*Limnol. Oceanogr.*, 49(3), 2004, 778–783


# Response of Lake Kivu stratification to lava inflow and climate warming

*Andreas Lorke*[1]
Applied Aquatic Ecology, Limnological Research Center (EAWAG), CH-6047 Kastanienbaum, Switzerland

*Klaus Tietze*
PDT GmbH, Physik-Design-Technik, Sensorik und Consulting, D-29227 Celle, Germany

*Michel Halbwachs*
Université de Savoie et Coordination de la Recherche Volcanologique, CNRS-INSU, F-73376
Le Bourget du Lac Cedex, France

*Alfred Wüest*
Applied Aquatic Ecology, Limnological Research Center (EAWAG), CH-6047 Kastanienbaum, Switzerland

*Abstract*

During the eruption of Nyiragongo Volcano in January 2002 about $10^6$ $m^3$ of lava entered Lake Kivu. The high concentrations of $CO_2$ and $CH_4$ dissolved in the deep waters of Lake Kivu raised serious concerns about a potential gas outburst with catastrophic consequences for the population in the Kivu-Tanganyika region. Therefore, 3 weeks after the volcanic eruption, we performed an ad hoc lake survey of the stability of the water column stratification. Vertical profiles of temperature and turbidity revealed signatures of the lava, which had penetrated to 100 m depth; however, there was no substantial warming or destratification of the gas-containing deep layers below. The deep double-diffusive structures also remained unaltered. Based on these observations, we conclude that a thermally driven gas outburst in Lake Kivu is not to be expected from future eruptions of comparable dimensions. In addition, the recent measurements allowed for an update and gave new insight into the stratification and double-diffusive mixing phenomena in Lake Kivu. A comparison with former measurements revealed a warming of the upper part of the lake of up to 0.5°C within the last 30 yr, which could be attributed to climate variability.

The Nyiragongo Volcano (Democratic Republic of Congo, East-Central Africa) erupted on 17 January 2002. Voluminous lava flows, originating from different eruptive vents, destroyed the central part of the city of Goma and about one million $m^3$ of lava entered Lake Kivu, which is located about 18 km south of the Nyiragongo Volcano in the western branch of the East African Rift Zone (Fig. 1). The lake covers a surface area of 2,400 $km^2$ and has a maximum depth of 485 m. There were serious concerns about a thermally driven limnic gas outburst due to the high concentrations of dissolved $CO_2$ and $CH_4$ at great depth in the permanently stratified water column (Deuser et al. 1973; Tietze et al. 1980). It was feared that the hot lava could warm up and locally destabilize the density stratification in those deep layers and, thus, trigger a massive gas outburst. Such a gas outburst in Lake Nyos (Cameroon) in 1986 (Kling et al. 1987; Tietze 1992), most probably triggered by rock fall, released a dense cloud of carbon dioxide that flowed through the surrounding valleys and asphyxiated approximately 1,800 people (Freeth et al. 1990). Taking into account that the total volume of dissolved gas in Lake Kivu is about three orders of magnitude higher than in Lake Nyos ($3 \times 10^9$ $m^3$ compared to $3 \times 10^6$ $m^3$, respectively) and given the considerably larger population density at Lake Kivu, a catastrophic disaster could not be excluded a priori. Paleolimnological evidence for such catastrophic events in the past 5,000 years was found in sediment cores from Lake Kivu by Haberyan and Hecky (1987).

Critical for a potential gas outburst is the relative saturation of the dissolved gases, which reaches a maximum at about 270 m depth with 8% carbon dioxide and 43% methane saturation relative to the hydrostatic pressure (Deuser et al. 1973; Tietze et al. 1980). The gases remain in the deep waters of the lake because the seasonal convective mixing of the surface boundary layer extends only to about 50 m depth (Tietze et al. 1980). Below, the water is anoxic and concentrations of dissolved carbon dioxide and methane increase continuously with depth. Although the carbon dioxide, which is thought to be mainly of mantle origin, enters the lake (in dissolved form) by groundwater inflow, the methane is thought to be produced within the lake by anaerobic bacteria, which can use both acetate from decomposing organic matter and magmatic $CO_2$ as a carbon source (Deuser et al. 1973; Tietze et al. 1980; Schoell et al. 1988).

[1] To whom correspondence should be addressed. Present address: Limnological Institute, University of Constance, Mainaustrasse 252, D-78464 Konstanz, Germany (Andreas.Lorke@uni-konstanz.de).

*Acknowledgments*

We thank Clément Mudaheranwa (Kigali, Rwanda) for the logistic support and for courtesy of the digital map (Fig. 1). We are grateful to J. J. F. de Vos (Bralirwa Heinecken, Gisenyi) for his kind assistance in the field. We also acknowledge Martin Schmid's comments on an earlier draft of this manuscript. The Humanitarian Aid Office of the European Community (ECHO) funded the study. A.L. was supported by EAWAG.

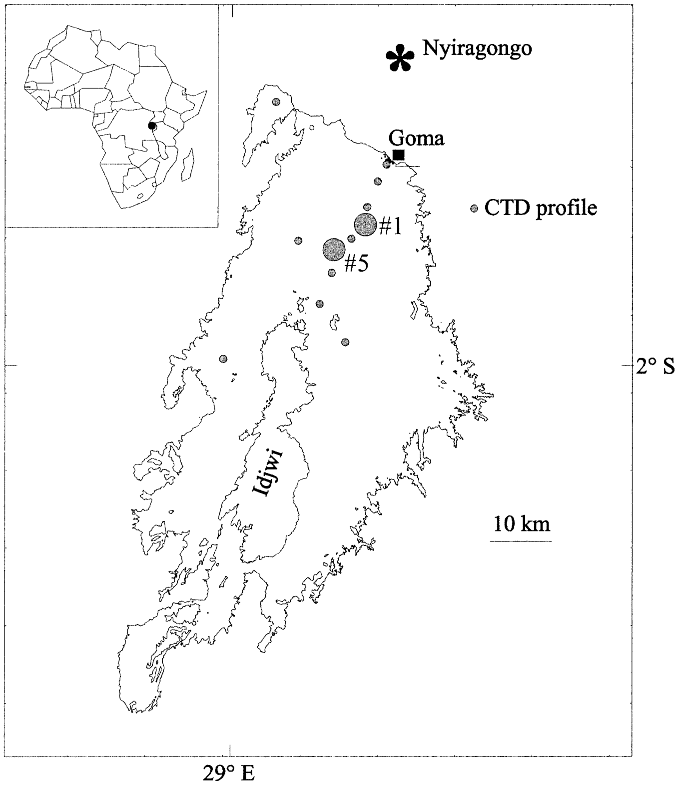


Fig. 1. Map of Lake Kivu with sampling stations and the location of the Nyiragongo volcano (18 km north of the city of Goma). The two bold and numbered CTD stations in the center of the northern basin are referenced within the text.

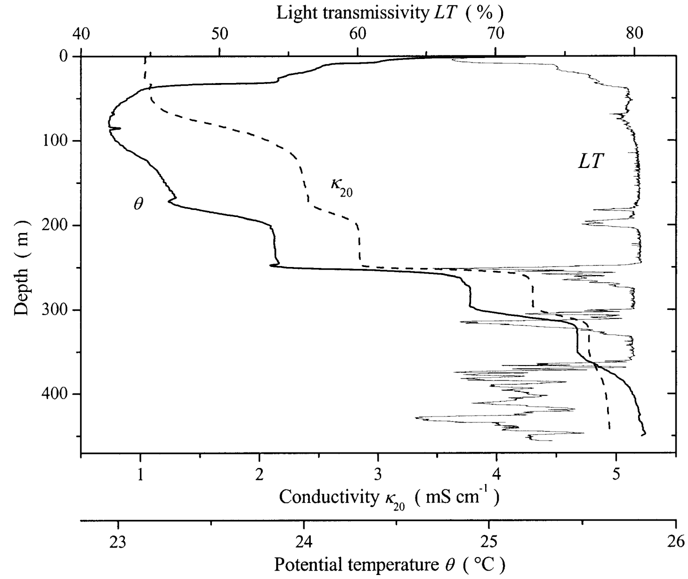


Fig. 2. Profiles of potential temperature, electrical conductivity corrected to 20°C, and light transmissivity, *LT*, from a CTD cast in the center of Lake Kivu northern basin (CTD Sta. 1 in Fig. 1).

## Measurements

A short-term investigation was carried out on 7–10 February 2002 to assess the potential of a gas outburst and to investigate the impact of the lava inflow on the density stratification of Lake Kivu. A SeaBird-19 conductivity–temperature–depth (CTD) probe was used to measure vertical profiles of electrical conductivity, temperature, and light transmissivity (at a wavelength of 660 nm along a light path of 0.1 m). Profiles were measured off the city of Goma in the vicinity of the lava inflow, as well as along a 26-km transect from Goma to the island of Idjwi in the southern part of the lake (Fig. 1).

Water samples were collected for calibration of the conductivity measurements at 10, 25, 150, and 450 m depth, using a 5-liter Niskin bottle; a remotely operated underwater camera (ROV) was used to visually inspect the lava intrusions at different locations and depths.

## Results

The general features of the CTD profiles are illustrated in Fig. 2, showing a cast from the center of the northern basin (Fig. 1). The stratification is characterized by a seasonal thermocline within the uppermost 50 m depth and a permanently stratified hypolimnion with increasing temperature stabilized by increasing salinity with depth. Below 200 m, the density stratification is further stabilized by increasing concentrations of dissolved $CO_2$ and weakly destabilized by increasing concentrations of dissolved $CH_4$ (Degens et al. 1973). The salinity was estimated to vary between 1.15 g kg$^{-1}$ at the surface and 4.47 g kg$^{-1}$ at 450 m depth, based on the ionic composition of the water samples and following the procedure described by Wüest et al. (1996). An interesting feature of the overall stratification is the step-like structure of the profiles within the permanently stratified part of the water column, showing three distinctive mixed layers (200–250 m, 275–300 m, and 320–355 m) separated by strong gradients. At least the lowest two mixed layers must have been actively mixed very recently and do not exhibit any diffusion-induced gradients of the measured parameters.

The light transmission has a constant bulk value of about 80% transmissivity (at 660 nm over 10 cm path length) and strong peaks of higher turbidity (i.e., lower light transmissivity) at each density gradient. However, based on our data it is not clear whether these turbidity peaks are related to the volcano eruption and subsequent lava inflow or whether they are a rather typical feature of Lake Kivu, since no detailed turbidity profiles were measured in the lake before. Lake Kivu has no significant river inflows (only one outflow, the Ruzizi River, which discharges into Lake Tanganyika); riverine intrusions can be excluded as alternative sources of the observed turbidity layers.

A signature of the lava inflow, however, is the thin layer of slightly increased temperature at about 80 m depth (Fig. 2). The layer thickness is approximately 5 m, and the temperature increase is less than 0.05°C. Temperature and light transmissivity profiles along the transect are shown in Fig. 3a and 3b, respectively. The temperature profiles close to the lava inflow show large fluctuations between 50 and about 120 m depth that resulted from the local heating by the lava and subsequent advective isopycnal transport. The observed

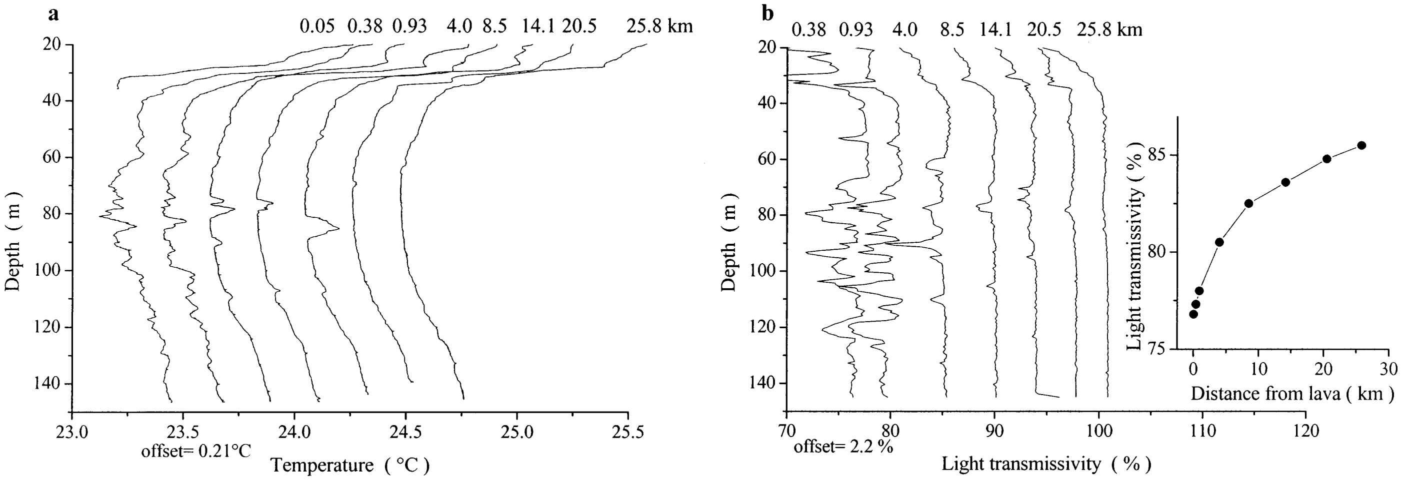


Fig. 3. Successive profiles of (a) temperature and (b) light transmissivity following a north-to-south transect from the lava inflow in Goma to the island of Idjwi (Fig. 1). Successive profiles are offset by 0.21°C and 2.2%, respectively. The numbers above the profiles indicate the distance from the lava inflow in kilometers. The inset plot in Fig. 3b shows the changes of the bulk values of the light transmissivity as a function of distance from the lava inflow. The bulk values were estimated as the mean background light transmissivity below 50 m depth (disregarding the fluctuations). All profiles were measured within 4 h, and temporal changes of the water column characteristics can hence be neglected.

temperature inversions do not cause local convection, since the strong salinity gradient within that depth range stabilizes the density stratification of the water column. Farther away from the lava inflow, the temperature fluctuations are increasingly smoothed out, and only one distinct peak of higher temperature remains. This peak is most pronounced in the profile taken 14.1 km away from the lava inflow with a temperature anomaly of ~0.14°C and disappears completely in the profiles measured farther away. This strongest peak can be related to the earliest and major inflow events. Taking into account the time between the Nyiragongo eruption and the profile measurements (23 d) and setting 14 km and 20 km as the lower and upper limits for the horizontal dispersion results in a horizontal advection velocity between 0.7 and 1 cm $s^{-1}$, respectively. This is in good agreement with tracer measurements in several lakes of comparable and smaller size (Peeters et al. 1996). Neither the main structure of the vertical stratification nor the heat content changed significantly as a result of the lava inflow.

The heat input of the $10^6$ $m^3$ of hot lava flowing into the lake can be estimated to about $2.4 \times 10^{15}$ J, assuming typical values (Dragoni et al. 2002) for temperature (1,000°C), heat capacity (850 J $kg^{-1}$ $K^{-1}$), and density (2,800 kg $m^{-3}$). The assumption that this heat was solely distributed within this thin, 5-m thick layer of water over an area spanned by a segment of a circle centered in Goma with an opening angle of 130° (following the shorelines) and a radius of 17 km results in a temperature increase of 0.3°C. This estimate is in reasonable agreement with the observed maximum temperature increase of 0.14°C, if we take into account that a significant amount of heat was probably lost to the atmosphere due to evaporation or was mixed into the rest of the water column. Since a local destabilization of a 5-m thick layer within the respective depth range would require a temperature increase of 0.5°C, even the total amount of heat introduced by the lava was not sufficient to force convective mixing of this layer over the observed spreading area.

The transmissivity profiles along the CTD transect are shown in Fig. 3b. Again, the profiles close to the lava inflow show strong fluctuations, which become smoother with increasing distance. A direct correlation between the strength of the well-defined temperature peaks and the peaks in the transmissivity could not be observed. However, in contrast to the temperature, the bulk or background transmissivity, as estimated from the profile sections, which show constant transmissivity values, increases with distance from the city of Goma (inset plot in Fig. 3b). Since this bulk transmissivity shows no vertical structure, contrary to what would be expected for settling particles, the changes can be attributed to the dilution of the particles only, with settling not being a dominant factor.

## Discussion

Several peculiarities have made Lake Kivu the subject of comprehensive investigations in the past. These include the high concentrations of dissolved gases (Tietze et al. 1980) and their origin (Deuser et al. 1973; Tietze et al. 1980; Schoell et al. 1988), the resulting density stratification (Newman 1976; Tietze 1981), the special underground geological situation (Degens et al. 1973), and the potential exploitation of the dissolved methane (Tietze 2000). In addition, double-diffusive mixing, caused by the geothermal heating at the bottom of the lake, was observed and analyzed (Newman 1976; Tietze 1981). All of those investigations were made more than 25 yr ago and provide the opportunity for a long-term comparison with the recent profiles.

A comparison of the recent temperature profile with the one measured by Tietze in 1975 (Tietze 1981) in Fig. 4 reveals a distinct warming of up to 0.5°C within the perma-

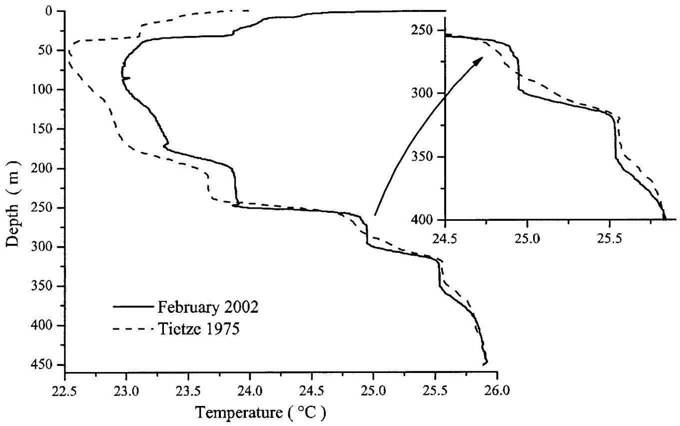


Fig. 4. Comparison of one recent (February 2002) temperature profile and one measured by Tietze in 1975 (Tietze 1981). Both profiles were measured with a vertical resolution of a few decimeters and a proven temperature calibration.

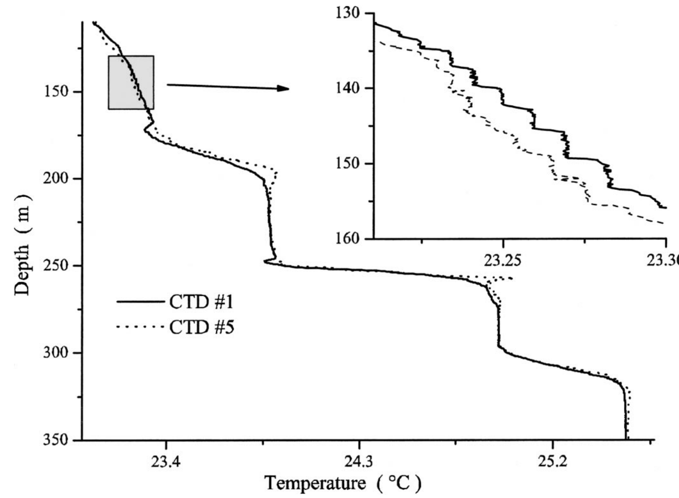


Fig. 5. Detailed structure of the temperature stratification around the mixed layers. The two profiles were measured 11 (CTD 1) and 17 km (CTD 5) off Goma (*see Fig. 1*) with 3 days time difference in between. The inset plot shows the double-diffusive staircase in CTD 1.

nently stratified part of the water column between 50 and 250 m depth. Both profiles were measured with a proven accuracy better than 0.01°C and a high vertical resolution of ~10 cm. A similar increase in temperature was already observed by Deuser et al. (1973) by comparing their profiles to earlier profiles of Damas (1937); however, their accuracy, which was stated to be ''within a few tenths °C'' in the original publication (Degens et al. 1973), is not sufficient to allow for a direct comparison. In contrast to Fig. 4, the formerly observed warming affected the entire water column, and Deuser et al. (1973) proposed the energy released by the formation of methane to be responsible. However, since the methane concentration increases significantly only below a depth of 250 m, where no recent temperature increase could be observed from Fig. 4, and the amount of warming increases toward the surface, this recent increase in temperature below the depth of seasonal mixing can most probably be attributed to climate variability or change. Comparable climatic warming was observed in other African lakes (Halfman 1993; Hecky 1993; Plisnier 2000; O'Reilly et al. 2003; Verburg et al. 2003), as well as in other lakes around the world (e.g., Scheffer et al. 2001).

Interestingly, the near-bottom temperature of the lake has not changed at all in the past 25 yr, indicating a stable equilibrium between the geothermal heat flux into the lake and the vertical transport within the water column. However, the stratification did change slightly. The most obvious change occurred between 250 and 300 m depth, where a well-mixed layer has been established in a former gradient region. The two other well-mixed layers, at 200–250 m depth and 320–350 m depth, became thicker compared to 25 yr ago. At least these two layers do not exhibit any gradients of the measured physicochemical parameters, as one would expect due to the smoothing effect of diffusion. Hence they can only be sustained by active mixing. The only possible mechanism for the local production of the necessary mechanical energy is convection, which may be driven by double diffusion or by layer merging of thin, 1- to 2-m thick double-diffusively mixed layers as shown in the inset plot in Fig. 5. The spikes in the recent profiles at the edges of those mixed layers (at 170 and 250 m depth in Fig. 5) also indicate convective activity, although they are stabilized by salinity. An unusual peculiarity is that the double-diffusive mixing in Lake Kivu is not solely characterized by the classical countergradients of temperature- and salinity-related density profiles, but also the increasing concentrations of dissolved gases must also be taken into account: whereas the carbon dioxide increases the density, the dissolved methane decreases the density with increasing gas concentration.

The density ratio $R_\rho$, an indicator for the susceptibility of the stratified water column to double diffusion, is defined as the ratio of the stabilizing forces and the destabilizing forces and can be expressed in this special case as

$$R_\rho = \frac{\beta_S \frac{\partial S}{\partial z} + \beta_{CO_2} \frac{\partial [H_2CO_3]}{\partial z}}{\alpha\left(\frac{\partial T}{\partial z} + \Gamma\right) - \beta_{CH_4}\left(\frac{\partial [CH_4]}{\partial z}\right)} \qquad (1)$$

where $S$ and $T$ denote salinity and temperature, $z$ is the depth, $\alpha$ and $\beta_S$ are the coefficients of thermal and haline expansion, $[H_2CO_3]$ and $[CH_4]$ are the dissolved gas concentrations, and $\beta_{CO_2}$ and $\beta_{CH_4}$ the respective expansion coefficients (note that $\beta_{CH_4}$ is negative), whereas $\Gamma$ denotes the adiabatic lapse rate. The ionic carbon species ($HCO_3^-$ and $CO_3^{2-}$) are included in $S$, and $[H_2CO_3]$ was estimated following Schmid et al. (2003), whereas $[CH_4]$ was taken from Tietze et al. (1980). Double-diffusive staircases have been observed in systems with $1 < R_\rho \leq 10$, but the susceptibility to double diffusion increases with decreasing $R_\rho$ until the density ratio approaches 1, where the stratification becomes unstable (Turner 1973).

Similar to that found by Schmid et al. (in press) in Lake Nyos, the density ratio is below 10 throughout the entire water column, but it falls below 3 within two depth ranges, at 130–170 m and 380–400 m depth. Double-diffusive staircases could only be observed within the upper range (*see Fig. 5*) and were best resolved in the profile shown in Fig. 2 (CTD Sta. 1 on the map in Fig. 1). This profile shows a typical double-diffusive staircase with layer heights between 2.4 and 4.3 m, increasing with depth. Applying the semi-empirical flux law from Kelley (1990) results in an average double-diffusive heat flux of 0.03 W $m^{-2}$. Although using the original flux law from Turner (1973) results in a higher heat flux of 0.05 W $m^{-2}$, these values are more than one order of magnitude smaller than those found by Newman (1976). Within the same depth range he estimated an average heat flux of 1.6 W $m^{-2}$, which could be attributed to the much smaller layer thicknesses of 0.85 m. Although in discrepancy with former measurements, the observed upward heat flux within the double-diffusive staircase is within the wide range of bottom heat flux estimates of 0.017–0.17 W $m^{-2}$ determined by Degens et al. (1973).

However, relating the bottom heat flux to the double-diffusive heat flux between 130 and 160 m depth raises questions about the heat transport within the intermediate depth range, where no small-scale staircases could be observed. The apparent diffusivity of heat $K_T$ within the double-diffusive regime can be estimated by applying Fick's law to the observed heat flux and the local temperature gradient and results in $K_T \approx 2 \times 10^{-6}$ $m^2$ $s^{-1}$, about one order of magnitude greater than the molecular diffusivity. Since it can be assumed that turbulent mixing in the interior of a strongly stratified and deep lake is comparably weak (Schmid et al. in press; Wüest and Lorke 2003), with vertical diffusivities close to the molecular ones, the enhanced vertical transport within the double-diffusive depth ranges could become transport-limited from below.

The continuing double-diffusive transport with limited heat flux at the lowest interface could lead to progressive layer merging until the entire depth range becomes mixed. This process could be responsible for the formation and maintenance of the special pattern in the Lake Kivu stratification, showing these three mixed layers separated by strong gradients (Fig. 2). In fact, the three mixed layers observed in the recent measurements correspond to the three distinct depth ranges with strong double-diffusive staircase formation in the measurements of Newman (1976). Further indication of active convective mixing across these layers is given by the observation of spikes (which are stabilized by salinity) in the temperature profiles adjacent to the layers (Fig. 5). The oppositional (bimodal) occurrence of these spikes in the two profiles shown in Fig. 5 reveals the either temporal or spatial dynamics of the fine structure of the stratification in Lake Kivu.

Beyond all the unique findings, it is important to point out that the overall stability of Lake Kivu is relatively high and that a much stronger impact than that of the January 2002 lava inflow would be necessary to trigger a gas outburst from the deep waters. Clear signatures of the lava inflow were found down to 100 m depth and as far as 14 km away from the location of the lava inflow, but the stability and the heat content of the gas-rich layers below were not affected. The observed general increase in temperature within the upper part of the permanently stratified part of the water column is related to long-term changes, which are most probably caused by climate variability. Even the double-diffusive structures, which are based on a small-scale equilibrium of molecular processes, were well preserved. Therefore, besides the acquisition of new concentration profiles of dissolved gases, future research should focus on the potential impact of lava outflows from submersed vents as well as on the potential impact of baroclinic dislocation in the water column triggered by the enhanced seismic activity at the lake bottom.

## *References*


DAMAS, H. 1937. La stratification thermique et chimique des lacs Kivu, Edourd et Ndalaga (Congo Belge). Verh. Int. Ver. Limnol. **8:** 51–68.

DEGENS, E. T., R. P. VON HERZEN, H.-K. WONG, W. G. DEUSER, AND H. W. JANNASCH. 1973. Lake Kivu: Structure, chemistry and biology of an East African Rift Lake. Geol. Rundsch. **62:** 245–277.

DEUSER, W. G., E. T. DEGENS, AND G. R. HARVEY. 1973. Methane in Lake Kivu: New data bearing its origin. Science **181:** 51–54.

DRAGONI, M., F. D'ONZA, AND A. TALLARICO. 2002. Temperature distribution inside and around a lava tube. J. Volcanol. Geotherm. Res. **115:** 43–51.

FREETH, S. J., G. W. KLING, M. KUSAKABE, J. MALEY, F. M. TCHOUA, AND K. TIETZE. 1990. Conclusions from Lake Nyos disaster. Nature **348:** 201.

HABERYAN, K. A., AND R. E. HECKY. 1987. The late Pleistocene and Holocene stratigraphy and paleolimnology of Lakes Kivu and Tanganyika. Paleogeogr. Paleoclimatol. Paleoecol. **61:** 169–197.

HALFMAN, J. D. 1993. Water column characteristics from modern CTD data, Lake Malawi, Africa. J. Gt. Lakes Res. **19:** 512–520.

HECKY, R. E. 1993. The eutrophication of Lake Victoria. Verh. Int. Ver. Limnol. **25:** 39–48.

KELLEY, D. E. 1990. Fluxes through diffusive staircases: A new formulation. J. Geophys. Res. **95:** 3365–3371.

KLING, G. W., AND OTHERS. 1987. The 1986 Lake Nyos gas disaster in Cameroon, West Africa. Science **236:** 169–170.

NEWMAN, F. C. 1976. Temperature steps in Lake Kivu: A bottom heated saline lake. J. Phys. Oceanogr. **6:** 157–163.

O'REILLY, C. M., S. R. ALIN, P.-D. PLISNIER, A. S. COHEN, AND B. A. MCKEE. 2003. Climate change decreases aquatic ecosystem productivity of Lake Tanganyika, Africa. Nature **424:** 766–768.

PEETERS, F., A. WÜEST, G. PIEPKE, AND D. M. IMBODEN. 1996. Horizontal mixing in lakes. J. Geophys. Res. **101:** 18361–18375.

PLISNIER, P.-D. 2000. Recent climate and limnology changes in Lake Tanganyika. Verh. Int. Ver. Limnol. **27:** 2670–2673.

SCHEFFER, M., D. STRAILE, E. H. NES, AND H. HOSPER. 2001. Climatic warming causes regime shifts in lake food webs. Limnol. Oceanogr. **46:** 1780–1783.

SCHMID, M., A. LORKE, C. DINKEL, A. WÜEST, AND G. TANYILEKE. In press. Double-diffusive convection in Lake Nyos, Cameroon. Deep-Sea Res. I.

———, ———, A. WÜEST, M. HALBWACHS, AND G. Z. TANYILEKE. 2003. Development and sensitivity analysis of a model for assessing stratification and safety of Lake Nyos during artificial degassing. Ocean Dynamics **53:** 288–301.

SCHOELL, M., K. TIETZE, AND S. M. SCHOBERTH. 1988. Origin of methane in Lake Kivu (East-Central Africa). Chem. Geol. **71:** 257–265.

TIETZE, K. 1981. Direct measurements of the in-situ density of lake and sea water with a new underwater probe system. Geophysica **17:** 33–45.

———. 1992. Cyclic gas bursts: Are they a ''usual'' feature of Lake Nyos and other gas-bearing Lakes? p. 97–107. *In* S. J. Freeth, K. M. Onuoha, and C. O. Ofoegbu [eds.], Natural hazards in West and Central Africa. Vieweg-Verlag.

———. 2000. Lake Kivu gas development and promotion-related issues: Safe and environmentally sound exploitation, Report PDT GmbH No. 520002, Republic of Rwanda, Ministry of Energy, Water and Natural Resources, Kigali (Rwanda).

———, M. Geyh, H. Müller, L. Schröder, W. Stahl, and H. Wehner. 1980. The genesis of methane in lake Kivu (Central Africa). Geol. Rundsch. **69:** 452–472.

Turner, J. S. 1973. Buoyancy effects in fluids. *In* G. K. Batchelor and J. W. Miles [eds.], Cambridge monographs on mechanics and applied mathematics. Cambridge Univ. Press.

Verburg, P., R. E. Hecky, and H. Kling. 2003. Ecological consequences of a century of warming in Lake Tanganyika. Science **301:** 505–507.

Wüest, A., and A. Lorke. 2003. Small-scale hydrodynamics in lakes. Annu. Rev. Fluid Mech. **35:** 373–412.

———, G. Piepke, and J. D. Halfman. 1996. Combined effects of dissolved solids and temperature on the density stratification of Lake Malawi, p. 183–202. *In* T. C. Johnson and E. O. Odada, [eds.], The limnology, climatology and paleoclimatology of the East African Lakes. Gordon and Breach.

*Received: 5 March 2003*
*Amended: 10 December 2003*
*Accepted: 11 December 2003*