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Paleoclimatic Estimates from Water and Energy Budgets of East African Lakes

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The Turkana, Nakuru–Elmenteita, and Naivasha basins in the Rift Valley of East Africa experienced high water levels during the period of 10,000–7000 yr B.P. Analyses of the modern hydrologic and energy budgets for these basins, along with that of Lake Victoria, were used to infer the amount of precipitation that would have been required to maintain the enlarged paleolakes of the early Holocene. Precipitation must have been at least 150–300 mm/yr (15 to 35%) above the modern average. The precipitation estimates were fairly consistent among the various basins, but no quantitative estimate was made for the additional precipitation required to account for overflow from the Rift Valley lake basins. Discharge from the Lake Victoria basin around 1880 A.D. was considerably above the more recent average, and the increased discharge into the White Nile for that period might have been similar to that of the early Holocene. A sensitivity analysis showed that temperature changes were probably not too important for changing the hydrologic-energy budget; changes of albedo, Bowen ratio, and cloudiness were likely to have been of greater importance.

INTRODUCTION

Geomorphic evidence of variations of water levels of tropical lakes has been known since the era of early exploration of East Africa (e.g., Nilsson, 1931; Butzer *et al.*, 1972). The climatic interpretation of these field observations has been hampered by the lack of absolute dating of the events, by insufficient information about the regional patterns of the environmental changes, and by uncertainties concerning the dominant climatic factors that might have accounted for the changes of the lake levels.

During recent years, a radiocarbon-dated chronology of water-level variations and land vegetation changes has been established for several East African lakes (Kendall, 1969; Washbourn-Kamau, 1971; Butzer, 1971; Richardson and Richardson, 1972; Butzer *et al.*, 1972; Butzer, 1980; Livingstone, 1980) and for lakes in southern Ethiopia (Street, 1979). Moreover, the broad-scale regional patterns of lake-level variations have been charted and summarized (Street and Grove, 1979); for example, Street and Grove showed that the raised lake levels of the early Holocene in East Africa formed part of a widespread pattern of enlarged lakes across parts of Africa, Arabia, and India. Finally, knowledge of the modern hydrologic budget of East Africa has accumulated (Jackson, 1961; East African Meteorological Department, 1970; Griffiths, 1972; Richardson and Richardson, 1972; World Meteorological Organization-UNDP, 1974; Gaudet and Melack, 1981) and methods for the quantitative treatment of the combined hydrologic and energy budgets of large catchment areas have been developed (Lettau, 1969; Kutzbach, 1980).

These advances provide a basis for the present reappraisal of the climatic implications of lake-level variations of equatorial East Africa. It will be shown: (1) that the estimates of increased precipitation (compared to present) for the early Holocene for four East African basins are generally consistent, and (2) that these estimates of increased precipitation in East Africa are also consistent with the results of other studies from neighboring regions. The conclusion that precipitation changes must have been primarily responsible for the lake-level changes is supported by an analysis of the possible role of other factors (e.g., temperature, cloudiness).

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BASIC THEORY

The equation for the steady-state hydrologic budget of closed-basin lakes involves a balance among three terms. The inputs are precipitation on the lake and runoff from the land to the lake; the output is evaporation from the lake (Leopold, 1951; Butzer *et al.*, 1972; Street, 1979). Subsurface inflow or outflow can also be included (if known), and surface discharge becomes a factor if the lake overflows its catchment. The accurate estimation of runoff from the land to the lake requires a knowledge of the hydrologic budget over the land. For the land hydrologic budget, the estimation of evaporation is particularly difficult. Because evaporation from the land surface appears as a factor in both the surface energy budget and the hydrologic budget, the estimates of evaporation should be consistent with both budgets. Kutzbach (1980) evaluated the combined hydrologic and energy budget equations for both lake and land surfaces of Chad Basin in order to estimate the present and past hydrologic conditions. That approach will be followed here and is reviewed briefly.

The steady-state equation for the mean-annual hydrological balance for a closed basin is

$$P = E_w a_w + E_l(1 - a_w), \qquad (1)$$

where P is the basin-average precipitation, E is the evaporation (the subscript w refers to the water surface of the lake; the subscript l refers to the land surface of the basin), and a_w is the fraction of the land area (A_l) and lake or water area (A_w) that is occupied by the lake: $a_w = A_w/(A_w + A_l)$.

The evaporation term may be estimated from the mean-annual energy budget at the water or land surface

$$E = \frac{1}{(1 + B)} \frac{R}{L}, \qquad (2)$$

where R is the net radiation, B is the Bowen ratio, and L is the latent heat of evaporation. The Bowen ratio is the ratio of the sensible heating of the atmosphere (proportional to the surface–atmosphere temperature difference) to the latent heating of the atmosphere associated with the addition of water vapor from evaporation. In the tropics, B is large for desert surfaces (B $\sim$ 3) and small for water surfaces ($B \sim 0.1$ to 0.2). By expressing the lake and land evaporation (Eq. (1)) in terms of net radiation and Bowen ratio (Eq. (2)), a diagnostic equation for basin-average precipitation is obtained

$$LP = \frac{1}{1 + B_w} R_w a_w + \frac{1}{1 + B_l} R_l (1 - a_w). \qquad (3)$$

Equation (3) is identical to Eq. (5) of Kutzbach (1980), except for a change of subscripts, and requires estimates of a_w, R_w, R_l, B_w, and B_l in order to solve for the basin-average precipitation.

For the case with discharge from the basin, the volumetric discharge per unit area of the basin, D, is added to the right-hand side of Eq. (1); the same discharge term multiplied by L, the latent heat of evaporation, would be added to the right-hand side of Eq. (3).

A fundamental assumption of this approach is that changes of lake level or discharge result entirely from climatic changes rather than in part from changes in the height of the landforms damming a closed-basin lake. Because of the near-synchronous high lake levels of these East African lakes in the early Holocene, Bishop (1971) has argued for a regional climatic explanation rather than local tectonic activity.

FIELD EVIDENCE

Limnological and paleolimnological information is available for Lake Victoria (Kendall, 1969; Livingstone, 1980), Lake Turkana (Butzer, 1971, 1980), Lakes Nakuru and Elmenteita (Washbourn-Kamau, 1971; Richarson and Richardson, 1972; Butzer *et al.*, 1972; Melack and Kilham, 1974), and Lake Naivasha (Richardson and Richardson, 1972; Butzer *et al.*, 1972; Gaudet and Melack, 1981), the three latter lakes are in the Eastern Rift Valley (Fig. 1).

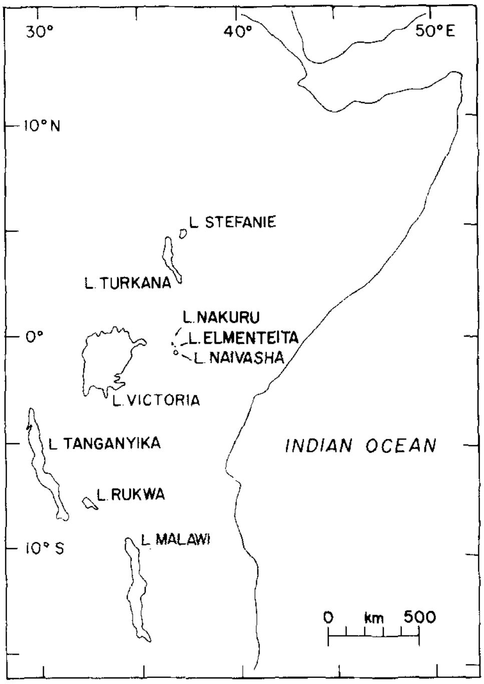

FIG. 1. Map of East Africa showing location of Lakes Victoria, Turkana, Nakuru, Elmenteita, and Naivasha.

Extracts from the detailed information contained in these references are reproduced in Table 1, and the present and past lake shorelines are shown in Figure 2. The different literature sources give somewhat different values for lake size (for example, the modern lakes vary in size from year to year); the climatological parameters are, however, also subject to considerable uncertainties so that basin and climatic information are perhaps of comparable accuracy.

The primary conclusion to be drawn from the abbreviated summary in Table 1 is that at times between about 10,000 and 7000 yr B.P., with the dates varying somewhat for each lake, the present closed-basin lakes of Turkana, Nakuru–Elmenteita, and Naivasha were considerably enlarged and, at times, overflowed their basins. Lake Victoria, which now discharges to the White Nile, had ceased its discharge for a period before 12,500 yr B.P. and around 10,000 yr B.P., but discharge was reestablished by 9500 yr B.P.

A key parameter in Eq. (3) is the fraction of the basin covered by water (a_w). That fraction increased for the high lake-level period (compared to modern) by a factor of almost 3 for Turkana (0.13 compared to a modern value of 0.05), by a factor of about 15 for Nakuru–Elmenteita (0.44 compared

TABLE 1. TOTAL BASIN AREA, LAKE AREA, FRACTIONAL LAKE AREA, LAKE-LEVEL CHANGE, AND DISCHARGE FOR FOUR EAST AFRICAN LAKES AT SELECTED TIMES

	Basin area (m^2)	Lake area (m^2)	Fractional lake area, a_w	Lake-level change, Δh	Discharge, D
Lake Turkana					
Modern	1463×10^8	73×10^8	0.05	0	0
10,000–7000 yr B.P.		185×10^8	0.13	60–80 m higher	Yes
Lakes Nakuru–Elmenteita					
Modern	2055×10^6	60×10^6 (combined)	0.03	0	0
10,000–8000 yr B.P.		900×10^6 (combined)	0.44	As much as 180 m higher	Yes
Lake Naivasha					
Modern	3185×10^6	114×10^6	0.04	0	0
At least 9200 yr B.P. to 5650 yr B.P.		600×10^6	0.19	60 m higher	Yes
Lake Victoria					
Modern	260×10^9	68×10^9	0.26	0	110 mm/yr[a]
Before 12,500 yr B.P. and about 10,000 yr B.P.		59×10^9	0.23	At least 25 m lower	0 (closed basin)
After about 9500 yr B.P.		68×10^9	0.26	?	Yes
About 1880 A.D.		68×10^9	0.26	2 m higher[b]	280 mm/yr[c]

Note. Discharge does not include estimates of possible subsurface flow.
[a] Based upon 1946–1970 discharge of 28.6×10^9 m^3/yr.
[b] Ravenstein (1901); Lamb (1966).
[c] Discharge of 73×10^9 m^3/yr corresponding to a raised lake level (WMO–UNDP, 1974, Vol. 1).

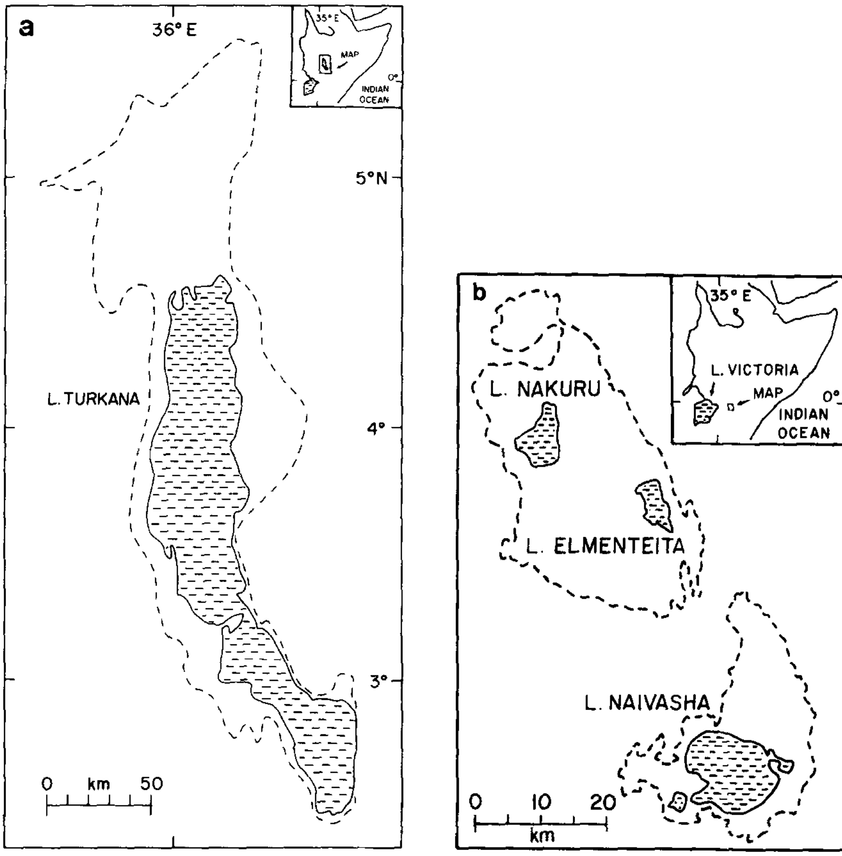

FIG. 2. Maps of (a) Lake Turkana, and (b) Lakes Nakuru, Elmenteita, and Naivasha. The present lake area is stippled, and the shoreline is shown by a solid line; the higher early Holocene shoreline is indicated by a broken line.

to 0.03), and by a factor of about 5 for Naivasha (0.19 compared to 0.04) (Table 1).

If past values of net radiation and Bowen ratio can be estimated, then the past values of fractional lake area provide the basis for minimum estimates of the increased precipitation required to maintain the expanded lakes (Eq. (3)). The actual value of increased precipitation would have exceeded these minimum estimates when overflow occurred.

MODERN HYDROLOGIC AND ENERGY BUDGETS

The modern values for net radiation and Bowen ratio for water and land surfaces (R_w and R_l, B_w and B_l) are determined with the aid of the climatic variables and the necessary equations (Table 2). The general climatic conditions for each basin are tabulated. The land surfaces of the four basins have considerable ranges of elevation and climatic conditions; temperature, precipitation, and radiation vary strongly with elevation. However, for the simplified hydrologic-energy balance model employed in this paper, these details are ignored.

The Lake Turkana basin receives the least precipitation (~750 mm/yr) of the four basins. Rainfall over the lake is much less than for the basin as a whole (Table 2). The ratio of lake area to basin area (a_w = 0.05; Table 1) is small. For Lakes Nakuru and Elmenteita, the basin-average precipitation is about 900 mm/yr; their combined fractional lake area is the smallest of the three Rift Valley lake basins (a_w = 0.03; Table 1). The climatic conditions for Lake Naivasha are similar to those for Nakuru–Elmenteita (P ~ 900 mm/yr), and the fractional lake area is comparable (a_w = 0.04; Table 1). Lake Naivasha differs from the other Rift Valley lakes described here in that it is essentially a freshwater lake (Richardson and Richardson, 1972); these authors noted that subsurface drainage may be a factor in preventing the concentration of salts in the lake via evaporation. This subsurface drainage would also need to be considered in the hydrologic budget if the rate were known quantitatively. Gaudet and Melack (1981) have estimated the hydrologic budget for Lake Naivasha and found that seepage losses were 5–20% of the total water loss. However, seepage gains were also estimated to occur, so that the net effect on the water budget of subsurface flow was small. (Lake Turkana also has relatively low salinity and may experience subsurface drainage as well; Beadle, 1974). The Lake Victoria Basin has a high basin-average precipitation (P = 1250 mm/yr) compared to Rift Valley lake basins; it has a rather large ratio of water surface to basin area (a_w = 0.26; Table 1) and significant discharge into the White Nile (D = 110 mm/yr; Table 1). A noteworthy climatic feature is the large rainfall over the lake at nighttime, associated with nighttime flow convergence (Flohn and Fraedrich, 1966; Fraedrich, 1972).

Radiation budgets were calculated from the shortwave (solar) and longwave (terrestrial) components of the radiation for each basin. Solar radiation (clear sky) was taken to be the same for each of the four basins (320 W/m^2). Fractional cloud cover was set at 0.50 for all basins except Turkana, where a value of 0.45 was used. The surface albedos of water (0.06) and land (0.22) were taken as representative for all basins with the exception of the more humid Victoria Basin, for which the land albedo was reduced by 10% (0.20). The estimated values of net shortwave radiation absorbed at the surface of water and land are very similar for each basin ($SW_w \sim 200$ W/m^2, $SW_l \sim 170$ W/m^2). The net loss of longwave radiation depends upon surface temperature, surface emissivity, vapor pressure, and cloudiness. The climatological values of surface temperature and vapor pressure differ from basin to basin and between lake and land within each basin (Table 2), and lead to values of net longwave radiation loss in the range of 60–80 W/m^2. For the three Rift Valley basins, the dry atmosphere above the land surface (relative to the water surface) allows greater longwave radiation losses from the land ($LW_l \sim$ 70–85 W/m^2) than from the lake ($LW_w \sim$ 60–70 W/m^2).

As a result of the greater longwave radiation losses from the land (compared to lake) and the smaller solar radiation gains for the land (compared to lake), the net radiation is significantly greater over the water surface than over the basin's land surface ($R_w \sim$ 140 W/m^2, $R_l \sim$ 90–100 W/m^2). The water–land contrast for net radiation is somewhat reduced for Lake Victoria Basin because of the relatively humid conditions of the basin. The climatic variables and equations used to estimate these radiation budgets are subject to some uncertainty, but the general result of a sizable difference in net radiation budget for land and water surfaces is well established.

Detailed hydrologic budget studies have led to an estimate of Bowen ratio of B_w = 0.21 for Lake Victoria (Table 2). For Lake Victoria, this amounts to an evaporation rate of about 1500 mm/yr. Lacking detailed hydrologic studies for the other lakes, a Bowen ratio of 0.21 has been assigned to each. Because the net radiation (R) is almost identical for each lake, the choice of a constant Bowen ratio for each lake implies (from Eq. (2)) almost identical evaporation rates of about 1500 mm/yr for each lake. As an alternate approach, evaporation was estimated according to a formula of Priestley and Taylor (1972), (see Table 2, footnote m). That approach yielded lake evaporation estimates ranging from 1500 mm/yr (Victoria) to 1800 mm/yr (Turkana). The relatively small range of these estimates suggests that our assumption of identical Bowen ratios and nearly identical evaporation rates is not unreasonable.

TABLE 2. MODERN ANNUAL MEAN RADIATION, EVAPORATION, AND HYDROLOGIC BUDGET TERMS FOR FOUR EAST AFRICAN LAKES AND THE SURROUNDING LAND SURFACES OF THE BASINS

				Turkana		Nakuru–Elmenteita		Naivasha		Victoria	
Climate parameter	Footnote	Symbol	Unit	Water	Land	Water	Land	Water	Land	Water	Land
Radiation budget											
Global radiation (clear sky)	*a*	G_0	W/m²	320	320	320	320	320	320	320	320
Fractional cloud cover	*b*	C	—	0.45	0.45	0.50	0.50	0.50	0.50	0.50	0.50
Global radiation (with clouds)	*c*	G	W/m²	226	226	216	216	216	216	216	216
Surface albedo	*d*	α	—	0.06	0.22	0.06	0.22	0.06	0.22	0.06	0.20
Net shortwave radiation	*e*	SW	W/m²	213	177	203	168	203	168	203	173
Surface emissivity	*f*	ϵ	—	0.96	0.92	0.96	0.92	0.96	0.92	0.96	0.92
Surface temperature	*g*	T	°K	302	295	292	289	291	289	298	295
Surface vapor pressure	*h*	e	mb	18	9	18	11	17	11	21	19
Net longwave radiation	*i*	LW	W/m²	72	85	62	71	64	71	60	59
Net radiation	*j*	R	W/m²	141	92	141	97	139	97	143	114
Bowen ratio and evaporation											
Bowen ratio	*k*	B	—	0.21	0.64	0.21	0.43	0.21	0.46	0.21	0.50
Evaporation (heat equiv.)	*l*	LE	W/m²	116	56	112	68	115	67	118	76
Hydrologic budget											
Evaporation	*m*	E	mm/yr	1500	730	1500	880	1500	870	1500	1000
Precipitation	*n*	P	mm/yr	250	770	900	900	600	900	1650	1100
Drainage into lake	*o*	N	mm/yr	1250	40	600	20	900	30	250	100
Discharge	*p*	D	mm/yr	—	—	—	—	—	—	400	—

[a] From Budyko (1974, pp. 46–47).

[b] From Atkinson and Sadler (1970); these satellite-derived cloudiness values tend to be somewhat larger than values obtained from ground observations (Barrett, 1974, pp. 177–178). Satellite imagery shows a relative minimum of cloudiness over Lake Victoria during daytime (NOAA–USAF, 1971, pp. 164–237) but this was not incorporated in our calculations.

[c] $G = (1-0.65C)G_0$; from Budyko (1974).

[d] From Sellers (1965, p. 21); Budyko (1974, p. 54–55). Rockwood and Cox (1978) measured a surface albedo of 0.22 for "mixed vegetation" (50% light, 50% dark) in West Africa; by contrast, savanna (70% light, dry grasses) had an albedo of 0.28, and moderate forest (70% dark vegetation) had an albedo of 0.17.

[e] $SW = G(1 - \alpha)$.

[f] From Sellers (1965, p. 41).

[g] From Butzer *et al.* (1972); Butzer (1980); Griffiths (1972); World Meteorological Organization–UNDP (1974, Vol. 1). Regression equations in Griffiths (1972, pp. 319–321) permitted estimation of temperature for various elevations.

[h] Estimates of e were based upon assumed relative humidities for the water and land areas, respectively, of 45 and 35% for Turkana, and of 80 and 60% for Nakuru–Elmenteita and Naivasha. For Victoria, direct estimates of e were available from World Meteorological Organization–UNDP (1974).

[i] $LW = \epsilon\sigma T^4 (0.39 - 0.05\sqrt{e})(1 - 0.53C^2)$; from Budyko (1974, pp. 58–60); $\sigma = 5.67 \times 10^{-8}$ W/m² °K⁴.

[j] $R = SW - LW$; values for the land surface of about 100 W/m² agree closely with the analysis by Budyko (1974, p. 155).

[k] Lake Bowen ratio (B_w) of 0.21 fixed for all lakes on basis of Lake Victoria hydrometeorological studies (World Meteorological Organization–UNDP, 1974). Land Bowen ratio (B_l) is that value required to satisfy Eq. (3) for modern values of P, R_w, R_l, a_w, and B_w.

[l] L is latent heat of evaporation. $L = 585$ cal/g^{-1} $= 246 \times 10^4$ J/kg at $T = 293$°K. Evaporation of liquid water column of 1000 mm/yr corresponds to the energy equivalent of about 77 W/m². Values of LE for the land surface of about 60–70 W/m² agree closely with the analysis by Budyko (1974, p. 161). The values for LE for the lake surface correspond, upon conversion to liquid water column equivalents, to E of about 1500 mm/yr (within 1–2%).

[m] The values of 1500 mm/yr for lake evaporation are rounded off from the exact solution of Eq. (2) for $B_w = 0.21$. The value of 1500 mm/yr has been given for Lake Victoria (WMO–UNDP, 1974), and values of about 1400 mm/yr have been given for Lake Naivasha and Lakes Nakuru and Elmenteita (Richardson and Richardson, 1972). Gaudet and Melack (1981) find for Lake Naivasha in 1973–1975 a lake and swamp evaporation of about 300×10^6 m³/yr or about 1670 mm/yr. The alternate formula by Priestley and Taylor (1972) that is discussed in the text is $LE = 1.26\, s(s + \gamma)^{-1}R$, where γ is the psychrometric constant and s is the rate of change of saturation vapor pressure with temperature at the specified surface temperature. This formula gives evaporation values for the lake surface ranging from 1500 mm/yr (Victoria) to 1800 mm/yr (Turkana). Land evaporation is also rounded to balance precipitation.

[n] These precipitation values are slightly rounded from the original sources for convenience. Turkana: East African Meteorological Dept. (1970, 1975); Jackson (1961). Nakuru–Elmenteita: Richardson and Richardson (1972). Naivasha: East Africa Meteorological Dept. (1970, 1975); Richardson and Richardson (1972); Gaudet and Melack (1981). Victoria: WMO–UNDP (1974, Vol. 1); East African Meteorological Dept. (1970). The exact values of basin-average precipitation used to solve Eq. (3) for B_l were: Victoria, 1243 mm/yr; Turkana, 754 mm/yr; Nakuru–Elmenteita, 900 mm/yr; Naivasha, 889 mm/yr.

[o] For the land budget, $N_l = P_l - E_l$. Runoff from land (N_l) is expressed per unit area of land surface of basin. Runoff into lake (N_w) is expressed per unit area of lake and the lake budget equation is $P_w + N_w = E_w$; $P_w + N_w = E_w + D$ for Victoria.

[p] Discharge: see Table 1; D is here expressed per unit area of lake.

The one remaining unknown, the Bowen ratio of the land surface of each basin (B_l), was estimated by solving Eq. (3) for B_l given the modern basin-average precipitation (P), the modern lake area fraction (a_w), and the modern estimates for R_w, R_l, and B_w. For the case of Victoria, the discharge term (D), expressed in energy units (LD), was added to the right-hand side of Eq. (3) before solving for B_l. As a consequence of obtaining E_l by first solving Eq. (3) for B_l, the combined lake and land evaporation for each basin exactly balances the basin-average precipitation (minus the discharge for the case of Victoria). The Bowen ratio that was obtained for the land surface ranged from 0.64 for Turkana (the basin with the lowest basin-average precipitation) to 0.45–0.50 for the other basins. We have verified, where possible, that our estimates of evaporation from the land surface (E_l) agree with previously published estimates (Table 2, footnote *l*).

In summary, the net radiation is markedly lower for the land surface than for the water surface, while the Bowen ratio is higher for land than for lake. As a result, the evaporation, which is proportional to $R/(1 + B)$, is substantially lower for the land surface than for the lake surface (Table 2). Paleoclimatic evidence of expanded lakes (decreased land surface) must therefore, in the context of Eq. (3), imply increased precipitation in order to compensate for the larger evaporative losses from water compared to land, (see Kutzbach (1980) for a more complete discussion and interpretation of the combined hydrologic and energy budget approach as applied to lake and paleolake studies within the context of Chad Basin).

SENSITIVITY STUDIES

A direct application of Eq. (3) to the estimation of past basin-average precipitation (P) based *solely* upon past values of fractional lake area (a_w) requires that past and present values of net radiation and Bowen ratio are identical. Some insight concerning the sensitivity of such precipitation estimates to variations of components of the radiation budget and the Bowen ratio was gained by expressing the net radiation in terms of the variables from which it was computed (temperature, vapor pressure, cloudiness, albedo; Table 2, footnotes) and then computing the change in precipitation (δP) as a function of individual variations of temperature, cloudiness, albedo, and Bowen ratio. A selection of these results for changes of land surface temperature (δT_l), lake and land cloudiness (δC), land albedo ($\delta\alpha_l$), and land Bowen ratio (δB_l) is summarized in Table 3. For comparison, the sensitivity of the precipitation estimate to variations in fractional lake area (δa_w) is also shown. The results differ slightly for each lake basin because of the somewhat different climatic conditions.

The first point to be noted is the relative insensitivity of the precipitation estimate to temperature changes: a δT_l of 1°K corresponds to a δP of only about 5-10 mm/yr; δP is positive if it is assumed that vapor pressure increases with temperature in such a manner that relative humidity remains constant, whereas δP is negative if vapor pressure is not permitted to adjust as the temperature is increased. In the latter case, the effect of the increased temperature is to increase the longwave radiation loss and therefore to decrease the net radiation and hence the evaporation. In the former case, which could occur if the atmospheric vapor pressure should increase as a result of increased evaporation from the warmer surface, the effect of the increased vapor pressure dominates over the temperature effect and produces the opposite result; the longwave radiation loss is reduced, and evaporation is increased. In either case, changes of temperature would have to be at least ±5°K to produce changes of precipitation similar to those produced by 10% changes of cloudiness, albedo, and Bowen ratio (~25 mm/yr or more). Moreover, the lake area fraction

TABLE 3. SENSITIVITY ANALYSIS

Basin	$\delta a_w(+10\%)$	$\delta B_l(+10\%)$	$\delta \alpha_l(+10\%)$	$\delta T_l(+1°K)$ e	$\delta T_l(+1°K)$ RH	$\delta C(+10\%)$
Turkana	4	−27	−36	−8	4	−42
Nakuru–Elmenteita	2	−26	−42	−8	7	−54
Naivasha	2	−26	−41	−8	7	−54
Victoria	14	−24	−28	−5	10	−63

Note. Changes of basin-average precipitation, δP (mm/yr) as a function of changes (δ) in lake area fraction (a_w), land Bowen ratio (B_l), land albedo (α_l), land surface temperature (T_l), and basin-average cloudiness (C). For T_l, e denotes fixed vapor pressure, RH denotes fixed relative humidity (see text).

(a_w) changed by a *factor* of 3 to 15 for the Rift Valley lakes (Table 1); increases of a_w of that magnitude are associated with changes of precipitation of the order of 100 mm/yr (Tables 3 and 4). As a result of this analysis, we conclude that temperature changes are likely to be of comparatively minor importance for estimating the magnitude of changes of precipitation.

Increases of land surface albedo ($\delta\alpha_l$) of 10% (e.g., from 0.20 to 0.22) are associated with δP of −30 to −40 mm/yr (Table 3); increased albedo implies decreased solar and net radiation and therefore decreased evaporation (and precipitation) according to Eq. (3). The numerical value of a 10% change was chosen as a useful benchmark because the albedo can undergo seasonal changes of at least 10% from wet to dry seasons (for savanna vegetation—wet season, $\alpha_l = 0.18$; dry season, $\alpha_l = 0.24$; average, $\alpha_l = 0.21$); Budyko, 1974). A Bowen ratio increase of 10% also corresponds to a precipitation decrease of about 25 mm/yr (Table 3). The higher Bowen ratio implies that less water is evaporated (Eqs. (2), and (3)) and, therefore, the required precipitation is correspondingly reduced.

TABLE 4. MODERN AND PAST ESTIMATES OF FRACTIONAL LAKE AREA (a_w), DISCHARGE (D), AND MINIMUM BASIN-AVERAGE PRECIPITATION (P) FOR FOUR EAST AFRICAN LAKE BASINS FOR SELECTED TIMES

	a_w	D (mm/yr)	P (mm/yr)	$\Delta P(1)$ (mm/yr)	$\Delta P(2)$ (mm/yr)
Turkana					
Modern	0.05	0	750		
10,000–7000 yr B.P.	0.13	+ ?	830–890	(+80)	(+140)
Nakuru–Elmenteita					
Modern	0.03	0	900		
10,000–8000 yr B.P.	0.44	+ ?	1160–1200	(+260)	(+300)
Naivasha					
Modern	0.04	0	900		
9200–5650 yr B.P.	0.19	+ ?	990–1055	(+90)	(+155)
Victoria					
Modern	0.26	110	1250		
Before 12,500 yr B.P., 10,000 yr B.P.	0.23	0	1125–1070	(−125)	(−180)[a]
About 1880 A.D.	0.26	280	1420–1470	(+170)[b]	(+220)

Note. The precipitation change (ΔP), compared to modern, is given for two assumptions: $\Delta P(1)$, changes in fractional lake area alone; $\Delta P(2)$, changes in fractional lake area plus 10% decreases of land albedo and land Bowen ratio. The range of the past estimate of P is equal to the difference between $\Delta P(1)$ and $\Delta P(2)$.

[a] This estimated change is based upon 10% increases of land albedo and land Bowen ratio.

[b] This estimated change is based upon increased discharge alone rather than increased lake area.

The equations used to simulate the radiation budget have the property that a cloudiness increase reduces the incoming shortwave radiation more than it reduces the outgoing longwave radiation. The net result of a cloudiness increase is therefore to decrease the net radiation and the required evaporation (and precipitation); Table 3, $\delta C = 10\%$, $\delta P \sim -50$ mm/yr. This climatological result does not agree with experience from individual weather events where cloudiness and rainfall should be positively correlated. However, from a climatic and energetic perspective rather than a current weather perspective, the nature of this correlation is not obvious. It depends, for example, on such things as the relative area covered by convective and stratiform clouds, and the diurnal character of the precipitation. Sunshine duration and rainfall records (unpublished series, courtesy of Kenya Meteorological Department) were evaluated for the Kabete and Dagoretti stations in central Kenya and Kisumu at Lake Victoria; these sources suggest a relation between precipitation and cloudiness that is of the same order of magnitude but of opposite sign to the precipitation–cloudiness relationship inferred from the combined hydrologic-energy budget approach (Table 3). While uncertainties of precipitation associated with cloudiness changes are not negligible compared to the magnitude of the precipitation changes estimated in the next section, we have not taken into account the possible role of changed cloud amount on past precipitation estimates (see Benson (1981) for a discussion of the sensitivity of lake evaporation to changes of cloud cover, based upon an energy-balance model).

Another method for estimating sensitivity relationships is to examine the observed changes of precipitation, lake level, and discharge during recent years. We have examined the hydrologic budgets for Lake Victoria for 1961–1964 compared to 1951–1960. The 1961–1964 period experienced high precipitation (about 300 mm/yr above that of 1951–1960), high lake levels, and a large amount of discharge compared to 1951–1960. We found that about three-fourths of the increased amount of precipitation was accounted for by the increased discharge and lake volume, leaving about one-fourth of the amount to be explained by increased evaporation (presumably from the wetter land surface surrounding the lake). A similar result was obtained by Kite (1981). These results reinforce the conclusion of the sensitivity analysis for the closed-basin studies (Table 3) that a change of lake size is likely to be the dominant term in the basin-average hydrologic budget, followed by the correction for the changed evaporation rate over land that would occur if the land Bowen ratio and land albedo changed.

In interpreting the results of the sensitivity analysis (Table 3) it is emphasized that the tabulated change of precipitation (δP) is for 10% changes of a_w, B_l, α_l, and C, and a 1°K change in T_l. The *known* lake area changes were 300 to 1500% (Table 1) whereas 10% changes are perhaps the correct order of magnitude for the other climatic variables (on the basis of their seasonal range, discussed previously). This explains our emphasis of lake-size changes as the dominant term.

PALEOCLIMATIC ESTIMATES

Because the sensitivity analysis showed that the increased fractional lake area during the early Holocene would be the dominant term contributing to estimates of increased precipitation, we first solved Eq. (3) by assuming no changes in radiation or in the Bowen ratio, and obtained an increase of basin-average precipitation of 80 mm/yr (~10%) for Turkana, 260 mm/yr (~30%) for Nakuru–Elmenteita, and 90 mm/yr (~10%) for Naivasha (Table 4). These are minimum estimates because overflow was not estimated quantitatively. For comparison, the increased discharge from Lake Victoria for the period around 1880 A.D. (compared to more recent values) (Table 1) would be equivalent to an increased precipitation of 170 mm/yr (~15%) if evaporation remained unchanged (Table 4).

If, as seems possible, the more humid climate of the early Holocene had been characterized not only by enlarged lakes but also by, say, 10% decreases of land albedo and Bowen ratio, then the resulting increase of evaporation from the land would lead to higher estimates of increased precipitation (Table 4): Turkana, 140 mm/yr (~20%); Nakuru–Elmenteita, 300 mm/yr (~35%); Naivasha, 155 mm/yr (~15%); and Victoria, 220 mm/yr (~20%). Because all of the Rift Valley lakes overflowed and probably also experienced subsurface seepage, the estimated increases of precipitation are minimum values. As previously noted, the direct effect of temperature changes on the net radiation (and therefore evaporation) should have been small.

By way of contrast to the early Holocene period of enlarged paleolakes, we estimate decreased precipitation (compared to modern) of 125–180 mm/yr (~10–15%) for the intervals before 12,500 yr B.P. and around 10,000 yr B.P. when discharge ceased from Lake Victoria (Table 4).

CONCLUSIONS

Based upon our combined hydrologic and energy budget model, the estimated minimum precipitation increases (compared to modern conditions) in East Africa in the early Holocene ranged from 80 to 260 mm/yr (10 to 30%) if only the increase of fractional lake area was considered, and from 140 to 300 mm/yr (15 to 35%) if the effects of possible 10% decreases in land albedo and land Bowen ratio were included. These estimates of minimum precipitation changes (i.e., neglecting overflow) were regionally consistent in the sense that a fairly narrow range of variation of the absolute and percentage increases was found for the three Rift Valley basins. The estimated precipitation increase is largest for Nakuru–Elmenteita, and this may be explained in part by our choice of fractional lake area for that basin, which included the area of the Menengai caldera, an area of possible overflow. If this area had not been included, the early Holocene precipitation estimate would have been about 30 mm/yr lower. On the other hand, if the discharges from Turkana and especially Naivasha were known quantitatively and had been included in the calculations, then the estimates for these basins would have been higher. Thus, the *possible* result of these adjustments would be to narrow the range of the estimates.

The amount of the estimated increase of precipitation for the Rift Valley basins is within the range of modern extremes; for example, the estimated conditions around 1880 A.D. as described for Victoria. Also, the modern rainfall variability, as characterized by the standard deviation of annual rainfall (1961–1979) in central Kenya, is of the order of 350 mm/yr. It is emphasized that the estimates of past precipitation are minimum estimates because overflow was not taken into account; moreover, the estimates would be increased if land albedo and land Bowen ratio were decreased by more than 10%.

Our estimates of the increased precipitation for these Rift Valley basins tend to be somewhat lower than those of others (Butzer *et al.*, 1972; Richardson and Richardson, 1972; see Street (1979) for a summary of the paleoprecipitation estimates for Eastern Africa). Outside of East Africa, and using a combined hydrologic-energy budget approach similar to that used here, Kutzbach (1980) estimated an early Holocene rainfall increase of 300 mm/yr for the Chad Basin in the semiarid fringe of tropical West Africa. From analysis of pollen in lake sediments, a rainfall increase of 200 mm/yr was obtained from Rajasthan, northwest India (Swain *et al.*, 1983).

The results from the sensitivity analysis indicated that temperature changes alone would most likely not have been an important factor in modifying the hydrologic budget. Changes of cloudiness, albedo, and Bowen ratio were shown to possibly have a substantial influence on precipitation estimates, but we did not attempt to estimate these parameters accurately from past conditions. Other paleoenvironmental information (e.g., vegetation, soils) might be helpful in this regard.

As noted in the Introduction, the charts of Street and Grove (1979) indicate that there is widespread evidence for enlarged lakes during the early Holocene across parts of Africa, Arabia, and India; the quantitative precipitation estimates from individual lake basins should be viewed within this broad-scale climatic pattern. The studies reported here for East Africa, coupled with the results from southern Ethiopia (Street, 1979), from the Chad Basin (Kutzbach, 1980), and from Rajasthan, northwest India (Swain *et al.*, 1983) serve to provide guidelines concerning the magnitude of the precipitation changes that must be explained. Street and Grove (1979) suggest that the reestablishment of monsoon rains (after 10,000 yr B.P.) over the Sahara and into the Indian subcontinent could account for the increased precipitation. The first and major of the two rainy seasons in central Kenya and the only rainy season in adjacent parts of East Africa, such as the northern basin areas of Lake Turkana, broadly coincide with the seasonal northward migration of the equatorial trough during the first half of the calendar year. It is within the context of these seasonal circulation changes that the enhanced precipitation—as inferred from the enlarged lakes—should be considered.

Regarding the cause of modifications of the monsoon circulation and precipitation of this magnitude, one factor of possible importance is the changed insolation regime produced by the values of obliquity and the time of perihelion for the early Holocene. This hypothesis is being explored with the aid of a general circulation model (Kutzbach, 1981; Kutzbach and Otto-Bliesner, 1982), and preliminary results show that an enhanced monsoonal circulation and increased precipitation, of the order reported here, could be expected over parts of the African–Eurasian land mass based upon the orbitally produced insolation geometry for that time. It is also of interest that the general circulation model simulation for the early Holocene indicates only small temperature changes for the tropics. Large-scale climate modeling work is however still in an early stage of development and it is likely that a combination of further detailed field work, basin budget studies, qualitative regional surveys, and modeling studies will be most effective for progress in the understanding of the large-scale climatic changes of the early Holocene.

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