Grey, D.R.C. & H.J.Cooke 1977. Some problems in the Quaternary evolution of the landforms of Northern Botswana. *Catena* 4:123-133.
Grove, A.T. 1969. Landforms and climatic change in the Kalahari and Ngamiland. *Geogr. J.* 135:191-212.
Jaeger, E.C. 1957. *The North American deserts.* Stanford, 308 pp.
Klaus, D. 1981. Klimatologische und klima-ökologische Aspekte der Dürre im Sahel. *Erdwiss. Forschg.* 16:1-175.
Lamb, H.H. 1974. Contributions to historical climatology: The Middle Ages and after; Christmas weather and other aspects. *Bonner Met. Abh.* 17:549-567.
Meckelein, W. 1959. *Forschungen in der zentralen Sahara. I. Klimageomorphologie.* Braunschweig, 181 pp.
Mensching, H. 1975. Die Sahelzone – Ursachen und Konsequenzen der Dürrekatastrophe in Afrika. *Umschau* 1975:305-308.
Mensching, H. 1978. Die Wüste schreitet voran. *Umschau* 1978:99-106.
Nicholson, S. 1978. Climatic variations in the Sahel and other African regions during the past five centuries. *J. of Arid Environment* 1:3-24.

PRELIMINARY RESULTS FROM SEDIMENT CORES FROM LAKE BOSUMTWI, GHANA

M. R. TALBOT[1] / D. A. LIVINGSTONE[2] / PATRICIA G. PALMER[3] / J. MALEY[4] / JOHN M. MELACK[5] / G. DELIBRIAS[6] / S. GULLIKSEN[7]

ABSTRACT

Results are presented from the analysis of three sediment cores from Lake Bosumtwi, Ghana, using mineralogy, sedimentology, palynology and grass cuticle content. The longest core is 16.9 m long and spans approximately the last 27 500 years of the lake's history. Apart from minerals of clastic origin, the sediments contain a variety of endogenetic and diagenetic minerals, particularly carbonates, with phosphates and zeolites also prominent in some sections. Mineral content and sedimentology are used to reconstruct the history of the lake. The oldest part of the sequence, from c.27 500 to 24 500 BP, is poorly understood, but seems to contain evidence of both high and low water levels. Between 23 000 and 19 500 BP the lake water was relatively dilute, but became more saline and alkaline between 19 000 and 16 000 BP when the lake level must have been low. Regressions of the lake are registered just before 13 500 BP, at c.3 750 BP and after 2 000 BP. Maximum lake level seems to have occurred between 9 000 and 4 000 BP, when the lake waters were dilute.

Grass pollen dominates the pollen assemblage before c.9 000 BP indicating the existence of grassland around Bosumtwi. A dramatic rise in the abundance of arboreal pollen suggests that forest rapidly replaced grassland after 9 000 BP. Patches of montane-like forest within the grassland are suggested by the

1. Geologisk Institutt, Avd. A, Universitet i Bergen, Allégt.41, 5014 Bergen, Norway.
2. Department of Zoology, Duke University, Durham, North Carolina 27706, USA.
3. Department of Biology, Louisburg College, Louisburg, North Carolina 27549, USA.
4. ORSTOM, Laboratoire de Palynologie du CNRS, Université des Sciences et Technique du Languedoc, Montpellier 34060, France.
5. Department of Biological Sciences, University of California, Santa Barbara, California 93106, USA.
6. Centre des Faibles Radioactivités, Labo. Mixte CNRS-CEA, Gif sur Yvette 91190, France.
7. Laboratoriet for Radiologisk Datering, Fysikkseksjonen, Sem Sælandsv.5, 7034 Trondheim, Norway.

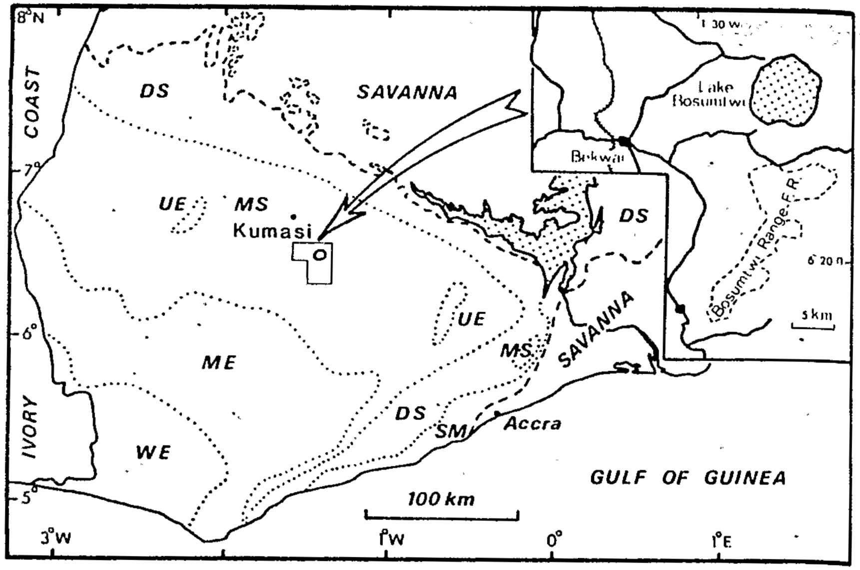


Figure 1. Map of southern Ghana, showing Lake Bosumtwi and position of major vegetation zones. WE – Wet Evergreen; UE – Upland Evergreen; MS – Moist Semi-deciduous; DS – Dry Semi-deciduous; SM – Southern Marginal (after Hall & Swaine 1976).

presence of *Olea hochstetteri* pollen in late Pleistocene-early Holocene sections of the cores.

Grass cuticle fragments are prominent in sediments deposited before c.9 000 BP, confirming the pollen evidence that grassland formerly existed around the Bosumtwi crater. Of particular interest is the occurrence of pooid grass cuticle in the Pleistocene-early Holocene sediments. Pooid grasses, and the pollen of *O.hochstetteri,* imply palaeotemperatures several degrees centigrade lower than today's in southern Ghana.

INTRODUCTION

During August 1976, seven piston cores were raised from Lake Bosumtwi, which occupies a meteorite impact crater in the forest zone of southern Ghana (Fig.1). Six cores are now in cold storage at the Zoology Department, Duke University; the other was left at Accra with the Geological Survey Department of Ghana. We present here preliminary results from the detailed sedimentary logging and mineralogical analysis of Cores B2, B6 and B7 and pollen analysis, at 0.5 m intervals, of B6. So far as we know, B7 is the longest core yet taken from any West African lake. It provides the longest continuous Holocene-Late Pleistocene record of palaeoenvironmental change currently available from this part of the continent.

The surface of Lake Bosumtwi is presently 100 m above sea level. The crater lies within the forest zone of Ghana, the region being characterised by moist semi-deciduous forest with patches of drier forest on some hill tops and ridges (Hall & Swaine 1976, Hall et al. 1978). The forest-savanna boundary occurs ca.65 km north of the lake (Fig.1). Mean annual temperature in the Bosumtwi region is about 26° and rainfall 1 500 m. A rather severe hot, dry season occurs from November to February. Pan evaporation rates at Kumasi average 1 370 mm/year (Ghana Met.Services data).

Lake waters have a pH of 9.1-9.6 at the surface with total dissolved solids amounting to 720 ppm, Na^+ and HCO_3 being the principal ions in solution. The Mg/Ca ratio of the surface water is about 4.4. Waters are usually anoxic below 10 m and permanently so below 40 m (Whyte, personal communication).

CORES AND CORE SITES

B6 and B7, from the centre of the lake, were taken in 78 m of water. B2 comes from 34 m of water, at a point about 1.2 km from the north-westlake shore.

B6 is 13.5 m long and B7 16.9 m long. Approximately 0.3 m is missing from the top of B7, but otherwise the two cores contain continuous sedimentary sequences and can readily be correlated via a number of distinctive stratigraphic features. B2 is 4.2 m long, and although it contains one major hiatus is nevertheless of considerable value in providing a link between the lake centre succession and sequences exposed above the present lake shore (see Talbot 1976, Hall et al. 1978, Talbot & Delibrias 1980, Talbot & Hall 1981).

RESULTS

The main features of the cores' stratigraphy, mineralogy, palynology and grass cuticle content are summarised in Figures 2 and 3.

Mineralogy

Sediment smears taken from the cores at irregular but closely spaced intervals (0.5-10 cm), have been examined under the petrographic microscope for mineral and biogenic component identification. Selected minerals have been further studied by scanning electron microscopy (SEM) and X-ray diffraction (XRD). Some semi-quantitative chemical analyses have been carried out using a Kevex energy dispersive X-ray system (EDS). As yet there has been no detailed investigation of the clay minerals.

Sediments recovered in the cores range in grain size from clay to granule sand. Muds of clay to fine silt grade predominate, but almost all of these con-

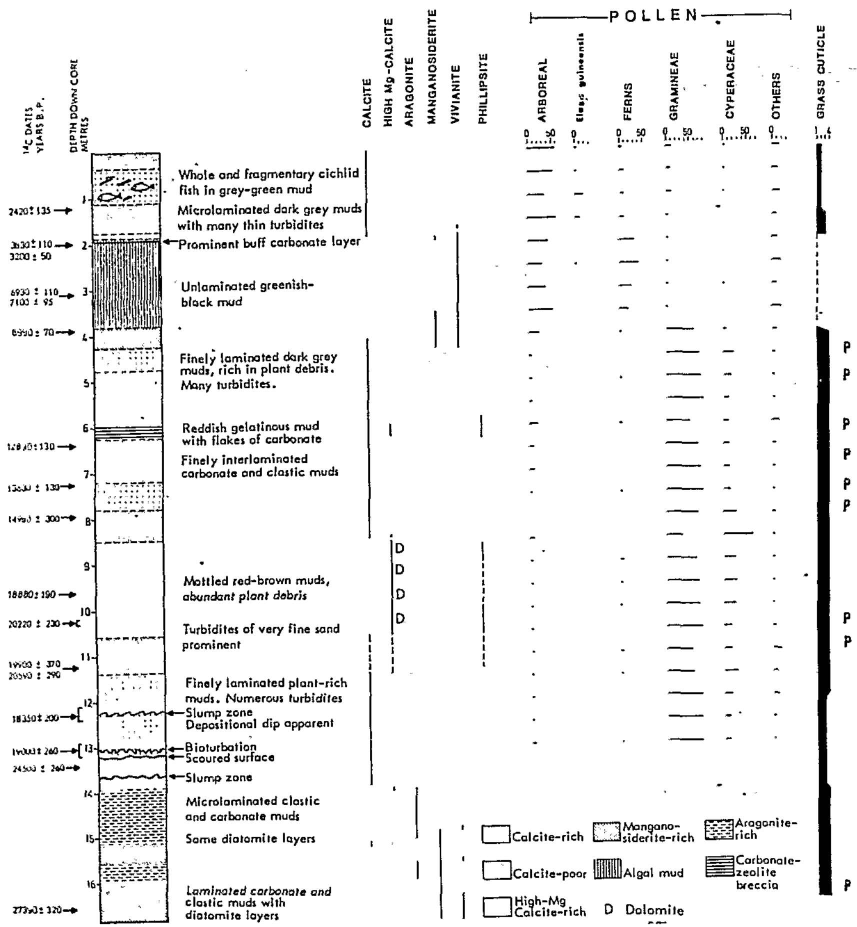


Figure 2. Log of cores B6 and B7 (from 78 m of water) with stratigraphic position of ^{14}C age determinations. Scale 1-4 shows relative abundance of grass cuticle fragments: 1 – present – 4 – abundant. P – pooid grass fragments.

tain a proportion of clastic, biogenic, endogenic or authigenic minerals with diameters in excess of 0.05 mm. The most important of these are:

Clastic

Quartz, mica and feldspar are the dominant clastic minerals, with tourmaline and rutile present as fairly constant accessories. Pale green chlorite is a particularly prominent mica. This seems mainly to be derived from Proterozoic metagreywackes and metavolcanics cropping out within the crater, but the presence of some authigenic chlorite cannot be ruled out.

Biogenic

Siliceous sponge spicules provide the most constant biogenic component, but they never form a significant proportion of the sediment. A variety of spicule types are present.

Diatoms occur, but are surprisingly scarce, except below a depth of 14.5 m in B7, where they form millimetre-thick laminae, composed of one or two diatom species. At some levels in the cores, diatoms have almost certainly been destroyed during diagenesis.

Fish remains are only prominent in the uppermost 1.1 m of B6, where there are some horizons rich in fish debris, including many whole individuals. Fragments are seen at a few levels further down, but in general the sediment pore waters seem to have been unfavourable to the preservation of bone and scales. Dissolution of these materials during diagenesis was probably a major factor in the development of authigenic phosphate minerals (see below).

Filamentous blue-green algae (cyanobacteria) can form an important non-mineral component of the sediment. In all cores, the mid-Holocene section is characterised by a distinctive greenish-black (5GY2/1-5G2/1), unlaminated mud (sapropel). The colour is in large part due to an abundance of blue-green algal pigment, and well-preserved algal filaments are often prominent in smear slides. These sediments have a total organic carbon content of 15-20 %.

Endogenic, authigenic

A variety of endogenic and authigenic minerals are present, the former precipitated within the water column, the latter from sediment pore waters. The stratigraphic distribution of the major minerals is shown in Figures 2 and 3. The following list is almost certainly not exhaustive.

Calcite. This is the commonest endogenic mineral and typically occurs as spear-head shaped crystals up to 0.25 mm long. The crystals often encrust plant debris. SEM examination shows the calcite crystals to be frequently corroded; corrosion presumably occurred when sinking crystals came into contact with deep lake waters of low pH, or within the sediment pore waters.

High magnesian calcite (Mg-Calcite). The calcite described above contains less than 5 mole percent $MgCO_3$. Calcite containing more than 5 % $MgCO_3$ occurs between 8.5 and 11.5 m, between 13.8 and 15.4 m. The $MgCO_3$ content of these magnesian calcites may be up to 25 %, but is usually from 6-18 %. Mg-Calcite typically occurs as discrete granules, but between 6.0 and 6.25 m, it forms a micritic matrix between the low Mg-Calcite crystals and aggregates of phillipsite.

Dolomite. Between 9.3 and 10.2 m, cores B6 and B7 contain a number of thin, buff-coloured laminae which are composed almost entirely of 2-5 μ diameter rhombs and granules of dolomite. The mineral is apparently well-

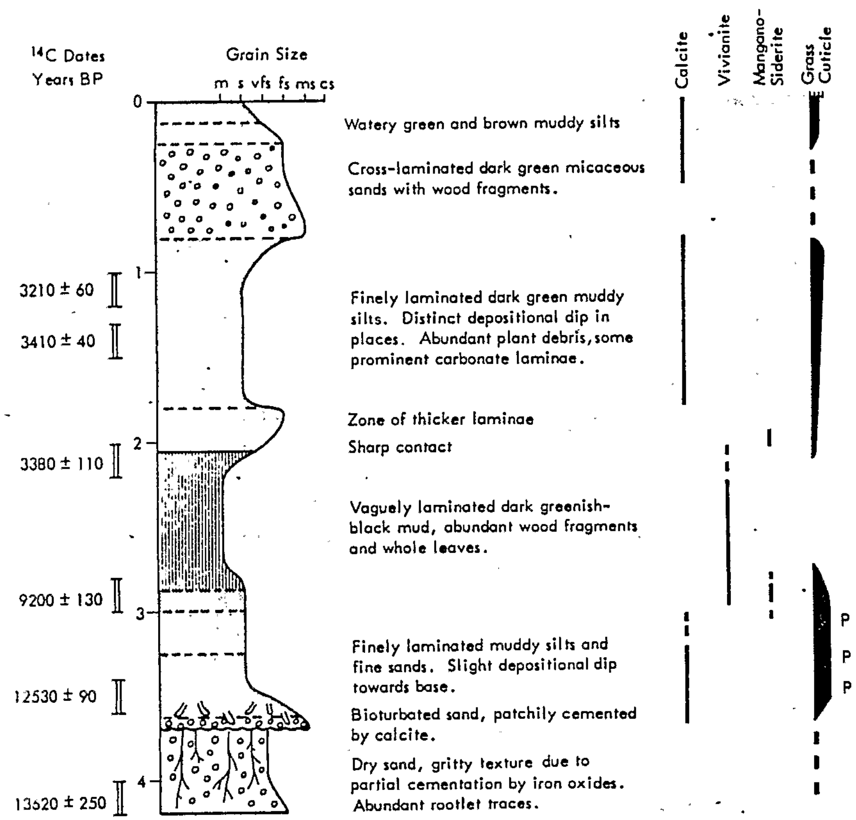


Figure 3. Log of core B2 (from 34 m of water). Note rootlet traces indicating colonization of sediment by littoral or terrestrial vegetation. Symbols as for Figure 2.

ordered; sharp XRD peaks suggest a stoichiometric structure. Dolomite granules also occur scattered through the intervening sediment, but here XRD analysis indicates that it is usually a proto-dolomite containing 46-48 mole percent $MgCO_3$.

Aragonite. Much less abundant than calcite, aragonite occurs only below 13.5 m in B6 and B7, where it forms monomineralic laminae up to 1 mm thick. The crystals are laths or needles, some twinned, and less than 50 μm long.

Manganosiderite, Fe Mn $(CO_3)_2$. This is a distinctive component of some parts of the cores (Figs.2 and 3), where anhedral granules, 1-5 μ in diameter, can form laminae up to 2 mm thick. XRD examination of samples from different stratigraphic levels suggests that the Fe/Mn ratio may be rather variable. In B6 and B7, a prominent band of manganosiderite marks the top of the sapropel. Apparently accumulation of this mud ended with an event favourable to the precipitation of manganosiderite. It has previously been reported from the sediments of Lake Kivu (Hecky & Degens 1973) and a few other lakes (Kelts and Hsü 1978).

Viviantie, $Fe_3 (PO_4)_2.8H_2O$. This accompanies the manganosiderite but also occurs where no manganosiderite is developed. It is, for example, the only visible authigenic mineral in the sapropel. Vivianite usually forms radial growths of prismatic crystals up to 0.3 mm long.

Phosphoferrite/Reddingite, $(Mn, Fe)_3(PO_4)_2.3H_2O$. Although it has not yet been identified optically, XRD analysis of the sapropel suggests that this mineral accompanies the visible vivianite. There are indications that another phosphate, anapaite, $Ca_2 Fe(PO_4)_2 .4H_2O$, may also be present.

Sulphides. A sprinkling of cubes, pentagonal dodecahedra and framboids of Fe, Mn-sulphides occurs throughout the cores, but are abundant at only a few horizons.

Zeolite. Small (<40 μm) crystals with an elongate, sometimes slightly curved and often complexly twinned pseudoorthorhombic habit occur through some core intervals. Semi-quantitative EDS analysis shows them to be an alumino-silicate with potassium as the principal cation. This composition and the distinctive crystal morphology suggest that the mineral is a potassium-rich variety of phillipsite.

Phillipsite has been reported from a number of alkaline lake sediments, but always in association with volcanic glass (Surdam & Sheppard 1978). It is interesting to note, therefore, that there is no evidence of vulcanicity in or around Bosumtwi. Here the silica necessary for phillipsite precipitation may instead have been of biogenic origin, since SEM examination reveals the presence of rare, highly corroded diatoms in intimate association with the zeolite.

Discussion

This is not the place to discuss in detail the extremely complex mineralogy of the Bosumtwi sediments, but it is worth emphasising one feature which has emerged. In many lacustrine successions, carbonate minerals occur in a predictable sequence related directly to changes in water salinity.

Normally the sequence is, with increasing salinity, calcite - Mg-calcite - aragonite and dolomite (Müller et al. 1972, Kelts & Hsü 1978, Eugster 1982, Eberli et al. 1982). In Bosumtwi, aragonite is abundant between 16 and 14 m in B6/B7, but there is very little Mg-calcite and no dolomite. Between 10.5 and 8.5 m, on the other hand, there is no aragonite, but abundant Mg-calcite and some dolomite. The aragonite in these cores is undoubtedly a primary precipitate, but the dolomite is probably diagenetic. It seems likely that previously unsuspected diagenetic reactions may have had an important influence upon the carbonate mineral assemblages. Recently Kelts & McKenzie (1982) have recorded dolomite of primary or diagenetic origin in organic-rich Tertiary marine sediments with pore waters of rather low Mg/Ca ratio (1-5), so it is possible that similar dolomite can occur in some organic-rich lacustrine sediments without the waters necessarily being particularly saline.

In this case, considerable caution will be required in the use of lacustrine Mg-bearing carbonate minerals as proxy palaeolimnological or palaeoclimatological indicators.

SEDIMENTOLOGY

Cores B6 and B7

For much of their length, these cores are finely laminated because of superposition of successive clastic units or alternation of clastic sediments with endogenic mineral laminae or plant debris. Deposition of clastic sediments seems mainly to have been from the distal portions of density flows. The proximal parts of these flows accumulated around the basin margin to form turbidite fans, some of which are now exposed above present lake level (Talbot 1976, 1983). In the cores, there are a few thin zones of lamination deformed by down-slope mass flow. Plant fragment-rich microlaminae between the clastic laminae were probably formed from material which settled from suspension during the interval between each density flow. Endogenic and diagenetic mineral laminae between the clastic laminae are composed of calcite, Mg-calcite, dolomite, aragonite or manganosiderite.

In view of the fine lamination the question arises as to whether the sediments are varved. In cores B6-4 and B6-5, where they can be measured and counted, the mean thickness of the laminae is between 0.5 and 0.6 mm. This is close to the mean sedimentation rate of 0.66 mm/year in this part of the core. On the other hand, detailed ^{14}C dating of sections in exposed proximal regions of the turbidite fans indicates that density flows were generated at frequencies of three to five per year during terminal Pleistocene and early Holocene times (Talbot & Delibrias 1980). We are unable to provide a general answer to this important question.

The sapropel displays few traces of lamination and is interrupted only by a few very thin turbidites. In general, the mud presents a rather massive appearance because of its homogeneity rather than post-depositional mixing.

Lamination is also poor between 6.0 and 6.25 m where three units of sticky, zeolite-bearing grey-green mud each passes upwards into a reddish breccia of carbonate and zeolite fragments in a gelatinous matrix. The breccia consists of flakes of Mg-calcite-cemented sediment a few millimetres thick

and up to 2 cm across. Some flakes show signs of abrasion. The flakes would thus seem to be clasts of reworked, partially lithified sediment. Formation of authigenic phillipsite indicates that saline, alkaline (pH>9) waters must have been in contact with the sediment. Zeolitic developments of this sort are typical of playa lakes (Surdam & Sheppard 1978), so Bosumtwi was possibly exceptionally low at this time, the reddened surface perhaps representing exposed or barely-submerged mud-flats. The associated gels provide additional support for this hypothesis, since in many alkaline playa lakes zeolites commonly develop from a gelatinous precursor (Surdam & Sheppard 1978, Eugster & Hardie 1978).

Despite the rather convincing mineralogical evidence there are, however, some serious problems in reconciling a very low lake level at this time with other sorts of palaeolimnological evidence from the crater (see below).

What appears to be a bioturbated zone occurs at about 13.0 m. This is concentrated around the remnants of a sharp but wavy contact between two muds, across which the upper mud penetrates the lower along irregular, tube-like structures. There are few traces of primary lamination for 5 cm above and below the contact; the muds instead have a mottled appearance. These structures seem best interpreted as bioturbation which, in view of its association with a sharp, perhaps scoured contact, appears to have occurred at a time of reduced sedimentation.

CHRONOLOGY

Seventeen ^{14}C dates have been obtained from organic matter contained in bulk sediment samples from cores B6 and B7 (Table 1). Most of the age determinations are stratigraphically consistent, but inversions occur at the top of the sapropel and at the c.12.8 m level (Fig.2).

Age-depth relations in B6 and B7 are plotted on Figure 4, using the top of the sapropel as the primary reference horizon. This plot clearly illustrates the anomalous nature of the two age determinations of 18 350 ± 200 BP and 19 000 ± 260 BP. If these dates are ignored, it can be seen that the rest of the points fall on two lines which intersect at about 9 000 BP, i.e. at the base of the sapropel. Sedimentation rates before this time were significantly higher (mean 0.66 m/1 000 years) than during the rest of the Holocene (mean sed. rate 0.39 m/1 000 years). As shown on Figure 4, a single line can be drawn through all data points, but this is rejected because it implies the absence of around 1 m of sediment from the top of B6, which seems improbable. If, on the other hand, allowance is made for the anomalous thickness imparted to the upper part of this core by the barely compacted fish layers (Fig.2), then the top of B6 obtained by us would lie very close to the Holocene line shown on Figure 4.

The change in sedimentation rate coincides with the transition from wooded grassland to forest vegetation in the Bosumtwi crater (see below)

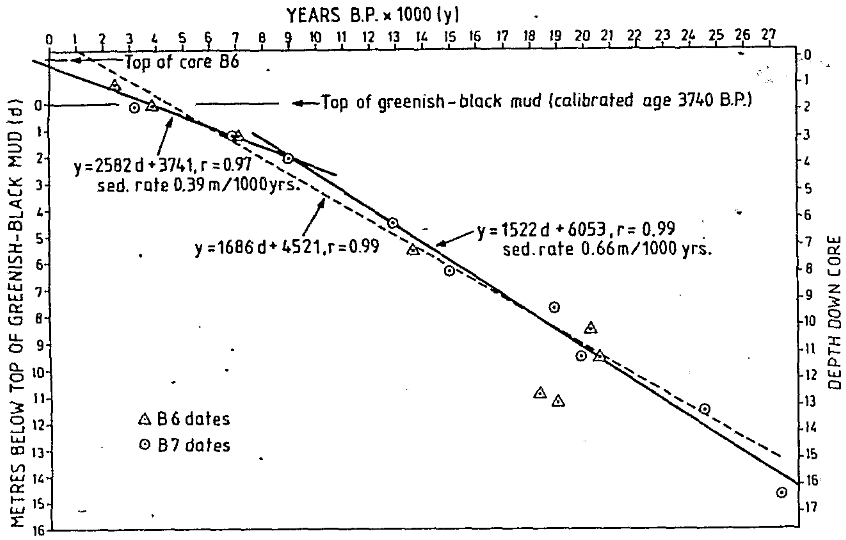


Figure 4. Sediment age depth relationship for cores B6 and B7, using the top of the greenish-black mud (sapropel) as the primary correlation horizon between the two cores. Sediment accumulation rates were clearly considerably reduced after ca.9 000 BP. (Broken line is drawn through all points except the anomalous 18 350 ± 200 BP and 19 000 ± 260 BP dates. This line implies the absence of about 1 m of sediment from the top of B6 and is therefore rejected in favour of the two solid lines.)

and thus appears to reflect a decrease in sediment yield induced by this major vegetational change.

At present, no satisfactory explanation can be provided for the two anomalous dates at c.12.8 m (Figs.2 and 4). The clustering of dates of the order of 20 000-20 500 BP between 10.2 and 11.2 m is probably due to very high sedimentation rates on a prograding turbidite fan.

Core B2: Chronology and sedimentology

This core comes from sediments that accumulated in shallower water, and so contains representatives of a more varied suite of depositional environments (Fig.3). Six ^{14}C ages have been obtained from bulk sediment samples (Fig.3, Table 1).

At the base of the core is a dry, crumbly-textured sand riddled with rootlets. Ferruginisation of the sands indicates pedogenetic processes, thus confirming that this unit represents subaerial exposure of a shallow water sand. A thin beach sand erosively overlying this soil records the subsequent transgression due to rising lake level. A ^{14}C age of 13 620 ± 250 BP from the soil

Table 1. Stratigraphical and analytical data on ^{14}C dated samples. All age determinations were carried out on organic carbon extracted from bulk sediment samples.

Lab. no.	Core no. and stratigraphic level of sample (metres)		^{14}C age, years BP ± 1 σ	$\delta^{13}C$ ‰ PDB-1	Corrected ^{14}C age, years BP
β – 1515	B-6-1	2.8-3.0	2 440 ± 135	–25.99	2 420 ± 135
β – 1516	B-6-2	0.6-0.7	3 900 ± 110	–29.28	3 830 ± 110
β – 1517	B-6-4	0.0-0.2	13 500 ± 130	–17.10	13 630 ± 130
β – 1518	B-6-5	2.7-2.9	18 915 ± 260	–19.61	19 000 ± 260
β – 1339	B-6-2	1.8	7 080 ± 95	–23.70	7 100 ± 95
β – 1349	B-6-5	0.0-0.1	20 185 ± 230	–22.90	20 220 ± 230
β – 1919	B-6-5	2.3-2.6	18 290 ± 195	–21.07	18 350 ± 200
β – 4184	B-7-2	0.1-0.3	3 220 ± 50	–26.32	3 200 ± 50
β – 4185	B-7-2	2.1-2.2	9 050 ± 70	–28.70	8 990 ± 70
β – 4186	B-7-3	1.5-1.7	12 890 ± 130	–24.65	12 890 ± 130
β – 4187	B-7-4	1.7-1.9	18 900 ± 190	–25.92	18 880 ± 190
β – 4188	B-7-6	2.7-2.9	27 360 ± 320	–23.42	27 390 ± 320
GIF – 5910	B-6-5	1.0-1.2	20 590 ± 290		
GIF – 5911	B-7-2	1.2-1.4	6 930 ± 110		
GIF – 5912	B-7-4	0.3-0.5	14 980 ± 300		
GIF – 5913	B-7-5	0.5-0.7	19 900 ± 370		
GIF – 5914	B-7-5	2.5-2.7	24 500 ± 400		
T – 4635	B-2-1	1.0-1.2		–27.43	3 210 ± 60
T – 4640	B-2-2	0.1-1.3		–26.76	3 410 ± 40
T – 4639	B-2-2	0.8-1.0		–17.19	3 380 ± 110
T – 4638	B-2-2	1.6-1.8		–28.85	9 200 ± 130
T – 4637	B-2-2	2.2-2.4		–20.70	12 530 ± 90
T – 4636	B-2-2	2.8-3.0		–10.99	13 620 ± 250

probably dates this transgression, as the bulk of the organic matter here seems to be from the roots of vegetation killed by the transgression.

The soil and overlying transgressive sand are of the same age as similar deposits seen at Banso and Obo (Talbot & Delibrias 1980). Thus the soil records a major period of low water in the Bosumtwi crater when the water level was below 38 m for long enough to allow significant pedogenesis of exposed sediments to occur. By 12 690 ± 230 BP the subsequent transgression had already reached an elevation of +16 m (Talbot & Delibrias 1980).

Only 0.6 m of sediment separates the horizon dated at 12 530 ± 90 BP from that with a ^{14}C age of 9 200 ± 130 BP (Fig.3). This is surprisingly little since it is known that this was a period marked by particularly high sedimentation rates around the basin margin (Talbot & Delibrias 1980). The most likely explanation is that the core site occupies a topographic high that was initially starved of sediment. Alternatively, it is possible that sediment was removed by erosion during a fall in lake level at c.10 500 BP (Talbot & Delibrias 1980), but there is no unequivocal evidence for this in the core and certainly no sign of subaerial exposure.

The sequence between 3.3 and 1.0 m is generally similar to the corresponding section in B6 and B7, although density flow deposits above and below the sapropel in B2 have thicker laminae and are of coarser grain size, turbidites of fine sand being relatively common. The sapropel also contains larger clastic grains than in B6 and B7, but accumulated at an even slower rate – c.0.15 m/1 000 years.

Age determinations from the top of the sapropel and at 1.1 and 1.4 m are all of the same magnitude (Fig.3), with a suggestion of inversion. A similar anomaly occurs in B6 and B7, the top of the sapropel yielding an age of 3 200 ± 50 BP in B7 and 3 830 ± 110 BP at a stratigraphically slightly higher level in B6. Some of these dates must be incorrect. An age of around 3 300-3 200 BP for the top of the sapropel seems improbable, since it is known from the exposed sediments that by this time the lake was recovering from a major regressive event (Talbot & Delibrias 1980). Accumulation of the algal-rich sediments was presumably terminated by this regression, which is probably marked in B2 by the zone of significantly coarser sediments just above the sapropel (Fig.3). That the date of 3 380 ± 100 BP for the top of the sapropel is incorrect seems at present to be the least unsatisfactory explanation inasmuch as the overlying dates of 3 410 ± 40 BP and 3 210 ± 60 BP would then be essentially correct. An age of c.3 750 BP for the top of the sapropel estimated from sediment accumulation rates (Fig.4) is considered to be most consistent with all presently available data. We can offer no explanation at this stage for the anomalously young dates from this part of the core.

Although there is still some uncertainty about precisely when the following regression reached its minimum level, this was clearly never low enough to cause exposure of the B2 core site. In view of the significant sand content at around 2 m in the core, the sediments here are thought to be of delta-front origin, in which case minimum water depth would, at a very rough estimate, have been of the order of 10-15 m.

Exposure did not occur during the more recent sub-historic regression either, but an upward coarsening into cross-bedded fine sands indicate progradation of a fluvio-deltaic unit over the core site. These delta-top sediments would probably have accumulated in no more than 1-2 m of water. The subsequent upward-fining succession (0.2-0.0 m, Fig.3) is a result of the present transgression.

PALYNOLOGY

An initial survey of the pollen has been carried out on B6, sampled at horizons 0.5 m apart (Fig.2), a spacing that represents intervals of the order of 1 000 years (Livingstone 1980). The main features of pollen distribution in this core are the dominance of grass pollen assemblages below 3.5 m (c. 8 500 BP), indicating the absence of a continuous forest cover. Above 3.5 m

arboreal pollen has maintained its dominance to the present-day.

A more detailed pollen study of the longer B7 core is now under way. First results show the presence of *Olea hochstetteri* pollen in quantities of between 3-8 % in the interval roughly dated between c.14 500 BP and 8 500 BP, with a maximum near 9 000 BP (Maley & Livingstone 1983). This wild olive, presently unknown from Ghana and the Togo mountains, is typical of montane forests now growing at altitudes above 1 100-1 200 m and no nearer Bosumtwi than 750 km to the west in the Massif des Dans (western Ivory Coast) and to the east in the Jos Plateau (Nigeria). The large percentages of its pollen at Bosumtwi prove an extension of montane forest at altitudes at least 300 to 400 m lower than today's, implying a lowering of mean temperature by at least 2 to 3°C during late Pleistocene and early Holocene times. A similar early Holocene extension of *Olea hochstetteri* has already been demonstrated in the southern part of the Chad basin during the early Holocene (Maley 1981).

Such a finding is in complete accord with the palaeotemperature evidence provided for the same period by grass cuticle belonging to the subfamily Pooideae (Palmer 1982, see Figs.2, 3 and below). During this time interval, the importance of grass pollen and the large concentrations of grass cuticle fragments in the sediments, together with the presence of *O.hochstetteri* pollen, indicate that the Bosumtwi region carried a wooded grassland of tropical montane type (Maley & Livingstone 1983, see also Talbot & Hall 1981). A further important point is to be noted; despite the presence of extensive grasslands, the arboreal pollen grains do not belong to the flora of the typical soudano-guinean savanna, but to the present-day semi-deciduous forest. Leaf fossils of similar age collected from lake margin sediments are from trees that are also typical for modern forests of the Bosumtwi region (Hall et al. 1978, Talbot & Hall 1981).

The large percentage of *Elaeis guineensis* pollen at about 1.5 m may be due to human cultivation around Bosumtwi 3 500-3 000 years ago. Archaeological excavations in Bosumpra cave, 75 km east of Bosumtwi, had previously suggested that the oil palm became an important food plant in southern Ghana sometime after 5 375 ± 100 BP (Smith 1975). Other archaeological evidence suggests that agriculture was introduced to Ghana around 3 300-3 000 BP (Flight 1976).

GRASS CUTICLE

Figures 2 and 3 show the relative abundance of grass cuticle fragments in the sediments, based on examination of smear slides under the petrographic microscope. A detailed study (by P.G.P.) of the cuticle using scanning electron microscopy is still in progress (Palmer 1976, 1982).

Cuticle occurs in all sizes from leaf segments several centimetres long down to tiny fragments of less than 1 mm. Some pieces are carbonised and could

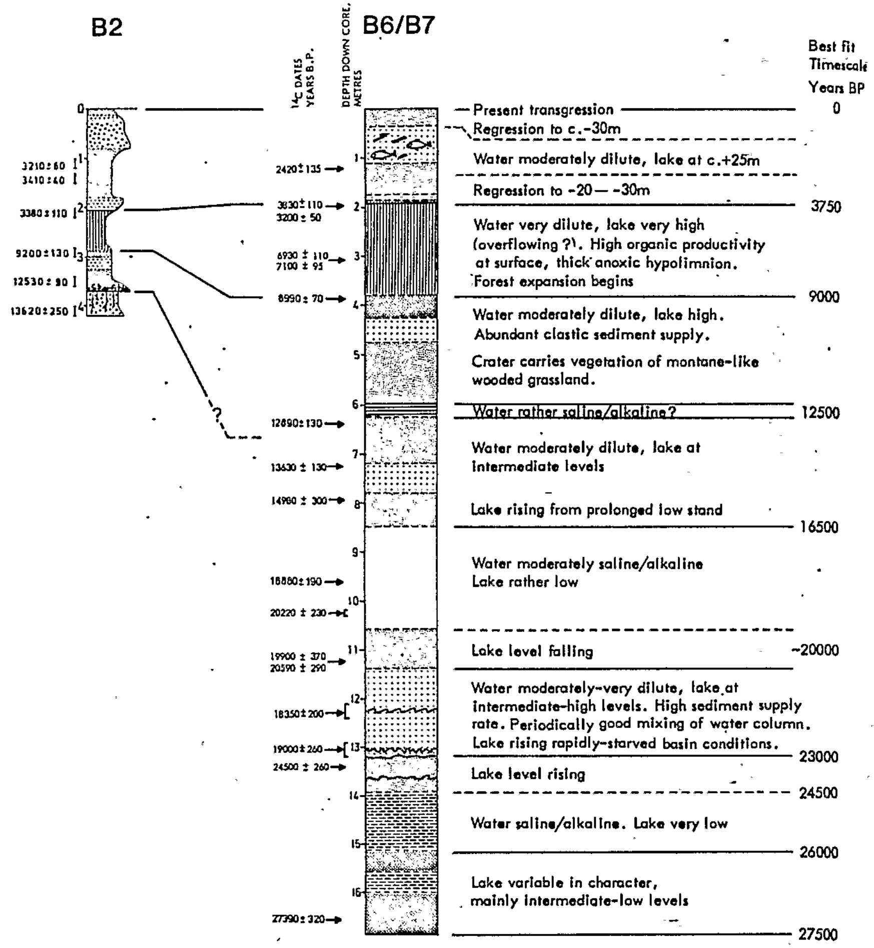


Figure 5. Correlation of cores B2, B6 and B7 with palaeolimnological and palaeoenvironmental summary and best-fit timescale based on all presently available chronological data.

be the remnants of partially combusted material charred by grass fires. Wafting of charred cuticle into the atmosphere by intense convection during bush fires may have transported many of the fragments to the lake from grassland outside the crater.

Grass cuticle is particularly abundant in B6 and B7 below 4.0 m and in B2 below 2.9 m, an abundance that reinforces the impression provided by the pollen spectra of extensive grasslands in the surrounding countryside. The dramatic decline in cuticle abundance at a time when arboreal pollen began

to increase must therefore reflect the disappearance of these grasslands as they were over-run by the expanding forest.

One particularly significant result has already emerged from the smear slide and preliminary SEM work. Amongst the cuticle present in some sections of the cores are fragments of species that appear to belong to the sub-family Pooideae (Figs.2 and 3). At present in tropical Africa, pooid grasses are confined to montane areas at altitudes exceeding 1 500 m, this distribution being primarily controlled by temperature (Livingstone & Clayton 1980, Palmer 1982). No pooid grasses are known from modern Ghana, their nearest occurrence being the Cameroun highlands. Recognition of probable pooid cuticle (including some large fragments) in the Bosumtwi sediments is thus of extreme palaeoclimatic importance, since it indicates the presence of these grasses around the lake. By analogy with their modern distribution, this implies mean annual temperatures considerably lower than today's.

SOME PRELIMINARY CONCLUSIONS

Although analysis of the cores is still far from complete, it is possible on the basis of the mineral stratigraphy, sedimentology and macrophyte remains to reach a number of preliminary conclusions about changes in the state of Lake Bosumtwi and in the climate and vegetation of the region over the past 27 500 years. These conclusions are summarised in Figure 5 and a revised lake level curve for the last 13 500 years, based on new data presented here, is shown in Figure 6.

1. 16.9-14.0 m (c.27 500-24 500 BP). The lowermost part of B6 and B7 seems to have accumulated in a lake rather different in character from that of any subsequent period. This section contains, for example, the only aragonite, indicating the existence of a rather saline lake with waters having an Mg/Ca ratio of at least 12 (Müller et al. 1972) – an almost three-fold increase over present-day values. Bosumtwi must thus have been very low during the periods of aragonite precipitation.

Below and between the two aragonite zones, the association of vivianite-bearing muds with manganosiderite laminae is similar to the early Holocene section. The older sediments also contain numerous diatom-rich laminae, however, which are absent from the Holocene. We cannot explain these differences.

2. 14.0-11.0 m (c.24 500-20 000 BP). The sediments in this interval are mainly of clastic origin and show slump features as well as primary depositional dip, suggesting accumulation on a turbidite fan. The only evidence of bioturbation in B6/B7 also occurs here, indicating a period when bottom waters contained sufficient oxygen to allow habitation by benthic organisms. This could

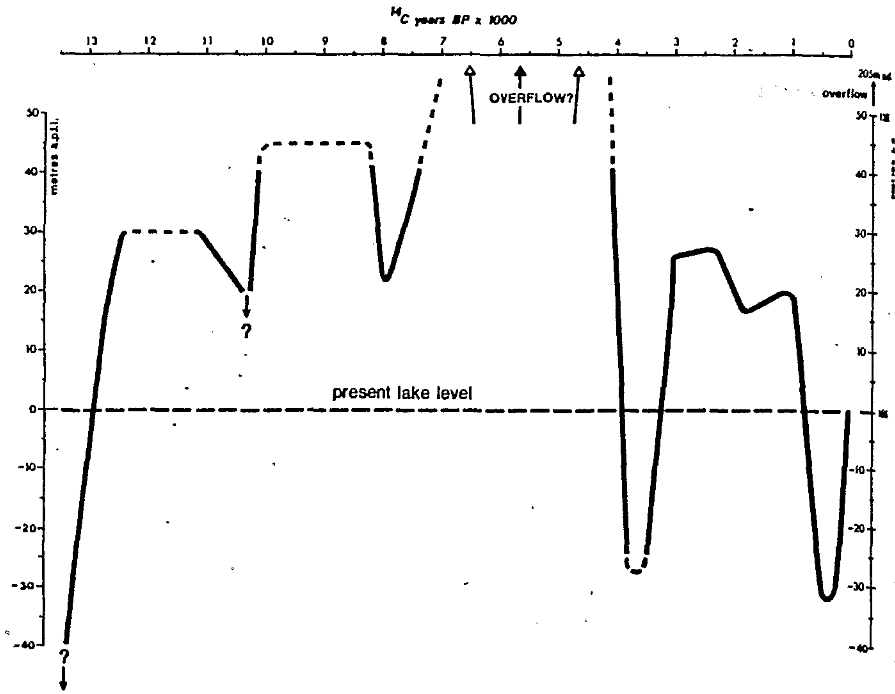


Figure 6. Revised lake level variation curve for the last 13 500 years. Based on Talbot & Delibrias (1980) and new data presented here.

be interpreted as an indication of particularly low water, but the only endogenic mineral present is calcite, mainly in just trace amounts, which suggests that the lake waters were relatively dilute. It is possible, therefore, that a particularly active circulation system prevented the development of anoxic bottom waters at this time.

The bioturbated and scoured section also seems to represent a period of reduced sedimentation, probably due to a particularly rapid transgression causing sediment to be trapped in newly formed estuaries. Under these conditions the centre of the lake would have been temporarily starved of sediment.

Above the bioturbated horizon comes an interval where the ^{14}C dates provide rather poor stratigraphic resolution. This is most probably due to particularly high sedimentation rates on a turbidite fan.

3. 11.0-8.5 m (c.20 000-16 500 BP). The calcite content of the sediments increases rapidly above 11.0 m and is joined above 10.5 m by Mg-calcite, dolomite and phillipsite. Clastic turbidites continue to be prominent.

At present this section of the cores must be interpreted with caution. The

appearance of Mg-calcite indicates increasing salinity and the association of this mineral with dolomite and zeolite initially suggests a saline, alkaline lake.

However, as noted earlier, aragonite is absent, so bearing in mind our earlier discussion on diagenetic effects, it is possible that salinities were not as high as would be implied by the typical lacustrine carbonate sequence. We hesitate, therefore, to suggest that the lake was exceptionally low during this time interval.

4. 8.5-5.95 m (16.500-12 500 BP). A change to more dilute waters is apparent above 8.5 m, the phillipsite disappearing and calcite taking the place of the Mg-bearing carbonates. Phillipsite reappears above 6.3 m, however, where the first of three thin brecciated carbonate crusts occurs. This section has proved particularly difficult to interpret. The red-stained, reworked crusts and associated Mg-calcite, zeolite and gel all suggest low water levels, high alkalinity and salinity and thus a rather abrupt regression of the lake. On the other hand, these sediments were apparently deposited c.13 000-12 500 years BP (Fig.2), at which time it is known from abundant and consistent evidence from both core B2 and several outcrops, that the lake was just recovering from a prolonged period of low water. We cannot explain this anomaly. The date of 12 890 ± 130 BP from just below the crustose horizon may be erroneous. Alternatively, diagenetic processes might be entirely responsible for the development of the crusts and their accompanying mineral assemblage.

5. 5.95-3.8 m (12 500-9 000 BP). Lake level evidence (Talbot & Delibrias 1980) indicates that Bosumtwi was high during the accumulation of this interval. Relatively dilute waters are reflected in the presence of calcite as the sole endogenic mineral. Above 4.6 m there is no calcite, but abundant vivianite and manganosiderite. The former is definitely of diagenetic origin and indicates anoxic conditions in the bottom sediments.

Pooid grass cuticle fragments and pollen of *Olea hochstetteri* are common in these sediments, suggesting a relatively low mean annual temperature. It is probable tht the high lake levels at this time were in part due to lowered evaporation rates.

6. 3.8-1.9 m (9 000-3 750 BP). Excellent preservation of abundant algal filaments and pigment indicates permanently reducing conditions at the lake bottom and high productivity in the surface waters. Anoxic bottom conditions are confirmed by the accompanying phosphate minerals, all of which form under reducing conditions. The homogeneity of this algal mud indicates a stable water body, especially in the deepest part of the lake. A more marginal facies of the sapropel is also known from one outcrop at 2 m above present lake level. The upper part of this sequence (which is truncated by an erosion surface) has been dated to 5 000 ± 120 BP (Talbot & Delibrias 1980). At this time, and for at least several hundred years before, anoxic conditions

characterised the bottom 80 m or more of the water column.

The complete absence of calcite from this section of the cores suggests dilute water, so the lake must have been particularly high, perhaps even overflowing, in which case it would have had a maximum depth of 185 m. Such depths would certainly have favoured a stable water column.

Accumulation of the sapropel began as forest started to replace the wooded grassland that had clothed the Bosumtwi crater for several thousand years (Fig.2). Establishment of a continuous forest cover had the effect of greatly reducing clastic sediment supply to the lake – hence the drastic reduction in accumulation rate after 9 000 BP. Contrasting climatic conditions may have been the reason for these changes. High sediment yields with the deposition of laminites suggests a climate of marked seasonal contrasts. The extreme homogeneity of the sapropel is probably the result of minimal seasonal contrasts. Evenly distributed, and presumably relatively abundant rainfall could have been a major reason for forest expansion at this time.

7. 1.9 m-top (3 750 BP-present). A dramatic decline in lake level occurred at about 3 750 years BP; core B2 indicates that the lake surface fell from its previously very high levels to between –20 or –30 m. This fall seems to have brought a sudden end to sapropel accumulation as laminated turbiditic sediments, initially rather rich in manganosiderite, abruptly overlie the sapropel. Although a new transgression to around +25 m followed (Talbot & Delibrias, 1980), Lake Bosumtwi never again achieved its early to mid-Holocene levels. A further period of aridity caused another major regression to about –30 m. The lake is still recovering from this last event.

One important point to note is that although a forest vegetation has persisted in the crater to the present day, sapropel accumulation has ceased. This demonstrates that vegetation cover has not been the only control on sedimentation in Bosumtwi (see 6 above). Resumption of laminite deposition presumably reflects the return to a more seasonally-contrasted climatic regime of the type that still exists in the region today.

ACKNOWLEDGEMENTS

We are indebted to many people inside and outside Ghana for assistance and encouragement with this work. M.R.T. gratefully acknowledges financial and material support for fieldwork from the University of Ghana, Legon. Subsequent support came from the University of Leeds, grants from the University's Sub-Committee for Research in Africa allowing him to visit Ghana and Duke University. Aase Scheie and Lillian Berberg, both of the Geological Institute, University of Bergen, are thanked for their very considerable assistance in the SEM and XRD examination of the core sediments.

D.A.L., P.G.P, and J.M.M. in particular thank Mr G.O.Kesse, Director of the Ghana Geological Survey Department, Mr Samuel Whyte of the Department of Zoology, University of Ghana and Dr John Bilson of the Allen Clinic, Kumasi. The University of Ghana at Legon and The University of Science and Technology at Kumasi provided living accommodation, herbarium facilities, library facilities, inspiration and encouragement. They wish to thank their field assistants at Abonu, Mr Alexis Osei Kwame Bonsu, Mr Patric Elliot Ofusu, Mr Johnson Kwame Opoku Afrifah and Mr John Bilson. They were much fortified by the kind hospitality of scientific colleagues and friends in Ghana, and by that of the village of Abonu, especially the extended family of the late Mr Kwaku Adasa and his wife Mrs Adwoa Serwah. Financial support was provided by grants from the National Science Foundation of the United States and from Duke University.

John Hall has proved an inspiring source of information and discussion on the vegetation of Ghana. Dr Kerry Kelts, ETH, Zurich, is thanked for comments on an earlier version of this paper.

REFERENCES

Eberli, G., J.A.McKenzie, K.Kelts & J.Pika 1982. Holocene sedimentation in Great Salt Lake, Utah: the relationship between carbonate mineralogy, percentage carbonate and salinity. *Abs. 3rd European Meetings IAS, Copenhagen:*10.

Eugster, H.P. 1982. Climatic significance of lake and evaporite deposits. In Climate in Earth History, National Research Council (US). Geophysical Study Committee. National Academy Press, Washington:105-111.

Eugster, H.P. & L.R.Hardie 1978. Saline lakes. In A.Lerman (ed.), *Lakes-Chemistry, Geology, Physics.* Springer Verlag, New York:237-239.

Flight, C. 1976. The Kintampo Culture and its place in the economic prehistory of West Africa. In J.R.Harland, J.M.J.de Wet & A.B.L.Stempler (eds.), *The Origins of African Plant Domestication:*211-221. Mouton, The Hague.

Hall, J.B. & M.D.Swaine 1976. Classification and ecology of closed-canopy tropical forest in Ghana. *Journal of Ecology* 64:913-952.

Hall, J.B., M.D.Swaine & M.R.Talbot 1978. An early Holocene leaf flora from Lake Bosumtwi, Ghana. *Palaeogeogr., Palaeoclimatol., Palaeoecol.* 24:247-261.

Hecky, R.E. & E.T.Degens 1973. Late Pleistocene-Holocene chemical stratigraphy and paleolimnology of the rift valley lakes of central Africa. *Woods Hole Oceanographic Inst. Tech. Rept.* WHOI-73-28, 93 pp.

Kelts, K. & K.L.Hsü 1978. Freshwater carbonate sedimentation. In A.Lerman (ed.), *Lakes-Chemistry, Geology, Physics:*295-323. Springer Verlag, New York.

Kelts, K. & J.A.McKenzie 1982. Diagenetic dolomite formation in Quaternary anoxic diatomaceous muds of DSDP Leg 64, Gulf of California. *Initial Rept. DSDP* 64 pt.2: 553-569.

Livingstone, D.A. 1980. Late Quaternary pollen from Lake Bosumtwi, Ghana. *Abs. 5th International Palynological Conference,* Cambridge, p.229.

Livingstone, D.A. & W.D.Clayton 1980. An altitudinal cline in tropical African grass floras and its paleoecological significance. *Quaternary Research* 13:392-402.

Maley, J. 1981. Etudes palynologiques dans le Bassin du Tchad et Paléoclimatologie de l'Afrique nord tropicale de 30 000 ans à l'époque actuelle. *Trav. et Doc. ORSTOM* 129, 586pp.

Maley, J. & D.A.Livingstone 1983. Extension d'une element montagnard dans le sud du Ghana (Afrique de l'Ouest) au Pleistocène supérieur et à l'Holocène inférieur: premières données polliniques. *C.R. Acad. Sci.,* série 3, 296:251-256.

Müller, G., G.Irion & U.Förstner 1972. Formation and diagenesis of inorganic Ca-Mg carbonates in the lacustrine environment. *Naturwissenschaften* 59:158-164.
Palmer, P.G. 1976. Grass cuticles: a new paleoecological tool from East African lake sediments. *Canadian Journal of Botany* 54:1725-1734.
Palmer, P.G. 1982. A stratigraphic analysis of fossil grass cuticles from Lake Bosumtwi, Ghana, West Africa, and its paleoecological significance. *Abs. 11th INQUA Congress* Moscow 1:244.
Surdam, R.C. & R.A.Sheppard 1978. Zeolites in saline, alkaline lake deposits. In L.B. Sand & F.A.Mumpton (eds.), *Natural Zeolites: Occurrence, Properties, Use.* Pergamon Press, Oxford:145-174.
Talbot, M.R. 1976. Late Quaternary sedimentation in Lake Bosumtwi, Ghana. *20th Ann. Rept. Res. Inst. Afr. Geol.*:69-73. Univ. Leeds.
Talbot, M.R. 1983. Late Quaternary sedimentation in a tropical meromictic lake – Lake Bosumtwi, Ghana. *Abs. 4th European Meeting IAS, Split*, p.162.
Talbot, M.R. & G.Delibrias 1980. A new late Pleistocene-Holocene water-level curve for Lake Bosumtwi, Ghana. *Earth Planet. Sci. Letters* 47:336-344.
Talbot, M.R. & J.B.Hall 1981. Further late Quaternary leaf fossils from Lake Bosumtwi, Ghana. *Palaeoecology of Africa* 13:83-92.

TRENDS OF PLEISTOCENE AND HOLOCENE RESEARCH ON THE SAHARA*

E. SCHULZ
Institute of Geography, University of Würzburg, W.Germany

Working one's way through the large volume of literature that has been published in recent years on the development of the Sahara, one gets the impression that detailed chronologies are already possible for individual regions of the Sahara. In a number of papers, the discussion is even concerned with the dating of events within decades (Gayh & Jäkel 1974, Jäkel 1979, Maley 1977, Pachur 1980, Commelin & Petit-Maire 1979, Rognon 1976, 1976a, Van Zinderen Bakker & Maley 1979).

A comparison of the various curves describing climatic developments indicates a rather high degree of uniformity for the whole Sahara. This may merely be a reflection of the fact, however, that most of the curves have been drawn to be representative for large parts of the Sahara. Evidence of differential developments in various regions of the desert is less common (Flohn 1980).

Parallel to the survey studies, there is an intense discussion going on concerning discrepancies in the description of individual climatic phases. This may be shown by two examples.

In 1976, Cour & Duzer arrived at the conclusion that the Holocene of the southern and central Sahara had been continuously arid, if not hyperarid. Their contradiction of former interpretations of the vegetational history of the region was based on the analysis of the present pollen rain in relation to the Holocene pollen spectra of the sebcha of Taoudenni (northern Mali).

Similarly, deep sea cores taken off the coast of West Africa are interpreted by Sarnthein (1979) to indicate arid conditions during the Würm and Holocene, whereas Diester-Haass (1979) finds evidence for relatively humid conditions. These opposing views are due to different opinions as to what extent climatic information can be read from grain size differences in sediments, from so-called desert quartz and from changes in the clay mineralogy of the cores. It has to be appreciated that the discussion has been carried on openly in one and the same journal.

* Re-printed with permission of the author and the chairman of the INQUA 'Regional Subcommission for the Study of the Holocene of the Circum-Mediterranean Area' from circular no.11 (July 1982) of the Subcommission.

extrait de PALAEOECOLOGY OF AFRICA, Rotterdam, 1984
vol.16, p.173-192.

PRELIMINARY RESULTS FROM SEDIMENT CORES FROM LAKE BOSUMTWI, GHANA

M. R. TALBOT[1]/ D. A. LIVINGSTONE[2] / PATRICIA G. PALMER[3]
J. MALEY[4] / JOHN M. MELACK[5] / G. DELIBRIAS[6] / S. GULLIKSEN[7]

1. Geologisk Institutt, Avd. A, Universitet i Bergen, Allégt.41, 5014 Bergen, Norway.
2. Department of Zoology, Duke University, Durham, North Carolina 27706, USA.
3. Department of Biology, Louisburg College, Louisburg, North Carolina 27549, USA.
4. ORSTOM, Laboratoire de Palynologie du CNRS, Université des Sciences et Technique du Languedoc, Montpellier 34060, France.
5. Department of Biological Sciences, University of California, Santa Barbara, California 93106, USA.
6. Centre des Faibles Radioactivités, Labo. Mixte CNRS-CEA, Gif sur Yvette 91190, France.
7. Laboratoriet for Radiologisk Datering, Fysikkseksjonen, Sem. Sælandsv.5, 7034 Trondheim, Norway.

ABSTRACT

Results are presented from the analysis of three sediment cores from Lake Bosumtwi, Ghana, using mineralogy, sedimentology, palynology and grass cuticle content. The longest core is 16.9 m long and spans approximately the last 27 500 years of the lake's history. Apart from minerals of clastic origin, the sediments contain a variety of endogenetic and diagenetic minerals, particularly carbonates, with phosphates and zeolites also prominent in some sections. Mineral content and sedimentology are used to reconstruct the history of the lake. The oldest part of the sequence, from c.27 500 to 24 500 BP, is poorly understood, but seems to contain evidence of both high and low water levels. Between 23 000 and 19 500 BP the lake water was relatively dilute, but became more saline and alkaline between 19 000 and 16 000 BP when the lake level must have been low. Regressions of the lake are registered just before 13 500 BP, at c.3 750 BP and after 2 000 BP. Maximum lake level seems to have occurred between 9 000 and 4 000 BP, when the lake waters were dilute.

Grass pollen dominates the pollen assemblage before c.9 000 BP indicating the existence of grassland around Bosumtwi. A dramatic rise in the abundance of arboreal pollen suggests that forest rapidly replaced grassland after 9 000 BP. Patches of montane-like forest within the grassland are suggested by the presence of *Olea hochstetteri* pollen in late Pleistocene-early Holocene sections of the cores.

Grass cuticle fragments are prominent in sediments deposited before c.9 000 BP, confirming the pollen evidence that grassland formerly existed around the Bosumtwi crater. Of particular interest is the occurrence of pooid grass cuticle in the Pleistocene-early Holocene sediments. Pooid grasses, and the pollen of *O.hochstetteri*, imply palaeotemperatures several degrees centigrade lower than today's in southern Ghana.

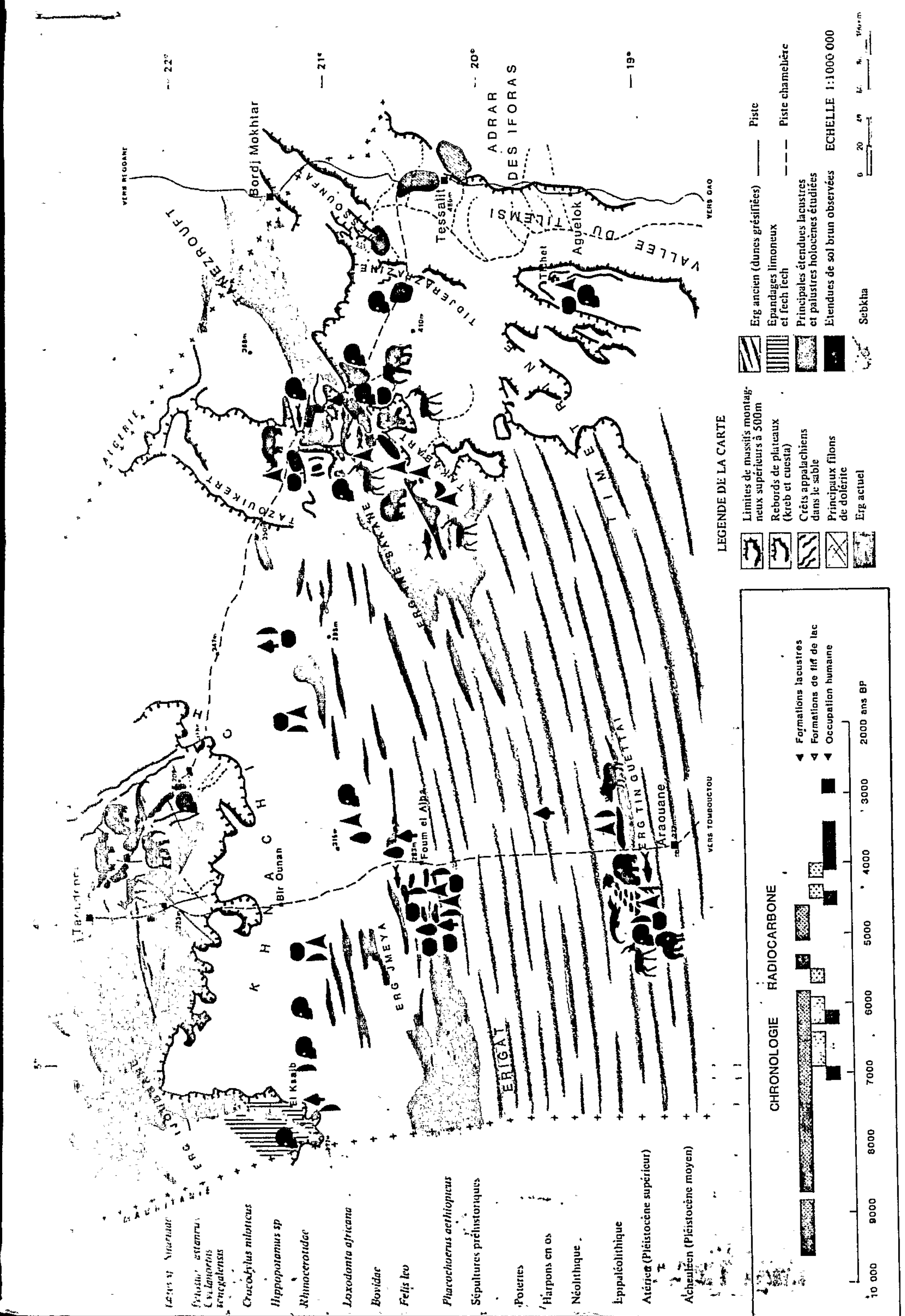
RESUME.

Des résultats sont présentés d'après l'analyse des sédiments de trois carottes prélevées dans le lac Bosumtwi au Ghana, faisant appel à la minéralogie, la sédimentologie, la palynologie et aux cuticules de Gramineae. La plus longue carotte est de 16,9m et couvre approximativement les derniers 27500 ans de l'histoire du lac. En plus des minéraux d'origine clastique, les sédiments contiennent une variété de minéraux endogènes et diagénétiques, particulièrement des carbonates, avec aussi des phosphates et des zéolites dans certaines sections. La partie la plus ancienne de la carotte, de c. 27500 à 24500 BP, est peu définie, mais semble contenir des preuves de niveaux lacustres hauts et bas. Entre 23000 et 19500 BP l'eau du lac était relativement diluée, mais elle devint plus saline et alcaline entre 19000 et 16000 BP lorsque le lac devait être bas. Des régressions lacustres sont notées juste avant 13500BP ,à c.3750 BP et après 2000 BP. Le niveau lacustre maximum semble s'être produit entre 9000 et 4000 BP, lorsque les eaux du lac étaient diluées.

Les pollens d'herbes dominent les spectres polliniques avant c.9000 BP, indiquant l'existence de prairie autour du Bosumtwi. Une augmentation remarquable de l'abondance des pollens d'arbres après 9000 BP suggère que la forêt a alors rapidement remplacé la prairie. La présence de pollens d'Olea hochstetteri dans les couches du Pléistocène supérieur et de l'Holocène inférieur suggère que des taches de forêt de type montagnard existait dans la prairie.

Des fragments de cuticules de Gramineae se rencontrent en quantité dans les sédiments déposés avant 9000 BP, confirmant les données polliniques suivant lesquelles une prairie existait alors autour du cratère du Bosumtwi. Il est particulièrement intéressant de noter la présence de cuticules de Pooideae dans les sédiments du Pléistocène et de l'Holocène inférieur. Des Gramineae de la tribu des Pooideae et les pollens d'Olea hochstetteri impliquent des paléotempératures de plusieurs degrés centigrades inférieurs à celles actuelles du sud du Ghana.

3



Le Sahara malien à l'Holocène, pp. 311-316.

Ved:

PALAEOECOLOGY OF AFRICA

AND THE SURROUNDING ISLANDS

VOLUME 16

edited by

J.A.COETZEE (D.Sc.)
E.M.VAN ZINDEREN BAKKER SR (Phil.Nat.D., D.Sc.h.c.)
Institute for Environmental Sciences
University of the Orange Free State, Bloemfontein

A.A.BALKEMA/ROTTERDAM/BOSTON/1984