

Deep-water renewal in Lake Issyk-Kul

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[1] The deep-water renewal rates of Lake Issyk-Kul are studied using the time-dependent anthropogenically produced tracers sulfur hexafluoride (SF₆), chlorofluorocarbons (CFCs) and tritium-helium-3. SF₆ and the CFCs are used to calibrate a mixing model from which the vertical age distribution is calculated and found to be comparable to the SF₆ apparent ages. Based on this model, the mean age of the water below 100 m depth is 6.1 yrs. The mean oxygen consumption rate for the same depth range is 6.4 μmol kg⁻¹ yr⁻¹ and the mean remineralization rates for nitrate, phosphate and silicate are 0.53, 0.003 and 0.67 μmol kg⁻¹ yr⁻¹, respectively. *INDEX TERMS*: 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4845 Oceanography: Biological and Chemical: Nutrients and nutrient cycling; 4239 Oceanography: General: Limnology; 1845 Hydrology: Limnology

1. Introduction

[2] Brackish Lake Issyk-Kul is located in Kyrgyzstan in the midst of central Asia's Tien Shan mountains at an altitude of 1606 m. With a maximum depth of ~665 m it is the world's fifth deepest lake. The local winters are relatively mild such that freezing only occurs in the shallow bays. Then the open surface water temperature is 4–4.5°C, always observed to be slightly higher than the deep water temperatures [Romanovsky, 1990] and well above the temperature of maximum density of 2.6°C for this lake's 6.06 g kg⁻¹ saline waters [Vollmer *et al.*, 2002].

[3] The lake does not overturn fully during the mixing season and the deep water is ventilated by local convection in addition to vertical turbulent mixing. The primary objective of our study is to determine the timescales of deep-water renewal. Knowledge of these renewal rates is important for understanding the cycling of nutrients and the fate of contaminants potentially entering the lake and for furthering the interpretation of past mixing conditions as recorded in the sediments. Measurements of the time-dependent atmospheric trace gas sulfur hexafluoride SF₆ have recently been used to estimate ventilation rates of young water masses in the oceans [Law and Watson, 2001; Tanhua *et al.*, 2002]. Here we present the first such measurements in a deep lake. With its rapid almost-linear rise in the atmosphere SF₆ has become a powerful time-dependent tracer to study recently-ventilated water masses. The chlorofluorocarbons (CFCs), which have been used successfully in previous deep lake ventilation studies [e.g. Weiss *et al.*, 1991], have become of limited use on shorter timescales because their atmospheric concentrations have stabilized or even begun to decline in recent years [Prinn *et al.*, 2000].

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2. Methods

[4] During a survey in September 2000 aboard R/V *Moltur* we collected a vertical profile of water samples near the deepest location close to the center of the lake (Station 8, 42° 23.6' N, 77° 13.1' E) using 10-liter Niskin bottles with standard precautions to avoid trace gas contamination [e.g. Bullister and Weiss, 1988]. Here we describe the results of our measurements of SF₆, CFCs, tritium (³H), helium (He) isotopes, neon (Ne), dissolved oxygen (O₂) and nutrients. The results of our major ion and conductivity-temperature-depth (CTD) measurements are reported elsewhere [Vollmer *et al.*, 2002].

[5] Samples for measurements of SF₆ and the CFCs CCl₂F₂ (CFC-12), CCl₃F (CFC-11) and CCl₂FCClF₂ (CFC-113) were flame-sealed into custom-made, 300 ml Pyrex glass ampoules [Busenberg and Plummer, 1992; Vollmer and Weiss, 2002] and analyzed at the Scripps Institution of Oceanography (SIO) using electron-capture detector gas chromatography [Vollmer and Weiss, 2002] with precisions (±1 std. dev.) of 1.4% for SF₆, 0.6% for CFC-12, 1.2% for CFC-11, and 0.7% for CFC-113.

[6] Water samples of ~20 g were collected in pinched-off copper tubes for the analysis of the He isotopes and Ne. Samples for ³H were collected in 100 ml glass bottles. These measurements were made at the Lamont-Doherty Earth Observatory (L-DEO) Noble Gas Laboratory [Ludin *et al.*, 1998] with typical precisions (±1 std. dev.) of 0.2–0.5% for He isotope ratios, 1% for the ⁴Ne and Ne concentrations and 2–3% for ³H.

[7] Dissolved O₂ was analyzed on board by a computer-controlled O₂ titrator using ultraviolet photometric endpoint detection with a precision of 0.06% (±1 std. dev.). The poisoned and unfiltered samples for nutrients were analyzed at SIO on a Skalar San-Plus continuous flow colorimetric analyzer [WOCE protocol, Gordon *et al.*, 1992] with absolute analytical precisions (±1 std. dev.) of 0.02 μmol kg⁻¹ for nitrate (NO₃⁻), 0.006 μmol kg⁻¹ for phosphate (PO₄³⁻), 0.001 μmol kg⁻¹ for nitrite (NO₂⁻) and 0.1 μmol kg⁻¹ for silicate.

3. Results

[8] The concentrations of the transient tracers and the nutrients are plotted in Figure 1 and listed in the electronically available Table 1¹. The SF₆ and CFC profiles show concentration maxima in the depth range of 40–90 m, below which they decrease to their lowest values near the bottom of the lake. Near the surface, these tracers show pronounced decreases in concentration reflecting their lower solubilities in the seasonal warm surface layer [Bullister *et al.*, 2002; Warner and Weiss, 1985; Bu and Warner, 1995]. At 3 m depth, the water is supersaturated with these gases by about 1 to 2% with respect to their northern hemisphere tropospheric background concentrations [Maiss and Brenninkmeijer, 1998; Walker *et al.*, 2000] as calculated from their solubilities as functions of temperature and salinity at a mean lake-surface barometric pressure of 0.818 atm.

¹Supporting material is available via Web browser or via Anonymous FTP from ftp://ftp.agu.org, directory "apend" (Username = "anonymous", Password = "guest"); subdirectories in the ftp site are arranged by paper number. Information on searching and submitting electronic supplements is found at http://www.agu.org/pubs/essup_about.html.

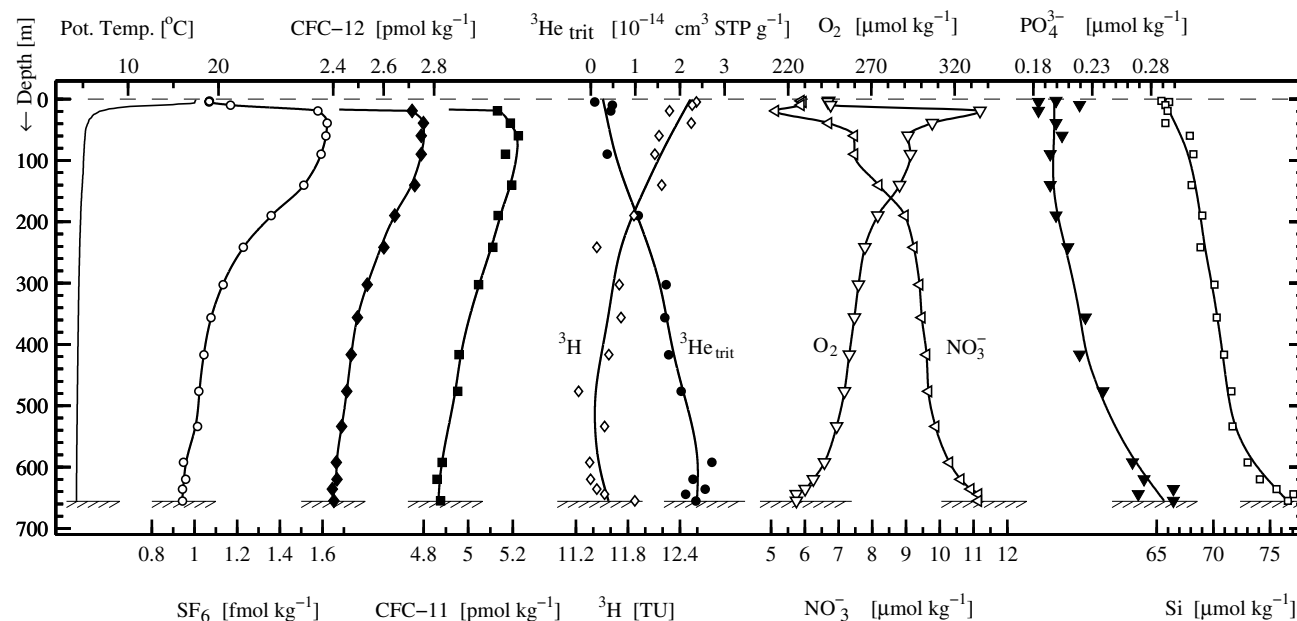


Figure 1. Selected profiles for Station 8 ($42^{\circ}23.6' \text{ N}$, $77^{\circ}13.1' \text{ E}$) occupied on Sept 22, 2000 on Lake Issyk-Kul. The near-surface samples for the CFCs are omitted to better show the deep-water structure.

[9] The profiles of dissolved O_2 and the nutrients are modulated by the lake's internal processes of biological productivity near the surface and the sinking and subsequent oxidation of organic matter at depth resulting in the consumption of deep-water dissolved O_2 . The dissolved O_2 concentration decreases steadily with increasing depth and reaches its minimum, corresponding to a saturation of 71% [Weiss, 1970], near the bottom of the lake. The concentrations of the nutrients increase with increasing depth. Below a depth of 100 m the mean slope of NO_3^- versus PO_4^{3-} in molar units is ~ 21 and that of O_2 versus PO_4^{3-} is ~ -480 .

4. Apparent Ages

[10] The age of a water sample is commonly taken as the mean time elapsed since its components have last been at the surface and thus been in contact with the atmosphere. For SF_6 and the CFCs, we reference our observations to their atmospheric concentration histories. We specifically define the *apparent age* of a water sample as the sampling date minus the date at which the surface water concentration was equal to that of the measured water parcel. This calculation is based on the solubilities of the tracers at the samples' potential temperatures and salinities [Bullister et al., 2002; Warner and Weiss, 1985; Bu and Warner, 1995] and on the assumptions on the degrees of saturation of the water for these gases with respect to their atmospheric concentrations. These saturation indices are generally variable with time because in most cases equilibration with the atmosphere is incomplete and dependent, among other aspects, on the rate of change of the atmospheric concentration of each tracer. In the absence of any knowledge of such saturation indices over the past decades we choose them as constants and derive them from the following observations.

[11] We interpret the 40–60 m depth range, where highest concentrations from our fitted SF_6 and CFC data are observed, as a remnant of the past winter's mixed layer. Because this portion of the water column is strongly stratified due to the seasonal thermocline, we assume that its tracer concentrations have changed insignificantly between our sampling time and the past winter.

Using an estimated winter mixed layer temperature of 4.5°C [Romanovsky, 1990], and based on the atmospheric histories of these tracers [Walker et al., 2000; Maiss and Breninkmeijer, 1998], we calculate saturation indices of 93% for SF_6 , 91% for CFC-12, 90% for CFC-11 and 85% for CFC-113. By making these assumptions, the mixed layer has a zero age for the winter preceding our observations.

[12] The apparent ages for SF_6 increase with increasing depth to a maximum of ~ 9 yrs (Figure 2). Because this tracer's atmospheric growth has been nearly linear in the recent past [Maiss et al., 1996], its apparent age is expected to be a good approximation of the *true age* of the mixture—the mean age of the water parcels contributing to the sample. The apparent ages of the CFCs are greater than those of SF_6 . This is because in recent years the CFCs have exhibited decreased growth rates. Under these circumstances interpolating a mean age for a mean CFC concentration inevitably leads to a bias to older water.

[13] The ^3H - ^3He ages [Torgersen et al., 1977; Jenkins and Clarke, 1976; Schlosser, 1992] were determined using tritium and noble gas measurements. He and Ne saturations [Weiss, 1971] were used to determine the amounts of ^3He and ^4He in the water that are related to equilibration with the atmosphere and to excess air by bubble injection, respectively. The ^3He results were then corrected for the fraction of terrigenic helium of hydrothermal origin based on the residual excess ^4He using the standard crustal $^3\text{He}/^4\text{He}$ ratio of 2×10^{-8} . In order to reference the ^3H - ^3He ages to the same winter mixed layer age as for SF_6 and the CFCs, the tritogenic ^3He concentrations ($^3\text{He}_{\text{trit}}$) were reduced by $0.9 \times 10^{-15} \text{ cm}^3 \text{ STP g}^{-1}$ ('preformed' $^3\text{He}_{\text{trit}}$) to give an age of ~ 8 months for the same 40–60 m depth range during our observations, or a zero age for this layer during the previous winter. The resulting ^3H - ^3He ages are close approximations of the actual mean age as the ^3H distributions in the lake are fairly uniform and are therefore insensitive to mixing effects.

5. Mixing Model

[14] To derive mixing and upwelling rates, a one-dimensional mixing model was used that is similar to those applied to Lake

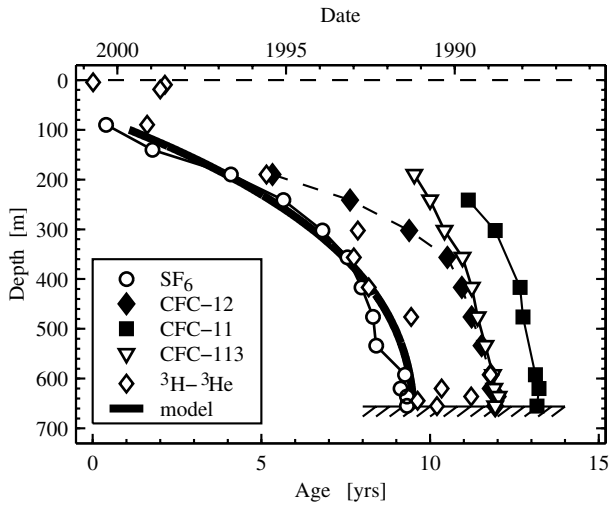


Figure 2. Apparent tracer ages and modeled ages.

Baikal [Killworth *et al.*, 1996; Peeters *et al.*, 2000] and which is described by:

$$A \frac{\partial c}{\partial t} + \frac{\partial(wAc)}{\partial z} = Qc_s + \frac{\partial}{\partial z} \left(K_z A \frac{\partial c}{\partial z} \right) + J \quad (1)$$

where A is the basin area at a given depth z , here chosen positive upward, and t represents time. The concentration of a specific compound in the deep water is given by c and that in the winter mixed layer by c_s . The mixing processes are described by convection, Q per unit depth, modeled as a third-order polynomial with respect to depth, of dense water plumes descending from the winter mixed layer. To preserve mass, this flux is compensated for by lake-wide upwelling with a velocity w . K_z is the vertical turbulent diffusion coefficient which we choose as a constant. Sources are represented by $J > 0$ and sinks by $J < 0$. The density plumes are related to the upwelling velocity by

$$Aw = \int_{z_{bot}}^z Q(z') dz' \quad (2)$$

where z_{bot} denotes the depth of the bottom of the lake.

[15] The mixing parameters Q and K_z are quantified by an inverse calibration technique using the SF₆, CFC-12 and CFC-11 atmospheric concentration histories and their saturation indices derived earlier. This yields a K_z of $3 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, a value that is similar to those found for Lake Baikal [Peeters *et al.*, 2000]. The upwelling across the 100 m isobath is 18 m yr^{-1} . CFC-113 was initially excluded from the inverse model calibration because of suspected degradation in oxygenated water [Roether *et al.*, 2001]. We have reconstructed all 4 tracer profiles using the model and the temporal evolution of the tracers in the troposphere. From this we found that the modeled concentrations deviated from the observations less for CFC-113 ($< +1\%$) than for SF₆ ($+2\%$) or CFC-12 (-2%) and CFC-11 (-2%). This has led us to conclude that CFC-113 degradation in Lake Issyk-Kul is insignificant. The inclusion of this tracer in the model calibration thereby strengthens our confidence in the model results.

[16] Using the model in a forward mode we calculate the true ages which, in contrast to the apparent ages, include subsurface mixing. This modeled true age profile (Figure 2) is comparable to the apparent SF₆ and ³H-³He ages and confirms that the absence of mixing in the calculations of the apparent ages results in an insignificant age bias for these 2 tracers. Based on a hypsographic

integration of the model ages we calculate a mean age of 6.1 yrs for waters below a depth of 100 m.

[17] We also calculated the O₂ utilization rate (OUR) by relating the model ages to the apparent O₂ utilization (AOU), shown in Figure 3 assuming O₂ saturation for 0 yr old water. Values of OUR in the upper parts of the deep water are about $5.6 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$, while below 500 m there is a dramatic increase in the OUR, suggesting a dominant contribution of biological degradation at the sediment-water interface. We have calculated a mean consumption rate of $6.2 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$ for the 100–500 m depth range by hypsographic integration of the slopes of the spline-fitted AOU versus the modeled ages. In this approach rather than in a second one which we employed for the lower part of the water column, the uncertainties in the above assumption on O₂ saturation are insignificant. For the lower part of the water column the first approach leads to large uncertainties caused by the uncertainties in the model age gradients at this depth. For this depth range we calculate a mean consumption rate of $7.8 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$ by dividing the volume-adjusted AOU by the volume-adjusted mean age of this layer. Our mean O₂ consumption rate for the water volume below 100 m depth is $6.4 \pm 0.9 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$, which is higher than that calculated for Lake Baikal ($4.5 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$, Weiss *et al.* [1991]) using similar techniques. This difference may be related to the larger sediment surface relative to lake volume for Lake Issyk-Kul.

[18] The corresponding mean nutrient remineralization rates are calculated similarly and by taking into account the surface water (preformed) concentrations. These are $0.53 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$ for NO₃⁻, $0.003 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$ for PO₄³⁻ and $0.67 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$ for Si. Assuming that the deep-water remineralization is balanced by an upward flux of these nutrients, we calculate yearly upward fluxes across the 100 m isobath of $6.5 \times 10^8 \text{ mol}$ for NO₃⁻, $3.9 \times 10^6 \text{ mol}$ for PO₄³⁻ and $8.3 \times 10^8 \text{ mol}$ for Si. Comparing these fluxes to those from rivers (Table 1, Kadyrov [1986]) and groundwater inflow [Zektser and Bergelson, 1989] shows that for NO₃⁻ the return fluxes to the 100 m surface layer are an order of magnitude larger than those associated with river inflow and 4 orders of magnitude larger than those associated with groundwater inflow. For PO₄³⁻, the deep-water return fluxes are comparable to those of the groundwater and an order of magnitude larger than those of the rivers. For Si the deep water return fluxes are comparable to those of the rivers and 3 orders of magnitude larger than those of the groundwater.

[19] Using the model ages and the terrigenous ⁴He concentrations of the deep water we calculate a mean accumulation rate of $2 \pm 0.3 \times 10^{-10} \text{ cm}^3 \text{ STP g}^{-1} \text{ yr}^{-1}$ for terrigenous ⁴He (Figure 3). From this we estimate a flux of terrigenous ⁴He of $4.8 \pm 0.8 \times 10^{10} \text{ atoms m}^{-2} \text{ s}^{-1}$ into the deep lake. This flux is of similar size than the average flux from the continental crust ($\sim 3 \times 10^{10} \text{ atoms m}^{-2} \text{ s}^{-1}$,

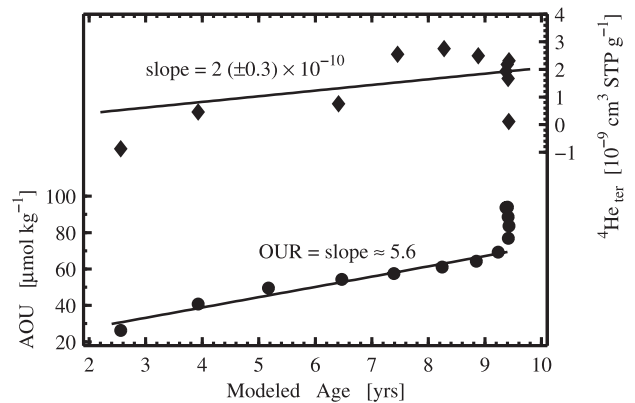


Figure 3. Oxygen utilization rates (OUR) and rates of terrigenous ⁴He accumulation below 100 m in Lake Issyk-Kul.

O'Nions and Oxburgh [1983]) and about one magnitude smaller than that calculated for Lake Baikal [*Hohmann et al.*, 1998]. It is likely that this flux is associated with a flux of natural SF₆ [*Harnisch and Eisenhauer*, 1998] which is present locally as found in our artesian well samples. However, based on the good agreement between the modeled and observed SF₆ and CFC concentrations, we assume that the flux of natural SF₆ into the lake can be ignored. We had planned to estimate this flux using the ratio SF₆ of to He in the well samples and our calculated terrigenous He flux into the lake. Unfortunately we could not pursue this plan because of experimental problems with the He well samples.

[20] **Acknowledgments.** We thank B. Turin for analyzing the tritium and noble gas samples, the W. M. Keck Foundation for the establishment of the L-DEO helium isotope laboratory, A. Grachev for help with the Russian literature and SIO and NASA for support.

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