Simulation of CO₂ concentrations, temperature, and stratification in Lake Nyos for different degassing scenarios

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[1] A large gas cloud erupted unexpectedly in 1986 from Lake Nyos, the larger of the two Cameroonian “Killer Lakes,” with devastating consequences. Regular monitoring subsequently revealed that the deep water of the lake was gradually recharged with CO₂. To preclude a similar event in the future, a degassing pipe was installed in the lake in 2001. In the present study a one-dimensional model is used to predict the effects of this pipe and other degassing options on the CO₂ concentrations and the stratification within the lake for the next 50 years. The results of the simulations show that without degassing, total CO₂ content would reach the preeruption value within a few decades. The presently installed pipe is sufficient to reduce CO₂ pressures in the entire water column above the pipe inlet to <5 bar within 10 years, and a steady state is reached within 50 years. Depending on the assessment of the risk due to the gas currently remaining in the lake and the costs involved, the installation of additional pipes could be considered (1) to remove the gas more quickly and (2) as a backup for long-term failures and maintenance. Once the steady state is reached, degassing with one pipe is a practicable long-term solution which can also be used for monitoring the approximate deep water CO₂ concentrations by nonprofessionals. Assuming a doubled deep water input in the future as an upper limit of the expected source strength, the pipe is still able to prevent a CO₂ accumulation. As a side effect, the degassing operation strongly changes the stratification in the lake. It transforms the lake from a meromictic to an oligomictic system and gradually removes the dissolved salts from the lake.

Components: 8417 words, 11 figures.

Keywords: carbon dioxide; degassing; gas eruption; k-epsilon; Lake Nyos.

Index Terms: 0458 Biogeosciences: Limnology (1845, 4239, 4942); 8488 Volcanology: Volcanic hazards and risks; 4255 Oceanography: General: Numerical modeling (0545, 0560).

Received 13 October 2005; **Revised** 23 March 2006; **Accepted** 10 April 2006; **Published** 28 June 2006.

Schmid, M., M. Halbwachs, and A. Wüest (2006), Simulation of CO₂ concentrations, temperature, and stratification in Lake Nyos for different degassing scenarios, *Geochem. Geophys. Geosyst.*, 7, Q06019, doi:10.1029/2005GC001164.

1. Introduction

[2] Meromictic lakes in volcanic areas set up the scene for a rare natural hazard: the limnic gas eruption. Gasses which are introduced into the deep permanently stratified part of these lakes can accumulate for decades or even centuries. Once the deep layer is sufficiently charged, a local supersaturation can be caused by a relatively small vertical displacement. A bubble plume can be formed which can gain additional momentum by entraining more gas-rich water and lead to the sudden release of a large part of the gasses stored in the lake [Tietze, 1992; Evans *et al.*, 1994]. This natural hazard had not been known before the catastrophic eruptions of carbon dioxide (CO₂) clouds from Lake Monoun in 1984 [Sigurdsson *et al.*, 1987] and Lake Nyos in 1986 [Freeth and Kay, 1987; Kling *et al.*, 1987; Sigvaldason, 1989], two of several crater lakes in the north-western part of Cameroon [Kling, 1988]. Except for some legends about maleficent lakes [Shanklin, 1989] there had been no indication that a similar disaster had happened in this region before.

[3] The only other evidence for previous limnic gas eruptions has been found in the sediment stratigraphy of Lake Kivu [Haberyan and Hecky, 1987]. In this large East African Rift lake the major part of the gas pressure is due to biogenic methane (CH₄), which seems to have increased in the past decades as a result of changes in the nutrient dynamics of the lake [Schmid *et al.*, 2005]. The probability of a gas release from Lake Kivu is currently small, but its consequences would be devastating in this densely populated region. Recently, the hypothesis has been put forward that a large-scale oceanic gas eruption could have been responsible for the mass-extinction at the Permian-Triassic boundary [Ryskin, 2003].

[4] In recent years degassing pipes have been installed both in Lakes Nyos and Monoun to prevent new gas eruptions from these lakes [Halbwachs *et al.*, 2004]. Water is withdrawn from the deep layer with a pipe and discharged as a fountain on the lake surface. Due to the high concentrations of dissolved CO₂, the water flow through the pipe is self-sustaining, driven by the expansion of the bubbles generated in the rising water.

[5] The aim of the present study is to quantify the effects of the degassing on the future development of stratification and CO₂ concentrations in Lake Nyos. The one-dimensional model previously developed for this purpose [Schmid *et al.*, 2003] is

calibrated with recent field data and then used to predict the evolution of the lake's stratification under different degassing scenarios.

[6] Previous simulation approaches for Lake Nyos include the application of a turbulence model to explain the evolution of temperature and stratification after the eruption [Kantha and Freeth, 1996], and the simulation of different degassing scenarios both with the simulation software DYRESM [McCord and Schladow, 1998; Kusakabe *et al.*, 2000], and using a mass balance approach [Kling *et al.*, 2005]. Compared to these previous studies the model presented here includes the following additional features: the algorithm for calculating vertical mixing takes account of the energy transfer via internal seiches; differential vertical transport of heat and salt and consequently local intensification of mixing due to double-diffusive convection are enabled; the flow rate through the degassing pipe is simulated with a hydraulic model as a function of the H₂CO₃ concentration at the inlet; the meteorological forcing contains daily and interannual variability, which allows simulating the large interannual variability of seasonal convective mixing intensity; the geochemistry of the lake is represented in more detail, including the formation of particulate iron at the lake surface and its effects on light attenuation.

2. Study Site

[7] Lake Nyos is situated 1091 m a.s.l. and has a surface area of 1.58 km², a volume of 0.15 km³ and a maximal depth of 208 m [Nojiri *et al.*, 1993]. Since the gas eruption in August 1986 which cost more than 1700 lives, the lake has been regularly monitored, and a continuous accumulation of dissolved CO₂ has been observed [Evans *et al.*, 1994; Kusakabe *et al.*, 2000]. The water column of the lake can be divided in three distinctly different sections which are separated from each other by two chemoclines (Figure 1). The top 55 m, the *mixolimnion*, are convectively mixed annually during the dry season. The exchange with the atmosphere and the dilution by the rainwater lead to low CO₂ concentrations and conductivities in this section. In the *hypolimnion* between 55 and 180 m depth, temperatures and CO₂ concentrations have remained almost unchanged since the eruption in 1986. The nearly linear increase of temperature and conductivity with depth, observed after the eruption, is still prominent, except for the effects caused by double-diffusive convection below the *mixolimnion* [Schmid *et al.*, 2004]. The *deep layer* from

Figure 1. Vertical profiles of temperature and conductivity (normalized to 25°C) measured in Lake Nyos on 17 March 2004 and definition of the different depth zones of the lake.

185 m to 208 m depth is fed by a deep water source of CO₂-rich saline water. The flow rate of this source has been quantified to 18 L s⁻¹ of water with a temperature of 26°C and a total CO₂ concentration of 420 mmol L⁻¹ [Schmid *et al.*, 2003].

3. Model Description

[8] In this study we used an improved version of the model described in detail by Schmid *et al.* [2003]. Here, only a brief summary is given. The modifications made during the calibration process are described in section 4.

[9] The model was developed using the lake module of AQUASIM 2.1 [Reichert, 1994, 1998], a specialized software for the simulation of technical and natural aquatic systems. A one-dimensional vertical grid with a resolution of 0.5 m was used. The horizontal variability of conductivity and temperature in the lake is extremely small [Schmid and Wüest, 2005], justifying the use of a one-dimensional model as has previously been discussed by McCord and Schladow [1998]. The integration algorithm automatically adjusts the time step depending on stability criteria for the numerical solution. A maximal internal time step of 1000 s was allowed.

[10] The model simulates the vertical transport of temperature, dissolved substances (carbonate species, alkali metals, oxygen and iron) and particulate iron. The vertical turbulent mixing coefficient, K_z ,

is calculated with a standard k - ϵ -model [Rodi, 1980], i.e., a model which solves the budget and transport equations for turbulent kinetic energy (k) and its dissipation (ϵ). In addition to the standard k - ϵ -model, energy transfer from the wind via internal seiches to the water column [Goudsmit *et al.*, 2002] is incorporated in the present model version.

[11] The water input at the surface is estimated as the difference between precipitation and calculated evaporation in the lake catchment of 7.2 km². The program does not allow changing the water surface level, i.e., the outflow is adapted to compensate for all inflows. In reality the lake level temporarily drops by ~1 m during the dry season.

[12] At the lake surface, CO₂ and O₂ are exchanged with the atmosphere. The carbonate chemistry and the pH are simulated using chemical equilibria. Oxygen is consumed at the sediment water interface. Dissolved reduced iron is oxidized to particulate iron oxides in the oxic part of the lake (especially when it is transported to the lake surface with the degassing system). The settling iron particles are then redissolved in the anoxic hypolimnion. A deep water source introduces warm water with elevated concentrations of CO₂, alkali metals and iron into the deepest part of the lake.

[13] The degassing system installed in the lake was simulated by removing water from the intake depth of the pipe (203 m) and releasing it with the same properties at the lake surface except that the CO₂ concentration is equilibrated with the atmosphere and the temperature is reduced by the endothermic process of CO₂ exsolution. The water flow through the polyethylene pipe with an inner diameter of 145 mm was calculated as a function of intake depth and H₂CO₃ concentration, with a hydraulic model that includes buoyancy due to progressive degassing of the rising water and friction inside the pipe (G. Kayser, personal communication, 2003). With an inlet depth of 203 m and a H₂CO₃ concentration of about 360 mmol L⁻¹, the resulting speeds are 3.7 m s⁻¹ at the inlet and about 30 m s⁻¹ at the surface outlet. At 100 m depth a lateral valve can be opened to reduce the flow rate. The degassing system started working in March 2001. Until September 2004, this lateral valve was open for a total of 18 months and the system was completely shut down for maintenance during about 4–5 months. Since then, the system has been continuously working with the lateral valve closed. The respective time schedule was implemented in the model.

[14] Compared to the model version described by Schmid *et al.* [2003], the source of the meteorological data to drive the model was harmonized: the average daily values of air temperature, wind speed, cloud cover, specific humidity, and precipitation for the years 1986 to 2004 were taken from the two nearest grid points ($6^{\circ}40'N$; $9^{\circ}23'E$ and $11^{\circ}15'E$) of the NCEP-NCAR reanalysis project [Kalnay *et al.*, 1996]. Air temperature was corrected by $-4.8^{\circ}C$, since the reanalysis value is valid for an elevation of about 300 m while the lake is located 1091 m a.s.l. Average values of the meteorological data for the entire time series are $22.1^{\circ}C$ air temperature, 1.4 m s^{-1} wind speed, 1787 mm annual precipitation, 60% cloud cover, and 21.8 mbar vapor pressure. Of course the NCEP-NCAR reanalysis data cannot be expected to represent the meteorological conditions at the lake for any given day, but they reproduce the average regional daily meteorological conditions and the typical interannual variability, which is more important for long-term simulations like those presented in this study. Systematic errors, which can be caused, e.g., by the special local wind conditions at the lake, can be largely compensated by calibrating the model parameters of the heat exchange at the lake surface and the turbulent vertical mixing (see section 4). Test simulations with the 6-hourly NCEP-NCAR reanalysis data needed much more calculation time while leading to only slightly different results. The mixing depth was somewhat shallower due to the stronger stratification during the daytime which more than compensated the stronger nighttime mixing. However, this effect can also be compensated by the model calibration and it was therefore decided to use the average daily data. Irradiation I at the lake surface was not estimated from the extraterrestrial radiation as done by Schmid *et al.* [2003], but from the clear sky radiation I_0 calculated for the given elevation with the method of Brock [1981], and corrected for cloudiness C using

$$I = I_0(c_1 + (1 - c_1)(1 - C))$$

with $c_1 = 0.43$, which is derived from the altitude-corrected value for Accra (Ghana) [Brutsaert, 1982].

4. Model Calibration

[15] The following data, measured in the lake, were used to calibrate the model: Water surface temperatures were recorded every 48 min from 5 November 2001 to 1 November 2002 and every 15 min

from 8 December 2002 to 5 March 2004, using a Vemco minilog temperature recorder which had been installed beneath a surface buoy. Vertical profiles of temperature, conductivity and pH were measured with a Sea-Bird SBE-19 on 3 November 2001, 8 December 2002 and 9 March 2004. Dissolved CO_2 concentrations were calculated from conductivity and pH as described by Schmid *et al.* [2004]. Vertical profiles of temperature and CO_2 concentrations for November 1993, March 1995 and April 1996 were obtained from Kusakabe *et al.* [2000].

[16] The parameters adjusted were (1) p_1 , which scales the incoming infrared radiation from the atmosphere, (2) p_2 , which scales the latent and convective heat fluxes, (3) α , which defines the energy transfer from the wind to the internal seiches [Goudsmit *et al.*, 2002], and (4) λ_0 , the light attenuation coefficient of the Lake Nyos surface waters without the influence of iron particles. The calibration simulation was started on 8 November 1986 with the initial conditions described by Schmid *et al.* [2003] and run until 30 June 2004. The degassing operation was included in the calibration run as described above.

[17] The calibration procedure resulted in the following values: $p_1 = 0.962$, $p_2 = 0.736$, $\alpha = 0.005$, and $\lambda_0 = 1.0 \text{ m}^{-1}$. The average simulated infrared radiation from the atmosphere H_A is around 350 W m^{-2} . The calibrated value of p_1 corrects H_A by -14 W m^{-2} , which is certainly within the error limits of the parameterization of this heat flux as has been demonstrated by a comparison of several empirical formulas [Henderson-Sellers, 1986]. The calibrated value of p_2 corrects the average latent (H_L) and convective (H_C) heat fluxes from -82 W m^{-2} to -60 W m^{-2} and from -33 to -24 W m^{-2} , respectively. This correction, though relatively large, is also well within typical uncertainties of these heat fluxes [Henderson-Sellers, 1986]. The effects of the two parameters partially cancel each other, and the average total heat flux correction is only 17 W m^{-2} . The main effect of the heat flux correction is to reduce the heat loss and consequently weaken convective mixing during the dry season between November and March, when the specific humidity is low and the latent heat fluxes from the lake reach their maxima. The need for correction of the heat fluxes is not necessarily due to errors in the formulae used to calculate the fluxes. Similar effects would also be caused by systematic errors in the meteorological forcing. The parameter α had to be increased from

Figure 2. Observed and simulated daily mean lake surface temperatures.

its original value of 0.0017 to 0.005 to avoid an underestimation of the turbulent mixing within the lake. This could also reflect a systematic underestimation of the local wind speed. A short time series of wind speeds observed at the lake indicated a higher average of 2.4 m s^{-1} [McCord and Schladow, 1998], compared to 1.4 m s^{-1} in the NCEP-NCAR reanalysis data. Finally, the light absorption λ_0 had to be increased from its original value of 0.4 m^{-1} , otherwise either the simulated surface temperatures were too low, or too much thermal energy was introduced into deeper layers.

The corrected total simulated light absorption coefficient for November 2001 was about 1.2 m^{-1} which agrees well with the Secchi depths of 1.5–2.2 m observed during this time (data provided by G. Tanyileke (2003)).

[18] Figure 2 compares the simulated and the measured daily mean lake surface temperatures. The simulated values lie on average 0.13°C above the measurements, mainly because the simulated cooling was slightly retarded in December 2001 and 2003. The standard deviation of the difference between the two is 0.58°C . The maximum temperatures during the dry season seem to be typically underestimated. This is probably caused by an underestimation of the stratification and an overestimation of mixing in the top few meters of the water column during strong irradiation due to the disregard of the daily variability in solar irradiation. This can also be seen in Figure 3, where the simulations did not reproduce the strong stratification observed in the top few meters. The model perfectly reproduced the observed seasonal dynamics with minimum temperatures of around 22°C at the end of January, a large maximum in April and May, a slight depression in August and a second small maximum in October/November. The minimal temperatures in January were very well reproduced, which is important because they largely determine the depth of mixing. Contrary to dimictic or monomictic lakes in temperate zones, where errors in the simulation during one year are largely

Figure 3. Vertical profiles of temperature, initial condition (measured in November 1986), and comparison between observations and simulations in 1996 and 2004 (data from 1986 and 1996 from Kusakabe et al. [2000]).

Figure 4. Vertical profiles of dissolved CO_2 , initial condition (measured in November 1986), and comparison between observations and simulations in 1996 and 2004 (data from 1986 and 1996 from *Kusakabe et al.* [2000]).

canceled out during the period of deep convection in winter, in the case of meromictic Lake Nyos, errors in the simulation of the mixing depth can propagate from one year to the next.

[19] It is obvious from the observed profiles that the temporal changes in the vertical structure of temperature (Figure 3) just below the thermocline between 50 and 100 m depth and just above the chemocline between 150 and 180 m depth were much larger than for CO_2 (Figure 4). This fact cannot be explained by turbulent mixing alone, because in that case the diffusion of temperature and CO_2 would be similar. The differential transport is due to double-diffusive convection, a phenomenon rarely observed in lakes. Double-diffusive convection can occur, when the effects of temperature and dissolved substances on density are opposed and when turbulent mixing is weak enough that the difference between the molecular diffusivities of heat ($1.44 \cdot 10^{-7} \text{ m}^2 \text{ s}^{-1}$) and dissolved substances ($\sim 1.0 \cdot 10^{-9} \text{ m}^2 \text{ s}^{-1}$) is not negligible for the vertical transport [*Kelley et al.*, 2003; *Turner*, 1965]. A detailed investigation in Lake Nyos in December 2002 showed a spectacular set of 26 well-mixed layers with thicknesses of 0.2–2.1 m and sharp interfaces (0.05–0.35 m) in between below the thermocline [*Schmid et al.*, 2004]. The formation of these layers had been triggered in February 2002 by a particularly strong

cooling event during that dry season. The vertical extent of the double-diffusive zone and the average layer thickness subsequently grew until they disappeared again around August 2004. From the measured temperature and CO_2 profiles [*Kusakabe et al.*, 2000], it is obvious that similar double-diffusive events must have occurred below the thermocline between 1988 and 1993 and probably also above the chemocline between 1988 and 1993 as well as between 1996 and 2001. This phenomenon could not be simulated with the model version described by *Schmid et al.* [2003], since molecular diffusivities were not included. In the meantime, the model code has been changed such that the molecular diffusivities of temperature and dissolved substances are added to the turbulent diffusivities. This new model version simulated mixed layers in both zones where double-diffusive convection occurred in reality (Figures 3 and 4). Of course the vertical resolution of 0.5 m of the model is by far not sufficient to simulate the real small-scale convective layers with interfaces of only a few cm thickness. Consequently, the thickness of the convectively mixed layers is strongly overestimated. Nevertheless, the simulated double-diffusion reproduced the differential transport between temperature and dissolved CO_2 , as can be seen from the good agreement between the

simulated and the observed large-scale gradients at these depths.

[20] In November 1986, three months after the eruption, the mixolimnion was only 8 m deep. The upper chemocline was then occasionally eroded during strong convective mixing events, finally leading to a deepening of the mixolimnion to 55 m depth in 2004. This behavior was very well reproduced by the model. The mixing depth accumulated during 18 years of simulation was overestimated by about 4–5 m as can be seen in the comparison of the CO₂ profiles from March 2004 (Figure 4). Since errors in the calculation of the mixing depth in one year are propagating, this difference can be due to an overestimation of mixing in just one cold dry season. The model also typically overestimates daily mixing depths within the mixolimnion which is the cause for differences between the observed and the simulated temperature profiles within the mixolimnion (Figure 3). Just above the bottom, mixing is overestimated by the model, while it is slightly underestimated in the region between 180 and 190 m depth. Due to constraints in the algorithms of AQUASIM, the production of turbulent kinetic energy by internal seiches could only be distributed depending on the ratio of sediment area to the water volume at a given depth. In reality however, a larger fraction of this energy is introduced and dissipated in layers where the stratification is strong, i.e., in zones of high density gradients [Goudsmit *et al.*, 2002]. We conclude that our model tends to distribute too much mixing energy to the bottom layers, where the ratio of sediment area to the water volume is large, and too little energy to the chemocline between 180 and 190 m depth. The average simulated energy dissipation from the seiches is $2.4 \cdot 10^{-5} \text{ W m}^{-2}$, which is about twice as large as previously estimated from measured vertical displacements of internal waves [Schmid *et al.*, 2004], and is only sufficient to sustain a turbulent transport on the order of the molecular diffusion coefficient of heat ($\sim 1.4 \cdot 10^{-7} \text{ m}^2 \text{ s}^{-1}$). However, as will become clear from the results presented below, all these differences between observed and simulated mixing intensities are small compared to the effects of the degassing system installed in the lake.

[21] As previously reported [Schmid *et al.*, 2003], the observed temporal development of temperature and CO₂ profiles between 1986 and 2004 could be best reproduced with a deep water source of 18 L s^{-1} with concentrations of total CO₂ (including bicarbonate and carbonate) of 420 mmol L^{-1}

(i.e., 395 mmol L^{-1} of dissolved H₂CO₃), and a temperature of 26°C. The total inputs of CO₂ and heat are well constrained from budget calculations. The error for the CO₂ input is maximally on the order of $\pm 50\%$, where the largest error source is the calculation of the CO₂ concentrations from the measured conductivities and pH, which is typically on the order of 10% for each profile. The simulated water flow of 18 L s^{-1} agrees very well with the 15.5 L s^{-1} estimated from budgeting the water flow through the degassing pipe and the observed subsidence of the water column [Halbwachs *et al.*, 2004]. Of course we cannot be sure that the CO₂ input remains constant. Observations indicate a higher input during the first two years after the 1986 eruption, then an approximately constant CO₂ input between 1988 and 1998 and possibly again an elevated input in the following years [Kusakabe *et al.*, 2000; Kling *et al.*, 2005].

5. Degassing Scenario Analysis

[22] The model described above was used to simulate the effects of different degassing scenarios (no degassing, one degassing pipe, two degassing pipes, different inlet depths, hypolimnetic withdrawal) on the stratification and the CO₂ concentrations in the lake. Simulations were started on 8 December 2002 with the profiles of temperature, dissolved CO₂ and salinity measured at this date, and run for 50 years. The sensitivity of the results to the meteorological forcing and to other relevant parameters was tested with additional simulations.

5.1. Sensitivity to Meteorological Forcing

[23] Since the sensitivity analysis of the model showed that the results are most sensitive to the heat fluxes at the lake surface [Schmid *et al.*, 2003], three meteorological scenarios covering the range of meteorological forcing in the past years were simulated. In the standard weather scenario, the meteorological data set from the years 1986 to 2003 was continuously repeated. In the “cold” scenario, only the data from the years 2001 and 2002 were used, when the minimal surface temperatures had been low and consequently seasonal mixing of the mixolimnion had been unusually deep. In the “warm” scenario, the meteorological data from 1988 was repeated, when simulated mixing had been very weak. The main difference between the meteorological forcing in the three scenarios is the water vapor pressure during the dry season (Figure 5). During the dry seasons 2001 and 2002, low water vapor pressures caused high

Figure 5. Daily range of vapor pressure and air temperature (average $\pm$ one standard deviation) for the entire meteorological data set used in this study compared to values used in the “cold” and “warm” scenarios.

evaporative heat losses. In 1988, on the other hand, the vapor pressure during the dry season did not sink far below the values usually observed during the wet. Besides that, in the cold scenario the average air temperatures during the dry season are lower than in the warm scenario. During the wet season, the interannual variability of water vapor pressure and air temperature is marginal and there is no significant difference between the two scenarios. All the scenarios described below were simulated with the three climate scenarios, but since the climate effects were not as large as to change any of the conclusions, the results for the climate scenarios are shown only for the cases without degassing and with one degassing pipe.

5.2. Scenarios Without Degassing

[24] Without degassing, the lower part of the lake is slowly filled up with the water from the deep water source, as it has been observed between 1986 and 2001. The chemocline slowly rises and the partial pressure of CO_2 just below the chemocline advances toward the hydrostatic pressure (Figure 6). However, saturation is not reached at any depth within the simulated time frame. The value nearest to saturation at the end of the simulation in 2052 would be 10.2 bar at 134 m depth. Since the lake bathymetry is very steep at these depths, the nearly linear trend of the simulated chemocline ascent can be extrapolated to estimate that saturation would be reached around the year 2080–2090. This means that the natural recharge time of the lake after a gas eruption is approximately one century. However, the energy input needed to lift water from below the chemocline to its saturation level strongly decreases when the chemocline rises. In 2052, a

vertical displacement of 30–40 m is needed to lift a water parcel to a depth where it would be gas saturated. An internal wave with this amplitude leading to local supersaturation could possibly be caused by a rockfall. The CO_2 content before the eruption was roughly estimated to $17.5 \cdot 10^9$ Mol [Evans *et al.*, 1994]. This CO_2 content is reached again in 2035, indicating a significant risk for an eruption already after 30 years (Figure 7).

[25] The influence of different meteorological scenarios on this simulation is relatively small. In the cold scenario, seasonal mixing removes the CO_2 down to 72 m, while in the warm scenario, mixing is weaker and reaches only 59 m. The difference between the cold scenario and the standard scenario is almost negligible, because the cold years have a dominating effect on the long-term thermocline deepening in the standard scenario. Simulated double diffusive convection is more intense just below the mixolimnion in the cold and standard scenarios, because at this depth the forcing for double-diffusion is regularly renewed by deep seasonal mixing. On the other hand, double-diffusion is similar in all scenarios just above the lower chemocline, because there the forcing is mainly due to the gradients caused by the rising CO_2 -rich and warm water mass.

5.3. One Degassing Pipe (Standard Scenario)

[26] The currently installed pipe removes about 60 L s^{-1} of water from 203 m depth. The pipe compensates for the estimated deep water input of 18 L s^{-1} and additionally removes 42 L s^{-1} of water from above, which leads to a slow subsi-

Figure 6. Simulated CO₂ gas pressures and temperatures without degassing system in (left) the standard, (middle) the warm, and (right) the cold scenario.

dence of the water masses above (Figure 8). The annual subsidence of the water column due to the degassing operation is 1.2 m yr⁻¹ at the lake surface and reaches 4 m yr⁻¹ at the current depth of the lower chemocline (185 m) if the pipe is continuously running. As shown by Schmid *et al.* [2003], this subsidence rate agrees very well with observed temperature profiles [Halbwachs *et al.*, 2004]. The occasionally mixed surface layer is entrained downward with the upper chemocline and reaches about 170 m depth after 50 years of degassing (see also section 6). CO₂ is efficiently removed from the lake, and around the year 2011, the CO₂ pressure at 200 m depth will be reduced from 10 bars at the beginning to less than 5 bars. The temporal development of the CO₂ concentrations in the lake is completely dominated by the effects of the degassing, and the influence of the different meteorological scenarios is almost negligible. The main effect in the simulated temperatures is that the mixolimnion is about 2.5°C warmer and much less intensely mixed in the warm than in the cold scenario, leading to the development of a new permanent stratification above the gas-rich zone (see also section 6).

[27] In theory, the degassing will run infinitely, since an equilibrium between CO₂ input and output

is reached at a flow of around 45 L s⁻¹ once the mixolimnion depth reaches the pipe inlet (Figure 9). In that case, the 18 L s⁻¹ from the deep source would be mixed with 27 L s⁻¹ of almost CO₂-free

Figure 7. Temporal development of total CO₂ content in the lake in different scenarios, compared to the amount of CO₂ contained in the lake before (estimated by Evans *et al.* [1994]) and after the eruption in 1986: (a) without degassing, with (b) one and (c) two degassing pipes, (d) with the inlet of the degassing pipe lifted to 180 m depth, and (e) with a doubled deep water source strength.

Figure 8. Simulated CO₂ gas pressures and temperatures with the currently installed degassing system in (left) the standard, (middle) the warm, and (right) the cold scenario.

Figure 9. Volume water flow [L s⁻¹] through the degassing pipe as a function of depth and H₂CO₃ concentration at the inlet. The shaded area at the top right marks conditions where the water at the inlet would be supersaturated with CO₂. The arrow shows the time development of the volume flow during the simulation with one degassing pipe from 2002 to 2052. The circle and the square mark the equilibrium flow through one single and each of two pipes, respectively.

“surface” water. Consequently, the lake water volume of 0.15 km^3 would be circulated through the pipe within a timescale of 150–200 years. In practice, the water flow through the pipe could stop running for several reasons: if an unusual mixing event strongly reduces the CO_2 concentration at the inlet depth, if a strong internal wave brings CO_2 -poor water, which is available only a short distance above the inlet, or when the raft-mounted pipe is lifted fast enough from its low level at the beginning of the rainy season.

[28] Besides the meteorological forcing, additional simulations were performed with the standard scenario to test the robustness of the degassing to the boundary conditions. A scenario with a doubled flow rate of the deep water source and consequently a doubled CO_2 input was used to estimate whether the degassing pipe would be able to cope with this situation. This corresponds approximately to the elevated CO_2 input observed during the first two years after the eruption [Kusakabe *et al.*, 2000], and it cannot be excluded that similar input rates will occur again in the future. The time needed for degassing the lake is about 50% longer in this scenario (Figure 7), but the presently installed pipe would still be able to largely remove the CO_2 and subsequently keep the total CO_2 amount at a safe level.

[29] A simulation with the source distributed to the lowest 25 m instead of just the lowest meter of the water column did not show any significant changes in the degassing efficiency. The total CO_2 content would be very similar to the results of standard simulation in Figure 7. The main difference in this scenario is that after a few years the simulated CO_2 gradient near the pipe inlet is much weaker. Consequently, the degassing operation is more stable because it would need a much larger internal wave to bring water to the inlet with a concentration low enough to stop the self-siphoning. On the basis of the observed vertical CO_2 profiles it can be excluded that a significant fraction of the CO_2 enters the lake above 185 m depth.

[30] A simulation without the molecular diffusivities of heat and salt, i.e., without the effects of double-diffusive convection was performed to check whether the overestimation of the convective layers thickness could have a significant effect on the results. However, the effect on the simulated CO_2 content in the lake was negligible. The main effects of double-diffusive convection are the formation of convectively mixed layers and a lower heat accumulation in the deep water due to en-

hanced vertical heat fluxes. This phenomenon is clearly visible in a simulation without degassing, but in all the degassing scenarios it is largely overridden by the effects of the degassing. Consequently, the overestimation of convective mixed layer thicknesses shown in the model calibration does not influence the conclusions of the present study. The mixed layers are the cause for the step-like structures visible in Figures 6, 8, and 10 which are produced when the temperature or CO_2 pressure within a mixed layer shifts across the limit between two different colors.

5.4. Two Degassing Pipes

[31] If a second identical pipe is installed in the lake, the degassing runs almost twice as fast, and the gas pressure at 200 m depth is reduced to 5 bars already after 4 years of operation (Figure 10). Simulations could only be run for 30 years, since the short circuit due to the mixolimnion approaching the pipe inlet caused numerical problems. The equilibrium flow through each pipe is only 34 L s^{-1} , significantly lower than for a single pipe (Figure 9). This value is not far from the point where the buoyancy from the bubble formation in the pipe is not sufficient anymore to keep up the self-sustaining mechanism. For this reason, the probability that the degassing system would be stopped by the processes mentioned in section 5.3 is much higher than with only one pipe.

5.5. One Degassing Pipe, Outflow out of the Lake

[32] In this scenario water from the pipe is removed from the lake instead of released to the lake surface. The influence both on the simulated temperatures and CO_2 pressures are negligible. The lake level would drop by about 0.5 m more during the dry season, but this would still be easily refilled during the wet. The main difference to the standard scenario is that the dissolved salts are much more efficiently removed from the lake (see below).

5.6. One Degassing Pipe, Inlet at Different Depths

[33] Compared to the standard scenario, the inlet of the pipe is set at 160 and 180 m instead of 203 m depth. In this case, the entire lake volume below the inlet of the tube would be filled with water which has about the same properties as the deep source water. Furthermore, the degassing would be much slower, especially during the first years,

Figure 10. Simulated CO₂ gas pressures and temperatures in the scenarios (left) with a double strength deep water input, (middle) with two degassing pipes, and (right) with one degassing pipe that discharges out of the lake for standard climate conditions.

because of the lower CO₂ concentrations at the inlet depth (Figure 7). Consequently, the degassing efficiency could not be improved by starting the degassing higher above bottom and subsequently lowering the pipe inlet. Such a schedule is also not necessary to preserve the stable stratification of the lake as it has been previously suggested for a much faster degassing operation using 12 pipes [Kusakabe *et al.*, 2000].

6. Side Effects of the Degassing System

[34] The degassing system has several side effects on Lake Nyos. The most obvious of these effects is the deepening of the seasonally mixed layer. According to the simulations presented above, the lake will be occasionally mixed almost down to the deepest reaches after 50 years of degassing (Figure 8). A stratification builds up every year during the wet season in the surface layer, and it can be maintained for several years in the standard meteorological scenario, but it is finally destroyed by strong convective mixing during the next cool dry season. The lake thus changes from meromictic (i.e., with permanently stratified deep water) to

oligomictic (i.e., occasionally mixed to the bottom). The model does not predict the formation of a new permanent chemocline at intermediate depth which could possibly emerge due to the dissolution of sedimenting material. However, the total observed sedimentation rates are on the order of 200–400 g m⁻² yr⁻¹ (B. Müller, personal communication, 2004). Even if this material were completely dissolved within a depth range of 100 m, this could lead to a maximal annual density difference on the order of a few mg L⁻¹ between the new hypolimnion and the mixolimnion. At the typical mixing temperature of 22°C, this density difference corresponds to a temperature difference of much less than 0.1°C and is not probable to cause permanent stratification.

[35] Furthermore, the degassing operation leads to a reduction of the total content of dissolved salts in the lake, because they are transported to the surface and then washed out by the throughflow (Figure 11). The average outflow of alkali ions for the 50 years of simulation is 2.9 g s⁻¹ without a pipe and 8.5 g s⁻¹ with a pipe. The estimated input with the deep water source is 4.1 g s⁻¹, which can

Figure 11. Concentration of alkali ions in the lake water in 2002 and for three different scenarios at the end of the simulation in 2052.

only compensate half of the losses caused by the washout process.

7. Conclusions

[36] The degassing system currently installed in Lake Nyos is sufficient to reduce CO_2 concentrations in Lake Nyos to a safe level, provided that the operation of the system is maintained in the long term. The system is adequate even if the future discharge of the deep water source should be double the estimated present value of 18 L s^{-1} . The degassing system will approach a steady state when the output of gas is equal to the input with the deep water source. As it is the case with the presently installed pipe, the inlet should be situated within the deepest layers of the lake, as a higher inlet would simply slow down the degassing process and finally leave more gas remaining in the lake. The possible variability of the meteorological forcing has no significant influence on the development of CO_2 concentrations in the different degassing scenarios.

[37] The installation of additional pipes is not necessary to remove the CO_2 from the system and to keep it at a low level, but it can be considered as an option to remove the gas more quickly from the lake. The time needed for degassing can be almost halved using two pipes. An alternative for achieving faster degassing would be the installation of a pipe with a larger diameter, since up to a certain limit the water flow at a given CO_2 concentration grows approximately with the fourth power of the pipe radius. Whether a faster degassing is necessary depends on the evaluation of the present-day situation in the lake. On the

basis of the fact that the total amount of CO_2 is already approaching the level after the 1986 eruption, a much more energetic trigger would be needed today to induce an eruption. Furthermore, since the vertical distance from the deep water to its saturation level is currently around 100 m, the self-amplifying mechanism proposed for the 1986 eruption would hardly be possible. However, in case of a massive rockfall from the steep cliffs into the lake, it cannot be excluded that a significant part of the deep water would be pushed up to a depth where it can be degassed. Consequently, the decision on the installation of additional pipes should be based on an assessment of the risk involved with the gas currently remaining in the lake and the economic costs of the additional installation. Finally, it should be kept in mind that at present the flooding that would be caused by a possible failure of the natural dam bounding the lake [Lockwood *et al.*, 1988] may be considered as the higher risk.

[38] For the long-term operation after initial degassing, it is certainly most efficient to run only one pipe, which should be constructed such that it requires as little maintenance as possible. Besides causing higher costs, parallel pipes would also increase the probability of interruptions in the steady state operation due to the lack of buoyancy of the rising water. Since the flow rate and consequently the fountain height depend strongly on the CO_2 concentrations at the inlet, the degassing pipe can also be used as a basic continuous monitoring system which is open to the public. However, care should be taken not to interpret an increased flow as an alarming signal, as it might be caused by the sinking of the pipe with the water level.

[39] The degassing operation has also some side effects on the lake. The dissolved salts are washed out of the lake because they are transported to the surface by the pipe and then removed by the surface water throughflow. The mixolimnion depth, which would probably reach an equilibrium level at about 60–70 m depth without degassing, will finally reach down to the pipe inlet with degassing. The lake will thus change from meromictic to oligomictic, i.e., it will be occasionally mixed down to almost the bottom during especially dry and cool dry seasons. Possible effects of these changes on the lake ecosystem cannot be assessed because the ecology of the lake has never been studied. It cannot be excluded, however, that some endemic species loses its habitat due to the degassing operation.

Acknowledgments

[40] The authors would like to thank Joseph Hell and Gregory Tanyileke (IRGM, Cameroon) for providing infrastructure and equipment for the field work at Lake Nyos, Michael Schurter and Michael Sturm (Eawag) for their support in the preparation and transportation of the mooring and the sediment traps, Issa and Emmanuel Nkem Take (IRGM, Cameroon), Christian Dinkel, Andreas Lorke, and Beat Müller (Eawag) for their invaluable help in the field, Peter Reichert (Eawag) for adapting the AQUASIM code for double-diffusive convection, Gaston Kayser for providing the code for calculating the water flow through the degassing pipes, and William C. Evans and Johan C. Varekamp for their constructive reviews.

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