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The physics of the warming of Lake Tanganyika by climate change

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Abstract

Climate warming over the 20th century has increased the density stratification and stability of Lake Tanganyika, a deep rift valley lake. Here we examine the physical processes involved in and affected by the warming of the lake, and we discuss effects on lake productivity. The rate of net heat absorption by Lake Tanganyika has been 0.4 W m^{-2} since 1913, twice the rate in the global ocean, indicating stronger climate forcing in the East African region. Lakes warm through increased incoming long-wave radiation. While lakes in general will increase heat outputs in a warming climate, heat outputs will increase more slowly in deeper lakes than in shallower lakes. Temperatures have increased by 0.2°C at 1000 m in depth, in part because of reduced cool marginal inflows, while water surface temperatures have increased by about 1.3°C. This differential heating over depth has increased the density gradient through the water column, reducing the potential for vertical mixing and thereby limiting nutrient fluxes to the phototrophic zone. An increase in transparency, indicating a reduction in productivity as a result of the reduced vertical mixing, occurred both in Lake Tanganyika and in Lake Malawi, a similar deep tropical lake in which warming has also been documented.

The consequences of climate warming for ecosystems have received the greatest attention in ecosystems at temperate and higher latitudes. Evidence for climate warming effects over the last century in the terrestrial tropics has been limited (Hecky et al. 1994; Tyson et al. 1999), but past and future warming could have substantial consequences for deep tropical lakes. Climate warming through the 21st century in East Africa is projected to be on the order of several degrees Celsius, and of similar magnitude to temperate regions (Mitchell and Hulme 2000; Hulme et al. 2001). Such climate warming could have significant effects on deep tropical lakes as warming surface waters lead to increasing differences in water density that will reduce vertical circulation in these deep, already-meromictic tropical lakes, resulting in reduced rates of nutrient renewal and reduced vertical penetration of oxygen. Observations in all major oceans indicate that the major cause of a reduced oxygen content in the oceans is reduced vertical circulation by global warming (Keeling and Garcia 2002), and deep tropical lakes are already anoxic through much of their depth, so decreasing depths of anoxia may be even more significant for the biota in tropical lakes.

Lake Tanganyika is a deep rift valley lake in East Africa (maximum depth 1470 m; Fig. 1), the second largest lake (by area) in Africa and the second deepest lake in the world. It is located between 3°S and 9°S, is 670 km long, and has a maximum width of 48 km. There are two deep basins (Fig. 1), one in the north and one in the south, each of which is over 1300 m deep. The lake is meromictic (i.e., seasonal mixing affects only the upper layers, while below 100–200 m in depth the lake water is permanently anoxic and relatively rich in nutrients) (Hecky et al. 1991; Edmond et al. 1993). Vertical exchange and especially upwelling driven by southeast trade winds provide the prime nutrient source to sustain primary production by internal nutrient loading (Coulter 1991). Upwelling and limited vertical mixing through metalimnetic entrainment occur primarily in the cool dry season (May–September), at the south end of the lake, driven by strong trade winds from the southeast. A warm, rainy season lasts from October to April. During the rainy season, stratification is similar along the length of the lake, and rates of vertical exchange are low (Coulter 1991). Verburg et al. (2003) compared temperature profiles taken in the north basin in 1994–1997 and in 2000 with historical data and showed the effects of global warming on productivity caused by the increasing vertical density gradient. Here we examine the physical processes involved in and affected by the warming of the lake and the development of the density structure in the water column; we also discuss effects on nutrient limitation and productivity.

Methods

We examined historical changes in air temperatures and wind speed, lake temperatures and heat gain, stratification (by inspecting potential density gradients and Schmidt Stability), surface heat fluxes and their contribution to heat gain, deep water formation, and underwater light attenuation. Historical data related to temperature profiles in the north basin are available for the warm seasons (January–March) of 1913 (Stappers 1913), 1938 (Beauchamp 1939), 1947 (Leloup 1949), 1973 (Craig et al. 1974), and 1975 (Edmond et al. 1993) as well as for the 1990s (to be presented here). In addition, temperatures at depths >500 m are available for the cool season of July 1913 (Jacobs 1914). Historical data from the south basin are more scarce and are not considered here. Water temperature was recorded at 11 different depths from 1 m to 300 m during the period extending from March 1994 to November 1996 by a thermistor string (Aanderaa Instruments) suspended from buoys in the north (recording every 30 min) and south basins (recording every hour). Wind speed and air temperature were recorded on top of the buoys, while

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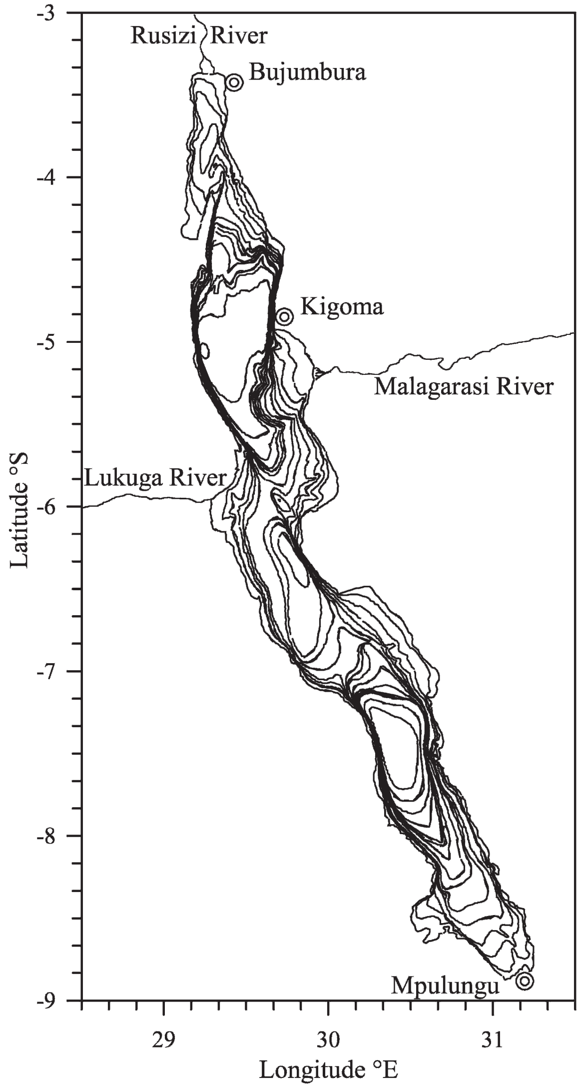

Fig. 1. Lake Tanganyika. Isobaths (every 100 m) were digitized from a map in Coulter (1991).

solar radiation, air pressure, and relative humidity were recorded every 10 min at weather stations (Aanderaa Instruments) at the north and south ends of the lake from March 1994 through November 1996. The recorded meteorological data were used to estimate heat fluxes through the lake surface using methods described in Brutsaert (1982). Two deep-temperature profiles (circa 1100-m depth) were collected in the north basin in August and September 2000. The historical temperature data sets and the methods for temperature data collection from 1994–2000 in the north basin offshore from Kigoma are described in more detail in Verburg et al. (2003).

Density calculations were made following the method of Chen and Millero (1977) using potential temperatures calculated as in Chen and Millero (1986) and salinity from Millero (2000). Conductivity in 2000 varied from 0.680 mS cm^{-1} in the surface waters to 0.706 mS cm^{-1} near the bottom. Beauchamp (1939) in 1938 and Kufferath (1952) in 1946–1947 found a similar range. Therefore, salinity (about 0.6‰) was assumed not to have changed during the past century, which is not surprising in view of the lake's great volume and slow flushing time. Water temperature data collected in January–February 1913 and in 1938 were potential temperatures and were converted to in situ temperatures following the method of Chen and Millero (1986).

Heat content in the warm season and net heat gain since 1913 were determined by integration over depth, with temperatures above 110-m depth in 2000 extrapolated from the change in the February–April temperatures between 1913 and 1996. Schmidt Stability was calculated following Idso (1973) from density while accounting for salinity.

Wind speeds for 1948–2004 were calculated from daily mean eastward and northward wind speeds made available by the National Centers for Environmental Protection/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996) for 12 locations around the lake along a 2.5° grid, between 2.5°S and 10°S latitude and between 27.5°E and 32.5°E longitude. In addition, monthly mean wind speeds measured at 12 m above the ground between 1965 and 1992 and monthly mean air temperatures from 1964 to 1991 were collected at Bujumbura Airport at the north end of the lake. Surface-water temperatures were measured daily at 08:00 h in 1993 to 1997 on a jetty in a small bay at Mpulungu (Fig. 1), at the south end of the lake.

Results

Air temperatures—Annual air temperatures at Bujumbura correlated well with global mean temperature anomalies (Fig. 2; $R^2 = 0.75$, $p < 0.0001$) and equally well with southern-hemisphere temperature anomalies ($R^2 = 0.75$, $p < 0.0001$). Warm-season temperatures increased more and more consistently than did cool-season temperatures (Fig. 2). On a monthly basis, increases in temperature and correlations were strongest for April and November and least for July. Annual rainfall did not correlate with warm-season air temperatures ($p > 0.05$).

Lake warming—There has been a clear warming trend in deep water in Lake Tanganyika over the last century (Fig. 3). Because of meromixis, great depth, and the related long flushing time (volume divided by outflow) of 4000 yr, the response to climate warming is attenuated with depth, and the effect of annual meteorological variability is smoothed by integration of variable surface heat fluxes over depth and time. Time series of water temperatures below 100 m in depth show a continuous increase over the 20th century. Densities decreased more in shallower water (Fig. 4), and consequently there has been a substantial increase in the density gradient with depth over the past century.

In Lake Tanganyika, the slight salinity gradient has a marked effect on stability. In spite of very low salinity and the small salinity change through depth from 0.57‰ to 0.63‰, the increase in the density gradient was smaller than the increase in the gradient in potential temperature. The gradients in potential temperature (Fig. 4) increased by 218% over the depth range of 110 to 200 m, 247% from 110 to 800 m, and 282% from 200 to 800 m in depth since 1913. The increases in the gradients in density, taking into

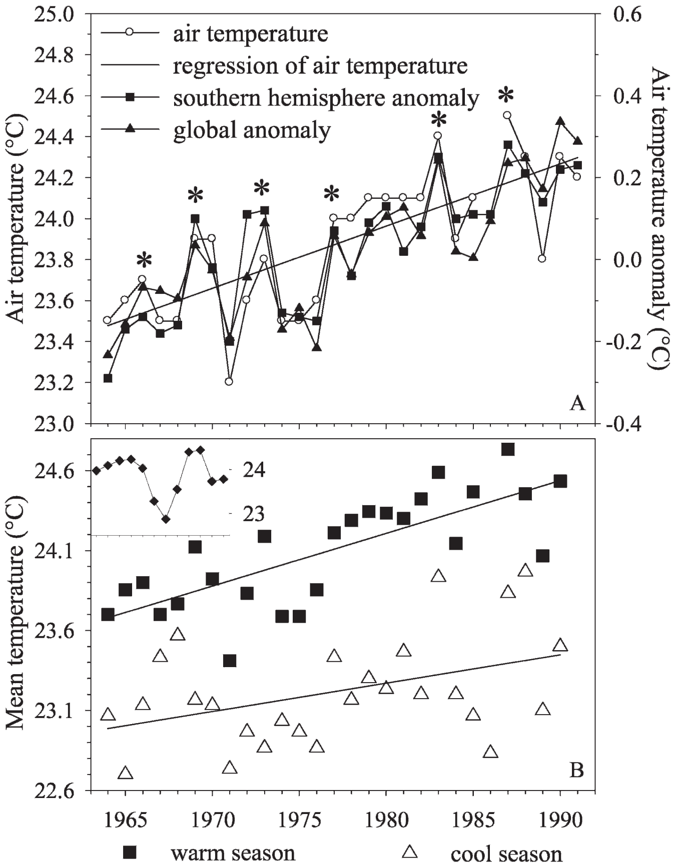

Fig. 2. Air temperatures. (A) Comparison of annual mean air temperatures at Bujumbura with global and southern hemisphere air temperature anomalies (Jones et al. 1999; $R^2 = 0.75$, $p < 0.0001$ for each comparison). High temperatures in El Niño years are indicated (asterisk). (B) Seasonal mean air temperatures at the north end of Lake Tanganyika (Bujumbura, Burundi). Cool season: June–August ($R^2 = 0.17$, $p < 0.05$); warm season: rest of the year ($R^2 = 0.58$, $p < 0.0001$). Inset: monthly mean air temperatures (January–December), Bujumbura.

account adiabatic effects on temperature and assuming salinity = 0‰, over the same depth ranges were 228%, 258%, and 296%, respectively. The increases in the gradients in density, assuming a uniform salinity = 0.6‰, over the same depth ranges were similar: 227%, 257%, and 295%, respectively. The increases in the gradients in density over the same depth ranges (Verburg et al. 2003), taking into account the change in salinity through depth from 0.57‰ to 0.63‰ (Millero 2000), were much smaller: 170%, 179%, and 188%, respectively.

Another indication of a warming trend is found in the comparison of the seasonal minima in surface temperature found in 1993–1996 in the south basin during the upwelling season with earlier literature values. The few published data (Cunnington 1920; Coulter 1963, 1968) consistently reported lower dry-season surface temperature minima offshore in the southern basin in the past (23.45–23.75°C), compared with those recorded with hourly measurements at 1 m in depth by recording thermistors throughout 1993–1996 (minimum 23.85°C, n = 33,630). These recent higher seasonal minimum surface temperatures can be related to an increase in the density gradient in the water column and a reduced vertical mixing potential.

Stability has increased since 1913, but not since 1973–1975 (Fig. 4C). Stability in February 1973 (318 kJ m^{-2}) was similar to the mean in January–March of 1996–2000 (313 kJ m^{-2}; below–300-m depth data of September 2000 were used, and temperatures in the layer above 300 m were extrapolated from data recorded in the warm season of 1996). The change in Stability between the warm seasons of 1913 and 1996–2000 (50% increase) was significant ($p < 0.05$). Daily mean Stability in January–March of 1996–2000 ranged from 282 to 347 kJ m^{-2}. In the cool months of August and September 2000, Stability was 254 kJ m^{-2} and 253 kJ m^{-2}, respectively, lower than the warm-season value of 1947 (257 kJ m^{-2}).

The heat gain over the century over the entire water column (Fig. 5) illustrates the increase in the temperature gradient with depth caused by a relatively lower heat gain as depth increases. The upper 25 m was unusually warm in

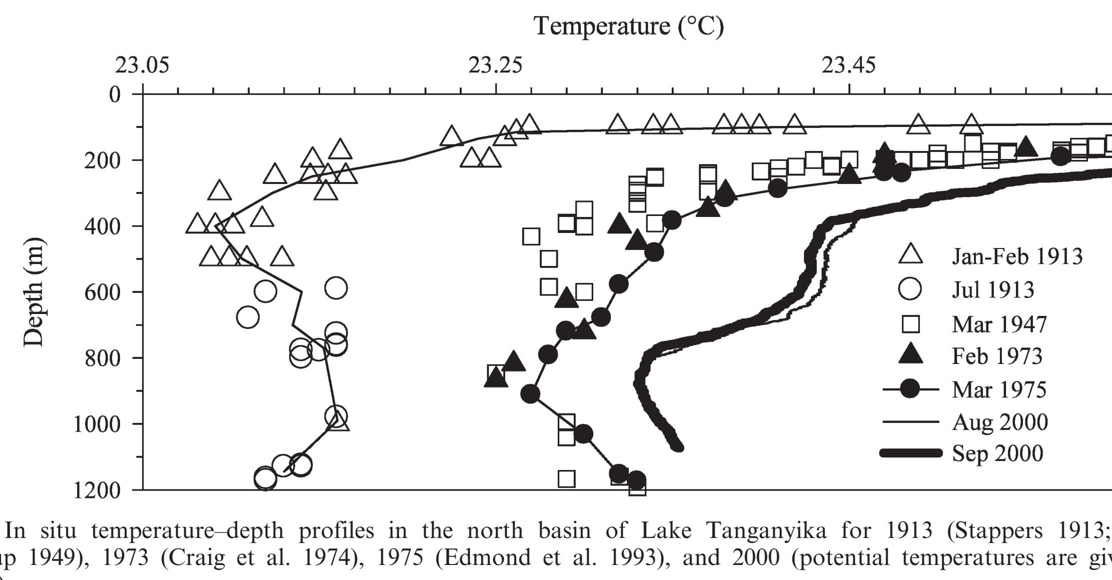

Fig. 3. In situ temperature–depth profiles in the north basin of Lake Tanganyika for 1913 (Stappers 1913; Jacobs 1914), 1947 (Leloup 1949), 1973 (Craig et al. 1974), 1975 (Edmond et al. 1993), and 2000 (potential temperatures are given in Verburg et al. [2003]).

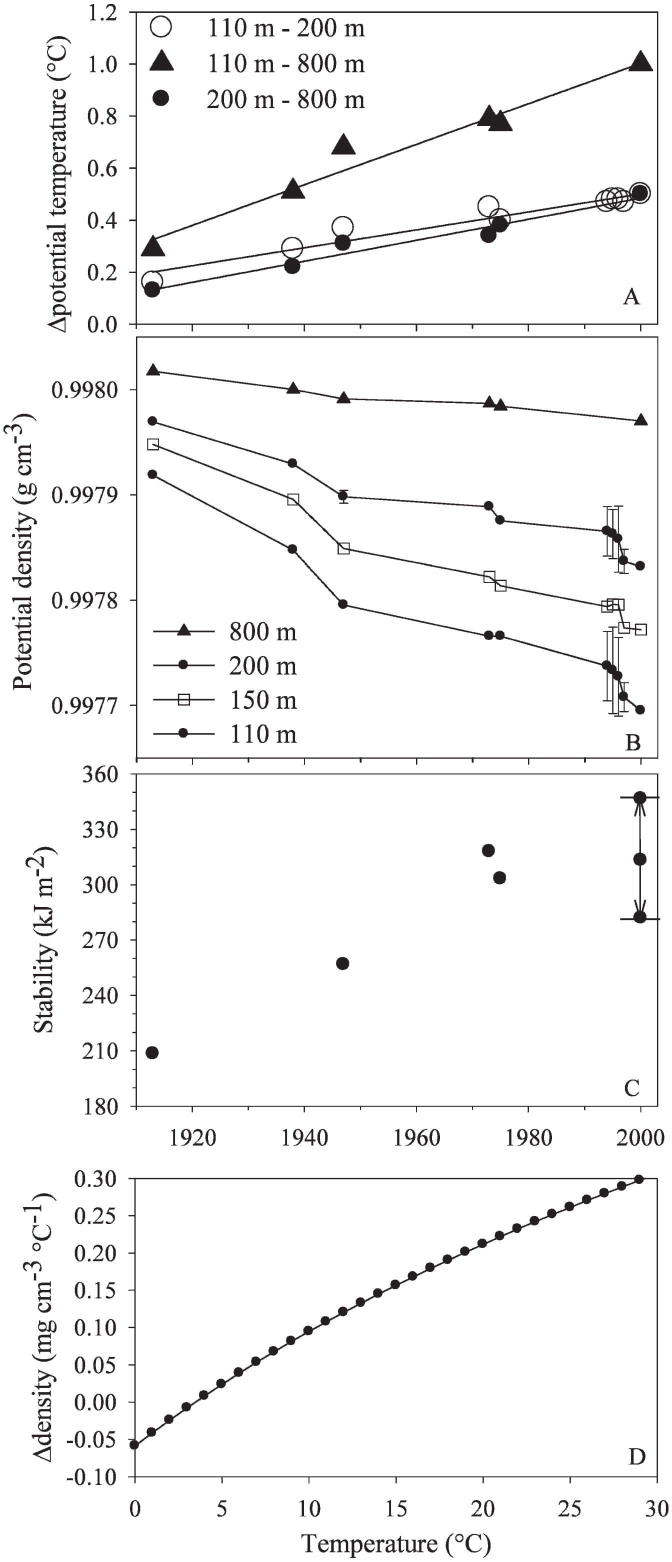

Fig. 4. Increased stratification in the 20th century. (A) Increases in the differences in potential temperature between 110 m and 200 m in depth ($R^2 = 0.93$), between 200 and 800 m in depth ($R^2 = 0.95$), and between 110 and 800 m in depth ($R^2 = 0.96$). (B) Densities at 110, 150, 200, and 800 m. Ranges (vertical bars) between minimum and maximum densities at 200 m are given for 1947 and 1994–1997 (daily means of half-hourly data) ← and at 110 m for 1994–1997. (C) Stability ($R^2 = 0.89$), calculated using only warm-season data (January–March) for 1913, 1947, 1973, and 1975. For 2000, data of the cool-season profiles were used below 300 m in depth, and extrapolation to the surface was carried out using the half-hourly thermistor string data of January–March 1996. The range of daily mean values and the mean are shown for 2000. (D) The relation between water temperature and the decrease in density per degree temperature increase.

the 1975 profile. However, heat gain per year at ≥50-m depth between 1975 and 2000 was similar or larger than that between 1913 and 1975. The mean heat gain in the water column (unweighted for bathymetry) has been 0.36°C since 1913, or 0.0042°C yr^{-1}, equivalent to 22.94×10^6 J m^{-2} yr^{-1} (0.73 W m^{-2}). With hypsometric correction (i.e., the heat gain at each depth multiplied by the area for that depth and the sum divided by the surface area), extrapolated lake-wide mean heat gain over the past century would have been 13.48×10^6 J m^{-2} yr^{-1} (0.43 W m^{-2}, for a total 1.2×10^9 J m^{-2} between 1913 and 2000). This amounts to 0.27% of current annual mean net radiation (Verburg and Hecky 2003). Lake-wide the increase in heat content between 1913 and 2000 was about 3.7×10^{19} J, and the upper 166 m of the lake has absorbed 50% of the heat gain since 1913, taking into account bathymetry (Fig. 5). In the offshore water column, 50% of the heat gain since 1913 has been absorbed in the upper 320 m, which is of relevance for the strengthening of the lake stratification.

Surface-water temperatures measured at a jetty at 08:00 h in a bay at the south end of the lake rarely fell below the potential temperature (23.15°C) at the bottom of the lake (Fig. 6). The water temperature at 08:00 h was below 23.15°C in 1993, 1994, 1995, and 1996 on 7, 5, 0, and 9 d, respectively. The lowest temperature measured was 21.7°C (in 1996), while in 1995 the temperature never dropped below 24.1°C.

There was no significant change in annual mean NCEP wind speeds ($p = 0.62$) or in annual mean wind speeds at 12-m height at Bujumbura ($p = 0.35$; Fig. 7). In particular, no trend was present in wind speeds ($p = 0.27$) during the cool, dry season, when wind speeds are higher and seasonal stratification is weakest.

Discussion

Regional climate warming—Surface air temperature data collection has been sparse in the East African region, and publications discussing the global distribution of climate warming trends over the past century often leave the area of East Africa blank (Rosenzweig et al. 2008 [for the period 1970–2004]). Others (Hansen et al. 2006) suggest climate warming rates in East Africa in the second half of the past century that are similar to those in North America and Europe. The air temperatures measured at the north end of Lake Tanganyika confirm a strong warming between 1964 and 1991, after which no continuous data are available near the lake. While the correlation between global and local air temperatures was strong ($R = 0.87$), the increase in

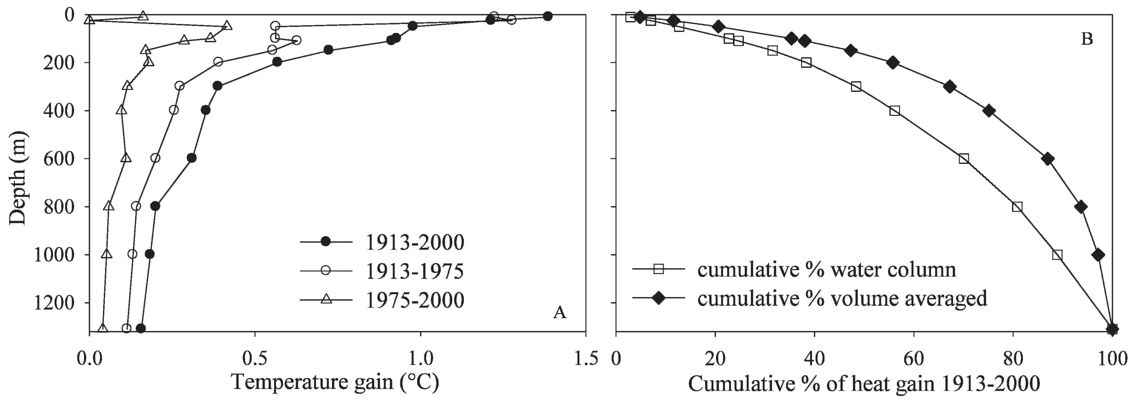

Fig. 5. Heat gain vs. depth. (A) Heat gain between 1913 and 2000 and between 1975 and 2000, using warm-season data. Temperatures above 110 m in 2000 were extrapolated from the trend between 1913 and March 1996, and bottom temperatures were derived by extrapolation. Heat gain between 1913 and 2000 below 600 m in depth may be slightly underestimated, because 1913 data below 600 m were collected only in the south basin, where temperatures below 600 m were found to be warmer than in the north basin in 1975 (by 0.03–0.05°C), because of higher deep mixing rates driven by upwelling at the south end during the southeast trade wind season (Edmond et al. 1993). (B) Cumulative heat gain since 1913 across depth, for volume-averaged heat content and for the center of the north basin. Half of the gained heat is located shallower than 166 m in depth (volume-averaged heat content).

temperature at Bujumbura was almost twice as high compared with the global mean (Fig. 2). By 2005 East Africa had warmed by about 1.2°C (range, 0.8–1.6°C) relative to the 1951–1980 mean (Hansen et al. 2006), much more than the global mean in that period (0.54°C). The widely cited increase of 0.6°C in global mean air temperatures in the last century is derived from both terrestrial air temperatures and ocean surface temperatures (Hansen et al. 2006). Because of thermal inertia and continued mixing with cooler deeper water, the ocean surface warms more slowly than land surfaces, which explains why the area around Lake Tanganyika warmed to a greater level than the global mean in the second half of the past century.

In Africa, temperature increases in the 21st century are expected to be similar to those of other continents (Fig. 8; Mitchell and Hulme 2000). Also, using five different climate scenarios, Hansen et al. (2007) predicted temperature increases in 2000–2100 in East Africa that were similar to or greater than those in North America and Europe in the 21st century. The expected temperature increases in East Africa in the 21st century are far greater (Fig. 8) than those that have already been experienced, indicating that any effects detectible over the last century will be dramatically increased in the present century.

Lake heat content increase—The temperature profiles measured down to the bottom of Lake Tanganyika in 2000 are the only deep temperature data published since 1975. This is the first paper with a climate warming focus to present the temperature–depth profiles of Jacobs (1914) for 1913 and of Leloup (1949) for 1947 as evidence for the warming of Lake Tanganyika. In particular, the data of Jacobs (1914),

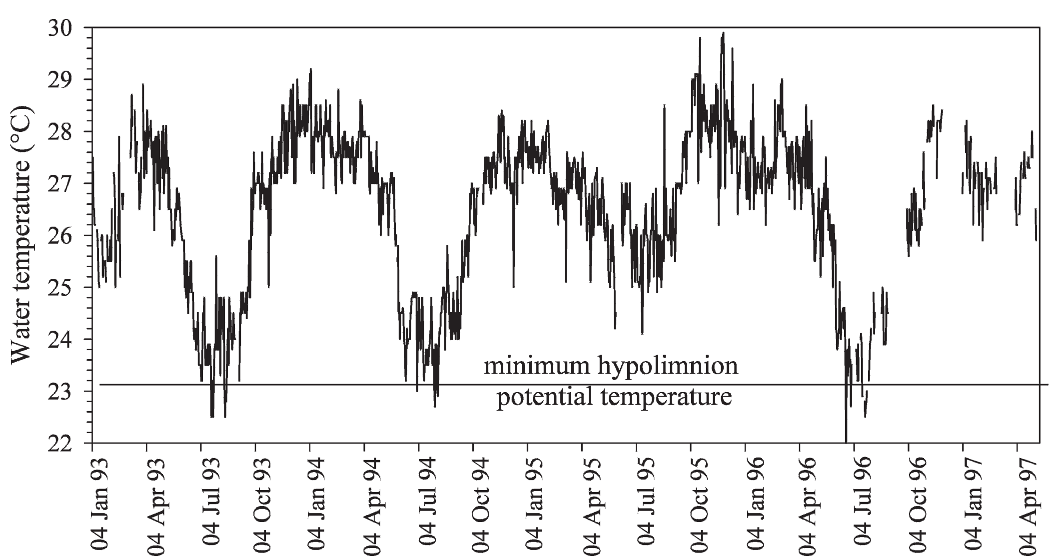

Fig. 6. Daily surface-water temperature in Mpulungu harbor at 08:00 h. The recent minimum potential temperature in the hypolimnion is indicated (23.15°C, about 0.2°C lower than the in situ temperature at 1200 m in depth).

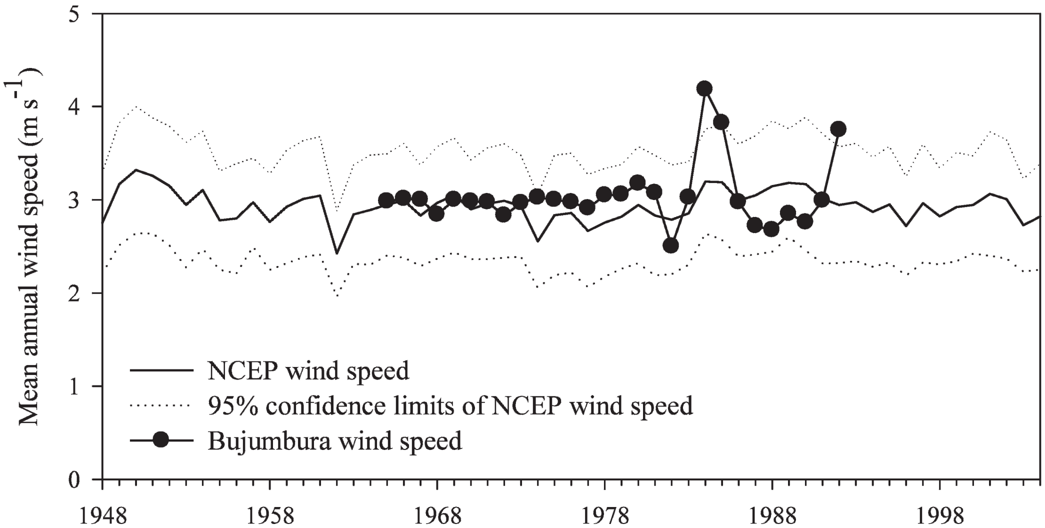

Fig. 7. Annual mean wind speeds (NCEP), averaged for 12 latitude–longitude positions in the area of the lake (between 27.5°E to 32.5°E and 2.5°S to 10°S), with 95% confidence limits, and annual mean wind speeds measured at 12-m height at Bujumbura airport.

collected by a German expedition, independent of the Belgian expedition of Stappers (1913) and only 5 months later, provide very strong support for the magnitude of the warming in Lake Tanganyika, because the data of Stappers (1913) and Jacobs (1914) are very similar (Fig. 3).

Lake Tanganyika contains about one sixth of the earth's liquid freshwater, and while small relative to the ocean, the amount of heat absorbed by the lake is substantial. The heat gain of the lake was 0.021% of the heat gain of the world ocean and 0.020% of the global heat gain. This amount was similar to the heat absorbed by the loss of sea ice from the northern hemisphere over the course of 40 yr: 2.4×10^{19} J m^{-2} (Levitus et al. 2001). The average rate of heating in Lake Tanganyika since 1913 (0.43 W m^{-2}) was greater than that in the global ocean (0.2 W m^{-2}; Hansen et al. 2005) since 1955.

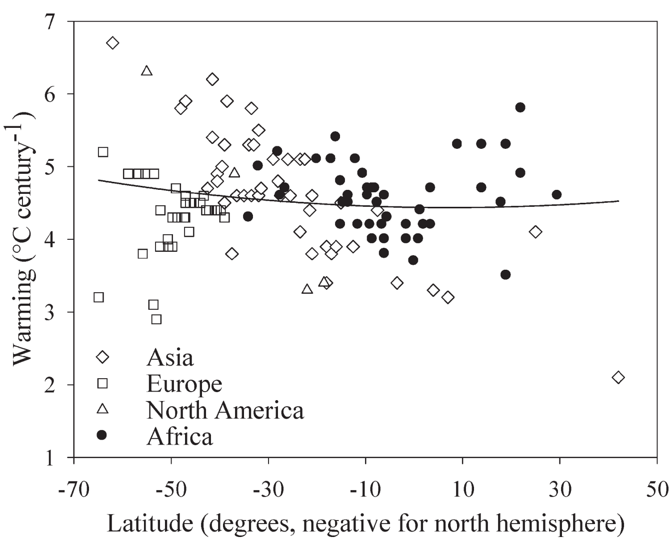

Fig. 8. Mean expected air temperature increases over the 21st century in 149 countries on four continents (mean estimate of five global climate models). Data from Mitchell and Hulme (2000).

Profiles of potential temperatures did not show an inversion (Verburg et al. 2003) but were fairly constant below 900 m in depth, with no indication of heating from the bottom. Therefore, the geothermal heat flux can be considered minimal, and we have no evidence that a geothermal heat flux might supply significant heating from below in this deep rift valley lake. Degens et al. (1971) reported geothermal heat flux rates for Tanganyika (mean = 0.04 W m^{-2}) that were near or below the world average, which is 10-fold smaller than the net rate of heat gain since 1913. In addition, while the lake warming is the result of *changes* in fluxes, there is no evidence for a change in the rate of geothermal heating, and therefore no evidence that it has contributed to the long-term warming of the lake.

Surface heat fluxes—Lakes warm and store more heat when heat inputs exceed heat outputs. Increased heat inputs will generally lead to increased heat outputs until the atmosphere and water surface are in equilibrium. Strong and complex negative feedbacks exist in the coupling of heat fluxes across the atmosphere–water interface that regulate the temperature of the water surface and heat content in lakes. Vertical mixing in the water column allows transport of absorbed heat to deeper water layers, at a rate determined by depth and water-density gradients, enhancing the potential of heat uptake by lakes.

An interesting question is how heat fluxes across the surface of lakes will trend with climate warming; another interesting question is to what extent the net warming of lakes is a result of increases in heat inputs and of changes in outputs of heat. The heat balance in lakes is determined by heat inputs by absorbed solar (S) and incoming long-wave radiation (L_{in}) and heat outputs by outgoing long-wave radiation (L_{out}), sensible (H) and latent heat (E) fluxes, and change in heat storage (the net surface flux, F_S), thus:

$$S + L_{in} = L_{out} + E + H + F_S \quad \left(\text{units W m}^{-2}\right) \quad (1)$$

In Lake Tanganyika, the average heat gain, F_S, since 1913, 0.43 W m^{-2}, while large on a global scale, amounts to only 0.27% of annual mean net radiation ($S + L_{in} - L_{out}$ = 161 W m^{-2}; Verburg and Hecky 2003) and 0.07% of mean radiative inputs ($S + L_{in}$ = 606 W m^{-2}). Therefore, the difference between changes in heat inputs and outputs that would account for the heat gain in the lake over the past century is very small. The required changes in surface heat fluxes are so small that they would be difficult to measure directly.

Using equations to determine L_{out}, E, and H from wind speed, air pressure, relative humidity, and temperatures of water and air, using the methods of Brutsaert (1982), the total amount of heat output through the lake surface can be estimated. L_{out} increases with the fourth power of water surface temperature and does not depend on other factors. Heat output through evaporation, the mass equivalent of E, to the atmosphere is proportional to the vapor pressure deficit, which is higher at higher temperatures but decreases when air temperatures increase substantially more than water surface temperatures. H can result either in a heat input or a heat output depending on the direction of the gradient in temperature between air and the water surface. H is proportional to the difference in temperature between water and air and will result in a heat output when air temperatures are lower than the water surface temperature. In Lake Tanganyika, as in many other lakes, surface-water temperatures are usually higher than air temperatures (Verburg and Hecky 2003), and as a result, under present conditions, H on average presents a loss of heat from the lake, and E and L_{out} always present a loss of heat. L_{out} always increases when a lake warms, but E and H may decrease, which would contribute to lake warming.

Modeling of heat output rates in a warming climate (Fig. 9A), using aerodynamic transfer functions for E and H that account for fluctuations in atmospheric stability (methods in Brutsaert [1982]), shows that as long as the increase in water surface temperature exceeds about 63% of the increase in air temperature, and with humidity and wind speeds held constant, the total outputs of heat (E + H + L_{out}) at Lake Tanganyika will increase with increasing air temperature (Fig. 9A). When E alone is considered, water temperature must increase by at least 68% of the increase in air temperature for heat output by E to increase. Heat output by H, on the other hand, decreases, unless water temperature increases by the same amount or by an amount greater than air temperature. Changes in humidity or wind speeds would modify the surface flux response to warming. However, relative humidity is not expected to change in a warming climate (Dessler and Sherwood 2009), and while an earlier report of a decrease in wind speed at Lake Tanganyika (O'Reilly et al. 2003) was based on a data series with many gaps, we find in contrast no evidence for a change in wind speed in a higher quality data set of local wind speed, one without gaps (Fig. 7).

If the combined heat output by E, H, and L_{out} increases in warming lakes, then heat inputs by the combination of solar and incoming long-wave radiation must increase more than the heat outputs if the lake is to continue to warm. Long-wave radiation from the atmosphere is expected to increase as a consequence of the increase in atmospheric greenhouse gases. From empirical relations between temperature and incoming long-wave radiation and long-wave radiation emitted from the lake surface

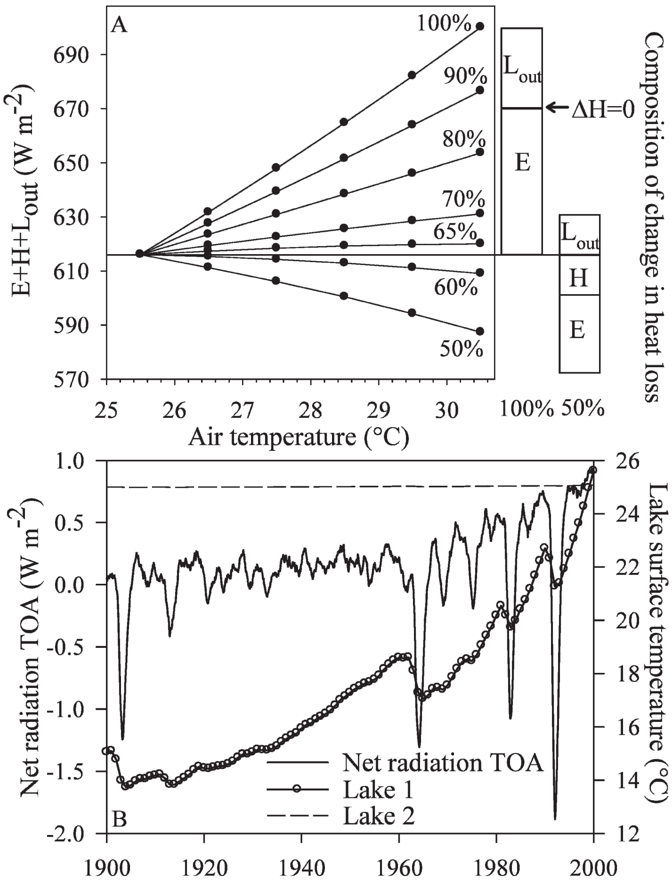

Fig. 9. Lake heat loss and gain. (A) Predicted changes in total heat outputs by latent and sensible heat fluxes and outgoing long-wave radiation with increasing air temperature, for various scenarios of the increase in water surface temperature, varying from 50% to 100% of the increase in air temperature. The contributions to the changes in heat loss by E, H, and L_{out} at water temperature increases of 50% and 100% relative to 5°C air temperature increases are shown by stacked bars. The modeling was initiated at an air temperature of 25.5°C and a water surface temperature of 26.6°C, and mean values for air pressure, wind speed, and relative humidity were used. The change in heat outputs at 5°C air temperature increase was −29 W m^{-2} for E, −15 W m^{-2} for H, and +15 W m^{-2} for L_{out} when water temperature increases by only 50% of the air temperature increase and +54 W m^{-2} for E, 0 W m^{-2} for H, and +30 W m^{-2} for L_{out}, when the increases in temperature of water and air are equal. The horizontal line indicates the level of no change in heat outputs. (B) Net radiation at the top of the atmosphere (Hansen et al. 2005) and surface temperatures of two hypothetical lakes. Heat inputs in the two lakes increase by the rate of net radiation TOA (mean 0.14 W m^{-2} between 1900 and 2000), without allowing change in heat outputs. Depths of lake 1 and lake 2 are 10 and 1500 m, respectively, and starting surface temperatures are 15°C and 25°C, respectively, and both lakes have fully mixed water columns.

(Brutsaert 1982), it follows that if air and water temperatures increase at equal rates, heat gain in lakes by incoming long-wave radiation will increase more than heat loss by outgoing long-wave radiation. The difference between the increase in L_{in} and the increase in L_{out} increases from about 1 W m^{-2} at 10°C to 2.5 W m^{-2} at 30°C, per degree temperature increase.

The net radiation at the top of the atmosphere (TOA) is the difference between heat entering and leaving earth's atmosphere. When net radiation TOA is positive, earth becomes warmer. Net radiation TOA has been positive over the past century, and the earth's energy imbalance is indicative of the rate of climate warming. The mean net flux TOA between 1900 and 2000 was 0.14 W m^{-2} (Fig. 9B). Between 1913 and 2000 the net flux TOA was 0.17 W m^{-2} (Hansen et al. 2005; the data of their fig. 1C are available on the NASA website), similar to the warming rate observed in the ocean (0.2 W m^{-2}; Hansen et al. 2005), which has absorbed >90% of the excess heat introduced to earth, but less than half of the warming rate of Lake Tanganyika. Three explanations can be suggested for heat absorption in Lake Tanganyika in excess of the global average, based on considerations of convection, changes in heat outputs, and local climate forcing: (1) Deeper and more vigorous mixing allows Lake Tanganyika to absorb more heat than the global average, similar to the situation in the ocean. This explanation does not seem sufficient, as Tanganyika does not benefit from the convective mixing that the ocean has in its polar regions. The best evidence for the weak convective mixing in Tanganyika is the lack of oxygen, despite its oligotrophic status, below about 150 m in depth, in contrast to the oxygenated bottom water in the ocean, which, of course, is much deeper. (2) A decline in heat outputs is partly responsible for the warming of Lake Tanganyika. This argument would indicate that apart from an increase in radiative heat inputs, a decrease in heat outputs by the sum of L_{out}, H, and E contributed to the increase in heat content. Such a decline in heat outputs seems plausible. It would require water temperature to increase on average by no more than 63% of the increase in air temperature (Fig. 9A). In view of the increase found in water surface temperature (Fig. 5), an increase of more than 2°C in air temperature since 1913 would be required, which cannot be ruled out considering that the rate of atmospheric warming at the north end of the lake was twice the global rate (Fig. 2). On the other hand, surface-water temperatures would not be expected to increase much more slowly than air temperatures because of reduced downward mixing of heat caused by increased density gradients. (3) Climate forcing in the East African region has been higher than the global average. This explanation seems the most viable. Because the local energy imbalance around the globe cannot be estimated with certainty (Hansen et al. 2005), the question of whether and how changes in heat outputs at Lake Tanganyika may have moderated the net heat uptake response to climate warming is left undecided.

It is clear that for the vast majority of lakes heat outputs must increase with climate warming (Fig. 9B). A hypothetical shallow lake of 10-m depth that warms by the global mean energy imbalance of 0.14 W m^{-2} since 1900 would increase by 11°C in 100 yr if no increase in heat outputs is allowed, in contrast to much deeper lakes, in which the absorbed heat can be transported to deeper water. Realistic surface temperature increases, far less than 11°C, require an increase in total heat output by E, H, and L_{out}, from which it follows that the warming of lakes, the majority of which are shallow, is the result of an increased heat input by incoming long-wave radiation, which is partly cancelled by increased heat outputs. The fact that L_{in} increases substantially more than L_{out} in a warming climate indicates that part of the increased heat outputs must be explained by an increase in E and perhaps in H. On the other hand, it is clear that in the deepest of lakes, unless stratification retains much of the heat nearer the surface, as has been the case in Lake Tanganyika, hardly any increase in surface temperature would be observed (lake 2 in Fig. 9B). As a result, surface temperatures in shallow lakes will probably tend to be slightly above the equilibrium temperature, while in deep lakes they are more likely to be, on average, below the equilibrium temperature. It follows that heat output by H is expected to increase in shallow lakes and decrease in deep lakes.

The recent net global forcing that drives climate change amounts to 0.85 W m^{-2} (Fig. 9; Hansen et al. 2005), six times the mean of the past century (Fig. 9B). Modeling indicates that lakes will absorb less net heat than is applied by local climate forcing (Fig. 9), by enhancement of their heat outputs through the interaction with the atmosphere. Deep lakes, however, have the capacity to absorb more heat than shallow lakes, and in a warming climate heat outputs will increase more slowly in deeper lakes than in shallower lakes. This inherent tendency toward greater heat absorption with little change in surface temperature in deeper lakes will be reduced if the lakes become increasingly stratified as warming progresses, as has been the case in Lake Tanganyika.

The global hydrological cycle is expected to accelerate with climate change (Wentz et al. 2007), and global evaporation has been predicted to increase (Ramanathan 2001), which agrees with our conclusion of increased heat outputs, in part through evaporation, by most lakes. The extent to which evaporation from inland water surfaces will change in a changing climate not only affects the heat budget and stratification of the water column but will also be of great importance to the water balance and water levels in lakes and to the availability of freshwater in a warmer world.

Depth of in situ temperature minimum—In 1975 and 2000 the depth of the in situ minimum temperature, below which temperature increases as an effect of pressure (adiabatic heating), was found at around 900 m in depth (Fig. 3). Capart (1952) found the minimum in 1946–1947 to be between 500 and 800 m (23.25–23.28°C). This value appeared shallower in 1913, at about 400 m in depth, probably as an effect of a more homogeneously mixed water column in the past, which decreased the depth of the lowest in situ temperature. The potential temperature gradient below 200 m in depth was much less steep in 1913 than in 1975 and 2000. When potential temperatures

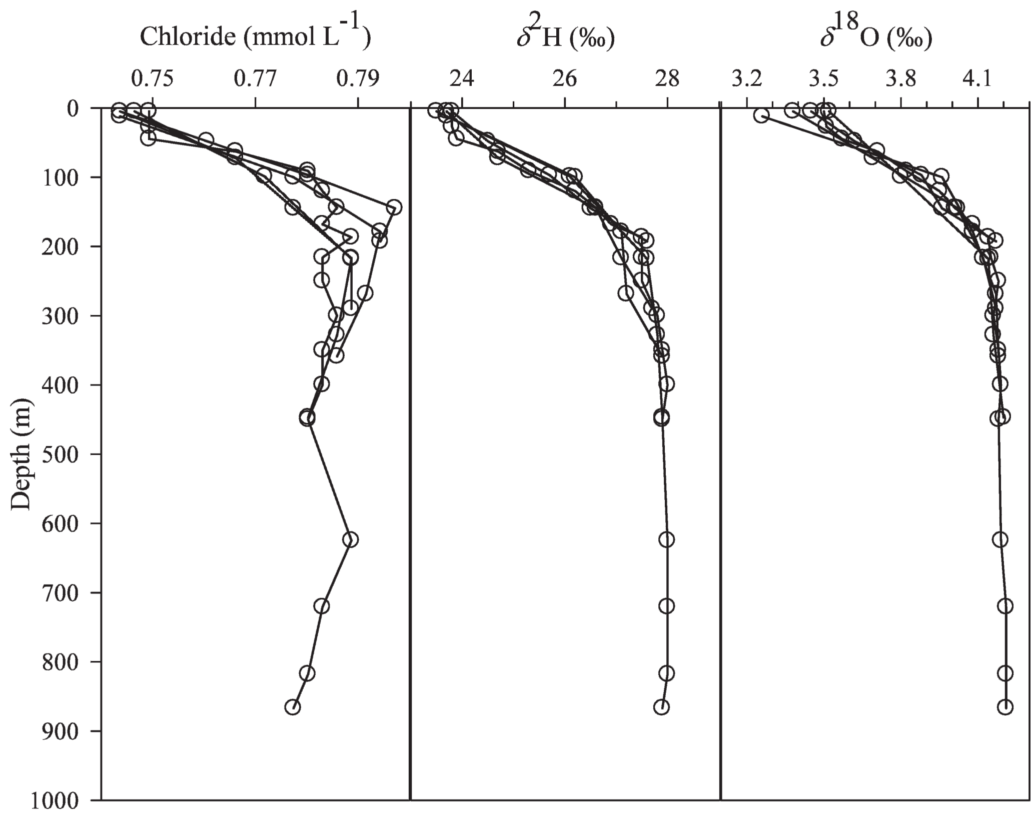

Fig. 10. Depth gradients of chloride, δ^{18}O, and δ^2H at five sites in the north basin. Data from Craig et al. (1974).

become vertically more similar, the depth at which in situ temperatures start to increase with depth will be shallower. A truly mixed water column would have the lowest in situ temperatures at the surface. Therefore, the observation of an increased depth of minimum in situ temperature over the past century is consistent with a warming lake and increasing density gradients that are retarding vertical circulation.

Deep water formation—Craig et al. (1974) described the lake as "isotopically upside-down" because they expected to find higher δ^{18}O and δ^2H values in surface waters than in deeper water as a result of isotopic enrichment, driven by high evaporation in the continuously warm and stratified surface waters. Instead, δ^{18}O and δ^2H increased between 0 and 200 m in depth (Fig. 10). In addition, chloride concentrations show a similar increase with depth (Fig. 10). As Craig et al. (1974) pointed out, river inflows cannot now contribute significantly to the formation of the hypolimnion waters, because the rivers entering the lake are too depleted in δ^{18}O (−2.7‰ [±1.2‰ standard deviations (SD)]) and δ^2H (−10.0‰ [±7.5‰ SD]) and have low chloride concentrations (0.14 [±0.20 SD] mmol L^{-1}) compared with lake water (Fig. 10). In addition, even mountain streams with headwaters several kilometers above the lake surface have mean temperatures over 24°C (Kimbadi et al. 1999), too high to contribute to deep water formation, while the largest inflow, the Ruzizi, has a mean temperature of 25°C (Vandelannoote et al. 1999), almost 2°C higher than the bottom water.

Craig et al. (1974) concluded that the deep water was a "relic hypolimnion," formed in a colder and drier time when evaporation was greater than at present. However, Craig et al. (1974) did not realize that in tropical Lake Tanganyika heat loss by evaporation is strongest not in the warmest season, as is typical for temperate lakes (except for deep Laurentian Great Lakes like Superior, which has maximum evaporation losses in winter), but in the cool season, when the trade winds are strong (Verburg and Hecky 2003). Therefore, the increase in δ^{18}O, δ^2H, and chloride with depth might be caused by convective downwelling of relatively cool surface water in the cool, dry season driven by strong evaporation, which concentrates heavier water isotopes and solutes.

The surface temperature found offshore in hourly records during 1993–1996 in the south basin, where surface temperatures have a lower minimum than in the north basin (Verburg and Hecky 2003), was never below 23.85°C, which is 0.7°C higher than the present potential temperature at the bottom, and was below 24°C only 0.8% of the time (n = 33,630). In contrast, 80% of the lake's volume (14,960 km^3 or 459 m averaged over the lake surface) is cooler than 24°C, and 76% is cooler than 23.85°C (potential temperatures). Deep cool water formation is not likely driven by evaporative cooling under current conditions occurring offshore (Peeters et al. 2003), but is more likely driven by cooling in shallow inshore areas, where surface cooling is more rapid than offshore (Wells and Sherman 2001), followed by convective downflows along the bottom, starting at the lake margins. Such cooling of the hypolimnion by marginal downflows does not require overturn mixing, although these downflows may entrain larger volumes, and does not disrupt stratification, but instead maintains it. However, surface temperatures measured on a jetty in a shallow bay at the south end of the lake (Fig. 6), where minimum surface temperatures are lower than in the north basin, and at a time of day when surface temperatures tend to be lowest (Verburg and Hecky 2003), show that recent atmospheric conditions are rarely conducive to the formation of water sufficiently cool to contribute to deep water formation. Consequently, the bottom water most likely did form in a cooler climate, when marginal downwelling was more frequent. This would agree with the age of the deep water of at least 700 yr, as estimated by Craig (personal communication in Rudd [1980]), substantially older than the mean residence time of 275 yr (derived as total volume divided by the sum of evaporation and outflow, with volume and outflow from Edmond et al. [1993], and an evaporation rate of 1986 mm yr^{-1} [P. Verburg and J. Antenucci unpubl.]). Edmond et al. (1993) estimate the age of the water in the hypolimnion as 800 yr based on the ^{14}C contents of dissolved CO_2. The upper water layer is much younger. Taking the mean age of water below 200 m in depth to be 700 yr, and based on the volumes below and above that depth and the mean whole-lake residence time, a residence time of 114 yr for the upper 200 m follows. During the Medieval Warm Period (circa A.D. 1000 to 1270) East Africa was very dry (Verschuren et al. 2000). During this period the lake was probably closed (Haberyan and Hecky 1987) as a result of strong evaporation. This may be the period when much of the water at present in the hypolimnion of Lake Tanganyika left the surface. The bottom water is now slowly increasing in temperature by downward diffusion of heat from overlying and warming waters.

As a result of warming since the Medieval period and especially rapid warming over the last century, the cool convective marginal flows have come to a virtual standstill. Therefore, the temperature of bottom water is expected to rise slowly and continuously, unlike the irregular 'sawtooth' behavior seen in temperate lakes, where deep mixing in occasional cold years resets the temperature of the bottom water (Livingstone 1997). In temperate lakes hypolimnion heat gain in the warming climate tends to be more irregular. 'Sawtooth' behavior occurs in temperate lakes that in some years overturn completely (Livingstone 1997), which does not happen in meromictic Lake Tanganyika. In the mechanism at work in Lake Tanganyika the ratio of the volume of deep water to the mixed layer volume (Wells and Sherman 2001) and the original temperature of bottom water before the inputs of cool downflows started to decrease will determine the change in the temperature gradient with depth in a warming climate. In other words, when the ratio of mean depth to maximum depth is high, such as is typically found in rift lakes like Lake Tanganyika, the bottom temperature would be expected to increase more slowly (than when the ratio is small), allowing the temperature gradient to increase. In lakes in which the ratio is small, the temperature gradient may instead become smaller, or would increase less quickly, as an effect of climate warming.

Stratification and vertical density gradient—The Stability, defined by Idso (1973) as the amount of energy needed to fully mix the lake without exchange of heat, is naturally enormous for such a deep lake as Lake Tanganyika: 3×10^5 J m^{-2}. The Stability is more than twice that predicted by the relation of Stability with maximum lake depth for smaller tropical lakes (Kling 1988). This is caused by the relatively steep bottom slope of Lake Tanganyika, which ensures that deep water layers contribute a relatively large share to the volume of the lake, compared with other lakes.

The water column between 400 and 800 m has been considered homothermal (Beauchamp 1939; O'Reilly et al. 2003) and has been assumed to vary by no more than 0.02°C (O'Reilly et al. 2003). However, there is a marked and persistent temperature gradient between 400 and 800 m, resulting in a difference of about 0.1°C for in situ temperatures (Fig. 3) and 0.2°C for potential temperatures. This gradient is crucial in determining rates of vertical mixing in this depth region and in determining the rates at which nutrients from bottom waters are returned to the surface layer, where they can become available to phytoplankton. Changes in temperature gradients have a stronger effect on density gradients and stratification in warm tropical lakes than in cool lakes because the decrease in density per 1° temperature increase is much larger at higher temperatures (Fig. 4D), a property of water that is important to density stratification in tropical lakes.

Salinity needs to be taken into account when determining changes in the vertical density gradient, even when salinity is very low and the change in salinity with depth is small. The vertical gradient in salinity, by increasing the density gradient, decreases the estimate of the proportional change in the density gradient resulting from warming. Because the potential for vertical mixing is inversely proportional to the density gradient, the salinity gradient, although small, has a marked effect on vertical mixing. The increase in the vertical density gradient by climate warming has been substantial since 1913, about 2.8-fold, but a larger estimate, about 3.5-fold, would have resulted if the vertical gradient in salinity would have been ignored.

Because there is no seasonality in temperatures below about 150 m in depth, which has also been reported by earlier workers (Capart 1952), it is safe to assume that the change seen in deep water temperatures is the result of a long-term trend such as climate warming and is not affected by seasonal and interannual variability. Shallower temperatures, and in particular surface temperatures, are affected by daily, seasonal, and interannual variability, as illustrated by the similar surface temperatures in 1975 and 2000 (Fig. 5). It appears that the measurements in 1975 were taken on an unusually warm day, although 1975 was not a warm year (below the trend in annual air temperatures; Fig. 2). A comparison of the Schmidt Stability across the century, especially when using only a few temperature profiles and ignoring daily, seasonal, and interannual variability, can be misleading because the calculation of the Schmidt Stability is weighted by volume per depth layer and is therefore especially strongly sensitive to variability in temperatures in the uppermost water layers. This is demonstrated by the large range in daily mean Schmidt Stability during the warm season and by the fact that the Stability in 1973 and 1975 was similar to the mean in 1996–2000 (Fig. 4C). Because of greater variability in shallow temperatures, it is more useful to examine trends in deep water temperature and density gradients (Fig. 4A,B) than to compare shallow temperatures or Schmidt Stability over the past century.

Climate warming, in synergy with the property of water involving high-density differences per °C at high temperatures, and the positive feedback of limiting the heat flux to a narrowing surface layer as meromixis becomes more pronounced and upwelling and convective downflows reduce have increased the density gradient over the course of the 20th century. Increased stability attributable to the steepened thermal gradient in the water column would reduce vertical exchange by turbulent diffusion between surface layers and the deep water of Lake Tanganyika.

Effects on productivity—Phosphorus controls productivity in the lake, and it increases sharply with depth (Fig. 11). Its availability for photosynthesis in the surface layer will determine changes in productivity with climate warming. In contrast, fixed nitrogen is lost by denitrification in the metalimnion and is replaced by nitrogen fixation in the surface mixed layer (Hecky et al. 1991). Because of the large volume and the long nutrient residence times (Hecky et al. 1991), together with the great depth and the anoxic monimolimnion, productivity in Lake Tanganyika depends almost entirely on nutrient supply by internal loading. External loading by inflows is, in addition, relatively unaffected by land use development, as most of the catchment remains sparsely populated (Coulter 1991; Bootsma and Hecky 2003).

If the increase in the density gradient has resulted in reduced vertical mixing in the entire water column, then reduced internal nutrient loading to the epilimnion would lead to a decline in productivity and an increase in transparency, as reported in Verburg et al. (2003). Dissolved silica concentrations have been increasing in the epilimnion since 1975, indicating a more oligotrophic and less productive lake, at least for diatoms, as silica demand and sedimentation by diatoms have evidently been reduced (Verburg et al. 2003). In addition, there is evidence that oxygen penetrates less deeply in the lake, as is indicated by lower SO_4 concentrations in the mixed layer and noticeable H_2S occurring at shallower depths than has been the case in the past (Verburg et al. 2003). There is no evidence for an effect of wind speed on the change in vertical mixing. Climate warming alone has most likely been responsible for the reduced vertical mixing. In Lake Malawi, a similar deep rift lake in East Africa that warmed in the past century (Vollmer et al. 2005), transparency increased at a similar rate as in Lake Tanganyika (Fig. 12), indicating that in Lake Malawi similar changes in the

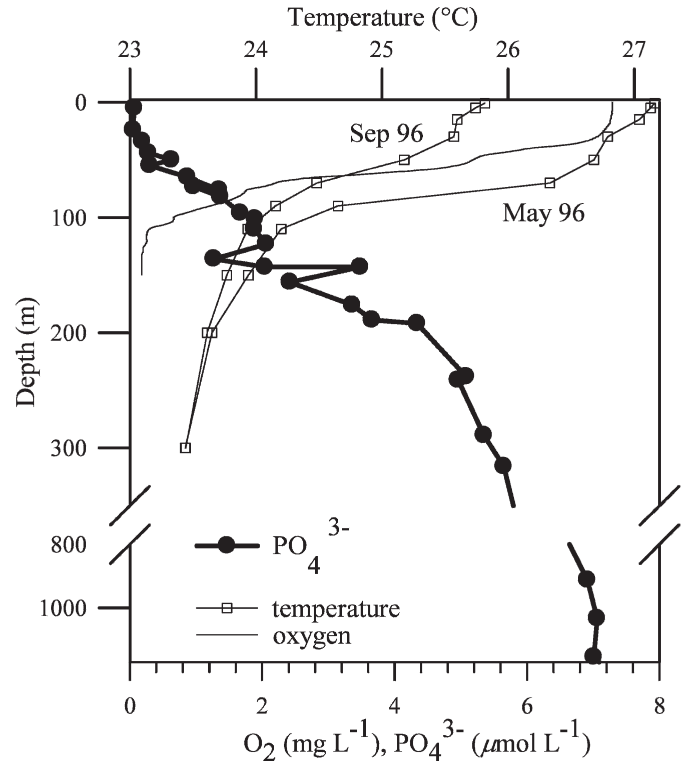

Fig. 11. Depth profiles of oxygen (April 1997), phosphate (March 1975: Edmond et al. 1993), and temperature on days of maximum (May 1996) and minimum (September 1996) heat content in the North basin.

pelagic ecosystem have occurred as a result of climate warming.

Purse seine fishery catch per unit effort in Lake Tanganyika has declined since 1962 and more steeply since 1985 (Sarvala et al. 2006), certainly in part as the result of an increase in fishing effort by other fishing technologies, in particular the light-lift net fishery, which now dominates catches (Sarvala et al. 2006). However, high fishing pressure alone may not be enough to explain the substantial decline in fish catch per unit effort at Lake Tanganyika, especially for short-lived zooplanktivorous *Stolothrissa tanganyicae,* which lives, on average, less than 1 yr and makes up the lion's share of the catch in Tanganyika. In short-lived species a reduced vulnerability to high fishing pressure is generally expected, because fast and massive recruitment can maintain heavily fished fish stocks. Reduced mixing, reduced algal biomass (Verburg et al. 2003), and greater light transparency all indicate that primary productivity may be declining. Such a decline would be expected to reduce fish biomass and yields and contribute to the declining catch per unit effort.

Climate warming has affected the physical structure of Lake Tanganyika and substantially affected the functioning of the pelagic ecosystem. It is important to note that these processes are caused by the fact that air temperatures are increasing as a result of climate warming and not by high temperatures per se. If the climate ceases to warm and stabilizes, although at a higher mean temperature, the lake

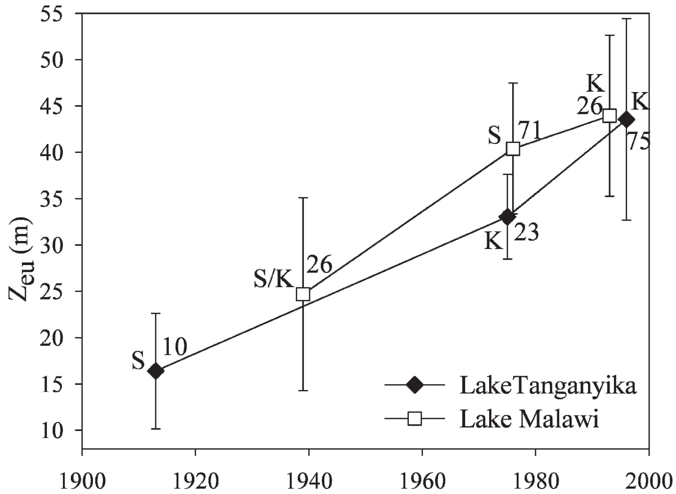

Fig. 12. The change in the mean depth of the euphotic zone (Z_{eu}) offshore was similar in Lake Malawi and Lake Tanganyika. Data for Lake Malawi were taken from Bertram et al. (1942) for 1939, from Beauchamp (1953) for 1939–1940, from Ferro (1977) for 1976, and from Patterson and Kachinjika (1995) for 1992–1994. Data for Lake Tanganyika were taken from Stappers (1913) for 1913 and from Verburg et al. (2003) for 1975 and 1996. Euphotic zone depth was calculated from attenuation K, which was either determined from the vertical absorption with depth of underwater light or from Secchi disk data using the equation in Verburg et al. (2003). Numbers indicate sample sizes, bars indicate standard deviations, and S indicates that original data were Secchi disk data, whereas K indicates that Z_{eu} was calculated from K as $Z_{eu} = \ln(100)/K$.

will reach a new equilibrium state, with, ultimately, a density gradient, internal nutrient loading rates, and levels of productivity that are more similar to those before the climate started to warm and imposed steeper density gradients on the lake. However, even if fossil fuel emissions and other processes affecting the greenhouse effect are substantially curbed in the near future, several centuries of further warming—at much higher rates than seen so far—have become virtually inevitable (Hansen et al. 2005). Although African ecosystems are remote from most of the anthropogenic activities that have driven climate warming, the Great Lakes of Africa and their riparian human populations have not and will not escape its effects.

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