekian, Yale University Press, 381-390.

Emery K. O., Lepple F., Toner L., Uchupi E., Rioux R. H., Pople W., Hulburt E. M., 1974. Suspended matter and other properties of surface waters of the northeastern Atlantic Ocean, *J. Sedim. Petr.*, **44**, 1087-1110.

Emiliani C., 1955. Pleistocene temperatures, *J. of Geology*, **63**, 538-578.

Flint R. F., 1971. *Glacial and Quaternary Geology*, J. Wiley and Sons Inc., New York, 892 p.

Faure H., Elouard P., 1967. Schéma des variations du niveau de l'Océan Atlantique sur la Côte de l'Ouest de l'Afrique depuis 40 000 ans, *C.R. Acad. Sc. Paris*, **265**, série D, 784-787.

Gameson A. L. H., Barrett M. J., 1958. Oxidation, reaeration and mixing in the Thames estuary, Public Health Service R.A. Taft Sanit. Engineering Center, *Tech. Rep.*, W 58-2, 63-93.

Gardner J. V., Hays J., 1976. Responses of sea-surface temperature and circulation to global climatic change during the past 200 000 years in the Eastern Equatorial Atlantic Ocean, *Geol. Soc. Amer. Mem.*, **145**, 221-246.

Giresse P., Odin G.-S., 1973. Nature minéralogique et origine des glauconies du plateau continental du Gabon et du Congo, *Sedimentology*, **20**, 457-488.

Goldberg E. D., 1958. Determination of opal in marine sediments, *J. Mar. Res.*, **17**, 178-192.

Goldthwait R. P., Dreimanis A., Forsyth L., Karrow P. F., White G. W., 1965. Pleistocene deposits of the Eirie Lobe, in *The Quaternary of the United States*, edited by H. E. Wright, D. G. Frey, Princeton Univ. Press, 85-98.

Griffin J. J., Windom H., Goldberg E. D., 1968. The distribution of clay minerals in the World Ocean, *Deep-Sea Res.*, **15**, 433-459.

Grove A. T., Street F. A., 1977. Global maps of Late Quaternary lake level fluctuations. *X INQUA Congress*, Birmingham, Abstr. vol., 185.

Heath G. R., Moore T. C., Dauphin J. P., 1976. Late Quaternary accumulation of opal, quartz, organic carbon, and calcium carbonate in the Cascadia Basin area, Northeast Pacific, *Geol. Soc. Amer. Mem.*, **145**, 393-409.

Heath, G. R., Moore T. C., Roberts G. L., 1974. Mineralogy of surface sediments from the Panama Basin, Eastern Equatorial Pacific, *J. Geol.*, **82**, 145-160.

Hecht A. D., 1976. An ecological model for test size variation in recent planktonic foraminifera: applications to the fossil record. *J. of Foram. Res.*, **6**, 4, 295-311.

Klingebiel A., Boltenhaguen C., Caratini C., Debyser Y., Delteil J. R., Durand B., Étienne J., Huc A., Latouche C., Millepieds P., Parra M.,

Pujol C., Sauvan P., Sommer F., 1975. Études stratigraphiques, sédimentologiques et géochimiques de sédiments carottés dans le golfe de Guinée au cours de la campagne Benin 1971, *Bull. Inst. Géol. Bass. Aquit.*, **17**, 101-232.

Labeyrie J., Duplessy J. C., Delibrias G., Letolle R., 1967. Étude des températures des climats anciens par la mesure de l'oxygène 18 du carbone 13 et du carbone 14 dans les concrétions des cavernes, in *Radioactive dating and methods of low level counting*. IAEA, 153-160.

Lemke R. W., Laird W. M., Tipton M. J., Lindvall R. M., 1965. Quaternary geology of northern Great Planes, in *The Quaternary of the United States*, edited by H. E. Wright, D. G. Frey, Princeton Univ. Press, 15-28.

Maley J., 1977. Paleoclimates of Central Sahara during the early Holocene, *Nature*, **269**, 573-577.

Manabe S., Hahn D. G., 1977. Simulation of the tropical climate of an Ice Age, *J. Geophys. Res.*, **82**, 27, 3889-3911.

Millot G., 1953. Minéraux argileux et leurs relations avec la géologie, *Rev. Inst. Franç. Pétr.*, **8**, n° spéc., 75-86.

Millot G., 1964. *Géologie des Argiles*, Masson, Paris, 499 p.

Molina Cruz A., Price P., 1977. Distribution of opal and quartz on the ocean floor of the subtropical southeastern Pacific, *Geology*, **5**, 81-84.

Moore T. C., Heath G. R., Kowsmann R. O., 1973. Biogenic sediments in the Panama Basin, *J. Geol.*, **81**, 458-472.

Ninkovich D., Shackleton N., 1975. Distribution, stratigraphic position and age of ash layer "L" in the Panama basin region, *Earth Planet. Sc. Lett.*, **27**, 20-34.

Paquet H., 1969. Évolution géochimique des minéraux argileux dans les altérations et les sols des climats méditerranéens et tropicaux à saisons contrastées, *Mém. Serv. Carte Géol. Als.-Lorr.*, **30**, 212 p.

Pastouret L., 1970. Étude sédimentologique et paléoclimatique de carottes prélevées en Méditerranée orientale, *Téthys*, **2**, 1, 227-266.

Pastouret L., Melguen M., Thiede J., 1974. Sedimentation pléistocène dans le golfe de Guinée. Comparaison des données micropaléontologiques et sédimentologiques. *VI^e Colloque Africain de Micropaléontologie*, Tunis (21 mars-3 avril 1974), in press.

Porrenga D. H., 1966. Clay minerals in recent sediments of the Niger delta, *Clays and Clay min.*, **14**, 221-233.

Richards F. A., 1965. Anoxic basins and fjords, in *Chemical Oceanography*, **1**, edited by Riley and Skirrow, Academic Press.

Ruddiman W. F., 1971. Pleistocene sedimentation in the Equatorial Atlantic: stratigraphy and faunal paleoclimatology, *Geol. Soc. Amer. Bull.*, **82**, 283-302.

Ryan W. B. F., 1972. Stratigraphy of late Quaternary sediments in the Eastern Mediterranean, in *The Mediterranean Sea, a natural sedimentation laboratory*, edited by D. J. Stanley, Dowden, Hutchinson and Ross Inc., Stroudsburg, USA, 149-170.

Ryan W. B. F., Hsü K. J. *et al.*, 1973. *Initial Reports of the Deep Sea Drilling Project*, **13**, 1-2, US Government Printing Office, Washington DC, 1-1447.

Sarnthein M., 1977. The sand deserts in peak glacial and interglacial, *Science* (in press).

Sarnthein M., Diester-Haass L., 1977. Eolian sand turbidites, *J. Sedim. Petrol.* (in press).

Schott W., 1966. Foraminiferen fauna und Stratigraphie der Tiefsee-sedimente in nordatlantischen Ozean, *Report Swedish Deep-Sea Exp.*, **VII**, 8, 357-469.

Schrader H. J., 1971. Ursache und Ergebnis der Auflösung von Kieselskeletten in den oberen Sedimentbereichen am Beispiel zweier Kern-Profile vor Marokko und Portugal, *Proc. 2nd Plankt. Conf.* (Rome), **2**, 1149-1155.

Servant M., 1973. Séquences continentales et variations climatiques : évolution du bassin du Tchad au Cénozoïque supérieur, *thèse d'État*, Paris, 348 p.

Servant M., Servant S., Delibrias G., 1969. Chronologie du Quaternaire récent des basses regions du Tchad, *C.R. Acad. Sc. Paris*, **269**, série D, 1603-1606.

Shackleton N. J., Opdyke N. D., 1973. Oxygen isotope and palaeomagnetic stratigraphy of Equatorial Pacific core V 28-238: oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale, *Quaternary Res.*, **3**, 39-55.

Siefermann G., Jehl G., Millot G., 1968. Allophanes et minéraux argileux des altérations récentes des basaltes du mont Cameroun, *Bull. Gr. franç. Arg.*, **20**, 109-129.

Street F. A., Grove A. T., 1976. Environmental and climatic implications of Late Quaternary lake level fluctuations in Africa, *Nature*, **261**, 385-390.

Thiede J., Pastouret L., Melguen M., 1974. Sédimentation profonde au large du delta du Niger (golfe de Guinée), *C.R. Acad. Sc. Paris*, **278**, série D, 987-990.

Till R., Spears D. A., 1969. The determination of quartz in sedimentary rocks using an X-ray diffraction method, *Clays and Clay min.*, **17**, 323-327.

Weyl P. K., 1968. Theory of ice ages, *Meteorological monographs*, **8**, 37-62.