Supplementary Materials and Methods:

There are numerous published hydrologic models that account for moisture balance in water bodies. Our modeling effort is a modified version of Yin and Nicholson (1998) for Lake Victoria, which we use to assess hydrologic balance for Lake Victoria and Lake Nasser. Using this modified model, we can then turn to estimating paleorainfall implied by the presence of proposed megalakes in the Sahara during the African Humid Period.

Precipitation falling on a lake basin can be estimated using a basic hydrologic mass balance model for a closed-basin lake assuming steady state conditions.

$$P = E_W * a_W + E_L * a_L \tag{1}$$

where P is the precipitation rate (mm yr^{-1}) over the entire basin, E_W and E_L are the evaporation rates (mm yr^{-1}) from the lake and land surface, respectively, and a_W and a_L represent the fractional area of the basin covered by water and land, respectively. The fractional lake area is expressed $a_W = A_W/(A_W + A_L)$, where A_W refers to the area of the lake and A_L refers to the land area of the basin excluding the lake.

To calculate P, we then need to know evaporation rates from the lake (E_W) and land (E_L). The water evaporation term is much higher than land evaporation in the basins we will model here, because of the higher evaporation rates over water versus land, and high fraction of the basin covered by water. Thus E_w is the most critical evaporation term in calculating P.

Energy balance models require a knowledge of the net solar radiation (R_n) absorbed by a water or land surface and available to evaporate water. R_n is calculated from:

$$R_n = (1-\alpha)R_{sw}-R_{LW} \tag{2}$$

This quantity is highly variable and depends on the net short-wave radiation incident on the surface (R_{sw}) not intercepted by clouds or reflected back due to surface albedo (α), or to long-wave thermal (R_{LW}) radiation. In Africa, R_{sw} has been directly measured by multi-year Meteosat using visible near-infrared data (Ba et al., 2001). Albedo is estimated to be 0.07 for water bodies (Abtew and Melesse, 2013), and R_{LW} can be estimated from (Berliand and Berliand, 1952, cited in Budyko, 1974):

$$R_{LW} = \varepsilon\sigma T_K^4\left(0.39 - 0.058\sqrt{e_{sat}}\right)(1 - 0.54c^2) \tag{3}$$

where R_{LW} is net wave radiation in MJ m^{-2}day^{-1}, ε is a unitless term for surface emissivity, σ is the Stefan-Boltzmann constant (4.9×10^{-9} MJ m^{-2} day^{-1} K^{-4}), T_K is the mean annual water temperature (K) obtained from Yin and Nicholson (1998), e_{sat} is the saturated vapor pressure (in kPa, see calculation under Evaporation section), and c is fraction cloud cover over the water body estimated from monthly 1° x 1° MODIS data from 2007 to 2017 (Fig. S1).

Calculation of Evaporation

We employed two commonly used methods, the Taylor-Priestly and Penman Combination, to calculate evaporation rates from water bodies.

Priestly-Taylor

This evaporation model was first presented in Priestly and Taylor (1972):

$$E = 365\frac{1}{\lambda}\frac{a\Delta R_n}{\Delta+\gamma} \tag{4}$$

where E is evaporation rate in mm yr^{-1}, a is the unitless Priestly-Taylor coefficient of 1.26, Δ is the slope of the saturation vapor pressure-temperature relationship (kPa °C^{-1}), γ is the psychometric coefficient (kPa °C^{-1}):

$$\gamma = \frac{0.001013P}{0.622\lambda} \tag{5}$$

and λ is the latent heat of vaporization (MJ kg^{-1}) calculated from:

$$\lambda = 2.501 - 0.002361(T_a) \tag{6}$$

where T_a is mean annual air temperature (°C). The value of Δ is calculated by:

$$\Delta = 0.04145e^{0.06088T_a} \tag{7}$$

where T_a is the measured mean annual air temperature (°C).

Penman Combination. The Penman Combination (Penman, 1948) equation widely used for modeling evaporation from water bodies is:

$$E = \frac{R_n}{\lambda}\left(\frac{\Delta}{\Delta+\gamma}\right) + E_a\left(\frac{\gamma}{\Delta+\gamma}\right) \tag{8}$$

where E is the evaporation rate (mm yr^{-1}), R_n is net short and longwave radiation (MJ m^{-2} day^{-1}), and E_a is evaporation due to wind (mm yr^{-1}) calculated from Brutseart (1982):

$$E_a = 0.26(1 + 0.54U)(e_{sat} - e_a) \tag{9}$$

where U is wind speed in m sec^{-1}, e_{sat} is the saturated vapor pressure (in kPa) is calculated from:

$$e_{sat} = 0.6198\ e^{(17.27+Ta)/(Ta+237.4))} \tag{10}$$

where T_a is mean annual air temperature (°C), and e_a, the measured mean annual vapor pressure, is reported in Yin and Nicholson (1998).

Estimating evapotranspiration and evaporation from land (E_L)

Evapotranspiration by plants and evaporation from soils is the other key moisture loss term in Equation 1. There are many ways to calculate this. The challenge is that E_L is not a fixed term but varies as a function of rainfall amount, soil and rock type, topography, and vegetation cover. Rainfall is a key determinant, and in general, the higher the rainfall rate, the lower the proportion of rainfall that is evapotranspired, and hence the higher the proportion of rainfall that is converted to run-off. An early quantification of this fundamental relationship between rainfall and run-off was presented in Budyko (1974), and it has been validated by many studies since then. We have chosen the empirical method suggested by Zhang et al. (2004), because it is based on a very large observational dataset (331 catchments from SE Australia) from a range of vegetation types:

$$E_L = P(1 + w\left(\frac{E_0}{P}\right))/(1 + w\left(\frac{E_0}{P}\right) + \frac{P}{E_0}) \qquad (11)$$

where E_L is the evapotranspiration rate (mm yr^{-1}), P is the precipitation rate (mm yr^{-1}), w is an empirical coefficient set at 2.53 for mixed forest and grassland, and E_0 is the maximum potential evapotranspiration rate (i.e. $E_0 = E_w$) determined from the Priestly-Taylor or Penman combination models. P can be solved for iteratively, in our approach by varying P until a_w calculated equals a_w observed.

Results and Discussion

Lake Victoria model

The Lake Victoria Basin as been measured and modeled previously by (Kite, 1982), Hastenrath and Kutzbach (1983), Howell et al. (1988), Yin and Nicholson, 1998), and Swenson and Wahr, 2009). Our intent in this paper is not to create yet another independent hydrologic model but instead generate a model patterned after Yin and Nicholson (1998) that incorporates more current hydrologic and radiation measurements of the basin and region. We present in Table S1 the model inputs and outputs for the Lake Victoria Basin. A few of the parameters require some further justification and discussion here. As described in the main body of the text, for modeling purposes in equation (1) we treat Lake Victoria as closed to overflow, with all losses due to lake evaporation, and a consequent calculated a_w of 0.47. Lake Victoria receives about 27.8 MJ m^{-2} (Budyko, 1974, Table 3, p. 46-47) solar radiation annually at the top of the atmosphere and this is reduced (R_{sw}) to ~ 19 MJ m^{-2} by cloudiness, as measured by Ba et al. (2001). Our calculation of R_{LW} in equation 3 uses fractional cloud cover (c) of 0.63 (Fig. S1), higher than that 0.5 estimated by Yin and Nicholson (1998).

Our modeled lake evaporation rates (E_w) for Lake Victoria vary between 1909 (Priestly and Taylor, 1992) and 1520 mm yr^{-1} (Penman Combination). For comparison, we included model calculations using the methods described in Hastenrath and Kutzbach (1983) approach, which yielded E_w = 1932 mm yr^{-1}. We favor the results of Penman Combination equation because it involves fewer assumptions and uses more direct meteorological measurements than the other models. It also yields the lowest estimates of evaporation rates, making our estimates of rainfall rates conservative. Finally, our estimate of E_w at 1520 mm yr^{-1} from the Penman Combination is close to that of Yin and Nicholson (1998) using an energy balance model (1551 mm yr^{-1}) and their own parameterization of the Penman Combination equation (1743 mm yr^{-1}). More recently, Swenson and Wahr (2009), using satellite rather than station data, and an energy balance model, estimate E_w of 1784 mm yr-1.

Our modeled rainfall estimate (P) for the Lake Victoria basin obtained by iteration of equations 1, 8, and 11 is 1280 mm yr^{-1}, which compares well to P = 1353 mm yr^{-1} from station data compiled for 1956-1978 (Yin and Nicholson, 1998). Our modeled run-off coefficient for the Lake Victoria basin is 0.17, can be compared to an estimate of ~0.2 from Awange et al. (2007) and <0.2 from Yin and Nicholson (1998).

In summary, our hydrologic model of the Lake Victoria basin using the Penman Combination equation for lake evaporation and equation (13) from Zhang et al. (2004) for basin precipitation rates and run-off agrees well with previous models and with the climatologic data available for the basin.

Lake Nasser Model

We use Lake Nasser to simulate the rainfall (P) required to support megalakes in the Sahara under current locally hot, dry and windy Saharan conditions, in the absence of any kinds of feedbacks such as observed around Lake Victoria. As feedbacks certainly did operate to reduce evaporation rates in the AHP, this approach sets an upper limit on the paleorainfall required to support the megalakes. We use the meteorologic measurements presented in Omar and El-Bakry (1983) to estimate evaporation rates from Lake Nasser. Our models (Table S2) yield evaporation rates of 2400 to 2700 mm yr^{-1}, which compares closely to 2700 mm yr^{-1} in Omar and El-Bakry, and to 2400 mm yr^{-1} from remotely sensed estimates of evaporation rates (Hassan et al., 2013), and somewhat more than 2200 mm yr^{-1} in Elsawwaf et al. (2010) for a broad range of lake stations. In the region today, there is virtually no natural run-off locally, and so P equals rainfall directly onto the lake, which also must be >2400 mm yr^{-1} to balance lake evaporation in Lake Nasser today and in the megalakes on the past.

Sudano-Saharan Model

Pollen studies suggest that Sudano-Sahel vegetation covered much of the Sahara south of 20°N during the AHP. Our model (Table S3) assumes vegetation and climate conditions in the modern Sudano-Sahel zone in order to examine the rainfall requirements to form the megalakes. We chose a zone in the heart of the Sahel at N12° and E22° as representative of AHP climate. R_n in this area is ~23 MJ m^{-2} day^{-1} measured from satellite data (Ba and Nicholson, 2004), and cloud cover is ~42% based on MODIS data (Fig. S1 and Table S3). To ensure that our analysis is conservative with respect to paleorainfall (P) and evaporation (E_w) estimates, we (1) adopted the temperatures and wind strength from the Lake Victoria model, even though the Sahel today is generally hotter and windier than Lake Victoria, and probably was more so in the AHP, and (2) used the lower evaporation rate estimate from the Penman Combination Equation. Even using these conservative assumptions, we estimate that E_w had to have been about 1800-2300 mm yr^{-1} to sustain megalakes Chad, Darfur, and Fezzan (Table S3). This in turn yields P estimates 1240 to 1440 mm yr^{-1} and run-off coefficients of 0.13-0.16, probably unrealistically high for the sandy substrate of much of the Sahara.

We would stress that these are minimum estimates of paleorainfall, given the conservative assumptions of our analysis. And they are much higher than the 200-400 mm yr^{-1} implied by pollen reconstructions of Sudano-Saharan vegetation.

Our estimates of paleorainfall required to sustain mega-Lake Chad are nearly double the rate (650 versus 1240 mm yr^{-1}) estimated by Kutzbach (1980). Our higher estimates mainly arise from: (1) the smaller catchment of the Chad Basin obtained from ourDEM analysis compared to that of Kutzbach (1980), and (2) higher R_n for the basin obtained by direct measurements by Ba et al. (2001).

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Figure captions

Figure 1 Fraction cloud cover for Africa was obtained from MODIS (see King et al., 2013), a monthly product of the fraction of clouds on a 1° x 1° grid based on mean of daily mean of pixels, averaged from 2007 to 2017. For hydrologic modeling of Lake Victoria (black dot and outlined in blue) we use the reported the mean above the Lake Victoria region is ~63%; and 42% for our Sudano-Sahelian model at ~N12°, E22°, indicated by filled dot, in the heart of the modern Sudano-Sahelian vegetation belt. The location of Lake Nasser is also shown.