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Physical and chemical properties of the waters of saline lakes and their importance for deep-water renewal: Lake Issyk-Kul, Kyrgyzstan

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Abstract—The relationships between electrical conductivity, temperature, salinity, and density are studied for brackish Lake Issyk-Kul. These studies are based on a newly determined major ion composition, which for the open lake shows a mean absolute salinity of 6.06 g kg^{-1} . The conductivity-temperature relationship of the lake water was determined experimentally showing that the lake water is about 1.25 times less conductive than seawater diluted to the same absolute salinity as that of the lake water. Based on these results, an algorithm is presented to calculate salinity from in-situ conductivity measurements. Applied to the field data, this shows small but important vertical salinity variations in the lake with a salinity maximum at 200 m and a freshening of the surface water with increasing proximity to the shores. The algorithm we adopt to calculate density agrees well with earlier measurements and shows that at 20°C and 1 atm Lake Issyk-Kul water is about 530 g m^{-3} denser than seawater at the same salinity. The temperature of maximum density at 1 atm is about 0.15°C lower than that for seawater diluted to the same salinity. Despite its small variations, salinity plays an important role, together with temperature changes, in the static stability and in the production of deep-water in this lake. Changes in salinity may have had important consequences on the mixing regime and the fate of inflowing river water over geological time. Uncharged silicic acid is negligible for the stability of the water column except near an $\sim 15 \text{ m}$ thick nepheloid layer observed at the bottom of the deep basin. Copyright © 2002 Elsevier Science Ltd

1. INTRODUCTION

Brackish Lake Issyk-Kul is the world's fifth deepest lake with a maximum depth of about 665 m. It is located in Kyrgyzstan at a present surface elevation of 1606 m in the midst of Central Asia's Tien Shan mountains. The lake is in a semiarid setting with precipitation highest in spring and summer while winters are relatively dry. The large seasonal air temperature variations of Central Asia are reduced near the lake to an annual amplitude in the monthly mean air temperature of about 20°C . As a consequence of the mild winters and warm summers, the open lake water never freezes and the winter mixed layer temperatures are greater than 4°C (Romanovsky, 1990). Past climate changes have caused shifts in the lake's hydrological budget which have led to lake level fluctuations with occasional basin closures (Rasmussen et al., 2000). Due to its basin closure during such a shift about 6900 yrs to 4900 yrs ago, the lake has evolved from a freshwater body to a saline one (Ricketts et al., 2001).

In this paper we discuss the vertical stability of the water column and possible mixing mechanisms and deep-water renewal processes using CTD (conductivity-temperature-depth) data. The understanding of these mechanisms is important in light of concern about increasing human influence on water quality (Savvaitova and Petr, 1992), the role of groundwater inflow (Zektser and Diordiev, 1979; Zektser and Bergelson,

1989), the use of the lake's sediment to study Central Asia's past climate and the potential effects that current climate changes may have on the mixing regime of this lake. Using anthropogenic time-dependent tracers it has been shown that the deep water of Lake Issyk-Kul is currently being renewed by mixing with the surface on the time scale of about a decade (Vollmer et al., 2002; Hofer et al., 2002).

For saline lakes the seawater-specific equations of state relating salinity and density (Unesco, 1981a, 1981b) to CTD measurements can generally not be applied because the major ion compositions in such lakes often differ from that of seawater. For Lake Issyk-Kul we have used measurements of the major ionic composition and of the dependence of conductivity on temperature together with an adapted conductivity-pressure relationship (Finger, 2002) to parametrize the relationships between the equations of state for this lake and those of seawater. We then apply these results to our Lake Issyk-Kul CTD data, thereby allowing us to build a framework for studying stability and mixing in the lake. This procedure can be adapted to other saline lakes if the ionic composition and the dependence of conductivity on temperature and pressure are known.

2. METHODS

In September 2000, a geochemical survey was conducted on Lake Issyk-Kul and some of the surrounding rivers and artesian wells (Fig. 1). During the shipboard work on R/V *Moltur*, which lasted from 22–24 September 2000, CTD data were collected along a north–south transect across the deep basin. On the second day, bottle samples together with a CTD profile were collected from Station 8 ($42^\circ 23.6' \text{ N}$, $77^\circ 13.1' \text{ E}$)

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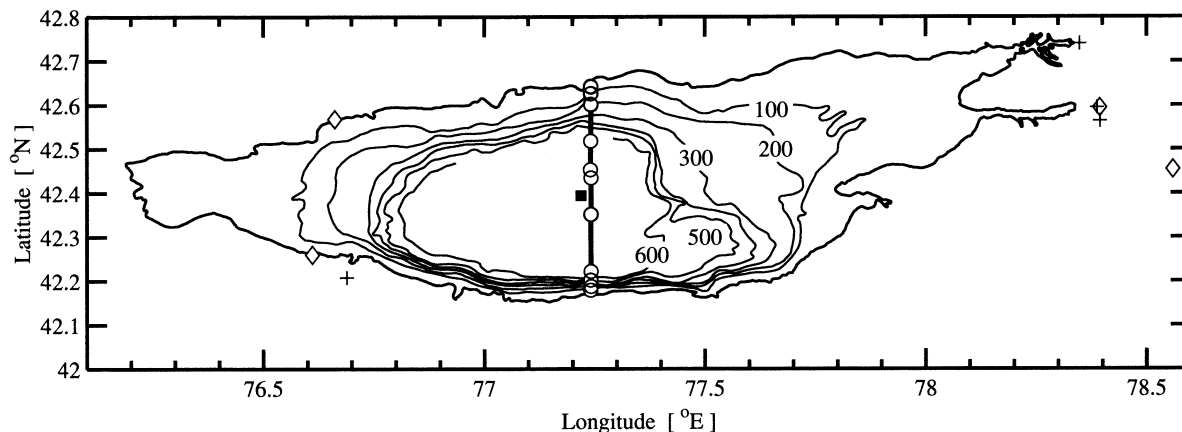


Fig. 1. Bathymetry of Lake Issyk-Kul showing the CTD-transect (○), Station 8 (■), rivers (+), and artesian wells (◇) sampling locations during the September 2000 geochemical survey. Isobaths are given in 100 m steps. The bathymetry data are provided by the Large Lake Observatory, University of Minnesota, Duluth. The 600 m isobath is not closed due to lacking data.

near the maximum depth of the lake. Results of our analyses of major ions and electrical conductivity ratios for these samples are presented here, while the results for nutrients, oxygen and a suite of time-dependent tracers are discussed separately (Vollmer et al., 2002). Vertical pressure, temperature, and conductivity profiles were collected using an *EG&G Smart-CTD* fitted with a *SeaTech* transmissometer. At most stations, pressure and temperature data were also collected using a *Seabird* SBE-39 self-contained recorder attached to the *Smart-CTD* frame.

Both pressure sensors were calibrated at the Scripps Institution of Oceanography (SIO) over the depth range of the lake and at various temperatures, and found to be well within the accuracy needed for this survey. The dynamic temperature compensations of the pressure sensors were tested with the help of a laboratory temperature “shock test” and the data were corrected accordingly. This was important for the SBE-39’s pressure sensor with an incorporated compensation time constant for its temperature response which poorly matched the response time of the sensor. After standard data processing and before our correction, this resulted in large errors in pressure of about 30 dbar for a significant time after rapid temperature changes, specifically in the lake’s sharp thermocline. The remaining uncertainty of about ± 3 dbar after data processing results in reduced but still significant uncertainties in the water temperatures assigned to the bottle samples taken from the thermocline.

The temperature sensors were checked against an SIO standard thermometer at several temperatures before the expedition and calibrated over the full temperature range of 4°C to 19°C after the expedition. Their accuracies are estimated $\pm 0.003^\circ\text{C}$ for the *Smart-CTD* and $\pm 0.001^\circ\text{C}$ for the SBE-39. In the isothermal deep water of the lake the SBE-39 temperature was typically 0.003°C higher than the *Smart-CTD* temperature sensor, an agreement which lies within expected uncertainties. Unless stated otherwise, temperatures are reported on the International Temperature Scale ITS-90. For some of the profiles, the conductivity sensor’s output locked at a certain value over considerable pressure and temperature changes, thereby creating artificial salinity excursions. These conductivities were refitted in a subjective manner by smoothing the conductivity profiles over these ranges.

Bottle samples were collected mainly from Station 8 using five 10-liter Niskin bottles fitted with epoxy-coated stainless steel springs and vacuum-baked Buna-N o-rings. For the bottle casts, the SBE-39 was attached to the wire near the hydrographic weight allowing us to reconstruct the water pressure and temperature at which the bottles closed.

Samples for major ions were collected in 125 ml high-density polyethylene bottles and analyzed ~ 4 months later at Oregon State University (Qiu, 2002). For these measurements the analytical uncertainties reported here are based on the relative standard deviation of replicate analyses at the reported concentration levels. Titration alka-

linities were determined by a potentiometric method with an uncertainty of less than 2% (Edmond, 1970). Major anions, Cl^- and SO_4^{2-} were measured by ion chromatography to an estimated precision of $\pm 3\%$. Cation contents (Ca^{2+} , Mg^{2+} , Na^+ , K^+) of the samples were measured by flame atomic absorption spectrophotometry (FAAS) using established methods (Perkin Elmer standard conditions for FAAS). A high purity CsCl solution was added to all samples and standards (to 1000 ppm) in order to buffer the flame ionization potential. In the case of Ca, a high purity LaCl_3 solution was also added to samples and standards (to 1000 ppm) to improve sensitivity in the presence of oxide-forming elements such as P and Al. Standards were purchased commercially and freshwater reference materials (National Institute of Standards and Technology SRMs 1643c and 1643d) were analyzed as checks on accuracy. Analytical errors are estimated to be 3%, 2%, 4%, and 1% or better for Ca^{2+} , Mg^{2+} , Na^+ , and K^+ , respectively. Results for the standard reference materials were within these estimated errors of the certified values and so no adjustment for systematic offsets were made to the cation data. The Ba concentrations were determined ($\pm 3\%$) by isotope dilution inductively coupled plasma quadrupole mass spectrometry as described previously (Falkner et al., 1994).

During a first measurement series, these ion samples were not filtered. During a second measurement series ~ 6 months later well and river waters were filtered through 0.45 micron Nucleopore membranes and reanalyzed for all species except barium, resulting in similar measurement precisions and accuracies compared to the first analysis. The reanalysis revealed strong precipitation of calcite in river samples as evidenced by lower Ca^{2+} and alkalinity concentrations by 15%–30% and 25%–29%, respectively. The alkalinity of some of the unfiltered lake samples was also rechecked and found lower by 2%–5% after storage for 6 months. Presuming that all the alkalinity change is due to calcium carbonate precipitation, an upper limit change of salinity of 0.015 g kg^{-1} is calculated for the lake samples.

Electrical conductivity ratios were measured ~ 2 months after collection at SIO on a *Guildline Autosol 8400A* salinometer using standard techniques. After standardization of the instrument at a practical salinity of 35 using an *Ocean Scientific International* IAPSO seawater standard, the instrument was checked at a salinity of 10 using a diluted seawater standard from the same provider. This showed that the salinometer measurements needed to be corrected downward. This correction was extrapolated linearly to Issyk-Kul water sample conductivity ratios and resulted in a mean correction factor of 0.9989 for the measured ratios of these samples. The overall precision for these measurements could not be determined as no duplicate samples were collected but it is assumed not to exceed significantly the analytical precision for our laboratory measurements. The instrument did not drift during the measurements and from measurements of three low-salinity standard samples we estimate an analytical precision, expressed as a practical salinity, of $\sim 2 \times 10^{-4}$, which is comparable to that found by

Table 1. Lake Issyk-Kul major ion data for Station 8, September 23, 2000. Units are [mmol kg⁻¹] unless stated otherwise. Data marked with * are excluded from averaging. Data at the depth marked # are averaged from duplicate samples. The volume-based concentrations of Cl⁻ and SO₄²⁻ were converted to weight-based concentrations using a density of 1.003 kg L⁻¹. For the calculation of the charge imbalance and the ionic salinity, we have approximated the titration alkalinity by HCO₃⁻ and CO₃²⁻ at molar fractions of 0.88 and 0.12, respectively (Qiu, 2002). The charge balance gives the relative excess of positive charge and is calculated as the difference between the total positive and the total negative charges, divided by the mean of the two and multiplied by 100. The salinity is the sum of the mass concentrations of all major ions. Separately collected samples were used for the measurement of the conductivity ratio at 24°C on a salinometer. The samples denoted § were also used for the laboratory conductivity measurements (see text).

Depth [m]	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	Cl ⁻	SO ₄ ²⁻	Alkalinity	Charge imbal. [%]	Salinity [g kg ⁻¹]	Ba ²⁺ [nmol kg ⁻¹]	Conductivity ratio
2.7	3.01	13.0	67.3	1.80	43.6	21.9	5.70	8.3	6.01	324	0.160 947
4.5	3.11	12.7	65.3	1.80	43.7	22.3	5.76	4.8	6.01	322	0.160 972
9.6	3.08	12.7	65.5	1.81	43.8	21.9	5.82	5.7	5.98	308	0.161 007
19.1	3.06	12.7	66.8	1.80	45.1	21.9	5.70	5.7	6.05	292	0.161 327
38.9	3.04	12.8	65.7	1.80	42.4	21.5	5.71	8.5	5.89	307	0.161 327
59.6	3.17	13.1	66.9	1.80	45.0	22.6	5.88	5.2	6.14	316	0.161 337
89.9	3.10	13.0	65.7	1.78	45.0	21.9	5.72	5.3	6.03	319	0.161 357
140.1	3.06	13.0	66.5	1.80	44.1	22.1	5.76	6.6	6.04	292	0.161 417
190.0	3.00	13.1	67.7	1.80	45.6	23.0	5.81	4.3	6.21	321	0.161 467
241.4	3.09	13.0	67.4	1.80	44.4	22.3	5.79	6.7	6.09	305	0.161 467
302.6	3.14	13.1	67.4	1.80	44.0	22.6	5.78	6.8	6.11	308	0.161 442
356.3	3.10	12.9	65.8	1.80	44.6	22.4	5.81	4.6	6.07	313	0.161 417
416.8	3.09	12.9	63.8*	1.80	45.2	22.9	5.81	*	*	310	0.161 397
476.3	3.11	13.1	65.8	1.79	45.0	23.0	5.86	3.2	6.15	320	0.161 382
533.7	3.12	13.1	64.8	1.80	45.3	22.9	5.78	2.3	6.12	293	§0.161 367
592.4	3.08	13.2	66.2	1.80	44.4	22.2	5.82	6.1	6.06	297	§0.161 367
619.9	3.03	13.0	66.2	1.80	44.3	22.3	5.83	5.4	6.06	306	§0.161 367
635.9	3.05	12.7	66.1	1.80	48.6	20.9	5.82	3.3	6.07	303	§0.161 367
644.3	3.07	12.9	67.2	1.82	44.4	22.4	5.75	6.1	6.09	300	0.161 372
#655.1	3.09	12.6	65.6	1.80	44.4	22.3	5.73	4.3	6.04	304	0.161 372
Average	3.08	12.9	66.3	1.80	44.6	22.3	5.78	5.4	6.06	308	0.161 324
Std. dev.	0.04	0.2	0.8	0.01	1.2	0.5	0.05	1.6	0.07	10	

Mantyla (1987). Common errors due to systematic offsets caused by discontinuities between suppression dial settings are absent for our measured samples as these covered a very small conductivity range and were therefore measured on the same suppression-range switch setting. The small observed vertical variability found for the Station 8 profile can therefore be assumed not to be an instrumental artifact.

The accuracy of our conductivity measurements is of lesser importance for this study. However it is crucial for future similar studies which may be carried out to determine the rates of salinification for this lake. The accuracy is estimated based on two possibly related uncertainties. The first is the above-mentioned discontinuity error. These discontinuities were measured for the suppression range settings covering the range between the sample and the low-salinity standard measurements and found to cumulate to a practical salinity of only ~4

× 10⁻⁴. The second uncertainty stems from the calibration of the salinometer for low salinities using a linear correction. We conducted some independent experiments on a different salinometer using an additional standard of salinity 30 and found salinity uncertainties of ~6 × 10⁻⁴ associated with the assumption of a linear correction. Thus our estimated overall accuracy is ~1 × 10⁻³ expressed as practical salinity.

3. MAJOR IONS

The results for the major ions are given in Table 1 for the Station 8 and in Table 2 for the rivers and artesian wells. In the lake, the molar cation concentrations are dominated by Na⁺

Table 2. Major ion data from September 2000 for rivers and artesian wells surrounding Lake Issyk-Kul. Units are [μmol kg⁻¹] unless stated otherwise. All results are averages of duplicate samples except for those marked with a ¶. The volume-based concentrations of Cl⁻ and SO₄²⁻ were converted to a weight-based concentration using a density of 1.000 kg L⁻¹. For the charge imbalance and salinity calculations see Table 1.

Sampling location	T [°C]	Ca ²⁺		Mg ²⁺		Na ⁺		K ⁺		Cl ⁻		SO ₄ ²⁻		Alkalinity	Charge imbal. [%]	Salinity [mg kg ⁻¹]	Ba ²⁺ [nmol kg ⁻¹]
<i>Rivers</i>																	
Dzhergalan [42° 35.7' N, 78° 23.2' E]	11.1	1010	193	307	33	161	232	2050	2.6	193	238						
Kara-Su [42° 33.8' N, 78° 23.7' E]	14.0	¶1240	391	419	115	140	312	3510	-11.8	299	410						
Ak-Terek [42° 12.5' N, 76° 41.3' E]	12.3	1310	459	400	79	182	342	3290	-3.7	295	277						
Tyup [42° 44.4' N, 78° 20.9' E]	14.2	1400	298	424	34	219	337	2990	-0.7	277	642						
<i>Artesian wells</i>																	
Tamchy [42° 34.1' N, 76° 39.7' E]	44.0	12	0	6670	15	1720	1600	1200	9.2	434	212						
¶Teplokluhenka [42° 27.3' N, 78° 33.6' E]	58.5	13	0	4880	45	2320	17	1160	33.9	261	10						
Bar Bulak [42° 15.9' N, 76° 36.7' E]	46.1	8630	66	57800	280	53300	7910	460	8.1	4362	306						
Dzhergalan Spa [42° 35.6' N, 78° 23.6' E]	38.9	447	64	13800	39	10900	774	1890	3.6	902	1781						

Table 3. Comparison of major ion data for Lake Issyk-Kul. Units are [mmol kg⁻¹] unless stated otherwise. Kadyrov (1986) reported Na⁺ and K⁺ as a sum which we have separated into the two components using the Na⁺/K⁺ ratio found in our study. By definition, these data are charge balanced but averaging data results in a charge imbalance. For the conversion from alkalinity to weight-based concentration, Kadyrov (1986) assumed HCO₃⁻ as the only component. We have adjusted this conversion using the same speciation as assumed for our data. The volume-based concentrations from the near-shore samples by Lyons et al (2001) were converted to weight-based concentrations using a density of 1.003 kg L⁻¹. Although Lyons et al (2001) reported the concentration of HCO₃⁻ in their data table, their description of analytical methods suggests that this number in their table is alkalinity instead. Our calculations of their HCO₃⁻ and CO₃²⁻ is based on this assumption. For the calculations of the salinity, the charge imbalance and HCO₃⁻ and CO₃²⁻, see Table 1.

Source	Sampling date	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	CO ₃ ²⁻	Salinity [g kg ⁻¹]	Charge imbal.[%]	Number of samples
Kadyrov (1986)	1958–1980s	3.0	11.8	62.6	1.7	45.0	21.9	4.1	0.6	5.89	1.7	>500
Lyons et al. (2001)	1996	3.3	11.7	58.2	1.6	40.9	20.3	5.9	0.8	5.62	0.8	3
This study	2000	3.1	12.9	66.3	1.8	44.6	22.3	4.5	0.6	6.06	5.3	20

followed by Mg²⁺ and Ca²⁺—the latter two are relatively more abundant than in seawater. The molar anion concentrations are dominated by Cl⁻ followed by SO₄²⁻, which shows a significantly higher abundance than in seawater and the same charge equivalents as Cl⁻ in the lake. The mean alkalinity is 5.8 mmol kg⁻¹, which is more than twice the value found in the oceans. At Station 8 the concentrations of these constituents are constant and the small vertical salinity gradients which we found from our salinometer measurements are too small to be detected in these major ion data.

We have examined (Qiu, 2002) the chemical speciation and mineral saturation conditions using a chemical equilibrium modeling program (MINEQL+, version 4.5). The calculations were carried out for atmospheric pressure, a pH of 8.7 (Kadyrov, 1986) and over a temperature range from 4°C to 25°C. These calculations show that the lake is strongly supersaturated with respect to Ca²⁺, Mg²⁺ and mixed Ca²⁺-Mg²⁺ phases such as calcite, dolomite, and magnesite. The logarithms of the saturation indices approach 2. Extensive ion-pairing of Ca²⁺ and Mg²⁺ with SO₄²⁻ and to a lesser extent with CO₃²⁻ is predicted to occur. At 25°C roughly 30% of Ca²⁺ is present as CaSO₄(aq) and 25% of Mg²⁺ as MgSO₄(aq) summing to ~30% of the available SO₄²⁻. Carbonic acid species are distributed as 82% and 4% in the form of uncomplexed HCO₃⁻ and CO₃²⁻, respectively, with an additional 6% and 8% present as Ca²⁺, Mg²⁺, and Na⁺ complexes, respectively. Based on our model calculations, the total carbonic acid species are apportioned into HCO₃⁻ and CO₃²⁻ at molar fractions of 88% and 12%, respectively.

We have compared our results with those found in other studies (Table 3). From 1958 to the 1980s, a vast collection of major ion data are given by Kadyrov (1986) with a spatial and temporal coverage that is far more extensive than any other existing data base for this lake. More recently, Lyons et al. (2001) collected a few near-shore samples. All three data sets presented in Table 3 are comparable in magnitude. The concentrations from our study exceed those by Kadyrov except for Cl⁻. However, the uncertainty in our Cl⁻ analysis does not preclude that even this concentration is larger than that reported by Kadyrov (1986). Concentrations found by Lyons et al. (2001) are smaller than those measured by us or by Kadyrov (1986) except for Ca²⁺ and HCO₃⁻, which are higher. Since the samples of Lyons et al. (2001) were all collected near shore, this is consistent with the presence of lateral surface gradients in salinity caused by river and groundwater inflow and up-

welling of more saline water in the center of the lake (Romanovsky, 1990). The fact that their Ca²⁺ and HCO₃⁻ are significantly higher, especially if scaled to less reactive species such as Na⁺ or Cl⁻, is consistent with precipitation of CaCO₃ occurring in the lake.

The open-lake salinity as derived from our major ion measurements is 0.17 g kg⁻¹ (~2.9%) higher than that of Kadyrov (1986). However, it is unclear if the lake has salinified during the past decades, as our measurement uncertainties are similar in magnitude to this salinity difference. If we assume that these ions added by river input (Kadyrov, 1986), are not removed in the lake and that the groundwater loading is of equal magnitude to that of the rivers (Zektser and Bergelson, 1989), this would result in a yearly increase in salinity of ~7 × 10⁻⁴ g kg⁻¹ (0.03%). A lake level decrease of ~6 cm yr⁻¹ (Romanovsky, 1990) would result in an additional yearly salinity increase of ~13 × 10⁻⁴ g kg⁻¹ and consequently the salinity would increase by ~0.3% per decade. Unless the subsurface input of ions is strongly underestimated, such salinity changes are difficult to detect over short time periods. Future accurate salinometer conductivity ratio measurements may help to estimate rates of salinification. Such measurements should be made immediately upon sample collection as precipitation of carbonates at rates up to ~2.5 × 10⁻³ g kg⁻¹ month⁻¹ contribute to uncertainties significantly larger than the measurement accuracy of the salinometer.

Our river and well measurements (Table 2) represent a one-time sampling of systems that the more comprehensive data of Kadyrov (1986) show to have temporal and spatial variability. Considering such variability, our results for the two major rivers Dzhergalan and Tyup agree well (better than 20%) with the results of Kadyrov (1986) and of Lyons et al. (2001). Our artesian well samples show generally low concentrations except for Bar Bulak, a spa located a few hundred meters from the southwestern shore. Its salinity is high, and its Ca²⁺ and Cl⁻ concentrations are higher than those found in the lake.

4. CONDUCTIVITY AND SALINITY

In order to calculate salinity from in-situ CTD conductivity for Issyk-Kul water with an ionic composition different from that of seawater, the dependence of conductivity on temperature, pressure, and salinity needs to be established for this lake. While for freshwater lakes with sufficiently diluted water, some of these relationships can be calculated (Wüest et al., 1996;

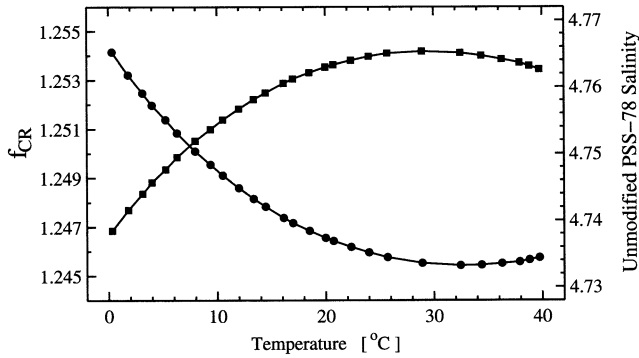


Fig. 2. Unmodified PSS-78 salinity (■) versus temperature for a laboratory conductivity experiment using Lake Issyk-Kul water. The ratio, f_{CR} , of the conductivity of seawater diluted to the salinity of Lake Issyk-Kul water to the measured lake water conductivity is drawn as ●.

Millero, 2000), they would generally have to be determined experimentally for the brackish water of Lake Issyk-Kul to define an Issyk-Kul specific salinity scale. Such a major undertaking is beyond the scope of our current study. Instead we have determined the conductivity-temperature relationship for a Lake Issyk-Kul water sample at its salinity and at atmospheric pressure. The Lake Issyk-Kul specific conductivity dependence on pressure is adopted in modified form from Finger (2002). Using these results and the ionic absolute salinity of the lake as calculated from our major ion measurements we are able to derive a procedure for computing salinity from our CTD measurements. In this procedure we convert the conductivity of Lake Issyk-Kul water to a conductivity that seawater would have at the same pressure and temperature, if diluted to the same salinity as that of Lake Issyk-Kul. This seawater-equivalent conductivity is then used in the Unesco (1981a) PSS-78 algorithm to calculate practical salinity. In this procedure second-order effects in these relationships are approximated by those for diluted seawater. (See Fig. 2)

We use several definitions of salinity throughout this paper. We refer to the *practical salinity* S as that calculated from conductivity using the Unesco (1981a) algorithm and report it as a unitless quantity on the Practical Salinity Scale 1978 (PSS-78). We follow the 1978 recommendations of the Joint Panel on Oceanographic Tables and Standards and define the *absolute salinity* S_A [g kg⁻¹] as the mass fraction of salts in a solution. Both of these salinities are referred to as *ionic salinities* as they are calculated from conductivity measurements of dissolved charged constituents in the water. Finally we will use the *nonionic* S_{Si} [g kg⁻¹] defined as the mass fraction of uncharged silicic acid in the solution.

4.1. Measured and Calculated Conductivity-Temperature Relationships

In order to measure the conductivity-temperature relationship of Issyk-Kul water at atmospheric pressure we conducted a laboratory experiment at SIO using a mixture of four deep water samples collected at Station 8 (Table 1). An aliquot of ~100 g of this mixture was introduced into a loop of vinyl tubing which connected a *Seabird* pump to a conductivity sensor (SBE-4C) which was bracketed by two temperature

Table 4. Electrical conductivity, κ_{EXP} of Lake Issyk-Kul water measured at different temperatures, T (ITS-90), and at 1 atm in a laboratory experiment using *Seabird SBE-3plus* temperature sensors and a *SBE-4C* conductivity sensor.

T [°C]	κ_{EXP} [mS cm ⁻¹]	T [°C]	κ_{EXP} [mS cm ⁻¹]
0.327	4.619	18.516	7.469
1.813	4.835	19.997	7.718
3.128	5.028	20.726	7.842
4.000	5.158	22.343	8.118
5.243	5.345	24.000	8.404
6.312	5.508	25.710	8.701
7.974	5.764	28.923	9.265
9.374	5.982	32.464	9.896
10.519	6.163	34.399	10.244
12.000	6.399	36.270	10.583
13.355	6.617	37.909	10.881
14.459	6.797	38.813	11.046
16.135	7.073	39.731	11.214
16.996	7.215		

sensors (SBE-3plus). A tee was installed to branch off a long tube to an air moisturizer, thereby equilibrating pressure without exposing the sample to evaporation. The temperature sensors were calibrated at SIO prior to the experiment and the conductivity sensor was calibrated by carrying out the same experiment with seawater diluted to low salinity, thereby covering the same conductivity range as that of Lake Issyk-Kul. The sensor assembly was first immersed in various baths of different salinities to test the potential influence of the external electrical field on the conductivity measurements of the sample. The salinities derived from the conductivity measurements during these preliminary experiments remained constant, thereby suggesting that the conductivity results were insensitive to the external field. The sensor assembly was then immersed in a controlled-temperature freshwater bath initially stabilized at 35°C to allow partial degassing of the sample. Air bubbles were then removed through the tubing tee. The SBE temperatures were slightly higher than the bath temperatures due to the heat added to the sample by the pump and sensors. However, the temperature decrease from the first to the second SBE temperature sensor was less than 1×10^{-3} °C, allowing us to accurately determine the temperature inside the conductivity sensor. Conductivities (κ_{EXP}) and temperatures were measured in the range of 0°C to 40°C. The measurements were conducted in variable sequences of ascending and descending temperatures and the measurements at 4°C, 12°C, and 24°C were repeated several times throughout the experiment. This showed the absence of hysteresis effects and leaks of freshwater into the loop. It revealed a small positive practical salinity drift of less than 7×10^{-4} over the course of the entire experiment, which we corrected by linearly detrending the results.

The resulting measured Lake Issyk-Kul water conductivities, listed in Table 4, were fitted by least-squares to a third-order polynomial as function of temperature

$$\kappa_{EXP}(S_{A,EXP}, T, P_0) = 4.5719 + 0.14342 \times T + 7.9652 \times 10^{-4} \times T^2 - 5.0066 \times 10^{-6} \times T^3, \quad (1)$$

where κ_{EXP} is the experimentally derived conductivity in mS cm⁻¹ for the salinity $S_{A,EXP}$ of the sample, for the temperature

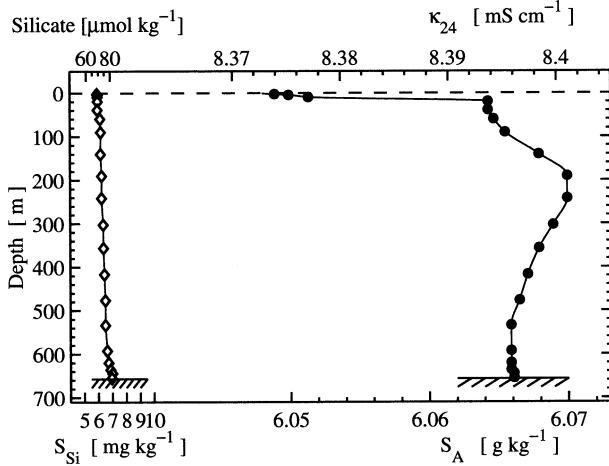


Fig. 3. Lake Issyk-Kul absolute silicate salinity, S_{Si} (lower left axis) and absolute ionic salinity, S_A (lower right axis) drawn on the same scale. The ionic salinity is based on electrical conductivity derived from measurements collected at Station 8 and analyzed on a salinometer at 24°C (upper right axis). Silicate salinity is calculated according to Wüest et al. (1996) from the concentration of uncharged silicic acid (upper left axis) which is taken from Vollmer et al. (2002). The water surface is drawn as a dashed line and the sediment-water interface 2 m below the deepest sample is drawn as a solid line.

T (°C, ITS-90) and for atmospheric pressure, P_0 . The experimentally derived conductivity at 24°C is 8.404 mS cm^{-1} which is slightly higher than that inferred from the salinometer measurements (8.396 mS cm^{-1}) of the original lake water mixture. This difference is because of possible evaporation during storage of the sample and because of some remaining uncertainties in the conductivity sensor calibration procedure.

To derive a temperature-conductivity relationship κ_{IK} for the Station 8 mean absolute salinity, $S_{A,IK} = 6.06 \text{ g kg}^{-1}$ (Table 1) we scale the experimentally derived conductivities, κ_{Exp} , to the Station 8 mean conductivities which we calculate using the salinometer results shown in Fig. 3

$$\kappa_{IK}(T, P_0) = 0.9988 \times \kappa_{Exp}(T, P_0). \quad (2)$$

To determine the scaling factor we have converted the Station 8 mean conductivity ratio at 24°C from our salinometer measurements, to an absolute conductivity by multiplying by the conductivity of the standard at the same temperature [$52.029 \text{ mS cm}^{-1}$, Unesco (1981a)]. This absolute conductivity was divided by the experimentally derived conductivity at the same temperature. By making the conversion in Eqn. 2 we assume that the temperature dependence of the conductivity, $d\kappa_{Exp}(S, T, P_0)/dT$ changes insignificantly over a small salinity range (Jellison et al., 1999; Wüest et al., 1996).

We have compared the experiment-based κ_{IK} with the conductivity-temperature relationship *calculated* from the charge unbalanced major ion data using the method of Wüest et al. (1996). This showed calculated conductivities to be lower by about 3% over the temperature range of 0°C to 25°C, either confirming the limitation of this procedure to low-salinity water only (Wüest et al., 1996) or suggesting that when balancing the charge the overall conductivity should be increased by adjusting the anionic concentrations. If we increase Cl^- by 2.1 mmol

kg^{-1} to balance the charge, then the calculated conductivity is still lower than the measured conductivity by 1.5%. More importantly, the change of the calculated, Cl^- -adjusted conductivity with changing temperature, $d\kappa/dT$, differs significantly from that of the measured conductivity by 8% to -2% over this temperature range, suggesting that such an approach to calculate the conductivities of this relatively saline water may indeed be invalid (Wüest et al., 1996).

4.2. Salinity using Equations for Seawater

In this approach we calculate salinity from conductivity by drawing on the extensive data for seawater. This allows us to approximate second-order effects in the relationships between conductivity and salinity, temperature, and pressure for Lake Issyk-Kul water by using those for diluted seawater. To make use of the Unesco (1981a) PSS-78 algorithm, we first convert the CTD in-situ conductivity to the conductivity a seawater sample would have at the same absolute salinity and temperature. We relate these two conductivities by a factor f_{CR} , which we calculate by dividing the conductivity of seawater, κ_{SW} (PSS-78 algorithm), diluted to the Station 8 mean absolute salinity ($S_{A,IK}$), by the Lake Issyk-Kul conductivity κ_{IK} (Fig. 2). This ratio was fitted to a third order polynomial, yielding

$$\begin{aligned} f_{CR}(T) &= \frac{\kappa_{SW}^{PSS-78}(S^*, T^*, P_0)}{\kappa_{IK}(S_{A,IK}, T, P_0)} \\ &= 1.2544 - 6.2513 \times 10^{-4} \times T + 1.3413 \times 10^{-5} \\ &\quad \times T^2 - 7.8671 \times 10^{-8} \times T^3, \end{aligned} \quad (3)$$

where T is the temperature [°C] on the ITS-90 scale and T^* is that temperature on the IPTS-68 scale. S^* is the unitless practical salinity of a standard seawater sample which corresponds to the absolute salinity $S_{A,IK}$ [g kg⁻¹] of Lake Issyk-Kul water. This conversion can be approximated (Millero, 1982) from

$$S_{A,IK} = 1.00488 \times S^*. \quad (4)$$

The factor $f_{CR} \approx 1.25$ (Fig. 2) shows that seawater is more conductive than Lake Issyk-Kul water by this factor, if diluted to the lake's salinity. To illustrate the importance of the implementation of the temperature-dependent f_{CR} , we plot both results, with and without multiplying the conductivities by f_{CR} , in Fig. 4. These results show that salinities using the unmodified Unesco (1981a) PSS-78 algorithm yields salinities of $\sim 4.77 \text{ g kg}^{-1}$, if converted to an absolute salinity using Eqn. 4. Also, some surface water salinities are nominally higher than the deep-water salinities. By converting the measured conductivities to seawater-equivalent conductivities, the overall salinity increases to $\sim 6.06 \text{ g kg}^{-1}$ and the surface shows a distinct freshwater layer.

Based on field experiments, Finger (2002) has shown that the dependence of conductivity on pressure for Lake Issyk-Kul is also different from that of seawater. His experiments have shown that over the full pressure range of the lake, a salinity error of $\sim 0.003 \text{ g kg}^{-1}$ is introduced when using the conductivity-pressure relationships for seawater (Unesco, 1981a; Bradshaw and Schleicher, 1965, 1980). We use the results by Finger (2002) to derive a pressure-dependent correction function f_P

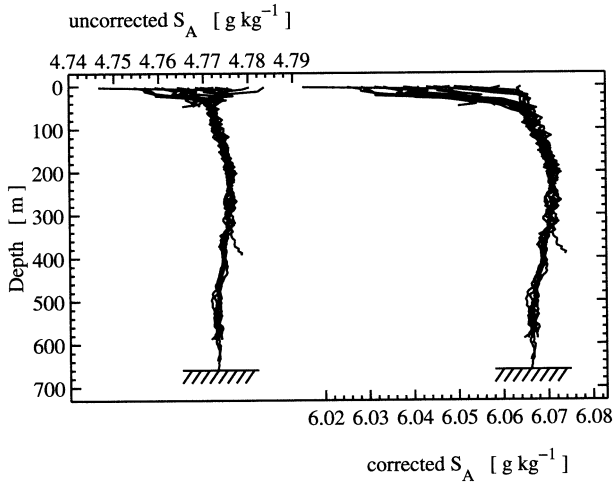


Fig. 4. Lake Issyk-Kul absolute ionic salinities, S_A , for 5 m averaged vertical profiles from all stations along a north-south transect occupied during September 22–24, 2000. The salinities were calculated from PSS-78 algorithm (Unesco, 1981a) and converted to units of $[g\ kg^{-1}]$ (Millero, 1982) and are plotted at the same scale. Salinities calculated from in-situ conductivities (left), assuming seawater composition, show lower salinities and inconclusive evidence of a freshwater surface layer. Salinities calculated using a correction specific to the major ion composition and their temperature-dependent conductivity (right) show higher salinities and a distinct freshwater surface layer. For both sets of profiles shown here, the pressure correction on the conductivity and the conductivity sensor calibration are included (see text).

$$f_p = 1 - 9.5 \times 10^{-7} \times P, \quad (5)$$

where P is the pressure [dbar].

Our in-situ CTD conductivities are finally multiplied by f_{CR} (Eqn. 3) and f_p (Eqn. 5), yielding a salinity-equivalent seawater conductivity, and are then used in the Unesco (1981a) PSS-78 algorithm along with pressures and temperatures to calculate practical salinity,

$$S = S_{SW}^{PSS-78}(C \times f_{CR} \times f_p, T^*, P). \quad (6)$$

This practical salinity, which will be used for density calculations and other Unesco (1981a) based equations, can be reconverted to an absolute salinity S_A using Eqn. 4, thereby allowing us to report salinity in $g\ kg^{-1}$. Equation 6 is in general applicable to any CTD data collected from the lake. In Table 5 we provide check values for this function.

4.3. Uncertainties in Salinity Calculation

Our approach to calculate salinity is, in good approximation, applicable to most Lake Issyk-Kul in-situ conductivities. How-

Table 5. Test values for calculating Lake Issyk-Kul specific practical salinities S_p (Eqn. 6), absolute salinities S_{abs} (Eqn. 4), and densities (Eqn. 7) using temperature T (ITS-90), electrical conductivity C , and pressure P .

T [°C]	C [mS cm ⁻¹]	P [dbar]	S_p [?]	S_{abs} [g kg ⁻¹]	Density [kg m ⁻³]
15	7.0	600	6.100	6.130	1007.093
10	7.0	600	6.964	7.000	1008.583
15	6.5	600	5.633	5.661	1006.696
15	6.5	0	5.681	5.708	1003.978

ever, whether this procedure can be applied to a mixture containing a significant fraction of river water remains to be tested. It is likely that in such cases the temperature dependence of the conductivity will deviate from that for open-lake salinities due to the strong dilution and the altered relative ionic composition. The error in our calculated salinity is of similar size to, and is dominated by, the errors in the major ion measurements. Fortunately this affects all salinities equally and is therefore of little relevance when comparing water masses for stability assessment. The large vertical changes in in-situ conductivity (mainly due to temperature changes) however require a careful expedition-specific calibration of the conductivity sensor to avoid significant errors in vertical salinity profiles. We have calibrated our conductivity sensor, as is typical for oceanographic studies, by relating the CTD measurements to the salinometer conductivity ratios using the Unesco (1981a) PSS-78 algorithm. However in our case the adaptation of this algorithm using Eqns. 3 and 5 is a crucial part of the calibration procedure.

In our approach to calculate salinity using a modified PSS-78 algorithm (Unesco, 1981a) we have approximated second order effects by using those for seawater. Such effects are among others, the salinity dependence on the conductivity-temperature relationship and the temperature dependence on the conductivity-pressure relationship. In order to assess the magnitude of these effects, we have also calculated salinity using a limnological approach as described in detail by Wüest et al. (1996) in which these second-order effects are ignored. This limnological approach yields absolute salinities for Station 8 that are lower by $0.5 \times 10^{-3}\ g\ kg^{-1}$ to $1.5 \times 10^{-3}\ g\ kg^{-1}$ compared with the algorithm based on Unesco (1981a) PSS-78. This comparison shows that these second order effects for seawater are small over the temperature and salinity range of our study. It has been shown that these effects are also small for other saline (Jellison et al., 1999) and freshwater (Wüest et al., 1996) lakes. Assuming that these effects are also small for Lake Issyk-Kul water, we conclude that they would have to be grossly different from those of seawater to be of significance for our data.

4.4. Nonionic Salinity

We have further calculated the salinity, S_{Si} (Fig. 3), due to silicic acid, $Si(OH)_4$ (McManus et al., 1992; Wüest et al., 1996), for the Station 8 profile collected during this survey (Vollmer et al., 2002). Using a pH of 8.2 at a deep-water temperature of 4.0°C (Kadyrov, 1986) we calculate 98% of the silicic acid to be uncharged (DOE, 1994). This fraction results in a nonionic salinity which increases from 6.0 $mg\ kg^{-1}$ at the surface to 6.9 $mg\ kg^{-1}$ at the maximum depth of Station 8. Even though this component of the salinity is smaller than the uncertainties in the major ion results, its vertical change is larger than our uncertainties in the vertical variability of S_A derived from precise salinometer results and it has non-negligible effects on the stability in the bottom mixed layer.

5. DENSITY

In order to derive a true equation of state for Lake Issyk-Kul, the density must be determined at various salinities and temperatures, a task which is again beyond the scope of this study.

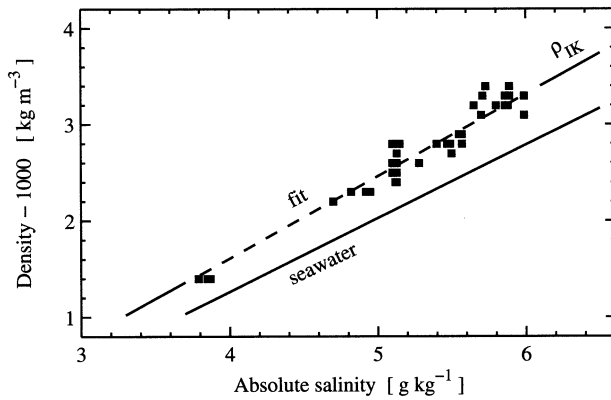


Fig. 5. Density anomaly for Lake Issyk-Kul water at 20°C as a function of absolute salinity. Data shown as squares are from the measurements by Kadyrov (1986). The dashed line is a least-square linear fit through these data and additionally forced through $S = 0$ and the density anomaly of pure water corrected for a small contribution of silicate density. The upper solid line, which is interrupted over the range of the measurements, is calculated from the Issyk-Kul specific density function, ρ_{IK} (Eqn. 7), and shows excellent agreement with the measured densities. The lower solid line is for seawater (Unesco, 1981b).

Alternatively, we make use of the international equation of state for seawater (Unesco, 1981b) and follow the approach used by Peeters et al. (2000) for the Caspian Sea by calculating the density of Issyk-Kul water ρ_{IK} as

$$\rho_{IK}(S, T, P) = \rho_{SW}(S = 0, T, P) + f_{\beta} \times (\rho_{SW}(S, T, P) - \rho_{SW}(S = 0, T, P)), \quad (7)$$

where ρ_{SW} is the density [kg m^{-3}] of seawater from the international equation of state for seawater, EOS-1980 (Unesco, 1981b), as a function of practical salinity, S , temperature, T ($^{\circ}\text{C}$ on IPTS-68), and pressure P , and f_{β} is the ratio of the haline contraction coefficient for the ionic composition of Issyk-Kul water (β_{IK}) to that of seawater (β_{SW}):

$$f_{\beta} = \frac{\beta_{IK}}{\beta_{SW}} = 1.12 \quad (8)$$

where values of both β_{IK} and β_{SW} are for 25°C and for the same salinity. The $\beta_{IK} = 8.335 \times 10^{-4}$ reciprocal salinity units was calculated according to a procedure described by Wüest et al. (1996) using partial molal volumes of the major ions at 25°C (Millero et al., 1977), which were charge balanced by increasing Cl^{-} by 2.1 mmol kg^{-1} . With a constant f_{β} , it is assumed that the values of β are independent of salinity and that they do not change relative to each other with changing temperature or pressure. In Fig. 5 we compare the calculated density anomaly, $\rho_{IK} - 1000$, with results from measurements of specific weight by Kadyrov (1986). Specific weight was converted to density anomaly from which a small density fraction for silicic acid (Wüest et al., 1996) was subtracted.

Values for ρ_{IK} at 20°C show excellent agreement with the fitted measurements of Kadyrov (1986). For an absolute salinity of 6.06 g kg^{-1} and a temperature of 20°C we calculate a density of 1003.316 kg m^{-3} which is 48 g m^{-3} less than the extrapolated value for the fit through Kadyrov's data. For the

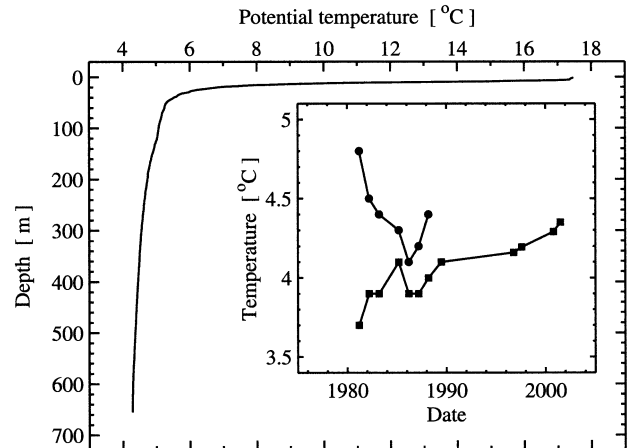


Fig. 6. Lake Issyk-Kul potential temperature (Smart-CTD) for Station 8 on September 23, 2000. Inset, interannual temperature variability in Lake Issyk-Kul: Potential temperatures near maximum depth (■) and winter surface temperatures at 5 m (●) are plotted against date. Temperatures prior to 1989 are from Romanovsky (1990), for 1989 is from the World Lake Database (<http://www.ilec.or.jp>), for 1996 is from Kipfer and Peeters (2002), for 1997 is from E. A. Ralph (unpublished data), for 2000 is from this study, and for 2001 is from Finger (2002).

current mean salinity at Station 8 and a temperature of 20°C, Issyk-Kul water is 530 g m^{-3} denser than seawater diluted to the same salinity, mainly due to the higher relative abundance of SO_4^{2-} in Issyk-Kul water. Using this density function we calculate a temperature of maximum density of 2.55°C for the surface water which is 0.15°C lower than that for seawater diluted to the same salinity.

6. STABILITY AND MIXING PROCESSES

The establishment in the preceding sections of relationships between temperature, conductivity, salinity, and density specific to Lake Issyk-Kul allows us to examine the stability of the water column and to discuss the interplay between temperature and salinity differences in the vertical mixing of the lake. During our September 2000 expedition, temperature profiles in Lake Issyk-Kul showed a thin surface mixed layer with temperatures varying between 16.5°C and 18.5°C, a sharp thermocline between ~10 m and ~60 m, and nearly isothermal deep water with temperature variations of less than 0.7°C between 150 m and the maximum depth (Fig. 6). The lowest potential temperatures of ~4.29°C were observed near the lake floor at several deep stations. As previously observed by Romanovsky (1990), our north-south transect temperature section showed a strong vertical doming effect with the isotherms being shallowest in the center of the lake. This effect is related to the strong counterclockwise divergent circulation in the lake. In winter, the open lake surface water cools to 4°C to 5°C while the water in the shallow bays may freeze. In Fig. 6 we plot the near-surface and bottom temperature evolution since 1980 showing considerable temporal variability, and sporadically, very small vertical temperature differences of only 0.2°C. Our deep water temperatures are significantly higher compared to these earlier observations but they are also higher than some of the earlier winter near-surface temperatures. This may be related to less intense deep-water renewal due to milder winters in the years

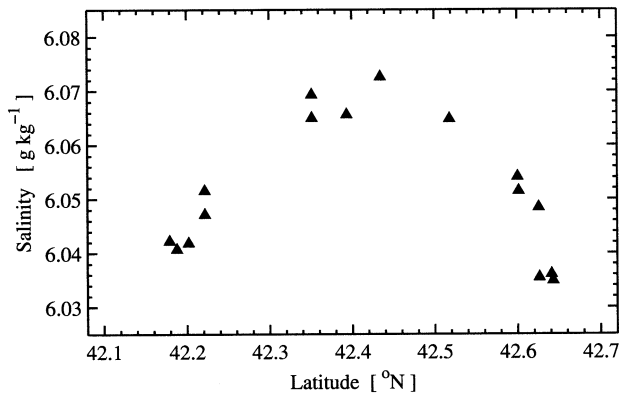


Fig. 7. Horizontal salinity gradient in Lake Issyk-Kul derived from CTD measurements. The mean absolute salinities for the upper thermocline in the depth range of 10 m to 30 m are plotted against latitude, showing a freshening towards the shores.

preceding our measurements (Romanovsky, private communication).

The overall absolute salinity variations in Lake Issyk-Kul for our transect are very small and do not exceed 0.06 g kg^{-1} (Fig. 4). The lowest salinities of about 6.02 g kg^{-1} are observed in the surface layer. At the surface a distinct freshening towards the northern and southern shores can be observed, which is most likely related to river inputs and the upwelling of more saline water in the center of the lake (Fig. 7). Absolute salinity reaches a maximum around 200 m, below which it decreases slightly. The cause of this salinity maximum is difficult to explain. It may be associated with the intrusion of high-salinity water from the sediment that was deposited during lower lake stands in the past when the water was more saline, and re-exposed to the lake through fissures opened during neotectonic block movements (Zektser and Diordiev, 1979; Romanov et al., 1989). Saline hydrothermal fluids such as those found in some wells surrounding the lake (Kadyrov, 1986) may also seep into the lake. Alternatively this salinity maximum may be caused by mixing to a depth of about 200 m in the winter when the surface water may become more saline than the water below due to enhanced evaporation and reduced freshwater inflow.

We were unable to find earlier conductivity or ionic salinity data that would allow us to assess the temporal persistence of the overall vertical salinity structure. However Finger (2002) showed that the local salinity maximum at about 200 m depth was still present 6 month after our study. At Station 11 near the southern shore ($42^\circ 12.13' \text{ N}$, $77^\circ 14.46' \text{ E}$) salinity is enhanced near 400 m depth compared to other stations (Fig. 4). For this profile temperatures were slightly elevated and transmissivity slightly reduced, but not so much as would be the case if the sensors had touched the sediment. We are unable to explain this feature by any other mechanism than the seepage of high-salinity water into the open lake.

There is evidence of a $\sim 15 \text{ m}$ thick bottom nepheloid layer in the central part of the lake, as suggested by decreasing light transmissivity. The ionic salinity appears to be slightly increased in this layer, as shown by the salinometer results (Fig. 3) but this increase is small and of similar magnitude to the precision of the measurement. The increase is not apparent in the corresponding CTD data because the conductivity sensor's

output signal locked on a single value over the lowest 150 m of the profile. The uniform potential temperature and the uniform oxygen and nutrient concentrations (Vollmer et al., 2002) suggest that this layer is well mixed.

To estimate the static stability of the water column at Station 8, we have calculated the Brunt-Väisälä frequency (Gill, 1982; Wüest et al., 1996)

$$N^2 = N_T^2 + N_S^2 + N_{Si}^2 = g\alpha \left(\frac{dT}{dz} + \Gamma \right) - g\beta_{IK} \frac{dS}{dz} - g\beta_{Si} \frac{dS_{Si}}{dz}, \quad (9)$$

where N^2 is the sum of temperature-dependent stability, N_T^2 , the salinity-dependent stability, N_S^2 , and the stability due to the vertical distribution of uncharged silicic acid, N_{Si}^2 (Fig. 8), g is the gravitational acceleration, α is the thermal expansivity, Γ is the adiabatic lapse rate, and z is the depth positive upward. The haline contraction coefficients are β_{IK} , defined earlier, and $\beta_{Si} = 0.36 \times 10^{-3} \text{ kg g}^{-1}$ (Wüest et al., 1996) for the major ions and for silicic acid, respectively. Our N^2 calculation is based on least-square spline-fitted salinometer and silicic acid data (Fig. 3) and on the SBE-39 temperature data for the deepest bottle cast. All these data were averaged into 10-m depth bins. In Fig. 8, the Brunt-Väisälä frequency is shown below a depth of 200 m. The combined N^2 decreases with increasing depth below 200 m to values below $1 \times 10^{-7} \text{ s}^{-2}$ between 300 m and 500 m. In this region the salinity has a destabilizing effect on the water column. Near the bottom nepheloid layer, temperature is nearly uniform and N_T^2 almost vanishes. However, salinity and silicate increase through the boundary to the nepheloid layer, thereby adding to the stability of the water column, which is most likely further strengthened by the suspended particles. In this layer, silicic acid contributes significantly to N^2 . The potential temperature near the bottom mixed layer has previously been observed to increase with increasing depth (E. A. Ralph, unpublished data), and can also be seen to increase for one of our profiles, possibly due to geothermal warming. Such a temperature structure could significantly contribute to a destabilization and therefore to enhanced mixing at this depth.

Turbulent diffusion cannot be the only mixing process, as the vertical temperature gradient seems to be permanent. Convective mixing processes are likely to contribute significantly to deep water renewal. With its intermediate salinity, the lake presents a fascinating example of delicate balance between salinity and temperature variations causing slight density differences which, in turn, may allow waters with certain T - S characteristic to sink to greater depth (Fig. 9). While in freshwater lakes of temperate climatic zones the surface water generally cools to or below its temperature of maximum density, $T\rho^{\max}$, of $\sim 4^\circ\text{C}$ during winter, the open Lake Issyk-Kul water never reaches its $T\rho^{\max}$ of 2.55°C , presumably because of the lake's large heat capacity and the relatively mild winters in this area. However in shallow bays the water has been observed to cool below $T\rho^{\max}$. Horizontal mixing of such water with warmer open lake water will create intense cabbelling (or thermal bars) which can produce water dense enough to sink (Romanovsky, 1990). However water in bays near river inflows may be considerably fresher than the open lake, which could prevent such a mixture from sinking even if it reaches $T\rho^{\max}$.

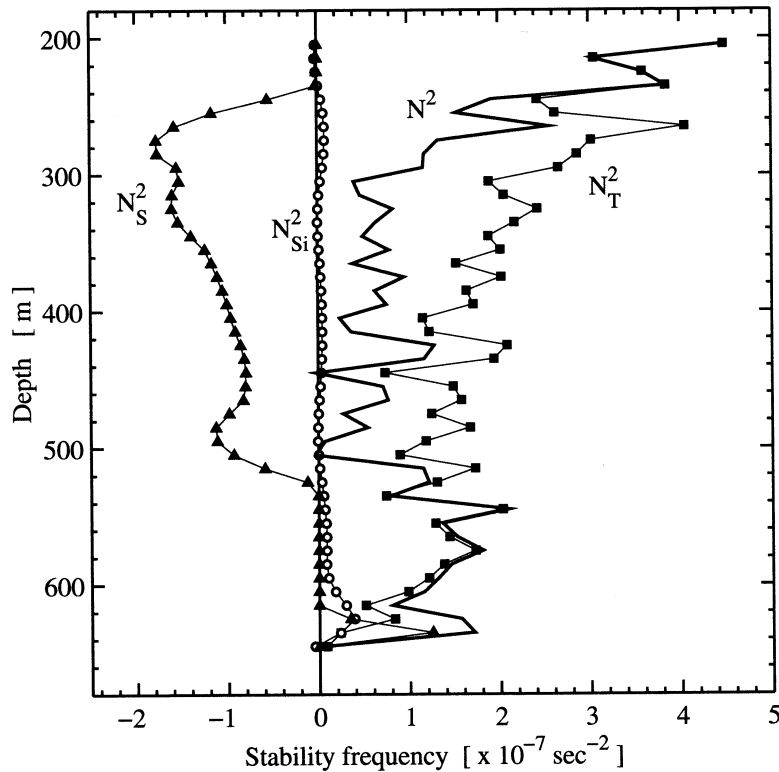


Fig. 8. Brunt-Väisälä frequency for Lake Issyk-Kul Station 8 on September 23, 2000 below 200 m. The combined frequency, N^2 , consists of a temperature-related fraction N_T^2 (■), an ionic salinity related fraction N_S^2 (▲), and a silicate related fraction N_{Si}^2 (○).

This is illustrated by the temperature-salinity-density relationship plotted in Fig. 9. Unlike freshwater lakes, where the enhanced salinity of inflowing rivers can cause density plumes (e.g., Hohmann et al., 1997), river inflow to Lake Issyk-Kul is too fresh to reach densities higher than those of the lake's saline water, regardless of the rivers' temperatures. Some major rivers may carry significant sediment loads, thereby increasing their densities. However it remains to be examined whether this extra load can cause a sufficient density increase to result in sinking. In the shallow eastern part of the lake where the majority of inflow occurs, the rivers may lose most of their larger particles to the sediment before reaching deeper water.

If our observed low-salinity surface cap, attributed to precipitation and river inflow, is a typical feature, it forms an effective initial low-density barrier that can barely reach the potential density of the deep-water even if cooled to $T\rho^{\max}$ (Fig. 9). In winter, such a freshwater cap would need to be eroded first by turbulent mixing and by increased evaporation compared to the relatively low freshwater input during this season. Then the surface could reach a salinity high enough that subsequent cooling could reach the potential density of the deep-water and consequently sink.

Another possible candidate for deep-water renewal is brine rejection during ice formation in the bays at the eastern and western ends of the lake. If a similar fraction as in seawater (~ 0.6) of salt is rejected by the formation of ice, then the surrounding water will become significantly denser than any water at the salinity of the open lake, even if the temperature is near -1°C and well removed from $T\rho^{\max}$. The salinity maxi-

um at 200 m (Fig. 4) and the T - S diagram (Fig. 9) for the CTD profiles below a temperature of 5°C suggest the existence of an endmember with higher salinity (by $\sim 0.01\text{ g kg}^{-1}$) and a temperature of $\sim 4.7^\circ\text{C}$. However this feature only reaches intermediate density and it appears unlikely that this feature is related to brine rejection. Interestingly, the anomalous near-bottom salinity at Station 11 points towards such an endmember which supports our alternate hypothesis that hydrothermal input may cause this salinity maximum instead.

7. CONCLUSION

Based on measurements of major ions and the temperature dependence of conductivity for Lake Issyk-Kul water, we have derived an algorithm that allows us to calculate salinities from CTD data. Using the major ion data and an adopted algorithm (Peeters et al., 2000) we have calculated densities which show excellent agreements with measured densities. The procedures are based on parametrizing the relationships between the equations of state for Lake Issyk-Kul to those of seawater, and we have shown that the errors involved with adopting second-order effects from seawater algorithms are small. These procedures can be applied to any saline lake if the ratios of the ions in the lake are constant.

Our September 2000 observations of Lake Issyk-Kul have revealed a delicate density balance determined by small gradients in temperature and salinity. Further research is necessary to explore the persistence of these gradients in time and space. Large salinity variations, as are thought to have occurred over

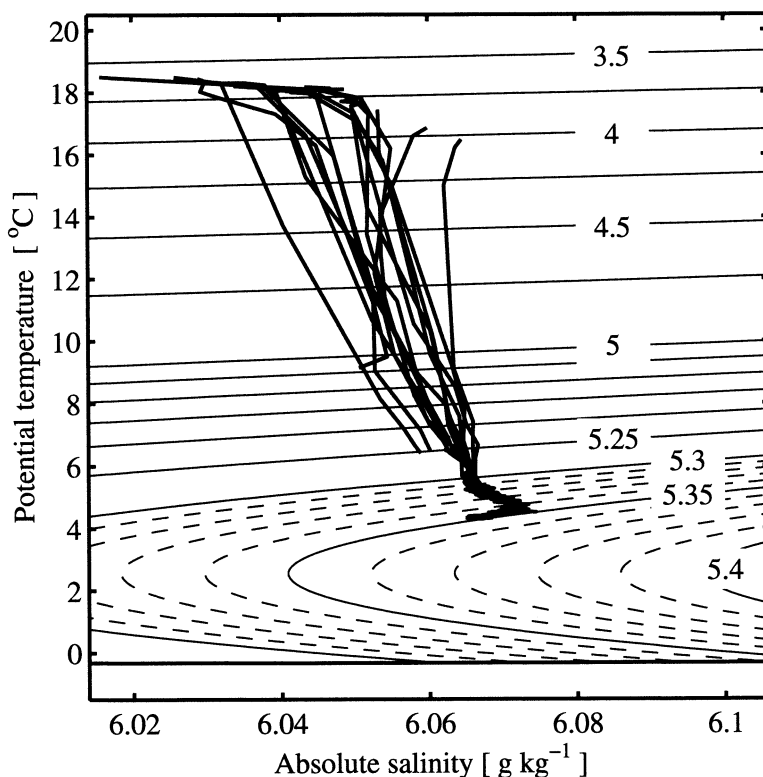


Fig. 9. Temperature-salinity diagram for Lake Issyk-Kul. CTD profiles for September 2000 are plotted as solid lines on isopycnals contours (solid thin lines and dashed lines) expressed as density anomalies $1000 - \rho_{TK}$ [kg m^{-3}]. The contour intervals change at 5 kg m^{-3} and at 5.3 kg m^{-3} . The freezing temperatures of seawater at Lake Issyk-Kul salinities are plotted as a solid horizontal line at the bottom of the panel below 0°C .

geological time, have important consequences on the mixing regime of the lake. In the past, when the lake was fresher, river water may have greatly enhanced deep-water formation through thermal bars and elevated salinities compared to the lake water. It is likely that the fate of the inflowing rivers was drastically different from present day and that therefore the nutrient fluxes in the lake may have been altered, leading to a different ecological setting. Due to a shift to higher temperature of maximum density which comes with lower salinity, the bottom water temperature during such times was unlikely below $\sim 2.7^\circ\text{C}$ – 4.0°C , which are the temperatures of maximum density for the pressures of the deep and surface water, respectively. On the other hand, with the lake's present salinity, the bottom water could potentially reach $\sim 1.2^\circ\text{C}$ – 2.6°C in a climate with colder winters than those of the present day. Such changes should ultimately be seen in the sedimentary record. Understanding the present-day mixing processes should therefore greatly help in interpreting these records, as well as predicting changes resulting from climatic and human impacts on the lake.

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