

Lake level and paleoenvironmental history of Lake Tanganyika, Africa, as inferred from late Holocene and modern stromatolites

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ABSTRACT

Fossil and living stromatolites are abundant around the margins of Lake Tanganyika, Africa, and provide a wealth of paleolimnologic and paleoclimatic information for the late Holocene. Six lines of evidence show that stromatolites and cements are precipitating in the lake today: (1) carbonate saturation state calculations, (2) documentation of living stromatolites and their depth distribution, (3) new stable isotope data showing the lake's present mixing state and ancient evaporation and inflow balance, (4) new radiocarbon data and a reevaluation of apparent ^{14}C ages derived from Lake Tanganyika carbonates, (5) the presence of modern Mg-calcite cements derived from lake waters, and (6) the presence of modern, biologically mediated Mg-calcite precipitates in the lake.

Lake Tanganyika's lake levels have been remarkably stable over the past 2800 yr, fluctuating around the marginally open to marginally closed level through most of this time period. Lake lowstands and high $\delta^{18}\text{O}$ values from the ninth century B.C. to the early fifth century A.D. indicate that the lake basin was comparatively dry during this time. However, the period prior to the most recent opening of Lake Kivu into the Lake Tanganyika basin (ca. A.D. 550) was not marked by major lake lowstands, nor was this opening accompanied by a dramatic lake-level rise. The Kivu opening was roughly coincident with a significant shift toward isotopically lighter ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) lake water, which persists today. The lake remained close to its outlet level between the sixth and thirteenth centuries A.D. Lake levels rose between the fourteenth and sixteenth centuries. At some

time between the late sixteenth and early nineteenth centuries, lake level fell to perhaps its lowest level in the past 2800 yr. By the early nineteenth century, lake level had begun to rise to the overflow level, apparently the result of a regional increase in precipitation/evaporation ratios.

Weak $\delta^{18}\text{O}/\delta^{13}\text{C}$ covariance for late Holocene carbonates suggests that the surface elevation of the lake has remained near the outlet level, with only occasional periods of closure. However, there is no simple relationship between solute input from Lake Kivu, isotope input from Lake Kivu, and lake levels in Lake Tanganyika. Lake Kivu waters are the primary source of major ions in Lake Tanganyika, but are much less important in controlling the $\delta^{18}\text{O}$ and the lake level of Lake Tanganyika. Because the Ruzizi River's discharge into Lake Tanganyika is largely derived from sources other than Lake Kivu, the overflow events in the two lakes have been uncoupled during the late Holocene.

INTRODUCTION

Numerous studies have demonstrated that laminated stromatolites and massive thrombolites can be used in paleolimnological reconstructions (Abell et al., 1982; Casanova, 1986; Hillaire-Marcel and Casanova, 1987; Casanova and Thouin, 1990; Straccia et al., 1990; Talbot, 1990). Stromatolites can provide information on lake-level fluctuations (based on their elevation), paleohydrology (through ^{13}C and ^{18}O analyses of stromatolitic carbonates), paleoenvironments (from their growth morphology), and chronology (through direct ^{14}C dating). Here we discuss new observations on the stromatolites of Lake Tanganyika, Africa, that shed light on the most recent

history of that lake, and on the late Holocene climate history of Central Africa.

Lake Tanganyika is located in the western branch of the African Rift system (Fig. 1). It is the largest of the African rift lakes and the second deepest lake in the world. The lake occupies a series of interconnected half-graben basins, the oldest of which is probably between 9 and 12 Ma (Cohen et al., 1993). The basinal bedrock lithology consists of Proterozoic metasedimentary rocks, Karroo (late Paleozoic–early Mesozoic) nonmarine sedimentary rocks, and late Tertiary volcanic rocks in the Ruzizi River watershed, occupying the north end of the lake basin.

The lake is currently an open basin, discharging into the Zaire basin via the Lukuga River. However, most water loss is from evaporation, and the outlet serves primarily to maintain a relatively constant lake level (Hecky and Degens, 1973). The lake has been closed as recently as 1878 (Camus, 1965).

Climate in the lake basin is semihumid tropical. The lake receives continuous, but seasonally variable, inflow from four major rivers (Malagarasi, Ruzizi, Lugufu, and Lufubu), as well as numerous small rivers. Most of the influent streams originate on rift escarpments, close to the lake, and therefore carry very low dissolved solid loads. In contrast, the Ruzizi River drains alkaline Lake Kivu to the north, and is the dominant source for most solutes in Lake Tanganyika, excluding Ca^{2+} (Craig et al., 1974). Lake Kivu surface waters are enriched in Mg^{2+} relative to Ca^{2+} , accounting for the unusually high Mg/Ca molar ratio (≈ 9) of the Ruzizi River where it enters Lake Tanganyika. Ruzizi waters are subsequently diluted by waters from the other influent streams to produce a low salinity, low alkalinity lake water (conductivity = 670 $\mu\text{mho}/\text{cm}$, $A_c = 6.6$ meq/l). Ground-water input from small hydrothermal springs has been documented in

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some parts of Lake Tanganyika (TANGHYDRO Group, 1992), but the volume of fluid flow is relatively small ($\approx 100\text{--}300$ l/s each, for the two fields discovered to date). Stable isotope (^{18}O and ^{13}C) and major ion chemistry measurements suggest that the surface waters are well mixed (Casanova and Hillaire-Marcel, 1992), although these epilimnetic waters mix much more slowly with the voluminous and anoxic monimolimnion of the lake (Coulter and Spigel, 1991). Surface water composition and major ion concentrations have been very constant, both seasonally and over several decades, because of the lake's size and the long residence time of its water (Hecky and Degens, 1973; Cohen and Thouin, 1987; Batungwanayo, 1987; Casanova and Hillaire-Marcel, 1992).

Calcareous sediments were first reported from the lake by Dubois (1959). Cores taken in the 1970s showed that aragonitic and calcitic muds have been deposited periodically during the Holocene (Stoffers and Hecky, 1978). Cohen and Thouin (1987) described numerous carbonate lithofacies in the lake, including lime muds, coated-grain and skeletal-fragment sands and gravels, and microbially cemented boundstones (stromatolites and thrombolites). Most of these sediments were either high-Mg calcite or aragonite, an observation that is consistent with the carbonate mineralogy predicted by Müller et al. (1972) based on the molar proportions of Mg/Ca in the lake water. Cohen and Thouin (1987) proposed that carbonates are currently being deposited in the lake, and that stromatolites and thrombolites have formed under essentially modern conditions.

Subsequent work has shown that many of the surficial carbonate deposits observed on the lake floor are quite old. Cohen (1989) showed that shell lags of up to 2000 yr in age are exposed on the lake floor at littoral-sublittoral depths, apparently the result of winnowing events during lowstands of lake level. On the basis of radiocarbon dating and ^{18}O evidence from stromatolites, Casanova and Thouin (1990) and Casanova and Hillaire-Marcel (1992) argued that all stromatolites found on the lake floor (and implicitly all littoral carbonate deposits other than mollusk and ostracode shells) are fossil. In their chronology, stromatolites formed between ca. 2000 B.C. and A.D. 750, during a period of late Holocene aridity and Lake Kivu closure. They proposed that an increase in the regional precipitation/evaporation balance and subsequent dilution of Lake Tanganyika waters during ca. A.D. 650–750 put an end to carbonate accumulation and stromatolite formation.

Here we present new data on the late Holocene history of Lake Tanganyika. Our findings are primarily based on the occurrence of both modern and Holocene stromatolites from parts of the lake which were not visited by earlier workers. We argue that microbially mediated and inorganically

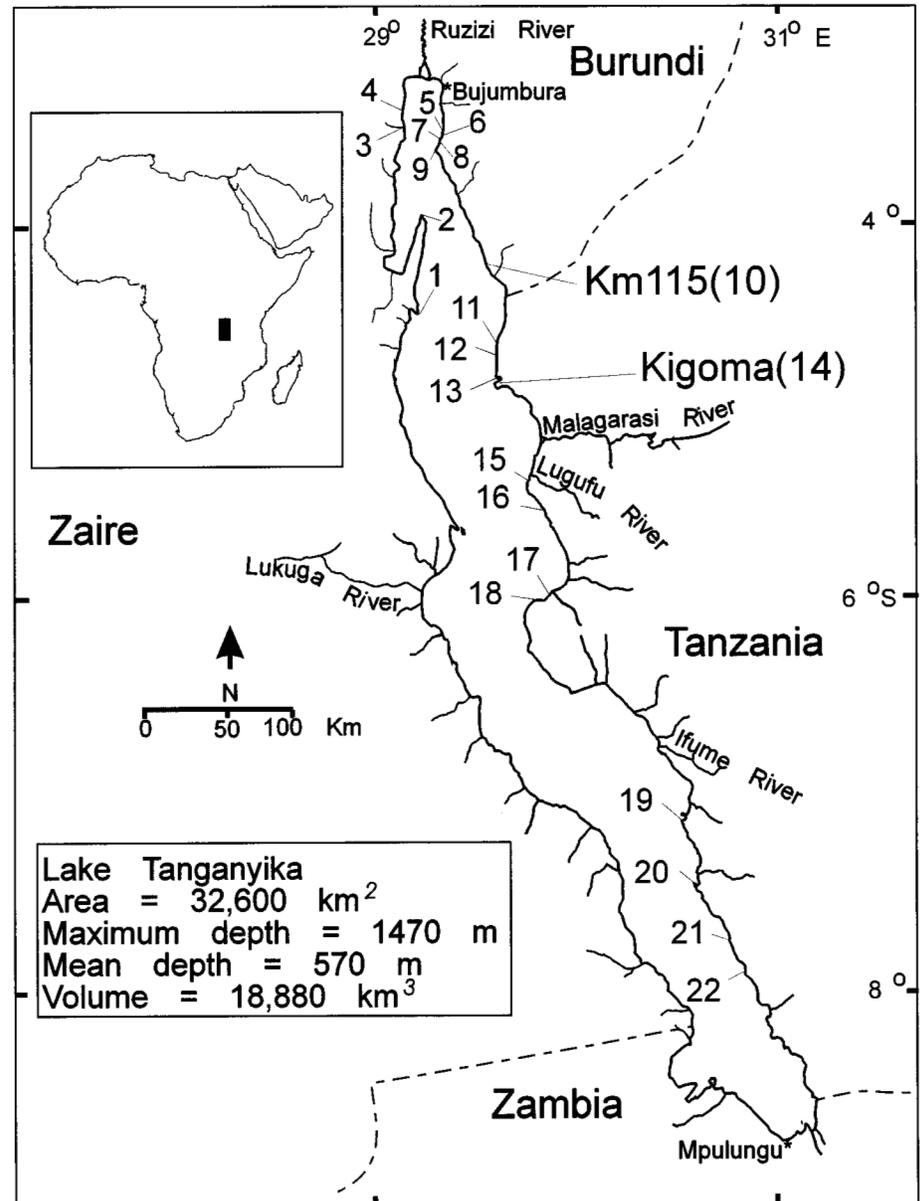


Figure 1. Location map of Lake Tanganyika showing known stromatolite localities. Much of the shoreline of Lake Tanganyika is unexplored. Thus, this map should not be construed to indicate that stromatolites are not present elsewhere. Numbers correspond to locality numbers in Table 2. Principal research sites discussed in text are no. 10 (Km 115 site, southern Burundi) and no. 14 (Kigoma site, Tanzania).

precipitated carbonate deposition did not cease in Lake Tanganyika at ca. A.D. 750, but has continued to the present.

METHODS

Saturation state, water balance, and isotopic equilibrium calculations were based on previously published water chemistry data (Table 1; Hecky and Degens, 1973; Craig et al., 1974; Tietze, 1981, 1982; Cohen and Thouin, 1987; Batungwanayo,

1987; Casanova and Hillaire-Marcel, 1992). Activity coefficients were calculated using the Debye-Hückel equation. Ion pairing in solution was estimated using the thermodynamic association constants preferred by Millero and Schreiber (1982, their Table 3). We ignored all Na^+ , K^+ , and SO_4^{2-} pairs in order to simplify calculations, because they had a negligible effect ($<1\%$) on the ions of interest (Ca^{+2} , Mg^{+2} , HCO_3^- and CO_3^{2-}). By the third iteration of these calculations, changes in free ion activities were less than 0.04%.

TABLE 1. SEASONAL VARIATION IN LAKE TANGANYIKA NORTH BASIN SURFACE WATER CHEMISTRY

Parameter	Wet season	Dry season
Na ⁺	53.5 mg/l	55.2 ± 2.3 mg/l
K ⁺	29 mg/l	28.5 ± 1.1 mg/l
Ca ⁺²	12.2 mg/l	12.4 ± 2.3 mg/l
Mg ⁺²	45.7 mg/l	45.4 ± 1.9 mg/l
Total cations	7.44 meq/l	7.49 meq/l
Alkalinity	6.5 meq/l	6.6 ± .01 meq/l
Cl ⁻	31.7 ± 1.3 mg/l	32.26 ± 0.9 mg/l
SO ₄ ⁻²	2 ± 1 mg/l	1.4 ± 1.2 mg/l
Total anions	7.44 meq/l	7.54 meq/l
pH	8.8–9.1	8.9–9.2
Temperature	27.2 °C	25.8 ± 0.7 °C

Note: Data source: Wet season: Casanova and Hillaire-Marcel (1992), 0–40 m. Dry season: Cohen and Thouin (1987). Cl⁻, SO₄⁻²; pH values for both wet and dry seasons are from Batungwanayo (1987); ranges are ±1σ.

Between 1985 and 1992, we identified stromatolite localities around Lake Tanganyika using SCUBA, echosounder profiling, and a remotely operated vehicle. At each locality water depth and gross stromatolite morphology were recorded by divers. Stromatolites were subdivided in the field into those that appeared to be growing and those that were either dead or ambiguous. We identified actively accreting stromatolites by the presence of HCl-reactive CaCO₃ crusts within a surficial mass of soft algae.

We studied two localities in detail. One is an area of fossil stromatolites in southern Burundi (also visited by Casanova and Thouin, 1990) and the other is an area of both fossil and modern stromatolites in northern Tanzania. The former locality lies 200 to 500 m offshore of the Burundi coast, in the southern portion of the Nyanza Lac structural platform (marked Km115[10] in Fig. 1 because of its proximity to kilometer point 115 on Burundi National Route 3). The second site is south of the town of Kigoma, ≈3 km south of the Kigoma Railway Hotel and between 100 and 400 m offshore, directly west of Bangwe Hill (labeled Kigoma [14] in Fig. 1). Supplementary observations and isotopic data were collected at other sites listed in Table 2.

Representative stromatolites were collected by SCUBA divers, and returned to the United States for analysis. We sectioned individual heads and examined surfaces and thin sections using light microscopy and scanning electron microscopy (SEM) to determine the microbial assemblages, structure, and composition of the stromatolites.

We obtained conventional ¹⁴C age dates from stromatolites and other carbonate samples. We cleaned our samples of organic and surficial material, and submitted the outermost portions of the stromatolites for analysis to the University of Arizona's Environmental Radioisotope Facility or the Norwegian National Radiocarbon Laboratory at Trondheim. Subsequent treatment at the

TABLE 2. KNOWN STROMATOLITE OCCURRENCES IN LAKE TANGANYIKA

Locality number	Long (E)	Lat (S)	Depth range (m)	Substrate	Growth morphologies present
1	29°10'	4°34'	4	Rocks	SC
2	29°13'	4°05'	23*	Cemented rubble and rocks	SC, LI
3	29°08'	3°36'	20*	Boulders in sand	SC
4	29°08'	3°30'	9–25	Rock walls and ledges	SC
5	29°20'	3°38'	9–30	Boulders in sand	SC
6	29°20'	3°39'	16*	Boulders in sand and rocks	SC
7	29°20'	3°40'	18*	Rocks	SC, LI
8	29°20'	3°41'	12–18	Loose boulders in sand	SC
9	29°19'	3°43'	8*	Loose boulders and rocks	SC
10 (Km115 site)	29°34'	4°17'	11–40	Cemented sand and cobbles, rocks	LI, LL, MB
11	29°37'	4°41'	30–46	Rocks	N.D. [†]
12	29°36'	4°45'	6–12	Cemented rubble and rocks	SC, LI
13	29°36'	4°52'	20–30	Rocks	LI, LL, MB
14 (Kigoma site)	29°36'	4°54'	6–12	Sand, cemented sand, rocks	SC, LI, LL
15	29°46'	5°24'	4–14	Rocks	SC
16	29°49'	5°31'	15–18	Rocks, sand	SC, LI
17	29°51'	5°58'	6–10	Shell lag	SC, LI
18	29°48'	6°00'	15–24	Rocks	SC, LI
19	30°31'	7°10'	24–30	Rocky ledges	LI
20	30°35'	7°27'	21*	Rocks	LI
21	30°47'	7°50'	24*	Rocks	LI
22	30°52'	7°59'	24–34	Rocks	LI, LL

Note: Locations in this table are keyed by number to Figure 1. Sites 10 (southern Burundi) and 14 (Kigoma, Tanzania) are the principal study sites from which most of our observations are derived. Morphology codes: SC—small columns and stratiform encrustations, LI—large isolated columns and domes, LL—large linked columns and domes, MB—major biohermal buildups.

*Indicates that a sample was collected from the given depth, but no range information was collected.

[†]N.D.—not determined.

University of Arizona involved pumping off the first 10% of the CO₂ evolved from acid digestion, thereby removing the remaining carbonate rinds and interspace area fill, and analyzing the remainder of the evolved gas. At Trondheim, we treated our samples by leaching the outer surface with dilute HCl to remove possible surface contamination. Between 60% and 90% of the original sample was then used for dating.

Prior to selecting samples for stable isotopic analysis we slabbled, polished, and thin sectioned the stromatolites to identify possible contaminants and diagenetic overgrowths. We examined both the polished slabs and thin sections using polarized transmitted light and (for thin sections) cathodoluminescence and blue/violet (400–440 nm wavelength) incident light microscopy.

Following inspection, we drilled samples (average 0.5 mg) from polished slabs using a jeweler's microdrill fitted with a 0.4 mm stainless steel bit. We heated our samples at 400 °C under vacuum, and then dissolved them in 100% phosphoric acid at 25 °C. The resultant CO₂ gas was analyzed for stable carbon and oxygen isotope ratios using a Finnigan MAT 251 mass spectrometer at the GMS Laboratory, University of Bergen, using standard corrections. Our analytical precision is better than ±0.1‰ for both δ¹³C and δ¹⁸O. We analyzed 27 carbonate samples taken from nine stromatolites and assorted oncolite, ooid beachrock, and fossil mollusk specimens.

The use of isotopic data from stromatolites for paleoenvironmental interpretations requires a

modern baseline for comparison. We analyzed living littoral-sublittoral gastropods from Lake Tanganyika for this purpose, because the isotopic composition of modern gastropods reflects both modern average conditions and seasonal variation in the factors affecting the δ¹⁸O of lacustrine carbonates. Most African gastropods have life spans of a year or more (Levêque, 1973), and sampling successive increments of shell growth provides a picture of the oxygen isotope variation encountered in the water column during the organism's lifetime. Temperature variation, which also affects the δ¹⁸O of shell, is very small (≈4 °C seasonally). Therefore, the isotopic composition of surficial and deep water, as well as evaporation of surface waters, dominates the isotopic history of the individual gastropod's shell.

We analyzed 34 living gastropods (*Lavigeria nassa*, *L. spp.*, *Paramelania damoni*, and *P. iridescens*) for δ¹³C and δ¹⁸O at the University of Rhode Island's Stable Isotope Laboratory. The shell of each snail was sliced longitudinally using a narrow kerf diamond saw, and samples were taken along the growth spiral to provide a sequence of isotope ratio measurements corresponding to roughly equal intervals in the accretionary process of shell growth. On the basis of mark-recapture experiments with *Lavigeria spp.*, all of these taxa are thought to live for at least several years (E. Michel, 1993, personal commun.).

Shell fragments were heated in a vacuum for 1–2 hr at 400 °C to destroy organic matter. Isotope measurements were made in a V.G. Micromass

602D instrument, with NBS-20 standards run at frequent intervals. The analyses were reproducible to ≈0.2‰. Data are corrected for isobaric interferences (Craig, 1957), and both stromatolite and snail data are reported in conventional δ notation relative to the PDB (Peedee belemnite) standard.

MODERN LAKE TANGANYIKA WATER CHEMISTRY

Whereas mollusks, ostracodes, and benthic microbial communities can thrive in water undersaturated with calcite, inorganic cements and the

nonenzymatically precipitated carbonates producing stromatolites will not normally form in undersaturated waters (Reddy et al., 1981; Casanova, 1986). Casanova and Hillaire-Marcel (1992) stated that shallow waters in Lake Tanganyika are “close to saturation with respect to calcite” and they speculated that seasonal calcium carbonate undersaturation precluded the growth of stromatolites in the modern lake. However, Cohen and Thouin (1987) argued, on the basis of the presence of apparently living stromatolites, machine-fabricated cloth fragments as “clasts” in surficial lake carbonates, and micrite rinds around

living *Chara* stems, that carbonates are forming in Lake Tanganyika today. Using the water analyses of Batungwanayo (1987) and Casanova and Hillaire-Marcel (1992) (Table 1), our calculations show that the lake is supersaturated with calcite (saturation state Ω = 6.54 times saturation), magnesian calcite (Ω = 5.9 at 10% Mg, and 3.0 at 20% Mg), and aragonite (Ω = 4.4). Furthermore, undersaturation does not occur seasonally. Monthly measurements of water chemistry at Station Km29 (northern Burundi, close to several stromatolite collection localities of both Casanova and Hillaire-Marcel, 1992, and ours) show similar

TABLE 3. ¹⁴C AGE DETERMINATIONS

Sample	Lab I.D. no.	Location and material*	Depth (m)	Modern carbon [†] (%)	δ ¹³ C (‰ PDB)	δ ¹³ C normalized age (yr B.P.)	Calibrated age cal. yr A.D. (B.C.) [§]	Final calibrated age cal. yr A.D. (B.C.) [#]
<i>University of Arizona analyses</i>								
86-DG-20	A-4782	KM115, Burundi, Stromatolite	30	87.2 ± 0.9	1.9	1100 ± 80	1500 ± 100	1440 ± 100
86-AC-6 (Surface)	A-5085	Kigoma, Tanzania, Stromatolite	12	95.4 ± 0.8	4.6	390 ± 70	Post-1950, ultramodern	Post-1950, ultramodern
86-DG-5	A-4772	Malagarasi Delta region, Tanzania, <i>Neothauma tanganyicense</i> fossil gastropod	20	85.5 ± 0.8	-0.2	1260 ± 70	1340 ± 90	1340 ± 90**
86-AC-1 #13	A-4773	KM26, Burundi, <i>Paramelania damoni</i> fossil gastropod	13	75.7 ± 1.0	-0.9	2240 ± 110	360 ± 125	460 ± 125
86-AC-1 #24	A-4775	KM26, Burundi, <i>Paramelania damoni</i> fossil gastropod	13	74.1 ± 2.0	-0.3	2410 ± 220	190 ± 230	290 ± 230††
86-AC-1A #90	A-4774	KM26, Burundi, <i>Paramelania imperialis</i> live gastropod	18	98.2 ± 1.4	-0.4	150 ± 30	Post-1950, ultramodern	Post-1950, ultramodern
86-DG-18	A-4771	KM115, Burundi, Ooids	4	96.1 ± 0.9	3.3	320 ± 75	Post-1950, ultramodern	Post-1950, ultramodern
<i>Trondheim analyses^{§§}</i>								
87-AC-2	T-9323	KM115, Burundi, Stromatolite	17	83.7 ± 0.5	2.2	1430 ± 50	1170 ± 80	1290 ± 80
87-AC-4	T-9324	KM115, Burundi, Stromatolite	15	83.4 ± 0.6	2.9	1460 ± 60	1140 ± 80	1290 ± 80
87-AC-5	T-9325	KM115, Burundi, Stromatolite	19	83.6 ± 0.8	2.4	1440 ± 80	1160 ± 100	1270 ± 100
87-AC-6 (core)	T-9326	Kigoma, Tanzania, Stromatolite	12	85.8 ± 0.8	3.6	1230 ± 80	1370 ± 100	1400 ± 100
87-AC-7	T-9327	Kigoma, Tanzania, Stromatolite	11	84.7 ± 1.0	3.8	1340 ± 100	1260 ± 120	1300 ± 100
87-AC-8	T-9328	Northern Zaire, Stromatolite	15	88.3 ± 0.7	4.2	1000 ± 70	1600 ± 90	1580 ± 90###
89-MS-211B	T-9331	Nyanza Lac, Burundi, Beachrock	0	75.4 ± 0.8	1.2	2270 ± 90	330 ± 110	430 ± 110
89-MS-142	TUa-253	Ruzizi Delta, Burundi, Ooids	10	78.2 ± 0.7	2.5	1980 ± 70	620 ± 90	680 ± 90
LT-M1	T-9332	Katibili Bay, Zaire, <i>Neothauma tanganyicense</i> collected alive in 1947	5	90.4 ± 0.7	-0.9	820 ± 60	Modern	Modern
<i>Casanova and Hillaire-Marcel (1992) data (modified from their Table 2)***</i>								
TA87-97	UQ1659	KM33, N. Burundi, Stromatolite	7	89.7 ± 1	4.5	1350 ± 200	1250 ± 210	1290 ± 210
TA87-89a	UQ1508	KM33, N. Burundi, Stromatolite	16	87.3 ± 1	3.8	1550 ± 200	1050 ± 210	1170 ± 210
TA87-104	UQ1660	KM33, N. Burundi, Stromatolite	30	84.6 ± 0.9	3.6	1800	800 ± 210	890 ± 210
TA87-105	UQ1661	KM33, N. Burundi, Stromatolite	35	84.0 ± 0.9	3.1	1850	750 ± 210	870 ± 210
TA87-55a	UQ1507	KM33, N. Burundi, Stromatolite	20	83.1 ± 0.6	3.6	1950	650 ± 120	680 ± 120
TA87-117b	UQ1714	KM33, N. Burundi, Stromatolite	10	83.0 ± 1.5	3.4	1950	650 ± 160	680 ± 160
TA87-121a	UQ1589	KM115, S. Burundi, Stromatolite	26	82.6 ± 0.6	2.3	2000	600 ± 120	660 ± 120
TA87-107	UQ1662	KM33, N. Burundi, Stromatolite	10	81.3 ± 1	3.6	2150	450 ± 210	600 ± 210
TA87-55c	UQ1715	KM33, N. Burundi, Stromatolite	20	80.7 ± 1.3	4.8	2200	400 ± 160	540 ± 210
TA87-113	UQ1664	KM33, N. Burundi, Stromatolite	35	79.4 ± 0.8	3.0	2300	300 ± 210	420 ± 210
TA87-110	UQ1663	KM33, N. Burundi, Stromatolite	30	79.3 ± 0.9	3.0	2350	250 ± 210	370 ± 210
TA87-117c	UQ1716	KM33, N. Burundi, Stromatolite	10	79.2 ± 1.3	3.1	2350	250 ± 160	370 ± 160
TA87-40	UQ1504	KM35, N. Burundi, Stromatolite	25	78.4 ± 0.7	3.6	2450	150 ± 120	240 ± 120
TA87-109	UQ1616	KM33, N. Burundi, Stromatolite	25	78.0 ± 0.5	4.0	2500	100 ± 120	140 ± 120
TA87-55d	UQ1506	KM33, N. Burundi, Stromatolite	20	76.4 ± 0.6	5.8	2650	(150 ± 120)	(100 ± 120)
TA87-117d	UQ1588	KM33, N. Burundi, Stromatolite	10	71.8 ± 0.5	5.5	3150	(550 ± 120)	(580 ± 120)
TA87-117e	UQ1713	KM33, N. Burundi, Stromatolite	10	69.3 ± 1.3	3.5	3400	(800 ± 160)	(880 ± 160)†††

*See Figure 1 for more precise locality information.

†Normalized to δ¹³C = -25‰.

§Calendrical calibration corrected for old carbon and Suess effect discussed in text (i.e., 1950-[A - 650 yr]) incorporating square root of sum of squares summation of standard errors.

#Final calibrated age using dendrochronologic correction factors of Stuiver and Pearson (1993).

**Multiple calendrical age solutions exist for this radiocarbon age. This date could also be 1300 or 1360. The middle value (1340 A.D.) is shown.

††Multiple solutions exist. 260 or 320 A.D. are also possible.

§§Sample 89-MS-142 (lab ref. no. Tua253) was analyzed by the Uppsala AMS facility on behalf of the Trondheim laboratory.

###Multiple solutions exist. 1520 or 1620 A.D. are also possible.

***Percent modern carbon values given for Casanova and Hillaire-Marcel (1992) data are presented as in their original table, without ¹³C correction. All other percent modern carbon values are given with ¹³C correction.

†††Multiple solutions exist. 830 or 920 B.C. are also possible.

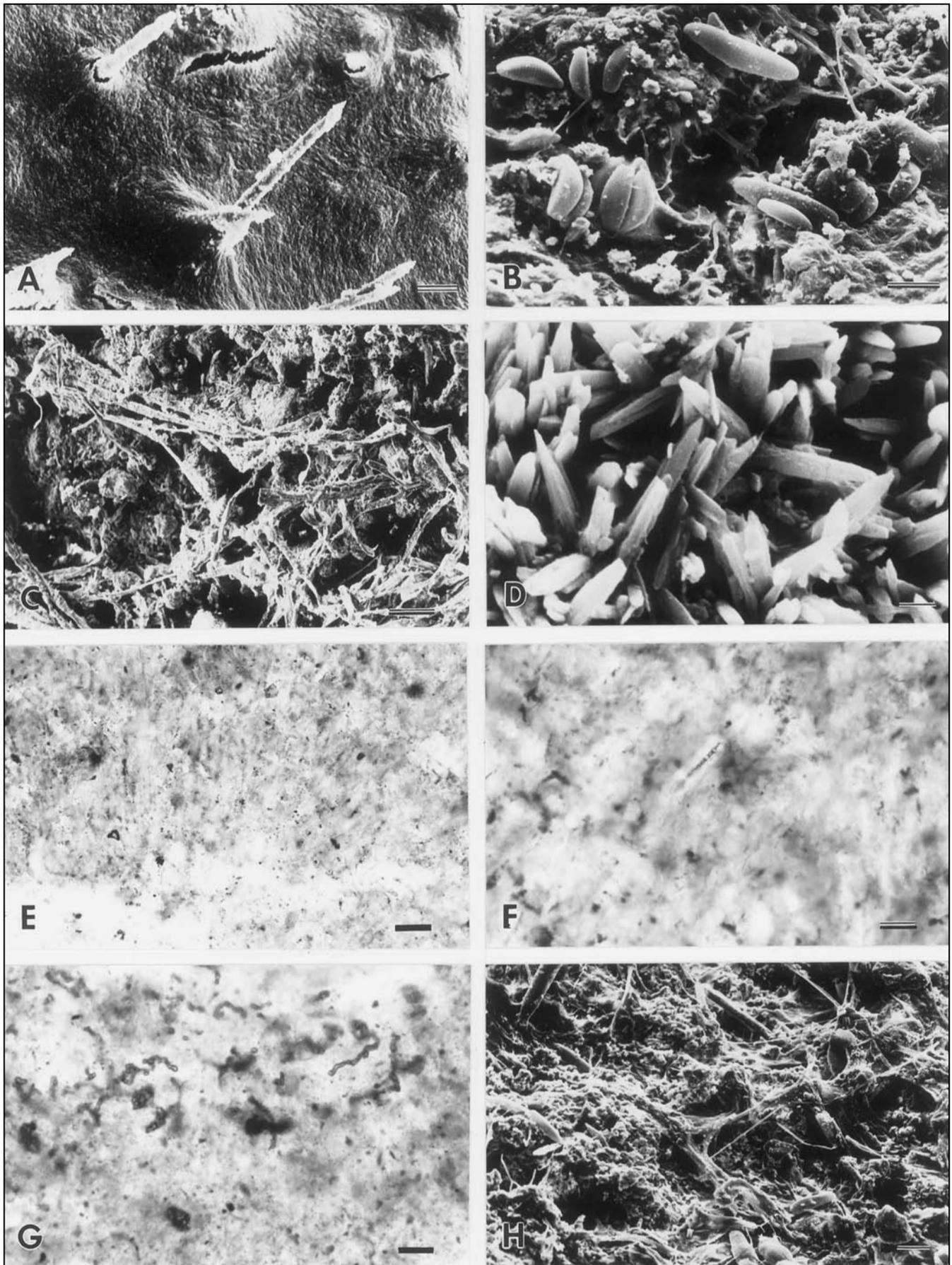


Figure 2. (Caption on facing page).

Figure 2. Photomicrographs of stromatolites from Lake Tanganyika. (A) Scanning electron microscope (SEM) image of encrusting sponges on the surface of stromatolite 87-5, from Km115. Bar scale = 10 μm . (B) SEM image of diatom community on surface of stromatolite 87-6c from Kigoma. Note abundant carbonate silt trapped on filaments. Bar scale = 10 μm . (C) Micrite trapped and bound between algal filaments on upper surface of stromatolite surface crust. Stromatolite 87-6a from Kigoma, Tanzania. Bar scale = 20 μm . (D) Aragonite needle mass on undersurface of surficial, unconsolidated crust of stromatolite 87-6b from Kigoma, Tanzania. Bar scale = 1 μm . (E) Radial arrangement of filament molds of approximately the same diameter (about 3 μm in diameter) in thin section of stromatolite from Km115 (sample 87-4). Bar scale = 25 μm . (F) Large molds of filaments with some remains preserved from a stromatolite at Kigoma. These filament molds are variable in diameter and often found in radial clusters; however, they are usually poorly preserved. They differ from the radial cluster in 2E in size and diameter variability. Bar scale = 25 μm . (G) Probable endoliths from stromatolite at Km115 locality. These filaments are somewhat tuberous and sinuous. Bar scale = 25 μm . (H) Microbial filament mass trapping carbonate particles and sponge spicules on upper surface of stromatolite crust. Stromatolite 87-6a, from Kigoma. Bar scale = 20 μm .

saturation states throughout the year (Batungwanayo, 1987). Because Lake Tanganyika waters are supersaturated with the relevant calcium carbonate phases, water chemistry does not limit modern stromatolite formation. This is consistent with the finding of ooids with ultramodern ^{14}C ages (sample 86-DG-18, Table 3).

MICROBIAL ASSEMBLAGES OF LAKE TANGANYIKA STROMATOLITES

Documentation of living stromatolite microbial assemblages and stromatolite morphology is critical for the paleoecological interpretation of fossil stromatolites. Tanganyikan stromatolites are covered by a complex community of benthic organisms, including filamentous and coccoid cyanobacteria, diatoms, green algae, fungi, and sponges (Fig. 2, A and B). The surficial filamentous cyanobacteria present are dominated by *Scytonema*, *Phormidium*, and *Microcoleus*. A diverse assemblage of diatoms (104 species) has been documented from the surfaces of fossil stromatolites in southern Burundi (Cocquyt, 1991), but these have not yet been compared with the diverse diatom floras of growing stromatolites. Chrysophyte cysts and nonmotile green algal zygotes are also incorporated into the surficial carbonate of living Kigoma stromatolites. Encrusting sponges are dominated by two common species, *Spongilla moorei* and *S. wivrae*.

At the Kigoma site, algal and cyanobacterial filaments bind loose particles of micrite, larger carbonate allochems, and clastic grains on stromatolite surfaces (Fig. 2C), similar to the stromatolites of Cuatro Ciénegas, Coahuila State, Mexico (Winsborough and Seeler, 1984). Although we cannot conclusively demonstrate the presence of crystalline calcite within the algal sheaths, aragonite needle masses can be clearly seen within the unconsolidated outer crusts of the stromatolite

heads (Fig. 2D). At present we do not know which species of cyanobacteria, algae, or sponges are involved in trapping and binding the bulk of the allochems. The cyanobacteria and diatoms have mucilaginous sheaths. Within the carbonate laminae and columns, at least two basic categories of microbial fossils are present: (1) small, sinuous, locally branched filaments (probably endoliths) at Km115 and Kigoma, and (2) straight, unbranched filaments (preserved microbial remains of constructing microbes) found only at Kigoma. The straight filaments fall into three size categories (they are all of similar shape and other morphological attributes). The smallest size class consists of straight to slightly curved filaments, about 1.5 to 3 μm in diameter, that have thin walls and are filled with microspar. These filaments often are found singly and not in clusters. In places, they are found in radial masses (Fig. 2E), which is not uncommon in other lacustrine stromatolites (e.g., Monty and Mas, 1981). Although not well preserved, they resemble filamentous cyanobacteria. The next two classes of filaments are found both in clusters or masses, oriented normal to laminae and resemble radial sparry fabric composed of microbial remains as described by Monty and Mas (1981, p. 105–116). The smaller class consists of small, straight, thin-walled, unbranched filaments, 3–4.5 μm in diameter, composed of spar. The larger class consists of straight, thin-walled, unbranched filaments, 9–22 μm in diameter, composed of spar (Fig. 2F). The larger class filaments are not common in any of the stromatolites examined.

Compelling evidence for microbial precipitation of calcium carbonate as a mechanism for stromatolite formation is lacking. Nowhere have we observed calcium carbonate encrusting microbes or in their sheaths and/or mucilage. *Phormidium* and *Scytonema* are known to calcify their sheaths (Pentecost and Riding, 1986) and are

present in our material; however, no carbonate in or on sheaths was observed in the microbes in our samples. Microbes collected from the surfaces of stromatolites were preserved in unbuffered formalin, which may have resulted in formic acid dissolving any sheath-associated carbonate. The absence of calcium carbonate encrusting microbes or in their sheaths and/or mucilage does not rule out the contemporaneous precipitation of calcium carbonate external to the microbes and their sheaths, as evidenced by Figure 2C, where micrite is enmeshed with filaments, and Figure 2D, where aragonite needles are found on the stromatolite's surface.

Probable endoliths are common in Kigoma stromatolites but are rare at Km115 (Fig. 2G). They are represented by small, sinuous filaments that are 1.4–4.8 μm in diameter. They appear to be spiral with no well-defined wall, and are composed of spar. The filament diameter is highly variable over short distances. Branching is rare and normal to the main filament. These filaments are randomly oriented. Filaments commonly have a very bright appearance in transmitted light. Because of their brightness, sinuous habit, nonuniform diameter, and branching, these filaments resemble endoliths; however, convincing examples have not been found in ooids or skeletal grains trapped in the stromatolites. Only the stromatolites have these filaments. No cellular remains are preserved. No terminal swellings or other structures resembling reproductive structures have been observed. The sinuous filaments exhibit an abrupt, short bend or kink at the region of curvature, similar to patterns produced by endoliths boring into calcite crystals (see Golubic, 1969, Figs. 3–5). Similar filaments from stromatolites of the Aldabra Islands were assigned to the cyanobacterium *Hyella* (Casanova, 1986).

Stromatolite Distribution and Morphological Features

Figure 1 and Table 2 illustrate the localities, depth ranges, and substrates where we have identified extensive stromatolite buildups. The distribution is almost certainly incomplete because much of the lake's littoral-sublittoral zone is unexplored. Many of these stromatolites are fossils, commonly encrusted by epiphytic and epilithic microbial communities but not actually depositing laminated or massive carbonate (Casanova and Thouin, 1990). Other stromatolites, however, appear to be actively growing, as evidenced by the presence of carbonate within the filament mass, stable isotope data, ^{14}C age dates, and internal morphological features discussed below.

We have observed stromatolites at depths of 6 to 46 m below the lake level of the early 1980s (775 m above sea level [masl]). The maximum

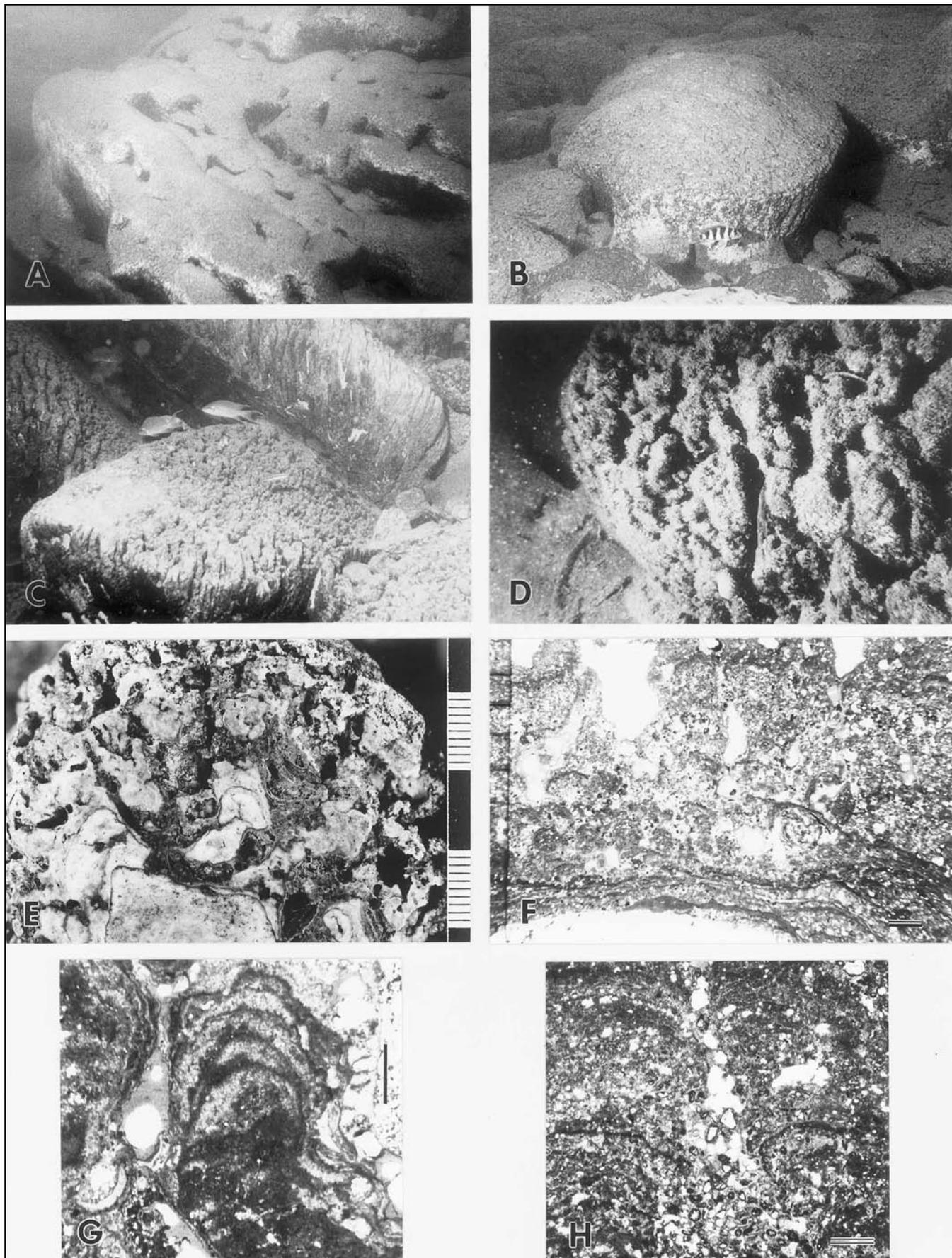


Figure 3. (Caption on facing page).

Figure 3. Stromatolites from Lake Tanganyika. (A) Large stromatolite promontories forming along rocky ledges at Km115, 20 m water depth. Deep, vertical fissures mentioned in text appear as dark, sinuous areas on the stromatolite's upper surface. Horizontal field of view is ≈ 2 m. (B) Domal stromatolites at Kigoma, 10 m water depth. Large dome in center of photo is ≈ 0.7 m wide. (C) Stromatolite encrustations and small columns growing over rocky blocks at locality 7, northern Burundi, 15 m water depth. Larger fish in center are ≈ 10 cm long. (D) Close-up of individual stromatolitic columns at locality 9, northern Burundi, 8 m water depth. Horizontal field of view is ≈ 20 cm. (E) Columns and columnar branching stromatolites encrusting cobble in sample 87-7, Kigoma site, 11 m water depth. Small scale divisions in millimeters. (F) Variable nature of mesostructure and microstructure of stromatolite from Km115. Small, poorly laminated columns formed on laminated fabric that encrusted crystalline rock, that in turn grade vertically into nonlaminated, peloidal carbonate. One millimeter horizontal scale in lower right of photomicrograph. Additional millimeter ruler scale on left. (G) Predominantly laminated mesostructure and microstructure of small stromatolite column from the Kigoma locality. Dark laminae are composed of micrite and light laminae are composed of microspar. Individual laminae are of variable thickness. Bar scale to right = 1 mm. (H) Coated grain and peloidal allochems infilling the interspace between small stromatolite columns. Stromatolite 87-2, Km115, 17 m water depth. Bar scale = 5 mm.

depths are in part artifacts of depth limitations on SCUBA divers, although on recent remotely operated vehicle dives to a maximum of 100 m at several sites, one of us (Cohen) did not see stromatolites at depths greater than 40 m. Echo sounder profiles over stromatolite localities indicate that stromatolites may be found as deep as 60 m. The upper limit of stromatolites is well established, except for oncolites, which we have found at the shoreline. We have rarely encountered stromatolites above 10 m below current lake level. However, the presence of littoral (0–3 m depth) snails commonly imbedded in small columnar stromatolites and stromatolitic encrustations (discussed later herein) may indicate that stromatolites have formed in shallower water in Lake Tanganyika's past.

In Lake Tanganyika stromatolites are extremely rare on delta fronts or adjacent to large rivers and are normally restricted to growth on hard substrates. This is in contrast to their common occurrence in the siliciclastic deposits of some Cenozoic paleolakes (Link et al., 1978). However, at several localities (Table 2) we have observed stromatolites growing in clusters on unlithified shelly or pebbly sands, similar in some aspects to those described by Straccia et al. (1990) from Miocene rift lake deposits of Idaho.

At both the Km115 and Kigoma sites stromatolites are mostly restricted to rocky substrates. On continuous rocky ledges or boulders stromatolites may grow over both horizontal or vertical faces, with concordant growth morphologies (Fig. 3A). At the Km115 site buildups have also formed on previously lithified beachrock or oolite.

The Tanganyikan stromatolites are highly variable in growth form (Fig. 3, A, B, and E). Shapes include stratiform encrustations on boulders and vertical walls, columnar-layered structures, col-

umns (Fig. 3E), branching columns, and oncolites. Several of these structures are found within bioherms that coalesce and are meters across (Fig. 3A). Columnar, columnar branching, columnar layered, and stratiform complexes, several centimeters in thickness, form on exposed bedrock or on cobbles. The oncolites (5–20 cm in diameter) occur commonly at water depths between 0 and 30 m. They typically develop as laminated encrustations around large mollusk shells or pebbles.

The stromatolites are described following guidelines proposed in Grey (1989). Four observational levels, megastructure, macrostructure, mesostructure, and microstructure, are used in these descriptions. Characteristic depth information is incorporated into this description.

Megastructure. The large-scale configuration of stromatolites forming a bed, or in a bed, is termed megastructure (Grey, 1989). Lake Tanganyika stromatolites form in isolated patches, as biohermal to biostromal aggregates of domes and columns several tens of meters across and < 1 m thick on rock ledges (Fig. 3B).

Macrostructure. Macrostructure is the shape and size of stromatolitic bioherms and biostromes in a bed (Grey, 1989). The bioherms or biostromes are variable in size and consist of large coalesced and discrete domes and columns, with rounded, gently convex to flat upper surfaces. Individual domes and columns can exceed 1 m in diameter, and most are < 1 m tall. Below depths of 12 m, large, individual stromatolites consisting of flat to gently convex topped bulbous columns, both linked and isolated, commonly occur (Fig. 3B). This shape appears to be a function of depth. Shallower water stromatolites tend to have a shape that conforms to the shape of the underlying rock (suggesting they are thin veneers; Fig. 3C). Deeper water structures assume their own shape regardless of

their substrate shape. Where bulbous columns are linked, the stromatolites form continuous ledges (at Kigoma site) or anastomosing ridges (at Km115).

The largest stromatolites encountered to date in Lake Tanganyika are found at depths > 30 m. At site Km115, stromatolite bioherms several meters across and > 1 m in height are present. These structures are extremely well indurated and have a variety of shapes, including large, narrow spurs and irregular promontories. Smaller, flat to gently convex-topped, bulbous columns similar to those found in shallower water continue below 30 m; however, they are restricted to horizontal ledges between the larger bioherms or on rock outcrop ledges.

Deep water stromatolites, in particular those in bioherms and biostromes, exhibit a different surface topography than shallower forms at Km115. Below ≈ 20 m, deep vertical fissures appear on the surfaces of many stromatolites. From 20 to 26 m these fissures increase in width and penetration with depth, and their edges become sharper and better defined. Below 26 m, almost all stromatolites have this fissured topography. At this depth, individual fissures average 5–10 cm across, and may be > 20 cm deep (most taper, so their true depth into the stromatolite is difficult to determine), and up to 1 m long in plan view. The fissures (1) are entirely contained within the surface of the stromatolite, (2) are present at the edge and terminate toward the center of the stromatolite's surface, or (3) completely bisect the surface. The fissures have rounded edges and their ends taper. The term "cereboid" has been used for this morphology (Casanova and Thouin, 1990; Casanova and Hillaire-Marcel, 1992).

In addition to the fissuring, deep water bioherms differ from shallow water bioherms in possessing much smoother, yet crenulated surfaces (possibly as a result of greater growth rates in shallower water). Small columns continue into deep water, but are uncommon, and are found only on small flat-topped to gently convex-topped bulbous columns confined to horizontal ledges.

Domes and bulbous columns are commonly composed of smaller domes and columns, some of which branch. Columns are generally 3 cm or less in diameter and vary irregularly in diameter from base to top. Columns commonly originate from basal, thin, stratiform laminae. Branching is complex and one sample can include furcate, umbellate, digitate, dendroidal, and anastomosed styles (see Hofmann, 1969, Fig. 13, for terminology). In plan view the columns are circular to elongate. The interspace area (the region between columns) approximately equals column diameter. Some columnar stromatolites have laminar, calcareous, white crusts of unknown origin that coat and cement boulders.

Mesostructure. Mesostructure refers to features intermediate between macrostructure and microstructure (Grey, 1989). In laminated structures, it describes the nature of the lamina, and in nonlaminated structures, mesostructure refers to the complex nature of the clotted, internal fabric. Unlike many Holocene and older lacustrine stromatolites (e.g., Holocene Walker Lake, Osborne et al., 1982; Pliocene Ridge basin, Link et al., 1978; Oligocene Marseilles rift, Casanova and Nury, 1989), most Tanganyikan stromatolites lack well-defined dark and light laminae (e.g., Fig. 3H).

Two types of mesostructure are found: one, moderately well laminated, is found at Kigoma, and the other, poorly laminated, is present at Km115. The poorly laminated stromatolites cannot be classified as thrombolites because they do not contain a microscopic clotted fabric, even though some are unlaminated (Aitkin, 1967). Stromatolites from Km115 are variable at the mesostructural level and consist of small columns, 3 cm or less in diameter, with a mesostructure consisting of no laminae, poorly defined thick laminae, and rare, moderately well-defined thin laminae (Fig. 3F; described below in more detail). All mesostructures can coexist in the same column over a distance of <1 cm along the axis of the column. The moderately well-defined thin laminae commonly are found at about the same level in two or more adjacent columns possibly indicating similar, synchronous processes and/or conditions. When viewed in tangential section, the better defined laminae are commonly gently convex, extending across the column. They also have moderately well-defined upper and lower boundaries, a moderate to high degree of inheritance (i.e., the degree to which a lamina follows the shape of underlying lamina; Hofmann, 1969), and only first-order curvature (Fig. 3G).

Some parts of the stromatolites have a somewhat radial fabric formed by molds of small, 3–5- μm -diameter filaments (Fig. 2E). The filaments are unbranched and have poorly defined, gray micritic walls that commonly merge imperceptibly with the gray mass of micrite that is present between the filament molds; the molds are filled with light gray to clear microspar. This fabric is rare, poorly preserved, and has few endoliths associated with it. Because of the poor preservation of mesostructure and microstructure within the Tanganyikan stromatolites, it is uncertain if this microbial fabric is common.

Although the stromatolites at Kigoma are better laminated (Fig. 3G), they have a complex mesostructure. Laminae are thick and variable in thickness within the column and along the same lamina. Second-order convexities are present. These second-order expressions of the laminae are crinkled, and a mixture of corrugate (small noncontiguous domes) and cumulate (contiguous

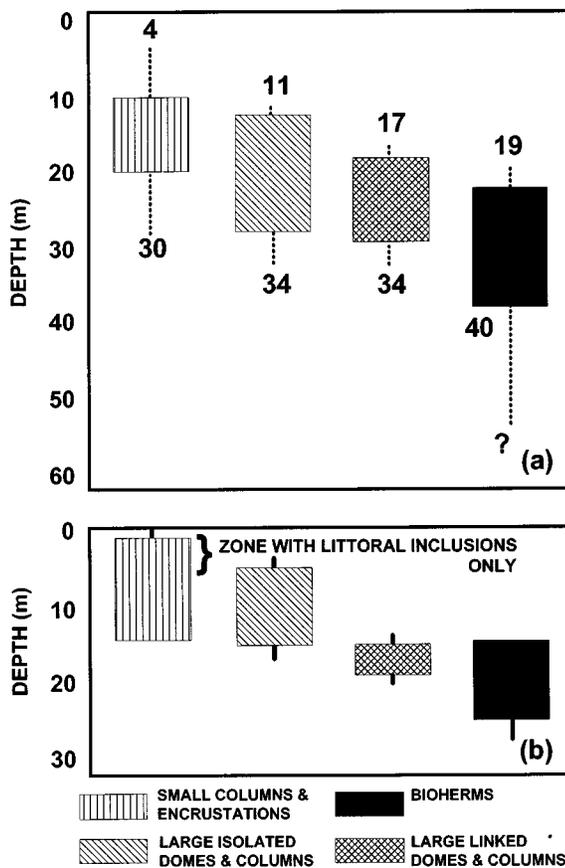


Figure 4. (a) Present depth ranges for stromatolite growth morphologies. The numeric values are the upper and lower depth limits from which we have retrieved stromatolites of each type in Lake Tanganyika. The heavy bars represent the depth ranges where the particular stromatolite morphology is most common. Bioherms have not been observed by SCUBA divers below 40 m. However, echosounder profiles taken from the same areas show bottom configurations that are suggestive of stromatolites at depths up to 55 m. (b) Probable original growth ranges for stromatolite morphologies. These inferred depth ranges are based on relative relationships to shoreline beachrock ridges, the presence of shells with known depth ranges that are incorporated into the stromatolites as allochems (note zone of littoral inclusions in small columns and encrustations), and depth-indicative sedimentary structures associated with stromatolites. The heavy bars are the ranges for which we have greater confidence, whereas the narrow bars represent possible greater extensions of vertical range.

small domes) forms within the same first-order lamina. Inheritance varies from low to high, but most laminae have a low to moderate degree of inheritance.

Stromatolites from both sites lack banding or macrolamination. Macrolamination, a higher order banding pattern produced by a grouping of laminae, is a common feature in lacustrine stromatolites (e.g., Johnson, 1937; Donsimoni and Giot, 1977) and its absence may reflect relatively uniform conditions in the lake at the site of stromatolite formation throughout the year.

Microstructure. Stromatolite microstructure refers to “the fine scale structure of the lamination, in particular the distinctness, continuity, thickness, and composition of the laminae” (Preiss, 1972, p. 93). Stromatolites from the Km115 site are composed primarily of peloids and coated grains (Fig. 3H). Where laminated, dark laminae are predominantly micrite (with peloids and other grains) and light laminae are microspar (with fewer peloids and other grains). The thin, well-defined, dark-light laminae discussed above have individual laminae of approximately equal thickness (dark lamina 18–60 μm ; \bar{X} = 34 μm thick, n = 17; light lamina 19–70 μm , \bar{X} = 33 μm thick, n = 17). The boundaries between the

laminae are moderately well defined, but not as sharp as in the Kigoma stromatolites. The thickness of the thickest laminae is more difficult to measure because boundaries are diffuse; however, thicknesses are on the order of 100–150 μm for dark laminae and 80–100 μm for light laminae. Grains in interspaces are of comparable size (40–100 μm), consisting of peloids, coated grains, invertebrate fragments, and mineral grains presumably from the underlying bedrock.

The laminae at Kigoma are better defined; however, they are still not as well defined, smooth, and sharply bounded as those commonly found in other lacustrine stromatolites (e.g., Bertrand-Sarfati et al., 1994). Dark laminae are thinner than the light laminae (59–965 μm ; \bar{X} = 226 μm , n = 42), and are composed of micrite and microspar. Light laminae are thicker (111–1210 μm thick; \bar{X} = 370 μm , n = 39) and are composed of microspar. Radial fibrous calcite is rarely observed in light laminae.

Lake Tanganyika stromatolites differ from other fossil and recent lacustrine stromatolites in the following features: (1) they commonly lack well-defined laminae; (2) light laminae rarely have radial fibrous textures; (3) macrolamination is lacking; and (4) second-order curvature (i.e., additional curves in the laminations across the

columns; Fig. 3G) is common. The reasons for these differences are unclear. The lack of macrolamination may result from the low degree of chemical and thermal seasonality undergone by the lake. This may result in only minimal seasonal changes in benthic microbial communities and other organisms that might affect stromatolite growth.

Nonmarine stromatolite growth morphology is a potentially useful tool for paleobathymetric interpretation. However, as Hillaire-Marcel and Casanova (1987) noted, our current knowledge of the actual depth range limits for particular growth forms is severely limited. Casanova and Thouin (1990) contended that it was impossible to relate growth morphologies to depth in Lake Tanganyika, and that the aspect of the substrate (horizontal, sloping, or vertical) was the primary determinant of differences in stromatolite form. Our data from a large number of localities show, however, that some useful paleobathymetric inferences may be extracted from fossil stromatolites in Lake Tanganyika. Figure 4 illustrates the present depth ranges at which we have found different growth morphologies of stromatolites. Because lake level has fluctuated during the late Holocene, we have estimated the original growth depths for individual stromatolites by one or more of the following approaches: (1) measuring the depth ranges of specimens that appear to be growing (see Methods section herein), (2) examining fossil gastropods that are associated with the stromatolites whose modern depth ranges are known, and (3) correlating stromatolite belts with clearly associated ancient strand lines, beachrock ridges, or sedimentary structures that are depth indicators. Figure 4 shows our estimates of original growth depths for stromatolites based on approaches 1–3. Stromatolites have grown during the late Holocene in Lake Tanganyika down to depths of at least 26 m, which closely corresponds to the lake's present photic zone (28 m as defined by the 1% light penetration method; Hecky, 1991).

¹⁴C GEOCHRONOLOGY

Our ¹⁴C results are presented in Table 3. We have normalized the initial ¹⁴C activity based on $\delta^{13}\text{C}$ measurements of the carbonates. It was also necessary to correct our age data for both atomic bomb radiocarbon effects and the effect of surficial mixing of old carbon derived from the lake's hypolimnion (cf. Craig et al., 1974). Casanova and Hillaire-Marcel (1992, p. 40) argued that ¹⁴C age dates from stromatolites in Lake Tanganyika can be considered "true" ages (i.e., without hard water effects) after normalizing the ratios for isotopic fractionation (ca. 450 yr B.P.) because the $\delta^{13}\text{C}$ of dissolved inorganic carbon (DIC) in surface water seems to be in isotopic equilibrium

with atmospheric CO_2 . They also pointed to ¹⁴C measurements of surface water in 1987 (102% \pm 7% modern, $n = 6$) as demonstrating atmospheric equilibrium. We believe that these surface water data show that a large reservoir correction is necessary, because of the absence of ¹⁴C enriched bomb carbon. Surface water seems to hold nearly equal portions of deep water carbon (89% modern; Harmon Craig, 1993, personal commun.) and atmospheric carbon (atmospheric $\text{CO}_2 = 115\%$ modern in 1987, interpolated from Nydal and Lövseth, 1983; in solution, HCO_3^- should be $\approx 117\%$ modern).

The reservoir effect can be estimated from the ¹⁴C composition of a snail (*Neothauma tanganyicense*) collected live in 1947 from 5 m depth in the lake (Katibili Bay, Zaire). Its age (after $\delta^{13}\text{C}$ correction) was 820 ± 60 B.P. ¹⁴C yr (Table 3, sample LT-M1). Carbonate of this age in atmospheric equilibrium is expected to be depleted in ¹⁴C by 2.1% (Stuiver and Pearson, 1986), giving an apparent age of 170 yr B.P. at the time of formation. The measured age of this snail shell is therefore 650 ± 60 yr too old ($\approx 7.8\%$ initial ¹⁴C). Following the marine carbonate calibration protocol of Stuiver et al. (1986), we adjusted our age dates for this 650 ± 60 yr residual old carbon effect. Correcting for the reservoir effect in the ¹⁴C content of a shell collected live in 1986 (Table 3, sample 86-AC-1A #90, $\delta^{13}\text{C}$ -corrected, apparent age = 150 ± 30 ¹⁴C yr B.P.) yields a ¹⁴C content of about 112% modern, consistent with the presence of bomb carbon in the shallow waters. The expected ¹⁴C concentration of shell aragonite in water exposed to 1986 air is 119% modern. This value is calculated by applying a fractionation of 1.021 (twice the ¹³C/¹²C fractionation between CO_2 gas and aragonite of 10.5‰ [26 °C], Romanek et al., 1992) to the atmospheric CO_2 composition of 117% modern. This discrepancy may result from the fact that we were only able to obtain a single prebomb live snail to date, coupled with the variability in Casanova and Hillaire-Marcel's (1992) water ¹⁴C values. After applying the reservoir effect correction, we converted radiocarbon years to calendar years using the dendrochronologic corrections of Stuiver and Pearson (1993). We summed all errors using the square root of the sum of squares method. Although studies by Vogel et al. (1993) have shown that ¹⁴C age dates from the mid-latitudes of the Southern Hemisphere are systematically ≈ 40 calendar yr too old (because of more efficient atmospheric mixing with the much larger ocean surface of the Southern Hemisphere), the implications of this work for the equatorial regions remain unclear. Therefore we have not applied this correction to our data.

We have recalculated ages for Casanova and Hillaire-Marcel's stromatolite data using the

same carbon reservoir corrections. This recalculation implies that the events described in Casanova and Hillaire-Marcel's (1992) lake level and isotope chronology (as illustrated in their Figs. 5, 6, 8, and 9) are ≈ 600 –700 yr younger than was originally proposed. Our ¹⁴C data (and renormalized Casanova and Hillaire-Marcel data) indicate that stromatolite formation and carbonate formation has continued in Lake Tanganyika throughout the late Holocene. The surficial carbonate on stromatolite 87-AC-6 yielded an ultramodern age, and was presumably growing at the time of collection.

LAKE-LEVEL HISTORY

Integrating historical data with paleodepth estimates from stromatolites, fossil gastropods incorporated in stromatolites, and wave-cut terraces, we have produced a preliminary lake-level curve for Lake Tanganyika for the past 3000 yr (Fig. 5). This figure combines our data with our ¹⁴C-normalized interpretations of Casanova and Hillaire-Marcel's data (derived from their Tables 2 and 3 and Fig. 5).

The combined data indicate that the total amplitude of lake-level fluctuations has been small throughout this time period (≈ 20 m). Lake levels were 5–10 m below present during the earliest interval studied (ca. 800 B.C. to A.D. 400). This minor lowstand correlates with generally low lake levels in East Africa between 3000 and 2000 yr ago (Hecky, 1971; Butzer et al., 1972). However, we do not see evidence for the considerably lower (≈ 75 m) lake levels that Hecky and Degens (1973) proposed (based on hydrologic considerations) for Lake Tanganyika prior to the Lake Kivu overflow event. This shows that Ruzizi inflow may have been less critical for maintaining Lake Tanganyika's water level during the late Holocene than the Hecky and Degens (1973) model would suggest. Vincens (1989) demonstrated a decrease in arboreal pollen and an increase in grass pollen at ca. 500 B.C. in cores from the northern part of Lake Tanganyika, and noted that this might represent a shift to more arid conditions, or the onset of deforestation. Both the lake-level curve and isotopic data (discussed below) suggest that the lake was closed at this time, indicating a relatively dry climate. Beachrock deposits dated at A.D. 430 ± 110 that are found close to the current lake level (775 to 776 masl) indicate a significant lake-level rise at this time. This rise may have resulted in a brief opening of Lake Tanganyika. Lake levels appear to have returned to the 765–770 masl range during the seventh century. We have no direct evidence of lake-level fluctuations for the next ≈ 500 yr. However, lake levels could not have fallen below 750 masl, because there is no evidence for exposure during this interval in cores

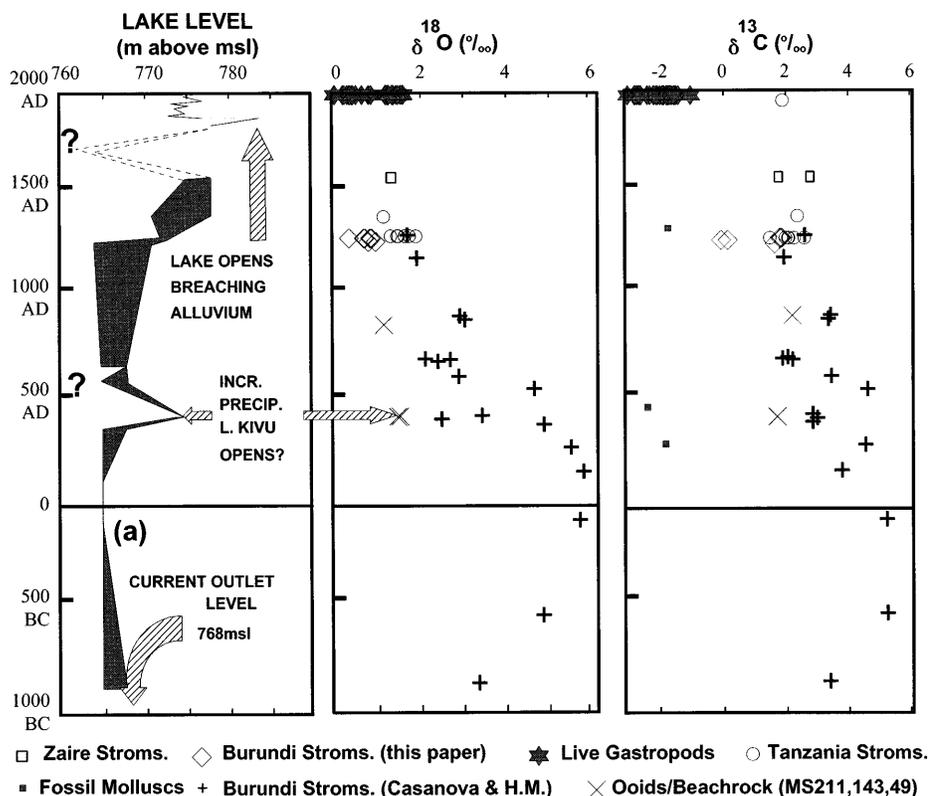


Figure 5. Lake Tanganyika lake level and stable isotope chronologies. Data are derived from Tables 3 and 4 (this paper) plus the isotopic and stromatolite data of Casanova and Hillaire-Marcel (1992), whose ^{14}C age estimates have been modified using correction factors discussed in the text. Question marks indicate probable lower stands of lake level, the elevations of which cannot be accurately constrained, except that they were not lower than 740 m. Solid narrow lake-level lines represent historic data. Dashed lines are possible fluctuations based on core and shell lag data. Shaded zones represent maximum and minimum lake-level estimates based on stromatolites. The probable Lake Kivu opening at ca. A.D. 500 is probably unrelated to the trend toward lower $\delta^{18}\text{O}$, which began in the fifth century. All plotted isotopic values are corrected for carbonate mineralogy (see Fig. 7 caption).

collected in 25 m water depth. The stability of oxygen isotope values during this time (discussed below) argues against any major lake-level fall. Winnowed, deep water shell lags that postdate the late fifth century A.D. (sample 86-AC-1, #13) indicate that lake levels may have fallen to the 760 masl level, but the minimum age of this event cannot be constrained.

Lake levels began to rise in the mid-thirteenth century. A second winnowing event occurred some time after the late sixteenth century, but prior to the early nineteenth century, and resulted from a significant lake-level drop. This fall is indicated by a shell dated to the fourteenth century (sample 86-DG-5) that was incorporated into a lag 20 m below the current lake level at some time after the phase of stromatolite growth during the fifteenth to sixteenth centuries, but before the beginning of the nineteenth century historic record. Evidence for this minor lake-level fall also comes

from shallow water sediment and ostracode indicators in several sublittoral cores (Park and Cohen, 1994; Wells and Cohen, 1994). However, we have no precise date for this lake-level fall. On the basis of the presence of dated shell lags at ≈ 20 m (which must have formed above ≈ 5 m wave base, and may have been exposed) and an absence of exposure surfaces in a core taken in 40 m water depth, this lake lowstand can only be bracketed at between 735 and 760 masl. At the onset of historical data (mid-nineteenth century), lake levels were close to those we observed in our youngest premodern radiocarbon dated materials (late sixteenth century), about 775 masl. Lake levels rose during the mid-nineteenth century, culminating in the 1877 opening of the lake. This was followed by a rapid downcutting of the Lukuga River outflow channel through alluvium, and the relative stabilization of lake level between 772 and 777 masl for most of this century (Evert, 1980). The

≈ 775 m level has evidently been revisited numerous times during the past 1300 yr.

The Lake Tanganyika record shows some similarities to the flood records of the Nile, showing that even these modest lake-level changes may be a response to regional climate change. Lowstands in Lake Tanganyika between the seventh and fourteenth centuries accord with the interpretations of low White Nile discharge during most of this period (Hassan, 1981). The only major period of high Nile floods during this time (twelfth century A.D.) is an interval for which we have no data in Lake Tanganyika. Decadal records of Nile floods indicate a major increase in discharge starting in the fourteenth century and continuing until the nineteenth century, which corresponds to the timing of lake-level rise in Lake Tanganyika. Similarly, the highstand of Lake Tanganyika during the mid-late nineteenth century is in close agreement with both White Nile discharge data and regional lake-level data (Butzer, 1971; Maley, 1973; Nicholson, 1976; Hassan, 1981).

These observations suggest that Central Africa was dry for much of the period between the ninth century B.C. and the fourteenth century A.D. A significant wet interval may have interrupted this dry period during the fifth to seventh centuries A.D. (discussed later herein). Starting in the fourteenth century A.D., a gradual onset of wetter conditions resulted in rising lake levels that accelerated during the mid-nineteenth century.

STABLE ISOTOPE HYDROLOGY OF LAKE TANGANYIKA AND THE TANGANYIKA-RUZIZI-KIVU SYSTEM

Lack of comprehensive rainfall data for the Tanganyika catchment precludes a detailed discussion of the isotope hydrology of the Tanganyika basin, but sufficient information is available to provide some guidelines for the interpretation of our results. The lake is characterized by an ^{18}O -enriched hypolimnion ($d \geq 200$ m; $\delta^{18}\text{O} \approx 4.2\text{‰}$ SMOW (standard mean ocean water); Craig et al., 1974; Tietze, 1981, 1982) and well-mixed, less ^{18}O -enriched surface waters ($d \leq 75$ m; $\delta^{18}\text{O} \approx 3.7\text{‰}$; Craig et al., 1974; Tietze, 1981, 1982). Between these two water masses lies a transition zone about 150 m thick, where the isotopic composition increases monotonically with depth as a result of mixing between surface and deep waters. Isotopic and other evidence shows that the hypolimnion may be a relict water mass formed during drier, cooler climatic conditions (Craig et al., 1974). The lake waters as a whole form a well-defined δD - $\delta^{18}\text{O}$ mixing line with river-supplied inflow (Craig et al., 1974; Fig. 6). This line is geometrically coincident with and statistically identical to a regression line computed from the lake data alone ($n = 83$, $r =$

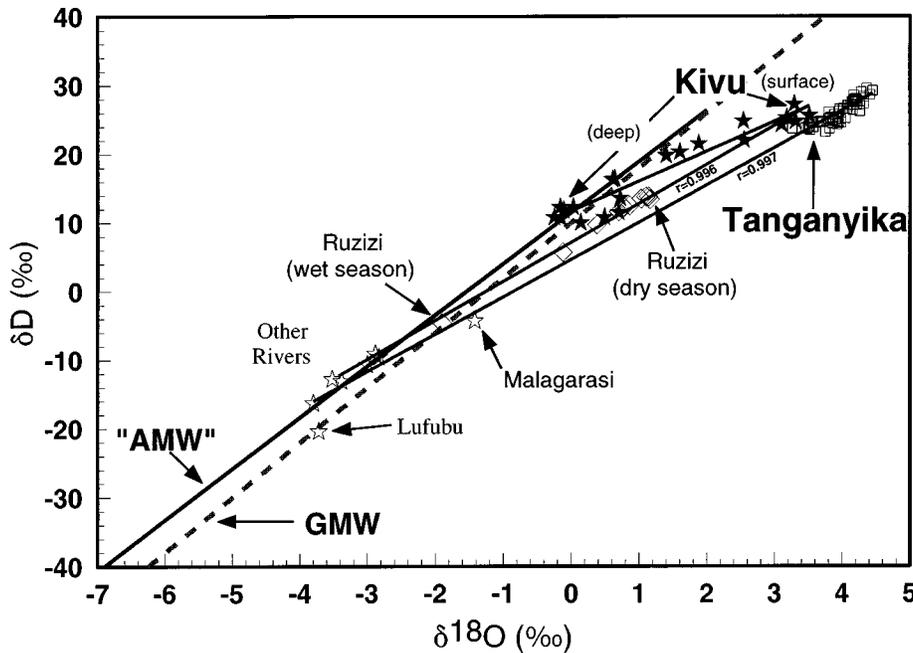


Figure 6. Variations in δD vs. $\delta^{18}O$ of waters from Lake Tanganyika, its tributaries, and Lake Kivu. Data are from Craig et al. (1974) and Tietze (1981, 1982). White squares—Lake Tanganyika data; white vertical diamonds—Ruzizi River, dry season; white horizontal diamonds—Ruzizi River, wet season; black stars—Lake Kivu; white stars—other rivers flowing into Lake Tanganyika. Global meteoric water line (GMW) from Craig and Gordon (1965); African meteoric water line (AMW) based on data for noncoastal tropical East and Central African stations from Rozanski et al. (1993). Note how smaller water courses (Other Rivers) in the Tanganyika basin and the deep waters of Lake Kivu lie on or close to the AMW. In the latter case, this coincidence confirms the importance of hydrothermal waters of meteoric origin to the hydrological budget of Lake Kivu (Degens et al., 1973).

0.908), indicating that both the surface waters and isotopically heavier deep waters evolved from the same primary water sources. Given the relative antiquity of the hypolimnion, we infer that despite changes in hydrological budget, there have been no major changes in bulk inflow composition during the history of the present water body. In the following discussions we therefore assume that there were no significant variations in the isotopic composition of regional rainfall during the period of stromatolite formation. The only significantly variable source is probably the Ruzizi River (see later discussion herein).

Rain falling on the lake surface is the single most important source of water to Lake Tanganyika and currently contributes about 63% of the total annual input (Haberyan and Hecky, 1987). The bulk composition of this rainfall can be estimated from isotopic analyses of local precipitation. Inspection of the most recent compilation of these data (Rozanski et al., 1993) indicates that stations in the interior of East and Central Africa define a regional compositional trend that is separate from the global meteoric water line (GMW; Craig, 1961). We have informally named this trend the

African meteoric water line (AMW; Fig. 6) and its validity for the Tanganyika basin is confirmed by the fact that smaller streams and rivers ("other rivers" in Fig. 6) all plot on or close to the line. Intersection of the Tanganyika-river mixing line with the AMW (Fig. 6) indicates the mean $\delta^{18}O$ of direct precipitation on the lake surface to be $\approx -3.4\text{‰}$ (rather than $\approx -2\text{‰}$ as predicted by the intersection with the GMW). In comparison with modern lake waters ($\delta^{18}O$ +3.7 to +4.2; Fig. 6), the relatively low values for local rainfall and river waters (Fig. 6) confirm that significant isotopic evolution has taken place in Lake Tanganyika, largely as a result of evaporatively driven fractionation in a water body that has a long residence time and has been occasionally closed.

In most prior paleohydrologic studies, the inflow from Lake Kivu via the Ruzizi River has been assumed to be very important to the water balance of Lake Tanganyika (Hecky and Degens, 1973; Haberyan and Hecky, 1987; Casanova and Hillaire-Marcel, 1992). Lake Kivu and the Ruzizi also fall along a δD - $\delta^{18}O$ compositional trend, but one that is separate from that defined by Lake Tanganyika and its sources (Fig. 6). In contrast to its

impact upon the salt budget (Haberyan and Hecky, 1987), the Kivu-Ruzizi system does not dominate the isotopic composition of Lake Tanganyika, although, together with the Malagarasi, it evidently does make a significant wet-season contribution. The seasonal isotopic variation of the Ruzizi is unlikely to reflect changes in the composition of Lake Kivu, because the mixing time of the epilimnion is ≈ 20 yr (based upon data from Haberyan and Hecky, 1987) and thus does not respond fast enough to record seasonal variations. Changes in the composition of the Ruzizi therefore reflect variations in the relative proportions of Lake Kivu water and runoff received from tributaries along its course from Kivu to Tanganyika. Mass-balance calculations, taking Kivu surface water $\delta^{18}O$ to be +3.4‰ (Tietze, 1981) and post-Kivu inflow $\delta^{18}O$ to the Ruzizi to be $\approx -2.5\text{‰}$ (Fig. 6), show that during the wet season the proportion of Kivu water at the mouth of the Ruzizi is only about 10% of the total flow, and $\approx 60\%$ during low flow, dry season conditions. Thus, during the wet season, the Ruzizi may receive as much as 90% of its inflow from catchments downstream of Lake Kivu. It is thus erroneous to assume that closure of Lake Kivu would automatically lead to dramatic loss of inflow from the Ruzizi (cf. Haberyan and Hecky, 1987; Casanova and Hillaire-Marcel, 1992). The Ruzizi would continue as a river system, obviously somewhat reduced in scale, but still capable of making significant seasonal contributions to Lake Tanganyika.

Closure of Lake Kivu would also have significant consequences for the isotopic composition of Lake Tanganyika, because the former is a source of ^{18}O -enriched water that is otherwise absent within the basin (Fig. 6). This would be the case irrespective of Kivu's hydrological budget. Regional inflow to this lake has a mean $\delta^{18}O$ of about 0‰ (Fig. 6), so even a less ^{18}O -enriched water body than today's would supply water to the Ruzizi with a significantly higher $\delta^{18}O$ value than any other known source around Lake Tanganyika. Under present conditions, of a total riverine input of $17.3 \text{ km}^3/\text{yr}$, Lake Kivu (rather than the Kivu-Ruzizi system as a whole) currently contributes $\approx 3.3 \text{ km}^3/\text{yr}$ (Haberyan and Hecky, 1987). Assuming a mean $\delta^{18}O$ composition of -2.5‰ (Fig. 6) for all riverine inflow other than that contributed by Lake Kivu, and a Kivu outflow composition of +3.4‰ (see above), mass balance considerations indicate that the total riverine inflow with Lake Kivu open has a mean composition of $\approx -1.4\text{‰}$. With Kivu closed and the source of isotopically heavy water cut off, the inflow composition is -2.5‰ , assuming that the isotopic composition of rainfall supplying the smaller Ruzizi tributaries remains constant (see above). Thus, rather than leading to isotope enrichment in Lake Tanganyika, closure of Lake Kivu would, at least initially, cause

TABLE 4. STABLE ISOTOPIC ANALYSES

Sample	Site	Depth (m)	Material	Mineralogy*	$\delta^{18}\text{O}^\dagger$ (‰ PDB)	$\delta^{13}\text{C}^\ddagger$ (‰ PDB)
Bergen, Norway analyses [§]						
87AC1A	KM26, northern Burundi	23	Stromtolite	Aragonite	1.13 (0.53)	2.10 (0.40)
87AC2A	KM115, southern Burundi	19	Stromatolite (core)	N.D.	1.01	1.68
87AC2B	KM115, southern Burundi	19	Stromatolite (outer rim)	N.D.	1.54	2.27
87AC3A	KM35.5, northern Burundi	15	Stromatolite (core)	Aragonite	0 (-0.60)	3.77 (2.07)
87AC3B	KM35.5, northern Burundi	15	Stromatolite (outer rim)	Aragonite	0.71 (0.11)	3.04 (1.34)
87AC4A	KM115, southern Burundi	15	Stromatolite (core)	70% HMC, 30% aragonite	1.35 (0.71)	2.19 (1.88)
87AC4B	KM115, southern Burundi	15	Stromatolite	72% HMC, 28% aragonite	1.38 (0.74)	1.91 (1.61)
87AC4C	KM115, southern Burundi	15	Stromatolite (outer rim)	80% HMC, 20% aragonite	1.50 (0.85)	2.24 (1.94)
87AC5A	KM115, southern Burundi	19	Stromatolite (core)	88% HMC, 12% aragonite	1.50 (0.90)	1.85 (1.65)
87AC5B	KM115, southern Burundi	19	Stromatolite (outer rim)	85% HMC, 15% aragonite	1.34 (0.74)	1.86 (1.66)
87AC6A	Kigoma, Tanzania	12	Stromatolite (core)	Aragonite, 12% HMC	1.70 (1.09)	3.95 (2.55)
87AC6B	Kigoma, Tanzania	12	Strom (middle)	Aragonite, 12% HMC	2.24 (1.63)	3.82 (2.42)
87AC6C	Kigoma, Tanzania	12	Stromatolite (outer rim)	Aragonite, 12% HMC	2.29 (1.68)	3.85 (2.45)
87AC7A	Kigoma, Tanzania	11	Stromatolite (core)	Aragonite	1.99 (1.38)	4.21 (2.51)
87AC7B	Kigoma, Tanzania	11	Stromatolite	Aragonite	2.13 (1.53)	4.57 (2.87)
87AC7C	Kigoma, Tanzania	11	Stromatolite	Aragonite	2.17 (1.57)	3.37 (1.66)
87AC7D	Kigoma, Tanzania	11	Stromatolite	Aragonite	2.31 (1.71)	4.02 (2.32)
87AC7E	Kigoma, Tanzania	11	Stromatolite	Aragonite	2.42 (1.82)	4.00 (2.30)
87AC7F	Kigoma, Tanzania	11	Stromatolite (outer rim)	Aragonite	2.64 (2.04)	3.91 (2.21)
87AC8A	Pemba, N. Zaire (Table 2, location 3)	15	Stromatolite (core)	N.D.	2.04	3.73
87AC8B	Pemba, N. Zaire (Table 2, location 3)	15	Stromatolite (outer rim)	Aragonite? (based on thin section)	2.06 (1.46)	4.73 (3.03)
89-MS-211A	South of Nyanza Lac, southern Burundi	0	Beachrock	HMC (11% MgCO ₃)	2.13 (1.47)	1.75
89-MS-211B	South of Nyanza Lac, southern Burundi	1	Beachrock	HMC (10% MgCO ₃)	2.15 (1.55)	1.73
89-MS-142	Ruzizi Delta, northern Burundi	10	Ooids	HMC (11% MgCO ₃)	1.80 (1.14)	2.26
89-MS-49	KM118, southern Burundi	8	Ooids	HMC (12% MgCO ₃)	1.74 (1.02)	2.06
TM-1	Bwana N'denge, Zaire (coll. Jan. 4, 1947)	8-20 [#]	<i>Paramelania damoni</i> gastropod (live)	Aragonite	2.28	-1.19
TM-2	Katibili, Zaire (coll. 1947)	? [#]	<i>Neothauma tanganyicense</i> gastropod (live)	Aragonite	2.22	-0.03

Figure 7. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data for Lake Tanganyika stromatolites, beachrock, and ooids. Values are normalized to calcite based on Tarutani et al. (1969) for $\delta^{18}\text{O}$ and Romanek et al. (1992) for $\delta^{13}\text{C}$ (See Table 5 for original and normalized data). Theoretical $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ ranges for calcite formed in equilibrium with various lake water masses are shown by bars. Estimates of $\delta^{13}\text{C}$ equilibrium values for atmospheric CO₂ are from Romanek et al. (1992). The range of calcite in equilibrium with measured lake dissolved inorganic carbon (DIC) is based on measurements from Degens and Kulbicki (1973), Craig et al. (1974), Hillaire-Marcel et al. (1989), and Hillaire-Marcel and Casanova (1987), and is estimated from the calcite-bicarbonate enrichment factor given by Romanek et al. (1992). Equilibrium $\delta^{18}\text{O}$ fields were calculated using equations from Friedman and O'Neil (1977). The data define a weak covariance, probably indicative of the alternately open and closed conditions of the lake.

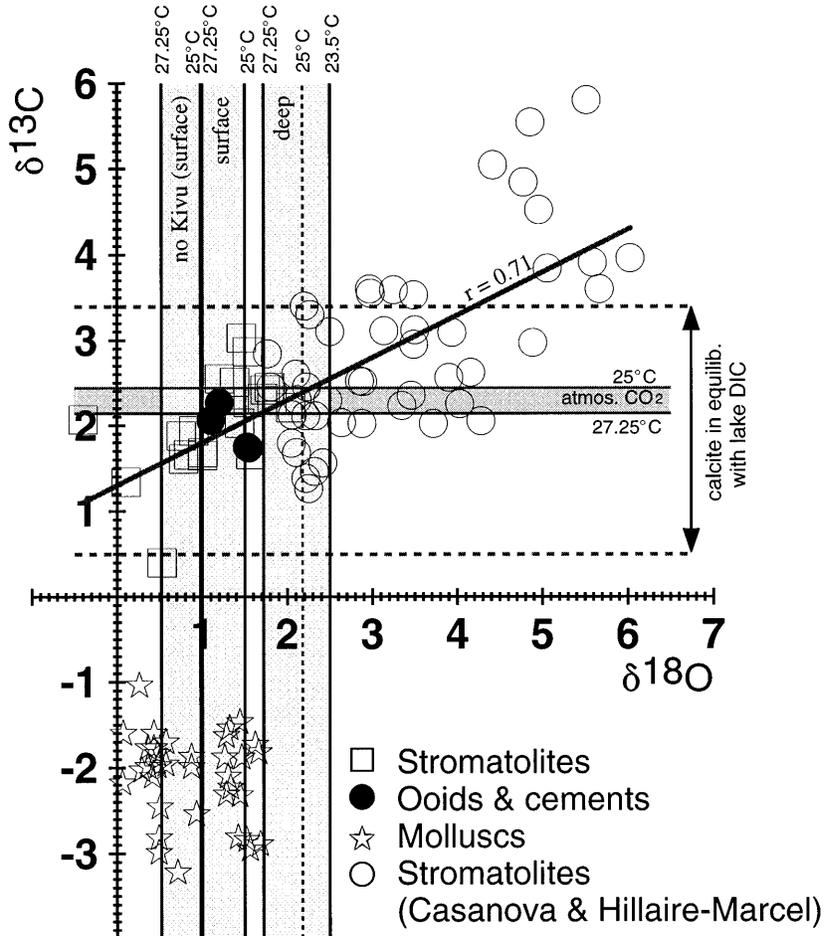


TABLE 4. (Continued)

Sample	Site	Depth (m)	Material	Mineralogy*	$\delta^{18}\text{O}^\dagger$ (‰ PDB)	$\delta^{13}\text{C}^\ddagger$ (‰ PDB)
University of Rhode Island analyses						
	Kaginga, Tanzania	24	<i>P. damoni</i> gastropod (live—outer part of aperture)	Aragonite	1.47	-0.29
	Same individual as above	24	<i>P. damoni</i> gastropod (live)	Aragonite	1.03	-0.07
	Same individual as above	24	<i>P. damoni</i> gastropod (live—apex)	Aragonite	1.10	-0.76
	Same individual as above	24	<i>P. damoni</i> gastropod (live—apex)	Aragonite	0.67	0.10
	Kasereka, Tanzania	36	<i>Lavigeria small dark sp.</i> gastropod (live—apex)	Aragonite	1.09	-1.29
	Same individual as above	36	<i>Lavigeria small dark sp.</i> gastropod (live—apex)	Aragonite	1.53	-0.84
	Same individual as above	36	<i>Lavigeria small dark sp.</i> gastropod (live—aperture)	Aragonite	1.31	-1.52
	Same individual as above	36	<i>Lavigeria small dark sp.</i> gastropod (live—aperture)	Aragonite	1.09	-1.12
	Kirando, Tanzania	1	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	1.17	-0.26
	Same individual as above	1	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	1.03	0.10
	Same individual as above	1	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	1.07	-0.14
	Same individual as above	1	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	1.47	-0.17
	Same locality as above, 2nd snail	3	<i>Lavigeria nassa</i> gastropod (live—apex)	Aragonite	1.01	-0.4
	Same individual as above	3	<i>Lavigeria sp.</i> gastropod (live—aperture)	Aragonite	0.86	0.66
	Same individual as above	3	<i>Lavigeria sp.</i> gastropod (live—aperture)	Aragonite	0.66	-0.48
	Same locality as above, 3rd snail	3	<i>Lavigeria sp.</i> gastropod (live)	Aragonite	0.96	-0.07
	KM26, northern Burundi	12	<i>Paramelania imperialis</i> gastropod (live)	Aragonite	1.17	0.00
	Same individual as above	12	<i>Paramelania imperialis</i> gastropod (live)	Aragonite	0.96	-0.29
	Same individual as above	12	<i>Paramelania imperialis</i> gastropod (live) (aperture)	Aragonite	1.08	-0.28
	Same individual as above	12	<i>Paramelania imperialis</i> gastropod (live—apex)	Aragonite	0.89	-0.33
	Nyanza Lac, southern Burundi	55	<i>Paramelania iridescens</i> gastropod (live—apex)	Aragonite	2.12	-1.14
	Same individual as above	55	<i>Paramelania iridescens</i> gastropod (live—aperture)	Aragonite	1.88	-0.61
	Same individual as above	55	<i>Paramelania iridescens</i> gastropod (live—apex)	Aragonite	2.16	-1.25
	Same individual as above	55	<i>Paramelania iridescens</i> gastropod (live—apex)	Aragonite	1.86	-0.18
	Same locality as above, new individual	73	<i>Paramelania iridescens</i> gastropod (live—apex)	Aragonite	1.88	0.07
	Same individual as above	73	<i>Paramelania iridescens</i> gastropod (live—apex)	Aragonite	2.02	-1.11
	Same individual as above	73	<i>Paramelania iridescens</i> gastropod (live—apex) (aperture)	Aragonite	1.93	-0.53
	KM29, northern Burundi	0.5	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	2.06	-0.18
	Same individual as above	0.5	<i>Lavigeria sp.</i> gastropod (live)	Aragonite	1.90	-0.40
	Same individual as above	0.5	<i>Lavigeria sp.</i> gastropod (live—mid-snail)	Aragonite	1.92	0.16
	Same individual as above	0.5	<i>Lavigeria sp.</i> gastropod (live—aperture)	Aragonite	2.04	0.23
	Same locality as above, new individual	0.5	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	2.04	-0.62
	Same individual as above	0.5	<i>Lavigeria sp.</i> gastropod (live—apex)	Aragonite	2.26	-0.12

*HMC—High magnesium calcite; N.D. = not determined.

[†]Values in parentheses are $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values normalized to calcite and used in Figures 7 and 8. Modern gastropod aragonite values are corrected in Figures 5 and 8, but not in Figure 2 or this table.

[‡]Letters following sample numbers for the University of Bergen analyses indicate multiple samples taken from the same stromatolite head, ordered from A..n, where A is nearest the core of the stromatolite and n is on the outer rim.

[#]Samples collected during 1946–1947 Belgian Hydrobiological Expedition. No further collection depth information given.

surface waters to become isotopically lighter. Conversely, reestablishment of open basin conditions in Lake Kivu should lead to a rise in $\delta^{18}\text{O}$ of the Tanganyika epilimnion.

Equilibrium and Modern Carbonate Isotope Composition

The isotopic composition of Lake Tanganyika carbonates varies over a range of several permil for

both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (Table 4, Fig. 7). We can identify the likely causes of these variations by calculating the expected composition of carbonate minerals precipitated in isotopic equilibrium with lake waters, and by analyzing the shells of living gastropods in the lake. We assume that all stromatolites grew within the top 30 m of the epilimnion (Fig. 4).

The oxygen isotopic composition of carbonates depends on temperature and water ^{18}O composi-

tion. Surface waters vary between 23.8 and 27.3 °C, but generally exceed 25 °C; the deep water has a very uniform temperature of ≈ 23.5 °C (Coulter and Spigel, 1991). The $\delta^{18}\text{O}$ composition of the lake water varies with depth, and stromatolites formed somewhere along the compositional trend between the deep waters, surface waters, and inflow waters (Fig. 6). Equilibrium calculations using the calcite-water fractionation equation given by Friedman and O'Neil (1977) indi-

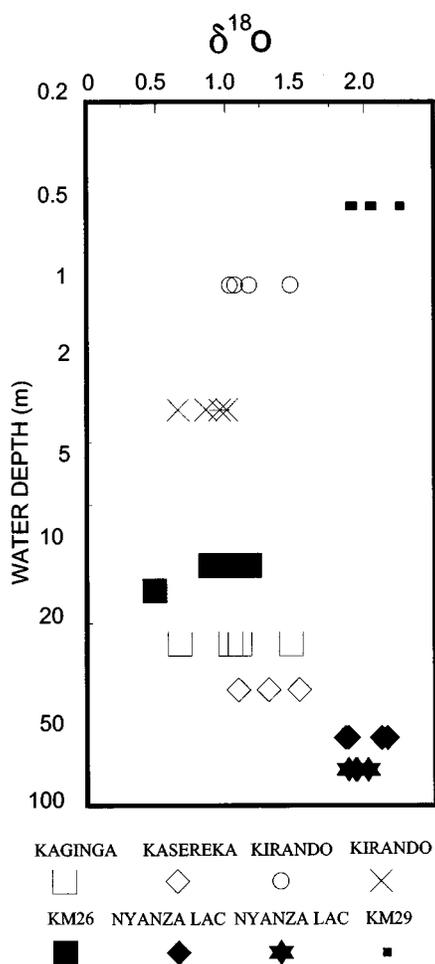


Figure 8. $\delta^{18}\text{O}$ trends with respect to depth for modern gastropods (without aragonite correction). Kasekera, Kazinga, and Kirando Points are located along the northern Tanzanian coast of the lake, close to localities 11, 12, and 15, respectively, in Figure 1. KM26 is located in northern Burundi, 3 km north of locality 5 in Figure 1, in northern Burundi. Nyanza Lac is located 7 km south of locality 10 in southernmost Burundi. The mollusks sampled are as follows: Kazinga—*Paramelania damoni*; Kasekera—*Lavigeria* sp.; Kirando Point (white circles)—*Lavigeria* sp.; Kirando Point (Xs)—*Lavigeria nassa*; KM26—*Paramelania damoni*; Nyanza Lac (black triangles and black stars)—*Paramelania iridescens*; and KM29—*Lavigeria nassa*.

cate a possible range of $\approx 0.45\text{‰}$ for calcite precipitated from modern surface waters ($T = 25\text{--}27.25^\circ\text{C}$) and potentially as much as $\approx 0.80\text{‰}$ for calcite precipitated from waters with a hypolimnic composition ($23.5\text{--}27.25^\circ\text{C}$; Fig. 7).

A wide range of $\delta^{13}\text{C}$ values are reported from

surface water DIC of Lake Tanganyika, mostly $>+1\text{‰}$ (Degens and Kulbicki, 1973; Craig et al., 1974), and in some cases as high as $+4.5\text{‰}$ (Hillaire-Marcel et al., 1989). However, in a 1987 survey of the northern basin, Hillaire-Marcel and Casanova (1992) found a mean $\delta^{13}\text{C}_{\text{DIC}}$ of only -0.6‰ ($n = 19$). Such values are much lower than those previously recorded from either surface or deep waters (cf. Degens and Kulbicki, 1973, p. 224–227; Craig et al., 1974, Fig. 4), suggesting their samples may have been affected by local or seasonal effects. Given the variation in reported DIC isotopic composition, the possible range of $\delta^{13}\text{C}_{\text{calcite}}$ in equilibrium with lake DIC is correspondingly large (Fig. 7).

Modern $\delta^{18}\text{O}$ trends vs. water depth in gastropod shell carbonate are shown in Figure 8 and Table 4. The range of values that we have obtained mostly lie within the predicted $\delta^{18}\text{O}$ range for carbonates formed in equilibrium with modern lake waters, although several littoral and sublittoral (3–25 m) samples were anomalously light. Between the lower littoral zone and the profundal zone, $\delta^{18}\text{O}$ increases, as we would predict from the increase in lake water $\delta^{18}\text{O}$ and the decrease in temperature with depth (Craig et al., 1974; Tietze, 1982). However, the shallowest water snails (*Lavigeria* spp.), collected from <1 m water depth, reverse this trend with relatively high $\delta^{18}\text{O}$ values. This may result from high evaporation rates in the surf zone.

ISOTOPE HISTORY AND INTERPRETATION

The stable isotopic history of Lake Tanganyika displays a complex relationship with the lake-level history. Modern lake water is much higher in $\delta^{18}\text{O}$ than river waters entering the lake, reflecting the importance of evaporation in the lake's water balance and the long residence time of deep lake water. Most late Holocene carbonates appear to have been precipitated from water with $\delta^{18}\text{O}$ values comparable to or higher than modern lake water (Table 4, Figs. 5 and 7). Between 900 and 100 B.C., $\delta^{18}\text{O}$ values increased, and then, after a brief period of stability, declined rapidly between the fourth and sixth centuries A.D. We do not have any direct lake-level fluctuation constraints for the period from 580 to 100 B.C. However, the high $\delta^{18}\text{O}$ values from this time period may reflect a prolongation of lake lowstands and closed basin conditions. Beachrock deposited at the current outlet level during the fifth century A.D. indicates that the lake was hydrologically open when $\delta^{18}\text{O}$ levels began to decline. Isotopic values of beachrock cements suggest they were precipitated from waters similar to those of today's lake (Table 4).

The marked decline in $\delta^{18}\text{O}$ values that occurred starting in the fifth century A.D. probably

represents a regional shift toward higher precipitation to evaporation ratios and consequently greater discharge from the tributary streams that feed Lake Tanganyika, including the Ruzizi River. Haberyan and Hecky (1987) proposed that Lake Kivu overflowed some time during the sixth century A.D. However, as noted earlier, the addition of Lake Kivu waters probably increased oxygen isotopic values in Lake Tanganyika. Therefore, the isotopic signal recorded at this time in Lake Tanganyika probably resulted from increased runoff in the Lake Tanganyika watershed other than Lake Kivu, such as streams that enter the Ruzizi River below the Lake Kivu outlet. The correspondence of Lake Kivu opening and isotopic lightening of Lake Tanganyikan waters at this time may both be consequences of a regional increase in precipitation, but the former is unlikely to be the cause of the latter. Increased precipitation at this time is also supported by data from Jolly and Bonnefille (1992), who showed an expansion of papyrus swamps and rising water tables in upland areas northeast of Lake Tanganyika starting ca. A.D. 720.

Lake levels only rose briefly during the fifth to sixth centuries and then declined to their previous 765–770 masl levels. If Lake Tanganyika opened during the fifth century, substantial discharge through the Lukuga River spillway may have prevented a major lake-level rise in response to increased precipitation. However, this does not explain the apparent stabilization of lake levels below the outlet level (≈ 770 m) during the seventh century. The precipitation/evaporation ratio may have hovered near 1 during this time, with brief intervals of overflow (on the order of a few decades), interspersed with closed basin conditions. Fluctuations of only a few meters would have allowed for the formation of stromatolites 10 m below modern lake level, and contemporaneous stromatolites forming just 10 m higher would have been eroded or prevented from forming as long-lived buildups.

Since the seventh century, oxygen isotope values of Lake Tanganyikan surface waters have been relatively stable (between 0‰ and $+2\text{‰}$) for all dated stromatolites, ooids, and beachrock. This variability is comparable to that observed in the modern gastropods collected between the surface and 73 m water depth discussed earlier (Fig. 8). The ≈ 10 m lake-level rise of the thirteenth and fourteenth centuries is not reflected by any significant shift in the isotopic composition of lake water.

Values of $\delta^{13}\text{C}$ for Lake Tanganyika also vary over a 5‰ range (generally $+1\text{‰}$ to $+6\text{‰}$) throughout the late Holocene. The only exceptions are the $\delta^{13}\text{C}$ values obtained from fossil and living gastropods. Snail shells were consistently 2‰ – 3‰ lighter than stromatolitic carbonate, an expected consequence of molluscan uptake of

isotopically light carbon from food, which is not in isotopic equilibrium with lake water (Tanaka et al., 1986; Dolan-Laughlin et al., 1992).

Figure 7 illustrates the relationship between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ for all stromatolites, ooids, and beachrock analyzed by us and by Casanova and Hillaire-Marcel (1992). We also plotted the equilibrium $\delta^{18}\text{O}$ values for modern surface and deep waters, and the equilibrium $\delta^{13}\text{C}$ values calculated from surface water DIC, assuming a pH of 9.0–9.1, and that DIC is overwhelmingly HCO_3^- (Romanek et al., 1992). A Pearson product moment correlation test on these data shows a significant correlation ($r = 0.71$, $p < 0.001$) between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. Talbot (1990) argued that covariant isotopic trends ($r > 0.7$) can be expected in hydrologically closed lakes, whereas open basins are typified by a lack of covariance. The covariance of isotopic trends during the late Holocene history of Lake Tanganyika may reflect a dominance of lake basin closure during this interval. Alternatively, it may reflect the mixing between long-residence-time deep waters and the shorter residence time of the epilimnion. Most of the carbonates that have formed in the lake over the past 2500 yr have $\delta^{18}\text{O}$ values intermediate between equilibrium with modern surface waters ($\delta^{18}\text{O}_c \approx +1.0\text{‰}$ to $+1.5\text{‰}$) and with deep hypolimnetic waters ($\delta^{18}\text{O}_c \approx +1.7\text{‰}$ to 2.5‰ , the lower value only being achieved if deep waters were warmed to modern lake surface temperature) (Craig et al., 1974; Tietze, 1982, Fig. 7). Similarly, most of the $\delta^{13}\text{C}$ values we found in these carbonates are comparable to those predicted from equilibrium with dissolved inorganic carbon values from the lake's surface waters ($\delta^{13}\text{C}_{\text{DIC}} = -0.6\text{‰}$ to $+4.5\text{‰}$, Hillaire-Marcel et al., 1989; Casanova and Hillaire-Marcel, 1992; Craig et al., 1974; Degens and Kulbicki, 1973).

CONCLUSIONS

Both modern and late Holocene stromatolites in Lake Tanganyika have been formed by complex microbial communities that include filamentous and coccoid cyanobacteria, diatoms, green algae, fungi, and sponges. These stromatolites have a wide range of shapes and sizes, and many morphological features are correlated with water depth. This has allowed us to determine lake-level fluctuations using dated fossil stromatolite distribution patterns.

We have combined new radiocarbon age data with the earlier data of Casanova and Hillaire-Marcel (1992), correcting for lake mixing effects, to produce a lake level and synoptic record of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ change in the lake. These data also have important implications for regional climate change during the Holocene.

Lake Tanganyika lake levels have been relatively stable over the past 2800 yr, fluctuating

around the marginally open to marginally closed (765–775 masl) level through most of this time period. Only during times when the Lukuga River outlet has become blocked by alluvium and debris has it been possible for Lake Tanganyika to rise significantly above this level. We find no evidence for the sharply lower lake stands (–75 m) predicted by some earlier hydrologic models for the time periods before ca. A.D. 550, when the Ruzizi was not flowing into Lake Tanganyika from Lake Kivu.

Relatively low lake levels and ^{18}O enrichment of lake water between the ninth century B.C. and the early fifth century A.D. suggest comparatively dry conditions. A marked decline in $\delta^{18}\text{O}$ during the fifth century A.D. was initially accompanied by a brief but rapid rise in lake levels. This event corresponds to increased regional precipitation and the opening of Lake Kivu, which spilled over into the Lake Tanganyika basin. Tributary discharges probably account for this $\delta^{18}\text{O}$ change, because the overflow of Lake Kivu alone likely would have led to isotopically heavier water entering the lake. Since the sixth century A.D. there has been little change in $\delta^{18}\text{O}$ of lake carbonates. Lake levels were at or just below the outlet level between the sixth and thirteenth centuries A.D. Lake levels started to rise during the fourteenth century, and then were punctuated by a fall of uncertain magnitude (at least 15 m but no more than 40 m below modern levels) at some time between the late sixteenth and early nineteenth centuries. Rising lake levels in the nineteenth century correlate with changes in discharge of the Nile, suggesting an increase in regional precipitation, and culminated in the opening of Lake Tanganyika in the late nineteenth century.

Trends of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ for biologically mediated carbonates (excluding mollusks) covary significantly during the late Holocene. This relationship, coupled with our lake-level data, suggests that Lake Tanganyika has hovered between open and closed basin conditions. The relationship between Lake Kivu solute input, isotope input, and the level of Lake Tanganyika is quite complex. Whereas Lake Kivu waters are very important in the regulation of major ion solute composition and concentration in Lake Tanganyika, their importance in regulating both lake level and $\delta^{18}\text{O}$ appear to be much less than previously assumed. Most of the water discharged by the Ruzizi River into Lake Tanganyika is probably derived from sources other than Lake Kivu itself. Therefore, overflows events in Lake Kivu and lake-level highstands in Lake Tanganyika may have been uncoupled throughout the region's Holocene history.

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