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# $^{14}C$ and Th/U Dating of Pleistocene and Holocene Stromatolites from East African Paleolakes

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During recent humid episodes, stromatolites were built along paleolake margins, some 60 m above the modern water level of Lakes Natron and Magadi (southern Gregory Rift Valley). Three generations of stromatolites are observed, the more recent ones frequently covering pebbles and boulders eroded from the older ones. The youngest one yielded $^{14}C$ ages ranging from approximately 12,000 to 10,000 yr B.P. Their $\delta^{13}C$ values (≥2.6‰) suggest isotopic equilibrium between the paleolake total inorganic dissolved carbon and the atmospheric $CO_2$, thereby lending credence to the reliability of the $^{14}C$. An initial $^{230}Th/^{232}Th$ ratio in the detrital component was determined by Th/U measurements on the $^{14}C$ dated stromatolites. Using this value a $^{230}Th/^{234}U$ chronology for the older stromatolites was calculated. Ages of ≥240,000 and 135,000 ± 10,000 yr were obtained for the first and second generations, respectively. A humid episode apparently characterized eastern Africa during each glacial–interglacial transition. $^{18}O$ and $^{13}C$ measurements on stromatolites, when compared to values on modern waters and carbonates, provide paleohydrological information. Long residence time of the paleolake waters and less seasonally contrasted regimes are inferred. 

## INTRODUCTION

Lacustrine carbonates may present significant anomalies in their $^{14}C$ content, anomalies which are often referred to as the "hard water effect." This is particularly true in the East African Rift lakes fed by carbonate-rich soda springs [see Eugster (1980) for examples]. Occasionally, the lake total inorganic dissolved carbon (TIDC) may achieve isotopic equilibrium with atmospheric $CO_2$ (e.g., Lake Abhé travertine pipes; Fontes and Pouchan, 1975); in such cases, a coherent $^{14}C$ chronology of paleolake fluctuations may be established. However, more frequently, TIDC does not show complete equilibrium with atmospheric $CO_2$ due to strong inputs of ground water carbon, which usually has an apparent $^{14}C$ age (i.e., $^{14}C$ activity <100% normalized NBS oxalic acid; cf. Broecker and Olson, 1959, 1961). As a consequence, paleolake carbonates yield $^{14}C$ ages which, after isotopic normalization, are older than those obtained, for instance, on terrestrial organic matter. In addition, it would be of value if one would be able to securely interpret the differences in $^{14}C$ ages between nearby paleolakes (Young and Renaut, 1979): do they reflect a real time difference in their respective high stands, due for instance to differential hydrological setting, or, are they purely the effect of distinct $^{14}C$ activities of the lakes' TIDC?

When dealing with carbonates which yield ages beyond the $^{14}C$ method limits (ca. >40,000 yr), the Th/U disequilibrium method may be used. Here again, difficulties arise. As opposed to the marine environment, which is characterized by a very uniform $^{234}U/^{238}U$ ratio (ca. 1.15; Cherdyntsev *et al.*, 1955), continental waters show highly variable uranium activity ratios (usually between 1 and 2). It is therefore difficult to assume that a freshwater carbonate has formed a radioactively closed system since the time it was depos-



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ited. Leaching with preferential removal of $^{234}U$ may have occurred, the subsequent $^{234}U/^{238}U$ ratio will eventually remain higher than 1.1, and no direct evidence of the open system will appear. When the carbonate was originally aragonitic, X-ray checks for the presence of replacement calcite are to some extent indicative of the system's relative stability. When the carbonate was calcitic, the problem appears insoluble, although several considerations may be of help [see Gascoyne *et al.* (1978) for examples of speleothems]. As concerns lacustrine carbonates, detrital particles (clay, sand) are often incorporated into the carbonate network. As a consequence, contamination by nonauthigenic $^{230}Th$ occurs. Since $^{230}Th$ is accompanied by $^{232}Th$, the presence of detrital thorium can be detected. However, $^{230}Th/^{232}Th$ ratios in detrital particles do vary from one site to the other. Only in exceptional cases has it been possible to evaluate this initial ratio (Schwarcz and Skoflek, 1982).

In the example we will discuss here, both the $^{14}C$ and Th/U methods have been applied successfully to the dating of late middle Pleistocene and early Holocene high lake levels in the East African Rift. The Lake Natron (Tanzania)–Lake Magadi (Kenya) basin (Fig. 1) has been filled at least three times in the recent past. Each time, stromatolitic construction occurred at the paleolake shorelines. Algal carbonate constructions have been found in most rift lakes: in Lake Abhé (Republic of Djibouti, Gasse and Rognon, 1973; Gasse, 1975); in Ethiopia, within the Hadar Formation (Taieb, 1975; Hillaire-Marcel *et al.*, 1982), and in Lakes Ziway–Shala (Grove and Dekker, 1976; Street, 1979), Chew Bahir (Grove and Goudie, 1971), and Hayq and Awasa (Grove *et al.*, 1975); in Kenya, in Lake Bogoria (Tiercelin, 1981); and, in Tanzania, in the lake Manyara area (Holdship, 1976). The most detailed studies are those of Vondra *et al.* (1971), Johnson (1974), Johnson and Reynolds (1976), and Abell *et al.* (1982), concerning the Lake Turkana area stromatolites.

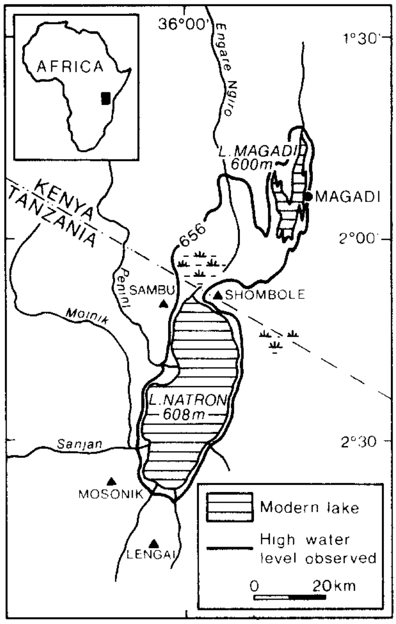


FIG. 1. Map of Lakes Natron and Magadi showing limits of paleolakes.

In the present example, we demonstrate that, with the help of careful geochemical studies of the modern and paleoenvironments, such material is suitable for $^{14}C$ and Th/U dating.

## LAKE NATRON–LAKE MAGADI BASIN

At the south tip of the Gregory Rift valley, NE of the Ngorongoro volcanic complex, the Natron–Magadi basin occupies a depression some 20 km wide and 100 km long, at an altitude of ca. 600 m. Along its western limit, the Nguruman fault escarpment shows successive layers of basaltic lava flows with zeolites, occasionally interrupted by Plio–Pleistocene lacustrine deposits. Several volcanoes surround Lake Natron. To the south, Oldoinyo Lengai shows frequent activity, erupting most recently in 1983. Its lava has a strong carbonatitic component (cf. Hay, 1983), whereas most of the other volcanoes (Gelai, Shombole, Sambu, Mosonik) range from trachytic to phonolitic types (Dawson, 1962).

Today, lakes Natron and Magadi are shallow water bodies with extensive trona crusts (up to 30 m thick; Eugster, 1980;

Jones *et al.*, 1977; Surdam and Eugster, 1976) which have been exploited by the Magadi Soda Company since the First World War. The precipitation–evaporation (potential) balance of the drainage basin is likely to be negative despite high inputs during the rainy season (from 400 mm $yr^{-1}$ at the bottom of the depression, up to 1750 mm $yr^{-1}$ at ca. 3000 m elevation) between November and May. Small pools of water persist throughout the year, caused by hydrothermal springs which have a very high content of sodium carbonate (up to 30 $\times$ $10^3$ ppm of dissolved solids). Lake Natron is also fed by a few perennial streams. Two of the rivers (Peninj and Enwase Ngiro) drain larger basins and have reasonably fresh waters (dissolved solids <200 ppm; cf. Jones *et al.*, 1977).

The depression has been filled by a larger water body on several occasions during late Pliocene and Pleistocene times. Two well-developed, intensively faulted, volcano–sedimentary formations (the Humbu and Moinik formations) are observed on the west bank of Lake Natron (Isaac, 1965). Paleomagnetic measurements indicate that the Olduvai (1.87–1.67 myr) and possibly the Jaramillo (0.97–0.90 myr) events are recorded in these deposits (Thouveny and Taieb, 1984). In the Lake Magadi area, the Oloronga beds reported by Baker (1958) are probably the correlative unit.

High lake levels also existed in more recent times. The latest corresponds to the deposition of the high Magadi beds (Baker, 1958) composed of clay and tuff layers with fossil fish (*Tilapia*) dated at 9120 ± 170 yr B.P. (N-862; Butzer *et al.*, 1972). This deposit has been associated with a "seasonal water level . . . higher and somewhat more stable" (Butzer *et al.*, 1972; p. 1073) than the modern one. Actually, the corresponding paleolake limit is observed at +48 m above the modern Lake Natron level (56 m above Lake Magadi): an almost continuous belt of biogenic calcareous concretions (ranging between 656 and ca. 645 m in altitude), with a few beach-type deposits, marks a continuous paleoshoreline around both lakes. These stromatolitic structures indiscriminately cover most of the bedrock surfaces and lithologies, including fossil logs and tree roots. Through detailed morphologic and petrographic examination of the stromatolites (Casanova, in press), it was possible to establish a precise littoral zonation and furthermore to identify three different phases of encrustation.

Two of these phases are observed at approximately the same elevation and at many sites. The most recent, third generation phase covers pebbles and boulders eroded from the older second generation phase. The oldest, first generation phase has been found *in situ* at a single spot some 80 m above the Lake Natron level, up the Moinik River. It has also been observed, reworked, at lower elevations. Occasionally the second generation phase developed on top of some bioherms eroded from the oldest phase (Fig. 2). Thus, three distinct lake level rises occurred in the recent past. The last two times, the paleolake reached some 2000 $km^2$ in area (vs 1200 $km^2$ at present). A strong erosional phase took place in between these two lacustrine episodes. In the Lake Natron area, the Humbu and Moinik formations, which are largely exposed in the Moinik and Peninj river deltas, have been significantly carved by running waters and possibly by wind action during the lake recession period: fluvial-type paleosurfaces are often covered by a layer of gravel and pebbles that are partly of foreign origin (quartzite from the Precambrian basement, cropping out some 30 km to the east). This paleosurface has been fossilized, in more recent times, by the third and last generation of stromatolites.

A last detail must be mentioned. To the NW of Lake Magadi, the third generation of stromatolites is exposed along the Nguruman fault escarpment at two different elevations (ca. 80 and 60 m above the modern lake level). A vertical movement of the fault seems likely to be responsible for

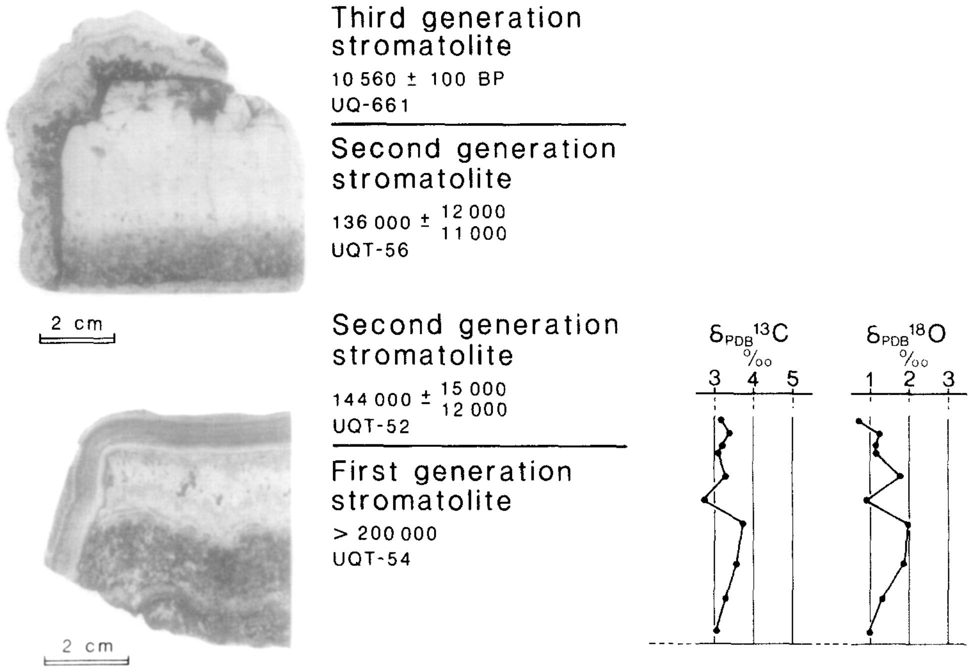


FIG. 2. Polished section showing the succession of the stromatolite generation in "stratigraphic" position, their age, and isotopic composition.

this anomaly because elsewhere a single water level is observed.

## LATE PLEISTOCENE–EARLY HOLOCENE LAKE AND THIRD GENERATION OF STROMATOLITES

Stromatolites formed during this high lake stand are found systematically at the former lake limits, i.e., at the margin of the depression and around paleoislands. They yield uncorrected $^{14}C$ ages ranging from 12,450 ± 100 (UQ-930) to 9650 ± 200 yr B.P. (UQ-907) (Fig. 3). When plotted on a frequency histogram (Fig. 4), three modes appear: a major one ca. 10,300 yr B.P., and two less pronounced ones ca. 11,300 and 9800 yr B.P.

Northwest of Lake Magadi, several spring mounds or travertine pipes (Fig. 5) are still preserved. They were no doubt formed during the same lacustrine episode at spring outlets fringing the margin of the lake, as shown by sedimentary structures (successive, perfectly horizontal layers supported by molds of *Typha*, etc.). $^{14}C$ measurements, from the base to the top of the highest of these pipes (ca. 9 m high), have an erratic distribution (Table 1) and, as a whole, show a clear deficit in $^{14}C$ when compared to the stromatolites (i.e., 21.2% of "modern" carbon $^{14}C$ activity, vs 27.7%). An input of deep groundwater carbon, depleted in $^{14}C$, explains this anomaly. For comparison, the $^{14}C$ activity of natural waters in the modern Lake Magadi area are shown in Figure 6. Hot and warm springs' TIDC usually has less than 10% $^{14}C$ activity. Despite their complex hydrogeological history (Eugster, 1980), one may suspect that most of the spring carbon has a deep origin in relation to the carbonatites which are present in this part of the rift. Very probably, the groundwater and TIDC budgets were different during the

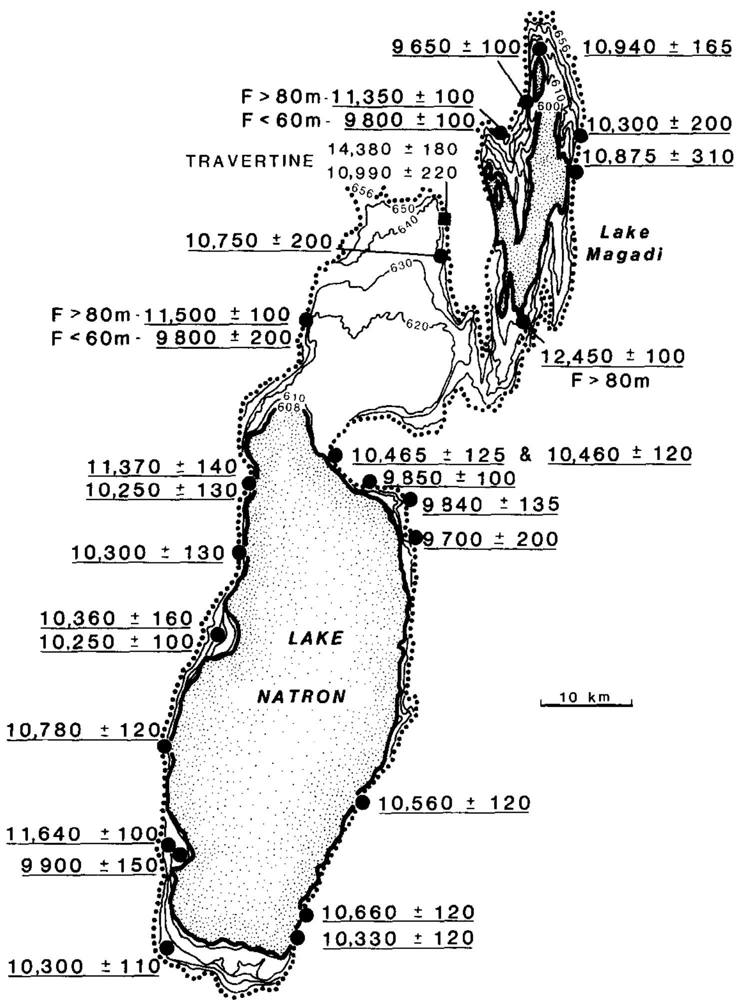


FIG. 3. $^{14}C$ ages of the third generation of stromatolites.

high lacustrine episodes and the paleospring $^{14}C$ content might have been much higher than 10%. However, the latter was certainly lower than 100%, i.e., lower than that of the atmospheric $CO_2$ at that time. The paleolake TIDC may thus be suspected not to have achieved isotopic equilibrium with atmospheric $CO_2$. Consequently, the average $^{14}C$ age of the stromatolites, the carbon of which derives from the lake TIDC, might be nothing more than an "artifact," and the same can be said of the three modes of the frequency histogram of ages.

Stable carbon isotope ratios ($^{13}C/^{12}C$) have to be examined to resolve this question. Modern springwater TIDC in the Lengai vicinity, i.e., the most abundant nearby carbonatite lava occurrences, has a $\delta_{PDB}{}^{13}C$ close to $-4.6‰$ (Fig. 7). Travertine carbonates, nowadays deposited at the walls of the Engare Ngiro Falls (which drains the area), show an average $\delta_{PDB}{}^{13}C$ of ca. $-1.8‰$ and a $\delta_{PDB}{}^{18}O$ of ca. $-5‰$ (Table 2). The respective isotopic compositions of TIDC and water are: $\delta_{PDB}{}^{13}C = -3.6‰$ and $\delta_{SMOW}{}^{18}O = -4.5‰$. These carbonates are therefore deposited in a condition quite close to isotopic equilibrium with the liquid phase and show minor kinetic effects. Their $^{14}C$ activity is 1.4% in relation to the deep origin of TIDC. The

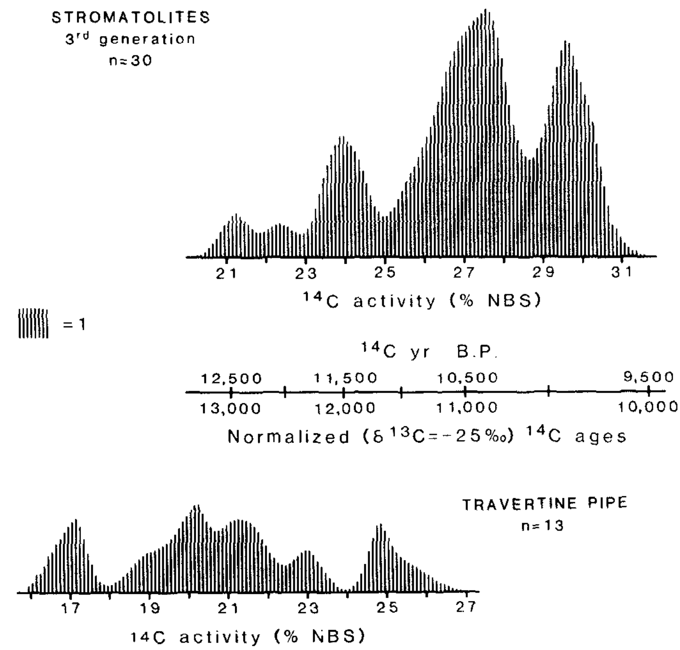


FIG. 4. Frequency histograms of $^{14}C$ ages: (a) stromatolites, (b) travertine pipe.

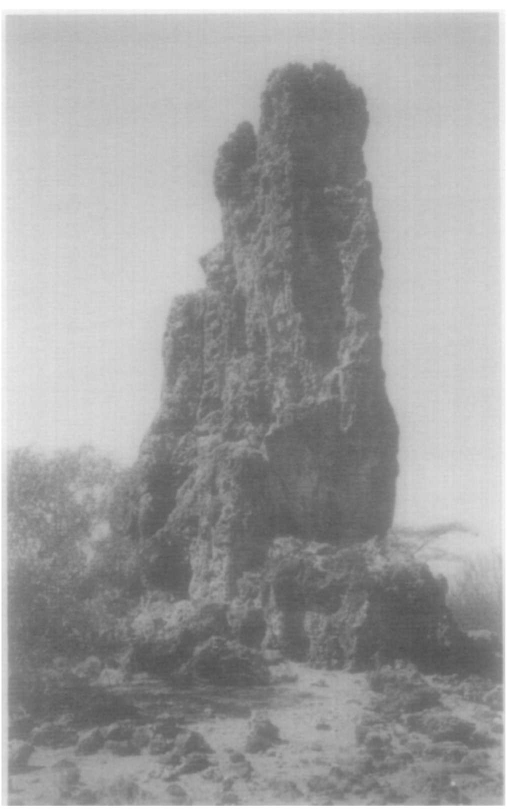

FIG. 5. Example of a travertine pipe NW of Lake Magadi.

paleolake travertine pipe NW of Lake Magadi shows a different carbon budget. Its positive isotopic composition (mean $\delta_{PDB}{}^{13}C$ = +3.3‰) rather suggests that significant exchanges with atmospheric carbon dioxide did occur. The high $^{18}O$ content of the paleowater, which may be derived from the $\delta^{18}O$ values of the paloelake carbonates (Table 2), also suggests a long residence time for the water, i.e., the possibility of long-term exchanges between TIDC and atmospheric $CO_2$. Despite the high amount of dissolved inorganic carbon in modern waters—and the drastically different hydrological budget—there is a rapid downriver shift in TIDC $\delta^{13}C$ (Fig. 7), indicating efficient exchanges with atmospheric $CO_2$. No stromatolite $\delta^{13}C$ value is lower than +2.6‰. That is, none is lower than the isotopic composition of a calcite deposited in equilibrium with atmospheric $CO_2$ ($\delta^{13}C$ ca. −7‰; cf. Craig and Keeling, 1963) at a temperature of ca. 25°C, which is certainly plausible for such a rift lake. We are tempted, therefore, to assume that the paleolake TIDC had probably reached equilibrium with the existing atmospheric

TABLE 1. ISOTOPIC COMPOSITION OF STROMATOLITES AND TRAVERTINE

| Laboratory number | Field number | Sample | $\delta_{PDB}{}^{18}O$ | $\delta_{PDB}{}^{13}C$ | $^{14}C$ Activity % ref. | Age ($^{14}C$ yr B.P.) | Comments |
|---|---|---|---|---|---|---|---|
| | | Stromatolites (third generation) | | | | | |
| UQ-636 | LN-8-10-3 | Stromatolite | 2.7 | 4.2 | 26.9 ± 0.4 | 11,950 ± 115 | |
| UQ-627 | LN-13-10-8 | Stromatolite | — | — | 27.6 ± 0.4 | 10,300 ± 115 | |
| UQ-618 | LN-6-10-7 | Stromatolite | 3.7 | 5.8 | 24.3 ± 0.4 | 11,375 ± 140 | |
| UQ-614 | LN-7-10-4 | Stromatolite | 2.0 | 3.0 | 27.9 ± 0.4 | 10.250 ± 110 | cf. UQT-78 |
| UQ-587 | LN-18-10-11 | Stromatolite | 2.2 | 3.2 | 23.5 ± 0.3 | 11,640 ± 100 | |
| UQ-564 | LN-3-10-1 | Stromatolite | 3.3 | 3.9 | 27.5 ± 0.5 | 10,360 ± 160 | |
| UQ-951 | LM-28a | Stromatolite | 3.0 | 3.7 | 29.3 ± 0.5 | 9,860 ± 150 | Base stromatolite |
| UQ-953 | LM-28b | Stromatolite | 3.0 | 4.2 | 29.9 ± 0.3 | 9,710 ± 100 | Top stromatolite |
| UQ-959 | LM-24 | Stromatolite | — | — | 36.5 ± 0.7 | 8,100 ± 200 | Regressive shoreline |
| UQ-917 | LM-42 | Stromatolite | 3.0 | 4.2 | 23.9 ± 0.3 | 11,500 ± 100 | |
| UQ-920 | LM-46 | Stromatolite | 2.9 | 3.4 | 34.7 ± 0.4 | 8,500 ± 100 | Regressive shoreline |
| UQ-943 | LM-28 | Stromatolite | 2.9 | 4.0 | 28.5 ± 0.5 | 10,090 ± 140 | |
| UQ-982 | MAG-82-D2 | Stromatolite | 4.4 | 3.8 | 37.5 ± 2.0 | 7,880 ± 450 | Regressive shoreline |
| UQ-877 | LM-18 | Stromatolite | −0.1 | 4.0 | 27.2 ± 0.4 | 10,460 ± 120 | cf. UQT-112 |
| UQ-911 | LM-37 | Stromatolite | 2.9 | 3.9 | 26.2 ± 0.5 | 10,750 ± 200 | |
| UQ-910 | LM-31 | Stromatolite | 3.1 | 4.2 | 29.5 ± 0.4 | 9,800 ± 100 | |
| UQ-909 | LM-30 | Stromatolite | 2.7 | 4.1 | 24.3 ± 0.4 | 11,350 ± 200 | |
| UQ-907 | LM-21 | Stromatolite | 2.4 | 3.8 | 30.2 ± 0.8 | 9,650 ± 200 | |
| UQ-904 | LM-14 | Stromatolite | 2.3 | 3.7 | 29.9 ± 0.9 | 9,700 ± 200 | |
| UQ-899 | LM-36 | Stromatolite | 3.0 | 4.2 | 27.8 ± 0.6 | 10,300 ± 200 | |
| UQ-894 | MAG-83-53 | Stromatolite | 3.1 | 4.0 | 29.4 ± 0.5 | 9,850 ± 100 | cf. UQT-130 |
| UQ-888 | LM-47 | Stromatolite | 2.8 | 3.6 | 25.6 ± 0.5 | 10,950 ± 200 | cf. UQT-111 |
| UQ-886 | MAG-83-44 | Stromatolite | 3.5 | 4.3 | 25.8 ± 1.0 | 10,850 ± 300 | cf. UQT-128 |
| UQ-938 | LM-24 | Stromatolite | 3.2 | 4.3 | 41.7 ± 0.5 | 7,000 ± 100 | Regressive shoreline |
| UQ-930 | LM-45 | Stromatolite | 2.8 | 4.2 | 21.2 ± 0.4 | 12,450 ± 100 | |
| UQ-932 | LM-43 | Stromatolite | 3.2 | 4.2 | 29.6 ± 0.6 | 9,800 ± 200 | |
| UQ-933 | LM-17 | Stromatolite | 2.8 | 4.1 | 26.6 ± 0.5 | 10,650 ± 100 | |
| UQ-936 | LM-27 | Stromatolite | 2.8 | 3.2 | 30.2 ± 0.3 | 9,650 ± 100 | |
| UQ-767 | LN-8-10-4 | Stromatolite | 2.6 | 4.1 | 22.4 ± 0.5 | 12,025 ± 175 | cf. UQT-57 |
| UQ-673 | LN-18-10-10 | Stromatolite | 2.9 | 3.8 | 29.1 ± 0.5 | 9,900 ± 150 | cf. UQT-53 |
| UQ-669 | LN-13-10-9 | Stromatolite | 1.6 | 3.3 | 26.5 ± 0.4 | 10,660 ± 115 | |
| UQ-661 | LN-14-10-11 | Stromatolite | 3.2 | 4.4 | 26.9 ± 0.4 | 10,560 ± 125 | cf. UQT-55 |
| UQ-659 | LN-16-10-1 | Stromatolite | 2.8 | 3.6 | 27.7 ± 0.4 | 10,300 ± 105 | cf. UQT-76 |
| UQ-652 | NA-82-40 | Stromatolite | 2.6 | 3.9 | 27.7 ± 0.5 | 10,300 ± 130 | |

TABLE 1—*Continued*

| Laboratory number | Field number | Sample | $\delta_{PDB}{}^{18}O$ | $\delta_{PDB}{}^{13}C$ | $^{14}C$ Activity % ref. | Age ($^{14}C$ yr B.P.) | Comments |
|---|---|---|---|---|---|---|---|
| | | Other relevant measurements | | | | | |
| | NA-82-67 | Travertine at falls | −5.0 | −2.8 | 1.4 ± 0.05 | | Modern deposition |
| UQ-671 | LN-14-10-10 | Travertine at spring outlet | −2.9 | 4.9 | 1.4 ± 0.05 | | |
| UQ-870 | LM-7 | Travertine pipe | 2.5 | 3.3 | 18.9 ± 0.5 | 13,390 ± 220 | |
| UQ-867 | LM-9 | Travertine pipe | 2.1 | 3.2 | 20.9 ± 0.5 | 12,560 ± 200 | Top |
| UQ-858 | LM-6 | Travertine pipe | 2.3 | 3.0 | 25.5 ± 0.7 | 10,985 ± 215 | |
| UQ-857 | LM-5 | Travertine pipe | 2.8 | 3.6 | 21.7 ± 0.5 | 12,290 ± 195 | |
| UQ-851 | LM-3 | Travertine pipe | 3.0 | 3.5 | 23.0 ± 0.4 | 11,810 ± 150 | |
| UQ-846 | LM-2b | Travertine pipe | 2.2 | 3.7 | 21.3 ± 0.6 | 12,420 ± 240 | |
| UQ-841 | LM-8a | Travertine pipe | 2.4 | 3.1 | 16.7 ± 0.4 | 14,380 ± 180 | |
| UQ-838 | LM-2a | Travertine pipe | 3.0 | 3.6 | 19.7 ± 0.4 | 13,060 ± 175 | |
| UQ-832 | LM-8b | Travertine pipe | — | — | 17.2 ± 0.3 | 14,150 ± 160 | |
| UQ-836 | LM-4 | Travertine pipe | 3.1 | 3.4 | 24.8 ± 0.3 | 11,190 ± 115 | |
| UQ-835 | LM-1 | Travertine pipe | 3.0 | 3.5 | 20.2 ± 0.3 | 12,845 ± 130 | Base |

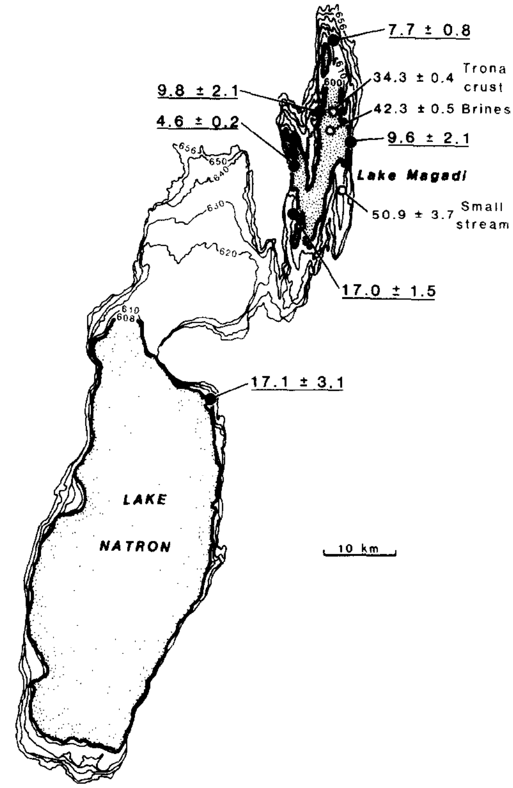


FIG. 6. $^{14}C$ activity (% NBS) (vs reference value) of Lake Magadi springs and brines.

carbon dioxide. The high $^{13}C$ content of some stromatolites (up to +7‰) may be explained by the preferential use of light carbon ($^{12}CO_2$) by algae for photosynthetic purposes, which induced enrichment in $^{13}C$ of the immediately surrounding TIDC, thus causing the precipitation of calcite laminae with high $\delta^{13}C$ values (Hillaire-Marcel and Casanova, in press).

This peculiarity in stromatolite construction is also evidenced by the detailed measurements of isotopic changes in laminae (Fig. 2). As Abell *et al.* (1982) have observed, parallel shifts often occur in $\delta_{PDB}{}^{18}O$ and $\delta_{PDB}{}^{13}C$ values. Changes in total lake alkalinity and pH were considered by Abell *et al.* (1982) to be controlling factors. We do not agree, because groundwater imputs ($HCO_3^-$ and $CO_3^{2-}$) are characterized by relatively low $\delta_{PDB}{}^{13}C$ values. Therefore, depending on temperature, the highest $\delta_{PDB}{}^{13}C$ value one may expect for a

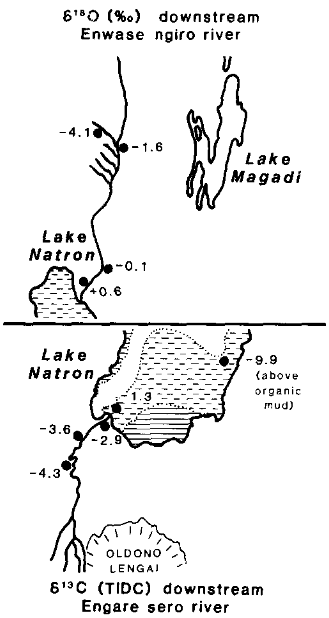


FIG. 7. $^{13}C$ enrichment downstream Engare Sero River.

solid carbonate, after equilibration of TIDC with atmospheric $CO_2$, remains in the +2 +3‰ range. Nor are water pH and the ionic forms of carbon to be considered, because of the successive isotopic equilibria which exist between inorganic carbon species. Moreover, the blue-green algae responsible for lacustrine stromatolite construction are not expected to develop in brackish waters; their presence rather suggests very moderately mineralized water (Casanova, 1981; Casanova and Lafont, 1985).

The parallel shifts of $\delta_{PDB}\,^{18}O$ and $\delta_{PDB}{}^{13}C$ values are better explained by biogenic effects, which may eventually be coupled with temperature changes. When temperature increases, higher evaporation rates will induce $^{18}O$ enrichment of lake water and higher photosynthetic activity will increase the relative enrichment in heavy carbon of the surrounding TIDC. Therefore, despite slightly lower isotopic fractionation factors between ($H_2O$ + TIDC) and calcite, an enrichment in heavy isotopes of the latter has to be expected (cf. Hillaire-Marcel and Casanova, in press).

It is also worth noting that there are no drastic shifts in $^{18}O$ content of laminae that would be associated with strong seasonal climatic differences. Similarly, micrographic examinations of Lake Natron and Lake Magadi stromatolites show no trace of the "couplets" characteristic of strong seasonal contrasts. Therefore, we postulate, for the high lacustrine episode of Lakes Natron and Magadi, a less seasonally contrasted regime than the modern one (cf. Vincens and Casanova, in press).

Figures 8 and 9 illustrate the relationship between the $^{13}C$ and $^{14}C$ contents and the $^{14}C$ and $^{18}O$ contents, respectively. They suggest a long residence time for the paleolake water and, consequently, a high probability for almost complete equilibrium between the lake TIDC and the atmospheric $CO_2$ (Fig. 10). As a result, we attribute a high degree of credibility to the $^{14}C$ activi-

TABLE 2. ISOTOPIC COMPOSITION OF MODERN WATERS (SPRINGS, BRINES, . . .)

| Laboratory number | Field number | Sample | $\delta_{SMOW}{}^{18}O$ | $\delta_{PDB}{}^{13}C$ | $^{14}C$ Activity % ref. | Comments |
|---|---|---|---|---|---|---|
| UQ-801 | LM-51 | TIDC/Water | −1.7 | 0.7 | 9.6 ± 2.1 | Fisch springs L. Magadi |
| UQ-800 | LM-49 | TIDC/Water | −1.3 | 0.4 | 7.7 ± 0.8 | Hot spring Little Magadi |
| UQ-796 | LM-40 | TIDC/Water | −2.8 | — | 17.0 ± 1.5 | Spring South L. Magadi |
| UQ-795 | LM-23 | TIDC/Water | −1.5 | −0.2 | 9.8 ± 2.1 | Spring Dredge 3, L. Magadi |
| UQ-798 | LM-44 | TIDC/Water | −1.4 | −0.9 | 50.9 ± 3.7 | Stream S.E., L. Magadi |
| UQ-817 | LM-25 | Trona Crust | 1.2 | 0.0 | 34.3 ± 0.4 | Dredge 3, L. Magadi |
| UQ-790 | LM-16 | TIDC/Water | −2.4 | −0.5 | 17.1 ± 3.1 | Spring, L. Magadi |
| UQ-793 | LM-26 | TIDC/Water | −3.5 | −5.4 | 4.6 ± 0.2 | Spring, L. Magadi |
| UQ-791 | LM-22 | TIDC/Water | 8 | 1.0 | 42.3 ± 0.5 | Interstitial brines, Dredge 3 |

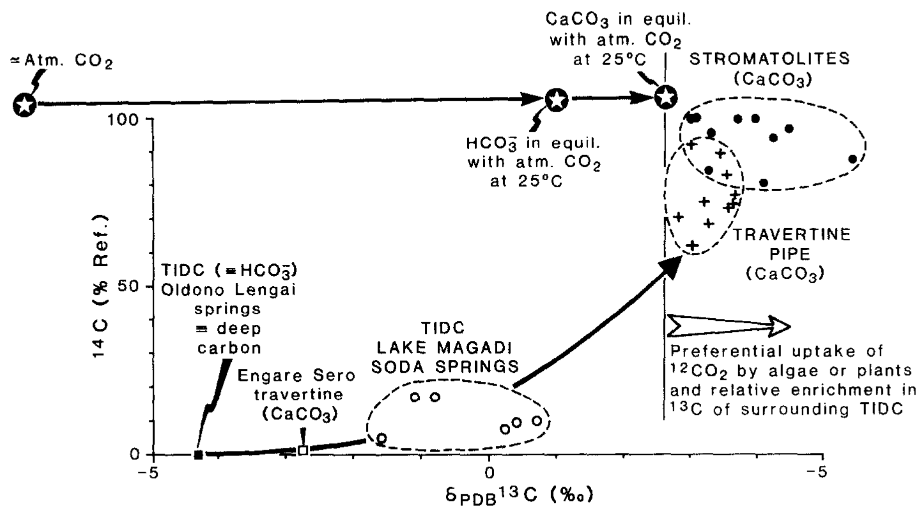


FIG. 8. $^{13}C$ and $^{14}C$ evolution of TIDC and carbonates in relation to isotopic exchanges with atmospheric $CO_2$ and to algal activity.

ties measured on stromatolites in terms of $^{14}C$ chronology. For purposes of comparison with other rift lakes, the dates have to be normalized to the same $^{13}C$ reference value of $-25$‰ (cf. Broecker and Olson, 1959, 1961). This represents a ca. 500-yr correction. The "corrected" ages are indicated on Figure 4.

The $^{14}C$ age obtained on the *Tilapia* beds is unfortunately of no help in determining the chronology of the highest lake level. There are two reasons for this. First, the fish carbon probably derived from lake organic matter, the carbon of which was eventually extracted from TIDC; consequently, if the latter had an apparent $^{14}C$ age, one may expect the same anomaly, with minor isotopic fractionation, to be present in the fish. Second, the relatively indurated tephra layers which fossilized the *Tilapia* usually occur on top of the lacustrine clay, suggesting that the *Tilapia* beds

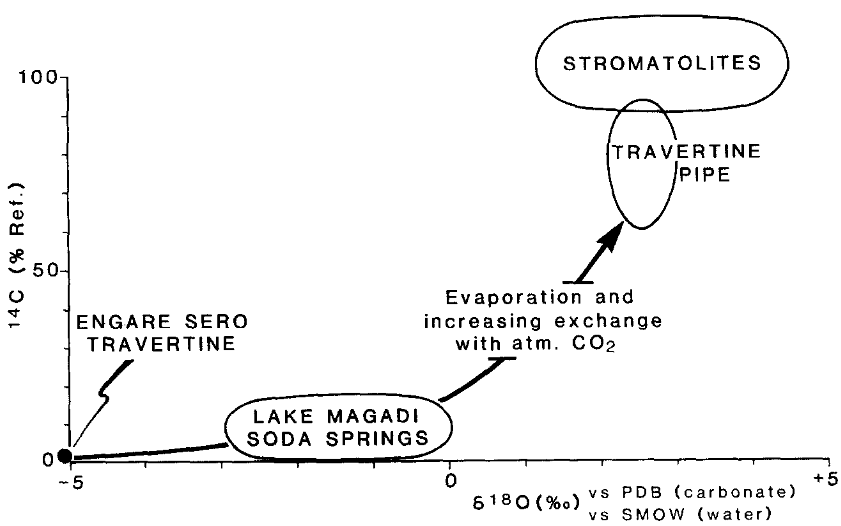


FIG. 9. $^{18}O$ and $^{14}C$ relationship of water–TIDC and carbonates in relation to evaporation and exchanges with atmospheric $CO_2$.

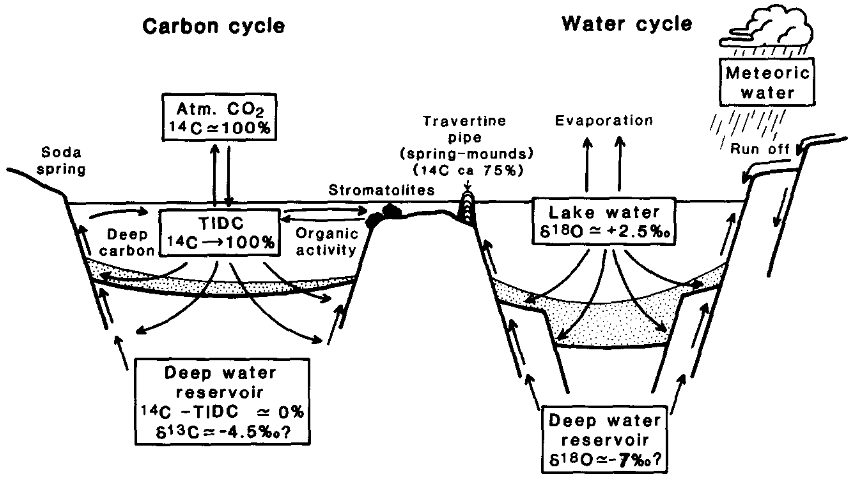


FIG. 10. Sketch of the isotopic hydrology of paleolakes Natron and Magadi.

instead correspond to the end of the high lacustrine episode.

On a normalized radiocarbon scale, we suggest the succession of events at the Pleistocene–Holocene transition period to have been as follows:

(1) Lake high stands ca. 11,800, 10,800, and 10,300 yr B.P. No direct evidence of regressions in between the high stands have been found. These three ages most probably correspond to phases of very stable lake level and chemistry, thus allowing stromatolite construction. Between ca. 11,800 and 10,300 yr B.P., vertical lifting of some blocks of the rift occurred (Fig. 11).

(2) The lake recession seems to correspond to a period of strong volcanic activity, ca. 9100 yr B.P.; the volcanic ashes induced massive death of fish in the lake and the *Tilapia* beds were deposited.

(3) During the Holocene, a relatively dry and seasonally contrasted regime probably persisted, while the abundant trona layers

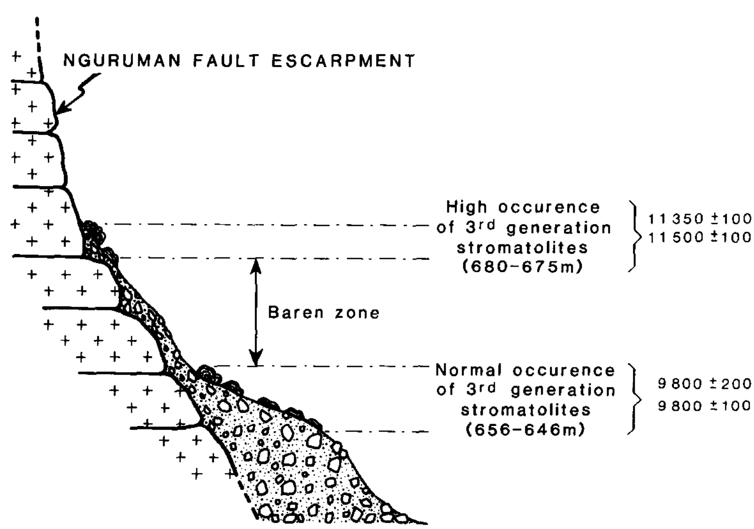


FIG. 11. Vertical distribution of the recent stromatolites along the Nguruman fault escarpment suggesting a ca. 20-m vertical displacement during the paleolake episode.

of Lake Magadi sediments (Baker, 1958) accumulated. However, periods with the water level rising a few meters above the modern one did occur: a $^{14}C$ date of 4650 ± 400 yr B.P. (UQ-927) has been obtained on fossil *Tilapia* collected at an elevation of ca. 605 m, SE of lake Magadi.

## LATE MIDDLE PLEISTOCENE LAKE AND SECOND GENERATION STROMATOLITES

For obvious reasons, the older stromatolites are not as well preserved as the early Holocene ones. They have been almost completely eroded on the steep walls of the fault escarpments which delimit most of the paleolake margins. However, in particularly favorable exposures, on flat or gently sloping surfaces, they are observed jointly with the discordant youngest generation concretions. Being morphologically quite different from the latter (Casanova, in press)—they often occur as thick bioherms—they are easily identified, and micrographic and geochemical studies confirmed the morphological diagnosis.

More than 12 sites, equally distributed around the basin, were sampled. Stable isotope measurements yielded values within the ranges already defined by the third generation stromatolites. It can therefore be assumed that the corresponding lacustrine episodes occurred under similar paleohydrological conditions. Both paleoshorelines may be surveyed within a very narrow altitudinal range (ca. 655–656 m). The basin morphology may account for the fact that the lakes were filled to the same elevation (Hillaire-Marcel and Casanova, in press).

The old generation stromatolites (Tables 3A and B and Fig. 12) yielded $^{14}C$ activities $\leq 1.4\%$. The traces of active carbon may simply be due to the incorporation of secondary calcite or to partial recrystallization. These carbonates are beyond the chronological limits of the $^{14}C$ method. We therefore attempted to obtain dates with the Th/U disequilibrium method.

Preliminary results showed a large spread of Th and U contents and activity ratios (Table 3). The relatively large range of $^{234}U/^{238}U$ ratios (from 1.2 to 1.36 for most samples) and absolute content in uranium indicate that partial leaching with preferential removal of $^{234}U$ may have occurred at least during the recent Holocene humid phase. Moreover, most samples contain a high proportion of $^{232}Th$, suggesting that a fair amount of $^{230}Th$ of detrital origin could have been simultaneously incorporated into the calcitic network.

It was thus initially impossible to determine significant $^{230}Th/^{234}U$ ratios representing radioactive decay which had occurred since the deposition of the calcite. Fortunately, the most recent stromatolites could be analyzed in order to estimate original isotopic ratios. Since both lakes were fed by waters collected within the same well-defined drainage basin, we expected that, after correction for the decay which had occurred since ca. 10,000 $^{14}C$ yr B.P., the $^{234}U/^{238}U$ and $^{230}Th/^{232}Th$ ratios measured on the late Pleistocene–early Holocene stromatolites would provide the required basic data. The results surpassed our expectations (Fig. 13 and Table 3). The 12 samples yielded very homogeneous values: 0.87 ± 0.025 for $^{230}Th/^{232}Th$ and 1.40 ± 0.03 for $^{234}U/^{238}U$. The detrital particles transported into the lake during both episodes were eroded from the same rocks, and for the first approximation there was no reason to suspect different $^{230}Th/^{232}Th$ initial ratios in the second and third generations of stromatolites. Therefore, most $^{230}Th$ in the second generation would derive from this "detrital" component and has to be subtracted from each sample to allow for detrital input. The subsequent corrected values coupled with the $^{234}U$ measurements made it possible to approximate an age for each sample. All fall into reasonably narrow brackets: 142,000 ± 25,000 yr (Table 3).

As a first step, this estimation of the age of the samples is quite satisfying. However,

TABLE 3A. THIRD GENERATION TH AND U ACTIVITY RATIOS IN STROMATOLITES

| Lab. No | Field No | Total U ppm ± σ | $^{234}U/^{238}U$ | $^{230}Th/^{232}Th$ | $^{230}Th/^{234}U$ | Uncorrected $^{14}C$ Age | Corrected $^{14}C$ Age ($\delta^{13}C$ = −25‰) | $(^{230}Th/^{232}Th)_0$ Initial Ratio | $(^{234}U/^{238}U)_0$ Initial Ratio | Remarks |
|---|---|---|---|---|---|---|---|---|---|---|
| THIRD GENERATION STROMATOLITES | | | | | | | | | | |
| 1) LAKE NATRON AREA | | | | | | | | | | |
| UQT-53 | LN.18.10.10 | 6.28±0.08 | 1.39±0.01 | 0.834±0.005 | 1.49±0.05 | 9,900±150 | ~ 10,400 | 0.85 | 1.40 | |
| UQT-55 | LN.14.10.11 | 2.18±0.02 | 1.40±0.01 | 0.890±0.006 | 2.15±0.06 | 10,500±120 | ~ 11,100 | 0.91 | 1.41 | |
| UQT-76 | LN.16.10.1 | 2.33±0.03 | 1.41±0.02 | 0.847±0.005 | 2.97±0.10 | 10,300±130 | ~ 10,800 | 0.89 | 1.42 | |
| UQT-78 | LN.7.10.4 | 3.19±0.03 | 1.35±0.01 | 0.861±0.005 | 2.64±0.08 | 10,250±130 | ~ 10,800 | 0.91 | 1.36 | |
| UQT-57 | LN.8.10.4 | 4.46+0.07 | 1.44+0.01 | 0.831+0.008 | 1.53+0.06 | 12,030+150 | ~ 12,500 | 0.85 | 1.45 | |
| UQT-130 | /.73.053 | 3.39±0.03 | 1.40±0.01 | 0.860±0.010 | 1.66±0.05 | 9,840±100 | ~ 10,300 | 0.88 | 1.41 | |
| 2) LAKE MAGADI AREA | | | | | | | | | | |
| UQT-111 | LM.47 | 1.39±0.03 | 1.34±0.02 | 0.797±0.007 | 5.10±0.19 | 10,935±165 | ~ 11,400 | 0.85 | 1.35 | |
| UQT-112 | LM.18 | 6.18±0.08 | 1.37±0.01 | 0.855±0.008 | 1.36±0.05 | 10,460±120 | ~ 10,900 | 0.87 | 1.39 | |
| UQT-128 | MAG.83.44 | 5.09±0.05 | 1.38±0.01 | 0.844±0.009 | 1.37±0.04 | 10,875±310 | ~ 11,400 | 0.86 | 1.39 | |
| UQT-132 | MAG.83.04 | 0.74±0.01 | 1.39±0.02 | 0.791±0.008 | 6.73±0.23 | 10,300±200 | ~ 10,800 | 0.85 | 1.40 | |
| UQT-79 | MAG.82.02 | 2.76±0.00 | 1.38±0.01 | 0.802±0.005 | 4.83±0.14 | — | ca 10,300 | 0.85 | 1.39 | Not dated by $^{14}C$ |
| 3) TRAVERTINE PIPE | | | | | | | | | | |
| UQT-195 | LM.8 | 59.0±0.6 | 1.24±0.01 | 5.75±0.17 | 0.105±0.004 | 14,380±180 | ca 11,000 | 0.85 | 1.25 | $12,000 ^{+500}_{-300}$ by corrected $^{230}Th/^{234}U$ dating |
| | | | | | | mean initial ratios (stromatolites) | | 0.87±0.025 | 1.40±0.026 | |

TABLE 3B. First and Second Generation Th and U Activity Ratios in Stromatolites

| Lab. No | Field No | Total U ppm ± σ | $^{234}U/^{238}U$ | $^{230}Th/^{232}Th$ | $^{230}Th/^{234}U$ | Uncorrected Age | $(^{234}U/^{238}U)_0$ Initial Ratio | Corrected Age * (± ca 10%) | Remarks * with initial $^{230}Th/^{232}Th$=0.87−0.05 |
|---|---|---|---|---|---|---|---|---|---|
| SECOND GENERATION STROMATOLITES | | | | | | | | | |
| 1) LAKE NATRON AREA | | | | | | | | | |
| UQT-52 | LN.9.10.8 | 0.80±0.01 | 1.25±0.02 | 0.291±0.005 | 3.91±0.16 | — | 1.38 | 144,000 | on top of UQT-54 |
| UQT-56 | LN.14.10.12 | 0.43±0.01 | 1.36±0.02 | 0.287±0.005 | 6.13±0.25 | — | 1.53 | 136,000 | |
| UQT-77 | LN.7.10.3 | 0.336±0.005 | 1.28±0.02 | 0.287±0.005 | 11.92±0.36 | — | 1.40 | 129,000 | |
| UQT-190 | LN.7.10.2 | 1.61±0.01 | 1.23±0.01 | 0.293±0.003 | 1.95±0.06 | — | 1.39 | 180,000 | on top of UQT-191 |
| UQT-193 | LN.7.10.5 | 1.98±0.03 | 1.25±0.02 | 0.306±0.003 | 2.71±0.10 | — | 1.38 | 152,000 | on top of UQT-192 |
| 2) LAKE MAGADI AREA | | | | | | | | | (1) completely |
| UQT-131 | LM.34 | 1.09±0.02 | 1.20±0.02 | 0.122±0.01 | 0.75±0.03 | $142{,}000^{+11{,}000}_{-10{,}000}$ | — | —(1) | recrystallized in relation |
| UQT-129 | MAG.83.54 | 0.69±0.01 | 1.24±0.02 | 0.309±0.002 | 2.56±0.10 | — | 1.36 | 154,000 | to basaltic intrusion |
| UQT-127 | MAG.83.42 | 1.20±0.01 | 1.31±0.01 | 0.315±0.002 | 2.00±0.07 | — | 1.49 | 170,000 | |
| UQT-115 | LM.35 | 1.35±0.02 | 1.22±0.02 | 0.342±0.003 | 1.88±0.08 | — | 1.34 | 164,000 | |
| UQT-113 | LM.48 | 0.80±0.01 | 1.21±0.02 | 0.361±0.003 | 3.71±0.10 | — | 1.29 | 129,000 | |
| UQT-133 | MAG.83.28 | 0.34±0.005 | 1.35±0.02 | 0.295±0.002 | 5.86±0.18 | — | 1.51 | 134,000 | |
| FIRST GENERATION STROMATOLITES (LAKE NATRON AREA) | | | | | | | | | |
| UQT-59 | LN.18.10.9 | 15.99±0.06 | 1.00±0.01 | 5.99±0.35 | 0.94±0.03 | $307{,}000^{+52{,}000}_{-34{,}000}$ | 1.00 | 290,000 | (2) correction with |
| UQT-54 | LN 9.10.7 | 1.90±0.04 | 0.87±0.02 | 0.226±0.003 | 1.35±0.07 | — | ca 1.6 | 304,000(2) | $^{234}U/^{238}U$ = 1.04 |
| UQT-58 | LN.2.10.1 | 0.373±0.003 | 1.07±0.01 | 0.045±0.002 | 1.23±0.08 | — | 1.28 | >300,000 | (mean value) |
| UQT-60 | LN.2.10.2 | 0.380±0.006 | 1.03±0.02 | 0.051±0.003 | 1.10±0.07 | — | 1.15 | >300,000 | |
| UQT-191 | LN.7.10.1 | 0.380±0.010 | 1.07±0.02 | 0.060±0.001 | 0.95±0.04 | $300{,}000^{+120{,}000}_{-56{,}000}$ | — | — | |
| UQT-192 | Na.82.45 | 3.90±0.04 | 1.19±0.01 | 0.470±0.006 | 1.12±0.04 | — | 1.44 | 295,000 | |
| OTHER RELEVANT MEASUREMENTS | | | | | | | | | |
| UQT-173 | Syenite (K-Feldspar) related to LENGAI LAVA FLOWS (RECENT) | | | | | | | | |
| | | 1.12±0.01 | 1.01±0.01 | 0.98±0.01 | 1.16±0.04 | — | 1.01 | — | |
| UQT-194 | TRONA CRUST OF LAKE NATRON (LATE HOLOCENE) | | | | | | | | |
| | | 2 10±0.02 | 1.28±0.01 | 0.643±0.001 | 0.26±0.01 | | 1.24 | | |

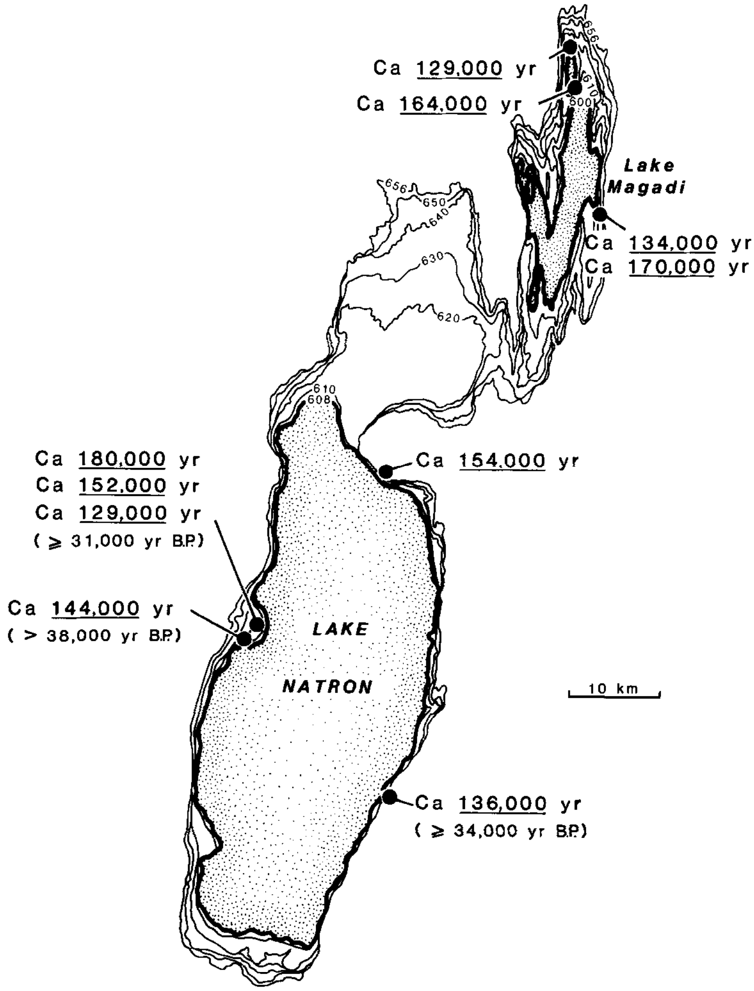


FIG. 12. Calculated Th/U ages of the second generation of stromatolites (the standard deviation represents a ca. 10–15% uncertainty on the ages).

consideration of $^{234}U/^{238}U$ ratios may allow greater accuracy in dating. The best preserved samples (UQT-56 and UQT-133) showed no evidence of secondary calcite and yielded a ratio of ca. 1.35. By extrapolating back some 140,000 yr, an initial ratio of ca. 1.52 can be calculated. It is likely that most other samples did not maintain a closed system for $^{234}U$ decay. Very probably the ages of samples UQT-56 and UQT-133 (ca. 135,000 yr B.P.) are the most representative of this generation of stromatolites. Interestingly, this age is similar to those obtained for late middle Pleistocene lakes in southern Libya (Gaven *et al.*, 1981) and in the Lake Turkana area (Butzer *et al.*, 1969). We are tempted to conclude that, from a paleoclimatic point of view, intertropical Africa is characterized by humid episodes during each glacial–interglacial transition between, for instances, stages 6–5 and 2–1 of the $^{18}O$ oceanic record (CLIMAP, 1984). The Lake Natron–Lake Magadi area is the third example of well preserved remains of the corresponding lacustrine episodes; others will likely be found as research continues.

## FIRST GENERATION STROMATOLITES

The six samples associated with the first

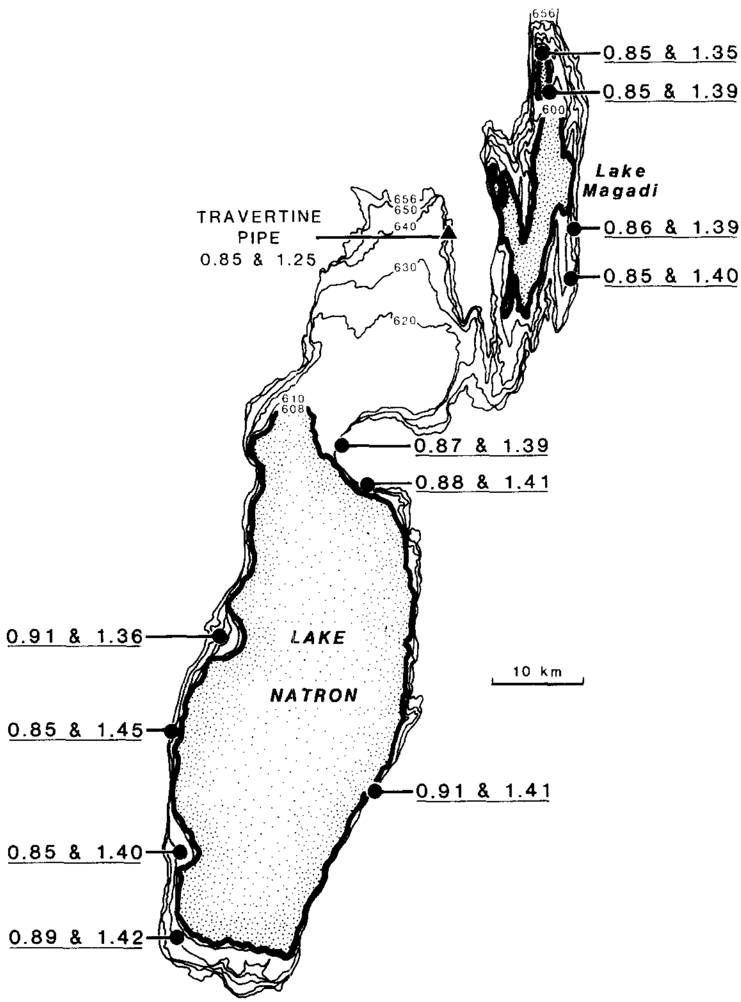


FIG. 13. Initial $^{230}Th/^{232}Th$ and $^{234}U/^{238}U$ ratios calculated for third generation stromatolites (ca. 10,300 yr B.P.).

generation stromatolites showed considerable depletion in $^{234}U$ due to prolonged leaching. Sample UQT-54 (Table 3) has a $^{234}U/^{238}U$ ratio of 0.87. This stromatolite, after erosion (Fig. 2), was covered by a younger one (UQT-52) dating from the late middle Pleistocene. What could be the age of the first stromatolite? Since we do not know when the main leaching phases occurred, it is difficult to make a meaningful estimate. An approach similar to that used above yields ages of ca. 300,000 yr (Table 3) which although coherent, are beyond the chronological limits of the method. However, sample UQT-59 (Table 3), which was found up the Moinik River, has a much higher uranium content. The uncorrected age of this sample is still in the 250,000- to 350,000-yr range; correction for detrital thorium inputs is almost negligible. Admitting an initial $^{234}U/^{238}U$ ratio similar to the maximum one calculated (ca. 1.52), and considering that the preferential leaching of $^{234}U$ occurred equally during both more recent lacustrine episodes, a corrected age of ca. 240,000 + 34,000/ −52,000 yr is obtained. The oldest generation of stromatolites could well be correlative of the transition between $^{18}O$ stages 8 and 7 (ca. 240,000 yr; Berggren *et al.*, 1980).

## CONCLUSIONS

The Lake Natron–Lake Magadi area had its latest maximum lacustrine extension some 10,800 yr B.P. (normalized $^{14}C$ age). The lake rise probably started earlier (ca. 11,800 yr B.P.) and the main recession was completed ca. 9,100 yr B.P. (*Tilapia* beds). This chronology fits reasonably well with the East African record of the late Pleistocene–early Holocene humid maximum. Similar high lake levels existed during the Pleistocene, some 135,000 yr B.P. and possibly ca. 240,000 yr B.P. Further studies in intertropical Africa should confirm the recurrence of humid episodes during glacial–interglacial transitions; the last episode (2–1 in the oceanic $^{18}O$ record) can be observed almost everywhere in Africa. The previous episode (6–5) now has been demonstrated in at least two and perhaps three areas: southern Libya (Gaven *et al.*, 1981), the southern Gregory Rift (present study), and the Lake Turkana area (Butzer *et al.*, 1969). The first episode (8–7) also seems to be recorded in the Lake Natron–Lake Magadi basin; however, its age is not as solidly grounded as that of the younger episodes.

The most significant conclusion we draw from our study concerns the potential use of stromatolites. Unlike other paleolake indicators, they have the advantage of being closely related to maximum lake levels. Their structure also provides an insight into paleolimnology and paleoecology (Casanova, in press), seasonality contrasts, and, through stable isotope measurements, paleohydrology.

Comparative studies of modern environments and geochemistry may help to determine whether paleowater inorganic dissolved carbon reached equilibrium with atmospheric $CO_2$. If it did, $^{14}C$ ages obtained on stromatolites are indicative of the age of the maximum lake level. Similarly, thorium and uranium isotope ratios in stromatolites may be used to date deposits older than ca. 40,000 $^{14}C$ yr B.P. It is, however, essential to determine the exact initial isotopic ratios ($^{234}U/^{238}U$; $^{230}Th/^{232}Th$) to take into account the detrital component of the $^{230}Th$ measured activity in fossils and also to estimate the proportion of uranium that might have been removed by leaching. In the present case, the occurrence of several generations of stromatolites within the same well-defined basin, the youngest of them being dated by the $^{14}C$ method, made it possible to examine these parameters and consequently to evaluate Th/U ages for the older generations.

A humid episode characterizes the early Holocene of intertropical Africa, and paleolake deposits are found almost everywhere. When they contain carbonates, the model used here may also be applied to the study of older lacustrine deposits.

## ACKNOWLEDGMENTS

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