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# Vertical turbulent diffusion and upwelling in Lake Baikal estimated by inverse modeling of transient tracers

R. Kipfer and M. Hofer
Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), Dübendorf, Switzerland

F. Peeters
Department of Physics, Swiss Federal Institute for Environmental Science and Technology (EAWAG), Dübendorf, Switzerland

D. M. Imboden
Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), Zürich, Switzerland

V. M. Domysheva
Limonlogical Institute, Siberian Division of the Russian Academy of Sciences, Irkutsk, Russia

**Abstract.** Vertical turbulent diffusion coefficients, upwelling velocities, and oxygen depletion rates are estimated by inverse modeling of the concentrations of CFC-11 ($CCl_3F$), CFC-12 ($CCl_2F_2$), $^3H$, $^3He$, and dissolved oxygen for the southern, central, and northern basin of Lake Baikal. A model is developed that considers two regions in each basin of Lake Baikal: (1) a surface mixed layer (SML) 400 m thick and (2) a deepwater column (DWC) below 400 m. The SMLs are assumed to be well mixed. In each of the DWCs, passive tracers are transported by vertical turbulent diffusion and upwelling. Upwelling is generated by a depth-dependent source of water because of density plumes propagating from the SML downward to larger depths. This water is considered to contain the same tracer concentrations as the SML. The tracer concentrations in the SMLs of the three basins are coupled to the atmosphere by gas exchange (including water vapor transport) and precipitation to the catchment by river inflow and outflow and to the neighboring basins via diffusive exchange and advection. SMLs and DWCs of the same basin are connected by vertical turbulent diffusion, density-driven water transport, and upwelling. Beginning at the turn of this century, the tracers CFC-11, CFC-12, $^3H$ and $^3He$ are modeled simultaneously to predict modern concentrations. On the basis of the tracer data the vertical diffusion coefficient $K_z$ is determined to be $4.6 \times 10^{-4}$ $m^2$ $s^{-1}$ ± 10% for the southern, $6.3 \times 10^{-4}$ $m^2$ $s^{-1}$ ± 10% for the central, and $1.7 \times 10^{-4}$ $m^2$ $s^{-1}$ ± 25% for the northern basin. The vertical advective flux of water at 400 m water depth is calculated as 110 $km^3$ $yr^{-1}$ in the southern, 70 $km^3$ $yr^{-1}$ in the central, and 290 $km^3$ $yr^{-1}$ in the northern basin. Concentration of dissolved molecular oxygen is modeled by using the estimated transport parameters and by fitting for the unknown consumption rate. Inverse modeling of oxygen suggests that $O_2$ depletion in the DWC can be described by a volume sink of 44 ± 3 $mgO_2$ $m^{-3}$ $yr^{-1}$ combined with an areal sink at the sediment water interface of 17000 ± 3000 $mgO_2$ $m^{-2}$ $yr^{-1}$.

## 1. Introduction

Lake Baikal (Siberia) is the habitat of ~45% of all known limnic species, most of them endemic to Lake Baikal [*Timoshkin*, 1994]. This unique ecosystem has evolved during the 20-40 million years of Lake Baikal's existence and is adapted to several months of ice cover, extremely large water depths (up to 1630 m), low nutrient concentrations, and high concentrations of dissolved oxygen everywhere in the water column (80% of saturation or more). The latter is a consequence of the low oxygen depletion rate in the deep water and the rapid vertical water exchange [*Weiss et al.*, 1991; *Peeters et al.*, 1997; *Hohmann et al.*, 1998].

Lake Baikal consists of three main basins (the southern (S), the central (C), and the northern (N) basin), which are separated by sills with a maximum depth of 350-400 m. The central and the northern basin are additionally connected by a very narrow strait between Cape Svyatoi Nos and Bol'shoi Ushkanii Island with a maximum depth of ~490 m. Figure 1 shows a map of Lake Baikal on which the deepest location in each basin is marked by a solid circle. On average the deep water below 300 m depth of the S and C basins of Lake Baikal is renewed by water from the seasonal mixed layer in ~10 years, whereas the deep water in the N basin is renewed in ~6 years [*Peeters et al.*, 1997].

Vertical distributions of transient tracers measured in Lake Baikal suggest that the fast renewal of deep water cannot be

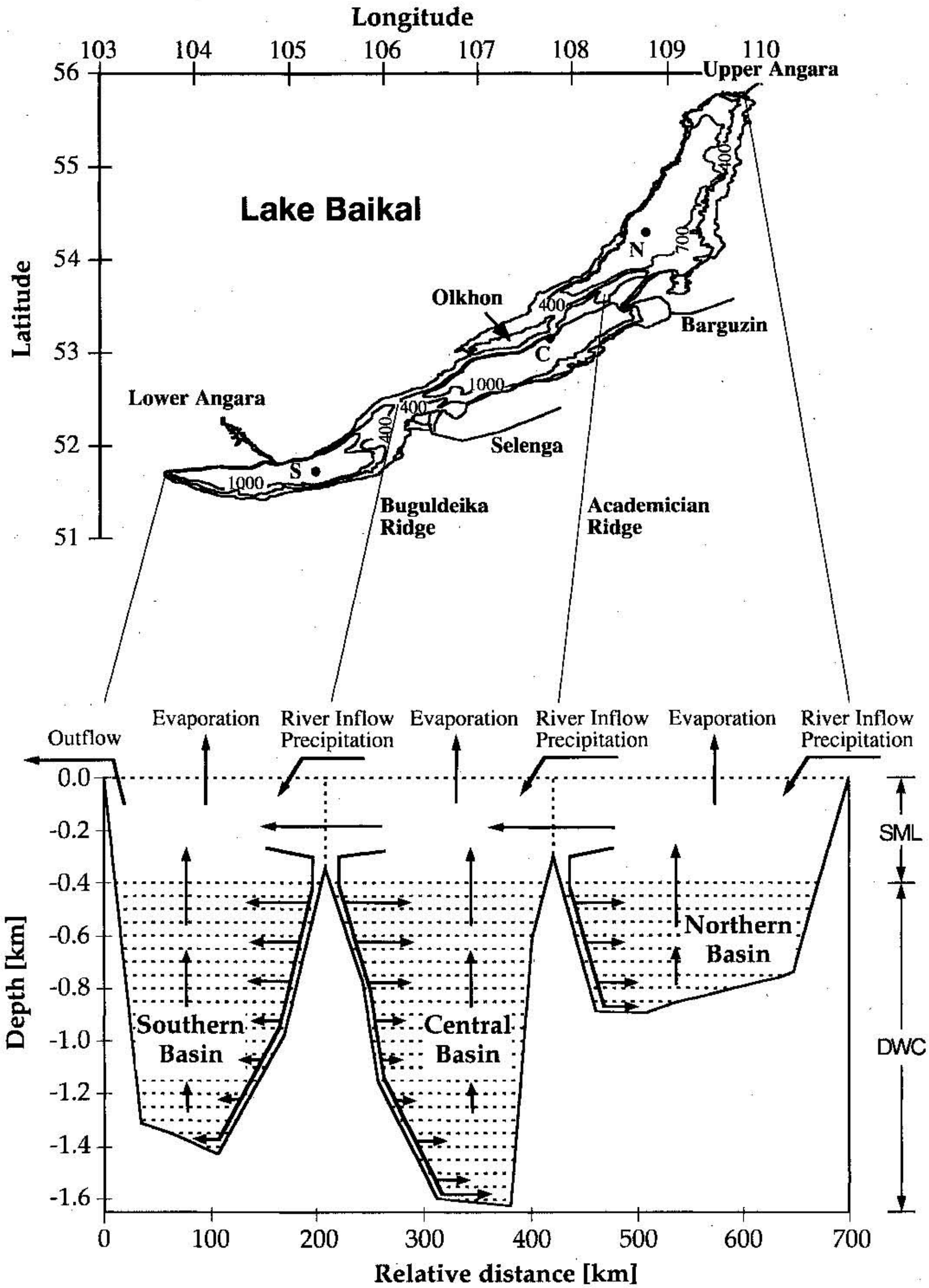


**Figure 1.** Map of Lake Baikal and cross-sectional view of the used transport model. Black dots in the map mark the deepest station of the southern (S), central (C), and northern (N) basin. Arrows indicate the direction of the advective water transport. The dashed lines show the employed discretization in the deepwater column (DWC). The surface mixed layer (SML) is defined by the water column above 400 m.

solely accomplished by vertical turbulent diffusion. Advective transport is required to generate the observed increase of CFC-11, CFC-12 and dissolved oxygen and the decrease of temperature and $^3$He in the deepest region of each basin [*Weiss et al.*, 1991, *Hohmann et al.*, 1998]. The conditions required for convection to occur down to maximum depth are controlled by the density anomaly of freshwater and the pressure dependence of the temperature of maximum density ($T_{md}$) in combination with water temperatures below 3.6°C in the deep water of the lake [e.g., *Shimaraev and Granin*, 1991; *Hohmann*, 1997]. At 250 m depth the water temperature ranges between 3.5 and 3.6°C during the entire year. At this depth, $T_{md}$ is ~3.5°C. Hence the water density in Lake Baikal at 250 m is always close to the maximum density possible according to temperature. Therefore temperature-driven convection by, for example, seasonal warming and cooling of surface water can only reach down to ~250 m. Below

500 m, $T_{md}$ is <3.0°C, and the temperature of the ambient water is >3.1°C. If water with a temperature of 3°C is transported down to depths below 500 m, its density would be larger than that of the ambient water. Consequently, convection could be triggered by processes that force cold water to penetrate through the potential barrier in the depth range between 200 and 500 m.

The potential barrier can be overcome by water with increased salinity or a large load of suspended particles, both leading to an increase in water density relative to the ambient water [e.g., *Peeters et al.*, 1996; *Hohmann* et al., 1997]. In addition, strong winds might force water downward and lead to deep water convection by thermobaric instabilities [e.g., *Weiss et al.*, 1991].

Several processes generating salinity- and/or turbidity-driven density plumes have been observed in Lake Baikal.

1. Because of its larger salinity and concentration of suspended particles, the water from the Selenga River (see Figure 1) flows along the lake bottom through the subsurface Kukui Canyon to the deep part of the central basin [*Hohmann et al.*, 1997].

2. Because of the nonlinearity in the equation of state for freshwater, mixing equal volume water masses of different temperatures will produce water that has a density greater than the mean density of the two constituents. If one water mass is colder and the other warmer than the $T_{md}$, the resulting water may have a density greater than either of the constituents. This process is called cabbeling. Cabbeling occurs at the thermal bar off the eastern shore of the central basin [*Shimaraev et al.*, 1993; *Peeters et al.*, 1996], but cabbeling alone could not reach below 250 m depth. It is aided by the increased salinity of the Selenga River water, which is trapped between the thermal bar and the shore [*Hohmann et al.*, 1997]. Density plumes at the thermal bar, which result from the mixing of open lake and trapped Selenga water, have been demonstrated to penetrate down to 400-500 m depth [*Peeters et al.*, 1996].

3. Interbasin exchange at the Academician Ridge (Figure 1) produces water of larger density, which then flows into the deep part of the N basin [*Peeters et al.*, 1996; *Hohmann et al.*, 1997]. This process is driven by the salinity difference between surface water from the C and the N basins.

4. Hydrothermal vents introduce warm and saline water at 200-400 m depth into Frolikha Bay and other regions in the northern basin. The hydrothermal water can propagate down to the bottom of the N basin [*Kipfer et al.*, 1996].

Wind forcing in combination with the thermal baricity was suggested by *Weiss et al.* [1991] to be responsible for the large-scale convection in Lake Baikal. This process is a consequence of the pressure dependence of the temperature of maximum density. If strong winds cause a large downward displacement of the water at the downwind end of the lake, the water column, which originally was stable, may become unstable. Situations favorable for this kind of instability are met in late fall or spring when water temperatures increase with depth to 3.5°C at ~250 m depth. In fact, a numerical model by *Killworth et al.* [1996] indicates that wind forcing can generate localized vertical plumes in the open water of Lake Baikal. *Walker and Watts* [1995] demonstrated by a three-dimensional model that a vertical displacement of surface water by several hundred meters leads to localized plumes reaching down to the bottom of Lake Baikal. According to their model these plumes would be largest near the lake boundaries. This kind of localized mixing process may have been observed in recent conductivity-temperature-depth (CTD) profiles (*N. G. Granin*, personal communication, 1998).

Each of the proposed processes has a different physical characteristic, but most of them cause water masses to sink along the bottom boundary of the lake. The sinking density plumes affect the mean tracer concentrations in the open water regions in that they act as a local source of water, which has the characteristics of surface water. Hence density plumes convectively transport tracers such as CFC-11, CFC-12, $^3$H, $^3$He, and $O_2$ from the surface layer directly to different depths of the lake. This may explain why apparent water ages are sometimes found to be smaller at the lake bottom than in intermediate water layers [*Weiss et al.*, 1991; *Hohmann et al.*, 1998].

Several authors have determined mean exchange rates of deep water with water from the seasonal mixed layer [*Weiss et al.*, 1991; *Peeters et al.*, 1997; *Hohmann et al.*, 1998]. The aim of this study is to separate the contributions of vertical turbulent diffusion and convection to vertical transport. *Killworth et al.* [1996] developed a quasi-stationary advective-diffusive model to distinguish between these two transport mechanisms. They assumed that the oxygen concentrations are at steady state, oxygen depletion rates are constant, and the time-dependent CFC-12 distribution can be approximated by $C(t,z) = a\,t + \hat{C}(z)$, with $a$ being constant and $\hat{C}(z)$ being independent of $t$. They obtained upwelling water volumes at 400 m depth of ~500 km$^3$ yr$^{-1}$ in the S, ~1000 km$^3$ yr$^{-1}$ in the C, and ~400 km$^3$ yr$^{-1}$ in the N basin. Vertical diffusivity below 400 m was of the order of a few times $10^{-6}$ m$^2$ s$^{-1}$ or smaller.

In this study we estimate mean vertical turbulent diffusion coefficients and upwelling velocities by inverse modeling of the vertical concentration distributions of CFC-11, CFC-12, $^3$H, and $^3$He. The method is based on a time-dependent advection-diffusion model. Because the CFCs can be considered as being conservative in the oxygen-rich environment of Lake Baikal and because the decay constant for $^3$H is well known, the parameter estimation can be restricted to those parameters that describe vertical transport. In a second step these transport parameters can then be used to model oxygen consumption in Lake Baikal. We believe that the chosen model assumptions are less restrictive than those by *Killworth et al.* [1996]; for instance, we neither assume that the CFC concentrations grow at a constant rate nor that the growth rate is the same at all depths.

## 2. Database

Water samples were collected between 1992 and 1997 during several expeditions on Lake Baikal. They were analyzed for the chlorofluorocarbons CFC-11 and CFC-12, tritium ($^3$H), $^3$He, and dissolved molecular oxygen ($O_2$). CFC-11 and CFC-12 data from 1995, 1996, and 1997 were determined by the analytical procedure of *Hofer and Imboden* [1998]. Additional CFC-12 data from 1988 and 1991 were taken from *Killworth et al.* [1996]. A detailed description and discussion of the $^3$H and $^3$He measurements are given by *Hohmann et al.* [1998]. Dissolved oxygen concentrations were measured by Winkler titration. The profiles of the oxygen concentrations were corrected for outliers by comparing them to profiles obtained from an oxygen sensor mounted on a CTD probe.

The tracer data (Figure 2) used to calibrate the model developed below were taken at the deepest station of each basin, indicated by solid circles in Figure 1. The errors for the different tracers are estimated to be 7% for CFC-11, 5% for CFC-12 and $^3$He, and 1 tritium unit for tritium (1 TU = 0.2488x$10^{-14}$ mL STP g$^{-1}$). The symbols in Figure 2 indicate individual measurements. In addition, the tracer data below 400 m depth are fitted by third-order polynomials (lines in Figure 2). Data and cubic fit predictions agree according to a $\chi^2$ test. The standard deviation of the relative error $(m$-$p)/m$ (where $m$ and $p$ stand for

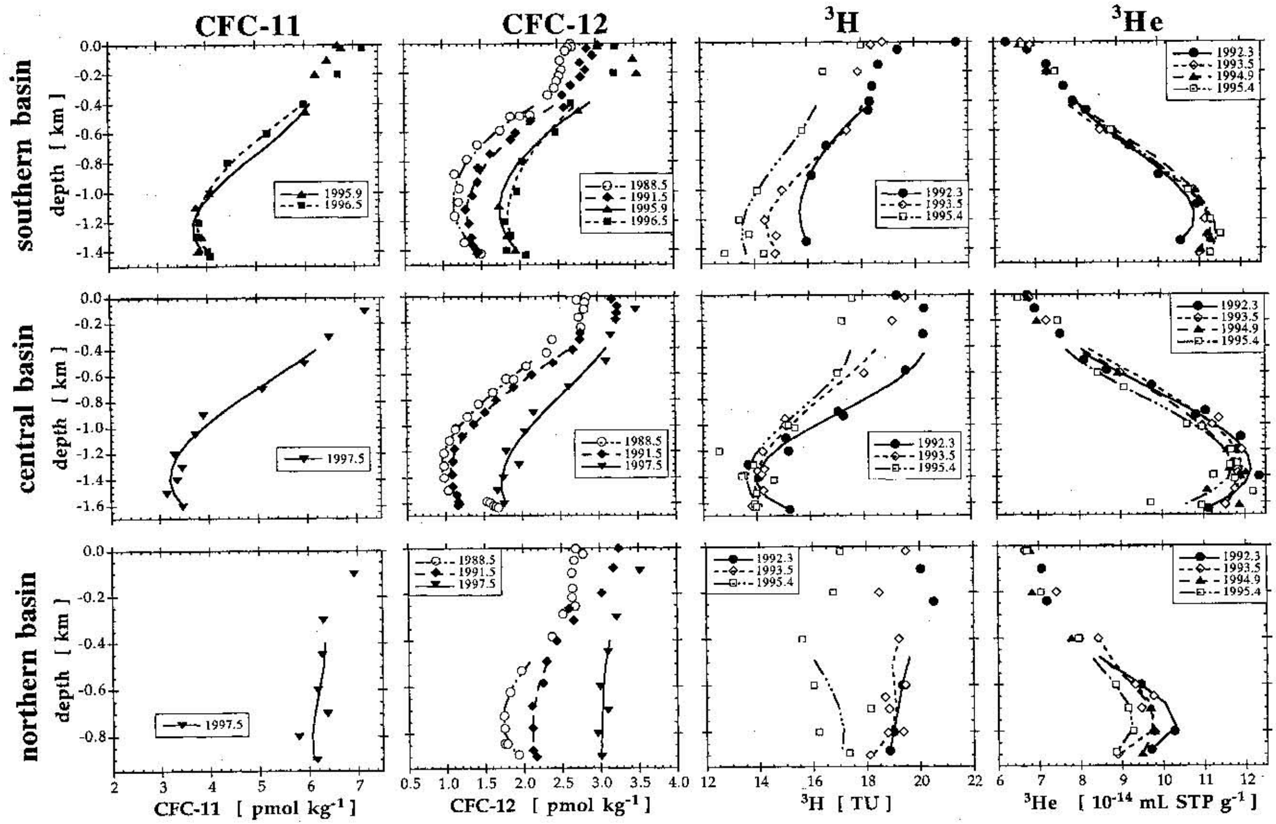


**Figure 2.** Vertical concentration profiles of the tracers CFC-11, CFC-12, $^3$H, and $^3$He in the three basins of Lake Baikal observed between 1988 and 1995 (see explanation of symbols). CFC-12 data from 1988 and 1991 are taken from *Killworth et al.* [1996]. Third-order polynomials (lines) have been fitted to the individual data points below 400 m water depth. CFC-12 data below 1500 m from the 1988 profile in C were excluded from the fit. The sampling time is represented by including a decimal fraction of the year.

measured and predicted concentrations, respectively) is ~0.03 for each tracer, implying a good agreement between data and cubic fits (see also Figure 2). Except for the CFC-12 profile taken in the central basin in 1988 all measurements below 400 m of the corresponding profiles are used to construct the smoothed profiles. The data points below 1500 m of the mentioned profile were omitted since they seem to be influenced by the transient effect of a density plume propagating along the lake bottom to the deepest part of the basin.

Oxygen profiles from the three basins are shown in Figure 5 (symbols). Again third-order polynomials are fitted to the data below 400 m (solid lines in Figure 5). Because in the following model calculation dissolved oxygen is assumed to be at steady state, for each basin, measured values from different years were combined into one single steady state profile. The differences between the steady state profiles and the measurements have a mean of $2x10^{-3}$ $mgO_2$ $l^{-1}$ and have a standard deviation of 0.14 $mgO_2$ $l^{-1}$. The former value indicates that steady state profiles and measurements are not shifted systematically, and the latter value gives the absolute error between steady state profiles and measurements. Therefore we consider the steady state profiles to represent the data within an error of ~±1%.

## 3. Model Description

The S and C basins of Lake Baikal are separated by Buguldeika Ridge and the C and N basins by Academician Ridge (Figure 1). At these sills, horizontal interbasin exchange below 400 m depth is prohibited. The model depicts each basin by a surface mixed layer (SML) extending from 0 to 400 m depth and by a deepwater column (DWC) below 400 m (see Figure 1). These depth ranges were chosen to allow an easy comparison to the study of *Killworth et al.* [1996], who considered in each basin a deepwater region below 400 m. The model assumes the SML to be well mixed. This may overestimate vertical transport in the SML because seasonal mixing reaches down to ~250 m water depth as is indicated by the location of the mesothermal temperature maximum. In addition to the seasonal mixing of the top 250 m by surface cooling and warming the depth range down to 400 m is strongly affected by thermal bar mixing at least in the central basin [*Shimaraev et al.* 1993, *Peeters et al.*, 1996]. Although the SML is not always completely mixed in the course of the year, the assumption of a fully mixed SML appears to be reasonable to describe the dynamics of the mean annual tracer concentration. To test the sensititvity of the model with respect to the depth of the SML we also employed a model in which the SMLs extend from 0 to 300 m depth and DWCs cover the depth range below 300 m.

Each SML is coupled to the atmosphere via gas exchange and evaporation, to the catchment via river inflow and outflow, and to the SML of adjacent basins by diffusive and advective horizontal transport. The SML and DWC of each basin are coupled by vertical diffusion and advection. Because of the sills between the basins, the model does not consider horizontal exchange between the DWCs of adjacent basins. The transport through the bottom 90 m of the narrow strait between Cape Svyatoi Nos and Bol'shoi

**Table 1.** Characteristic Properties of Lake Baikal

| | South Basin | Central Basin | North Basin |
|---|---|---|---|
| Surface area, $km^2$ | 7381 | 10469 | 13621 |
| Area of cross section in 400 m depth, $km^2$ | 5461 | 7484 | 9971 |
| Volume of surface-water region (0-400 m depth), $km^3$ | 2452 | 3446 | 4585 |
| Volume of deep-water region below 400 m depth, $km^3$ | 3776 | 5497 | 3259 |
| Maximum depth, m | 1430 | 1630 | 920 |
| Mean annual wind velocity, m $s^{-1}$ | 4.0 | 4.2 | 3.0 |
| Total annual evaporation, $km^3$ $yr^{-1}$ | 3.24 | 4.61 | 5.99 |
| Total annual precipitation, $km^3$ $yr^{-1}$ | 2.89 | 4.12 | 5.35 |
| Total annual river inflow, $km^3$ $yr^{-1}$ | 20.8 | 26.3 | 14.1 |
| Total annual outflow, $km^3$ $yr^{-1}$ | 59.8 | 39.3 | 13.5 |
| Mean of surface water temperature from May to December, °C | 5.8 | 5.6 | 4.9 |
| Mean of air temperature from May to December, °C | 3.1 | 2.1 | 0.9 |
| Date of freeze up from long-term observations | Jan. 10 | Jan. 5 | Jan. 2 |
| Date of open up from long-term observations | May 4 | May 11 | May 18 |
| Period of ice coverage (long-term mean), days | 114 | 126 | 136 |

Data are from *Shimaraev et al.* [1994].

Ushkanii Island (maximum depth 490 m) is assumed to have a negligible effect on the tracer concentrations in the C and N basins. Data collected in the strait during two expeditions did not indicate deepwater formation at the strait.

The morphometric, hydrological, and meteorological parameters such as cross-section $A(z)$ and volume $V(z)$ for each basin, river inflow and outflow, evaporation, precipitation, air and surface water temperature, duration of ice cover, and wind speed are taken from *Shimaraev et al.* [1994] (Table 1). For all these parameters, annual mean values are used since the focus of the model lies on the description of tracer concentrations over several years or even decades. No attempt is made to resolve seasonal effects.

### 3.1. Modeling Vertical Transport in the DWCs of Lake Baikal

Because the individual processes leading to convection from the surface down to the deepwater regions of Lake Baikal are still not very well understood and the database is small, it is necessary to reduce the description of vertical transport to a few parameters. For simplicity it is assumed that density currents along the lake bottom transport water with the signature of the SML to different layers of the DWC. This process called nonlocal mixing has been described by *Imboden* [1981]. The depth-dependent addition of water to the DWC causes upwelling in the open water column with depth dependent vertical velocity $v(z)$. Except for these boundary currents each basin is assumed to be horizontally mixed. Thus the vertical concentration profile in the DWL is described by the following one-dimensional diffusion-advection equation:

$$A\frac{\partial C}{\partial t}+\frac{\partial (C\,v\,A)}{\partial z}=\frac{\partial}{\partial z}\left(K_z\,A\,\frac{\partial C}{\partial z}\right)+C_s\,\frac{\partial (v\,A)}{\partial z}, \qquad (1)$$

where $A$ is the basin cross section, $C$ is the concentration of the substance of interest, $t$ is time, $K_z$ is the coefficient of vertical diffusion, $C_s$ is the concentration in the corresponding SML, and $v$ is the upwelling velocity. The vertical coordinate $z$ increases in the upward direction, and $z = 0$ at the water surface. The terms on the left-hand side of (1) describe the in situ concentration change with time and the effect of vertical advection (upwelling). On the right-hand side we have the effect from vertical turbulent diffusion and from the addition of water at depth $z$ because of density currents.

Assuming that the water is incompressible and the bottom of the lake is impermeable for water, mass conservation relates at depth $z$ the input of water from the SML per unit depth and time, $q(z)$ and the upwelling velocity $v(z)$:

$$q(z)=\frac{\partial (v\,A)}{\partial z}, \qquad (2)$$

Integration of (2) with $v(z_{md}) = 0$ yields

$$Q(z)=\int_{z_{md}}^{z} q(z')\,dz'=v(z)\,A(z)\,, \qquad (3)$$

where $Q$ is the total vertical water flux rate and $z_{md}$ is $z$ at the maximum depth of the basin.

Rewriting (1) yields for a conservative tracer

$$A\frac{\partial C}{\partial t}+v\,A\frac{\partial C}{\partial z}=\frac{\partial}{\partial z}\left(K_z\,A\,\frac{\partial C}{\partial z}\right)+C_s\,\frac{\partial (v\,A)}{\partial z}-C\,\frac{\partial (v\,A)}{\partial z}. \qquad (4)$$

Dividing by $A$ and rearranging leads to

$$\frac{\partial C}{\partial t}=K_z\,\frac{\partial^2 C}{\partial z^2}-v_{\text{pseudo}}\,\frac{\partial C}{\partial z}+rate_{\text{pseudo}}\,(C-C_s), \qquad (5)$$

with

$$v_{\text{pseudo}}=v-\frac{1}{A}\frac{\partial A}{\partial z}\,K_z-\frac{\partial K_z}{\partial z}$$

$$r_{\text{pseudo}}=-\frac{\partial v}{\partial z}-v\frac{1}{A}\frac{\partial A}{\partial z}\,, \qquad (6)$$

where $r_{\text{pseudo}}$ can be interpreted as a pseudo rate. These equations can be used to characterize the temporal development of the concentration of any conservative tracer such as CFC-11 or CFC-12. It is assumed that because of the large oxygen concentration, decomposition of the CFCs is negligible in Lake Baikal.

Not all of the measured tracers are conservative. For instance, tritium decays into $^3He$. Thus, for these tracers the term $\lambda C_{\text{tritium}}$ has to be added (for $^3He$) or subtracted (for tritium) on the right-hand side of (5). The radioactive decay constant of tritium is $\lambda = 0.05576$ $yr^{-1}$ [*Unterweger et al.*, 1980]. Furthermore, for the case of dissolved molecular oxygen a consumptiom term $\Psi_{VO2}(z)$ has to be subtracted. According to *Livingstone and Imboden* [1996] it is reasonable to define this term as the sum of a volume sink for oxygen in the open water, $\Psi_{VO2}$ [$mgO_2$ $m^{-3}$ $yr^{-1}$], and an areal sink for oxygen, $\Psi_{AO2}$ [$mgO_2$ $m^{-2}$ $yr^{-1}$], because of oxygen consumption within or near the sediments:

$$\Psi_{O2}(z)=\Psi_{VO2}+\Psi_{AO2}\,\frac{dA(z)}{dV(z)}. \qquad (7)$$

Note that the ratio between sediment surface and water volume per unit depth increment, $dA(z)/dV(z)$, strongly increases at the deepest part of the basin.

The described model for vertical transport in the DWC is similar to that by *Killworth et al.* [1996]. However, *Killworth et al.* assumed that the temporal increase of the CFC-12 concentrations is equal at all depths. This greatly restricts the structure of future profiles once the initial profiles are fixed. They also assumed that oxygen depletion is independent of depth.

In order to solve (5), appropriate boundary conditions are needed. At $z = z_{md}$ the boundary conditions are defined by the assumption that the sediments are impermeable for both water and tracers. At the upper boundary ($z = -0.4$ km) the DWC is directly coupled to the SML by vertical turbulent diffusion. In addition, the two regions of each basin are coupled via the nonlocal exchange of water from the SML directly into different depths of the DWC. Thus the model has to be completed by expressions describing the mass balance in the SML in each of the basins.

### 3.2. Modeling the SMLs

Tracer concentrations in the SML (0-400 m depth) are influenced by gas exchange with the atmosphere, by river inflow and outflow, by precipitation and evaporation and by interbasin horizontal exchange. Furthermore, the SML is coupled to the underlying DWC by vertical turbulent diffusion and by nonlocal exchange described by (2) and (3) and the source $C_S\, \partial(v\,A)/\partial z$ in (1). The SML of each basin is modeled as a completely mixed box. The corresponding equations are discussed by *Peeters et al.* [1997].

Gas exchange is described by a linear model [see *Schwarzenbach et al.*, 1993]. For the case of tritium the effect of water vapor exchange is included according to the model by *Herczeg and Imboden* [1988]. Both processes are assumed to be completely supressed during the period of ice coverage which typically lasts from early January to middle May. The tritium concentration in precipitation and rivers in the drainage area of Lake Baikal, $^3H_{\mathrm{Baikal}}$, is related to tritium in precipitation measured at Ottawa (Canada). On the basis of time series of tritium in precipitation measured within the International Atomic Energy Agency (IAEA) network [IAEA, 1975] at several Russian stations (Irkutsk, Yakutsk, Novosibirsk, Omsk, Habarovsk and Enisejsk) over shorter time periods, *Peeters et al.* [1997] have derived the following empirical relation: $^3H_{\mathrm{Baikal}} = 23 + 1.5\ ^3H_{\mathrm{Ottawa}}$. Amospheric concentrations of the CFCs as a function of time are taken from *Elkins et al.* [1993], *Katz et al.* [1995], and *Montzka et al.* [1996].

The equilibrium concentrations of the CFCs in water in contact with the atmosphere are determined from the equilibrium solubilities of *Warner and Weiss* [1985]. The corresponding equilibrium concentration for $^3$He is calculated from the solubilities of helium by *Benson and Krause* [1976] and the $^3$He/$^4$He fractionation determined by *Clarke et al.* [1976] and *Benson and Krause* [1980].

Horizontal water exchange between adjacent basins consists of two components: (1) the net flow determined from the water balance of each basin (see Table 1) and (2) the horizontal turbulent exchange. The latter is estimated from the relation between length scale $L$ and horizontal diffusivity $K_h$ given by *Okubo* [1971]. With $L$ = 100 km the horizontal diffusion coefficient is $K_h \approx 100$ m$^2$ s$^{-1}$. This corresponds to a horizontal exchange velocity ($K_h / L$) of ~30 km yr$^{-1}$. As it turns out, the outcome of the model is not very sensitive to the exact value of the intrabasin exchange velocity.

### 3.3. Initial Conditions

Because of their recent dynamic history, man-made tracers (tritium and its decay product $^3$He, CFC-11, and CFC-12) are not at steady state in Lake Baikal. The above model has to be analyzed in terms of its time-dependent behavior. Therefore a starting time and the corresponding initial conditions have to be defined to initiate the model runs. In order to minimize the influence of the initial conditions the model is started in the year 1900.

At the beginning of the century, concentrations of the man-made CFCs were still zero since their production only began in the late 1930s [*Busenberg and Plummer*, 1992]. Initial tritium and $^3$He concentrations are assumed to be in equilibrium with the external input of tritium and $^3$He into the lake. Prior to the atmospheric nuclear bomb testing in the 1950s the prebomb concentration of $^3$H in precipitation was ~5 TU [*Clarke et al.*, 1969]. The natural atmospheric concentration of $^3$He is $7.25\times10^{-12}$ ppV [*Clarke et al.*, 1976]. On the basis of these concentrations the initial $^3$H concentration is determined to be 0.5 TU in the SML and to decrease with increasing depth down to ~0.2 TU. In the SML the initial $^3$He concentration is ~$6.20\times10^{-14}$ mL STP g$^{-1}$ and increases up to $6.25\times10^{-14}$ mL STP g$^{-1}$ at maximum depth. The intial concentrations do not differ signifcantly between basins.

### 3.4. Estimation of Transport Parameters

Since the DWCs of the three basins are indirectly coupled by vertical exchange with the corresponding SMLs and by interbasin exchange between neighboring SMLs, the complexity of the numerical model is fairly large. Although information on four different tracers is available, the transport mechanisms are still too complex to allow a meaningful parameter identification unless additional assumptions are made to reduce the number of free parameters. Therefore the description of vertical transport within each basin is reduced to five parameters. One free parameter is the vertical turbulent diffusivity, and the other four describe the advective transport of water from the SML to the DWC, $q(z)$. It is assumed that $q(z)$ is positive and can be described as a third-order polynomial

$$q(z) = a_0 + a_1 z + a_2 z^2 + a_3 z^3 \qquad (8)$$

in m$^2$ yr$^{-1}$ defined by the coefficients $a_0$, $a_1$, $a_2$ and $a_3$. Since the tracer profiles, that are used to calibrate the model parameters can be described quite well by third-order polynomials (Figure 2), (8) seems appropriate to resolve the spatial variations contained in the available data set.

The estimation of the transport parameters is conducted in two different ways. On the one hand, the individual tracer data are used to optimize the model outcome. On the other hand, the third-order polynomials (see section 2) are used to generate a new smoothed data set with a regular vertical resolution of 100 m in each profile, which approximately corresponds to the average vertical resolution of the original measurements. Inverse modeling based on the constructed data set allows a direct comparison with the inverse model of *Killworth et al.* [1996] in which transport parameters were derived from smoothed data (cubic spline) instead of the original measurements. The parameter fitting is made by minimizing $\chi^2$, i.e., the sum of the weighted squared differences between modelled and measured (or constructed) tracer concentrations. Summation includes all transient tracers; the deviations between modeled and measured (or constructed) concentrations was weighted according to the uncertainty of the measurements (see section 2).

### 3.5. Oxygen Depletion in the DWCs

Once the transport parameters are obtained, the model can be employed to estimate oxygen depletion in Lake Baikal according to the model given by (7). The following assumptions are made: (1) In the SML annual mean values of oxygen depletion and oxygen production are equal. Thus net oxygen depletion in the SML is zero. (2) In both, the SML and the DWC concentrations of dissolved oxygen are at steady state. In the DWC of each basin the steady state oxygen concentration profile is approximated by a third-order polynomial, that has been fitted to the measured dissolved oxygen concentrations (lines and symbols in Figure 5). (3) Areal and volumetric oxygen depletion rates are equal in all three basins.

Oxygen profiles are calculated from model runs, that begin in 1900 with an initial profile that is equal to the steady state $O_2$ profile. The areal and volumetric oxygen consumption rates are varied until the modeled profiles in 1994 minimize the sum of the weighted squared differences between model predictions and the steady state profiles (assumed errors for dissolved oxygen are 1%; see section 2).

### 3.6. Model Implementation

The numerical model is implemented by using the modeling tool Aquasim [*Reichert*, 1994a]. Aquasim has been designed as a flexible tool for describing various transformation rates of chemical substances as well as the coupling of compartments with different transport characteristics. Aquasim solves the system of partial differential equations describing a model by spatial discretization, which leads to a system of ordinary differential equations. This system of ordinary differential equations is integrated using the implicit algorithm Dassl [*Brenan et al.*, 1989]. The vertical structure in the DWC of Lake Baikal is depicted by horizontal layers each 50 m thick. The SML (top 400 m) of each basin is a mixed compartment, that is considered to be characterized by a very large diffusion coefficient providing vertical homogeneity of the tracer concentrations. Parameter estimation was performed by using both a simplex and a secant algorithm. Errors of the parameters were obtained from the variance-covariance matrix of the estimated parameters (for details, see *Reichert* [1994b]).

## 4. Results

### 4.1. Transport Parameters Obtained from Modeling Transient Tracers

The result of the dynamic modeling of the transient tracers (CFC-11, CFC-12, $^3$H, and $^3$He) in the three basins of Lake Baikal is shown in Figure 3 (solid and dashed lines depict simulations; symbols depict measurements). The transport paramaters are determined so as to minimize the weighted squared deviations of the modeled values from the third-order polynomial profiles shown in Figure 2 (lines). According to a $\chi^2$ test the model is consistent with the smoothed data set. The probability $P(\chi^2 \geq \chi^2_{mod})$, that a $\chi^2$-distributed variable with the same degree of freedom is equal or larger than the minimum $\chi^2$ obtained by the model, $\chi^2_{mod}$, is ~85%. If the volume-weighted mean concentrations from the SML are included in the goal function of the fitting procedure, $P(\chi^2 \geq \chi^2_{mod})$ is ~40%. Fitting the model to the original set of individual measurements (symbols in Figure 2) drops $P(\chi^2 \geq \chi^2_{mod})$ to <1%. However, the bottom five CFC-12 data points measured in the central basin below 1500 m in 1988 contribute ~50% of the total $\chi^2$. If the model is run without these five data points, $P(\chi^2 \geq \chi^2_{mod})$ reaches 25%. Note that these five data points also pose a serious problem to the model by *Killworth et al.* [1996] because their model is based on the assumption of constant growth of CFC-12 at all depths. This assumption is in clear contradiction to the strong decrease of the CFC-12 concentration from 1988 to 1991 in the near-bottom region of the central basin.

Although the results from the $\chi^2$ test vary depending on the goal function employed, in all cases the fitted transport parameters are essentially the same. Vertical turbulent diffusion coefficients $K_z$ obtained from the optimization procedure are $4.6 \times 10^{-4}$ m$^2$ s$^{-1}$ ± 10% in S, $6.3 \times 10^{-4}$ m$^2$ s$^{-1}$ ± 10% in C, and $1.7 \times 10^{-4}$ m$^2$ s$^{-1}$ ± 25% in N. Note that the error estimates give the uncertainty of the model parameters within the framework of the model. They neither include uncertainties in the model formulation (e.g., in the design of the SML and DWC or in the description of the transport processes), nor do they include errors in the input functions (e.g., in the atmospheric CFC concentrations or in the $^3$H concentration in precipitation). $K_z$ in N seems to be significantly smaller than in the other basins. A possible explanation may be found in the fact that the mean wind velocity in N is ~25% smaller than in the southern and central basins (Table 1). Since turbulent kinetic energy input varies with the third power of wind speed, turbulent diffusive mixing in N could well be reduced by a factor of 2 or more. In addition, the duration of ice cover is longest in the N which, on an annual basis, also reduces the mean input of kinetic energy.

Model predictions for the nonlocal transport of water from the SML into the DWC, $q(z)$, the total vertical advective transport of water in the DWC, $Q(z)$, and the corresponding upwelling velocity $v(z)$ are depicted in Figure 4. While the size and the vertical structure of advective transport look similar in C and S, the situation is different in N. First, both total vertical volume flux and upwelling velocity are significantly larger in N. At a depth of 400 m the vertical transport of water is 110 km$^3$ yr$^{-1}$ in S, 70 km$^3$ yr$^{-1}$ in C and 290 km$^3$ yr$^{-1}$ in N. Second, in S and C the vertical distribution of advective input into the DWC, $q(z)$, reaches a minimum at depths between 1000 and 1200 m. In contrast, in N, $q(z)$ steadily decreases with depth. Note that the maximum depth of N is only ~900 m.

Although experimental evidence for the long-term temporal evolution of the discussed tracers is missing, it is instructive to look at the reconstruction of the tracer profiles for earlier times (Figure 3, dotted lines). Both the profiles of CFC-11 and CFC-12 steadily increase with time from 1960 to the present. The relative shape of the profiles remains virtually the same. In contrast, tritium concentrations peak around 1970 at levels well above the concentrations in recent years. Since then, decreasing tritium concentrations in precipitation and inlets combined with radioactive decay and vertical exchange cause the tritium concentrations in the lake to drop and make the vertical distribution more homogeneous. Finally, because of $^3$H decay, $^3$He concentrations have increased in the DWC since the 1950s. They reach a maximum around 1990 in S and C and around 1980 in N. Since then, $^3$He concentrations have dropped because of

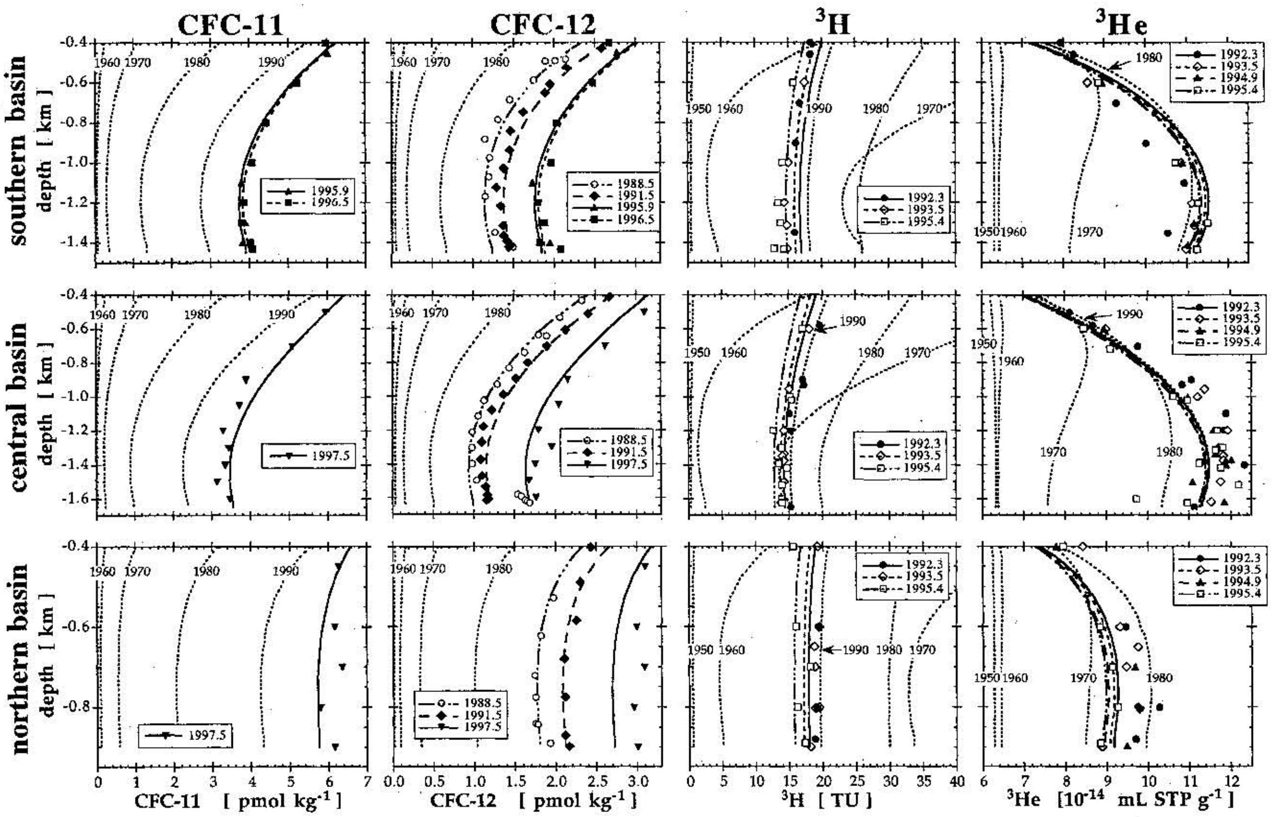


**Figure 3.** Comparison of tracer profiles from the numerical model (dashed, dash-dotted, and solid lines) with measurements (symbols). In order to show the temporal evolution of the tracer profiles, modeled profiles for the year 1990 and earlier are included (dotted lines), although no data exist for this time period. Note that each model simulation starts in 1900. The sampling time is represented by including a decimal fraction of the year.

vertical exchange with the surface layer from which $^3$He is lost to the atmosphere via gas exchange.

In the SML, CFC concentrations are undersaturated compared to the atmospheric equilibrium concentrations at annual mean surface temperatures. According to the model the degree of undersaturation is about the same for CFC-11 and CFC-12 and decreases with time. The model suggests an undersaturation for CFC-12 of ~30% in 1970, 16% in 1988, and 7% in 1996. The seasonally resolved model by *Peeters et al.* [1996] predicted similar undersaturations for the annual mean CFC-12 concentrations in the top 300 m compared to the annual mean equilibrium concentration. For CFC-12 measurements from 1988, *Weiss et al.* [1991] report an undersaturation of 18% in the top 250 m compared to the atmospheric equilibrium concentration at 3.7°C. An undersaturation of the CFCs and a decrease in this undersaturation with time in the SML is reasonable because limited gas exchange, in combination with the transport of water low in CFCs from the DWC to the SML, should result in undersaturation in the SML [see *Weiss et al.*, 1991]. Because in the DWC the ratio of the CFC concentration to the atmospheric equilibrium concentration increases with time because of exchange with the SML, the influence of deepwater exchange on the degree of undersaturation in the SML decreases with time. In addition, the increase of the atmospheric CFC concentration has slowed down considerably within the past 10 years.

A variant of the above model was employed in which the SMLs extended only down to 300 m instead of to 400 m. The results obtained were very similar to those obtained by the original model; values of the turbulent diffusivity were $6.4\times10^{-4}$ m$^2$ s$^{-1}$ in S, $7.3\times10^{-4}$ m$^2$ s$^{-1}$ in C and $2.9\times10^{-4}$ m$^2$ s$^{-1}$ in N. These are of the same order as the values discussed above and support the earlier finding that the turbulent diffusivity in N is significantly smaller than in the other basins. The advective transport $q(z)$ in the three basins determined by inverse modeling is essentially the same as above. Consequently, the model outcome with respect to the characteristic features of vertical transport in Lake Baikal is independent of whether the SMLs are assumed to extend from 0 to 300 m or from 0 to 400 m.

The original transport model can be run not only from the start of this century, but it can also be employed to model shorter time periods. For instance, each tracer can be modeled by using the earliest measured profile as an initial condition. The volume-weighted means of the concentrations measured in the SML can be employed as upper boundary conditions. The deviation between model calculation and measured profiles at later dates serves as a measure for the performance of the model. Although the resulting values for $\chi^2$ are even smaller than for model runs that begin at the turn of the century, the variation in the measured profiles during the short periods for which observations exist (4 years for $^3$H and $^3$He and ~ 9 years for CFC-12) are so small that the model prediction depends only weakly on the magnitudes of the transport parameters. Thus inverse modeling over short time periods is not a sensitive tool for determining mixing in Lake Baikal.

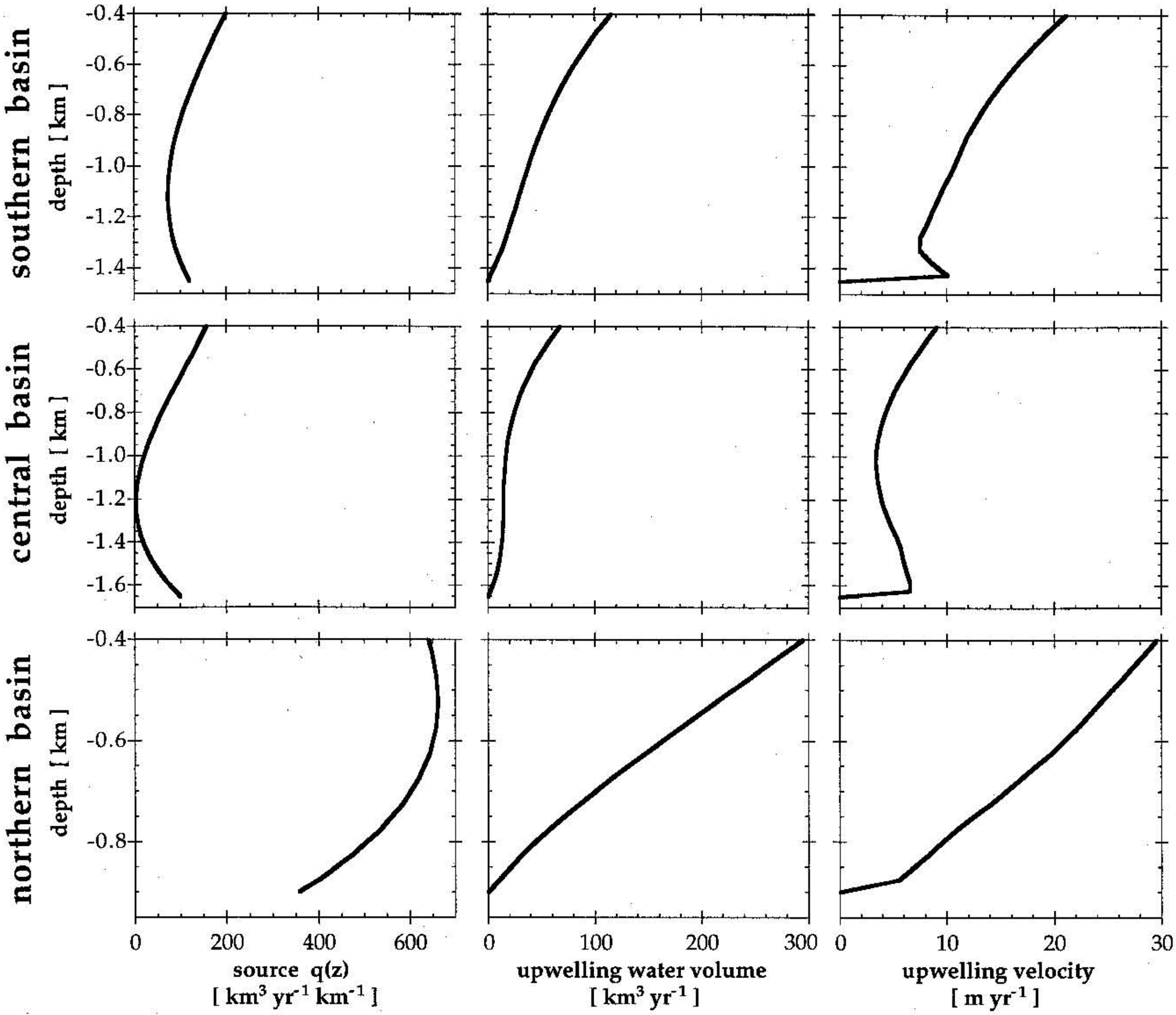


**Figure 4.** Input of water per unit time and depth from the SML into depth $z$ of the DWC, $q(z)$. $Q(z)$ is the corresponding upwelling flux of water (equation (3)), $v(z)$ is the upwelling velocity, and $v(z) = Q(z)/A(z)$, where $A(z)$ is the basin cross-sectional area.

### 4.2. Oxygen Depletion

On the basis of the fitted transport parameters, volumetric and areal oxygen depletion rates in the DWC of the three basins are estimated by inverse modeling of the smoothed profiles (lines in Figure 5): $\Psi_{VO2} = 44 \pm 3$ $mgO_2\ m^{-3}\ yr^{-1}$ and $\Psi_{AO2} = 17000 \pm 3000$ $mgO_2\ m^{-2}\ yr^{-1}$. The $\chi^2$ test indicates a good agreement between model and steady state $O_2$ profiles derived from the measurements ($P(\chi^2 \geq \chi^2_{mod}) = 92\%$). In Figure 6a (model 1: dotted lines), model calculation and measurements of dissolved $O_2$ are compared. Figure 6 also shows total oxygen depletion rates calculated from (7) (Figure 6b; dotted lines). It should be remembered that oxygen is assumed to be at steady state in the DWCs of Lake Baikal. Note that because of the different morphology, each basin has a different total depletion rate, although $\Psi_{VO2}$ and $\Psi_{AO2}$ are assumed to be the same in each basin. Typically, total oxygen depletion rates sharply increase near the bottom, which is the result of the large sediment area-to-volume ratio in this region. For the entire DWC of Lake Baikal the mean oxygen depletion rate is ~100 $mgO_2\ m^{-3}\ yr^{-1}$, and the volume-weighted mean oxygen depletion rate is 75 $mgO_2\ m^{-3}\ yr^{-1}$.

Allowing for individual depletion rates, $\Psi_{VO2}$ and $\Psi_{AO2}$, in each basin (Figure 6; model 2: solid lines) results in an excellent agreement between model prediction and smoothed measurements ($P(\chi^2 \geq \chi^2_{mod}) = 99\%$). However, since both models, one with indiviual and one with equal $\Psi_{VO2}$ and $\Psi_{AO2}$ in each basin, are consistent with the data according to a $\chi^2$ test, we prefer the latter because it requires only two fit parameters.

We have also tried to model dissolved oxygen by assuming that total oxygen depletion is independent of depth but different in each basin. In this case the model predicts an oxygen depletion rate of 82 $mgO_2\ m^{-3}\ yr^{-1}$ in S, 76 $mgO_2\ m^{-3}\ yr^{-1}$ in C, and 113 $mgO_2\ m^{-3}\ yr^{-1}$ in N (Figure 6, model 3: dashed lines). Yet, according to a $\chi^2$ test the model with depth-independent depletion rates is not consistent with the smoothed measurements. Finally, if oxygen depletion is assumed to be the same in all three basins, the model predicts a mean oxygen depletion of ~80 $mgO_2\ m^{-3}\ yr^{-1}$. Again, the fit is not consistent with the data.

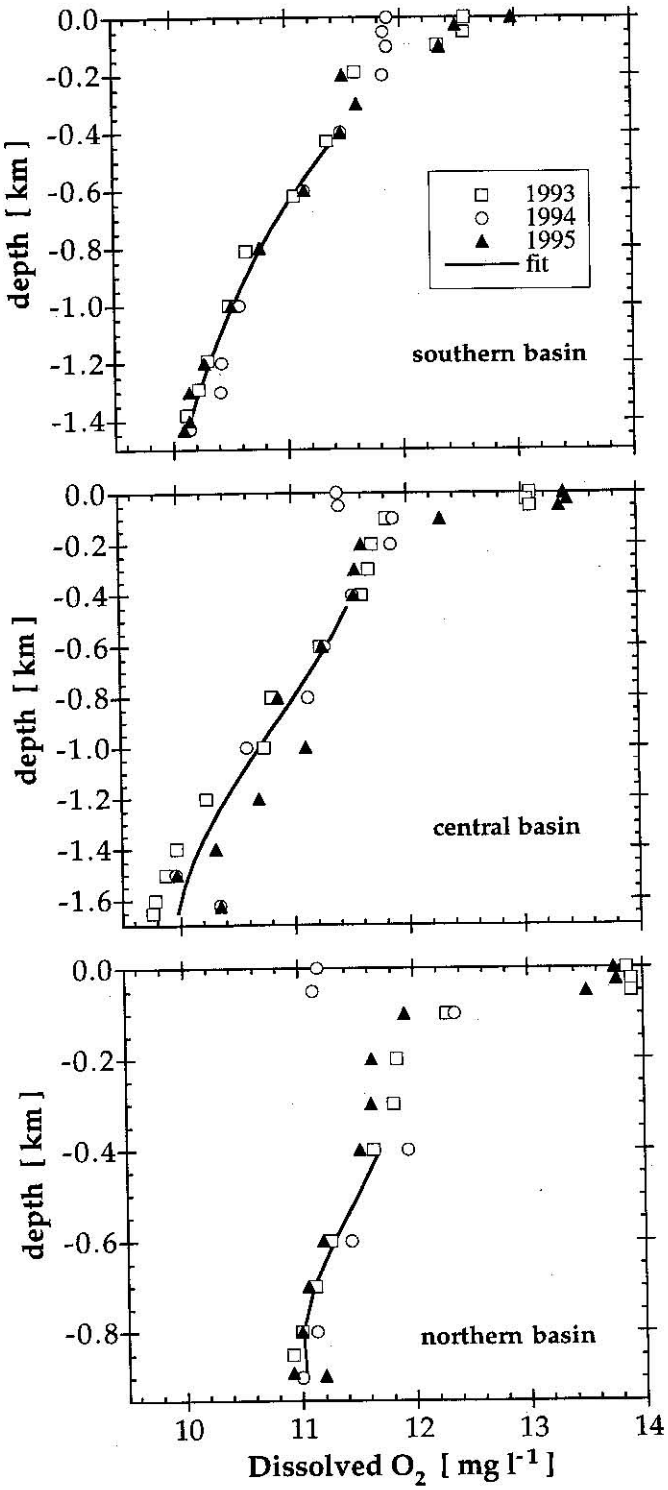


**Figure 5.** Concentrations of dissolved oxygen measured in the years 1993 to 1995 (symbols). Since oxygen is assumed to be at steady state, third-order polynomials (solid lines) are fitted to all data of each basin.

## 5. Discussion and Conclusion

### 5.1. Vertical Turbulent Diffusion and Advective Transport

In the past years several experimental and theoretical attempts have been made to explore the mechanisms of vertical water exchange in Lake Baikal and to quantify these processes [*Weiss et al.*, 1991; *Shimaraev et al.*, 1993, 1994; *Walker and Watts*, 1995; *Kipfer et al.*, 1996; *Peeters et al.*, 1996; *Killworth et al.*, 1996; *Hohmann et al.*, 1997, 1998]. Because of the particular vertical structure of some paramaters reported by *Weiss et al.* [1991] (e.g., water temperature, concentration of dissolved oxygen, CFC-12, silica, and others), most of these investigations have been guided by the assumption that convection is the dominant process of vertical transport between SML and DWC of Lake Baikal.

In a sense this investigation makes no exception: A model is presented in which vertical exchange is described by two mechnisms, (local) turbulent diffusion and (nonlocal) advection. Yet the outcome of our study significantly differs from the results of other investigations, especially the one by *Killworth et al.* [1996]. The average vertical turbulent diffusion coefficients determined by simultaneous inverse modeling in all three basins of the tracers CFC-11, CFC-12, $^3$H, and $^3$He are $4.6\times10^{-4}$ m$^2$ s$^{-1}$ ± 10% in S, $6.3\times10^{-4}$ m$^2$ s$^{-1}$ ± 10% in C, and $1.7\times10^{-4}$ m$^2$ s$^{-1}$ ± 25% in N. These values are more than 2 orders of magnitude larger than those calculated by *Killworth et al.* [1996] (a few times $10^{-6}$ m$^2$ s$^{-1}$). With vertical turbulent diffusion contributing more to vertical transport our model consequently predicts a smaller contribution of convection than *Killworth et al.* [1996]. Furthermore, we suggest that upwelling is largest in N and smallest in C (290 and 70 km$^3$ yr$^{-1}$, respectively, at 400 m depth), whereas *Killworth et al.* [1996] state the opposite (~400 km$^3$ yr$^{-1}$ in N and 1000 km$^3$ yr$^{-1}$ in C).

There are two important reasons why we think that this investigation is, in fact, making an important and more realistic step toward the understanding of vertical mixing in Lake Baikal. First, our model is more complete than any other investigation before. It combines more data and more tracers, extends over a time period of nearly 100 years, and keeps the number of adjustable paramaters to a minimum (five transport parameters in each basin). As shown above, a model calculation over a time period of just a decade or so is not very sensitive to the actual parameter choice since Lake Baikal is a fairly inert system. By starting the model in the year 1900 with initial conditions derived from independent considerations (no CFCs present and tritium and $^3$He in equilibrium with input at prebomb background level), we are able to show that the chosen transport parameters are consistent with the long-term dynamics of Lake Baikal and that model and data do not diverge during that period. In fact, running our model with $K_z = 5\times10^{-6}$ m$^2$ s$^{-1}$ (a value at the upper limit of those suggested by *Killworth et al.* [1996]) and with the appropriate adjustment of the advective transport parameters does not, according to the $\chi^2$ test, lead to an agreement between data and model predictions.

The second reason why we believe the outcome of our model to be realistic is based on what is known about the physics of lakes and oceans. Vertical turbulent diffusivity is related to the input of turbulent kinetic energy by wind stress and inversely related to the stability (Brunt-Väisälä) frequency $N^2$ [*Gregg*, 1987, *Imboden and Wüest*, 1995]. In the DWC of N, $N^2$ is $<5\times10^{-8}$ s$^{-2}$ [*Peeters et al.*, 1996]. Similar values can be calculated from CTD measurements in the other basins. The absence of a strong thermocline during most parts of the year suggests that kinetic energy can easily penetrate into the deep regions of the lake, primarily via basin-sized waves. Records from current meters moored in S at ~1400 m depth show that horizontal current velocities reach ~10 cm s$^{-1}$ [*Ravens et al.*, 1999]. Given the extremely small vertical stability of the water column such currents inevitably lead to turbulence.

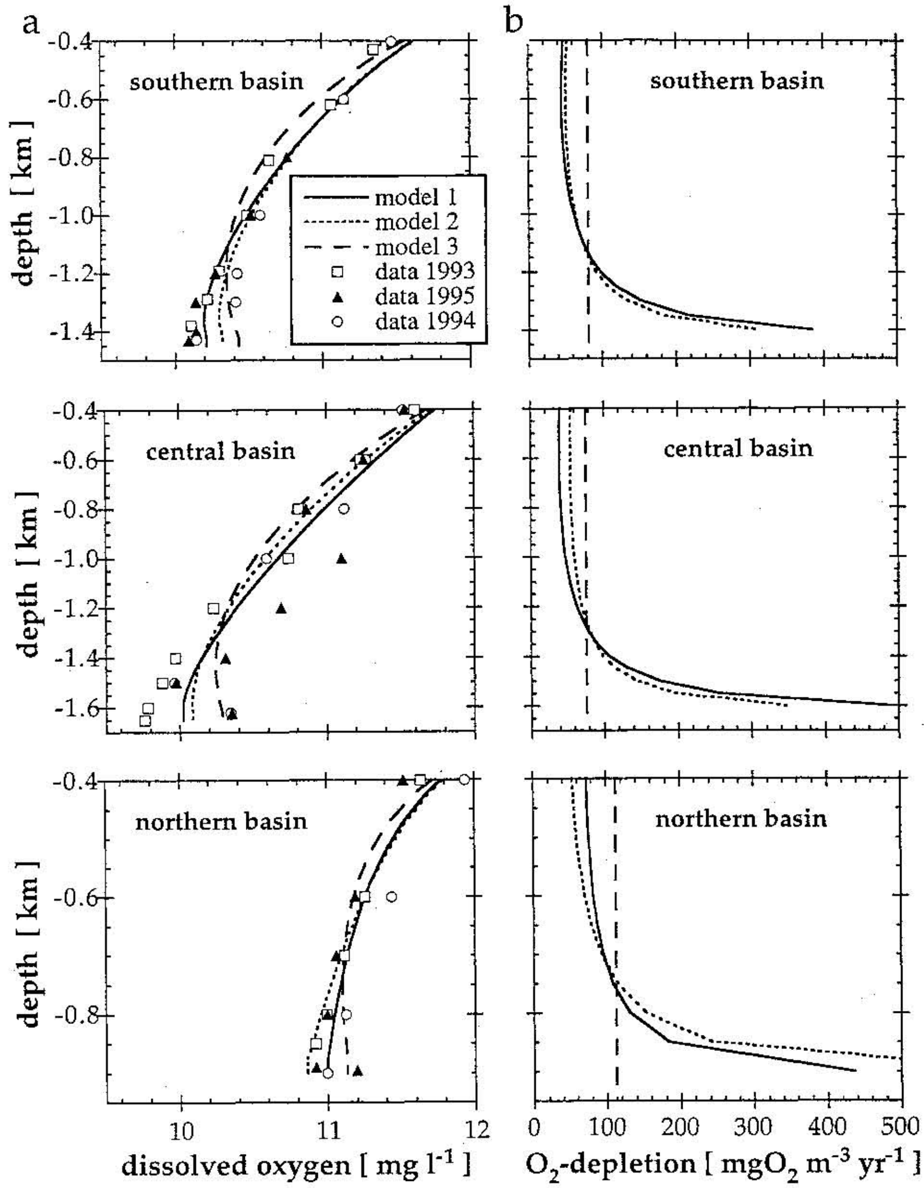


**Figure 6.** (a) Comparison of different model calculations to measured dissolved oxygen concentrations. In model 1, volumetric and areal consumption rates are equal in all basins, in model 2, volumetric and areal oxygen depletion rates differ between basins, and in model 3, total oxygen depletion rate is independent of depth but different in each basin. (b) Total oxygen depletion rates for model 1 to 3.

On the basis of temperature microstructure measurements, *Ravens et al.* [1999] have calculated the vertical turbulent diffusivity between the water surface and 600 m depth in the interior part of of Lake Baikal. In the depth range between 400 and 600 m depth they found $K_z$ to range in the S basin between $1\times10^{-4}$ and $5\times10^{-4}$ m$^2$s$^{-1}$. It has been demonstrated that average basin-wide vertical turbulent diffusion coefficients, as reflected by the long-term behavior of tracers, primarily result from the combined action of strong vertical mixing at the boundaries and subsequent horizontal mixing into the open water column [e.g., *Ledwell et al.*, 1993; *Ledwell and Bratkovich*, 1995; *Wüest et al.*, 1996]. According to *Goudsmit et al.* [1997], in lakes the difference between basin-wide and open water vertical diffusivity can reach an order of magnitude since closed water basins have relatively large boundary-to-open-water ratios. Hence the open water diffusivities derived from temperature microstructure by *Ravens et al.* [1999] are fairly consistent with basin-wide values of ~$5\times10^{-4}$ m$^2$s$^{-1}$ in the S basin determined from the inverse model.

On the basis of temperature measurements made between 1972 and 1988, *Shimaraev et al.* [1994] estimated for the depth range below 400 m vertical exchange rates of $7\times10^{-4}$ m$^2$ s$^{-1}$ in S, $6\times10^{-4}$ m$^2$ s$^{-1}$ in C and $2\times10^{-4}$ m$^2$ s$^{-1}$ in N of Lake Baikal. These exchange rates excellently agree with our estimates for the vertical turbulent diffusion coefficients. Note, however, that the heat budget method employed by *Shimaraev et al.* [1994] only provides a measure of vertical turbulent diffusion coefficients if the effects of advective transport on the heat content below a certain depth can be neglected or corrected for. In addition, in the deepwater region of Lake Baikal the vertical heat flux by turbulent diffusion is very small even at large diffusivities because vertical temperature gradients are extremely small (of the order $2\times10^{-4}$ °C $m^{-1}$ at 600 m depth). Thus the determination of $K_z$ for the deep water of Lake Baikal from changes in the heat content below a certain depth requires very precise temperature measurements over long time periods: for example, the expected heat increase below 600 m due to turbulent heat diffusion at a $K_z$ of $5\times10^{-4}$ $m^2$ $s^{-1}$ corresponds to an increase of the mean temperature below 600 m of only ~0.003°C within half a year.

Size and structure of vertical advection as computed by our model also need some physical explanation. According to our result the advective input of water, *q(z)*, into S and C has a minimum around 1200 m (Figure 4). To explain this feature one can argue that the vertical distribution of *q(z)* reflects the probability of a density plume to become neutrally buoyant at depth *z*. On one hand, it is plausible that this probability decreases with increasing depth since water with characteristics not too different from the SML is likely to sink not very far from the surface. On the other hand, all plumes with densities larger than any water in the water column are "assembled" at the deepest part of the basin. This explains the increase of *q(z)* toward the bottom.

In C, and most likely also in S, density plumes are generated by water from the Selenga River [*Hohmann et al.*, 1997]. The density difference between open water and Selenga water is relatively large. In fact, measurements in the Kukui Canyon off the Selenga delta have shown that density currents flow along the lake bottom to the deepest part of the central basin. The situation is different in N. In the N, convection is mainly driven by density-induced flow across the Academician Ridge of water from C into the N [*Hohmann et al.*, 1997]. The total flow rate across the ridge (which is fairly long) is large, which may explain why advective transport *q(z)* into N is larger than in the other basins. However, the density excess due to the enhanced salinity of the surface water from the C compared to N water is rather small. Thus one can speculate that in contrast to C and S, where salinity and turbidity of the Selenga River significantly contribute to density differences, very cold surface water in N, which would assemble at the basin bottom in C and S, is not able to penetrate through the barrier at the mesothermal maximum. In addition, the bottom plume structure might be obscured by the generally much larger advective fluxes in N at intermediate depths.

It should be mentioned that the discussed model as well as the one by *Killworth et al.* [1996] have a shortcoming; they both neglect entrainment of ambient water into the plumes of water penetrating into the DWC. Adequate information to justify a quantitative treatment of entrainment is missing, since each of the potential processes (formation of river water plumes, interbasin exchange, cabbeling, thermal bar, and thermal baricity) has its own entrainment characteristic, and data on these processes is scarce. Therefore entrainment was not incorporated in the model so far. Yet, we can qualitatively discuss how entrainment may affect the identification of the two transport mechanisms, turbulent diffusion and advection. On one hand, entrainment increases the volume of the sinking density plumes [*q(z)* and thus also *Q(z)*]. On the other hand, entrainment reduces the difference between the tracer concentration in the plume and the ambient water, i.e., decreases the effect the advected water has on the concentration in a given layer of the DWC. Thus, in cases where the influence of entrainment is significant, neglection of entrainment would cause an underestimation of *q(z)* and *Q(z)* and, in turn, an overestimation of transport by vertical turbulent diffusion.

### 5.2. Balance of Dissolved oxygen in the DWCs

The transport parameters determined by inverse modeling of tracer concentrations allow the relative importance of the two major mixing mechanisms on the transport of dissolved oxygen into the DWC to be assessed. According to the model, turbulent diffusion is more important than advection for the transport of dissolved oxygen. Turbulent diffusion contributes ~70% in S, 90% in C, and 40% in N to the net input of oxygen into the DWC below 600 m. This differs from the result by *Killworth et al.* [1996] according to which the coefficient of turbulent diffusion is extremely small and thus convection dominates the oxygen flux. In our model this only applies to the near-bottom regions of the basins where the ratio of *q(z)* to the cross section is large.

According to our model the volumetric sink for oxygen is $44 \pm 3$ $mgO_2$ $m^{-3}$ $yr^{-1}$, and the areal sink for oxygen is $17000 \pm 3000$ $mgO_2$ $m^{-2}$ $yr^{-1}$. The areal sink agrees reasonably well with the oxygen consumption at the water sediment interface determined by *Mizandrontsev* [1990], who gives a range from 4000 to 35000 $mgO_2$ $m^{-2}$ $yr^{-1}$. A comparison of the volumetric sink with the results on oxygen consumption of the investigations of *Shimaraev et al.* [1996], *Killworth et al.* [1996], and *Hohmann et al.* [1998] is not straightforward because they did not distinguish between a volumetric and an areal oxygen sink. Equation (7) implies that overall oxygen depletion significantly increases at the bottom of each basin (Figure 6). By vertical integration of the depth-dependent oxygen depletion rate the volume-weighted mean oxygen depletion in the DWC of Lake Baikal was determined to be 75 $mgO_2$ $m^{-3}$ $yr^{-1}$. This value agrees well with the value for oxygen depletion of 37-100 $mgO_2$ $m^{-3}$ $yr^{-1}$ determined by *Shimaraev et al.* [1996] for the deepwater region of Lake Baikal. As a comparison, the value for oxygen depletion estimated by *Killworth et al.* [1996] ranges from 130 to 180 $mgO_2$ $m^{-3}$ $yr^{-1}$, and that given by *Hohmann et al.* [1998] is ~140 $mgO_2$ $m^{-3}$ $yr^{-1}$.

### 5.3. Transport of Oxygen Under Different Climatic Conditions

An increasing interest in the evolution of the Lake Baikal ecosystem [*Grachev et al.*, 1998] as well as the intensification of agricultural and industrial activities in the catchment area of Lake Baikal provide motivation to speculate on the distribution of oxygen in the past as well as in the future. Recently, a sediment coring program was launched in Lake Baikal to reconstruct climatic conditions in the past. In this context the question arises how changes in climatic conditions and in the catchment affect vertical transport of water, dissolved oxygen concentrations, and other parameters.

According to the outcome of our inverse model, vertical turbulent diffusion significantly contributes to the vertical transport of oxygen. Most likely, a moderate drop or rise in the mean surface water temperature does not significantly alter the stability of the DWC provided that a mesothermal temperature maximum develops regularly. Thus, as long as the mean wind conditions and the duration of the ice coverage are not altered, the average turbulent diffusivity is expected to remain at present levels. In contrast, if the climate gets substantially warmer and temperatures in the deep water of Lake Baikal rise significantly, the density gradients and thus stability in the deep water might increase and reduce the role of turbulent diffusion.

Transport by convection depends on the occurence of cold water at the lake surface. Such conditions are met as long as the surface water regularly freezes and thaws. As described before, convection of cold surface water to the DWC is hindered by the potential barrier between 200 and 500 m. Salinity, concentration of suspended particles, or strong winds to produce a thermobaric instability are needed to overcome the barrier. Thus the chemical and mineralogical composition of the major inlets (mainly the Selenga River) and the occurrence of strong winds are the most important factors in determining the future convective deepwater mixing in Lake Baikal.

Salinity and suspended particles of the inflows most likely will increase in the future because of land use and erosion in the catchment area of Lake Baikal. Thus, the probability for convection to occur will be larger than today, leading to a larger input of oxygen into the DWC. However, input of salts and suspended particles into the DWC might also increase the stability of the water column in the DWC. In this case one would expect a reduction of vertical turbulent diffusion and thus of the downward transport of oxygen.

In conclusion, we suggest that the transport of oxygen would not be severely affected by moderate climatic temperature changes. However, changes in the hydrological regime, the salinity, and the load of suspended particles in the Selenga and other rivers might eventually cause the transport of oxygen by turbulent diffusion to become smaller. In the most extreme case, concentrations of dissolved oxygen in the DWC of Lake Baikal may decrease. Although at this point this remains fairly speculative, careful observations are recommended to increase our understanding of the relevant mechanisms of deepwater mixing in Lake Baikal and to provide early information in case changes might occur.

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R. Kipfer and M. Hofer, Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), Dübendorf, Switzerland. (kipfer@eawag.ch; hofer@eawag.ch)

F. Peeters, Department of Physics, Swiss Federal Institute for Environmental Science and Technology (EAWAG), CH-8600 Dübendorf, Switzerland. (fpeeters@eawag.ch)

D. M. Imboden, Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), CH-8092 Zürich, Switzerland. (imboden@novatlantis.ch)

V. M. Domysheva, Limonlogical Institute, Siberian Division of the Russian Academy of Sciences, Irkutsk 66403, Russia (domysheva@lin.irkutsk.ru)



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# Correction to "Vertical turbulent diffusion and upwelling in Lake Baikal estimated by inverse modeling of transient tracers" by F. Peeters, R. Kipfer, M. Hofer, D. M. Imboden, and V. M. Domysheva

In the paper "Vertical turbulent diffusion and upwelling in Lake Baikal estimated by inverse modeling of transient tracers" by F. Peeters, R. Kipfer, M. Hofer, D. M. Imboden, and V. M. Domysheva (*Journal of Geophysical Research, 105*(C2), 3451-3464, 2000), the author list was inadvertently printed out of order. The correct author list with affiliations is shown below.

F. Peeters, R. Kipfer, and M. Hofer
Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), Zürich, Switzerland

Department of Environmental Physics, Swiss Federal Institute of Environmental Science and Technology (EAWAG) Dübendorf, Switzerland

D. M. Imboden
Department of Environmental Physics, Swiss Federal Institute of Technology (ETH), Zürich, Switzerland

V. M. Domysheva
Limonlogical Institute, Siberian Division of the Russian Academy of Sciences, Irkutsk

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