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LATE PLEISTOCENE-HOLOCENE CHEMICAL STRATIGRAPHY AND PALEOLIMNOLOGY OF THE RIFT VALLEY LAKES OF CENTRAL AFRICA

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ABSTRACT

The interaction of climate and geology in Central Africa during Late Pleistocene and Holocene is examined. The study is based on sedimentological and limnological work on the main lakes of the Western Branch of the East African Rift Valley, particularly Lake Kivu. Changes in sediment chemistry, mineralogy and diatom assemblage provide a detailed histogram of lake level oscillations. Calculations indicate that the drop in lake level could be as high as 600 m for Tanganyika and 400 m for Kivu. Fluctuations in water levels are the means for reconstruction of climatic events in tropical Africa of the last 15,000 years. Paleoclimatic comparison between tropical and temperate zones reveals that pluvial times coincide with the prominent interstadials in Europe, e.g. Bølling, Allerød, Climatic Optimum, and reversely, cool and dry periods in equatorial Africa with ice ages in the Northern Hemisphere.

The African climatic sequence of pluvials and interpluvials is accompanied by corresponding periods of hydrothermal activity and quiescence. This may suggest that rain water exercises control on hydrothermal activities.

INTRODUCTION

Eastern Africa since Miocene times (Bishop, 1963) has been a focus of intense tectonic and volcanic activity which has produced a dramatic topography. Broad regional uplift has concomitantly resulted in large rifted troughs formed by the relative downward movement of crustal blocks along normal faults (Cloos, 1939). This pattern of rifting can be traced across the African continent for nearly 3,000 km from the Zambesi River in the south to the Red Sea in the north. Recent characterization of the Red Sea as a juvenile oceanic rift with active spreading has directed interest to the present and future status of the African rift system and its relationship to the world-wide system of mid-ocean rifts (Degens et al., 1971). Uplift and rifting in eastern Africa as elsewhere have been accompanied by profuse volcanism which has deeply buried large areas of Precambrian basement under Tertiary, Pleistocene, and Recent lava and ash. Active volcanoes are still a prominent feature of the landscape.

The African rift valleys are generally grouped into four separate sections: (1) the Nyasa section, (2) the Western Rift, (3) the Eastern or Gregory Rift and (4) the Lake Rudolf and Ethiopian section (Holmes, 1965). These Rift valleys delimit drainage basins, and they contain lakes the size of which is partially dependent on the climatic regime over the respective areas. Lake Malawi occupies a large area of the Nyasa section, while Lakes Tanganyika, Kivu, Edward, and Albert are strung out along the arcuate Western Rift valley (Fig. 1). (Lakes Edward and Albert have recently been renamed as Lake Idi Amin Dada and Lake Mobutu Sese Seko respectively, but they will be referred to in this paper by their former names to facilitate referring to the earlier literature.) These two rift valleys have comparatively high rainfall over their catchment, and they are presently regions of external drainage. The lakes of these rifts are tributary to three of the largest rivers in the world, the Nile (Lakes Edward and Albert), the Zaïre (Lakes Kivu and Tanganyika), and the Zambesi (Lake Malawi). The arid Eastern Rift and Ethiopian Rift are large basins of internal drainage where only small saline lakes dot the vast area. Lake Rudolf in the Ethiopian Rift is an exception, but it has been gradually shrinking for the last 5,000 years (Livingstone and Kendall, 1968). The other large East African lake, Victoria, occupies a gentle downwarp between the Eastern and Western Rift valleys, and likewise, it is a product of the same large-scale earth movements which led to the formation of the rifts.

The antiquity of these rift basins, their tropical position, and their geology present earth historians a unique opportunity to examine the interaction of climate and geology on the continents through a significant portion of geological time. Only in the oceans are substantially longer records to be found. Knowledge of the history of the oceans is growing explosively; however, the correlative continental record is poorly understood. There is still controversy over the correlation between the evidence for numerous Pleistocene glacial cycles found in the oceanic record and the concept of only five or so major glaciations which the continental record so far examined has preserved. To reconcile such differences a continuous record from the continents is required and a tropical setting is preferred for maximum climatic information. Boreholes and seismic records in the Albert Basin have demonstrated the presence of over 1.2 km of sediments (Bishop, 1963; and Wong, unpublished).

Seismic profiles from the other lakes show accumulations of similar magnitude (Table 1; Degens et al., 1971, 1973; Wong, unpublished). Thick stratigraphic sections are also found in the Eastern and Ethiopian Rifts, but the aridity of the climate and the shallowness of the modern lakes reduce the probability of recovering a continuous record.

Thus far, the lakes of the Eastern Rift have received the greater attention of paleohistorians because of the abundance of human cultural sites and the accessibility of the sedimentary record in the shallow lakes to the unsophisticated coring techniques which can be brought to these remote areas. Such efforts have been successful in reconstructing a rather coherent view of the history of that rift over the past 30,000 years (Butzer et al., 1972; Richardson and Richardson, 1972; Richardson, unpublished). By comparison the paleolimnology of the Western Rift is very poorly known. Published information is limited to the description of a 12 m core recovered from Lake Tanganyika (Livingstone, 1965) and descriptions of exposed stratigraphic sections in the Albert basin (see Bishop, 1963, for a summary). The diatom stratigraphy of a 12 m core from Lake Albert which covers the past 25,000 years is also presently under study (Thomas Harvey, personal communication).

Recent expeditions from the Woods Hole Oceanographic Institution in 1970 to Lake Tanganyika, 1971 to Lake Kivu, and 1972 to Lakes Kivu, Edward and Albert, have returned gravity and piston cores of varying lengths from 1 m to 8 m. In addition a 1972 expedition by the Laboratorium für Sedimentforschung of Heidelberg University has recovered a suite of gravity cores up to 3 m in length from Lake Malawi.

This paper presents the chemical stratigraphy of representative cores from each of these lakes along with the biochemical and diatom stratigraphy of Lake Kivu. Microfossil studies on the other cores are still in progress. Based on the available data, a provisional reconstruction of the history of these lakes for the past 15,000 years is also attempted. Although the available material represents a miniscule fraction of the potential stratigraphic record in these lakes, it does cover the Late Glacial-Holocene transition and shows the effects of Pleistocene climatic fluctuations on these lakes. In addition, interpretations of the stratigraphic evidence will have significance to theories which have been proposed to explain various unusual aspects of the biology and limnology of the present lakes, such as the species flocks in Tanganyika and Malawi (Brooks, 1950; Fryer, 1959) and the immense methane reservoir and depauperate fauna in Kivu (Degens et al., 1973; Deuser et al., 1973; Verbeke, 1957).

MODERN LAKES

To provide a framework for the historical reconstructions to follow, a brief synopsis of the limnology of the modern lakes is necessary. The information presented in Table 1 and Table 2 and reviewed below was extracted from a diverse literature; and the interested reader is directed to the original sources for a more detailed discussion of the various lakes (see for Lake Albert: Talling, 1963; Verbeke, 1957; for Lake Edward: Damas, 1937; Verbeke, 1957; for Lake Kivu: Damas, 1937; Schmitz and Kufferath, 1955; Verbeke, 1957; Degens et al., 1973; Deuser et al., 1973; for Lake Tanganyika: Beauchamp, 1939; Capart, 1952; Coulter, 1963; Degens et al., 1971; and for Lake Malawi: Beauchamp, 1953; Eccles, 1962).

WATER CHEMISTRY

The chemical composition of lake waters from Albert, Edward, Kivu, and Tanganyika (Table 1) is remarkable for the high potassium and magnesium values relative to the other cations. Kilham (1971a) proposes that these unusual ionic ratios find their origin in the volcanic rocks of the area. The Virunga volcanic field, which forms the drainage divide between Edward and Kivu, and the Toro-Ankole field, of Southwestern Uganda in the Edward basin, are characterized by their ultrabasic, potassic rocks. These rocks have $K_2O \geq Na_2O$, and the Ca and Mg contents are much higher than for average crustal material (Higazy, 1954; Bell and Powell, 1969). These volcanic rocks are also rich in TiO_2 and P_2O_5, and have relatively great abundances of Sr, Ba, Rb and Zr, which are uncommon for such ultrabasic rocks. These facts will later be important to interpretations of sedimentary chemistry.

Kilham (1971a) and Degens et al. (1973) believe that intense weathering of these rocks by hot, CO_2-charged ground waters with subsequent loss of Ca by precipitation will yield the observed composition in a straightforward manner for Kivu and Edward. Albert and Tanganyika will, in turn, take on these characteristics as their major tributaries, the Semliki and Ruzizi respectively, arise in Edward and Kivu. These tributaries supply a considerable portion of the water budget and an even greater proportion of the salts to Albert and Tanganyika. Other tributaries deriving from the Precambrian granitic and metamorphic rocks of the plateaus above the rift escarpments are more dilute and of a quite different composition (cf. Malagarisi and Ruzizi in Table 1). The relative contribution of salts by these tributaries must be minor, as Albert and Tanganyika strongly resemble Kivu and Edward in their major ion composition.

Lake Malawi farther to the south and with no connections with the aforementioned volcanic provinces has a quite different water chemistry. The volcanism, unusual petrology, and hydrothermal spring activity which so dominate the lakes in the Western Rift are less significant in the Nyasan Rift. Hot springs are known along the rift valley, but they are unaccompanied by active volcanism (Von Herzen and Vacquier, 1967). The lake is much more dilute than those of the Western Rift, suggesting perhaps that weathering rates are lower because the ground waters of the region are not enriched in volcanic-derived CO_2. However, rainfall is generally higher, temperatures lower, and the lake volume smaller than Tanganyika, and the dilution may simply reflect lower residence times. The ionic composition of the waters is representative of most African waters, although Talling and Talling (1965) do note that the chloride is relatively low. According to Hecky and Kilham (1973), low Cl/Na ratios in many African lakes are indicative of fairly humid conditions in the drainage basin.

LIMNOLOGY

Kivu

Damas (1937) first described the unusual hydrography of Lake Kivu which has since been studied in ever increasing detail (Degens et al., 1973). Below 70 m, temperature and salinity increase with depth. The density increase produced by higher salinities just offsets the density decrease inherent with the higher temperatures resulting in a stable thermohaline density structure. The greater diffusivity of heat relative to salts yields a series of convecting cells (Turner, 1969), and the increases in temperature and salinity with depth occur in a steplike fashion. Density differences between these cells are minute and can often be resolved at a decimeter level (F. Newman, unpublished data), and the density differential between even surface water and water below 350 m is not large, being on the order of .002 g/cm^3 (Schmitz and Kufferath, 1955). At present, Kivu would seem to be a delicately balanced system which has persisted for at least the 40 years of observation. The history of this phenomenon over longer periods is an important consideration of the present stratigraphic work.

Meromixis in Kivu greatly reduces the exchange of water between the upper 70 m of water which circulate at least seasonally and the anoxic water of greater depths. The effect of this lack of mixing is dramatic, as the nutrient values in Table 1 attest. Surface waters are depleted in phosphorus and nitrogen by plankton production and subsequent sinking and decay below the thermohaloclines where regenerated nutrients are effectively trapped. Silicon, the other nutrient in Table 1, also increases with depth, but it gives a linear regression on Na, indicating that diatom production and resolution has little to do with the observed trends. As well as trapping nutrients, the deep waters also accumulate gases of volcanic (CO_2) and predominantly biological origin (H_2S and CH_4) (Deuser et al., 1973), which are held in solution by hydrostatic pressure. Waters below 300 m contain two liters of gas for each liter of water.

The loss of nutrients to the deep water trap profoundly influences the primary productivity of surface waters (Table 2). Primary production is less than one-fourth that of Edward and Albert, which are lakes of a similar size and chemical composition. In addition to the deep thermohaloclines, Kivu has a seasonal thermocline which is established at approx. 20 m depth during the warm, rainy seasons. Although temperature differences between the epilimnion and hypolimnion are slight, a very stable stratification can develop, as the density changes at these elevated tropical temperatures are relatively greater than at lower temperatures. Cool, dry seasons lead to full circulation to 50 m. Biological production is increased after these periods of mixing, and the transparency of the waters is reduced by the increased populations of plankton. The values listed in Table 2 pertain only to the pelagic communities of the lake. In Kivu as in Tanganyika (Coulter, 1963), the inshore waters are often characterized by intense plankton blooms; whereas such events are uncommon offshore in the greater part of the lake.

The fauna and flora of Kivu have several aspects which have caused prior comment (Verbeke, 1957). The fish fauna is depauperate with only 11 species, none of which is considered a pelagic, planktophagous species. Chaoborids are absent from the lake as they are from Tanganyika. The molluscs, esp. pelecypods, seem to have suffered recent extinctions, judging from the abundance of shells of species which are no longer extant in the lake. The planktonic diatom flora (Hustedt, 1949) is remarkable for the domination by the genus *Nitzschia*, a feature which is also shared by Tanganyika. Various possible causes, including volcanism, catastrophic overturns, unusual water chemistry, etc. have been suggested, and an historical perspective on the limnology of the lake should provide insight.

Tanganyika

Tanganyika is divided by a shallow structural sill (mean depth 500 m) into two basins, of which the southern is the deepest with a maximum depth of 1,470 m. It is thermally stratified with a perennial thermocline lying at approx. 100 m. Although the thermocline lies at the same depth in both basins, the lower limit of detectable dissolved oxygen is 200 m in the south basin of the lake and 100 m in the north, indicating that mixing phenomena for the basins are somewhat different. Temperature differences across the seasonal and perennial thermoclines are slight, ranging from 2-3°C in the wet season to 0.5-1°C across the perennial thermocline in the dry season. This minor density differential combined with the enormous volumes of water to be mixed (Coulter, 1963) is sufficient to maintain the thermal stratification over at least periods of several years. The persistence of stratification over longer periods is unknown. Brooks (1950) suggests that the thermocline stood somewhat deeper in 1895. Beauchamp (1939) believed that the lake might mix completely irregularly over long periods of time when prolonged intervals of cooler climate allowed mixing. Others (e.g. Talling, 1963) have suggested that cooler tributaries and nocturnal or seasonal cooling of littoral waters might yield downslope density currents which would continuously introduce cooler waters to the depths. In offshore areas internal waves, generated by tilting of the thermocline under meteorological stresses, may account for much of the mixing of hypolimnetic waters into the epilimnion (Coulter, 1963).

Despite the perennial thermocline, mixing processes are sufficient to prevent any significant stratification of the major chemical constituents. Only those minor consitutents subject to biological uptake and release exhibit stratification. The deep waters constitute an enormous store of nutrients which are unavailable to planktonic populations in the greater part of the lake. Coulter (1963) and others consider the offshore waters to be ultra-oligotrophic, whereas the littoral regions and areas of seasonal upwelling at the southern end of the lake are eutrophic. No primary productivity measurements have been done in the lake, but the diatom flora is similar to Kivu and the transparency is even greater, suggesting that the offshore areas are even less productive than Kivu. Alteration of the present hydrographic regime which would permit greater mixing across the thermocline could substantially alter the productivity of the lake, and evidence of such might be found in the stratigraphic record.

The Ruzizi River is the major tributary to the lake and determines its chemical composition. The lake has been closed in historical times, and most authors agree that without the Ruzizi and the Kivu drainage, Tanganyika would be closed and more saline today. The acquisition of the Ruzizi coincided with the blockage of the formerly northward flowing drainage in the Kivu basin by the imposition of the Virunga volcanoes which created Lake Kivu. Moore (1903) originally ventured a date of 15,000 years B.P. based on the rugged appearance of the volcanoes. Subsequent work has moved the date of the Virunga event substantially back into the Pleistocene (Brooks, 1950). Hydrological linkage via the Ruzizi between events in the Kivu basin and the status of Tanganyika is a most interesting problem which will find resolution in stratigraphic studies on the sediments of the two lakes.

Biologically, Lake Tanganyika is exceptional among the East African lakes for the large number of endemic genera and species (Brooks, 1950). Although species flocks among fish are present in other lakes of the region, especially Malawi and Victoria, at the generic level they are much more restricted. In addition to fish, nearly all the major faunal groups have a high degree of endemicity, which is not the case in the other lakes. The antiquity of the lake is most responsible for the evolutionary diversification, but temporal variability in lake levels and water chemistry are thought to have played a role. Stratigraphic studies will contribute greatly to understanding the relative importance of geographical and temporal isolating mechanisms to speciation in Tanganyika.

Edward

Lake Edward lies to the north of the Virunga volcanic dam. It receives runoff from that area via the Ruchuru River as well as from Lake George to the east via the Kazinga Channel. The lake's water chemistry is nearly identical with that of Kivu, but its mean depth is only forty meters, which has important consequences for productivity. In addition to the seasonal thermocline, there is a feeble deep thermocline which generally lies between 70-80 m. Below this depth, anoxic, nutrient-rich waters are encountered. However, depths greater than 80 m occupy only a very small area on the western side of the basin. Consequently, most of the basin is mixed to the bottom seasonally, and nutrient regeneration is quite efficient. The productivity of Edward is four times as great as that of Kivu and probably Tanganyika. The diatom flora of the offshore areas consists primarily of *Nitzschia fonticola* with *Stephanodiscus damasi* of secondary importance. In view of the productivity, this must be considered a eutrophic assemblage.

The fish fauna of the lake is depauperate when compared to its northern neighbor, Albert. Fossil-rich stratigraphic beds exposed in the basin indicate the fauna was more diverse in the early Pleistocene (Brooks, 1950) and was of Nilotic origin. Extinction is attributed to desiccation during the Pleistocene. Most recent faunal invasions have been from Lake Victoria, and recolonization by the Nile fauna in Lake Albert is precluded by the cataracts on the Semliki River.

Albert

Lake Albert receives a large portion of its total inflows from Lake Edward via the Semliki River. The major constituents are similar in proportions to Edward, but the nutrients are strikingly different (Talling and Talling, 1965). Albert waters are very rich in phosphate but impoverished in silica, whereas the opposite is true in Edward. These nutrient relations may account for the differences in the diatom flora between the lakes, as sediment assemblages from Albert are dominated by *Stephanodiscus astraea* as opposed to the *Nitzschia fonticola* in Edward. This is consistent with Kilham's (1971ab) hypothesis that *S. astraea* is a low silica specialist among planktonic diatoms. Primary production is also high in Albert, and again this is related in part to efficient nutrient regeneration. There is no well defined seasonal stratification in Albert, but Talling (1963) reports a peculiar short-lived stratification of sufficient magnitude to allow detectable accumulations of nutrients and depletion of oxygen in the deepest waters. This stratification evidently is initiated by the sinking of cooler inshore waters. Contrary to the other lakes, then, Albert achieves the ephemeral stratification it does have in the cool-dry season, and is most thoroughly mixed during the warm-wet seasons, when inshore waters tend to be warmer than offshore waters.

Malawi

Malawi is the least studied of the lakes considered. As mentioned above, its water chemistry is quite different from the other lakes. Its nutrient chemistry is most similar to Tanganyika in that all the nutrients are low in the offshore waters. The lake has a perennial deep thermocline as in Tanganyika, but its depth is much more variable and generally deeper than in Tanganyika. The depth of detectable oxygen (250 m) is also greater than Tanganyika; and, because the average depth is only 273 m, substantial areas of the lake are mixed to the sediment surface. This in conjunction with the apparent mobility of the thermoclines under the strong wind regime (Beauchamp, 1953) should mean a fairly high productivity. Although no primary productivity estimates are available, the dominant diatoms, *Melosira nyassensis* and *Stephanodiscus astraea*, suggest at least mesotrophic conditions.

Malawi has some species flocks among the fish, but these are largely restricted to one genus. The lake is generally considered to be much younger than Tanganyika, with a Middle Pleistocene origin considered most reasonable (Brooks, 1950). As is the case in Tanganyika, the lake has fallen below its outlet in modern times; and it is conceivable that more saline phases occurred in the past. However, according to Brooks there are no faunal elements as there are in Tanganyika which require or suggest such a phase.

GROSS STRATIGRAPHY AND CHRONOLOGY

Kivu

The gross stratigraphy of sediment cores from stations 10, 4', 15', and 13' are presented in Fig. 2. These are considered representative of the greater number of cores collected in 1971 and 1972 as they are from different depths and basins. Stations 10 and 4' (Fig. 1) are from the deep northern basin of the lake where depths exceed 400 m. Core 10 has three distinct sedimentary units. From 0-138 cm the sediment is finely laminated with alternating dark brown and white layers. At several loci, particularly from 24-29 cm and 100-138 cm, the white laminae are generally absent and the resulting dark brown material has a gel-like consistency. The brown material is almost entirely organic, while the white laminations are carbonate. The water content of this section is very high, 90-95%. The second unit is also finely laminated with alternating dark grey and yellowish-white layers. The brown material of the upper section is not present. The laminations of this section consist of organic-rich material alternating with diatomites. Several intervals of pure diatomite up to several cm thick are also found. The third stratigraphic unit, volcanic ash, terminates the core at 320 cm. Over a meter of this material was recovered at a nearby station: towards the bottom of the second unit, from 260 to 320 cm, numerous thin intercalations of ash occur. Also, 3 cm of ash occur in the top unit from 75 to 78 cm. The massive ash was not encountered in numerous other 5 m cores from the north basin, suggesting that it is of local occurrence and pinches out rapidly. Capart (from Verbeke, 1957), as a consequence of a detailed bathymetric study of the lake, identified three sublacustrine volcanic cones in the north basin in the vicinity of our station 10. These are the deepest members of a set of volcanic cones along the northern shore of the lake, which erupted underwater, although they are now exposed (DeNaeyer, 1955, 1963). The ash at station 10 is probably associated with the activity of these sublacustrine cones. A subaerial origin with subsequent sedimentation seems to be ruled out by their limited extent.

Core 4', from several kilometers to the southeast of station 10, is an 8 m section. The corer penetrated over 10 m of sediment, and core-shortening (or extrusion past the core catcher) of the section by expanding gases on retrieval probably occurred. Numerous gas spaces between sections of core were present upon opening, but the separations were along bedding planes and the fine structure of the sediment was not disturbed. The section has essentially only one unit. It is laminated as the second unit in core 10, but major diatomite intercalations (>0.5 cm) are more frequent. The first unit of core 10 is absent.

Cores from stations 13' and 15' are significant because their stratigraphy indicates lower water levels in the past. Station 13 was in 86 m of water at the southern end of the main lake. This core had the first and second units of station 10 present, but it terminated in a third unit which extended from 160 cm to the bottom at 421 cm. This unit was dark grey, texturally stiff, and structureless, except for the top 30 cm, which were white because of a high calcium carbonate content. There was also a thin white band at 195. The very bottom of the core had a dry, crumbly texture similar to a soil. Some of this material may have been lost on retrieval. Organic and water content were low throughout this bottom unit. The stratigraphy suggests that the lake stood much lower than its present level during the deposition of this third unit, and it was probably lower than the coring site before then.

than its present level during the deposition of this third unit, and it was probably lower than the coring site before then.

Station 15' was in the Bukavu Basin of Lake Kivu at a depth of 90 m. This basin is separated from the main lake and station 13' by a shallow sill, 35 m deep. The first 150 cm are dark brown to black, very organic and structureless. From 150 to 260 cm the sediment has frequent thin laminae of calcium and magnesium carbonates. Below 260 cm to the bottom of the core the sediment has a light grey appearance and organic material and water content are relatively low. Microscopic examination reveals that aragonite needles are a primary constituent of the sediment. The high carbonate and low organic content suggest a shallow environment and lower water levels.

Several other cores recovered from Kivu contained evidence of lower lake levels. A core from station 9 in a depth of 338 m had the same two upper stratigraphic units as core 10, but it terminated in an unstructured clay-rich facies which extended from 337-357 cm. The lower 20 cm almost bridge the stratigraphic gap between core 10 and 4'. Clays are virtually absent from sediments in depths greater than one hundred meters in the present lake, and sediments with the clay content of station 9 (>60%) are restricted to much shallower depths. Verbeke (1957) reports that the silts and clays derived from the rivers are generally restricted to depths of less than 40 m. Presence of the clay-rich sediments at 338 m suggests a profound drop in water level. Station 1' in 229 meters of water also terminates in a clay-rich, unstructured sediment. The top two units of station 10 are present from 0-180 cm, but from 180 cm to 369 cm clays become much more important. From station 12' in 254 m of water a 257 m core was recovered. From 150 to 190 cm a clay facies was present, and this was underlain by an even coarser, mica-rich bed from 190-250 cm, suggesting that the mouth of a larger creek was very close by at the time of deposition. The bottom 7 cm of this core were finely laminated and diatomaceous, which indicates that a much higher lake level had preceded the shallow water phase.

The most dramatic evidence for a much lower lake level was found at station 14, where a 6 m core was taken in 310 m of water. The top 90 cm of this core contained stratigraphic units correlative with the upper two units of station 10. Below 90 cm the core consisted of poorly sorted, well-rounded pebbles of metamorphic provenance, the largest of which was 5.5 cm in diameter. The lack of graded bedding and the presence of shell fragments suggest that this is a beach deposit. Further support for a former lake stand at more than 300 m below the present level comes from the seismic profiles (Degens et al., 1973). These records show that at depths of less than 300 m, which are found mostly in the southern part of the lake, there is only a thin veneer of sediments overlying the structural basement. These sediment depths contrast sharply with the greater than 0.5 km of sediments found in the deep northern basin.

Radiocarbon dates for the cores from Lake Kivu are listed in Table 3. Sediment from a depth of 170 cm in station 10 is assigned an age of 6,200 years. The bottom of 10 is correlated with the dated bottom of core 9 by microfossil and chemical criteria, and it is assigned an age of 10,000 years. The base of core 9 has an age of 10,700 years. These dates indicate a sedimentation rate of 0.3 mm/yr for sediments between 9 and 170 cm and 0.4 mm/yr for sediments between 170 and 320 cm.

A date of 11,200 years B.P. was obtained from the top of station 4'. Such an age is reasonable as the distinctive brownish upper unit of core 10 is absent, and microfossil and chemical data to be presented later indicate that there is no temporal overlap between cores 10 and 4'. The absence of the upper units is explained either by a depositional hiatus or superpenetration of the corer. As all other cores in the basin recovered the brown unit, superpenetration seems most likely. The other dates from 4' indicate a very high sedimentation rate of 3-4 mm per year, or nearly 10 times the rate for the past 10,000 years at station 10. Although the high rate is surprising, it is similar to the high rate measured on a 1 meter core from 310 m in Lake Malawi (Von Herzen and Vacquier, 1967).

Other samples for dating were chosen in order to estimate the date of the former lowered lake levels and the rapidity of the subsequent rise. The dates from core 13' and core 15' indicate very rapid deposition of the relatively inorganic, shallow water facies at the bottom of the core, i.e., 1.6 mm/yr at station 13' and 5.2 mm/yr at station 15'. The laminated sediments in 15' were deposited at a lower rate of 2.6 mm/yr and the uppermost unlaminated sediments at 0.15 mm/yr, assuming no zero error for the surface sediments. An assumption of atmospheric equilibration for the CO_2 in the modern lake may be unfounded, as there is a sizable contribution of volcanic CO_2 to the alkalinity of the lake, but we have no pertinent data to estimate the possible error. The dates from 13' and 15' seem to show that as much as 1,000 years may have elapsed between the initiation of predominantly biogenic sedimentation over 13' (11,070 years B.P.) and biogenic sedimentation over 15' (9,725 years B.P.).

This implies that the Bukavu basin was separated from the main lake up until approximately 10,000 years B.P. As the Ruzizi exits the lake from the Bukavu basin, Kivu had no surface outlet until after that time. The timing of these rising lake levels and eventual overflow are consistent with dates established on other African lakes (Butzer et al., 1972; Kendall, 1969), and this suggests that the organic carbon incorporated in the sediments was quite near equilibrium with the atmospheric reservoir, assuming that climatic events in East Africa were contemporaneous. If the carbon dates are affected by "old" carbon, then the dates given are maximum estimates.

Tanganyika

The stratigraphy and radiocarbon chronology of the Tanganyika cores have been previously presented (Degens et al., 1971), and only a short review will be given here. A series of gravity cores up to 250 cm in length are available from various depths in the lakes. A core from the north basin, 12, and one from the south, 3, are depicted in Fig. 2. Sedimentation rates are about 0.5 mm/yr at station 12, and 0.4 mm/yr at station 3. Rates which were 10 times lower are found on the sill between the two basins. Both cores have two stratigraphic units. The upper one is dark grey in color and finely laminated, while the lower section is lighter in color and the laminations less regular. Turbidites up to 3 cm thick are present through both sections in both cores, but are more frequent in the southern basin core. These turbidites explain in part the higher sedimentation rates in the deep basin. All the Tanganyika cores contain a much greater admixture of clays, principally kaolinite with some illite and chlorite, than the deep cores from Kivu. A 200 cm core from station 19 over the sill in 550 m of water recovered two additional stratigraphic units not encountered in the deep basin cores. These were blue-grey and grey in color, and montmorillonite dominated the clays. The change in the clay mineralogy combined with the absence of pyrite in these lower units suggested that these sediments were laid down under oxidizing conditions, which implies a hydrography markedly different from the present regime. A date from the bottom of core 19 gave an age of greater than 28,200 years B.P.

Edward

The sediments of Lake Edward varied quite dramatically with depth. At station 5 (see Fig. 2), 120 m, and station 4, 102 m, the sediments were laminated, but the thickness of the light laminae (diatomites) was quite variable; numerous bands, 1-2 cm thick, were encountered. These two cores are strikingly similar visually to that of station 4' in Kivu. At several intervals throughout the cores very fine volcanic ash was disseminated over as much as 10 cm bands. Also discrete bands of coarser ash were present, but these coarser strata were not directly associated with the finer material. These two deep cores were very organic and had a quite pronounced green color, a result of their high pigment content. The three other cores from Edward were from depths of 80 m or less. These showed no trace of the lamination present in the deeper cores. They were of a homogeneous brown color and still very organic. Presumably seasonal mixing of waters to these depths brings sufficient turbulence to destroy the fine lamination present at greater depths below the more permanent deep thermocline. The role of organisms in sediment mixing cannot be ignored, but Verbeke (1957) states that the benthic fauna is restricted to depths of less than 40 m. A radiocarbon date from the deepest core indicates a sedimentation rate of 0.9 mm/yr.

Albert

The sediments of Lake Albert are quite different from those of the other lakes considered (Fig. 2). The core from station 5 at the maximum depth of the lake is a nearly uniform grey sediment, and clays are the most abundant component. There is a slight color change over the length of the core as it becomes darker towards the top. A carbon date from near the base of the core yields a sedimentation rate of 0.7 mm/yr at that site. The core from station 3, water depth 47 m, is more noteworthy. From the top to 185 cm the sediment is dark grey to black in color, and from 185 to 460 cm it is a uniform light grey. The sediment from 460 to 480 is very sandy, and the core terminates in a dry, brittle material at 481 cm. T. Harvey (personal communication) encountered a similar layer at 660 cm in his 12 m core taken from Albert. His carbon dates which bracket the layer indicate that this layer is 13,000 years old. Harvey is studying the diatoms of Albert 3 and has informed us that an age of 7,500 years B.P. could be provisionally assigned to 340 cm by a microfossil correlation with his well dated core. Thus, sedimentation rates at this site, 0.25 mm/yr from 13,000-7,500 years B.P. and 0.45 mm/yr from 7,500 to the present, have been quite variable and much lower than at station 5. Harvey also finds the rates in his core to be labile. The shallowness of this lake and the large fetch of the wind allows mixing to the bottom throughout much of the year, which results in fairly well oxidized sediments. The variable sedimentation rates in space and time may indicate fairly strong bottom currents which combine with the topography of the sites to yield the different rates.

Malawi

A 252 cm core from 553 m of water of station 15 at Malawi was made available by G. Müller of Heidelberg. The sediments appear to be similar to those of Tanganyika in that the sediment is fine and organic, and diatomaceous laminations are often apparent. Mica-rich turbidites are also frequent, especially below 143 cm. No carbon dates are yet available for this core, but the possibility of rather high rates (3 mm/yr) in the lake is demonstrated by Von Herzen and Vacquier (1967).

LAKE KIVU SEDIMENTS

DEEP NORTHERN BASIN (>400 m water depth)

Chemical Stratigraphy

Nineteen trace metals as well as the major chemical constituents of the sediments, Fe_2O_3, SiO_2, Al_2O_3, CaO, MgO, and TiO_2, were determined by uv spectroscopy. Details of the method as well as the raw data are presented in separate reports (Degens and Kulbicki, 1973ab; Baradat, 1970).

The cores from stations 10 and 4' are considered, because of their radiocarbon chronology, to give a nearly complete record of sedimentary events in the northern basin of Kivu through the past 14,000 years. However, as we have constructed the chronology, a gap between

10,000 and 11,000 years B.P. results. Fortunately, about 700 years of this gap are present in nearby core 9. Other evidence presented later supports such an interpretation. The highlights of the chemical stratigraphy are presented in Fig. 3; the distribution of only those elements which exhibit large variation is presented. On the basis of this chemical stratigraphy we recognize five geochemical units in the cores. The lowest from 710 cm to 1,200 cm is characterized by high but variable Al and strongly oscillating Mn. The second unit from 710 cm to 500 cm has declining Al and the Mn fluctuations are reduced in intensity and eventually frequency. From 500 cm to 410 cm, Fe accompanied by Ni is prominent in the sediment. A similar facies is present from 320-275 cm in core 10. From 275 to 130 cm and from 90-35 cm in the latter core the $CaCO_3$ content is high, while from 130 to 90 cm and 35-0 cm $CaCO_3$ is again absent, and organic matter and Pb are greatly elevated.

Spencer et al. (1968) have demonstrated the efficacy of factor analysis for reducing the large, unstructured raw data which sedimentary analyses generally yield, to a simplified matrix which is more amenable to interpretation. Factor analysis with orthogonal rotation (Spencer, 1966) was performed on the elemental data of cores 10 and 4', and five factors were found to explain 82% of the variance. Table 3 presents the varimax factor matrix for each element which was present in greater than just detectable concentrations. These factor loading scores permit, within limits, the identification of the factors. Also, as all the elements are not depicted in Fig. 3, this table will allow insight into the behavior of the other elements.

The high loadings of Al_2O_3, TiO_2, Ga, Cr, Cu, Zn, and V suggest that this is the detrital component consisting primarily of clays (kaolinite and illites) and weathered volcanic ash. The movement of the first three elements with detrital minerals is not surprising, as they are generally residuals of weathering (Cowgill & Hutchinson, 1970) and the movement of the other four elements with this group can be traced to the unusual volcanic rock sequence of the basin. The rocks of the area are thought (Higazy, 1954) to have resulted from the reaction of a carbonatite magma with crustal rock. These distinctive geochemical characteristics imparted by the volcanic process to the lava and ash of the area are brought to the sediment by the weathering and erosion process. The strong association of Cr, Zn, Cu and V with this factor indicates their lesser mobilization by weathering than the other metals.

The high negative loading of Ca, Sr, and Ba on factor 2 suggests that the carbonate phase is involved with this factor. The highest positive loading is with SiO_2. As this factor is certainly not detrital, the SiO_2 of factor 2 is biogenic, being contributed primarily by diatoms. The positive loading of V and Mo on this factor is in line with a biological origin, as both elements are essential micronutrients for algal growth, and they are probably incorporated with the organic matter of the sediments. The negative relationship between the carbonate phase and the biogenic phases suggests a mutual dilution. This is partially true, but it will be shown later that temporal and spatial patterns of carbonate deposition have probably changed fundamentally in the lake. As a general rule, though, high biological productivity has probably been accompanied by low carbonate deposition to the coring sites, and the observed loading on factor 2 is the result. The negative loading of Mn hints at a manganous carbonate phase which, it will be seen, was reasonable for an early period of the lake's history.

Factor 3 has high positive (>0.400) loadings for Fe, Ni, Co, Zn, and Cu. The close association of these elements in lacustrine sediments has previously been noted (MacKereth, 1966; Cowgill and Hutchinson, 1970). Their behavior is largely determined by redox phenomena and the availability of a sulfide phase for precipitation and co-precipitation. Therefore, factor 3 might be characterized as the redox potential of the depositional environment.

Magnesium responds almost entirely to factor 4 with a very high loading. Manganese also has a sizable loading on this factor. Mg in waters of this region (Table 1) remains in solution, while Ca is removed. This is consistent with its known chemical behavior. Experiments with magnesium bicarbonate solutions (Christ and Hostetler, 1970; and Sayles and Fyfe, 1973) seem to show that aqueous Mg^{2+} is very unreactive because of its highly hydrated structure. In fact, increasing Mg^{2+} concentration in itself inhibits crystallization of a magnesium carbonate phase; however, crystallization is catalyzed by increasing ionic strength. Thus, salinity may be a primary determinant of Mg deposition in Kivu, and factor 4 would then reflect the salinity of the lake. The specificity of factor 4 for Mg does not make an alternative hypothesis involving sepiolite or chlorite formation as attractive, because the loading for SiO_2 and Al_2O_3 is very low (although it must be admitted that a low loading does not rule out the possibility of such compounds). The relatively high loading of Mn is more compatible with a carbonate phase. The only appreciable negative loading is that of Mo, which may result from decreased planktonic productivity or at least dilution by the Mg-rich phase. Either interpretation would be consistent with the hypothesized high salinity.

Only one element, Pb, has significant loading on factor 5, and this makes it difficult to establish the identity of the factor. Fig. 3 shows that the Pb concentration is generally low, but has extreme values from 85 cm to 120 cm with lesser pulses above and below that. Organic carbon and pyrite are also high between 85 cm and 120 cm, suggesting that the Pb is removed either with organic matter, perhaps by chelation and incorporation in humic compounds, or as a sulfide. But the upper unit of organic-rich material from 0-30 cm does not show the Pb enrichment, suggesting that Pb becomes available in rather discrete pulses. Cowgill and Hutchinson (1970) found that the sediments they studied were enriched in lead when compared to source materials which suggested continual leaching and precipitation, but they did not observe the pulsing or the very high Pb content (up to 5,800 ppm) which we find. In Kivu sediments Pb is always at least 3 to 4 times enriched relative to its concentration (<10 ppm) in neighboring volcanic rock (Higazy, 1954).

With the insight provided by factor analysis, it is now possible to characterize the hydrogeochemistry of the north basin of Lake Kivu through nearly 14,000 years of its history. Between 13,700 years and 12,500 years, the lake stood at a much lower level which, based on the gross stratigraphy discussed earlier, may have been as much as 300 m below the present level. Al_2O_3 was between 10 and 20% (average = 12%) throughout this period, and detrital components, especially kaolinite and mica, constituted over 50% of the sediment by weight. Although Al_2O_3 is abundant, its value fluctuates quite markedly, reflecting changing water levels and distances from source rivers and streams to the coring site. Lake Kivu during this period was a closed basin lake; and astatic water levels, reflecting short-term balances of precipitation and evaporation (Langbein, 1961), would be expected.

The waters of a closed Lake Kivu would have been saline and alkaline; and, except for their somewhat higher K and Mg derived from distinctive source rocks, they would have strongly resembled the sodium bicarbonate-carbonate lakes of the Eastern Rift Valley. In these alkaline waters Ca precipitates peripherally as it enters the lakes, and Ca deposition in offshore areas is low (Kilham and Hecky, in press). This explains the negligible Ca content of core 4'. Na, K, and Mg would accumulate and alkalinity increase with time, but a fluctuating water level would allow salt wastage by deflation when the lake was down (Langbein, 1961). Lake Kivu would differ from the Eastern Rift Valley lakes in one significant aspect: water depth. The closed basin lakes of East Africa, with the exception of Rudolf, are shallow or even ephemeral. Lake Kivu, even when it stood 300 m below its present level, had 100 m of water over the 4' coring site. The shallow lakes are usually mixed (oxygenated) to the sediment surface diurnally, but the greater water depth of the Late Pleistocene Kivu suggests that prolonged stratification might have occurred as it does in the deep, modern lakes.

The oscillating Mn reflects the stratification history. Prolonged stratification leads to deoxygenation of deep waters in productive lakes. After deoxygenation, redox potentials in deep waters and sediments decline; and, as they do, many elements become reduced and the solubility of their compounds may change. Mn and Fe as well as many other transition metals are chiefly affected. Mn will be reduced first from its sedimentary oxide or silicate and will enter deep waters. If the redox potential continues to drop, Fe^{2+} can become available. In the alkaline, saline Kivu, the concentration of these two reduced species would probably increase until solubility with respect to the carbonate phase was exceeded. Thus their behavior would also be pH dependent. Surface waters would have higher pH values than deep waters and, therefore, more $CO_3^=$ available to remove Mn or Fe.

Kivu during this period was a closed system, so that simply changing the Mn from an oxide to carbonate would not produce the oscillations. Two scenarios are offered which could bring about the required redistribution. One is largely temporal and one spatial, and they are not mutually exclusive. In the first case stratification occurs; Mn^{2+} diffuses from the sediment and accumulates in relatively low pH deep waters. Subsequent mixing because of weakening or obliteration of the stratification would deposit $MnCO_3$, as the slow kinetics of Mn^{2+} oxidation should favor precipitation as a carbonate (Garrels and MacKenzie, 1971). The second case involves Mn^{2+} diffusing from sediments, mixing across the thermocline and precipitation. Precipitation in deep water would lead to resolution because of the lower pH, while precipitation in shallow water permits incorporation in the sediments. In this case deep water sediments become impoverished in Mn while shallower sediments are enriched. Fluctuations in Mn content at the 4' site in this case would record the position of the thermocline. When Mn is high, the 4' site was above the thermocline; and, when Mn is low, it was below the thermocline. Thermal stratification seems most likely, as a salinity stratification (meromixis) would be much more stable and longer lived, in which case Fe^{2+} and sulfide production should have occurred. In fact Fe is rather constant throughout this period, showing enrichment only 3 times, at 13,500, 13,200, and 12,500 years B.P. The first of these is represented in the core by a bed, almost 1 cm thick, of nearly pure manganosiderite; the chemical composition of this material is not known, since our routine 5 cm spacing between samples missed this bed.

If the depth of mixing of the epilimnion during this low water period is considered to be more or less constant over several years (our sampling interval is on the order of 10 years because of the high sedimentation rate), then these fluctuations of Mn can be related to water level. High water levels would place the thermocline well above the 4' site and Mn-depleted sediment would accumulate. Lower water levels would bring Mn-rich sediment to the site. This interpretation is presented in Fig. 4, which shows how lake levels might be reconstructed. Prior to 13,500 years, when the lake was shallowest as suggested by the Al_2O_3 content of deposited sediment, there are no significant oscillations, which implies that the lake may have been completely mixed throughout with no significant stratification. Subsequently the lake level was somewhat higher, stratification was more persistent, and redistribution of Mn occurred. The redox potential, however, only infrequently dropped to values sufficient to liberate Fe^{2+}.

In the discussion of the factor analysis a rather high positive loading of Mn on factor 4, the Mg factor, was noted. Such a covariance could be generated by fluctuating lake levels in a closed, alkaline, saline lake. Periods of Mn deposition at 4' imply falling lake levels and higher salinities. Increasing ionic strength and alkalinity should favor precipitation of magnesium carbonates, although never on a massive scale. By the models presented above, the correlation would not have to be particularly high, i.e., a slight shift in the position of the thermocline could alter Mn deposition without involving a volume change sufficient to modify salinity significantly. It is of note the the manganosiderite contains on average between 50 and 80% $MnCO_3$ as ascertained by x-ray diffraction analyses.

Beginning at 12,500 years B.P., a period of generally rising lake levels ensued. The Al_2O_3 and Mn contents of the deposited sediment decrease and the iron content increases. High Al_2O_3 values are accompanied by TiO_2 values which are relatively even higher. Thin laminae of volcanic ash are present, indicating that increases in Al_2O_3 are related to volcanism rather than changes in lake level. No extreme fluctuations in Mn occur, and the increases observed correspond to the TiO_2 peaks. Therefore, this higher Mn results from greater inputs of unweathered detrital material. Iron shows an appreciable increase in sediments deposited between 12,000 and 11,700 years B.P., suggesting more strongly reducing conditions and pyrite formation. The availability of sulfide in the deep waters will trap Fe and other transitional metals, while the more soluble Mn will continue to migrate to shallow environments. A higher lake level and a more stable stratification presumably engendered the more reducing conditions. The fossil diatoms in these strata, to be discussed later, also suggest a stable thermocline or possibly a weak halocline. Between 11,700 years and 11,500, Fe deposition decreased slightly in relation to other components, and water levels continued to rise. This implies a less stable stratification.

After 11,500 years B.P. pyrite deposition was relatively greater as detrital and biogenic sedimentation decreased. High lake levels, but not sufficient to open the lake to the south, and a stable stratification are inferred. The core from station 10 begins with Fe and Ni similarly high; however, the bottom of this core is assigned an age of 10,000 years B.P., which leaves a gap of 1,000 years in the north basin record. The geochemistry of sediments deposited at station 9 in the critical time interval between 10,000 and 11,000 years B.P. resembles closely those formed at the base of core 4' (high Al and low Fe). A lower lake level stand is inferred.

The beginning of $CaCO_3$ deposition in offshore waters is effected when the lake rose above the Ruzizi outlet, which stood at a higher elevation than now. As the lake overflowed, the HCO_3^{1-}-CO_3^{2-} alkalinity decreased. Ca was no longer removed peripherally as it entered, but it was mixed into the open lake, where high pH values because of photosynthetic removal of CO_2 caused its precipitation. The decrease in iron and nickel results from dilution by the $CaCO_3$. The lake had open drainage and a hydrography similar to Edward and Tanganyika, i.e., a deep, perennial thermocline with no salinity stratification up to 5,000 years B.P., when catastrophic events occurred in the lake's basin.

From 5,500 to 5,000 years B.P., TiO_2 and Al_2O_3 deposition rose sharply, which indicates increasing volcanic activity in the basin and sedimentation of ash. At 5,000 years B.P., $CaCO_3$ deposition ceases while organic carbon and Pb increase. The organic carbon emerges as the contribution of other sedimentary constituents, especially $CaCO_3$ and diatoms, is reduced. The higher organic carbon does not mean higher productivity. Rather, planktonic production dropped (to be discussed with the biostratigraphy), and the contribution of terrestrial organic matter was relatively increased. In contrast, the Pb increase by a factor of nearly 100 suggests a large input at that time.

$CaCO_3$ is not being deposited at the station 10 site today, either. The lake is presently meromictic, and the deep waters charged with CO_2 (Degens et al., 1973). No pH measurements on the deepest water have been accomplished, but the trend is noticeably acidic with depth. Surface waters have pH values between 9 and 10, but the pH drops below 7 as the waters become anoxic. Therefore $CaCO_3$ particles falling into the deep water redissolve, and incorporation of $CaCO_3$ into the sediment is only possible in depths above the first halocline. Many areas of Kivu's shoreline today exhibit a nearly continuous pavement of $CaCO_3$ (Verbeke, 1957); and cores from shallower regions have carbonate-rich surface sediments. The absence of carbonate from 130 cm to 90 cm implies meromixis similar to that existing today. It does not mean a lower lake level and extremely alkaline surface waters, as the carbonate-free sediments of the core from 400 cm to 1,200 cm, because the Al_2O_3 content is low.

The warm, saline, deep waters of the present lake are attributed to hydrothermal springs (Degens et al., 1973), and this spring activity became significantly stronger 5,000 years B.P. High Pb values for the sediments are a good marker of spring activity, and peaks prior to 5,000 years indicate that the springs were present but on a reduced scale.

The Pb derives from intense leaching of the metal-rich volcanic lavas and ash by warm, ground waters charged with volcanic CO_2. This leaching should enrich the lake waters in all the metals, but the geochemical fate of the metals and their sedimentation differ markedly. Factor analysis showed that the heavy metals are split among three factors. The first factor contained those metals which were associated with the detrital phase, and it contained metals which were either more resistant to leaching or, if leached, were not incorporated into the lake sediments. Zn, Cu, and Cr had their highest loadings on this factor. Fe, Ni, and Co had their highest loadings on factor 3 while Zn and Cu also had significant loadings. This represents part of a widely recognized covariant suite of metals (Spencer et al., 1968) which migrates together under oxidizing and reducing conditions. In Kivu the oxides and heavy metal carbonates predominate prior to 11,500 years B.P., while the sulfides dominate after that time. Pb exists virtually alone in the last factor, indicating a quite different behavior from the other metals.

With the increased hydrothermal activity, the input of all the metals increased. Pb was the first to be removed and added to the sediment. The removal was effected by sequestration with organic material and subsequent sedimentation. Degens et al. (1972) describe a similar process for Zn in Lake Kivu, but with a different result. They identified resinous spheres ($\sim$1 micron in diameter) which chelated Zn and precipitated sulfide to form sphalerite crystallites (5 to 50 Å in size) which covered the sphere. These spheres were found in the water column, and failure to find these spheres in the sediment suggested that they were buoyant and were being lost to the lake. Such is not the case with the chelated Pb, however, and it is rapidly deposited after entry to the lake. Cowgill and Hutchinson (1970) found that Zn and Pb behaved very similarly during weathering and sedimentation in a basin which has a petrologic provenance similar to Kivu, i.e., abundant potassic alkalic lavas and ash. Both Zn and Pb were greatly enriched in sediments relative to soils and rock; no difference in mobilization between Zn and Pb during weathering was observed. The high loading of Zn on the detrital factor in Kivu indicates that it is being lost from the basin by the above mechanism rather than being totally deposited in the lake. Pb is being sedimented, but it is being removed in a manner similar to that of Zn.

The absence of Fe and other metals frequently associated with Zn from the sphalerite spheres suggested to Degens et al. (1972) that the organic resins were highly specific, excluding those metals which could not, for structural reasons, coordinate properly. These metals for which no organic chelator was present would have to be removed in deep waters as sulfides or in shallow waters as oxides. Precipitation of these metals in deep waters would depend on the production of sulfide by sulfate-reducing bacteria. These organisms use sulfate as a hydrogen acceptor in metabolic oxidation. They are dependent on organic matter as a source of free energy; and, thus, sulfide production in deep water will be tied to organic production in surface waters. Evidence to be presented later demonstrates that plankton production sharply fell during these hydrothermal pulses; this in turn should also have affected microbial activity at depth. The magnitude of this decrease can be estimated in Fig. 3. At 9,000 years B.P., $CaCO_3$ rose from nearly 0 to 30% of the sediment by dry weight. This necessarily decreased all the other major components by approximately one-third to maintain a zero sum. Iron did decrease by this amount, i.e., from 15% to 10%. At 5,000 years B.P., $CaCO_3$ disappeared from the sediments, but there was no response by the Fe. Diatoms also decreased at this time, and the only relative increase in a major component is registered by organic carbon. As Fe did not rise by 5% at 5,000 years B.P., at least a one-third reduction in sulfide production had occurred. Sulfide production did not increase to meet the greater supply of metals, and these metals were lost to the deep waters as they migrated to shallow oxygenated sediments or out the Ruzizi River, unless they could coordinate to the organic matter.

Approximately 4,000 years B.P., $CaCO_3$ was again being deposited in the deep north basin. The deep waters were no longer acid because mixing processes were more efficient. The thermohaline stratification which was present from 5,000-4,000 years B.P. and which occurs today had been destroyed. The density changes involved in the Kivu system are small, and a combination of cooling or concentrating surface waters and/or heating or diluting the deep waters could lead to a breakdown. Assuming heuristically that the relative contribution of heat and salt to the hydrothermal waters is relatively constant, i.e., hotter waters leach sufficient salts to offset thermal density decreases, then the maintenance of crenogenic meromixis (Hutchinson, 1957) is dependent on two processes: (i) a sublacustrine source of denser water must exist, and (ii) surface waters must be constantly diluted by runoff and loss at the effluent. If the lake became closed, surface waters would become more saline; and, as coolness and aridity are correlated in the African climate, surface cooling might be expected also. Such a state of events would lead to a breakdown of the convecting water cells as density differences are annihilated. The depth of the well-mixed layer (near atmospheric equilibrium) would grow from above in a gradual cascade of collapsing cells. If the process extended over many years, the mixing of the immense reserves of hydrothermal gases and heavy metals into the surface waters need not be catastrophic. Evidence from other cores and lakes will be presented below, which indicates that Kivu was without a surface outlet during this interval of $CaCO_3$ deposition. Despite the fact that the lake was closed and saline as it was prior to 9,000 years B.P., Ca was deposited in the deep north basin because the source of Ca was the sublacustrine springs, not surface runoff as was the case previous to the onset of spring activity. The occurrence of high Pb concentration within the period of $CaCO_3$ deposition means that the hydrothermal springs simply changed in water chemistry.

The above results would contrast sharply with what would happen if mixing was propagated from below by the injection of less saline and/or hotter water. These circumstances would lead to convecting cells growing from below and not in atmospheric equilibrium. In this case toxic substances of the deep waters could be introduced rapidly into the surface waters with catastrophic consequences for the biota. There is no evidence in the stratigraphic record for a situation like this, but it is questionable whether such short-term events could be resolved in the fossil record. What evidence there is for conditions deleterious to the biota coincides with the onset of intense hydrothermal activity and stratification rather than the termination of the thermohaline structure. This evidence will be discussed later.

Events similar to those that happened at 5,000 years B.P. occurred again at 1,200 years B.P. High TiO_2, Al_2O_3, and Mn content from volcanic ash are found in sediments immediately underlying Pb-rich, $CaCO_3$-depleted sediments. The Fe and Ni subsequently increase more rapidly, suggesting that the productivity of surface waters and hence sulfide production was not as drastically affected by the second event. The strong similarity of the two events, i.e., ash followed by evidence of increased hydrothermal spring activity in both cases, seems to imply a causative relationship between volcanism and spring activity. As the water of these springs has been shown to be meteoric (Degens et al., 1973), climatic processes may be involved also, namely climatically determined availability of water for deep percolation through the porous lavas and ash of the Virunga system might regulate the volcanic activity. The water, by mixing with hot magma, would be transformed to steam, and the consequent volume changes could fracture or dislodge confining overburden and allow the release of volcanic ejecta and lava. The latest lavas from the Virunga system are undersaturated with water (Holmes, 1965, p. 1024); and, therefore, water may be especially critical to this system.

The chemical stratigraphy of the deep, northern basin of Kivu supplies the following information about the lake history:

(i) Prior to 12,500 years Lake Kivu stood as much as 300 m below its present level; the lake necessarily was closed and more saline and alkaline;

(ii) Station 4' site was under at least one hundred meters of water; and, judging from the fluctuating Mn deposition at 4' and the depth of the perennial thermoclines in the modern lakes, it may have marked the depth of the thermocline through much of that time;

(iii) at 12,500 years B.P. water levels began to rise; the thermocline moved well above the 4' site and sharpened; the deep waters and sediment became more reducing and pyrite formation increased; there was some volcanic activity in the area around 12,000 years B.P.;

(iv) the lake level dropped sharply around 10,700 years B.P. but not quite to the low level it occupied prior to 12,500 years B.P.;

(v) the lake achieved the elevation of the Ruzizi outlet 9,500 years B.P. and began to freshen;

(vi) at 5,000 years B.P. hydrothermal spring activity increased dramatically following or coinciding with increased volcanic activity; a complex hydrography similar to that extant in the lake today was established;

(vii) the thermohaline density structure was subsequently obliterated, probably by the lake level falling below the Ruzizi outlet; and

(viii) a wetter climate reopened the lake and another pulse of hydrothermal and volcanic activity 1,200 years B.P. reestablished the thermohaline structure which has persisted up to the present.

Biostratigraphy

Siliceous microfossils. Characterization of the biostratigraphy of the north basin cores, 10 and 4', is based on siliceous microfossil analyses for both cores and synoptic determination of amino acids and hexosamines for core 10. Diatom frustules are by far the most abundant microfossils in cores from all of the Rift Valley lakes. States of preservation vary from lake to lake and within lakes, but in the north basin of Kivu preservation is excellent (Fig. 3). Very few species attain relative abundances greater than 5%, as might be expected in off-shore planktonic populations in large, deep tropical lakes; as hydrographic variability is greatly reduced relative to more temperate lakes. Analyses of the diatom stratigraphy of other Kivu cores and the other lakes is still in progress. For the purposes of this paper, only a brief outline of the information derived from diatom analyses will be necessary. A more detailed presentation of the diatom stratigraphy will be given when all the work on Kivu is completed.

From the bottom of core 4' to 130 cm in core 10, only 3 species comprise well over 90% of the flora at any sample level. These three are *Stephanodiscus astraea* v. *minutula* (Kütz.) Grun., *Nitzschia fonticola* Grun., and *Nitzschia spiculum* Hustedt. Among the other diatoms only *Nitzschia acicularis* W. Smith achieves abundances over 5% for short periods of time. Transmission electronmicrographs of carbon replicas are presented in Figs. 5-7. Comparative availability of nutrients and, indirectly, stratification of the water column appear to control the relative productivity of these species. Understanding of their ecology has been achieved from: (1) the distribution of the species in the modern lakes, (2) simple models of the three-dimensional distribution of nutrients in lakes, (3) relationships between species productivity and sinking rates, and (4) the behavior of the species within the present core. An elaboration of the argument will be presented elsewhere (Hecky and Kilham, in prep.), and only the general conclusions are given here in order to interpret the distribution of the diatom species through time in the north basin.

Stephanodiscus astraea is a low silica specialist among planktonic diatom species. Kilham (1971ab) states that it only becomes prominent in the plankton of lakes as dissolved silica concentrations fall below 1 mg/l. In Lake Albert, which has very low silica and high phosphorus, *S. astraea* dominates the microfossil assemblages from surficial sediments. In Lake Edward, where dissolved silica in surface waters is much higher, *N. fonticola* predominates. Surficial sediments of Lake Tanganyika and modern Lake Kivu contain long, thin (length to width ratio > 25:1) *Nitzschia* species to the virtual exclusion of other species. *N. spiculum* (Figs. 5 and 6) is quite representative of this morphology. All nutrients are extremely low in offshore Tanganyika waters, while dissolved phosphorus is nearly vanishing in Kivu surface waters (Table 1). These long, thin *Nitzschia* appear to be able to sustain production under nutrient conditions which are unsuitable for other diatom species. The low nutrient supply in Kivu and Tanganyika has a marked effect on photoplankton productivity, which was remarked on earlier and can be seen in Table 2. The long, thin morphology results in slow sinking rates and high surface area to volume ratios. These attributes reduce the probability of sinking out of the nutrient-depleted euphotic zone before reproduction can be achieved, and maximize the surface area available for nutrient uptake. Contrary to the situation in Edward and Albert, diatoms are not the most prominent constituent of the phytoplankton in Kivu and Tanganyika today. Rather, blue-green algae (esp. *Lyngbya* sp. in Kivu) and other algal groups with slower sinking rates or self-motility dominate. This is especially clear in Kivu, which has been most extensively studied. Biological processes reduce phosphorus to a negligible level, yet dissolved silica is virtually unaffected in the north basin surface waters, as Si gives a linear regression on Na through all depths. Diatoms must account for only a small proportion of the primary productivity in the north basin today, and the diatom production accomplished is due solely to the *Nitzschia* spp. Conversely, when *S. astraea* is prominent, diatoms are accounting for a much greater proportion of primary production as dissolved Si is driven very low.

In tropical lakes two quantitatively significant nutrient sources for surface waters can be identified. One is surface runoff from the drainage basin, and the other is regeneration of nutrients from the water column and sediments. The salt contribution of tributaries is primarily derived from the weathering of rocks except in arid regions where atmospheric fallout becomes important (Hecky and Kilham, 1973). Silicate rocks are exclusively present in the Kivu basin, and weathering will yield dissolved SiO_2 and HCO_3^{-1} as the most prominent constituents. Runoff from the basin can be characterized as rich in dissolved SiO_2. In contrast, nutrients derived from regeneration will be poor in silica. Phosphorus and nitrogen are efficiently regenerated from plankton detritus, yielding the very low turnover times for these elements which are often measured in lakes (Rigler, 1964). Silica, however, is only slowly released, and judging from the vast accumulations of diatomites in these lakes, regeneration of SiO_2 can probably be considered negligible on an ecological time scale. Thus water masses receiving their nutrient content predominately from regeneration processes will be comparatively rich in P and N but poor in Si.

The relative contribution of these two types of water, Si-rich runoff and Si-poor deep water, will be determined primarily by lake size and stratification. Generally speaking, as lake area and volume increase, replacement times increase and the nutrient contribution of surface runoff to the lake decreases proportionally. Where tributaries enter the lakes, production can be intense, and this tends to remove peripherally the nutrients from weathering. Therefore, the productivity of large tropical lakes, as that of the ocean, depends to a large extent on regeneration processes; and is relatively independent of tributary inputs over significant periods of time. In such lakes, diatoms with silica uptake kinetics (Guillard et al., 1973) attuned to low silica concentrations should predominate. The availability of regenerated nutrients to the euphotic surface waters depends on mixing processes, as much of the regeneration occurs in deep water. A continuum from molecular diffusion across very stable interfaces to advection in upwelling areas is possible in aquatic systems, but in all the lakes under consideration only eddy diffusion across more or less stable thermoclines and chemoclines is significant for most of their area. As a first order approximation, mixing will be inversely proportional to the density differential across a thermocline or halocline. This difference in conjunction with meteorological stresses, i.e., seasonal variation in temperature, winds, rainfall, etc., will determine the depth range or mobility of the thermocline. This depth range, in turn, gives an indication of the amount of deep, nutrient-rich water which will be mixed into the nutrient-poor surface waters. Lake Albert, which is mixed completely most of the year, and Lake Edward, which is likewise thoroughly circulated except for a small area constituting less than 10% of its total surface area, should exhibit a high productivity, and they do. Productivity should also be reasonably high in Malawi, as its deep thermocline is fairly mobile (Table 2). The deep thermocline of Tanganyika and the thermohaloclines of Kivu have been nearly stationary over the periods of observation, and thus mixing and productivity are low. Formation and obliteration of the shallow seasonal thermocline will engender fluctuations around a low mean value determined by mixing across the perennial thermocline.

With the above considerations in mind, the changes in diatom abundances for the period from 13,700 years B.P. to 5,000 years B.P. can be interpreted. At 13,700 years B.P., *N. fonticola* is prominent. As the Al_2O_3 content is near its highest value in the core and low lake levels are indicated, the presence of *N. fonticola* at this time indicates that the waters are relatively rich in Si, and a nearby tributary or at least near-shore conditions are implied. *N. fonticola* subsequently becomes relatively less abundant, and *S. astraea* increases. This indicates somewhat higher water levels and reduced influence of land-derived nutrients. The mobile thermocline which the fluctuating Mn reflects, permits relatively efficient regeneration of low Si nutrient-rich water. Around 13,000 years B.P., *N. fonticola* and *N. spiculum* increased. This suggests rising water levels and more stable stratification as the relative contribution of deep water to the phytoplankton is reduced. The low Al_2O_3 and Mn agree with this interpretation. These elements subsequently rise, and *S. astraea* returns to dominance, reflecting lower lake levels and better circulation.

Again at 12,000 years B.P., *N. spiculum* became prominent, implying rather stable stratification. Geochemically the stabler stratification made the deep water and sediments more reducing, and Fe-deposition (as pyrite) and organic carbon deposition increased relative to the other sedimentary constituents. The association of these conditions with the evidence for volcanism (high TiO_2) during this period could indicate sublacustrine volcanic activity and the introduction of saline spring water at depth. However, Pb is low, and equally plausible is a climatic causation. The climate was becoming warmer and moister through this time, and a closed, saline lake was being freshened. Prolonged intervals of warm, wet years might favor the establishment of a relatively warm, dilute surface layer of water overlying cooler, saltier water. The resulting sharp, stable thermocline would reduce mixing, and *N. spiculum* would be favored. These conditions endured for nearly two hundred years before *N. spiculum* declined as stratification became less stable because of climatic change or reduced volcanic activity.

At 11,500 years B.P., *N. fonticola* increased relative to *S. astraea*. High water levels and a fairly stable thermocline are suggested. Increased pyrite deposition implies a strongly reducing hypolimnion. A deep thermocline and high runoff is similarly implied as the lake opened and became diluted at 9,000 years. From 9,000 years to 5,000 years B.P., *S. astraea* declines in abundance, to be replaced by *N. fonticola* and subsequently *N. spiculum*. This reflects a general warming trend in the climate (recognized world-wide), and the thermocline is becoming more stable as the temperature (and density) differential of surface and deep waters increases. Reduced mixing and lower productivity result. It must be emphasized that shifts between *S. astraea* and *N. fonticola* reflect only the availability of Si, and do not necessarily imply any changes in the total production of the phytoplankton or even of the diatoms. Gross productivity will be set by the supply of P and N, which will be utilized by non-siliceous algae as well as diatoms. A relative increase in *N. spiculum*, however, does imply, under the model presented, that overall phytoplankton production has decreased as it reflects low mixing rates.

Hydrothermal activity established a thermohaline density structure approximately 5,000 years B.P., and the consequences for diatoms were rapid and extreme. At a sediment depth of 130 cm in core 10, there is a sharp contact both in lithology and biofacies. In the space of 5 mm, the sediment changes from a carbonate-rich, finely laminated sediment, dark greenish grey in color, to a thin layer of fine volcanic ash and diatomite, yellowish to olive in color, and finally to an organic, unlaminated material, dark brown in color, with a gel-like consistency. The diatoms change from *Nitzschia spiculum* in the first unit to purely *Stephanodiscus astraea* v. *minutula* in the second, to relatively short and wide species of *Nitzschia* such as *N. palea* (Kütz.) W. Smith and *N. accomodata* Hustedt (Fig. 6). Even more striking are the changes in absolute abundances, as diatoms are scarce in the brown unit, and minute, siliceous scales of the chrysophyte, *Paraphysomonas vestita* (Stokes) de Saedeleer (identified by R. W. Norris) become the most abundant microfossil. This assemblage of short, wide *Nitzschia* and *Paraphysomonas* scales recurs within each brown layer (Fig. 6), but the elimination of the long, thin species which are abundant throughout the past 5,000 years is not nearly as extreme as it was in the first event. Among non-siliceous microfossils, species of the green algal genera *Tetraedron* and *Cosmarium* are prominent in these brown layers, but incidence of identifiable algal remains is quite low compared to the other sedimentary units.

The brown units are interpreted as periods of greatly reduced phytoplankton productivity. Planktonic diatoms were virtually eliminated. Many of the species present, e.g., *N. palea*, have a demonstrated ability for an epiphytic planktonic existence (Gessner, 1956) and using organic substrates as nutrient and/or energy sources. Hustedt (1949) found these species occupying benthic habitats in Kivu. *Paraphysomonas vestita* is a motile, colorless phagotroph which ingests bacteria and small algae (Manton and Leedale, 1961). The emergence of these species as the most prominent microfossils in the sediment may simply be due to the elimination of the more numerous producers, i.e., these elements are ubiquitous throughout the core, but at abundances so low that the standard enumeration techniques (counting 400 diatoms at each level) would not reveal them; but the restriction of this element to the *Nitzschia* spp. and the *Paraphysomonas* suggests that they are remnants of a distinctive phytoplankton which flourished at the time.

By the models discussed earlier, the establishment of an extremely stable density interface could lead to greatly reduced mixing and extremely low productivity. As the chemical stratigraphy implies that the hydrothermal activity stratified the lake sufficiently to trap CO_2 and dissolve calcium carbonate at that time, such low productivity is expected. The rapid onset of this activity displaced the pre-existing deep water upwards and led to a constant advection of water low in Si but rich in N and P into the surface waters. Production was intense and the layer of *Stephanodiscus astraea* was deposited. As the warm, saline water filled the deep basin and rose near the surface, a sharp density difference would be established between the cool, dilute surface water from runoff and the hydrothermal water. The exact depth of this interface would be determined by the relative inputs of the two water sources. Initially a stable two-layer system may have been present with very low productivity the result. Diatoms, because of their sinking habit, would be selected again in the low nutrient environment. Only those diatom species capable of an epiphytic existence would survive by attaching to larger, more buoyant algae such as blue-greens, flagellates, and green algae. The *Nitzschia* spp. would also be favored because of their ability to utilize organic resources, as a great proportion of the nutrients in the system will be tied up in dissolved organic material; we have found up to 50 mg C per liter. The small *P. vestita* (10-20 μm) would be present, feeding perhaps on microalgae, e.g., small flagellates. This species would be unable to ingest the larger diatoms which dominated the phytoplankton previously and subsequently. The *Tetraedron* and *Cosmarium* would not be inconsistent with the proposed oligotrophic conditions.

From the initial two-layer system, warm salty water under cool, dilute water, a multi-layer system proliferated because of the heating from below. This heating may have been significantly augmented because of biogenic heating by fermentative bacteria (Deuser et al., 1973). Proliferation of multiple layers reduced density differences between layers, and greater mixing resulted. As the hydrothermal water mixed into the surface waters, the salinity and productivity of the surface waters increased, and euplanktonic diatoms reappeared. Under the low nutrient conditions, the long, thin *Nitzschia* were expected (Fig. 7). *N. spiculum* was joined by *N. bacata* and *N. mediocris* as new elements in the diatom plankton. The ecological relationships of these three is unclear, as little is known about any of them, but they all do exhibit the long, thin morphology. Proliferation of *Nitzschia* spp. of this morphology probably represents occupation of planktonic niches left vacant by the exclusion of all other diatom species. The available record is insufficient to determine whether this is a result of immigration of new species or evolutionary diversification within the lake. *N. mediocris* and *N. bacata* have their greatest relative abundances in the core where $CaCO_3$ is low or absent, which should indicate that stratification was most stable and of a thermohaline character. *N. spiculum* has a maximum for the past 5,000 years at approximately 2,000 years B.P., while $CaCO_3$ was being deposited and perhaps only a deep thermocline was present.

Another new element in the diatom plankton was a *Chaetoceros* sp. (Fig. 7). This genus is rarely recorded from inland waters, as it is typically marine. All previous reports of significant abundance in continental waters have been from lakes much more saline than the present Lake Kivu, where we have found it in surficial sediments (cf. Anderson, 1958). Its abundance through the past 5,000 years in Kivu peaked approximately 4,000 years B.P. Its presence is probably related to somewhat higher salinities combined with low nutrient conditions. Its persistence in the modern lake may reflect adaptation by a strain to the more dilute conditions as salinity changed slowly over time. It exhibits a very strong negative interaction with the stubby *Nitzschia* spp. of the brown layers, indicating that it, as all the planktonic species of diatoms, is virtually eliminated during the ultra-oligotrophic periods of extremely stable stratification.

Although the lake has experienced notable changes in salinity because of climatic and volcanic related changes in hydrography, the diatom flora has responded primarily to varying nutrient regimes. Only the *Chaetoceros* sp. can be interpreted in terms of salinity alone. The diatom stratigraphy is entirely consistent with the chemical stratigraphy in reconstructing the paleolimnology of the lake. Together they provide powerful paleoecological tools, as they yield evidence on quite different portions of the lacustrine environment. The chemical stratigraphy reflects in part conditions in the surface waters, but the imprint of surface water conditions can and will be strongly modified by conditions in the deep water and burial (witness the dissolution of $CaCO_3$ in the deep waters of the modern lake). In contrast, diatoms yield information primarily on surface waters, and respond principally to changes in nutrient chemistry which are caused by lake level fluctuations and stratification. In addition, diatoms can yield information on relative biological productivity by species shifts, as seen above. Only by integration of the chemical and biological evidence can full understanding of the paleoenvironment be achieved.

Organic matter. Organic carbon is rather constant below 130 cm in core 10 and throughout core 4' (Fig. 3). Values range from 1.5% to 12% C on a dry weight basis, with an average around 7.5%. Above 130 cm in core 10, organic carbon rises sharply to 16% and remains over 12% through the upper section except for the interval between 90 cm and 30 cm. Changes in sedimentary organic carbon in the north basin are not simply interpretable in terms of productivity. Much of the pattern can be accounted for by the varying contributions of the other principal sedimentary constituents, i.e., diatoms, $CaCO_3$, and detrital minerals. For example, at 130 cm, organic carbon increases from 5% to 15%, but inspection of Fig. 3 shows that the detrital minerals represented by Al_2O_3, TiO_2, $CaCO_3$ and Fe_2O_3, the other major components, all decrease or vanish. As discussed above, planktonic diatoms became scarce, and total phytoplankton production probably also fell. The deep waters became more strongly reducing and more conducive for preservation of organic material as a salinity stratification was established. These events taken together can easily account for the dramatic rise in organic carbon without necessitating an increase in productivity. In fact, characterization of the organic carbon yields information which also suggests that planktonic productivity fell.

Mopper and Degens (1972) characterized the sugars in the brown layers and compared them with sugars from other strata. They found the brown layers to be greatly enriched in glucose relative to other sediments. This could be interpreted as reflecting a terrigenous source which would contribute large amounts of cellulose. However, galactose was too high to attribute all the material to an allochthonous origin. The enrichment in glucose is compatible with the hypothesized low planktonic contribution to this layer. In this case the land-derived organic component remained constant, and phytoplankton production fell, which created a relative enrichment in glucose from terrigenous cellulose. It is also possible that the unusual phytoplankton present during deposition of these brown layers had a distinctive pattern of sugars, but brown organic sediments are usually associated with relatively high allochthonous contributions

(Hutchinson, 1957, p. 882). This likewise is consistent with a low phytoplankton production. Earlier it was noted that sulfate-reduction was probably less during deposition of these sediments; otherwise, the Fe and other heavy metals from the hydrothermal activity would have been removed in the deep water. Reducing sediments are characteristically black because of FeS precipitation. The very fact that these sediments are brown reflects the lowered activity of the sulfate-reducing bacteria. As a substrate, terrigenous organic matter which had already been subjected to bacterial degradation in soils would not be suitable for the sulfate-reducing bacteria as plankton detritus.

In conclusion, all the evidence is compatible with very low phytoplankton productivity during periods of intense hydrothermal activity. Separation of toxic effects from nutritional ones is not possible from the stratigraphic record, but it is clear that the stronger density stratification imposed by spring activity does reduce phytoplankton production even when spring activity and the input of toxic metals is reduced.

Amino acid and hexosamine concentrations are given for core 10 in Fig. 8 on a carbon basis to eliminate the variability created by inorganic sediment constituents. The individual amino acids exhibited no recognizable trends through the cores, but total amino acids demonstrated a rising concentration initiated at 260 cm (8,500 years B.P.) and continuing to the surface of the core. This rise may reflect in part the changing diagenetic conditions in the deep water. The climate is warming and the thermocline becoming sharper, which leads to stronger reducing conditions in the hypolimnion. Better preservation of organic material would be expected. Hexosamines increase even more sharply at 260 cm, but they subsequently level off and then decrease sharply at 130 cm and continue to fall towards the surface. Zooplankton, insects and bacteria contribute hexosamines to lacustrine sediments, but we concur with Kemp and Mudrochova (1972) that chitin will be more abundant and more resistant to breakdown than bacterial cell walls. By trophic considerations, i.e., biomass produced, algal proteins will be the most prominent source of amino acids. A ratio of these two components, therefore, can yield information on the balance of the plankton community through time, neglecting diagenetic effects (Degens and Hecky, 1973). Diagenesis will be most critical for amino acids, as ammonification by heterotrophic bacteria proceeds rapidly in oxidized sediments while reducing conditions are more conducive to preservation. Chitinous particles appear to be less affected as hexosamines are enriched relative to amino acids in deep ocean sediments (Mopper and Degens, 1972).

In Lake Kivu sediments the amino acid:hexosamine (AA:HA) ratio increases threefold across the contact which marks the onset of strong hydrothermal activity. Better preservation of amino acids would account for the change in part, but the hydrothermal activity could be detrimental to zooplankton in two ways. The first would be direct, as the introduction of heavy metals into surface waters could poison the zooplankton. The second would be indirect, as stratification caused a drop in productivity and change in algal species composition sufficient to eliminate some zooplankters. In modern Lake Kivu calanoid copepods are the predominant zooplankters at all seasons, while cladocerans are abundant only after seasonal mixing when productivity is highest (Verbeke, 1957). In Edward and Albert, cladocerans and copepods are abundant throughout the year, although their relative numbers wax and wane with phytoplankton productivity. The latter lakes have high productivity, high standing crops, and a diatom-dominated phytoplankton, and it may be that one or all of these characteristics are necessary for cladocerans to flourish in these large African lakes, and the absence of these characteristics may explain why cladocerans have never been found offshore in ultraoligotrophic Lake Tanganyika.

Observation of diatom frustules indicates that a shift in the zooplankton population occurred when the AA:HA ratio changed. Below 130 cm nearly 50% of all the *Nitzschia spiculum* frustules are broken. The frequencies depicted in Fig. 8 were generated by counting only those frustules which were greater than one-half an entire frustule. Preparation techniques were identical for samples, and thus breakage was not due to sample handling, which does include rapid agitation. Breakage cannot be attributed to benthic organisms, as the entire core is finely laminated, which requires that bottom fauna be absent. Therefore the breakage must be due only to the activity of zooplankton. The decrease in the percentage of breakage can be explained by either an overall decrease in the feeding activity of zooplankton or a reduction in the specific zooplankton element producing the breakage during feeding. No microfaunal analyses have been done on the core, but it is tempting, in view of the present distribution of zooplankton in the lakes, to conclude that cladocerans decreased in the Kivu about 5,000 years ago as a response to either toxic substances from hydrothermal activities or a change in the quantity and nature of food sources. Large cladocerans harvest large particles (>15 μm) more effectively than small zooplankton (Brooks, 1969).

Most of the diatoms of this study would qualify as large particles, and a reduction in diatom production 5,000 years B.P. might be sufficient in itself to explain the paucity of cladocerans in the offshore plankton of Kivu. As there are no planktivorous fish in Lake Kivu today, reduction of large zooplankters by predation, which is commonly observed in temperate lakes (Brooks, 1969), would not be a factor.

If conditions of preservation do not change markedly in a core as they did 5,000 years B.P., the AA:HA ratio could serve as an indicator of trophic status. A characteristic of eutrophication is that algal production is markedly increased, while zooplankton increase to a lesser degree. These circumstances would lead to an increase in the AA:HA ratio. In the Kivu core, productivity was highest based on the diatom evidence in the shallow, saline lake phase of 13,000 years B.P. Only one analysis is available from that time, and the AA:HA ratio is 20. As the lake was in much better vertical circulation at that time, and deep waters less reducing, this high ratio cannot be attributed to better preservation of amino acids. In this case the high ratio reflects extremely high phytoplankton production and eutrophic conditions.

SHALLOW SOUTHERN BASINS (<100 m water depth)

South Idjwi Basin (Station 13')

At the southern end of the main lake, there is a shallow, shelf-like basin which has a maximum depth of 220 m, and deep sills at 180 m separate the deeper water of this basin from direct exchange with the main lake. The 421 cm core from station 13' in 86 m of water ended in a soil horizon which developed during the low lake stand. A radiocarbon date near the bottom indicates that lacustrine sedimentation was initiated at 12,130 years B.P. This correlates well with the high stand hypothesized from evidence in the north basin. A high sedimentation rate, 1.6 mm/yr, was indicated up to 11,000 years B.P., and this would be consistent with a near shore environment. After 11,000 years B.P. a much lower average sedimentation rate is observed. The chemical stratigraphy of the core (Fig. 9) from this site should reveal the shallow water hydrogeochemistry which was correlative with the deep water record.

Varimax factor analysis indicates that the same five factors are controlling elemental distribution at station 13' as were operating in the north basin. Factor scores for each level in the core are given in Fig. 10. The loading on the detrital factor (factor 1) is especially instructive, as it indicates a relatively deep water stand between 10,000 years and 5,000 years B.P., followed by the return to somewhat shallower conditions. This is explained by downcutting at the Ruzizi which created a gradual lowering of the lake after it opened. The deposition of detrital material as indicated by the TiO_2 and Al_2O_3 content was very high initially. No diatoms were observed in the sediments below 330 cm, which also suggests a very rapid rate of deposition.

The negative loading of over most of the core on factor 2 means that deposition of $CaCO_3$ was high relative to diatom sedimentation, and this correlates well with events of this time period in the northern basin. The very high values for Ca at 11,000 years B.P. support the hypothesis that in the closed lake Ca was removed peripherally. The lower values prior to that represent dilution by the high detrital contributions. Mn has maxima with the Ca, confirming the precipitation of a manganous carbonate in shallow water environments. Fe_2O_3 (not shown in Fig. 9) does not increase, as it was being removed as a sulfide in the deep water during this time. Mn is relatively high in the lowest portion of the core, but it oscillates independently of Al_2O_3 and TiO_2, suggesting it is authigenic. The subsequent period of low deposition suggests that at least the seasonal thermocline was over the site at this time, and weakly reducing conditions were common. In the upper region of the core, most of the Mn present is in the detrital phase. The lake was dilute during this phase, and $MnCO_3$ would not be expected.

Negative loadings on factor 3 are also low over most of the core, and a sulfide facies was never abundant at this site; this is demonstrated by the behavior of Ni (Fig. 9). The Mg factor is likewise negatively loaded through the core, indicating that the lake waters over this site have always been dilute.

The increase of Pb deposition (factor 5) at 5,000 years B.P. reflects the initiation of hydrothermal activity. Not all the Pb was trapped in the deep waters of the north basin by organic chelation and precipitation, but the amounts incorporated in the shallow-water sediments are much lower than in the north basin sediments. The geochemistry of the sediments at station 13' does not appear to conflict with the reconstruction of the lake based on the north basin record.

Bukavu Basin (Station 15')

The Bukavu Basin is isolated from the main lake and station 13' by a shallow sill, approximately 30 m deep at its maximum depth. The Ruzizi River, the only outlet, exits from the southern end of this basin.

The stratigraphic record in the Bukavu Basin should be the best evidence for the opening of the lake and the existence of an outlet. Factor analysis again implicates the same factors as for the other cores, and factor scores for each sample are given in Fig. 10. The highlights of the chemical stratigraphy are depicted in Fig. 11.

The lowest portion of this core, from 290 cm to the bottom, was a light grey color; and carbon dating suggested extremely rapid deposition, >5 mm/yr. Clays and $CaCO_3$ account for most of the sediments, with the latter predominating, often accounting for 70% of the dry weight. Shallow and alkaline saline waters covered the site. The lake was below the Ruzizi outlet during this time. If it were connected with the main lake, 60 m of water would overlay the coring site. The gross stratigraphy suggests shallower conditions approximating an evaporating pan. The first indication of higher lake levels is about 10,000 years B.P. Carbonate deposition fell at this site because the Ca-source (land) was farther away and/or the waters were more dilute. Connection with the main lake would not significantly dilute the waters in itself, as that was also a saline body of water, but it was probably not as saline as the smaller body of water which existed in the Bukavu Basin prior to connection. Regardless of salinity, alkalinity would remain high and Ca-removal would be peripheral until the lake was considerably freshened. Therefore the initial decrease was perhaps not due to overflow at the Ruzizi. Terraces in the Kivu area 100 m above the present lake level suggest that a rather deep lake would have existed prior to overflow. The decrease in $CaCO_3$ deposition at the coring site allowed relative increases in all the other sedimentary components, such as Al_2O_3, TiO_2, and Ni in the detrital phase, and Mo and organic carbon in the biogenic phase. These increases are due solely to the decrease of carbonate sedimentation at site as water depth increased. A very strong negative correlation between the CaO concentration and the detrital and biogenic phases is evident from Fig. 11.

The Mn as usual appears to behave differently than the other elements. Throughout the lower portion of the core it is probably present as a carbonate or oxide, as oxidizing conditions were possible in the shallow water stage. The initial decrease in CaO is recorded in Mn, and it rises again when CaO does also, indicating that conditions were still quite alkaline. At approximately 9,600 years B.P., it begins to decrease sharply, and for several hundred years Mn deposition was very low, even though Ca-deposition achieved high values. This means that the lake had become diluted, and Ca-deposition was a consequence of photosynthesis in offshore waters. A similar change in the nature of Ca-deposition was recorded in the north basin at approximately the same time. The low Mn values suggest slight reducing conditions and decreased Mn-deposition, and the low Mn interval marks the highest lake level in the record. These sediments are not diatomites, and slow accumulation rates are observed. Oxidizing conditions resulted as the river downcut its outlet and lowered the lake level. From 5,000 years B.P. the basin has been completely circulated, and MnO_2 remained in the deep sediment. Even during the high lake stage, the sediments were not sufficiently reduced to allow sulfate reduction. This means that the deep thermocline in the main lake must have been below the sill separating the basins at water depths greater than 130 m during the high stand.

Trace metals in the carbonate phase likewise record the opening and freshening of the lake. In the aragonite deposited in the shallow, alkaline, saline phase, Sr and Ba are very high. They drop sharply when Mn does as a response to a change in the nature of carbonate sedimentation. As Ca-deposition changed from being purely inorganic to being mediated by algal photosynthesis, the carbonate mineralogy changed from aragonite to calcite, hydrocalcite and protodolomite (P. Stoffers, personal communication). The structural change excluded Sr and Ba from the crystal lattice. At 5,000 years B.P., aragonite began to be deposited once again, and the Ba and Sr increase. This change in carbonate deposition represents a return to inorganic processes as hydrothermal activity increased the input of Ca and alkalinity and made deposition in the deep north basin impossible. Lead also increases at the same time, substantiating the correlation with the north basin. Another line of evidence that implicates the springs as the cause of the observed changes is an attempt to date sedimentary organic carbon between 115-170 cm. This material gave an anomolously old date of 9,460 ± 240 years B.P. This was the only attempt made to date sediments deposited after the hydrothermal springs determined the water composition of the lake. The attempt confirms that the carbon from the springs is not atmospheric, and only partial equilibration with the atmosphere occurs. Further dating of sediments deposited in the past 5,000 years would be futile. As a result, age assignments during this interval are only approximate.

There is a pronounced maximum in Mg at a depth of 60 cm. At its peak Mg is nearly equal in concentration to Ca, suggesting that authigenic dolomite may be present. For reasons discussed earlier, this must imply a substantial increase in salinity. The lake dropped below the Ruzizi, and may have had only an intermittent connection with the main lake. Such a connection allowed recharge of waters from Kivu which were already low in Ca and high in Mg. The waters when subjected to evaporative concentration might be favorable to dolomite precipitation directly from solution because of the unusual ion ratios (Liebermann, 1967). There is no evidence that the basin actually dried out during this interval, but neither did it overflow. The shift from aragonite to high magnesium calcite and dolomite once again excluded Sr and Ba from the carbonate phase. This Mg maximum is correlative with the $CaCO_3$-rich sediments deposited between 3,500 years and 1,200 years B.P. in the north basin. The decline in Mg is attributed to regaining the Ruzizi outlet. The consequences for Tanganyika of the intermittent overflow will be discussed later.

LAKE LEVEL OSCILLATIONS AND WATER BUDGET

Kivu

The radiocarbon-dated cores (Table 4) allow an absolute chronology of lake levels to be established. The most reliable datum points are considered to be:

(i) The thick beach deposit at station 14 which marks the lowest level the lake recorded in the time interval sampled;

(ii) the clay facies at station 9;

(iii) the initiation of sedimentation at station 13';

(iv) the first decline in aragonite deposition at station 15'; and

(v) the terraces which surround the lake.

The beach deposit at -310 has not been directly dated as a depositional hiatus is thought to have occurred, but it is thought to reasonably correlate with the bottom of core 4', which has a carbon-dated age of 13,700 years B.P. The clay facies in station 9 has a date of 10,670 years B.P., which means it was deposited during the time not sampled by cores 10 and 4'. This implies that the lake dropped significantly between 11,000 and 10,000 years B.P. by perhaps as much as 250 m. The clay facies extends over 30 cm of depth, and there is no evidence of coarser material or graded bedding. Diatoms are abundant, and *Stephanodiscus astraea* constitutes over 90% of the assemblages. This deposit is not a turbidite, and it is indicative of a shallow water environment. There is no reason to assume that the age is too young. Prior to the injection of volcanic CO_2, the alkalinity of the water was derived from the atmosphere by weathering reactions. There is no carbonate rock exposed in the basin, so there was no source of "fossil" carbon. The only possible error could arise from a long residence time for the HCO_3^{1-}-CO_3^{2-} anions in the closed lake (Broecker and Walton, 1959). Precipitation of Ca and $^{14}CO_2$ invasion across the lake surface would reduce the residence time. All our ^{14}C analyses were performed on organic carbon, and this fraction will be much less subject to incorporation of old carbon than will the carbonates (Broecker and Walton, 1959). Therefore, all the dates on the cores are considered to be reasonably accurate except for the one from station 15' which was discussed earlier.

The initiation of sedimentation at station 13' occurred at 12,130 years B.P., at which time the lake level exceeded -86 m. However, it appears unlikely that the lake overflowed before 9,500 years B.P., when carbonate deposition and the mineralogy changed at station 15'. The record at station 15' extends only to 10,200 years B.P., so it is not possible to tell if the lake overflowed into the Bukavu Basin prior to 10,000 years B.P. It may have before it sank down at 10,500 years B.P. There is no noticeable evidence of the subaerial exposure of station 13 at that time. The low stand was of short duration, and this, combined with the unconsolidated sediments and sloping site, would favor erosion rather than soil development. The terrace at +100 m has been dated at 12,400 years B.P. (Olson and Broecker, 1959), but the date is on shell material and is likely too old. A date of 9,500 years B.P. seems more appropriate, as this high stand probably marked the opening of the lake. Olson and Broecker also dated shell material from a beach at the present level of the lake. A date of 14,000 years B.P. suggested to them this was a fossil beach. This date is likely to be in error by nearly the same factor as the higher terrace, and an age of 11,000 years would be reasonable. This would coincide with the high stand prior to the drop in level at 10,500 years. If the lake stood that high, then it was connected with the Bukavu Basin, but the Ruzizi outlet was still nearly 100 m above the lake.

The only unequivocal evidence for lower stands since 9,500 years B.P. are the Mg-rich sediments in station 15'. The drop in level was not extreme, and an intermittent connection with the main lake appears probable. The sill is at -30 m, so a drop of much more than 30 m would have been only temporary.

The reconstruction of lake levels is presented in Fig. 12, where it is compared to a record of lake level changes inferred by Kendall (1969) for Lake Victoria. A rather good correlation exists, but the Kivu record, because of its high sedimentation rate prior to 12,000 years B.P., exhibits more detail. There is no clear evidence for a low stand in Victoria between 2,000 and 3,000 years B.P., but the closed basin lakes in the Eastern Rift valley record lower levels at that time (Butzer et al., 1972; Hecky, 1971). The level changes in Victoria are slight compared with the tremendous changes in Kivu. This is a result of the morphometry of the basins, with Victoria being a shallow saucer, and Kivu a deep cup. The volume changes would be more comparable. The hydrologic parameters which permit these lake level changes in Kivu will be discussed later in connection with Lake Tanganyika.

Aside from the firm datum points discussed above, the station 4' core suggests somewhat higher levels at approximately 12,000 years B.P. and 13,000 years B.P. on the basis of the fossil diatoms and sediment geochemistry. The prominent 12,000 years B.P. event is probably correlative with the "Bølling" interstadial of Europe. The lower lake which followed would then correlate with the "Older Dryas" stadial, and the succeeding high lake with the "Allerød" interstadial. The very low lake which station 9 records would represent the "Younger Dryas", and the opening of the lake would initiate the Holocene. The 13,000 year event is not identified in the European chronology, but O^{18} fluctuations in the Camp Century ice core (Dansgaard et al., 1971) reveal minor oscillations with approximately a 940-year period which precede the Bølling. These oscillations will be termed "Kivu Oscillations", because the Lake

Kivu record presents the most complete and readily correlated history of the late Pleistocene yet available in tropical Africa. It confirms in great detail that climatic events between the equator and the poles were nearly synchronous through the past 14,000 years.

Tanganyika

The cores available from Tanganyika are short gravity cores. The oldest sediments recovered are in a core taken from the sill (station 19) separating the two basins. The sedimentation rates implied for this core are extremely low, and the possibility of a hiatus seems likely. Microfossils are rare in the core, suggesting dissolution or non-deposition. The cores from the deep basin have much higher rates and abundant fossil diatoms, and one of these was chosen for detailed analysis to gain better resolution for the chemical history and to allow correlation with the diatom stratigraphy. Station 12 is in the north basin in 1,200 m of water. The recovered core is 195 cm long, and is readily correlated with nearby station 15 which was radiocarbon dated. The bottom of the core is assigned an age of 4,000 years. The highlights of the chemical stratigraphy are presented in Fig. 13. In addition, the core from station 2 in the south basin also had radiocarbon dating, organic carbon, and $CaCO_3$ analyses performed. This core was somewhat longer and older and covers over 6,000 years. The $CaCO_3$ stratigraphy is given in Fig. 14 and the organic carbon in Fig. 15, where it is compared to that of stations 15 and 19.

The TiO_2 and Al_2O_3 content are nearly uniform over the length of the core. The Al_2O_3 concentrations presented in Fig. 13 show some variation, as they are given on an expanded scale; but the variability reflects primarily the sedimentation of the other components, especially diatoms, organic carbon, and carbonates. Likewise, conditions appear to have been stable for the Mn system. Lead is relatively invariant except for a pronounced peak at 95 cm. Only CaO concentrations show that the hydrogeochemistry of the lake has changed markedly in the past 4,000 years. From initially very low concentrations, there is a rather sharp rise at 130 cm which extends 30 cm, representing approximately 600 years of deposition. High Ca-deposition rates were maintained for a short time, after which they fell sharply to rates which were still higher than those prior to 3,000 years B.P. Another sharp increase was recorded in the recent past. The south basin core (Fig. 14) demonstrates similar changes, but they are somewhat less abrupt. Organic carbon in the south basin core increased from much lower levels about 4,500 years B.P., while Ca-deposition increased more than 1,000 years later. The record in the north does not extend over this period, but other correlatable cores which cover longer periods (Fig. 15) do have very low organic carbon contents (Degens et al., 1971).

The ionic similarity of surface waters from Kivu and Tanganyika was noted above, and most previous investigators (Beauchamp, 1939; Capart, 1952; Coulter, 1963) have suggested that the Ruzizi accounted for up to one-half the salts delivered to the lake annually. We have seen that the high salinity of the present river is the result of enforced hydrothermal activity which began in Kivu 5,000 years B.P. The increases in Ca-deposition in Tanganyika are related to the events in Kivu, but with an appropriate delay which reflects the large residence times of water and salts in Tanganyika; e.g., it would take nearly 1,000 years for the tributaries of Tanganyika to renew the volume of the lake, neglecting evaporation losses. The Ruzizi is more saline than Tanganyika waters, and hence it flows downslope on entering the lake without immediate effect on surface waters. Only after homogenization with the deep water will any pulse in salts become noticeable in surface waters by mixing across the thermocline. Thus the initial point source of salts (the Ruzizi) becomes a broad diffuse source (the thermocline) and wide areas of the lake become affected nearly synchronously well after the initial input increases.

In Fig. 14 the $CaCO_3$ content of station 10 in Kivu is juxtaposed with the Tanganyika records. The maximum in deposition in the north basin of Kivu occurred during an interval when the lake was closed. If the Kivu basin is the primary source of the calcium in Tanganyika as postulated, then a hiatus in the overflow from Kivu would be reflected after a sufficient lag as a lower deposition in Tanganyika. Such a sequence seems consistent with the available $CaCO_3$ stratigraphy. The increase in $CaCO_3$ deposition in Tanganyika 2,000 years ago cannot be due to evaporative concentration, as the tremendous volume of the lake buffers salinity changes. The peak in Pb concentration indicts Kivu water as the source of the Ca. The correspondence of the maxima in Ca-deposition 2,000 years B.P. in Kivu and Tanganyika is fortuitous, as the mechanisms eliciting the increases are quite different.

Further confirmation of the effect on Tanganyika of the hydrothermal activity in Kivu and the climatically-induced intermittency of the Ruzizi is obtained from analyses of interstitial waters. The pore waters of a core from station 2 have been analyzed (Greene and Jones, 1970), and the results are given in Fig. 16. A general increase in the salinity of the lake over the past 5,000 years is demonstrated. The two ions which are most sensitive to events in Kivu are K and Mg, which are disproportionately over-represented in Ruzizi waters when compared to other inflows to Tanganyika. The effect on the Na and Ca would be less dramatic, as the other rivers carry large amounts of these cations, and Ca is buffered with a solid phase. Both Mg and K exhibit a similar pattern, i.e., an interval of increase, an interval of constant concentration, and a second interval of increase extending to the surface. A similar pattern of increase is seen for Na, but the pause so evident for Mg and K is only a change in slope in Na. As Tanganyika must become closed when the Ruzizi is not flowing, the salinity of the lake would continue to increase, although slowly, without the Ruzizi. That Mg and K do not increase through the period as Na does, only emphasizes that rocks of the volcanoes of the Virunga, which is the second largest tributary to Lake Tanganyika, are the primary source for these cations (cf. the chemistry of the Ruzizi and the Malagaris; in Table 1). It appears from Fig. 16 that the salinity of the lake is still increasing, and this is supported by calculations presented later, which indicate that more salts enter the lake via the Ruzizi than leave the lake via the Lukuga. As the lake is not in equilibrium, calculations of salt residence times have little meaning. In general it is concluded that the Ruzizi is the primary source of salts to the lake, and that Lake Kivu was not overflowing into the lake between approximately 4,000 years and 1,000 years B.P. This chronology for the Ruzizi at Tanganyika agrees well with that constructed for the river at its source.

The initiation of hydrothermal activity in the Kivu basin increased the salinity of the Ruzizi River, which began to increase the salinity of the deep waters of Tanganyika, as the stratigraphy of interstitial water shows. The immense volume of the deep waters allowed salinity to change very slowly. As the increases were slow, unlike the situation in Lake Kivu, mixing processes generally kept pace and no persistent chemocline developed. The Ruzizi River is also cooler than the surface waters (Beauchamp, 1939) and nearly isothermal with the deep waters of the lake. The increased salinity of the Ruzizi after 5,000 years B.P. would carry the cool water plume to greater depths and reduce somewhat thermal equilibration. The thermocline would be sharpened and mixing and productivity reduced. As previously discussed, such conditions would favor long, thin *Nitzschia* sp. Although the diatom stratigraphy is not complete in any detail, it is clear that prior to 5,000 years B.P. in the north basin the relative abundance of *Nitzschia* spp. was lower and *Stephanodiscus astraea* was present. In the south basin *S. astraea* remains prominent in the microfossil assemblage up to the present, indicating better mixing in the south. The depth of detectable O_2 is also greater in the south (Degens et al., 1971). The greater distance from the Ruzizi, and the effects of seasonally strong south winds (Coulter, 1963) presumably account for these differences. A more stable stratification would lead to more strongly reducing conditions and better preservation of organic material. $CaCO_3$ precipitation from surface waters would not begin until the Ruzizi-derived Ca began to increase in surface waters. The temporal succession of these events, i.e., deep water salinity beginning to increase immediately at 5,000 years B.P., organic carbon increasing at 4,000 years

B.P., and $CaCO_3$ precipitation beginning at 2,000 years, reflects the buffering to change which the immense size of the lake engenders.

The above discussion has shown that Lake Kivu and Lake Tanganyika are closely linked, and that the history of the latter cannot be separated from the former. In view of this, a reconstruction of Lake Kivu's late Pleistocene hydrology may allow insight into that of Lake Tanganyika, although we do not have a record for that period from the latter lake. Hydrological reconstructions require detailed knowledge of climate, basin geology as it affects runoff, and lake morphometry. In open lakes, response to changes is reflected primarily in discharge rates while water level changes are subdued. In closed-basin lakes, response is mediated by changes in lake level. The only discharge of water from a closed basin is by evaporation, which will be directly proportional to surface area. An increase in inputs (rainfall and runoff) will cause lakes to rise until losses by evaporation from the increased surface area equal the inputs; and lower inputs will lead to a reduction in surface area and lake level. Thus, when Kivu was closed it is possible to reconstruct, within limits, most probable climates to generate the lake levels observed, following the hydrological relations given in Langbein (1961).

Morphometric and hydrological parameters for the Lake Kivu basin and the Lake Tanganyika basin exclusive of the Kivu basin are given in Table 5. These were partially abstracted from various sources and partially represent newly generated data. For Lake Tanganyika, hydrological estimates for rainfall, evaporation, and inflow to the lakes are those of Theeuws (1920), which appear to agree fairly well with subsequent measurements (Capart, 1952). Morphometric data are based on the map of Capart (1949) and Hutchinson (1957; p. 168). The runoff coefficient is calculated using 900 mm as the average annual rainfall over the basin exclusive of the Lake Kivu drainage area, and dividing this value into the total volume of inflows less the Ruzizi. The data for Kivu are abstracted from Verbeke (1957) except for discharge rate, which was obtained from Amenagement Hydro-Electrique des Chutes de Mururu, which monitors flow over a dam on the Ruzizi. Data for the past 10 years indicate that discharge is nearly 50% larger than previously believed. Evaporation rates for Kivu were estimated from a graph in Langbein (1961). Of special interest in Table 5 are the basin slope parameter and runoff coefficients. Volume (V) is a nearly linear function of area (A) in both lakes because of their trough-like morphology (Figs. 17, 18). In most lakes, area is proportional to some fractional power of volume, because most lakes have a curvilinear cross-section, $A \propto V^{0.4-0.75}$ (Langbein, 1961). For simplicity, Langbein's equations relating these two parameters assume a linear relation, but this will not introduce any significant error for Kivu and Tanganyika, as it will for other lakes. The basin slope, $\Delta V/\Delta A$, for Tanganyika is nearly double that of Kivu, which dictates that to accomplish a given proportional change in area, the volume (i.e., depth) must decrease nearly twice as much. The runoff coefficients for the Kivu basin and the Tanganyika basins are double and one-half the world average respectively (Hutchinson, 1957, p. 230). Rainfall is quite high in the Kivu basin, relief is steep, and rapid infiltration into the porous lavas and ash to the north of the basin effectively reduce evapotranspiration losses. For Tanganyika, exclusive of the Kivu basin, much of the basin area is in the semiarid plateau region of Central Tanzania, and evaporative losses are extreme. The water budgets for the two lakes based on the data in Table 5 are given in Table 6. The estimates for refill time are much lower than those of Beauchamp (1939), which are often cited because he overestimated the mean depth of Lake Tanganyika. Also given in Table 6 are estimated salt losses and storage times. As mentioned earlier, these are not completely valid, as the lakes are probably not in any long-term equilibrium. It can be seen that the Kivu basin is currently contributing twice as much Na and over three times as much K to Tanganyika as is lost via the Lukuga outlet. These values support the interpretation given to the interstitial water stratigraphy. Both water and salts have long storage times in Tanganyika and can be expected to respond slowly to changes in input.

From Tables 5 and 6 it can be seen that the Ruzizi accounts for 18% of the tributary waters, and its volume discharge from Kivu exceeds the Lukuga (using either Theeuw's estimates {1920} or Capart's {1952}). Thus, loss of the Kivu overflow will be sufficient to cause Tanganyika to become closed. Within the past 14,000 years this has happened twice. The question of interest is how far did the level of Tanganyika drop, and what possible effects did it have on the biota of the lakes.

When Lake Kivu stood at -300 m, the lake had an area of 948 km^2 (Fig. 19) and the tributary area was 6,192 km^2. With these values it is possible to generate a family of curves giving possible combinations of mean annual temperature and mean annual precipitation which will be sufficient to maintain a lake at a given level by using graphs in Langbein (1961) and assuming different runoff values. Two sets of curves are given in Fig. 20. One set gives the combination sufficient just to maintain a closed lake at the modern level, and the other set those conditions which would maintain a lake at -300 m. From these sets of possible conditions it is necessary to pick a combination of "most likely" conditions. Schumm (1965) gives a family of curves relating mean annual runoff to precipitation and temperature for numerous basins. From this it can be seen that the modern runoff coefficient is nearly 50% greater than expected. This indicates the effect that topography and substratum porosity have on runoff. The runoff coefficient would be lower under a more arid climate, but it would probably not approach the value for Tanganyika, 0.08, because of these unique basin factors. A value of 0.20, i.e., one-half the modern value, should be reasonable. With a curve selected by runoff coefficient, a selection of mean annual temperature or precipitation determines the other parameter. Temperature is less likely to be highly variable over wide areas than precipitation, and the probable magnitude of temperature changes in Africa during the late Pleistocene have been estimated. Osmaston (1967) calculated the elevation of the firn line during glacial maxima in the Ruwenzori and determined that a decrease in mean temperature of 4°C under modern precipitation conditions would explain the depression of the firn line. Drier conditions would require lower temperatures. Flint (1959) calculated a value of 5°C for Mt. Kenya and 7°C for Mt. Kilimanjaro under similar assumptions. Extrapolation of these values to lower elevations requires an assumption about the lapse rate. A temperature decrease of 3°C in the Kivu basin in the late Pleistocene should be conservative. For a mean annual temperature of 17°C, only 70 cm/yr annual precipitation would be required to maintain the lake at -300 m. If the temperature were lower and/or the runoff coefficient higher, then even less rainfall would be adequate. Present annual precipitation is 130 cm/yr, which shows that precipitation was nearly half the modern value at that time.

If climatic conditions in the Tanganyika basin, exclusive of the Kivu basin, which was closed, exhibited changes of similar magnitude, then the probable level of Tanganyika can be estimated. Zinderen, Bakker and Coetzee (1972) summarize evidence for the African climate of the Late Pleistocene, and conclude that conditions were cooler and drier all over Africa, so that extrapolation of changes from the Kivu basin to nearby areas is reasonable. Assumptions are:

(i) Rainfall fell to 60 cm/yr over the greatest part of the basin, which is two-thirds of the modern value, 90 cm/yr;

(ii) the runoff coefficient was the modern value (this should lead to overestimation of tributary inflows and underestimation of resulting volume changes in the lake);

(iii) mean annual temperature was 21°C cf. modern 24°C; and

(iv) net annual evaporation for the assumed precipitation and temperature was 114 cm/yr (from Langbein, 1961).

For the lake at equilibrium with these conditions, i.e., no volume changes, the following equation can be evaluated:

$$\Delta V = I - E(A_1)$$

where ΔV is the volume change, I the volume of inflows, E the net annual evaporation, and A the lake area at equilibrium. When

$$\Delta V = 0$$

then

$$A_1 = \frac{I}{E} = 7{,}550 \text{ km}^2.$$

This area can be converted to a depth by means of Fig. 21, which relates area and volume to depth. The above conditions imply that the lake was lowered by over 600 m during the Late Pleistocene, which would be sufficient to split the lake into separate basins. The relatively large lowering of the lake is imposed by the high basin slope, which requires relatively great volume reductions to effect a change in surface area. The drop is nearly twice that in Kivu, which has a basin slope only one-half that of Tanganyika. It is drastically greater than the changes Kendall estimated for Victoria and the higher lake levels several investigators have identified in closed-basin lakes in the Eastern Rift Valley (Butzer et al., 1972).

Lake level depressions of this magnitude have been postulated before for Lake Tanganyika. Capart (1949) reported that echo sounding in soft sediments between 150 m and 850 m recorded multiple echoes. He interpreted these multiple returns as indicative of alternating layers of light and dense sediments. In depths of less than 150 m the bottom was relatively hard, and no multiple echoes were observed. In depths greater than 850 m only a single diffuse trace was recorded, indicating a very soft bottom. He hypothesized that oscillations of lake level had allowed the deposition of denser sediments at depths much greater than now possible. At depths less than 550 m as many as eight traces appeared, while at depths between 550 and 850 m only two appeared. Support for such lowerings was also found by the identification of many drowned river courses extending down to 550 m.

Livingstone (1965) recovered a 10.74 m core from 440 m of water at the southern end of the lake in an area where his echosounder recorded one multiple echo in the neighborhood (the fathometer did not have sufficient range to reach bottom at the sampling locality). Livingstone found no soil horizons in his core, but he did find several ash layers, one being quite thick. A radiocarbon date for the ash at approximately 575 cm yielded an age of 11,690 years B.P. Extrapolation of the inferred sedimentation rate to the bottom of the core suggested an age of 22,000 years B.P. On the basis of this, Livingstone could offer no explanation for why the traces would only be found in depths of less than 850 m, for even at his site the volcanic source must have been quite remote. He did suggest, however, that the sublacustrine valleys might have been eroded by density currents.

Our hydrological reconstructions would disagree with Livingstone; and therefore, either our estimations are grossly in error or alternative explanations of his results are possible. Our experience with sedimentation rates in Kivu suggests that extrapolation of the sedimentation rate for the past 11,000 years to the previously deposited sediments could lead to gross overestimates of sediment ages. Livingstone's core is extremely rich in diatoms, as were the Kivu cores. A lowering of lake levels by the magnitude that Capart and we suggest would bring the productive upwelling system (Coulter, 1963) at the south end of the lake near Livingstone's site, as lake levels rose after reaching their minimum. Therefore, the bottom of his core may be younger than supposed and may not record the events of the lowest lake stands. The ash layer may explain Livingstone's multiple echo, but the congruity of that return with those of Capart is a supposition which Livingstone admits. At the least, the possibility of much lower levels during the Late Pleistocene is revived by the demonstration of such levels in Kivu. These lower lake levels keep viable the hypothesis that multiple lake lowerings and separation of the lake into two basins would allow allopatric speciation to account for the endemic species flocks in the lake. The lower lake stand would also explain the very low sedimentation rate at station 19 on the sill as a depositional hiatus or even erosion by wave action.

The more recent loss of the Kivu drainage between 4,000 and 1,000 years B.P. is less easily qualified. Little evidence is available about probable temperatures, as the pollen data becomes complicated by cultural alteration of the vegetation. It can be stated that loss of the Ruzizi alone would cause the level of Tanganyika to fall to -75 m, even if no climatic changes are assumed. If changes in precipitation are assumed to be responsible for the loss of the Kivu overflow, even lower levels are implied. It may well be that the very thin sediment cover at depths of less than 150 m result from the last decline in lake level, as Capart (1949) suggested. To recover from a level of -75 m to overflow at the Lukuga requires that 2,200 km^3 be added to the lake (Fig. 21). If it is assumed that the Ruzizi discharge accumulated without loss, nearly 700 years would be required to cause Tanganyika to overflow. As the lake was certainly lower, longer times would be required to effect overflow at the Lukuga. When first discovered in 1854 A.D., the Lukuga outlet was completely blocked, and the lake did not overflow until 1878 A.D., at which time its level fell markedly before minor oscillations about the modern level prevailed. Climatic oscillations of the past century have been insufficient to close the basin again. This in conjunction with the long response time of the lake suggests that 1878

A.D. marked the first overflow at the Lukuga since approximately 4,000 years B.P.

In view of the high probability of lower lake levels and periods of internal drainage, the salinity stability of the lake comes into question. The "thalassine" aspect of the biota in the lake has been explained as an adaptation to previously much more saline waters (Brooks, 1950). The present lake receives perhaps 50% of its salts from the Ruzizi, and even before the hydrothermal activity began the Kivu drainage probably supplied a disproportionate amount of salts to the lake, as the rocks in the drainage are relatively soluble. During arid periods the composition and salinity of all incoming waters were probably similar to the Malagarasi (120 mg/l). Under the climatic assumptions which placed Lake Tanganyika at -600 m, inflows were 8.3 km^3/yr, which means that approximately 1 x 10^{12} g/yr of salt was added to the lake. At -600 m, the lake's volume, 6,000 km^3, was only one-third the present volume, 18,800 km^3. A threefold increase in salinity would be imposed if no salt loss occurred, and salinity would increase at 0.2 mg/l/yr. Assuming a salinity of the present lake (which is unusually high), a threefold concentration would give a salinity of 1.2 g/l. If the lake is assumed to have been closed for 15,000 years, which appears reasonable from correlation with Lake Albert (Harvey, personal communication) and the Eastern Rift closed-basin lakes (Butzer et al., 1972), a maximum salinity of 4.2 g/l would be generated before dilution began. Lower values are more likely. Therefore, it is clear that Tanganyika has not been more than slightly brackish through its more recent history. As long as the moist Kivu drainage is available in pluvial periods, a saline Lake Tanganyika is not possible.

Truly saline periods must antedate the acquisition of the Kivu drainage and the Virunga event. The seismic work in Lake Kivu puts this well back into the Pleistocene.

Edward

A chemical stratigraphy for the past 6,000 years in Lake Edward, based on a core from station 4, is presented in Fig. 22. The detrital components, TiO_2 and Al_2O_3, covary and reach maxima during intervals where fine ash was noted in the gross stratigraphy. Aside from the ash layers, deposition of detrital material appears to be invariant. Ash deposition has been prominent throughout the time represented by the core, but the exact source is uncertain. The Virunga volcanoes to the south and the Toro-Ankole field to the east and southeast have similar volcanic rocks and would affect water chemistry in a similar fashion. The most recent activity in the Virunga field has been copious lava flows, while numerous explosion craters are present in the Toro-Ankole field, and these may have supplied the ash. The Pb maxima appear to correspond well with the hydrothermal spring activity in Kivu, with peaks at 5,600 years B.P., an interval of reduced activity, and greater activity since 1,200 years B.P. In the latter case the Pb precedes the ash, perhaps indicating a different source. As this core begins with a Pb peak, indicating that hydrothermal activity may already have commenced, it is possible that the radiocarbon date for this core includes some volcanic CO_2, and it may be slightly too old. Nickel also demonstrates maximum concentrations during periods of high Pb and ash concentrations; therefore, it is difficult to attribute the Ni to any particular facies.

Diatoms comprise the bulk of the dry sediments, along with organic material, as might be expected in this highly productive lake. The high Mo values are indicative of the biogenic nature of the sediment. Low values for Mo mark higher deposition of detrital components. The high productivity of Edward through this period of volcanic and hydrothermal activity contrasts strongly with the conditions of Kivu. In Edward the nutrient-rich runoff and springs contributed peripherally to the lake and never stratified it with a chemocline. The shallower mean depth also allows mixing to the sediment surface and relatively efficient mixing. In Edward it is possible that volcanic activity had a positive effect on productivity. In Kivu the hydrothermal springs were injected under the lake and stratified it, and low productivity was the result.

Calcium carbonate demonstrates the most marked changes in deposition of any component in the core. A significant increase occurred approximately 200 years B.P., and subsequent decreases can be attributed to dilution by ash. In the very recent past, Ca-deposition again increased dramatically to 43% and 54% $CaCO_3$ by dry weight at the surface of the core. This tremendous increase requires a much greater Ca contribution to the lake. The Ca is probably derived from surficial weathering of the copious amounts of fresh ash which have recently been added to the basin. This would imply that the lake is somewhat more saline than previously. Manganese appears to covary with $CaCO_3$ with slight increases at 2,000 years B.P. and quite sharp increases at the top of the core. Again, increased inputs seem to be involved. The situation is analogous to Lake Kivu, where increased supply by subsurface weathering led to increased deposition of Ca and Mn in shallow waters. In Lake Edward such sediments need not be restricted to shallow waters, because the lake is not strongly stratified and the salts are added peripherally to surface waters.

Albert

Lake Albert is downstream from Lake Edward, and the connecting Semliki River accounts for a large proportion of the inflows to the lake (Talling, 1963), and thus a close correspondence between the chemical stratigraphies of the two lakes might be expected. However, the sediments of the two lakes are quite different, as the biogenic components are greatly reduced in Lake Albert. Organic carbon is low throughout Albert sediments, and the diatoms are highly fragmented and dissolution may be significant. Oxidizing conditions through all depths permit an active benthic fauna to efficiently regenerate organic materials. As a consequence, the Lake Albert stratigraphy allows a clearer insight into variations in the inorganic phases.

The chemical stratigraphy of two cores from Lake Albert are presented in Figs. 23 and 24. Station 5 is from the maximum depth of the lake, 58 m. Sedimentation rates have been higher at that station than at station 3, where a much longer record is available. Station 3 is somewhat closer to shore also. The core from station 3 terminated in a dry, carbonate-rich layer. Ca, Sr, Mg, and Mn are all enriched in this layer, and dolomite and calcite have been identified by X-ray mineralogy. This layer is underlain and mixed with sands of high quartz and feldspar content (U. Briegel, personal communication). This layer undoubtedly marks a lower stand in Lake Albert correlative with that of Kivu.

The detail for this period which Lake Kivu supplies is not evident here, as desiccation caused one or more hiatuses. However, the occurrence of a secondary $CaCO_3$ and Mn at 400 cm may imply that the lake did not open until then. Dating of this core is by correlation, and possible hiatuses make the age assignments imprecise. Much higher lake levels followed, and the lake overflowed at the Albert Nile. Dilute conditions are implied from the very low rates of $CaCO_3$ deposition. Clays are by far the most prominent sedimentary constituent up to 5,000 years B.P., as both the station 3 and station 5 cores indicate. In both cores there is evidence of fluctuations in Mn concentration through this period. The high concentrations represent oxidizing conditions, whereas the low concentrations indicate Eh's sufficiently low to liberate Mn from the sediment. No effect was seen on Fe; therefore, the sediment-water interface was only weakly reducing. The lake was dilute throughout this period, and it is unlikely that a manganese carbonate phase was involved.

At 5,000 years B.P. the hydrogeochemistry of the lake changed, as did all the other lakes in the Western Rift valley. In both cores Pb shows a peak, and the carbonate and organic carbon content increase. At station 5 the increases are gradual, with the highest concentrations of carbonate near the top of the core. At station 3 the increase in carbonate is abrupt, and $CaCO_3$ achieves higher values than those in the dry layer at the bottom of the core. This high carbonate layer at 5,000 years B.P. does differ from the earlier one in that Mg does not increase. The waters were ionically richer after 5,000 years B.P. but not highly saline. The Mg content does rise, but it levels off at around 3%. Much of this may be in the clays. The clay facies in Albert differs from that in Kivu and Edward, as montmorillonite is present and accounts for nearly one-half the clay by weight. The montmorillonite incorporates more Mg than the kaolinite and illites which predominate in the other lakes. The sharp peak in $CaCO_3$ in station 3 contrasts noticeably with the gradual rise in station 5, but this is best explained by different sedimentation rates and the nearshore position of station 3. The $CaCO_3$ precipitation is mediated by algal photosynthesis driving up the pH of the water. In large African lakes the highest productivity is measured nearshore, as nutrient input and regeneration are greatest there. Therefore, the highest Ca-deposition would be expected in shallower, nearshore waters (Kendall, 1969). The $CaCO_3$ content of station 5 does eventually exceed that of station 3, so that differing sedimentation rates are also involved. As we have reconstructed the chronology, vastly different rates are evident at station 3, while station 5 yields a nearly constant rate.

Organic carbon rises significantly in the station 3 core after 5,000 years B.P. As the absolute contribution of the clays is thought to have remained the same while $CaCO_3$ increased, this rise in organic carbon probably reflects increased productivity. Greater nutrient inputs from the Virunga hydrothermal center and the weathering of fresh ash in the Toro-Ankole field caused a natural eutrophication of the lake, perhaps increasing productivity by a factor of two. The higher organic content made the sediments slightly reducing, and the Mn fluctuations which were noted earlier ceased after 5,000 years B.P. This is quite clear for station 3, and it is true for station 5, although the expanded scale of Fig. 24 obscures it somewhat. The effect of the hydrothermal waters on the productivity of Lake Edward was not noticeable, as we may not have recovered sediments which predated the onset of spring activity and/or the lake was quite productive previously and was not as sensitive to increased nutrient input. The Semliki is thought to be the primary source of salts and nutrients to the lake, so that the Albert stratigraphic record suggests that nutrient "overflow" from Lake Edward significantly increased 5,000 years ago. Diatom analysis of the Albert station 5 core may demonstrate the effect of the natural eutrophication on the phytoplankton of the lake.

Malawi

Lake Malawi has no hydrological connection with the Western Rift valley and could not be directly affected by the events in the Virunga hydrothermal center which have dominated hydrogeochemistry of the lakes of that rift for the past 5,000 years. However, Malawi is part of the same rifting system, and it offers the opportunity to see if similar events have occurred in the Nyasan rift. If they have, it becomes probable that the Virunga hydrothermal center is not a localized phenomenon, but merely an expression of the initiation of rifting activity on a much greater scale. The chemical analysis of the sediment is presented in Fig. 25.

The presence of several Pb peaks in the sediments suggests that hydrothermal activity is present in the basin. The detrital elements are quite variable, as would be expected from the frequent turbidites observed. Mn and Fe_2O_3 have maxima together between 40 and 80 cm, suggesting an oxidized phase, but at other core intervals they do not covary, implying reducing conditions. Organic carbon and $CaCO_3$ achieve maximum concentrations, while Pb is high between 120 and 150 cm, which may indicate that hydrothermal activity supplied more Ca and nutrients, thereby increasing productivity. However, the C:N ratio reaches a maximum at that time, which could be caused by relatively lower algal production or greater input of allochthonous organic material, as was discussed earlier. Alternatively the high C:N might indicate warmer bottom temperatures, and more rapid deamination of amino compounds as suggested by Mopper and Degens (1972).

Radiocarbon dates of a nearby core (Von Herzen and Vacquier, 1967) give an age of about 2,000 years for the topmost sediment and a slight but consistent increase in age with depth. Assuming that the sediments of this core have been "contaminated" by dead carbon to the same extent, the figures would translate to sedimentation rates in the order of 3 m per thousand years. This rate is not unreasonable in view of the lithological similarity to Kivu core 4'. These high rates would imply that core 15 collected by the Heidelberg expedition is less than 1,000 years old. We tentatively conclude that the hydrothermal activity noticeable in this core corresponds to the last Kivu event, which started about 1,200 years ago. It is conceivable that correlations of hydrothermal events recorded in older East African rift lake sediments reveal a synchronous pattern.

There is already sufficient evidence of periodic hydrothermal activity to allow a hypothesis of related rifting activity all along the African and Red Sea rift system. Such behavior would be similar to that proposed for mid-ocean ridges, where rifting and hydrothermal activity are considered complementary processes.

CONCLUSIONS

The stratigraphic records for five rift valley lakes in Central Africa provide insight into the interplay of climate and geology in determining the hydrogeochemistry of the lakes during Late Pleistocene and Holocene times. The Late Pleistocene lakes were closed and more saline, as a cooler and much more arid climate prevailed. Climatic oscillations which marked the transition from this period into the warmer and wetter Holocene appear to be readily correlated with the temperate European record and ice cores from Greenland. Worldwide cooling is indicated, with aridity occurring in continental areas with much higher modern precipitation as evaporation from a cooler and more saline ocean was reduced.

Volcanic and hydrothermal activity increased sharply in the Virunga volcanic region 5,000 years B.P. The copious supply of saline water bearing the unmistakable chemical "fingerprint" of the distinctive Virunga rocks now dictates the chemical composition of all the Western Rift valley lakes. Similar hydrothermal activity may have begun in the Nyasan Rift at the same time, possibly marking a renewal of rifting activity. Water may be a necessary prerequisite for volcanic and rifting processes, and the African climatic sequence of pluvials and interpluvials may have caused corresponding periods of activity and quiescence in the geological record.

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LIST OF FIGURES

Fig. 1 Location map of Lakes Malawi, Tanganyika, Kivu, Edward and Albert, showing core sites discussed in present work. Kivu stations occupied in 1972 are marked with prime (e.g., 4') to distinguish them from the 1971 sites.

Fig. 2 Gross stratigraphy and sedimentological description of sample material.

Fig. 3 Chemical stratigraphy of Kivu stations 10 and 4' (Deep Northern Basin). Only those elements are shown which show wide ranges in concentration and are of particular use in the reconstruction of the paleoenvironment. A complete file on all elements has been published in the form of a Technical Report (Degens and Kulbicki, 1973). The upper 4 m of the profile are from station 10, and the lower 8 m from station 4'. There is a hiatus of about 1,000 years between the base of station 10 and the top of station 4'. Nearby station 9 (not shown in graph) bridges about 700 years of this gap.

Fig. 4 Distribution of manganese in Kivu sediments deposited during the time interval between 13,700 and 11,000 years B.P. High manganese values go on account of manganosiderite, which is deposited in shallow water environments. Fluctuations in Mn contents are explained by oscillations of the thermocline which shifted the area of manganosiderite deposition away or in the direction of the present core site. High water stands are

Fig. 4 (cont.) indicated by the emergence of sulfides and specific diatom assemblages. Mineral facies and reconstructed thermocline oscillations are indicated. The water depth at low stand was *ca.* 300 m lower than today; the high water mark stood approximately at -100 m. The fluctuations in water level before 12,300 years B.P. are referred to as "Kivu Oscillations" (Degens and Hecky, 1973); they exhibit one major high between 12.9 and 13,100 years B.P.

Fig. 5 Dominant diatoms of Lake Kivu sediments deposited prior to 5,000 years B.P. (2 micron scale).

Stephanodiscus astrea (a-c)

Nitzschia fonticola (d,g,h)

Nitzschia spiculum (e)

Nitzschia acicularis (f)

Fig. 6 Dominant diatoms and a crysophyte in brown layers (2 micron scale).

Nitzschia accomodata (a)

Nitzschia sp. (b)

Nitzschia palea (c)

Nitzschia bacata (d,g)

Nitzschia subcommunis (e)

Paraphysomonas vestita (f,g)

Fig. 7 Dominant diatoms of Lake Kivu sediments deposited since 5,000 years B.P. (2 micron scale)

Chaetoceros sp. (a,b,g)

Nitzschia spiculum (f)

Fig. 7 (cont.) *Nitzschia bacata* (c)

Nitzschia mediocris (d,e)

Fig. 8 Distribution of amino acids (AA), hexosamines (HA), organic carbon, AA/HA ratios and percentages of broken frustules of *Nitzschia* sp. in Kivu core 10. The AA/HA ratios, AA and HA contents are expressed relative to organic carbon (Mg/g C).

Fig. 9 Chemical stratigraphy of Kivu station 13 (South Idjwi Basin).

Fig. 10 Factor scores of Kivu stations (a) 10 and 4' (Deep Northern Basin) and (b) 13 and 15 (South Idjwi and Bukavu Basin).

Fig. 11 Chemical stratigraphy of Kivu Station 15 (Bukavu Basin).

Fig. 12 Reconstruction of lake level in Kivu for the last 15,000 years.

Fig. 13 Chemical stratigraphy of Tanganyika station 12 (Northern Basin).

Fig. 14 $CaCO_3$ stratigraphy of Kivu and Tanganyika sediments.

Fig. 15 Organic carbon content in Tanganyika sediments from stations 15 (North), 19 (Central sill), and 2 (South).

Fig. 16 Interstitial water analyses (Lake Tanganyika, station 2; after Greene and Jones, 1970).

Fig. 17 Lake Kivu: linear volume/area function indicating trough-like morphology.

Fig. 18 Lake Tanganyika: linear volume/area function indicating trough-like morphology.

Fig. 19 Lake Kivu: volume/area/depth relationships.

Fig. 20 Runoff coefficient: relationship between precipitation and evaporation for Kivu (A) and Tanganyika (B). Indicated are values for modern precipitation and evaporation.

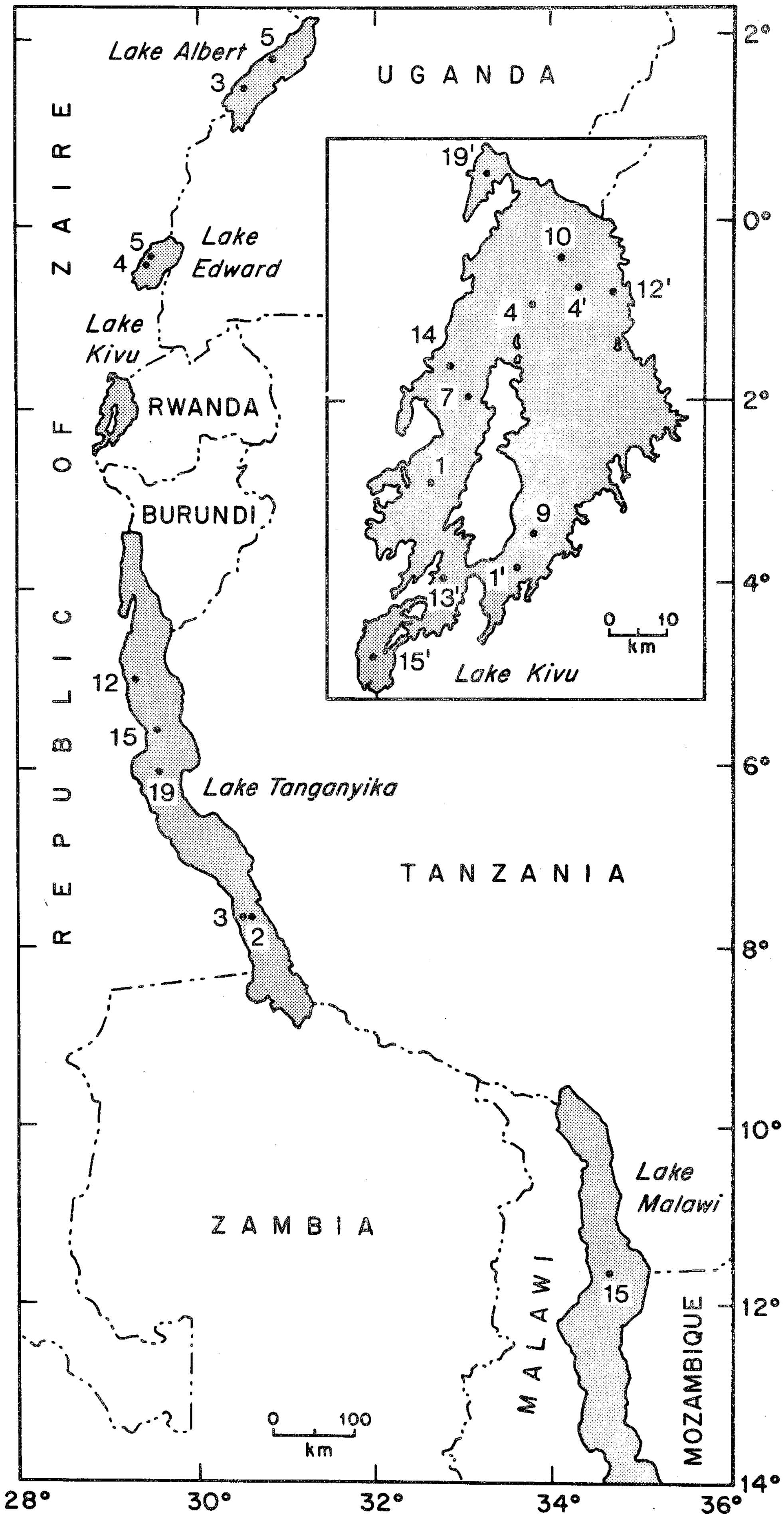
Fig. 21 Lake Tanganyika: volume/area/depth relationships.

Fig. 22 Chemical stratigraphy of Edward station 5.

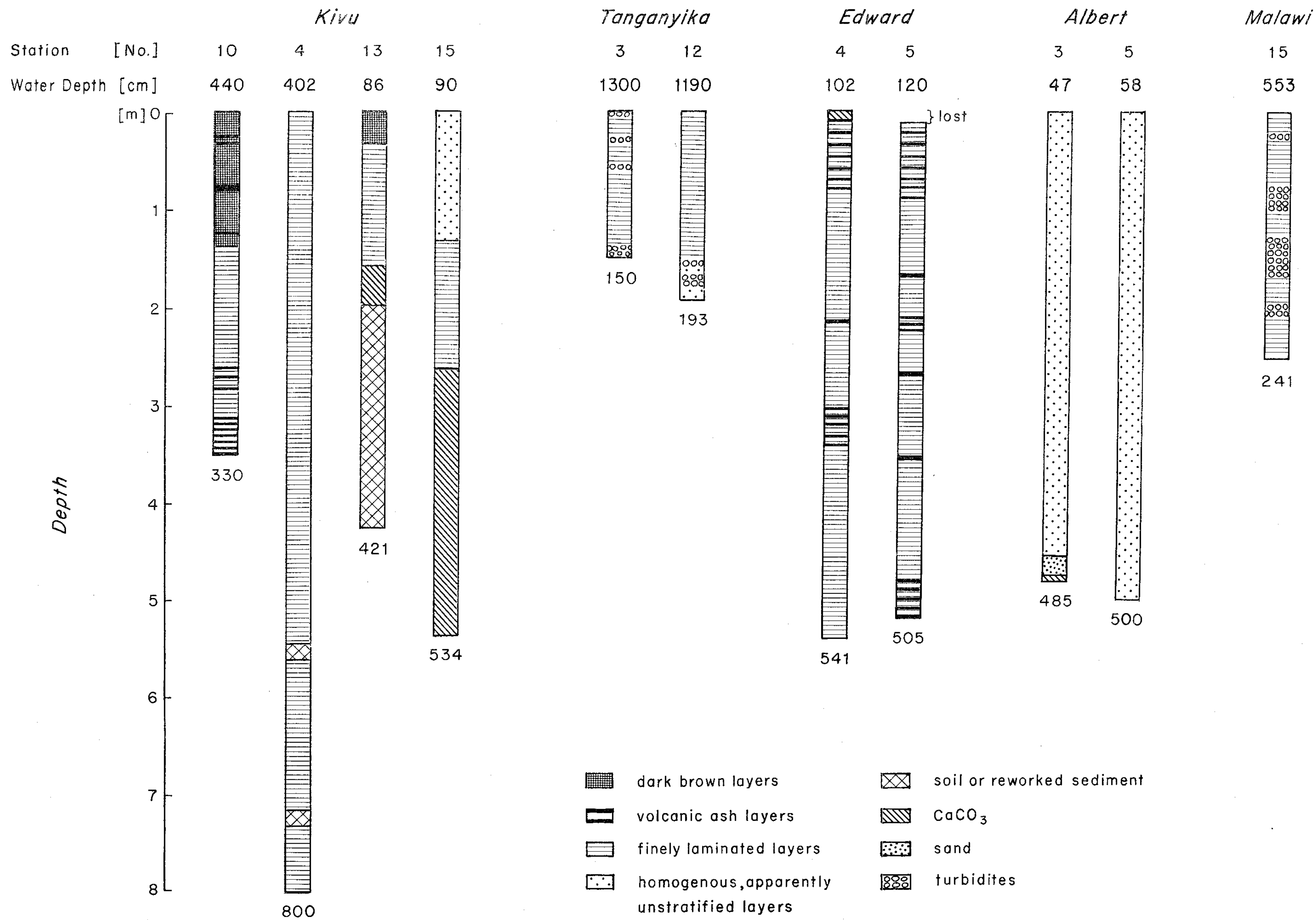
Fig. 23 Chemical stratigraphy of Albert station 3.

Fig. 24 Chemical stratigraphy of Albert station 5.

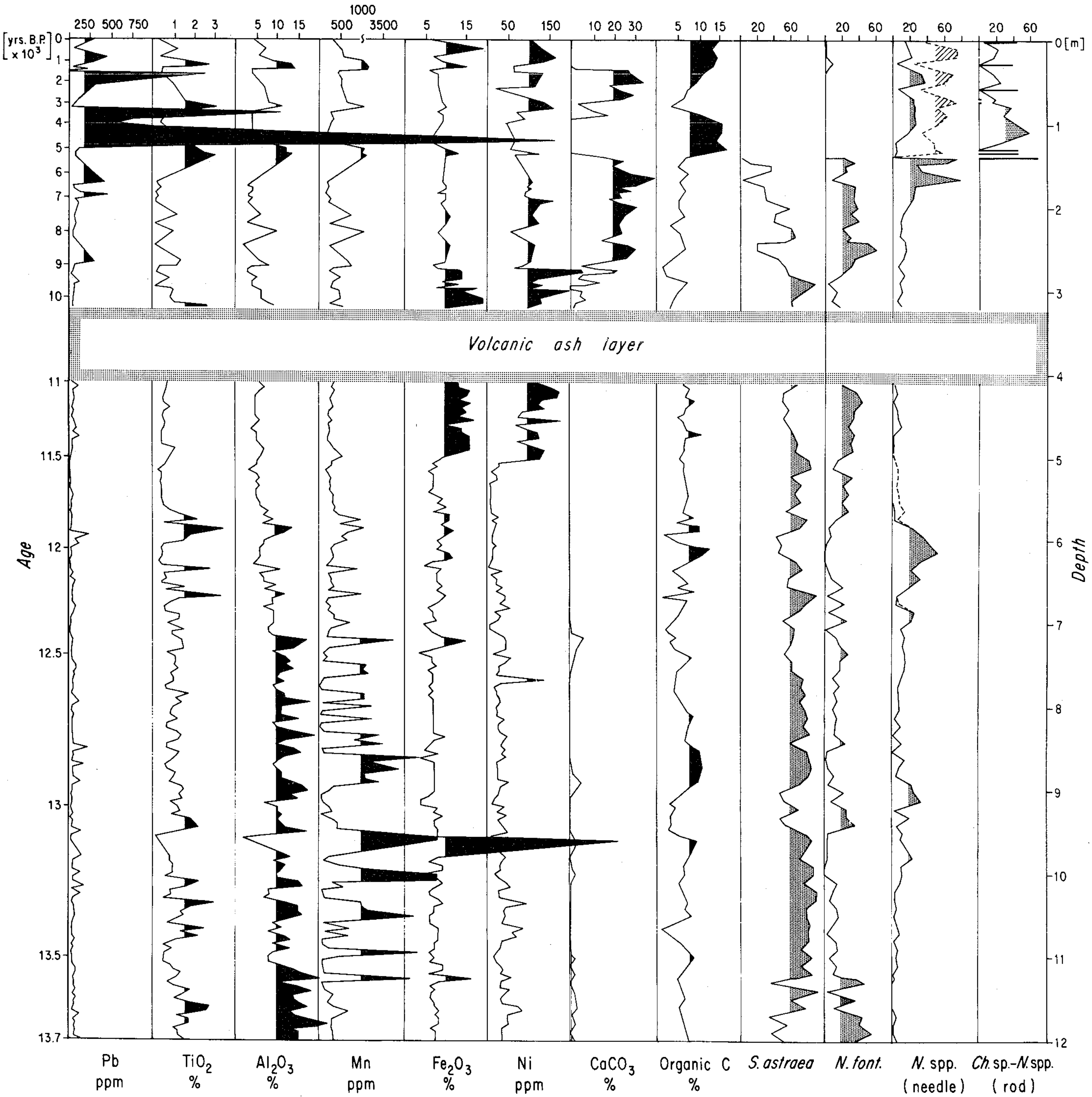
Fig. 25 Chemical stratigraphy of Malawi station 15 (Heidelberg expedition, 1972).



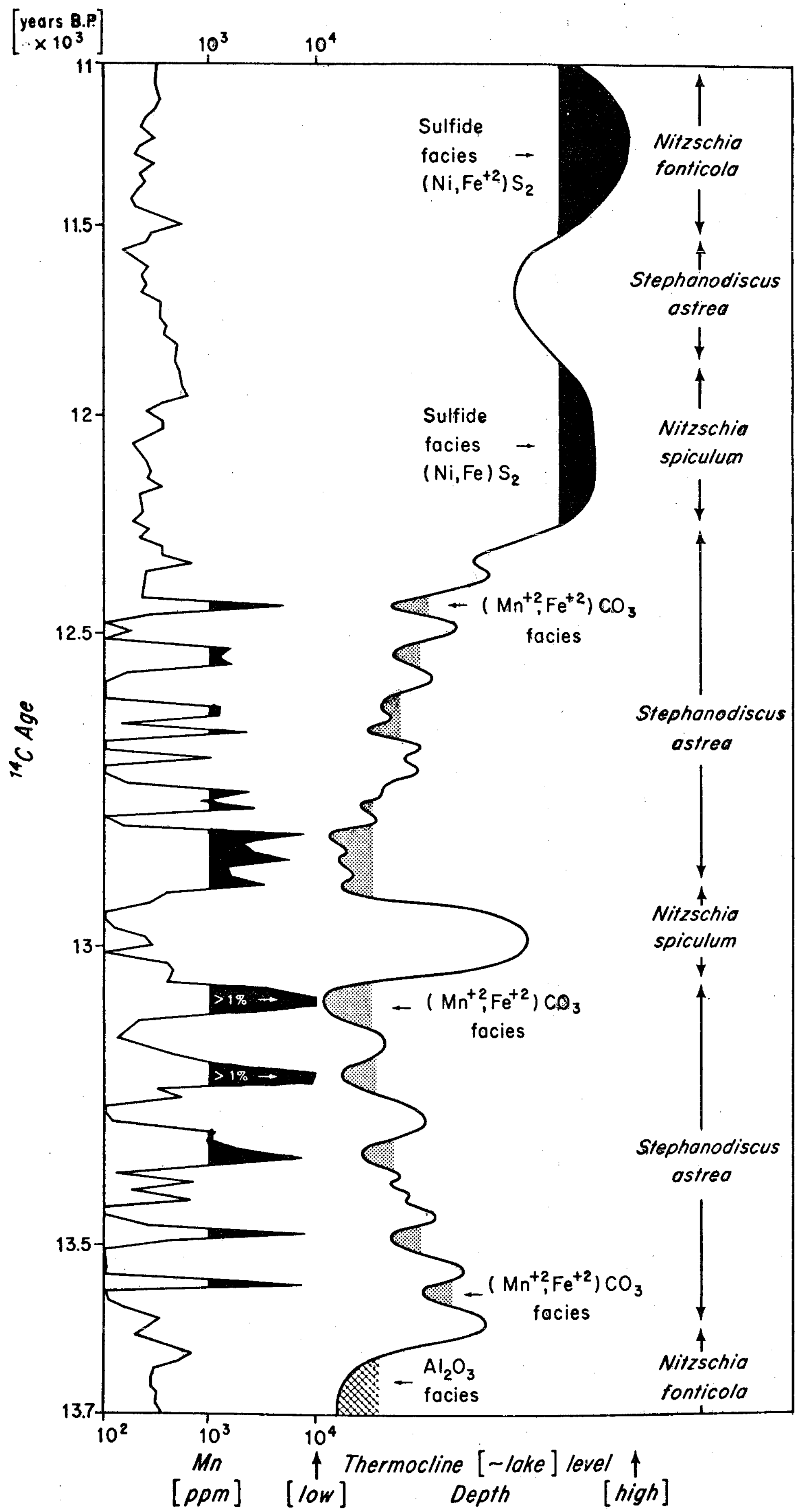
Lake Albert
5
3
UGANDA
ZAIRE
REPUBLIC OF
Lake Edward
5
4
Lake Kivu
RWANDA
BURUNDI
12
15
19
Lake Tanganyika
TANZANIA
3
2
ZAMBIA
MALAWI
Lake Malawi
15
MOZAMBIQUE
0 100
km
19'
10
12'
4
4'
14
7
1
9
1'
13'
15'
0 10
km
Lake Kivu
2°
0°
2°
4°
6°
8°
10°
12°
14°
28°
30°
32°
34°
36°



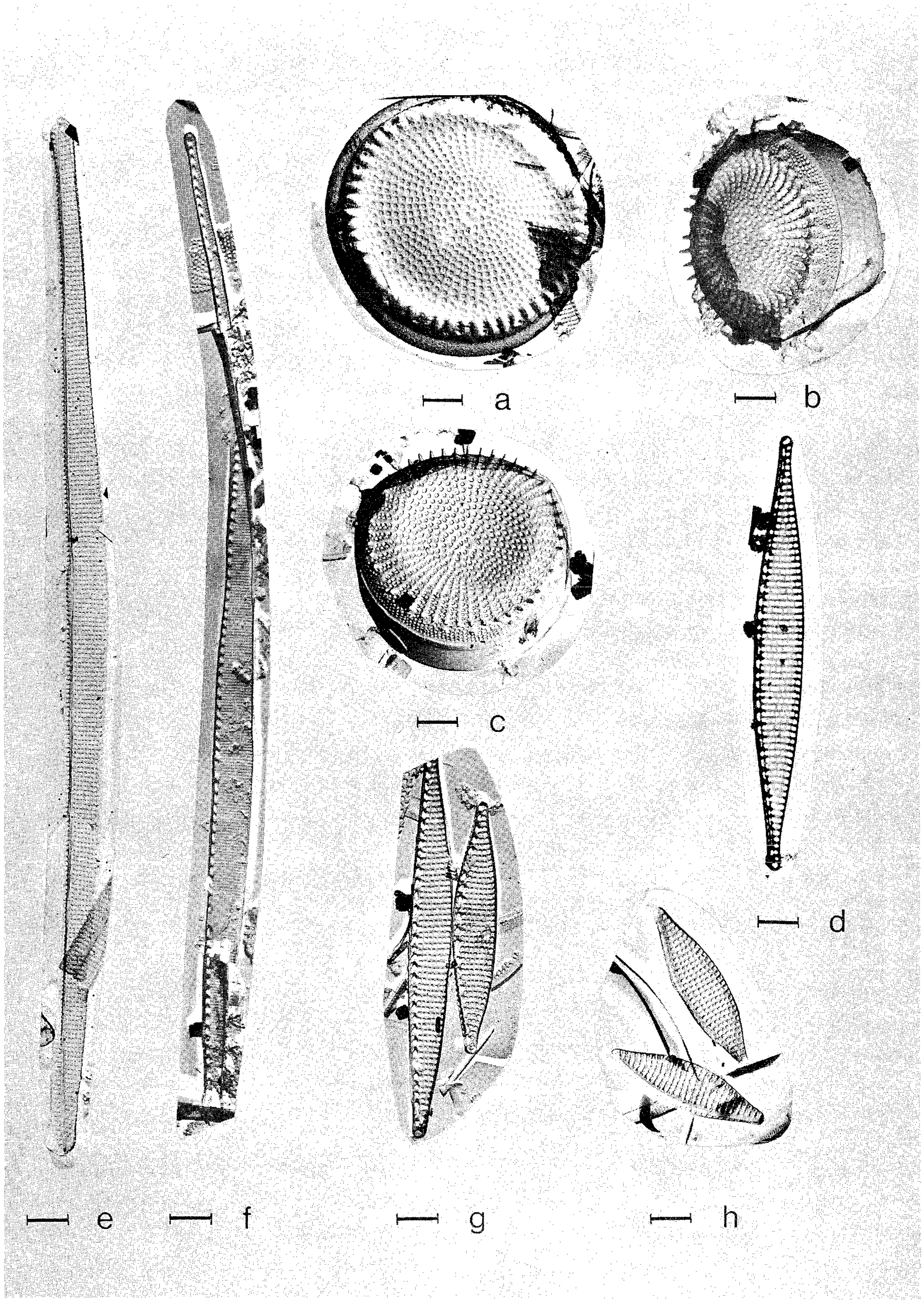
Kivu
Tanganyika
Edward
Albert
Malawi
Station [No.]
10
4
13
15
3
12
4
5
3
5
15
Water Depth [cm]
440
402
86
90
1300
1190
102
120
47
58
553
[m] 0
1
2
3
4
5
6
7
8
Depth
lost
330
800
421
534
150
193
541
505
485
500
241
dark brown layers
volcanic ash layers
finely laminated layers
homogenous, apparently unstratified layers
soil or reworked sediment
$CaCO_3$
sand
turbidites



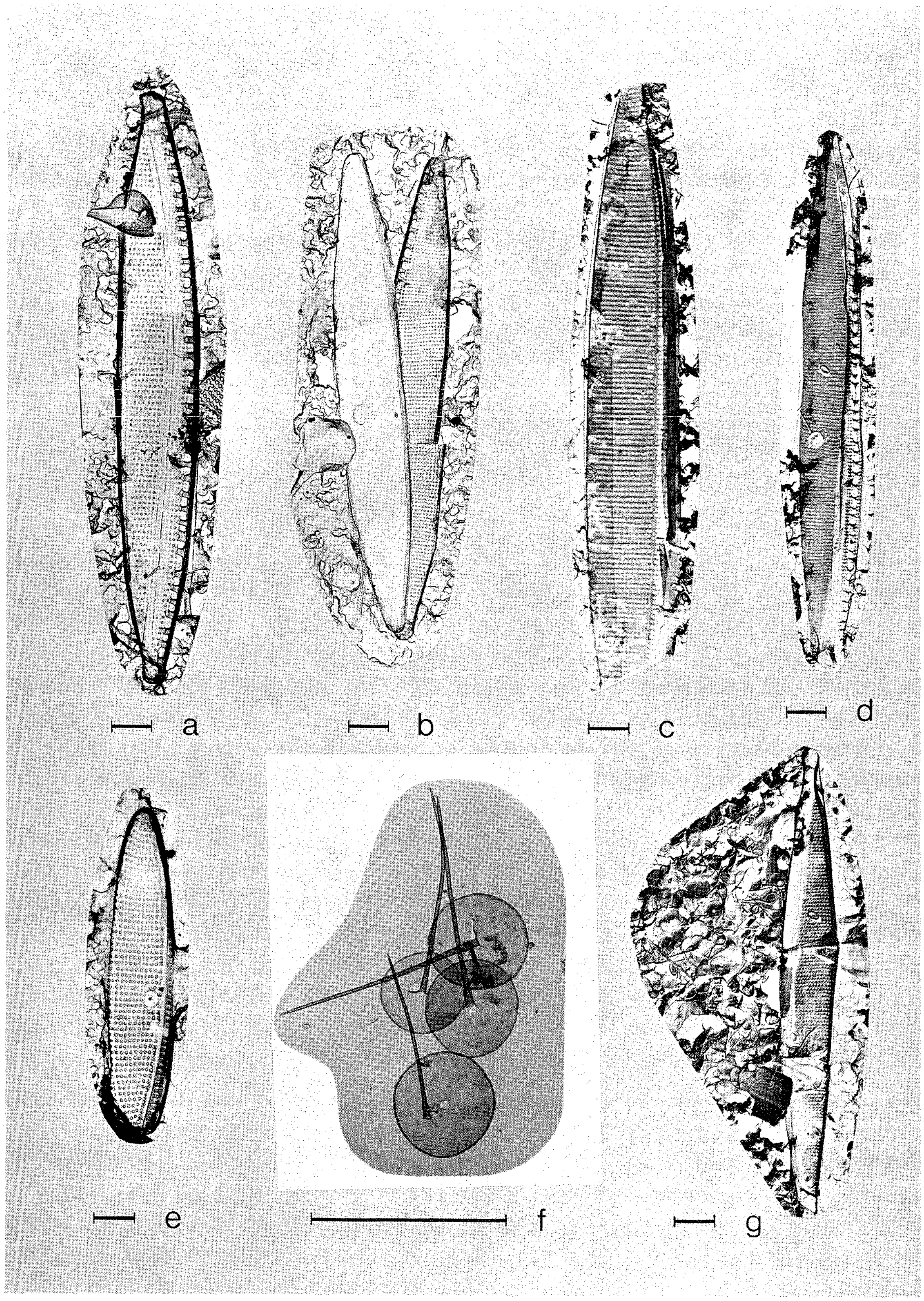
[yrs. B.P. x 10^3]
Age
Depth
0[m]
Volcanic ash layer
Pb ppm
TiO2 %
Al2O3 %
Mn ppm
Fe2O3 %
Ni ppm
CaCO3 %
Organic C %
S. astraea
N. font.
N. spp. (needle)
Ch. sp.-N. spp. (rod)

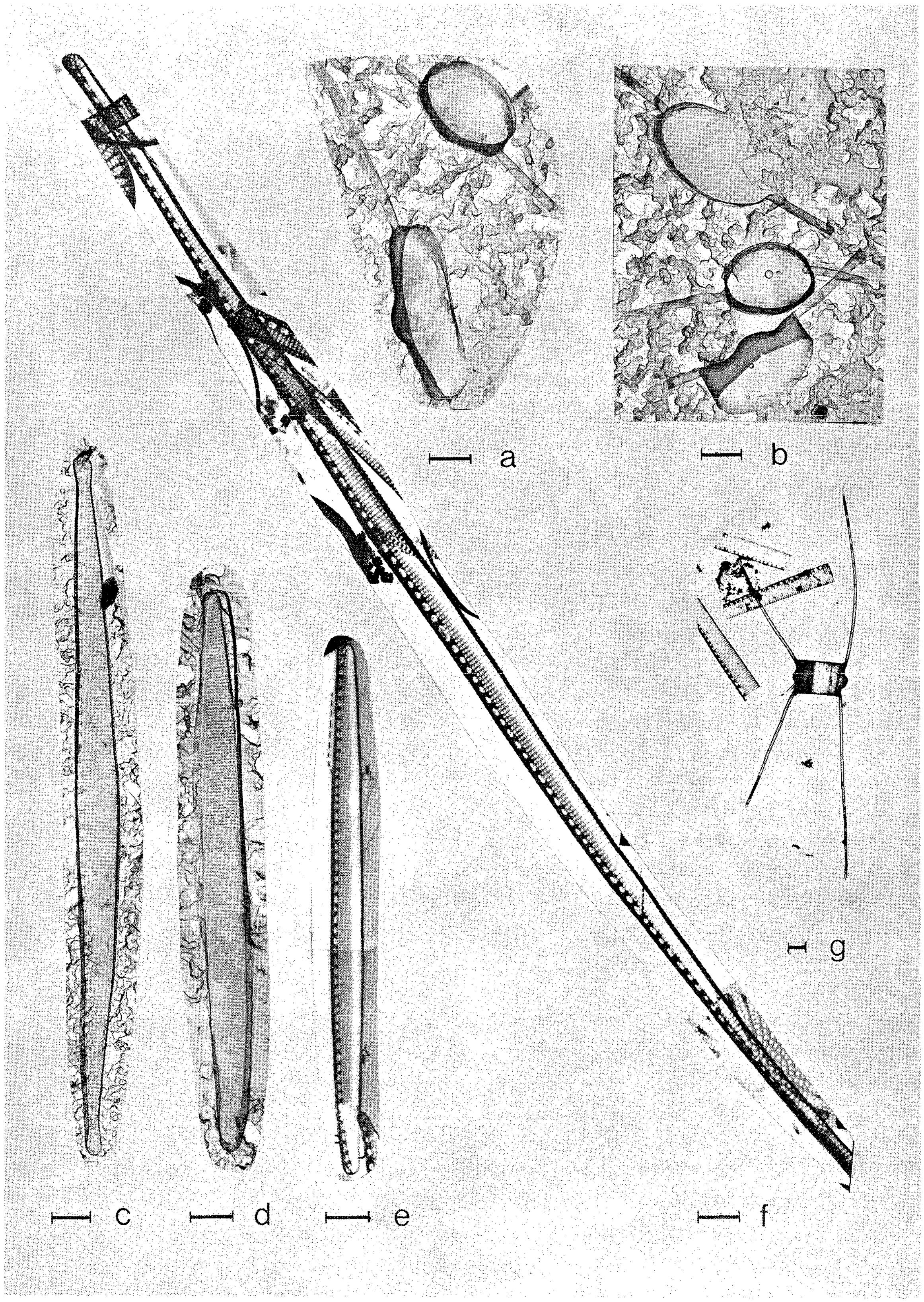


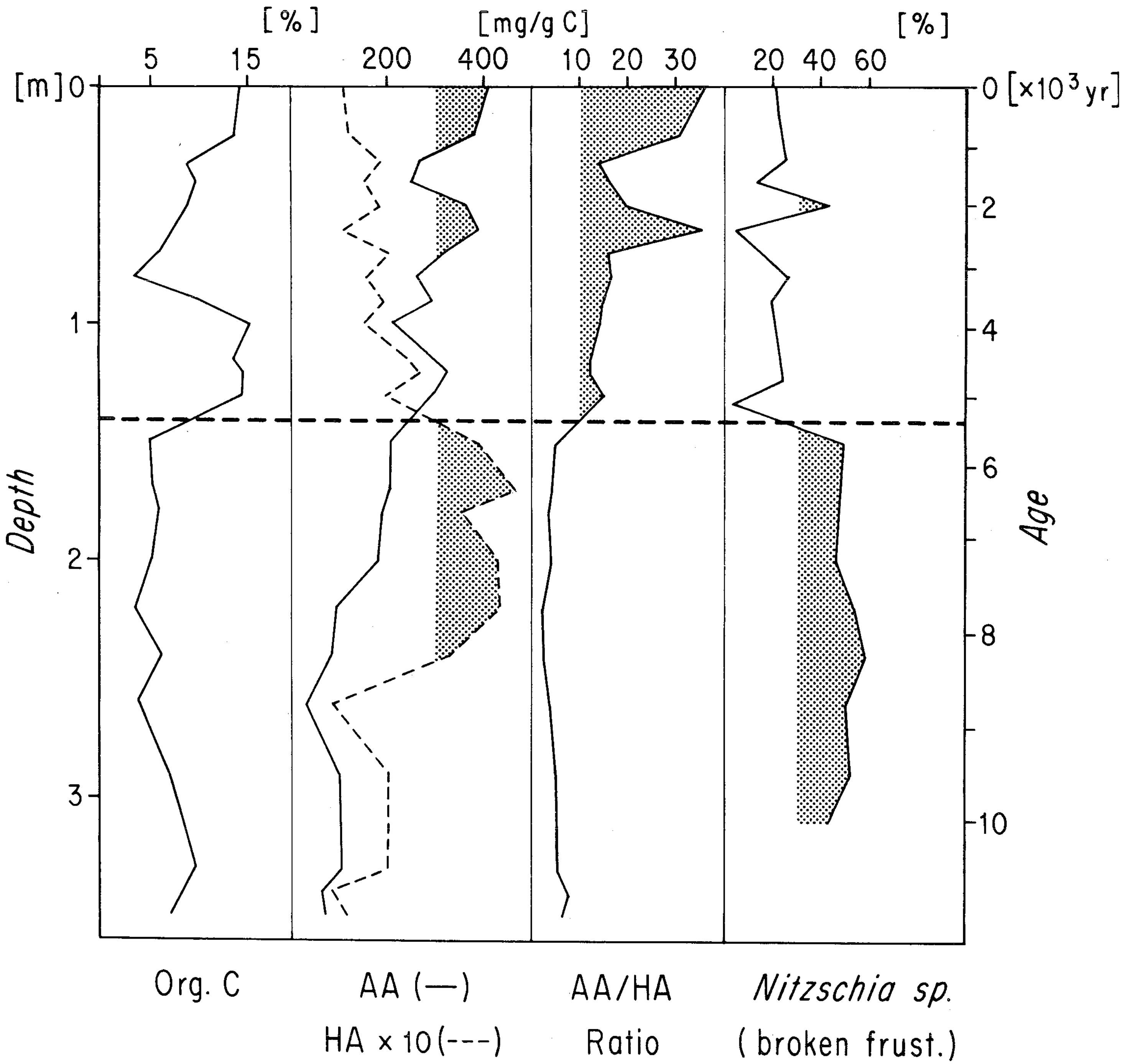
[years B.P. × 10^3]
10^3
10^4
11
11.5
12
12.5
13
13.5
13.7
^{14}C Age
Sulfide facies $(Ni,Fe^{+2})S_2$
Sulfide facies $(Ni,Fe)S_2$
$(Mn^{+2},Fe^{+2})CO_3$ facies
$(Mn^{+2},Fe^{+2})CO_3$ facies
$(Mn^{+2},Fe^{+2})CO_3$ facies
Al_2O_3 facies
>1%
>1%
Nitzschia fonticola
Stephanodiscus astrea
Nitzschia spiculum
Stephanodiscus astrea
Nitzschia spiculum
Stephanodiscus astrea
Nitzschia fonticola
10^2
10^3
10^4
Mn [ppm]
Thermocline [~lake] level
[low] Depth [high]

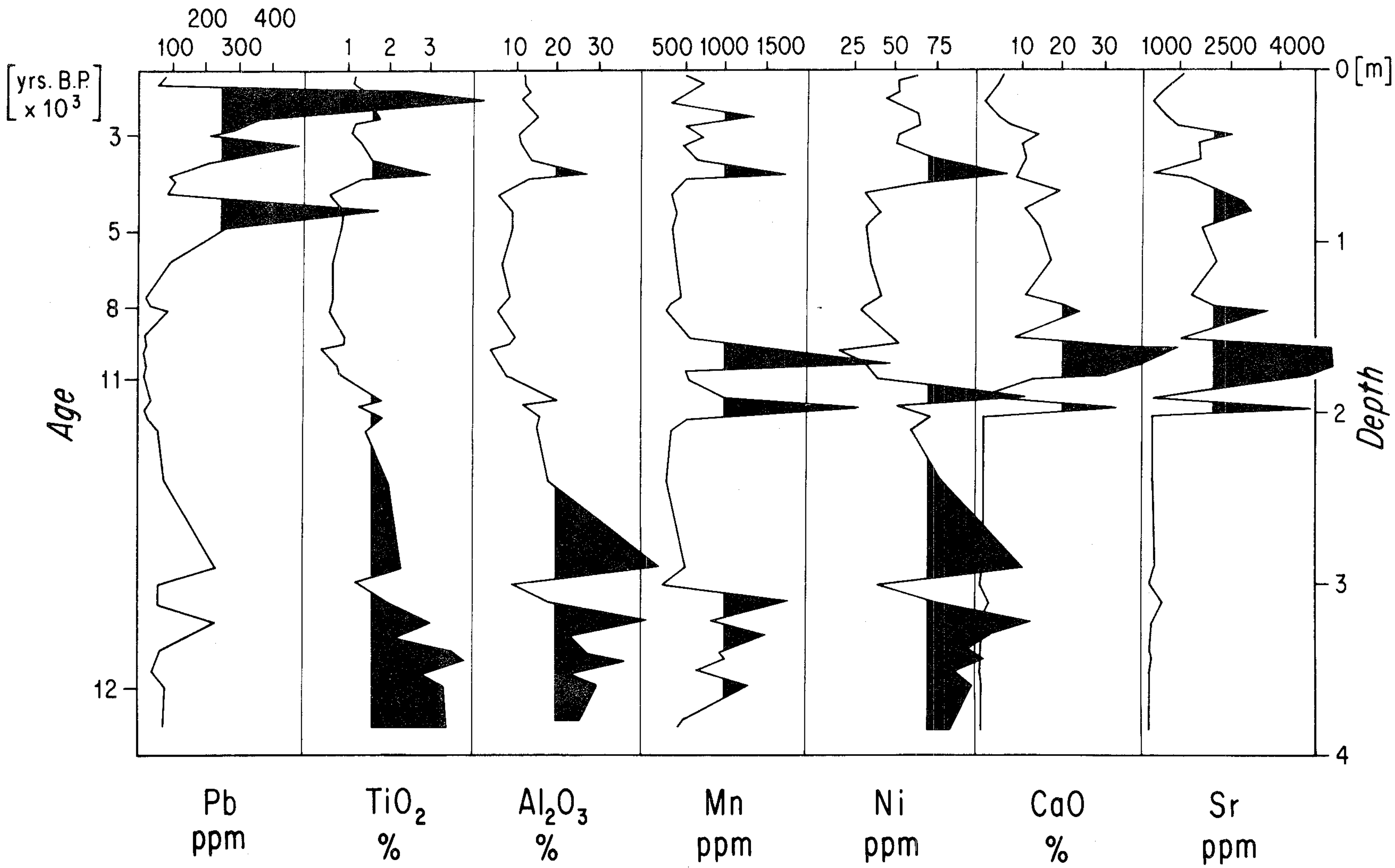


a
b
c
d
e
f
g
h

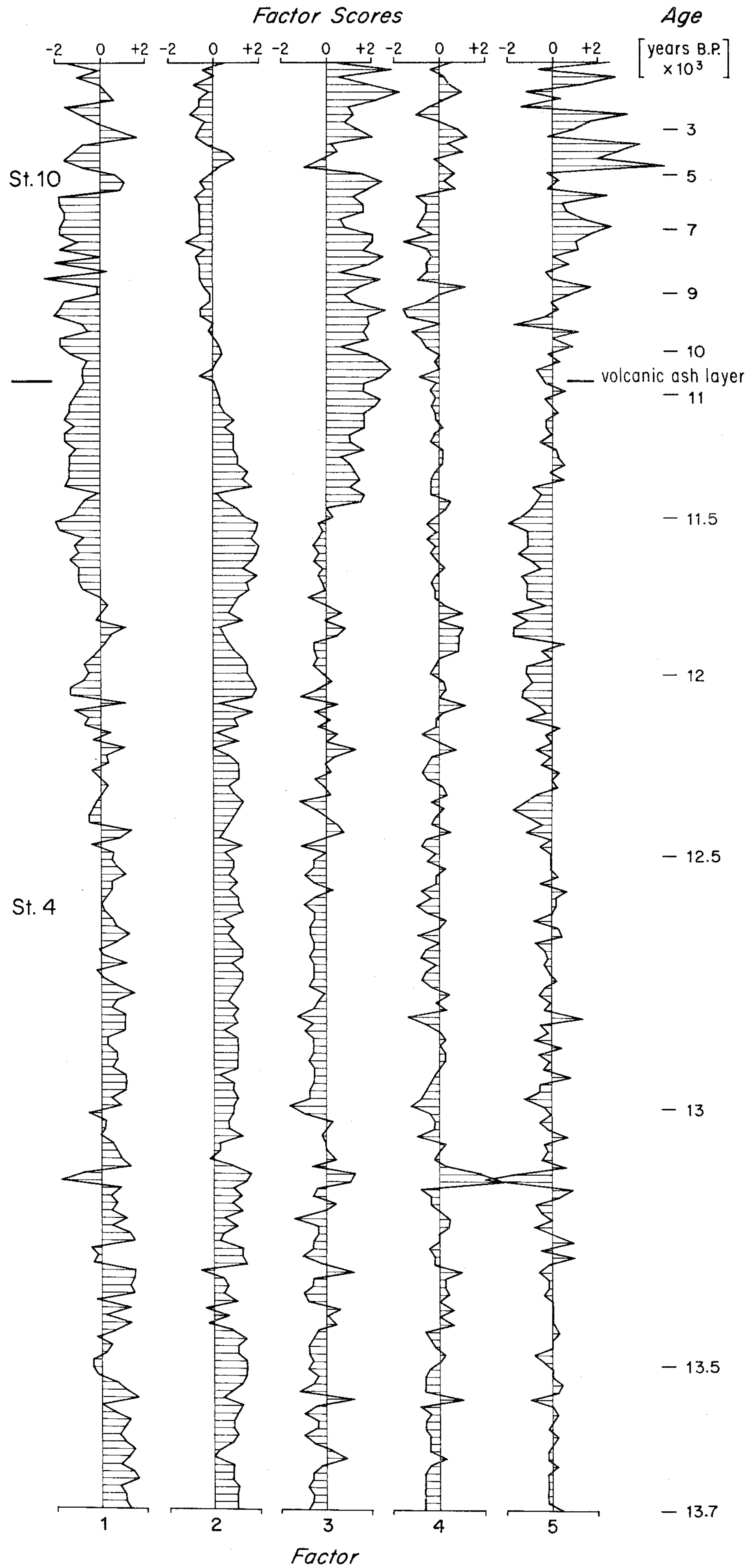

a
b
c
d
e
f
g


a
b
g
c
d
e
f

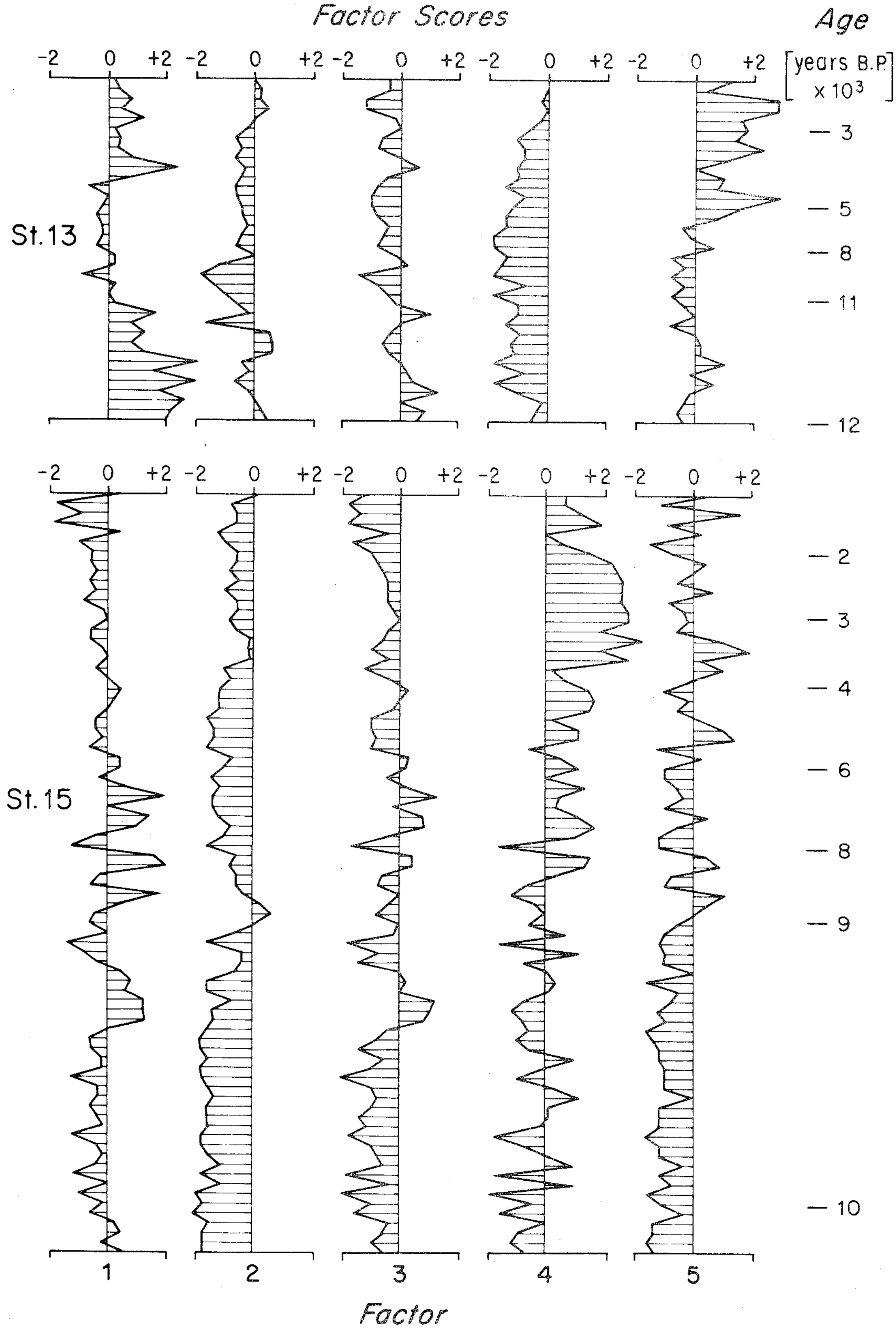

[%]
[mg/g C]
[%]
5
15
200
400
10
20
30
20
40
60
[m] 0
1
2
3
Depth
0 [x10³ yr]
2
4
6
8
10
Age
Org. C
AA (—)
HA x 10 (---)
AA/HA
Ratio
Nitzschia sp.
(broken frust.)



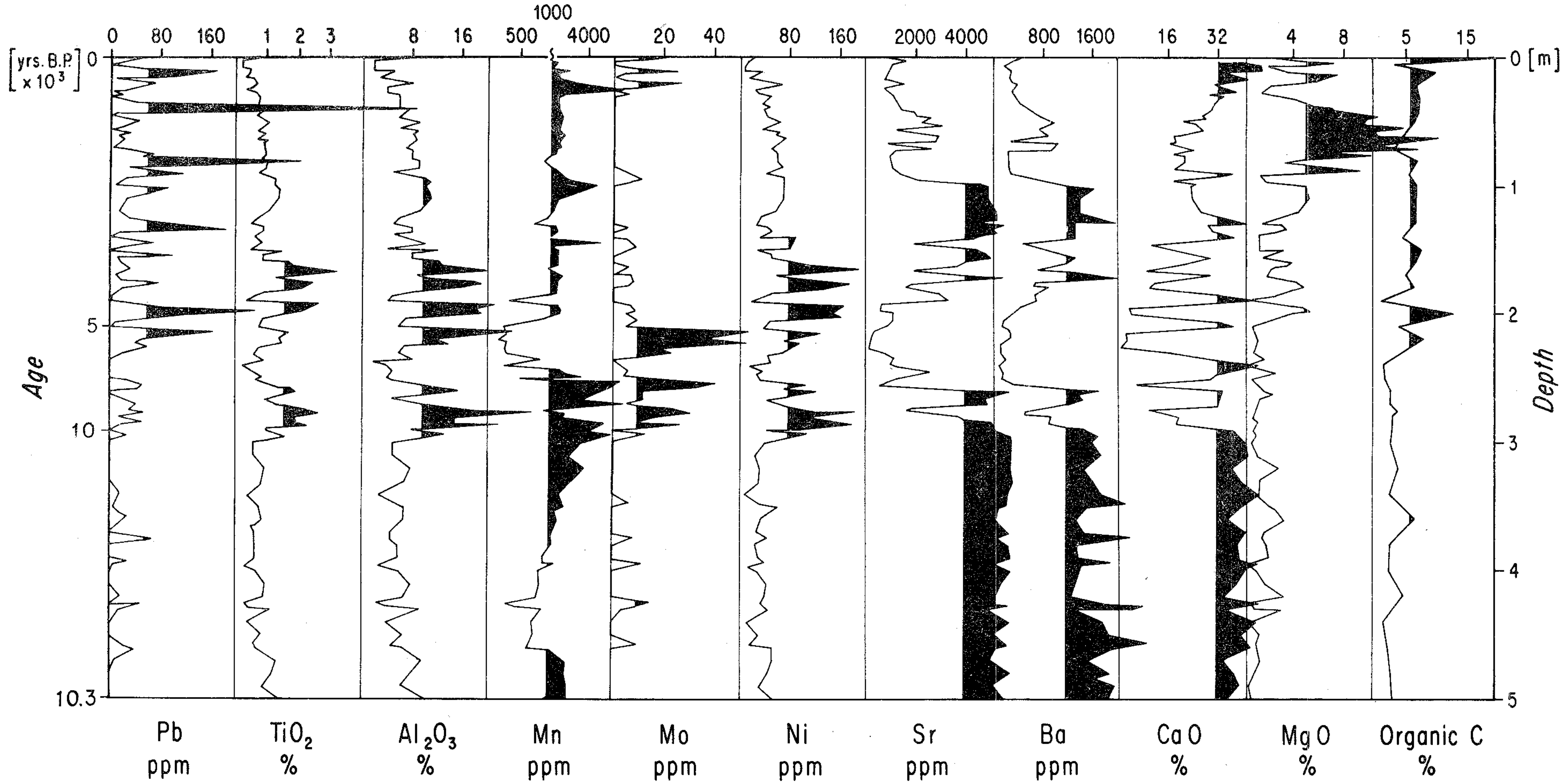
[yrs. B.P. x 10^3]
Age
3
5
8
11
12
Pb ppm
100 200 300 400
TiO$_2$ %
1 2 3
Al$_2$O$_3$ %
10 20 30
Mn ppm
500 1000 1500
Ni ppm
25 50 75
CaO %
10 20 30
Sr ppm
1000 2500 4000
Depth
0 [m]
1
2
3
4

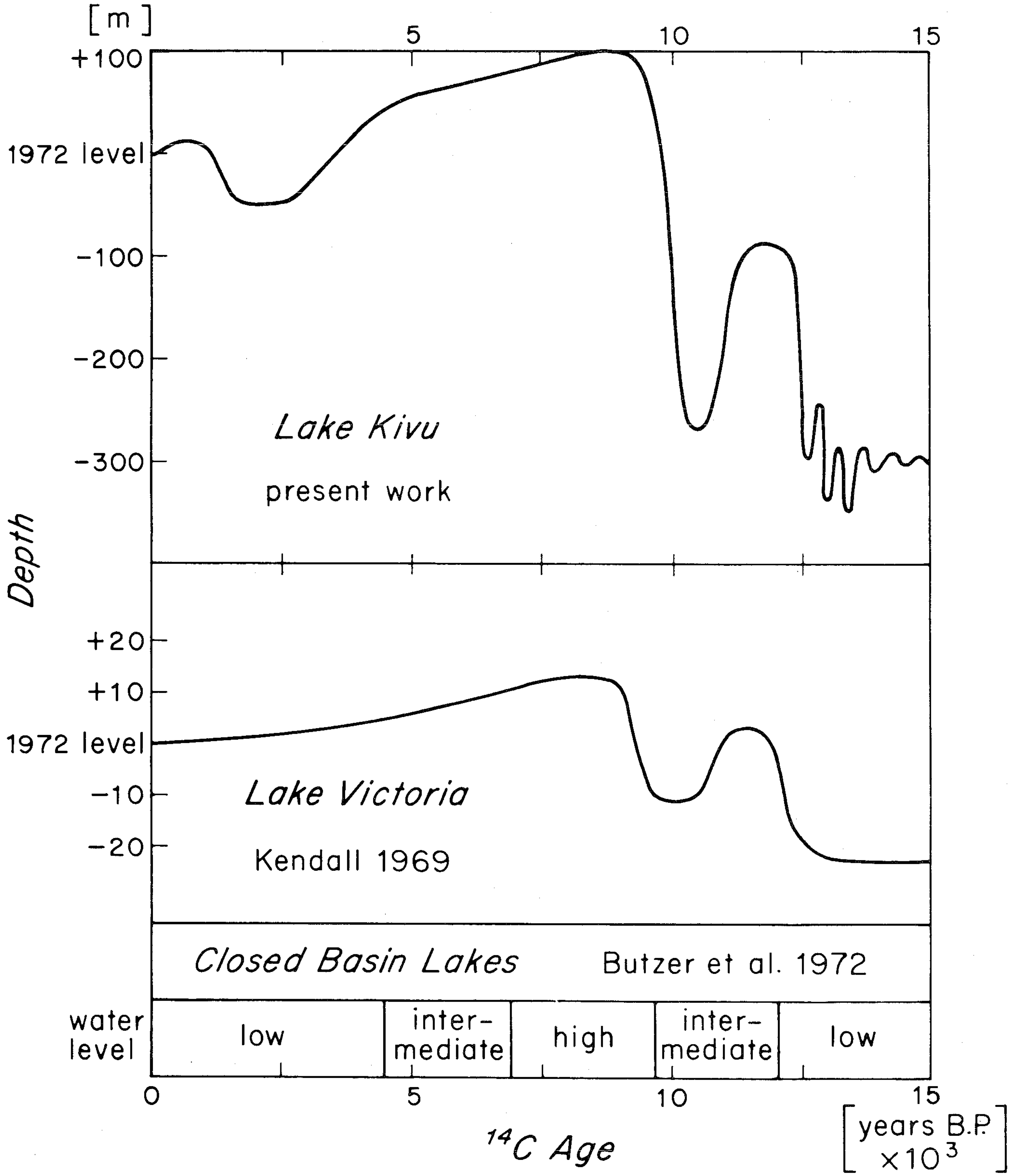


Factor Scores
Age
-2 0 +2 -2 0 +2 -2 0 +2 -2 0 +2 -2 0 +2
[years B.P. x10³]
3
5
7
9
10
11
11.5
12
12.5
13
13.5
13.7
St.10
St. 4
volcanic ash layer
1 2 3 4 5
Factor

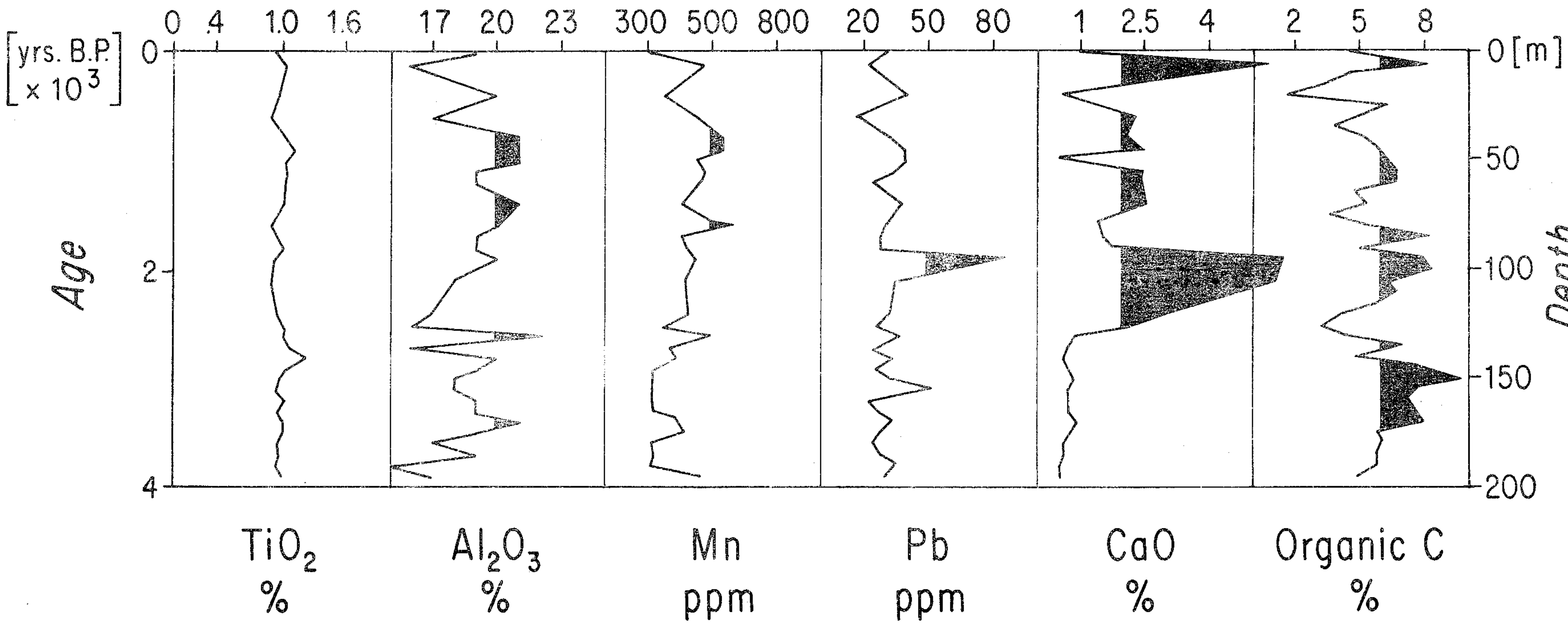


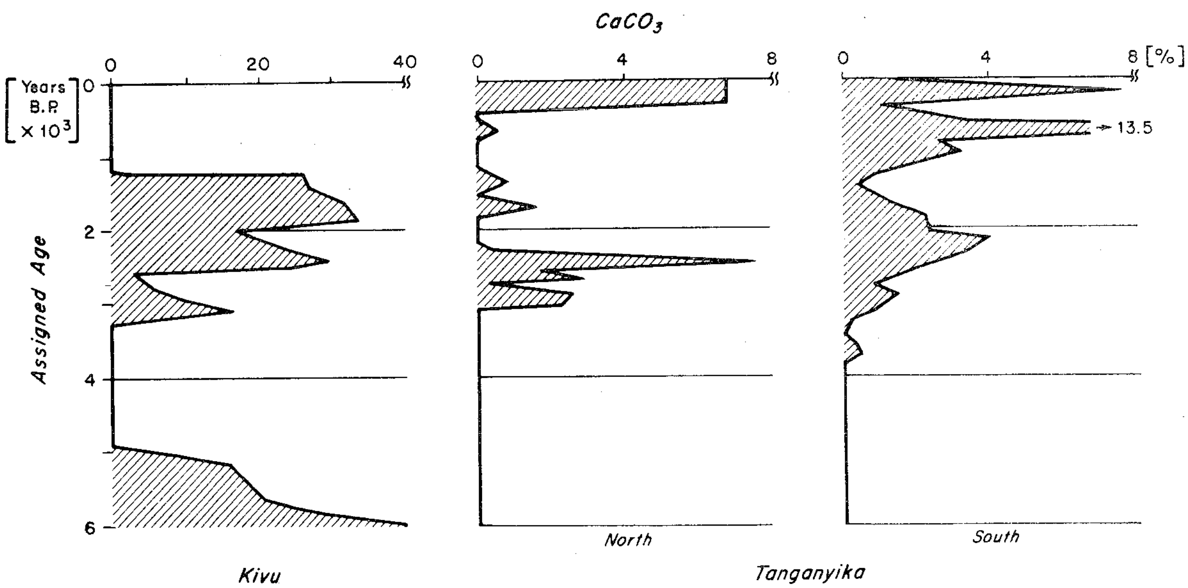
Factor Scores
Age
[years B.P. x 10³]
-2 0 +2
St.13
St.15
3
5
8
11
12
2
3
4
6
8
9
10
1
2
3
4
5
Factor


[yrs. B.P. x 10^3]
Age
0
5
10
10.3
Depth
0 [m]
1
2
3
4
5
Pb ppm
0 80 160
TiO2 %
1 2 3
Al2O3 %
8 16
Mn ppm
500 1000 4000
Mo ppm
20 40
Ni ppm
80 160
Sr ppm
2000 4000
Ba ppm
800 1600
CaO %
16 32
MgO %
4 8
Organic C %
5 15

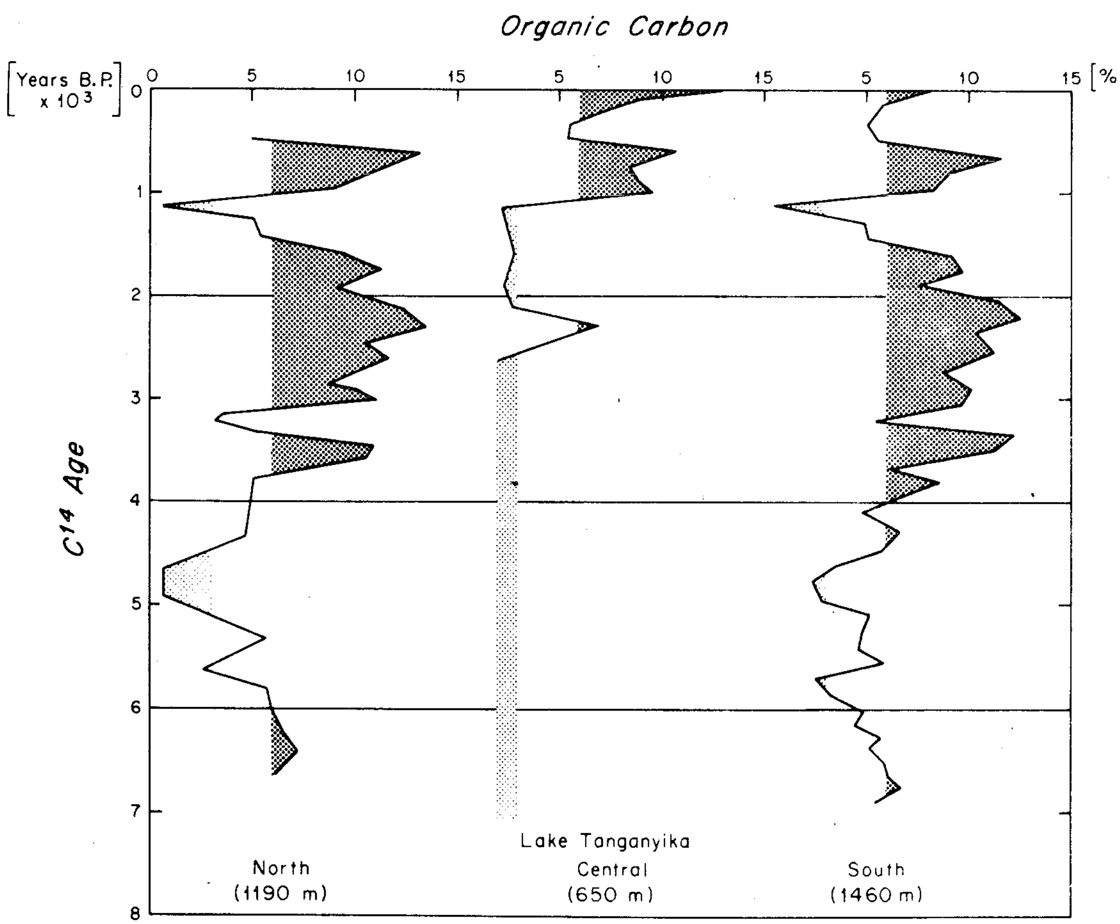


[m]
5
10
15
+100
1972 level
−100
−200
−300
Lake Kivu
present work
Depth
+20
+10
1972 level
−10
−20
Lake Victoria
Kendall 1969
Closed Basin Lakes
Butzer et al. 1972
water level
low
inter-mediate
high
inter-mediate
low
0
5
10
15
^{14}C Age
[years B.P. ×10^3]

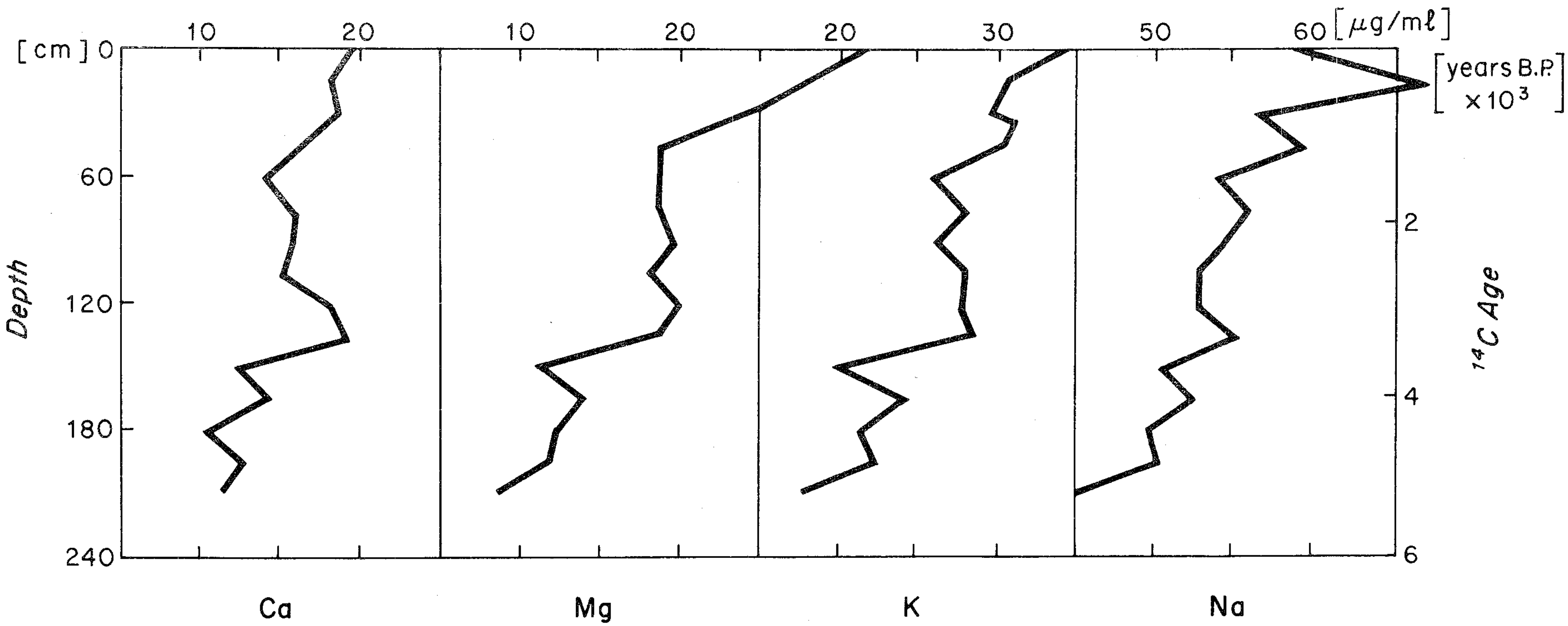

[yrs. B.P. x 10^3]
Age
0
2
4
0 .4 1.0 1.6
17 20 23
300 500 800
20 50 80
1 2.5 4
2 5 8
0 [m]
50
100
150
200
Depth
TiO2 %
Al2O3 %
Mn ppm
Pb ppm
CaO %
Organic C %



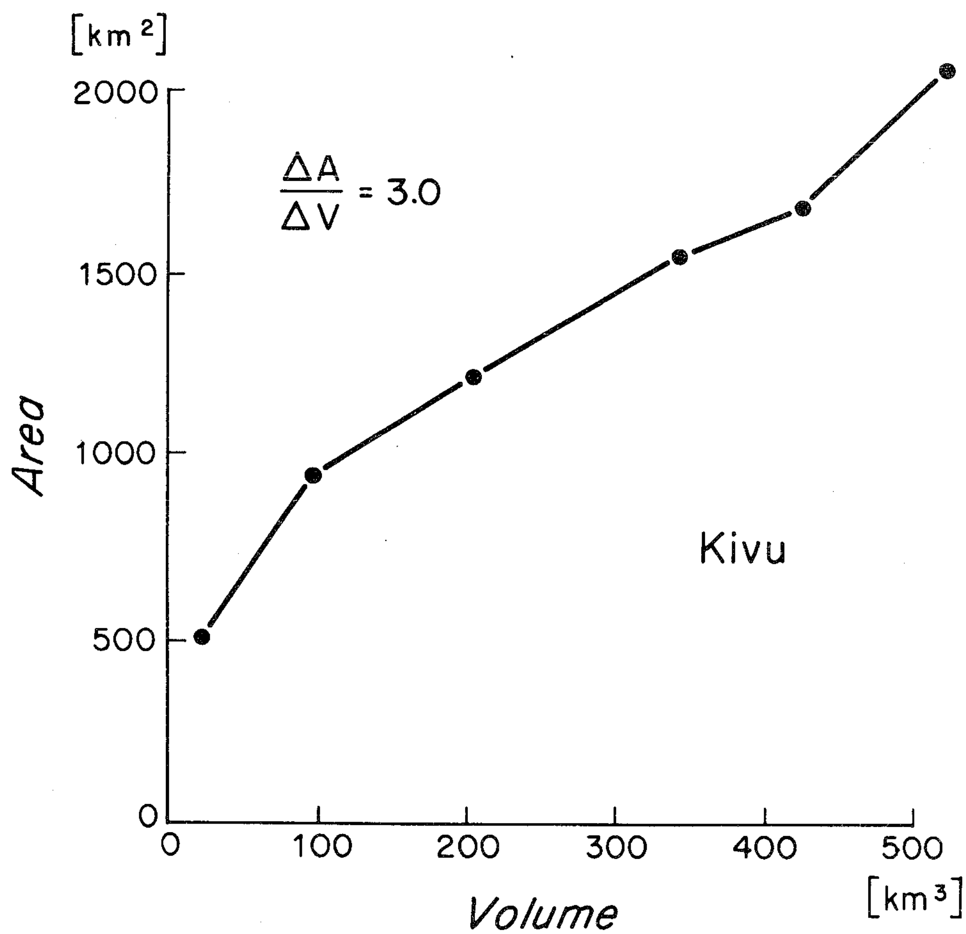
CaCO3
Years B.P. × 10^3
Assigned Age
0
2
4
6
0
20
40
0
4
8
0
4
8 [%]
→ 13.5
North
South
Kivu
Tanganyika



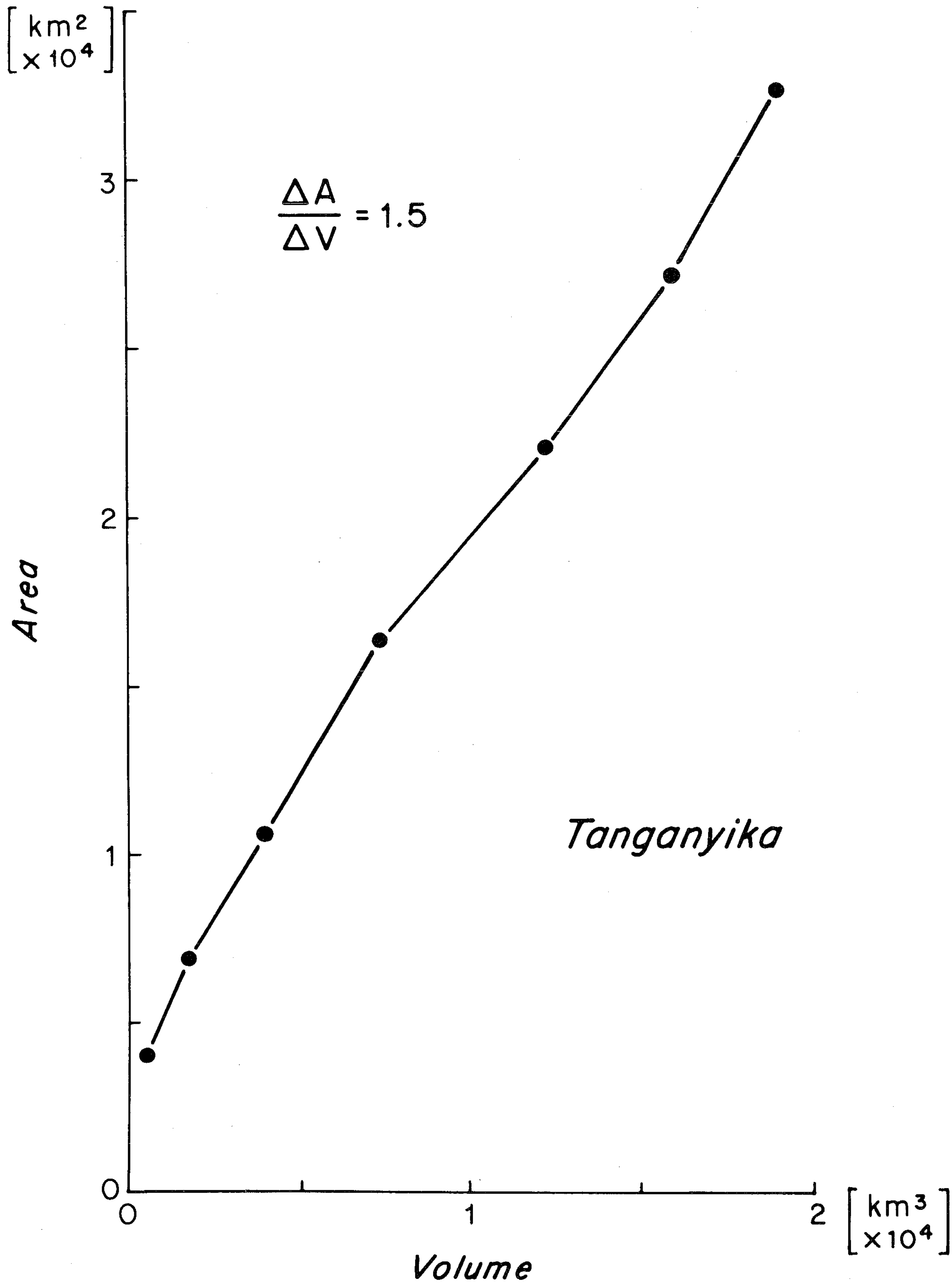
Organic Carbon
[Years B.P. x 10^3]
0
5
10
15
5
10
15
5
10
15 [%]
1
2
3
4
5
6
7
8
C^14 Age
Lake Tanganyika
North
(1190 m)
Central
(650 m)
South
(1460 m)



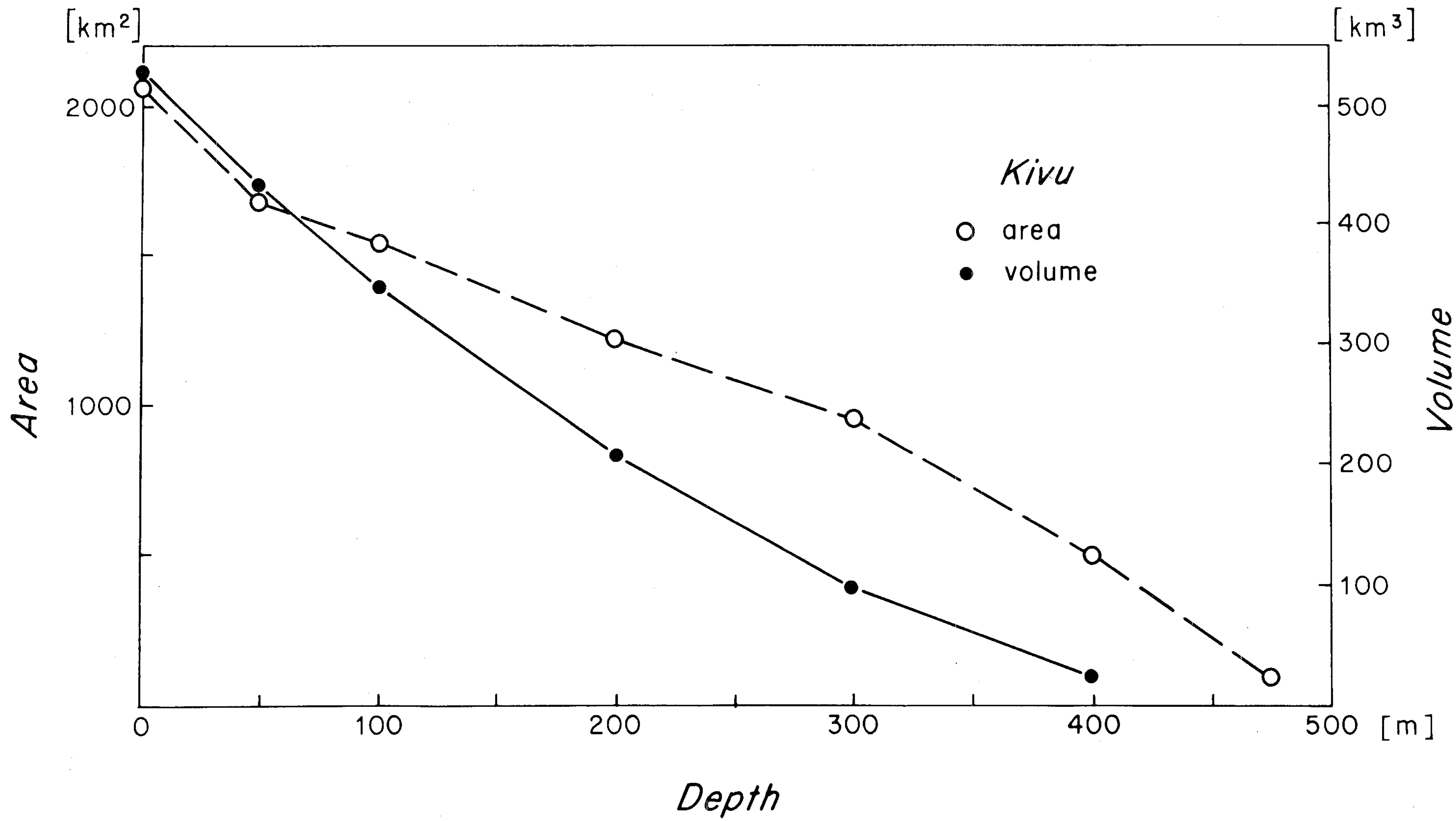
[cm] 0
60
120
180
240
Depth
10
20
10
20
20
30
50
60 [μg/mℓ]
[years B.P. ×10³]
2
4
6
¹⁴C Age
Ca
Mg
K
Na



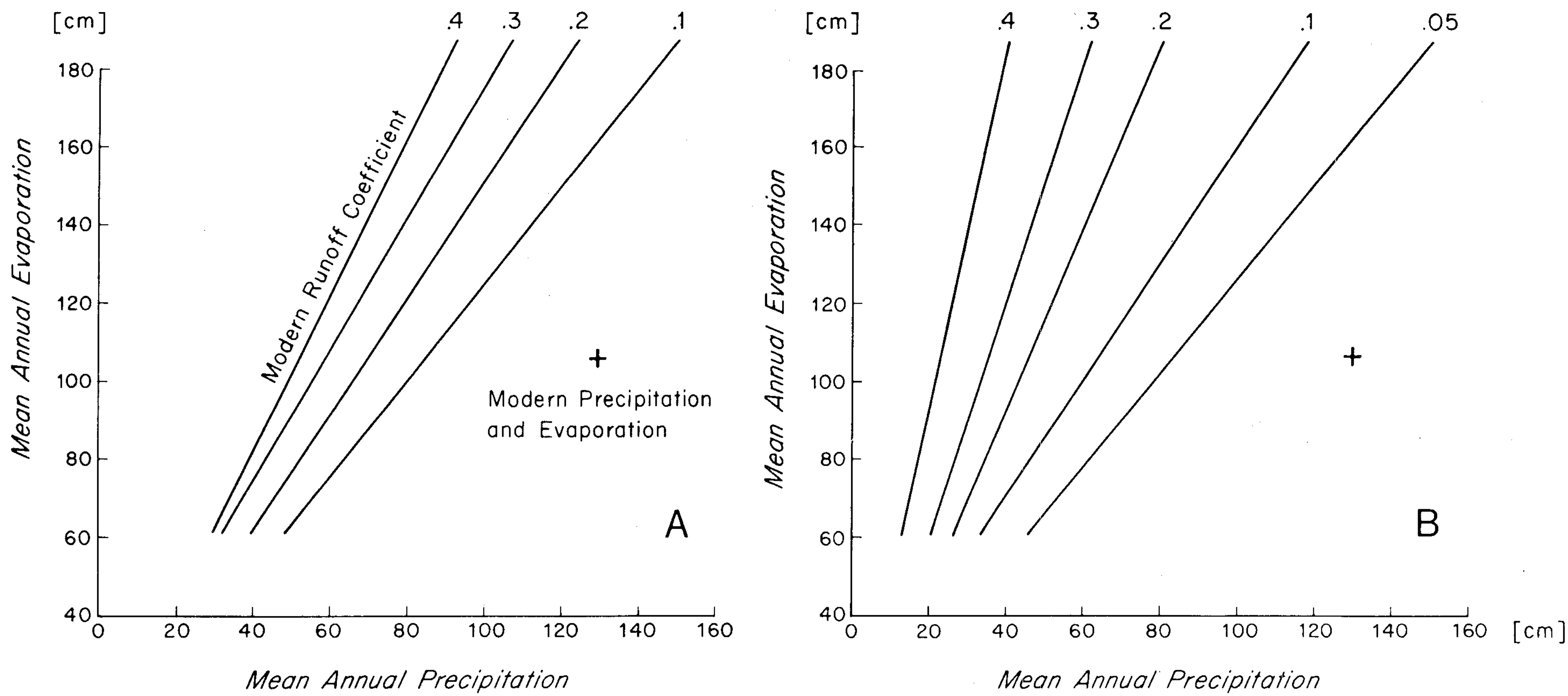
[km²]
2000
1500
1000
500
0
Area
ΔA/ΔV = 3.0
Kivu
0
100
200
300
400
500
[km³]
Volume

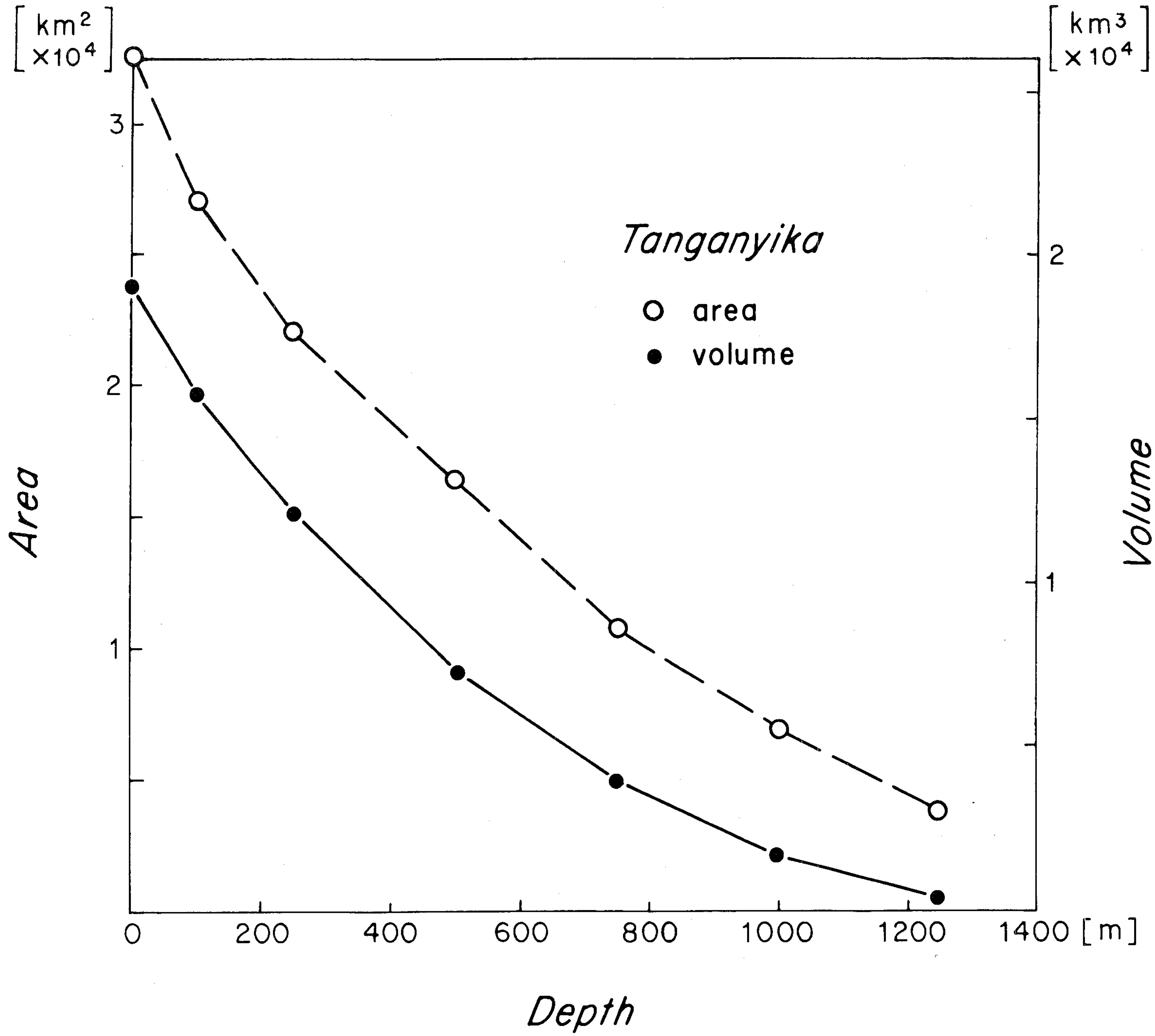


[km²
×10⁴]
3
2
1
0
Area
ΔA/ΔV = 1.5
Tanganyika
0
1
2
[km³
×10⁴]
Volume

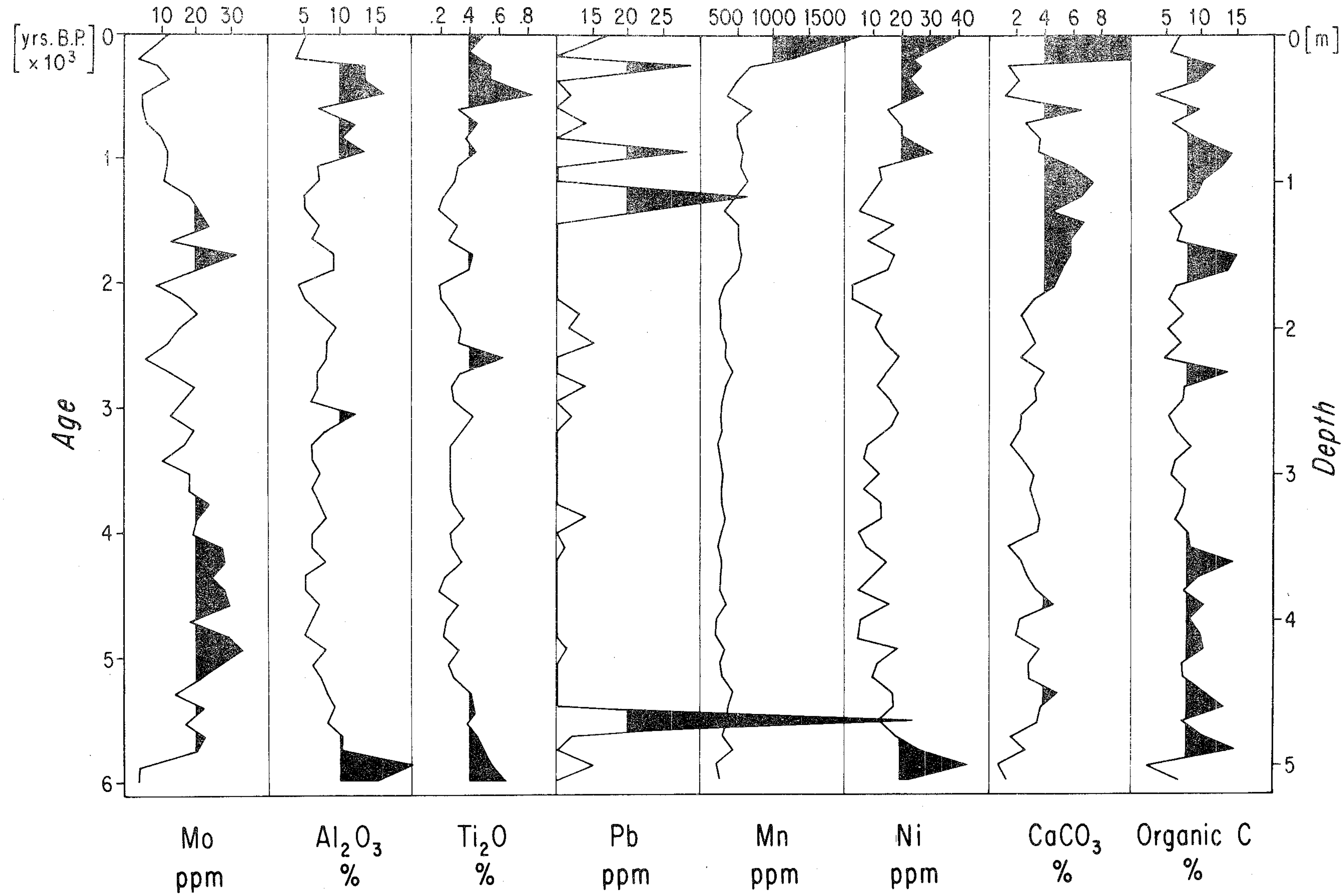


[km²]
[km³]
Area
Volume
Depth
Kivu
area
volume
2000
1000
500
400
300
200
100
0
100
200
300
400
500 [m]

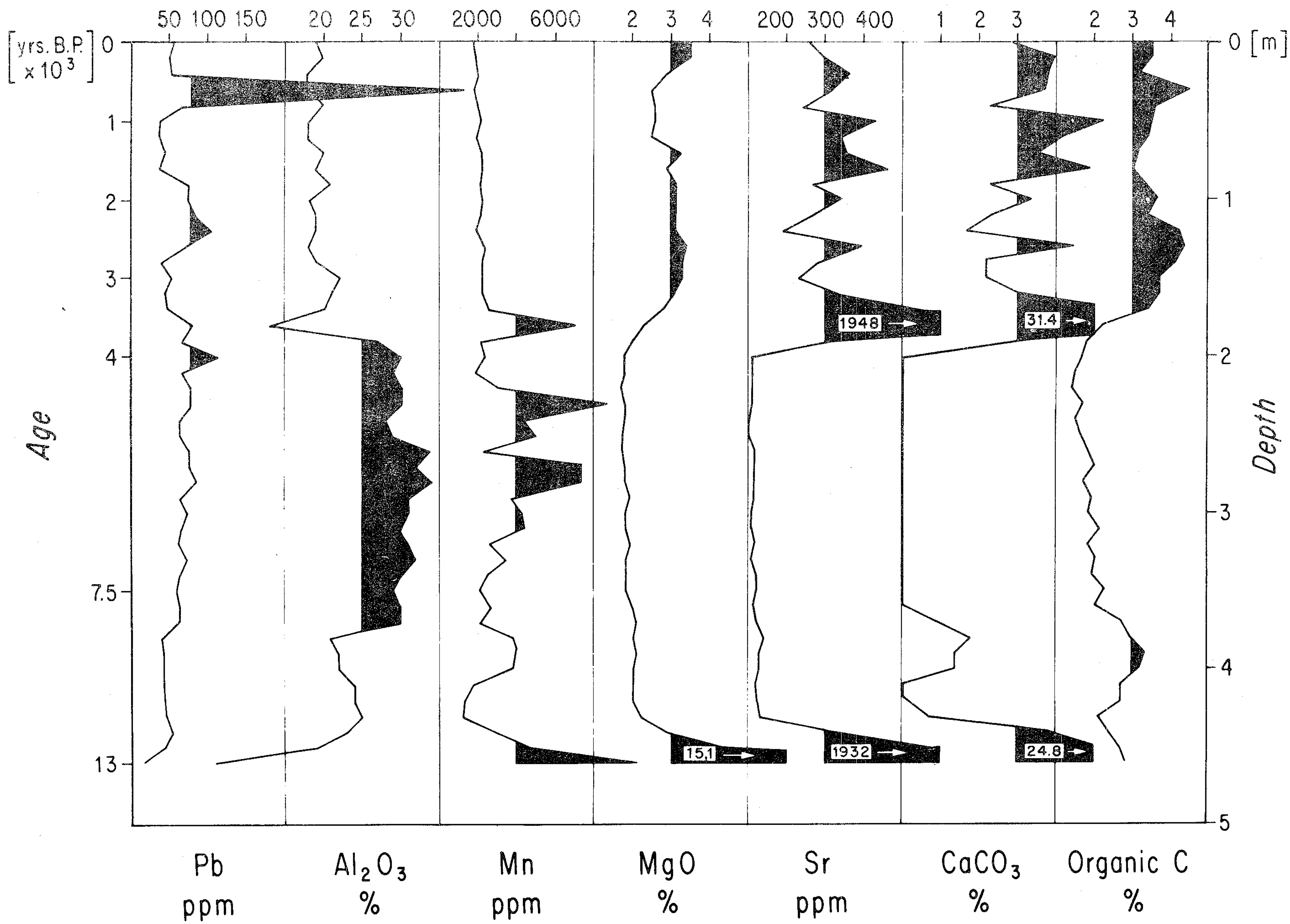

[cm]
.4
.3
.2
.1
Modern Runoff Coefficient
Modern Precipitation
and Evaporation
A
Mean Annual Evaporation
Mean Annual Precipitation
[cm]
.4
.3
.2
.1
.05
B
Mean Annual Evaporation
Mean Annual Precipitation
[cm]



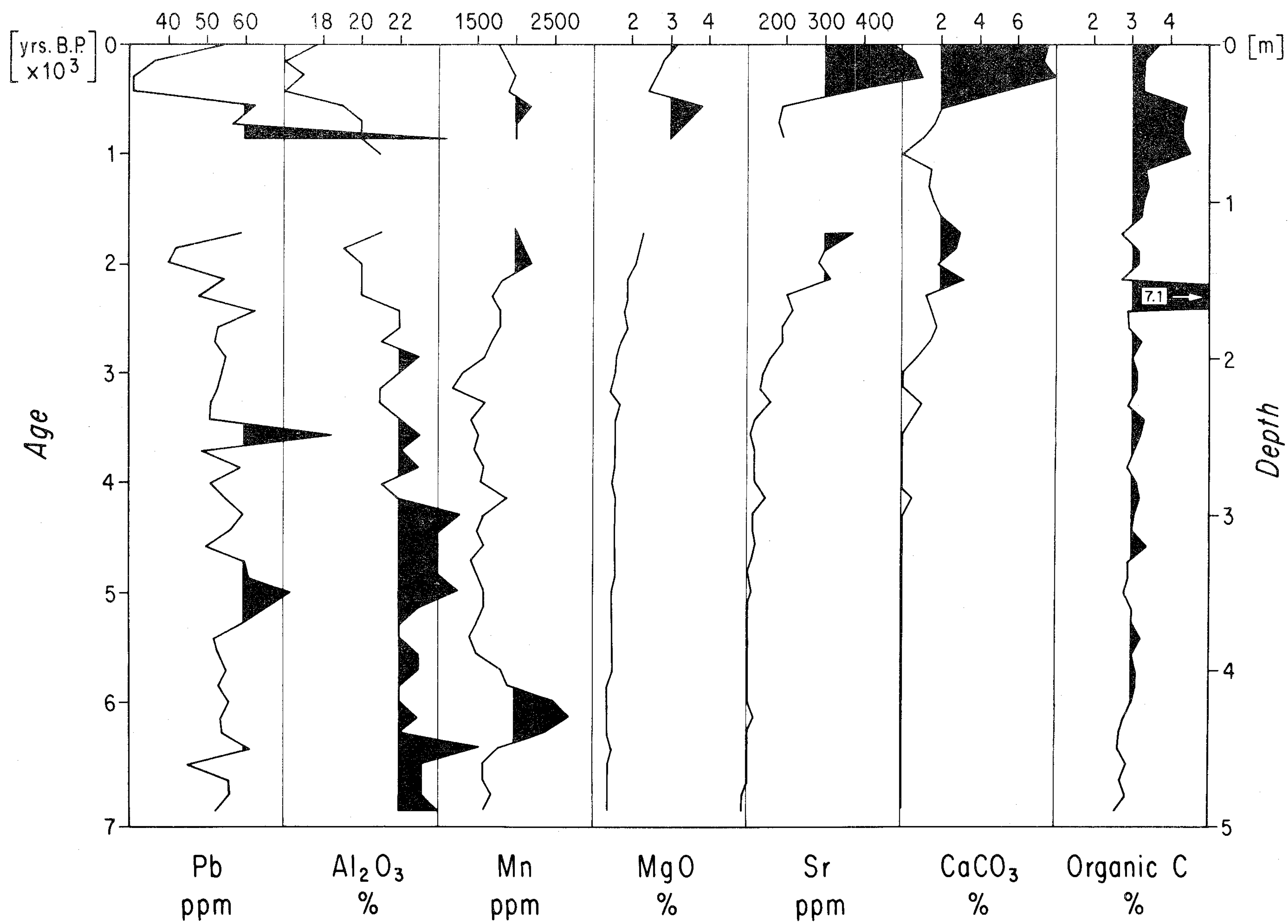
[km2 ×10^4]
[km3 ×10^4]
Area
Volume
Tanganyika
area
volume
0
200
400
600
800
1000
1200
1400 [m]
Depth
3
2
1

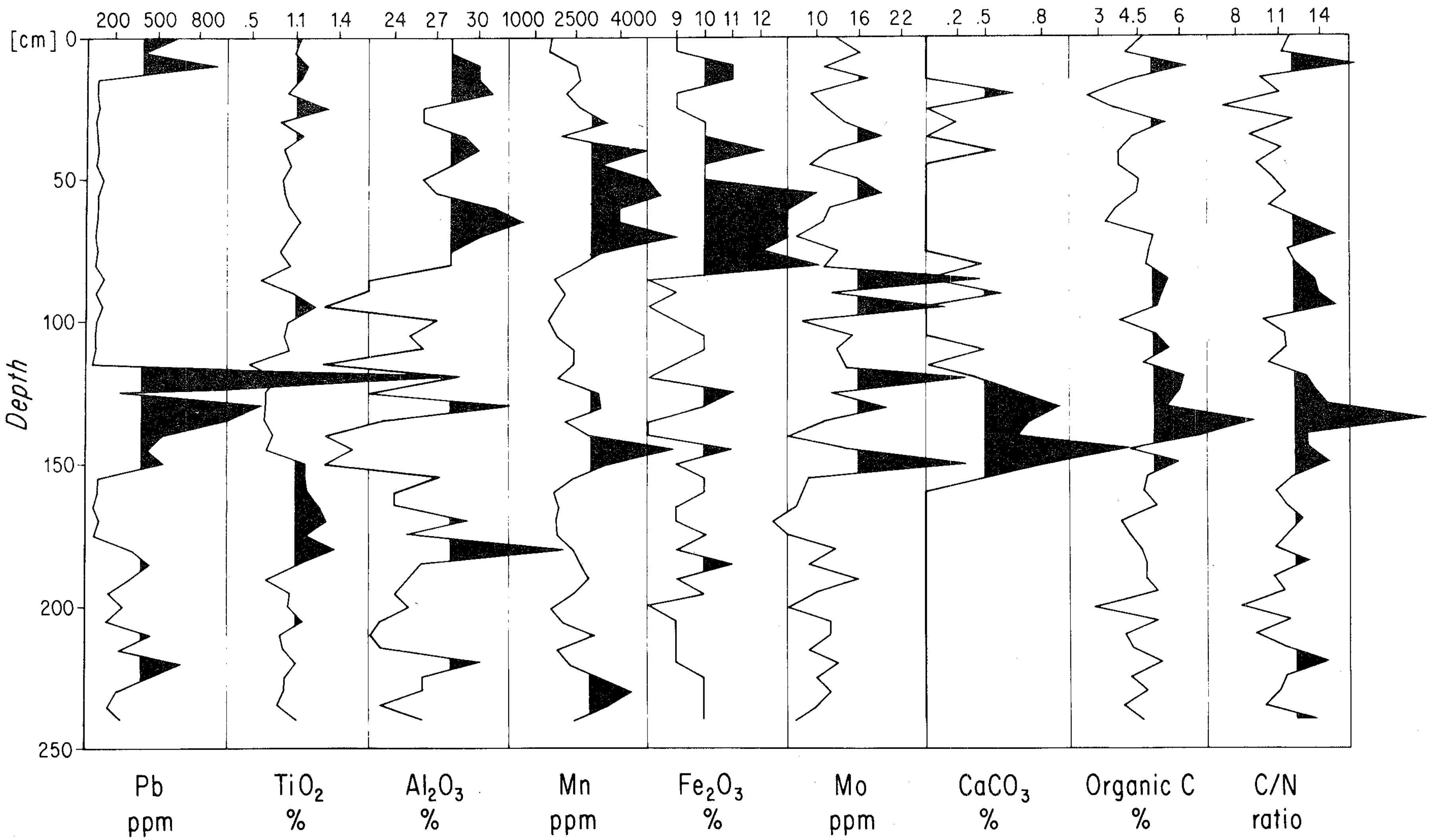


[yrs. B.P. × 10³]
Age
0
1
2
3
4
5
6
10 20 30
5 10 15
.2 .4 .6 .8
15 20 25
500 1000 1500
10 20 30 40
2 4 6 8
5 10 15
0 [m]
1
2
3
4
5
Depth
Mo
ppm
Al₂O₃
%
Ti₂O
%
Pb
ppm
Mn
ppm
Ni
ppm
CaCO₃
%
Organic C
%



[yrs. B.P. x 10³]
Age
0
1
2
3
4
7.5
13
Depth
0 [m]
1
2
3
4
5
50 100 150
20 25 30
2000 6000
2 3 4
200 300 400
1 2 3
2 3 4
1948
31.4
15,1
1932
24.8
Pb ppm
Al₂O₃ %
Mn ppm
MgO %
Sr ppm
CaCO₃ %
Organic C %


Pb ppm
Al2O3 %
Mn ppm
MgO %
Sr ppm
CaCO3 %
Organic C %
Age
[yrs. B.P. x10^3]
Depth
[m]
7.1



200 500 800
.5 1.1 1.4
24 27 30
1000 2500 4000
9 10 11 12
10 16 22
.2 .5 .8
3 4.5 6
8 11 14
[cm] 0
50
100
150
200
250
Depth
Pb
ppm
TiO2
%
Al2O3
%
Mn
ppm
Fe2O3
%
Mo
ppm
CaCO3
%
Organic C
%
C/N
ratio

Table 1

Chemical Composition of Lake and River Waters from Central Africa

	Na	K	Ca	Mg	Cl	SO_4	ΣCO_2	Si	P	$N_{(NH_3)}$	N (NO_3 + NO_2)
				mg/l			(m mole/Kg)			μg/l	
Tanganyika[1] (St. 3; surface)	66.3	34.2	8.2	41.5	21	3.4	5.64	0.034	35	71.4	2.8
Tanganyika[1] (St. 3; 1460 m)	66.3	32.8	5.2	41.0	23	2.5	6.38	8.8	166	28	2.8
Kivu[1] (St.4; surface)	121.6	97.4	4.8	87	55	23.8	12.5	6.5	26	252	27
Kivu[1] (St. 4; 100 m)	192.0	145.0	64.0	147	int.*	25.0	29.3	11.9	601	6818	5.3
Kivu[1] (St. 4; 200 m)	244.6	178.8	83.1	182	int.	166.4	42.0	12.0	1046	18396	3.8
Kivu[1] (St. 4; 350 m)	465.2	315.2	110.9	344	int.	214.0	133.0	29.6	1702	76440	6.7
Kivu[1] (St. 4; 440 m)	487.4	338.0	112.6	417	int.	220.0	121.0	34.3	1754	99470	4.5
Ruzizi River near Bukavu[1] ***	117.0	100.2	4.6	88	n.d.**	n.d.	13.1	0.045	35.2	28	144.2
Ruzizi River at Tanganyika[2]	94.8	63.0	8.4	67.0	23.8	17.8	10.46[4]	9.0	n.d.	n.d.	n.d.
Malagarisi River at Tanganyika[2]	16.4	2.4	12.9	9.1	15.5	2.1	1.55[4]	22.4	n.d.	n.d.	n.d.
Lake Edward[3]	110	90	12.4	47.8	36	43	9.85[4]	6.5	18[5]	n.d.	24
Lake Albert[3]	91	65	9.8	32.1	33	36.5	7.33[4]	0.4-0.9	120-150[5]	n.d.	10-33
Lake Malawi[3]	21	6.4	19.8	4.7	4.3	5.5	2.36[4]	1.1	n.d.	n.d.	n.d.

[1] from Degens et al., 1973
[2] from Beauchamp, 1939
[3] from Talling and Talling, 1965
[4] meq/l
[5] P·PO_4

* int. = interference
** n.d. = not determined
*** determined on supernatant after setting of suspended load

Table 2 Tabulation of Limnological Data

	ALBERT[1]	EDWARD[1]	KIVU[1]	TANGANYIKA[2, 3]	MALAWI[4, 5]
AREA (Km^2)	6,800	2,325	2,055	34,000	30,800
MEAN DEPTH (m)	25	40	240	572	273
MAXIMUM DEPTH (m)	58	117	485	1,470	700
VOLUME (Km^3)	140	90	583	18,900	8,400
TEMPERATURE (EPILIMNION) (^oC)					
WET SEASON	28.5-29	26.8-27.2	24.0-24.5	26-27	27-28
DRY SEASON	27.4-28.2	25.2-26	23.0-23.5	24-25	23-24
TEMPERATURE (HYPOLIMNION) (^oC)					
WET SEASON	26.6-27.2	25.0-25.2	22.5[6]	23.2-23.5	22.1-22.3
DRY SEASON	26.6	24.7-24.8	22.8	23.2-23.5	22.1-22.3
DEPTH OF THERMOCLINE (m)					
WET SEASON	Astatic	15-20	20-30[6]	50-60	50
DRY SEASON	Astatic	40-50	40-50	100-120	200
RANGE	Astatic	25-30	20	50	150
OXYGEN:MAXIMUM DEPTH LIMIT	58	80-117	70	100-200	250
MIXING SEASONS	Dec.-Aug.	Mar.–Apr.; July–Aug.	Feb.–Mar.; June–July	June-Aug.	June-Aug.
TRANSPARENCY (m) (SECCHI DISK)	2-6	1.4-3	3.5-6.0	10-20	10-30
PRODUCTIVITY (NET) mg O_2 $m^{-2}hr^{-1}$	1000[7]	1000[7]	230[8]	Unknown	Unknown
DIATOMS	*Stephanodiscus astraea* *Nitzschia sp.*	*Nitzschia fonticola* *Stephanodiscus damasi*	*Nitzschia spp.*	*Nitzschia spp.*	*Melosira nyasseus* *Stephanodi[scus] astraea*

[1]Verbeke, 1957 [2]Capart, 1952 [3]Coulter, 1963 [4]Beauchamp, 1953 [5]Hutchinson, 1957 [6]50 m depth considered; water temperature increases with depth below 60 m in Kivu

[7]Calculated from Talling, 1965 [8]Calculated from Degens et al., 1973

Table 3

Varimax Factor Matrix of Kivu Sediments (No. of Samples: 381)

ELEMENT	FACT. 1	FACT. 2	FACT. 3	FACT. 4	FACT. 5	COMMUNALITY
Vanadium	0.701	0.505	0.293	-0.018	-0.007	0.834
Molybdenum	0.029	0.641	0.391	-0.451	0.295	0.856
Lead	0.157	0.212	0.142	-0.086	0.882	0.876
Zinc	0.706	0.261	0.467	0.003	0.282	0.864
Copper	0.751	0.180	0.415	0.029	0.228	0.822
Chromium	0.827	-0.225	-0.012	0.097	0.161	0.770
Nickel	0.217	-0.299	0.724	-0.003	0.352	0.784
Cobalt	0.387	-0.303	0.679	-0.015	0.022	0.703
Strontium	-0.222	-0.922	0.154	0.045	0.049	0.927
Baryum	0.114	-0.821	0.175	0.022	-0.119	0.732
Gallium	0.829	0.179	0.223	-0.116	-0.032	0.784
Manganese	0.260	-0.511	-0.003	0.437	-0.240	0.577
Fe_2O_3	0.197	0.379	0.809	0.041	-0.041	0.840
SiO_2	0.218	0.775	0.313	-0.106	0.143	0.778
Al_2O_3	0.945	0.164	0.058	0.005	0.011	0.924
CaO	-0.295	-0.900	-0.011	0.164	-0.026	0.925
MgO	0.044	-0.148	0.044	0.951	-0.017	0.930
TiO_2	0.878	0.007	0.186	0.177	-0.034	0.839
% Variance						
	28.44	24.80	14.05	7.80	6.93	
% Accumulated Variance						
	28.44	53.24	67.29	75.09	82.03	

Table 4

Radiocarbon Dates of East African Rift Sediments

Locality	Core	Sample depth (cm)	Age (yr B.P.)
Lake Kivu	9	330-350	10,670 ± 280
	10	160-180	6,200 ± 215
	4'	0- 84	11,200 ± 165
		195-230	12,300 ± 330
		350-400	12,650 ± 190
		665-700	13,150 ± 330
		742-800	13,560 ± 165
	12'	170-210	11,950 ± 250
	13'	168-205	11,070 ± 160
		340-382	12,130 ± 170
	15'	115-170	9,460 ± 240
		190-250	9,725 ± 165
		425-480	10,170 ± 140
Lake Tanganyika	3	120-140	3,670 ± 120
		235-255	6,350 ± 160
	15	85-105	1,925 ± 100
		175-195	3,750 ± 105
	19	10- 30	1,300 ± 95
		180-200	>28,200
Lake Edward	5	460-505	5,665 ± 100
Lake Albert	3	140-180	2,660 ± 90
	5	450-495	6,910 ± 115
Lake Malawi	NY-19*	0- 8	2,400 ± 200
		46- 53	2,540 ± 140
		86- 94	2,670 ± 200
Lake Green	LV1	~150 m above Kivu water level	13,300 ± 1150
Nyamuragira (volcano)	N745	Charcoal in lava flow	515 ± 120

*(Von Herzen and Vacquier, 1967)

Table 5

Morphometric and Hydrological Parameters

for Lakes Kivu and Tanganyika

		Kivu	Tanganyika (exclusive of Kivu)
Area:	A_b = catchment (Km^2)	7,140	231,000
	A_l = lake (less islands)	2,060	32,600
	A_t = tributary area	5,080	198,400
Depth:	$\bar{Z}$ = mean (m)	240	570
	Z_{max}	485	1,470
Volume:	V (Km^3)	583	18,880
Basin slope = $\Delta V/\Delta A_l$		0.33-0.4	0.66
Discharge:	Volume Km^3/yr	3.2	2.7
	over A_l cm/yr	160	53
Precipitation:	rate (P) cm/yr	130	90
	volume on A_l (Km^3/yr)	2.7	29
	volume on A_t (Km^3/yr)	6.6	180
Runoff:	coefficient (r)	0.4	0.08
	volume Km^3/yr	2.7	14
	over A_l cm/yr	134	44
Air Temperature: °C		19.5	24
Evaporation:	gross rate (Eg) cm/yr	106	135
	volume loss Km^3/yr	2.1	44
	net rate (En=Eg-P) cm/yr	-24	45
	net volume Km^3/yr	-0.5	15

Table 6

Calculations on Water and Salt Budgets of Lakes Kivu and Tanganyika

		Kivu	Tanganyika
Water			
Volume	km^3	517	18,800
Precipitation + Runoff	km^3/yr	5.3	44.0
Residence time for water	yr	110	430
Runoff	km^3/yr	2.7	18
Refill time (runoff alone)	yr	190	1,000
Salts			
Total Na^{+1}	kg	13×10^{10}	12×10^{11}
Na^+ loss at effluent	kg/yr	3.7×10^8	1.7×10^{8}*
Storage time**	yr	350***	7,000
Total K^{+1}	kg	9.5×10^{10}	6.3×10^{11}
K^+ loss at effluent	kg/yr	3.2×10^8	9.2×10^7
Storage time	yr	300	6,800

*Based on calculated Lukuga discharge of 2.7 km^3/yr and composition of river identical to that of lake.

**Storage time assuming losses of salt to sediment and evaporation are negligible.

***Storage time for whole lake; multiple water layers of Kivu will have their own storage times.

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2. Chemical Stratigraphy
3. Rift Valley Lakes
4. Central Africa

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