Research Article

Physical characteristics of Acidic Mining Lake 111

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Abstract. Measurements of physical properties have been conducted in Mining Lake 111 (ML111), located in Lusatia, Germany over the time period 1996–2002. In the deepest area of the ML111, a monimolimnion was observed, that persisted for the years 1996–1999. It disappeared in 2000 and again formed in 2001. The definition of the main physical properties, such as the temperature compensation for electrical conductivity, in acidic mining lakes required a lake specific approach. The relation between conductivity, temperature and density was determined for the acidic ML111. The variation in dissolved substances affected these relationships such that conductivity varied with temperature even in different layers of the water column and the limitations for a lake wide correlation was evident. Variation in the conductivity of the epilimnion could be verified, and agreed with the estimates of evaporation from the lake surface during summer stratification. Calculations, following the gradient flux method, indicated vertical transport coefficients between 10^{-7} and 10^{-6} m²/s throughout the hypolimnion. The heat budget indicated that heat was transferred into the lake bed or the ground during spring.

Key words. Acidic lakes; density; conductance; conductivity; vertical transport.

Introduction

Acidic mining lakes are becoming a world-wide environmental concern as the number of abandoned open cast mines increases. The voids created during mining are filled either by groundwater or flooded from nearby external sources. In many cases groundwater carries acid mine drainage, which eventually causes the acidification of the lake waters. Acid mine drainage is formed when pyrite, a sulphur bearing mineral, is exposed to air and water, resulting in the formation of sulphuric acid and iron hydroxide. Pyrite is commonly present in coal layers and in the strata overlying coal layers.

There are over 100 acidic mining lakes in Germany, counting only those having a surface area larger than

0.5 km² (Nixdorf et al., 2001), some of which are still emerging. A number of these lakes are located in close proximity to residential areas and therefore demanded for recreational purposes. Several strategies have been considered to neutralise acidic mining lakes. As part of these strategies, Mining Lake 111 (ML111) was chosen for the implementation of in-situ biogenic alkalinisation experiments due to its small size, high acidity, and high iron and sulphate concentrations (Klapper and Schultze, 1995; Stottmeister et al., 1999; Frömmichen et al., 2001; Koschorrek et al., 2002). In order to evaluate which remediation method might be implemented successfully, one needs to know the level of vertical transport and predict the evolution of stratification (Stevens and Lawrence, 1997; Stevens and Lawrence, 1998; Böhrer et al., 1998; Boehrer, 2000; von Rohden and Ilmberger, 2001; Fisher, 2002).

The specific composition of dissolved substances makes acidic mining lakes ecologically marginal aquatic systems. For example, high concentrations of iron and

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sulphate have an effect on water density and light absorption rate, which eventually shape stratification (Friese et al., 1998; Schimmele and Herzsprung, 2000). The density of water depends on the amount of dissolved materials present in the water, and most of those substances can be traced by measuring electrical conductivity (κ). However, conductivity cannot indicate how the ionic composition is changing and it does not measure non-ionic substances. Thus, the chemical composition of the water in these lakes has a substantial effect on the measurement of conductivity. The conductivity of water is also affected by temperature changes. Hence, to use conductivity as a measure of dissolved substances and for the calculation of density ("equations of state"), it must be referred to a standard temperature. The standard temperature is usually 25°C and this is called the conductance (κ_{25}) (Schimmele and Herzsprung, 2000).

Water density is affected by the composition of dissolved and suspended materials. Established methods used to calculate the density of seawater (Fofonoff and Millard, 1983) and freshwater (Bührer and Ambühl, 1975) from temperature and conductivity are unable to give correct values for mining lake waters. The above mentioned methods usually yield an underestimation of the actual densities and density gradients (Gräfe and Boehrer, 2000; Schimmele and Herzsprung, 2000). Therefore, temperature dependence of density in mining lakes needs to be redefined until performance of the direct density measurements is improved to accomplish high enough accuracy (Gräfe et al., 2002).

In this contribution we develop the relationship between conductivity, temperature and density for ML111 to illustrate the physical characteristics of the lake. We reconstruct and analyse the seasonal variation and the long-term development of conductance in the basin over the last 6 years. We also use this relation to calculate density gradients vertically and point out some difficulties with this approach when dealing with the deeper monimolimnion. The vertical transport coefficients during the stratification period can be evaluated by the gradient flux method and compared to the hypolimnia in other lakes.

Study site and methods

ML111 is located in the Lusatian Mining District of East Germany (51°29′ north, 13°38′ east). It was formed after the cessation of mining activity in 1956 and filled with groundwater inflow by 1967. Acid mine drainage increased the acidity of the lake down to pH 2.6. There is no surface inflow nor outflow (Friese et al., 1998). Any water entering the lake, besides precipitation, will be from groundwater. An outflow would require a very high water table and was never observed.

ML111 is a narrow lake, elongated in the north-south direction, with a surface area of 0.107 km^2 and an average depth of 4.6 m. The lake consists of two sub-basins. These sub-basins are separated by a sill, which is about 2 m below the surface. Water depth reaches a maximum of 10.2 m in the north of the southern basin, which is narrower but longer than the northern basin. The southern basin is divided into two parts by a contraction about 65 m wide. The topography of the basin is shown in Figure 1.

Vertical profiles of pressure, temperature and conductivity were measured at the deepest point of ML111, station R12 (Fig. 1) with a multi-parameter probe (Ocean Seven 316, Idronaut). The measurements were carried out during daytime at frequent intervals between spring and autumn, whilst once every two months in winter. We occasionally anchored an automatically profiling multiparameter probe at R12 as well. Data were recorded at each 10cm depth. The accuracy of the temperature and conductivity sensors was 0.003°C and 0.003 mS cm^{-1} with a resolution of 0.0005°C and 0.001 mS cm^{-1}, respectively (Idronaut, 2001).

A self-contained meteorological buoy (M1, Aanderaa) was deployed at station R12. The buoy recorded air temperature, wind speed and direction, relative humidity, incoming short-wave and outgoing long-wave radiation every 10 minutes. Daily values of precipitation were obtained from the German Meteorology Service for the station Grünewalde, approximately 3 km from the lake.

In order to determine the relationship between conductivity and temperature, water samples were collected from various depths of ML111 on 8 July 1997 and 21 August 2001 (Tables 1 and 2) and taken to the laboratory in 2 litre bottles. The samples were initially cooled to 2°C and then gradually heated up to 30°C. Conductivity and

Table 1. The relationship between conductivity (mS/cm) and temperature (°C) of Mining Lake 111 water collected on 8 July 1997 at several depths.

Sampling depth (m)	κ/κ_{25} in 1997
0.5	0.0167T + 0.5833
5	0.0171T + 0.5738
9	0.0175T + 0.5619

Table 2. The relationship between conductivity (mS/cm) and temperature (°C) of Mining Lake 111 water collected on 21 of August 2001 at several depths.

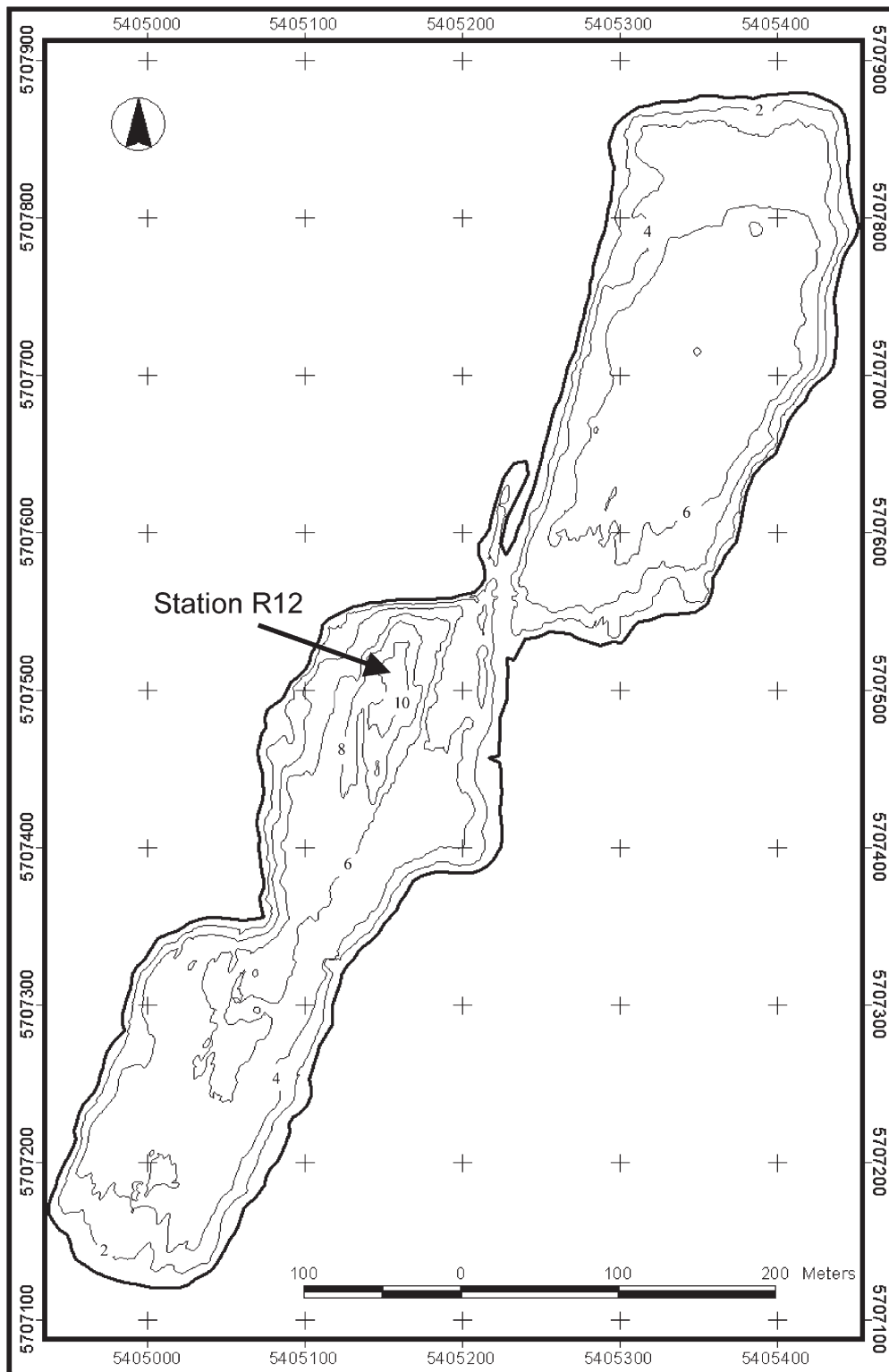| Sampling depth (m)	κ/κ_{25} in 2001
0.5	0.0163T + 0.5932
5	0.0165T + 0.5873
8	0.0173T + 0.5666

Figure 1. Bathymetric map of Mining Lake 111 with the location of station R12 where the meteorological buoy was deployed and the CTD data were recorded (contours from 2 m to 10 m by 2 m intervals) (after Büttner et al., 2000).

temperature were measured during the heating with a multi-parameter probe (Ocean Seven 316, Idronaut).

The reference density of the water was produced under laboratory conditions. The density of ML111 water was measured by injecting 7 ml of the water sample into a U-tube oscillator DMA 5000 (Anton Paar GmbH) with a syringe. The sample was cooled from 30°C to 2°C and density was recorded at intervals of 1°K.

Results and discussion

Temperature stratification

Figure 2 illustrates the temperature time series in ML111 over the observation period. During summer, temperatures ranged from 26°C at the surface to 6°C at the bottom of the lake. It was observed that deep mixing in winter did not extend to the bottom of the lake and overturn did not occur. The temperature profile in February 2001, for instance, (Fig. 3) showed that temperature had a uniform value of 4.1°C from the surface to 5.5 m depth and increased almost linearly from this depth to the lake bottom by more than 1°C, reaching 5.4°C. The overall stability of the water column was maintained by the dissolved substances as indicated by conductance, which rose to 2.62 mS cm^{-1} at 9 m depth. This pattern was in contrast to summer where temperature was the control-

ling agent of stratification stability. The conductance decrease at around 6 m was due to changing ionic concentrations (i.e., evaporation at the surface). The density (Fig. 9) was balanced by the temperature gradient. Gradients in the composition of dissolved substances only played a role below 6.7 m. The change in the water density in the bottom 50 cm of the profile is discussed below.

Conductance and its seasonal variation

Conductivity of water samples was found to increase linearly with increasing temperature (Fig. 4). The conductivity-temperature relatonship and fit parameters are given in Table 1 for 1997 and Table 2 for 2001. Conductance (κ_{25}) resulted from the regression curve from measured conductivity, κ against temperature, T (°C): $\kappa_{25}(0.5\,\text{m}) = 2.603$ mS cm^{-1}, $\kappa_{25}(5\,\text{m}) = 2.501$ mS cm^{-1}, $\kappa_{25}(9\,\text{m}) = 2.899$ mS cm^{-1} in 1997 and $\kappa_{25}(0.5\,\text{m}) = 2.573$ mS cm^{-1}, $\kappa_{25}(5\,\text{m}) = 2.472$ mS cm^{-1}, $\kappa_{25}(8\,\text{m}) = 2.529$ mS cm^{-1} in 2001.

The slope of the κ-T relationship is higher for the 8 m and 9 m samples than those of the shallower depths. There was also some variation between years, which might be connected to the temporal change in water composition. The spatial and temporal variation in water composition was a major limitation of implementing the same evalua-
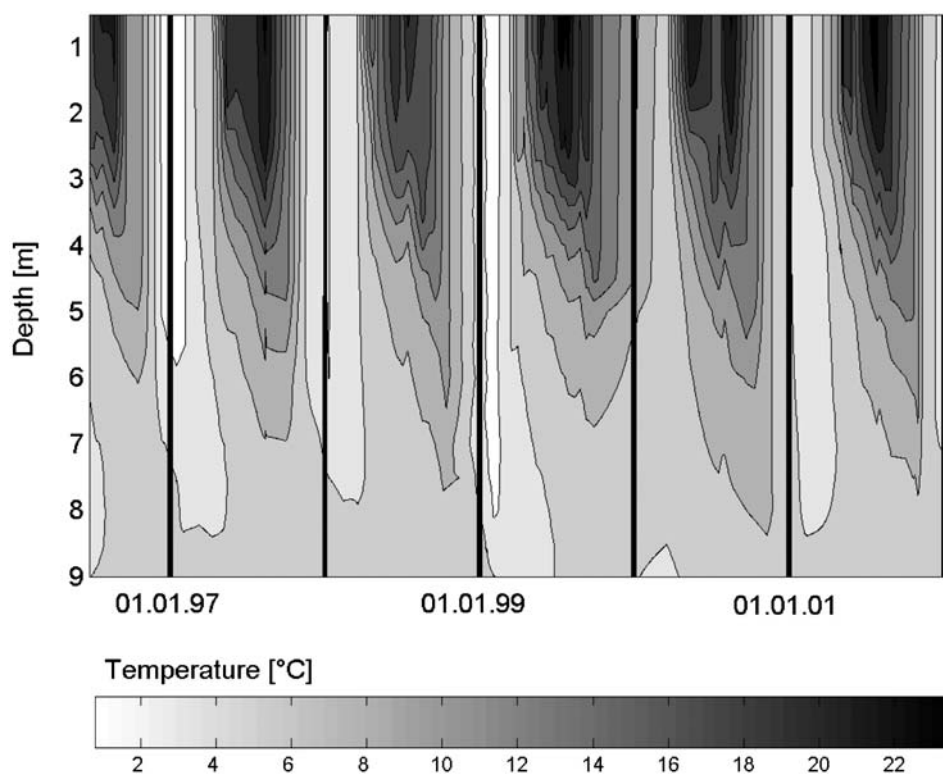

Figure 2. Time series of temperature (°C) in the central basin over the observation period 1996–2002 (contours from 2°C to 22°C by 2°C).
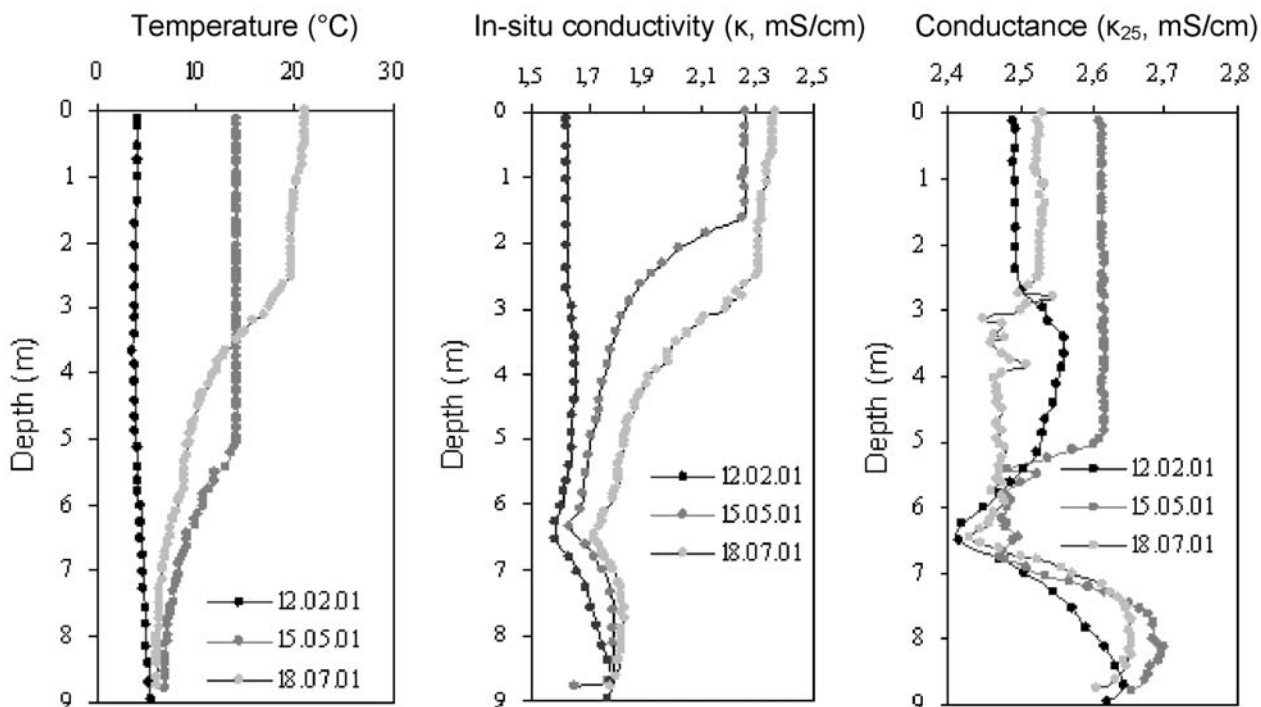
Figure 3. Temperature, in-situ conductivity and conductance (κ_{25}) profiles taken at station R12 at three occasions in 2001.
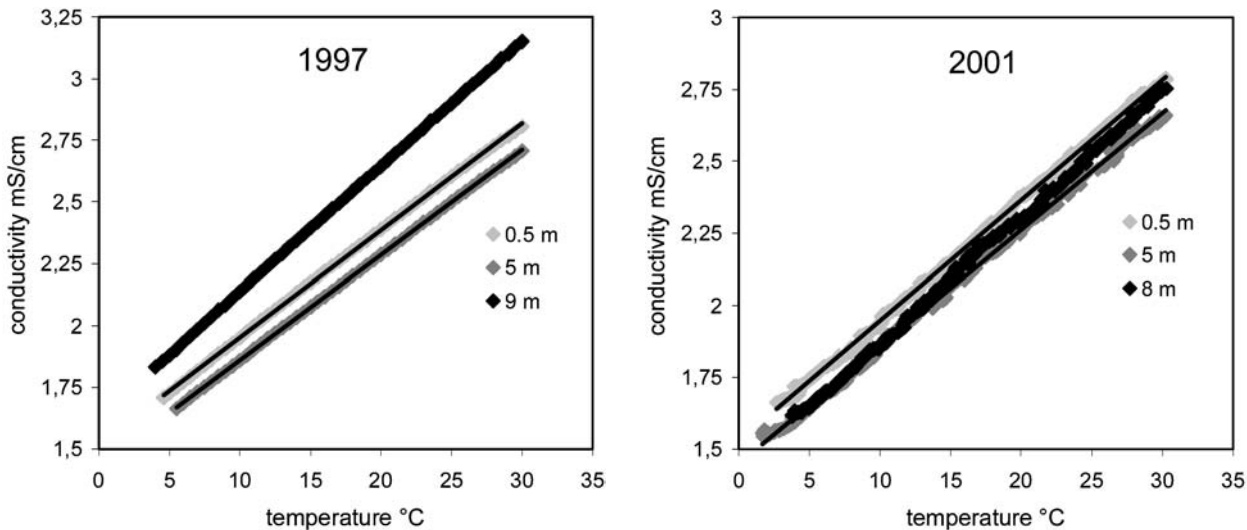

Figure 4. Conductivity versus temperature of water samples from different depths from years 1997 and 2001.

tion for conductance at all depths over the entire observation period. To gain values of conductance for the field data, we implemented a temperature compensation as an average of mixolimnion values (0.5 m and 5 m depth)

$$(\kappa/\kappa_{25})_{2001} = 0.0166T + 0.5844 \qquad (1)$$

This introduces a maximum error of 0.028 mS cm^{-1} at the surface and 0.058 mS cm^{-1} towards the bottom waters

during winter. In Figure 5 we plotted the variation of conductance in the basin over the observation period, based on the profiles taken at station R12.

A layer of high conductance was observed below 7 m depth from the beginning of measurements in 1996 to the last quarter of 1998. This layer persisted through all seasons and indicated the presence of a monimolimnion in the small deep depression of the lake around station R12. The layer then disappeared but emerged again in the
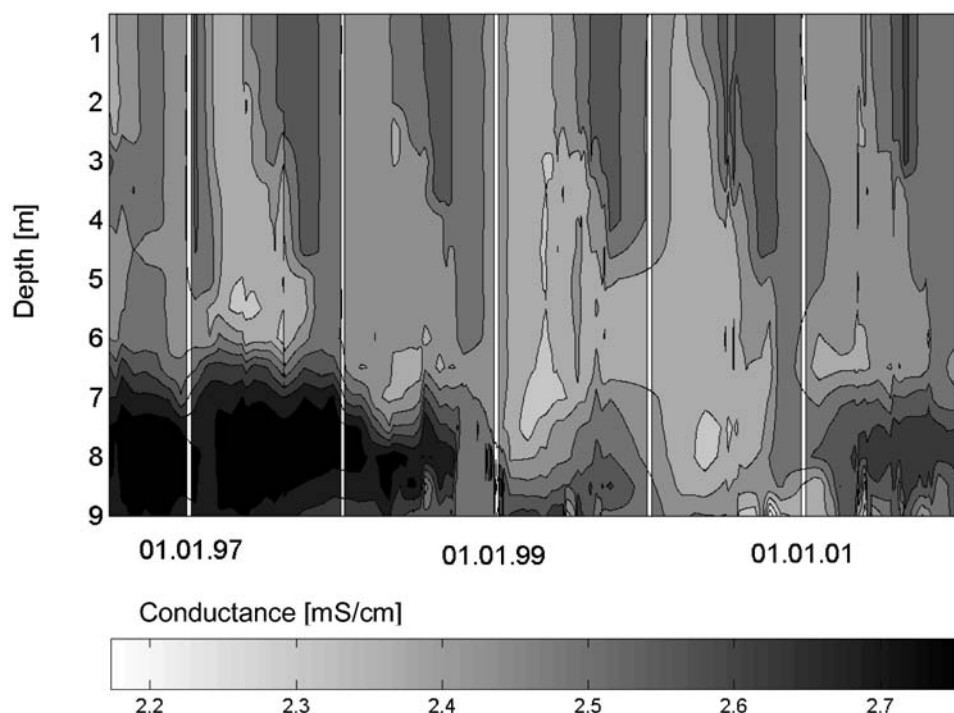
Figure 5. Time series of conductance (κ_{25}) in the central basin over the observation period 1996–2002 (contours from 2.2 mS cm^{-1} to 2.7 mS cm^{-1} by 0.05 mS cm^{-1}).

spring of 2001, although not as strongly. As Stevens and Lawrence (1998) already suspected for Brenda Mines pit-lake, we believe that saline ground water inflow created the monimolimnion.

For the preservation of the monimolimnion, the annual formation and the duration of ice cover on the lake was important. During periods of ice cover, the wind was prevented from acting on the lake surface and thus from introducing turbulent kinetic energy into the lake, which subsequently would have been available for the erosion of the haline stratification. Without ice cover, the possible stable density stratification below the temperature of maximum density, i.e., near 4°C "inverse" stratification, was too weak to withstand an even low wind for a considerable time period, as density differences below the temperature of maximum density were small and the halocline in the lake was located rather near the surface.

A striking feature in the conductance variation was a slight decrease between 8 m, where it reached its peak values, and 9 m depth, which could be attributed to gradients in the chemical composition of dissolved substances.

Moreover, there appeared to be a seasonally high conductance at the surface. It persisted regularly throughout the period of temperature stratification that stabilised the density (Fig. 2). The conductance of surface water changed significantly from 2.4 mS cm^{-1} to 2.65 mS cm^{-1}.

We tested the assumption that this rise in conductance was created by the difference in the accumulated precipitation-evaporation balance during this time of year. We used the following equation to calculate evaporation, E_p (mm d^{-1}) (Richter, 1973):

$$E_p = (a + bu_2) \cdot (E_o - e_L) \tag{2}$$

where u_2 was the wind speed (m s^{-1}) at 2 m above the water surface, E_o (Torr) was vapour pressure, and a and b are constants with values of 0.3 and 0.2, respectively.

E_o and e_L are derived from:

$$E_o = 4.581 \cdot 10^{7.5T/(235+T)} \tag{3}$$

$$e_L = E_o \cdot h \tag{4}$$

where T (C°) was air temperature and h was humidity.

Based on the above equations and precipitation data from the German Weather Service, we show the daily balance in precipitation and evaporation (p-e) in Figure 6. The 4-year mean was found as a loss of 0.1639 mm d^{-1}. A minimum p-e was observed in July 1997 at –37 mm d^{-1}, whilst the maximum was recorded in June 1998 (8.7 mm d^{-1}).

In order to relate the p-e to the increase in conductance, we took the year 1997 as an example. In 1997 p-e
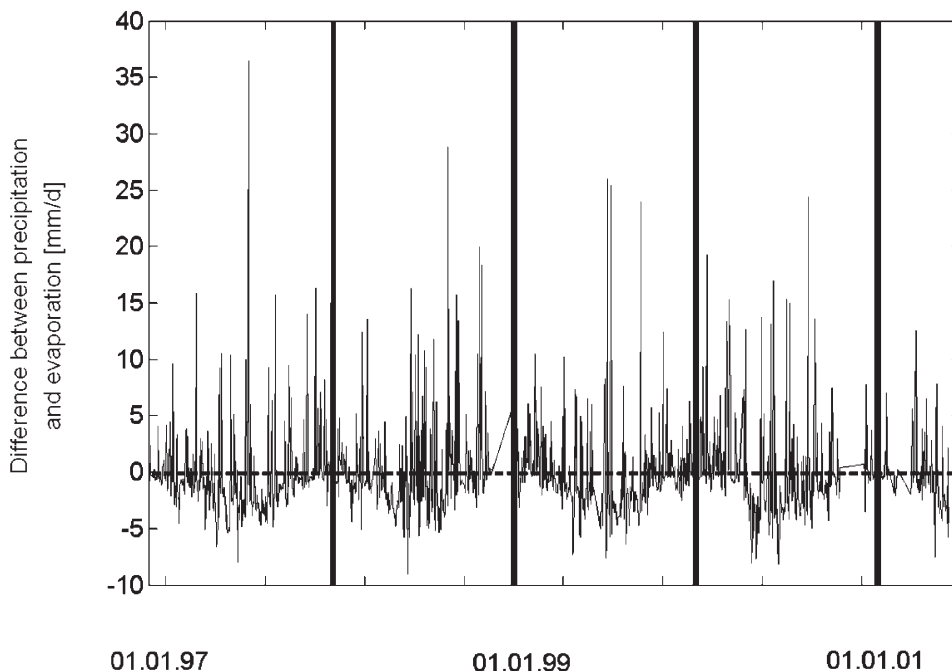
Figure 6. Daily balance of precipitation and evaporation in Mining Lake 111 (evaporation calculated after Richter (1973), precipitation measurements from the German Weather Service).

balance summed to -168 mm between April and October. This value (168 mm) corresponded to 6% of a roughly 3 m thick epilimnion, which was lost to the atmosphere. During the same period, conductance of the epilimnion increased by 0.25 mS cm^{-1} (i. e., by 0.25 mS cm^{-1}/2.4 mS cm^{-1} = 10%). The signal in conductivity was similar to that anticipated from the calculation of evaporation loss and thus indicated that evaporation might account for the conductance increase in the epilimnion. The difference reflected the inaccuracy of our epilimnion depth assumption, the difficulty of measuring a representative wind velocity over small lakes, and the limited applicability of the evaporation formula at the observed low wind speeds. In addition the micro-meteorology over the lake further influenced the accuracy of the results as the precipitation record was taken from a site 3 km away.

Density calculation

In a first order approach for freshwater lakes, the influence of dissolved substances on water density could be included by extending the temperature correlation for pure water with a term adding conductance with a lake specific factor γ (see e. g. Bäuerle et al. (1998) for the water density in Lake Constance):

$$\varrho(T, \kappa_{25}) = \varrho(T) + \gamma \cdot \kappa_{25} \qquad (5)$$

where

$$\varrho(T) = 999.8429 + 5.9385 \cdot 10^{-5}\, T^3 - 8.56272$$
$$\cdot 10^{-3}\, T^2 + 6.54891 \cdot 10^{-2}\, T \qquad (6)$$

T being in °C.

The measured variation in density with temperature was shown for ML111 water and for pure water (eq. 6) in Figure 7. The difference was attributed to dissolved substances that revealed in the conductance. To gain further insight, this density difference between the lake water and pure water measurements was plotted in Figure 8. A pronounced temperature dependence became visible. The difference of as much as 1.95 kg m^{-3} was seen at low temperatures between the two density functions in Figure 8, while at higher temperatures this difference was considerably smaller. Using a constant value for γ would be of limited numerical accuracy, while a third order polynomial (Fig. 8) matched the data over the entire range of considered temperatures within 0.002 kg/m^3, and corresponded to the accuracy of the density measurements.

$$\gamma(T) = -7.522 \cdot 10^{-6}\, T^3 + 2.613 \cdot 10^{-4}\, T^2$$
$$- 5.146 \cdot 10^{-3}\, T + 0.7918 \qquad (7)$$

In conclusion, inserting eq. 7 into eq. 5 reflected the density variation of the investigated water parcel of ML111 to the best of our knowledge.
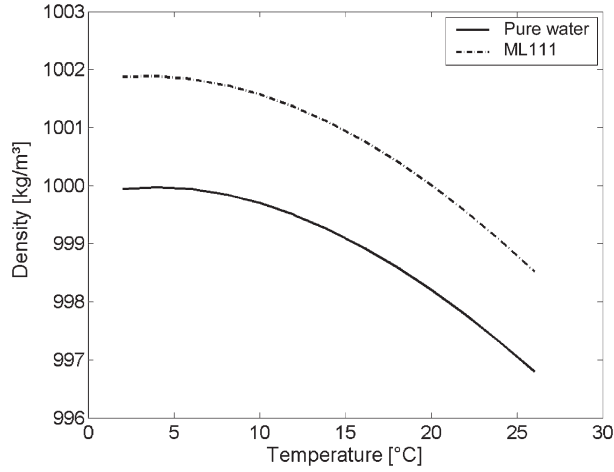
Figure 7. Variation in density with temperature for Mining Lake 111 water and pure water (error bars are within the size of the line thickness).
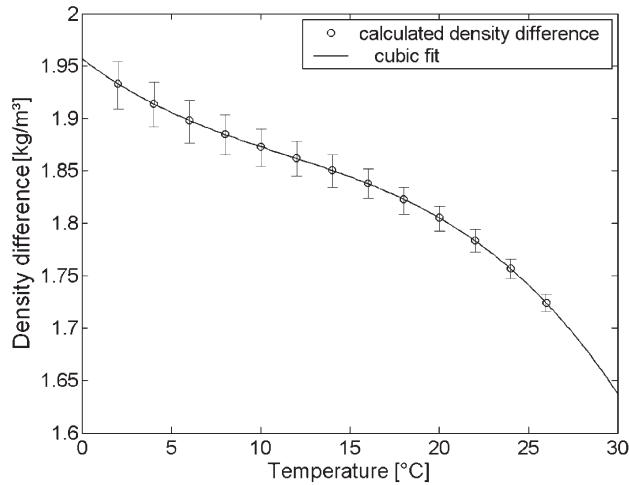

Figure 8. Temperature dependence of γ-coefficient, relating conductance with its contribution to density.

In the next step, the coefficient for thermal expansion (α) defined as:

$$\alpha = -\frac{1}{\varrho} \frac{\partial \varrho}{\partial T} \tag{8}$$

could be evaluated as

$$\begin{aligned}
\alpha = & (-1.7797 \cdot 10^{-7} + 2.2543 \cdot 10^{-8} \, \kappa_{25}) \, T^2 \\
& + (1.783 \cdot 10^{-5} - 5.2208 \cdot 10^{-7} \, \kappa_{25}) \, T \\
& + (-6.5425 \cdot 10^{-5} + 5.1409 \cdot 10^{-6} \, \kappa_{25}) \tag{9}
\end{aligned}$$

giving us $\alpha = 2.368 \cdot 10^{-4} \, K^{-1}$ at 22 °C with a κ_{25} value of 2.55 mS cm^{-1}. This value was 5 % higher than the α for pure water at the same temperature.
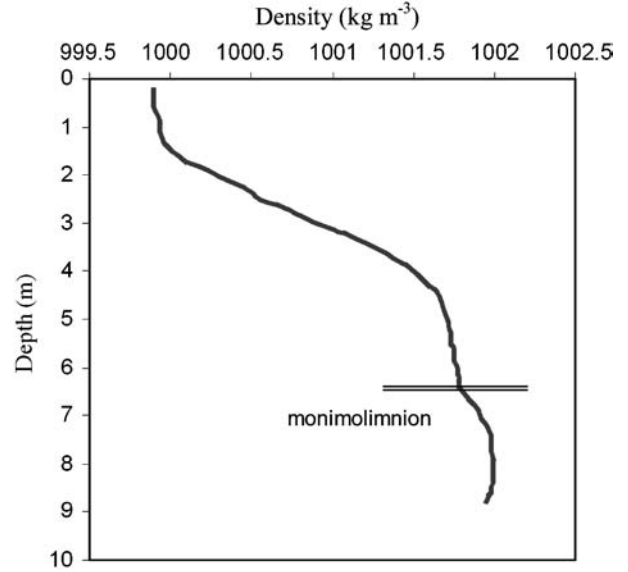

Figure 9. Averaged density profile in Mining Lake 111 (station R12) over the period used for calculating vertical transport coefficients.

One should note that the accuracy of using such a procedure was limited by the variation in water properties within ML111. Using the above formulas, the density profile (Fig. 9) illustrated that a monimolimnion was clearly separated from the hypolimnion below 6.7 m depth; a feature also seen in the time series in conductance variation. However, a slight decrease in density at the bottom reflected the difficulty of calculating density above the sediment layer. We do not believe that an unstable density stratification could be established under the given conditions, and we consider it an artifact attributed to variations in the composition of dissolved substances.

Vertical transport

Vertical exchange was limited during the stratification period. This was important for nutrient supply and, especially so, for Mining Lake 111 because of the planned remediation based on biogenic alkalinisation. Knowledge about the vertical transport of substances, e.g. oxygen, is essential when implementing remediation concepts that rely on an oxygen free environment.

Every 6 hours, profiles of fine vertical resolution of several physical parameters were measured by the automatically profiling multiparameter probe at R12, of which we concentrated on the temperature field (Fig. 10) in the stratification period from 16 May to 29 July 2001, when convective circulation could be excluded for the deeper waters.

To eliminate short-term changes in the temperature profiles that could be caused by internal waves, we aver-
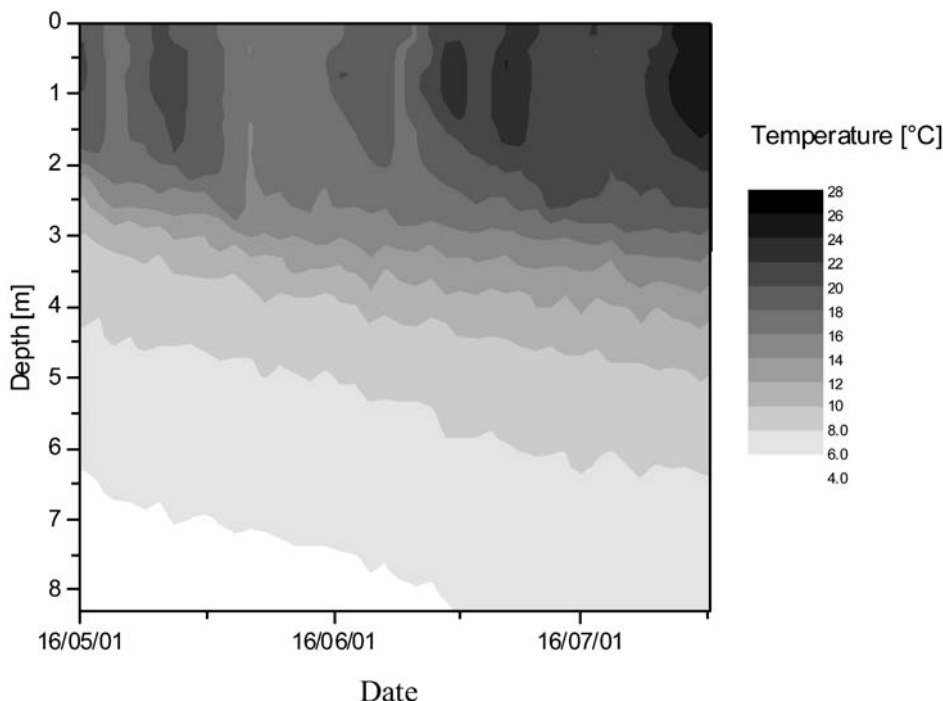
Figure 10. Temperatures in Mining Lake 111 (station R12) over the period used for calculating vertical transport coefficients.

aged 8 consecutive profiles corresponding to mean temperature profiles T_i over 2 days. Over the time period Δt those profiles showed changes in the heat budget of each layer i in the lake of equal thickness Δz, cross sectional area A_i, and thus volume $V_i = A_i \Delta z$. The heat flux, controlling the temperature profile through the specific heat C_p, could quantitatively be evaluated by assuming all heat was being introduced from above unless a separate source term Q_i, such as geothermal heat flux or residual solar irradiation at the respective depth, was specified for the respective layer i (Heinz et al., 1990; von Rohden and Ilmberger, 2001).

$$K_i = \frac{\sum_{k=1}^{i} \left\{ \frac{\Delta T_k}{\Delta t} V_k - \frac{Q_i}{c_p \varrho} \right\}}{\frac{\Delta T_i}{\Delta z} A_i} \qquad (10)$$

Under the assumption of no heat source $Q_i = 0$ W, we found vertical transport coefficients ("coefficient of diffusivity") K_i in the range of 2 to $4 \cdot 10^{-7}$ m²/s (Fig. 11), which lie just above the molecular transport for heat within water. The results for the epilimnion (above 3 m depth) were meaningless, as the stratification broke down during the observation period. For the hypolimnion (4.5 to 6.5 m depth) the resulting vertical transport coefficient of $4 \cdot 10^{-7}$ m²/s corresponds to a stratification of about $N^2 = \frac{g}{\varrho} \frac{\Delta \varrho}{\Delta z} = 0.4 \cdot 10^{-3}$ s⁻² (i. e., the square of the so-called

stability frequency or Brunt-Väisälä frequency), and for the chemocline/upper monimolimnion (6.5 to 8 m depth) a transport coefficient of $2 \cdot 10^{-7}$ m²/s corresponds to a bulk stratification of $N^2 = 1,5 \cdot 10^{-3}$ s⁻². Rohden and Ilmberger (2001) published a plot of vertical transport coefficients against stratification which they had attained from measurements of a spreading artificial SF6 spike in the salinity stratified mining lake Merseburg-Ost 1b about 110 km west of Mining Lake 111. Rohden and Ilmberger (2001) showed that a higher transport coefficient was connected with a lower stratification. Their values were at the same order of magnitude but 2–4 times higher, which might be connected to the fact that wind speeds in Merseburg-Ost 1b were 1.9 times higher (Boehrer et al., 2000). Considering the small size of ML111, i.e., weak influence by winds, these results fit into the general picture (Heinz et al., 1990) we have of transport coefficients in stratified deep waters of lakes from earlier investigations (Jassby and Powell, 1975; Imboden et al., 1983; Wüest, 1987).

In their numerical approach for the vertical transport in Merseburg-Ost 1b, Walz (1997) and von Rohden and Ilmberger (2001) used a geothermal heat flux $Q = q (A_k - A_{k-1})$ at $q = 0.23$ W/m² through the contact area $A_k - A_{k-1}$ of each water layer i with the sediment which reflected the usual value for deeper lakes in the respective area. In ML111, such a value would cause smaller transport coefficients smaller than the molecular diffusivity and even partly smaller than 0, which in general is not acceptable.
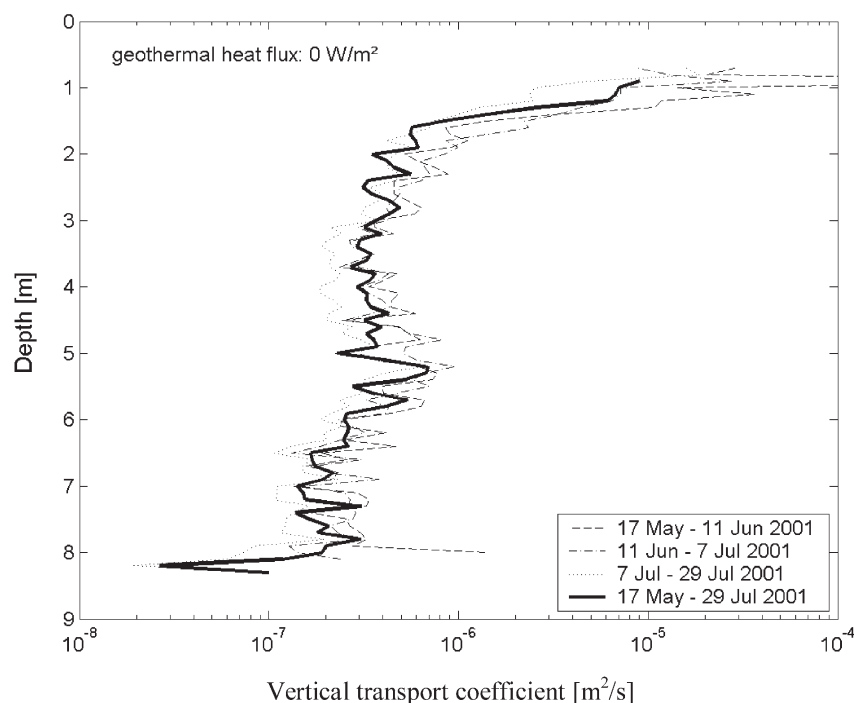
Figure 11. Profile of vertical transport coefficients in Mining Lake 111 calculated from temperature profiles over the entire time period in early summer 2001 under the assumption of 0, −1 and 0.23 W/m² geothermal heat flux.

Conditions were different in ML111, as the lake was small and the maximum depth was only about 10m. At this depth, groundwater could still show a pronounced seasonal temperature variation. In addition, the lake bed might have cooled during winter and in early summer it acted as a heat sink for the lake.

ML111 was very turbid. Attenuation depths measured previously (Koschorreck and Tittel, 2002) showed seasonal variability at about a mean attenuation depth of 1.2 m. Thus, the heat input at greater depths was small. However, including a small limited value for the radiative heat input would even require a higher heat sink than concluded above, and consequently, the arguments for the heat sink are strengthened.

Groundwater inflow was hard to estimate quantitatively. It would contribute to the heat sources Q. The vertical shift of the entire water column by inflowing groundwater was assumed to be small and hence ignored. Consequently, we did not include the expected error connected to the numerical evaluation of the vertical transport coefficient nor did we consider the error by the measurement. However, we performed the calculation of vertical transport coefficients under the assumption of 0.23 W/m² and −1 W/m² to indicate the accuracy of the evaluation. In conclusion, during the evaluation period the heat transport by groundwater inflow and heat conduction was probably directed more out of the lake than into the lake. Being aware of the difficulties of quantify-

ing groundwater inflow, the vertical transport coefficient between 8 m and 4.5 m depth lay in the range of $1 \cdot 10^{-7}$ m²/s to $1 \cdot 10^{-6}$ m²/s for early summer 2001.

Conclusions

In this paper, we presented the physical characteristics of an acidic mining lake, Mining Lake 111, showing a distinct relation between temperature, conductivity, and density. The dependency of conductivity on temperature might vary even between different layers of the lake. A general definition that suits the entire range of lakes of this kind is virtually impossible due to their different properties associated with water composition. Besides, dissolved non-ionised substances, which contribute to density, are not reflected in conductivity. Therefore, lake-specific definitions are required for density and conductivity.

Vertical transport calculations were carried out by using the gradient flux method under different assumptions for geothermal heat flux. The calculations indicated that Mining Lake 111 lost heat to the lake bed rather than gaining heat from it. During early summer the transport coefficient throughout the hypolimnion ranged between 10^{-7} and 10^{-6} m² s^{-1}.

Acknowledgments

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