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Effect of salt exclusion from lake ice on seasonal circulation

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Abstract

We describe the circulation of Tailings Lake, located 220 km north of Yellowknife, Northwest Territories, Canada. From Jul 2004 to Sep 2007, temperature and conductivity profiles were collected together with data from three moored temperature chains and a meteorological station. The salt content of the lake was elevated (salinity 1 g L^{-1}) relative to surrounding natural waters (~0.1 g L^{-1}). Winter ice cover included 60–80 cm of transparent "black" ice, which excluded up to 99% of the salt from the water as the ice formed. During winter, the salt excluded from the ice was mixed throughout the water column under the ice. In spring, ice melt and snow runoff resulted in a freshwater cap; the density contrast between this fresh epilimnion and the more saline hypolimnion was sufficient to inhibit spring turnover. In summer, the epilimnion was deepened by storms, mixing anoxic water from the hypolimnion into the surface layer. Because spring turnover was suppressed, oxygen replenishment of the hypolimnion occurred only in fall. Although Tailings Lake has been subject to numerous anthropogenic changes, similar effects can be expected in other natural and manmade water bodies with elevated salinity and significant ice cover.

Brine rejection from sea ice and its importance to ocean circulation has long been appreciated (Carmack 1990). Salt exclusion has also been recognized as a potentially important process in the circulation of ice-covered fjords (Gibson 1999*a*) and saline lakes (Gibson 1999*b*; Willemse et al 2004). The effect of salt exclusion on the circulation of fresh and brackish lakes has generally been considered minor, has been used to explain anomalies in temperature profiles (Ellis et al. 1991), and has rarely received detailed study. One exception is Welch and Bergmann (1985) who measured under-ice circulation in a small Arctic lake driven by heat released from the sediments and cryoconcentration of salts from the ice cover.

In the present paper we examine the importance of salt exclusion on a brackish lake in the Arctic in which thick ice cover and a high degree of salt exclusion play an important role. Although the lake we describe is part of a mine site, and as such has undergone significant alteration, the results of this work apply to other high-latitude lakes with elevated salinity, whether natural or manmade.

Study site

Tailings Lake (64°26.9′N, 115°03.4′W) is part of the Colomac gold mine, 220 km north of Yellowknife, Northwest Territories, Canada. The mine site lies in rolling bedrock hills in the Slave Craton of the Canadian Shield at an elevation of 340 m a.s.l. During mine operation from 1990 to 1997, wastewater and 11×10^9 kg of solid tailings were piped to a zero-discharge tailings containment area. Tailings completely filled the adjacent Spruce Lake and formed extensive beaches on the south and east sides of Tailings Lake. After bankruptcy, remediation of the site fell

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to the federal government of Canada and has been ongoing since 1998, in preparation for eventual discharge of lake water to the environment.

The primary contaminant of concern in Tailings Lake was cyanide and its byproducts; fortunately, the waste rock and ore from the mine site were relatively low in sulfides, and acid rock drainage has not been a significant problem. Although cyanide concentrations in Tailings Lake had been decreasing (from ~100 mg L^{-1} in 1997 to 1 mg L^{-1} in 2001), the concentrations of thiocyanate remained substantial (120 mg L^{-1} in 2001), and concentrations of ammonia, a thiocyanate byproduct, were also high (45 mg L^{-1} in 2001).

To enhance the natural removal of both thiocyanate and ammonia, Tailings Lake was treated with mono-ammonium phosphate in 2002 (1 mg L^{-1} P) and 2003 (0.7 mg L^{-1} P). This treatment stimulated phytoplankton productivity, which enhanced both the degradation of thiocyanate and the removal of ammonia. Thiocyanate decreased below detection in summer 2004 and ammonia was reduced to <0.5 mg L^{-1} by fall 2005. The primary remaining concern is release of phosphorus from the sediments.

Tailings Lake has a total drainage area of 3.16 km^2; uncontaminated runoff from 1.48 km^2 of the total drainage was diverted during the study period. During the study, the lake was below full pool, with a net increase in water level of ~0.3 m $year^{-1}$. With average annual precipitation, outflow is expected in 2009. Tailings Lake has a single basin 1.4 km long, with a surface area of 0.50 km^3, a volume of 3.4 Mm^3, and a maximum depth of 14 m during the study period.

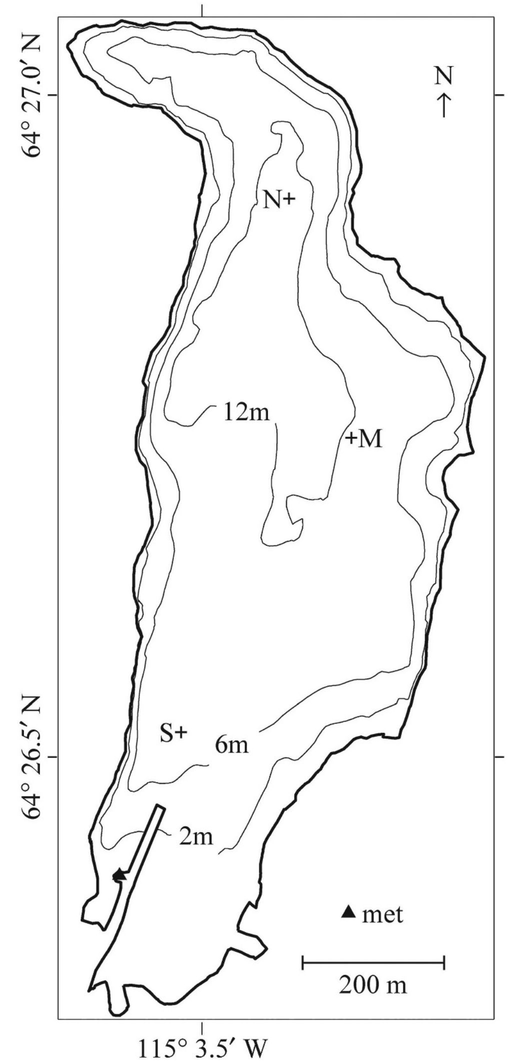

Fig. 1. Bathymetry of Tailings Lake. N, M, and S mark stations TLK-N, TLK-M, and TLK-S, respectively. The triangle marks the location of the meteorological station (met) on the causeway.

Methods

Data collection—The main axis of the lake lies roughly north–south; along it, three sampling stations were located: TLK-N, TLK-M, and TLK-S at the north, mid, and south ends of the basin, respectively (Fig. 1). Three moorings, one at each sampling station, consisted of internally recording temperature loggers attached to a line at 1-m intervals to 10 m, except to 7 m at TLK-S. This line was suspended from a 50-liter buoy and anchored with chain; a backup ground line ran from the anchor chain to shore. The loggers were Onset Hobo Water Temp Pros with an accuracy of $\pm0.25°C$. Temperature was recorded every 20 min from 01 Jul 2004 to 19 Sep 2007 and uploaded each fall.

Profile data were collected monthly during open water and once each winter during ice cover. Two types of profile data were collected. As part of ongoing monitoring, data were collected with a Hydrolab Minisonde-4 giving temperature ($\pm0.1°C$), conductivity (±15 μS cm^{-1}), and dissolved oxygen (±0.2 mg L^{-1}). Although the Hydrolab was routinely calibrated for 100% oxygen saturation, the accuracy was poorer than the instrument specification at low oxygen levels. Oxygen readings of 0–2 mg L^{-1} were often recorded for water smelling of H_2S or containing Fe in predominantly dissolved form, both of which indicate the absence of oxygen.

High-accuracy temperature ($\pm0.005°C$) and conductivity (±1 μS cm^{-1}) were obtained with a Seabird SBE19plus conductivity–temperature–depth (CTD) profiler. The Seabird conductivity was calibrated in the laboratory against a Guildline Portasal salinometer and agreed to better than 0.1% throughout the study period. In situ conductivity was converted to conductivity at 25°C, C_{25}, and salinity, $S = 0.67 \times 10^{-3}\ C_{25}$ [μS cm^{-1}], following Pawlowicz (2008).

Ice samples were obtained by collecting chips from an auger hole that was drilled progressively deeper, carefully clearing the hole and surrounding ice to separate chips from different depths. Chips were collected in Ziploc bags, melted, and stored in high-density polyethylene bottles, and conductivity was measured using a Guildline Portasal.

Meteorological data were collected by a Davis Vantage Pro weather station located on the causeway at the south end of Tailings Lake. Averaged wind speed, wind direction, air temperature, relative humidity, and solar radiation were collected every 2 h from 18 Aug 2004 to 19 Sep 2007.

In this paper we will describe in detail the data from 01 Jul 2004 to 15 Oct 2005. The behavior of Tailings Lake from Nov 2005 to Sep 2007 was similar unless noted.

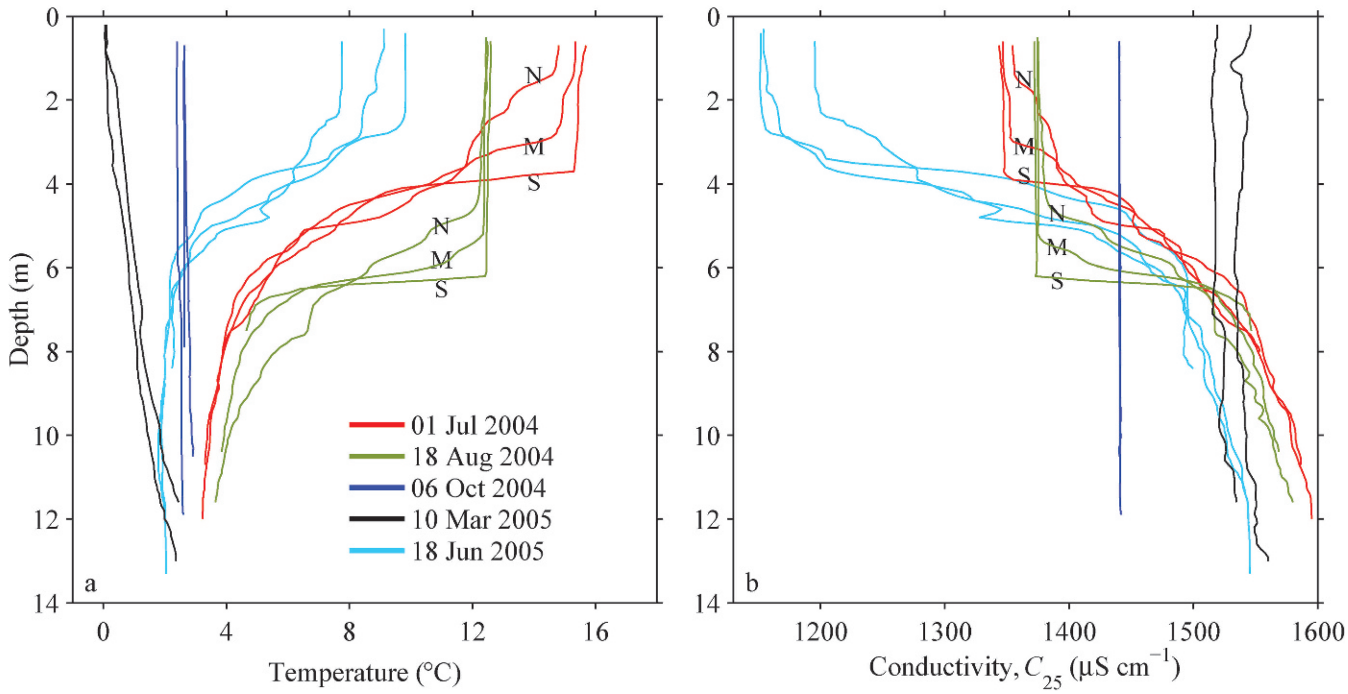

Fig. 2. Selected profiles of (a) temperature and (b) conductivity, 2004–2005. N, M, and S mark TLK-N, TLK-M, and TLK-S, respectively.

Calculations—To examine the effect of salt exclusion during ice formation in winter it is convenient to have a simple equation for density valid over a limited range of temperature and salinity. We adapt the equation by Farmer and Carmack (1982) as follows:

$$\rho = (\rho_o + bS)\left[1 - a(T - T_{MD})^2\right] \quad \left(\mathrm{g\ cm^{-3}}\right) \qquad (1)$$

where S is salinity (g L^{-1}), T is temperature (°C), $T_{MD} = 3.98 - 0.215\,S$ (°C) is the temperature of maximum density, $\rho_o = 0.999975$ g cm^{-3}, $a = 8.25 \times 10^{-6}$ °C^{-2} and $b = 8.05 \times 10^{-4}$ g cm^{-3}. Equation 1 agrees with the equations for both freshwater (Chen and Millero 1986) and seawater (Millero et al. 1980) to 10^{-5} g cm^{-3} in the range $0 \le S \le 2$ and $0 \le T \le 8$°C. When $T > 8$°C, during the open water season, we use Chen and Millero (1986).

To calculate stability just after ice-off and through the open-water season, we need density, which depends not only on temperature but also on conductivity. Although the moorings provide continuous temperature data, the only conductivity data comes from profiles, the first of which were taken 10–20 d after ice-off. We estimated the conductivity of the surface mixed layer throughout the open-water season using a simple budget based on the first CTD profile after ice-off. The depth structure of the conductivity was assumed to be the same as that of temperature, an assumption supported by the available CTD profiles. The predicted surface-layer conductivity showed good agreement with that from CTD profiles collected throughout the rest of the open-water season.

The potential for mixing after ice-off is evaluated using the Wedderburn, $W = g' h_1^2 / (u_*^2 L)$, and Lake, L_N, numbers (Imberger and Patterson 1990). Both represent the ratio of the stabilizing effect of density stratification to the destabilizing effect of wind, but L_N accounts more accurately for the stratification and the geometry of the basin. For two-layer stratification in a square basin with a shallow surface layer $W = L_N$ (Stevens and Imberger 1996). Here h_1 is the depth of the surface layer, L is the length of the lake, and $g' = (\Delta\rho/\rho_o)g$ is the reduced gravity, where $\Delta\rho$ is the density difference across the pycnocline, ρ_o is the mean density, and g is gravity. The wind is expressed as a shear velocity, $u_* = \sqrt{\rho_a C_D u_{10}^2/\rho_o}$, where ρ_a is the density of air, $C_D = 1.3 \times 10^{-3}$ is the drag coefficient; and u_{10} is the 2-h wind velocity 10 m above the lake, estimated from wind speed measured at 5.5 m using law of the wall scaling.

We also compute the stability during the open-water season as the energy needed to mix the entire lake,

$$St = \left(\frac{g}{A(0)}\right) \int_0^H A(z)(\rho(z) - \bar{\rho})z\,dz \quad \left(\mathrm{J\ m^{-2}}\right) \qquad (2)$$

where z is the depth from the surface, $A(z)$ is the area, and H is the maximum depth.

Results

We begin with an overview of seasonal changes by looking at the profile data from Jul 2004 to Oct 2005, followed by a more detailed look at each season using the moored temperature and meteorological data.

Annual cycle—The conductivity in Tailings Lake was brackish, averaging 1440 μS cm^{-1} (salinity 0.96 g L^{-1};

Table 1. Lake and ice characteristics, including f_w, fraction of snow in white ice, and f_b, fraction of excluded salt from black ice.

	Max. depth (m)	Area (km^2)	Date	Depth (cm)		C_{25} (μS cm^{-1})				f_w white ice*	f_b black ice*
				White ice	Black ice	White ice	Black ice	Winter water	Summer		
Tailings Lake	14	0.46	Mar 05	10	80	485	17±11	1530	1440	0.68	0.99
Zone 2 Pit§	110	0.15	Mar 05	10	75	440	10±4	1155	1140	0.62	0.99
			Apr 06	16	59	230	140	1110	—	0.79	0.87
Spot Lake†	5	~0.01	Mar 05	32	63	77	7	205	120	0.62	0.97
Paddle Lake†	5	0.065	Mar 05	19	56	46	5	165	61	0.72	0.97
Faro‡	~90	0.51	Jan 06	11	44	600±50	90±40	1200	—	0.50	0.93
Grum‡	~50	0.095	Jan 06	12	40	570±50	32±4	1000	—	0.43	0.97
Vangorda‡	~50	0.059	Jan 06	6	39	160±30	24±7	1600	—	0.90	0.98

* Using observed under-ice water conductivity during winter sampling.
† Natural lake at Colomac mine site.
‡ Pit lakes at Faro mine site, 200 km north of Whitehorse, Yukon (62.353°N, 133.364°W).
§ Pit lake at Colomac mine site.

with major ions sodium and sulfate). In contrast, the conductivity of 17 surrounding natural lakes ranged from 23 to 177 μS cm^{-1} and averaged 94 μS cm^{-1} (salinity 0.06 g L^{-1}; with major ions calcium and carbonate).

At the time of the first CTD profile on 01 Jul 2004, Tailings Lake was stratified in both temperature and conductivity; a warm, fresh surface layer overlaid cooler, more saline water (Fig. 2). The conductivity stratification resulted from ice melt and freshet runoff. Note that a north wind on 01 Jul 2004 resulted in a surface layer that was shallower at the north station, TLK-N, and deeper at the south station, TLK-S. The effect of wind on the pycnocline will be examined in further detail below.

Besides tilting the pycnocline, wind also mixes the surface layer down, entraining water from below. On 01 Jul 2004 the surface layer was ~4 m deep, but by 18 Aug the surface layer had been mixed to ~6 m (Fig. 2). Through early fall, the surface layer continued to both cool and mix down until the entire lake overturned, as shown by the uniform vertical profiles on 06 Oct 2004. Shore ice had begun to form by 06 Oct 2004 and freeze-up was complete by mid-Oct 2004.

Characteristics of the lake ice are summarized in Table 1; data for six other water bodies (four pit lakes and two natural lakes) are given for comparison. For all seven water bodies, the ice consisted of a thin layer of white ice on top of a thick layer of black ice. Of particular note is the high degree of salt exclusion from the black ice in Mar 2005: 97–99% of the salt was excluded.

Under-ice profiles from 10 Mar 2005 are shown in Fig. 2. The temperature was reverse stratified, increasing linearly from 0°C below the ice to 2°C at the bottom. Unlike temperature, the conductivity profile in Mar was relatively uniform with depth (Fig. 2b). The exclusion of salt from the ice raised the conductivity of the entire basin from 1440 μS cm^{-1} on 06 Oct 2004 to 1530 μS cm^{-1} on 10 Mar 2005 (Fig. 2b). This increase in conductivity is approximately consistent with a simple salt balance. Note that salt excluded from the ice also increased the conductivity of Zone 2 Pit and Spot, and Paddle Lakes (Table 1). For the shallow natural lakes, the increase in conductivity from summer to winter was particularly large as a result of high surface area relative to volume. The conductivity of Spot Lake increased from 120 μS cm^{-1} in summer to 205 μS cm^{-1} in Mar 2005, and that of Paddle Lake increased from 61 to 165 μS cm^{-1}.

Ice-off was complete by early Jun 2005. Profiles from Tailings Lake on 18 Jun 2005 show stratification in both temperature and conductivity (Fig. 2). The surface mixed layer was shallow with a depth of about 4 m. Near the bottom (>10 m), the temperature and conductivity were essentially the same as observed under ice on 10 Mar 2005. This is the first indication that spring turnover did not occur. Note that the deep temperature on 18 Jun 2005 was ~2°C, less than the temperature of maximum density adjusted for salinity, 3.8°C, another indication that salinity stratification prevented spring turnover. We now examine each season in further detail using the moored data.

Fall turnover—The temperature at TLK-N is given as a line plot in Fig. 3a showing two open water seasons, 2004 and 2005, and the intervening period of winter ice cover. Data at TLK-N are shown because this was the deepest station; data at TLK-M and TLK-S were similar unless noted. In 2004, fall turnover began during cool and windy weather on 17 Sep 2004 when the lake became isothermal (all the temperature lines overlaid). The entire lake remained isothermal as the lake cooled from 6.5°C on 17 Sep 2004 to just below 2°C on 5 Oct 2004. Early in fall turnover, there were days when fine weather caused a slight warming at the surface, but this slight stratification was lost at night. Later, when the lake was below 4°C, there were nights when the surface cooled to give slight reverse stratification, with colder, less-dense water at the surface, but this reverse stratification was lost during the day. After 10 Oct 2004 (day 284), reverse stratification persisted, and complete ice cover appeared around 15 Oct 2004 (day 289), as suggested by reduced temperature fluctuations at all depths.

In 2005, fall turnover and ice-on were later than in 2004. Fall turnover began on 27 Sep 2005, and the lake cooled from 5.2°C to 1.9°C by 15 Oct 2005 (not shown). After mid-Oct, reverse stratification persisted, with ice cover

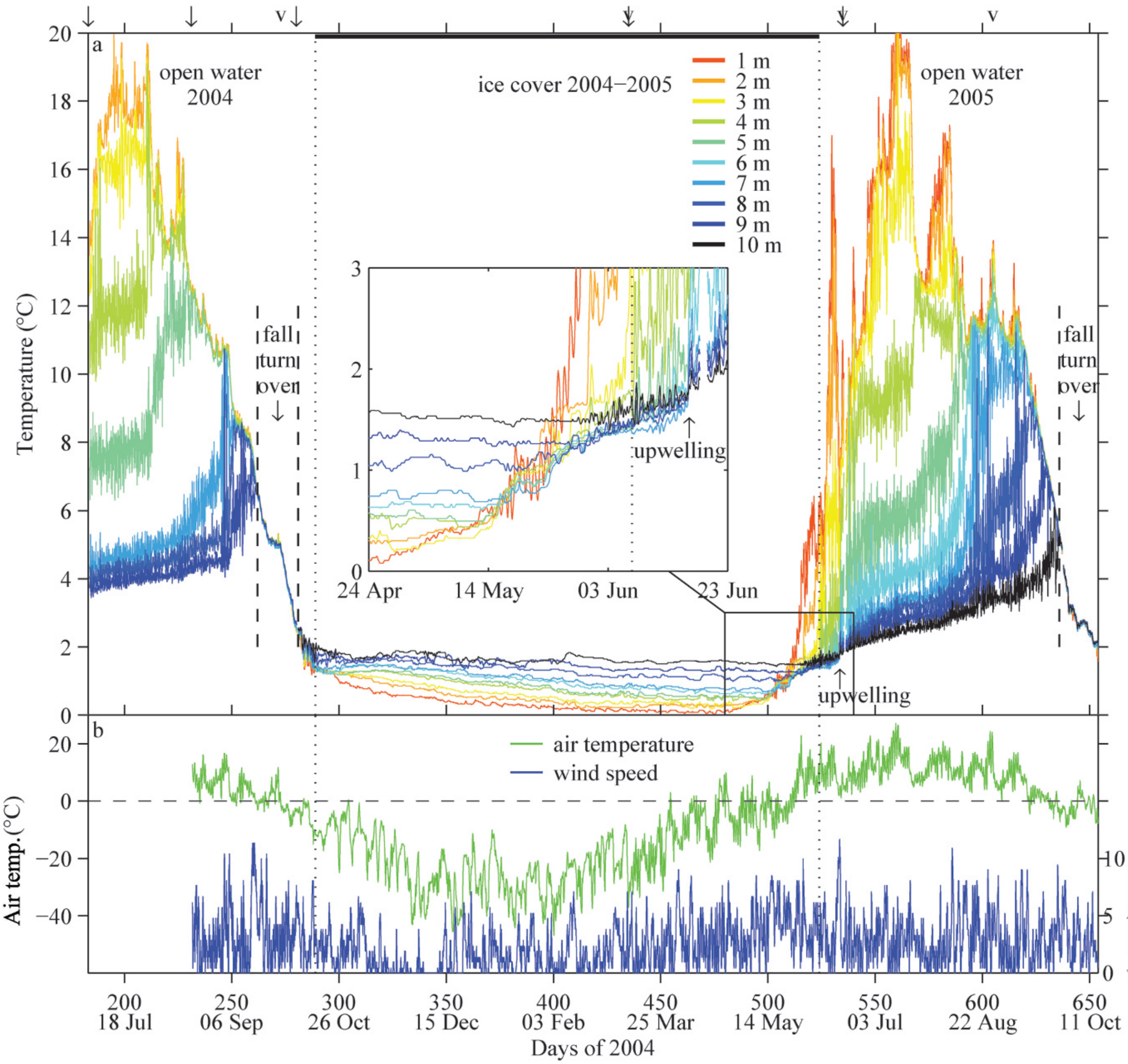

Fig. 3. (a) Line plot of temperature at TLK-N, and (b) wind speed and air temperature from 01 Jul 2004 to 15 Oct 2005. The inset shows the water temperature around the time of ice-off on expanded scale. The approximate period of ice cover is marked with a thick black line; estimated times of ice-on, 15 Oct 2004, and ice-off, 07 Jun 2005, are marked with dotted lines. Arrows (CTD) and v (oxygen) mark the dates of the data shown in Figs. 2 and 6, respectively. Data were plotted shallow first and deep last so that shallow data lines may lie under deeper ones. The moorings were installed 01 Jul 2004 and the met station on 18 Aug 2004. No data during mooring service on 06 Oct 2004 (day 280), 18–19 Jun 2005 (day 535–536) and 28 Sep 2005 (day 637). Temperature data at 10 m began 06 Oct 2004.

complete about 29 Oct 2005. However, unlike 2004, there were several windstorms between mid-Oct and mid-Nov 2005 strong enough to disrupt the ice.

Winter ice formation—The water temperature under ice cover is shown in the midsection of Fig. 3a. Just after ice-on, the largest temperature gradient occurred between the bottom of the ice at 0°C and the sensor at 1 m. As the winter continued, the temperature gradient was spread more evenly through the water column as the entire lake gradually cooled (cf. the Mar 2005 profile in Fig. 2a). However, the water column remained temperature-stratified throughout the winter, though there were times when the temperature was the same at adjacent sensors, suggesting localized mixing. During winter, consistently cold air temperature (Fig. 3b) resulted in the growth of 0.9 m of ice by Mar 2005.

Spring ice-off—In the Colomac region, spring melt generally occurs as a series of overlapping events. The exact timing of each event in 2005 is not known for Tailings Lake, but we infer the timing from local knowledge, water level, and the moored temperature data as follows. A few warm and sunny days can melt the snow on the ice as early as April, which allows increased penetration of sunlight through the ice. In 2005, the temperature at the 1-m sensor began to increase at the end of April (Fig. 3a), suggesting the absence of snow on the ice at this time. Because the water just below the ice was at a temperature less than T_{MD} = 3.8°C, this water became denser as it warmed, and progressively mixed downward by radiative-driven convection (Mironov et al. 2002). This occurred from 28 Apr to 16 May 2005 (day 484–502); at the end of this time the surface mixed layer penetrated to about 5 m depth and was about 0.7°C (Fig. 3a inset).

Next, runoff from the drainage, fed by both snowmelt and spring rain, slotted into Tailings Lake as a relatively fresh lens under the ice. In 2005, a rapid increase in water level (0.3 m) indicated that freshet began in mid-May and lasted through mid-Jun. In mid-May, the temperature at 1 m began to show a strong diurnal signal, independent of the temperature at 2 m. This suggests that the 1 m temperature sensor was within a freshwater lens just under the ice. By ice-off, the temperature at 1 m reached 6.5°C, and the fresh lens mixed down to about 2 m (Fig. 3a). The warming of the fresh lens under the ice was likely the result of sunlight penetrating through the ice and heat transported by the freshet inflow. Temperatures of 5–7°C under the ice were also seen in 2006 and 2007; a similar effect was also observed in Store Saltsø, Greenland, by Willemse et al. (2004). During this time, water below the fresh lens (>3 m), began to warm slowly, reaching ~1.5°C at the time of ice-off.

By the last week of May, the ice is typically "rotten" (candled) and ice-off usually occurs in the first week of Jun, often quite suddenly. Although the exact date of ice-off in 2005 is not known, we can infer the date of ice-off from the movement of the TLK-M and TLK-S moorings. Comparison of the deep temperature of all three moorings (not shown) indicates that the moorings at TLK-M and TLK-S moved during a windstorm on the afternoon of 07 Jun 2005 (day 524). The moorings were found shifted by 50–150 m, into slightly shallower water as constrained by their ground lines to shore. Because wind-driven ice movement during breakup can exert significant forces, we assume this as the date of ice-off.

After ice-off, there was no time during which the surface and deep temperatures changed in lockstep as they did during fall turnover (Fig. 3a). In the first 5 d after ice-off, the surface layer temperature increased by over 10°C, whereas the deep water temperature increased by only 0.1°C. The independence of the deep and near-surface temperatures confirms that spring turnover did not occur.

Summer stratification—The evolution of stratification during the open water season is shown in contour plots of water temperature for 2004 (Fig. 4b) and 2005 (Fig. 4e). Through summer and early fall, the thermocline deepened in response to episodic wind events. For example, on 02 Sep 2004 (day 246), wind mixed the surface layer down from 5 to 7 m (Fig. 4b). In 2005, a series of windy days in mid-Aug mixed the surface layer from 5 to 8 m (Fig. 4e). Over the summer, the hypolimnion warmed slowly (~0.5°C month^{-1}; Fig. 3a) and became slightly fresher (cf. 01 Jul–18 Aug 2004; Fig. 2b), likely a result of a small degree of vertical transport as often observed in lakes.

To evaluate the potential for spring mixing, the Wedderburn and Lake numbers are computed for open-water stratification in 2005 (Fig. 4d). The Wedderburn and Lake numbers were very similar except during fall, and the Lake number will be discussed in what follows because it accounts more accurately for stratification and geometry. Throughout the open water season, L_N was generally >1 except during windstorms immediately after ice-off and during fall turnover. Although L_N was <1 during two spring storms (Fig 4d; 08–09 Jun 2005 and 15–16 Jun 2005), these storms did not last long enough to induce spring turnover. As the open-water season progressed, thermal stratification quickly came to dominate the stability. The rapid rise in surface layer temperature immediately after ice-off results from the high solar insolation experienced at this latitude during Jun and ensures that the period of time during which the lake is vulnerable to turnover is short. In summer, L_N was generally much greater than 1, except during the Aug storms, when L_N approached 1 and the surface layer deepened. Finally, in Sep, L_N was <1 for long enough to achieve fall turnover.

We now examine the response to the large windstorm that occurred on 15–16 Jun 2005 (days 532–533, just before the period of missing data during mooring upload, 18–19 Jun; Fig. 4e). During this storm, wind from the north pushed the surface water to the south. The effect of the storm can be seen at TLK-N (Fig. 4e), where cold deep water was briefly upwelled to the surface. Upwelling was observed in 2003 during a windstorm from the north (W. Coedy pers. comm.) with the appearance of "black" water smelling of H_2S at the north end of the lake, in contrast to the usual "green" surface water colored by high chlorophyll.

The effect of the storm of 15–16 Jun 2005 is also shown in temperature contours along Tailings Lake (Fig. 5). At the start of 13 Jun (day 530) there was little wind and the thermocline was relatively level (Fig. 5b). A weak wind from the south during the day on both 13 Jun and 14 Jun resulted in slight upwelling during midday at the south mooring (Fig. 5c,e). When the wind switched to the north–northwest toward the end of 14 Jun (day 531) the thermocline began to upwell in the north (Fig. 5f). During the height of the storm on 16 Jun, deep water was upwelled to the surface at TLK-N and surface water was downwelled below 7 m at both TLK-M and TLK-S (Fig. 5i). After the wind subsided on 17 Jun, a horizontal thermocline was reestablished (Fig. 5j,k).

Although the storm of 15–16 Jun 2005 resulted in significant disturbance, the lake resisted turnover as a result of temperature and salinity stratification. Nevertheless, the storm had a variety of effects. It mixed the surface layer down from approximately 3.5 m before the storm to 4.5 m after (Fig. 4e; 16 Jun, day 533). Both this mixing of the surface with cooler deep water and the storm itself resulted in significant cooling of the surface layer from ~11 to 8°C. The upwelling event also resulted in a transfer of surface

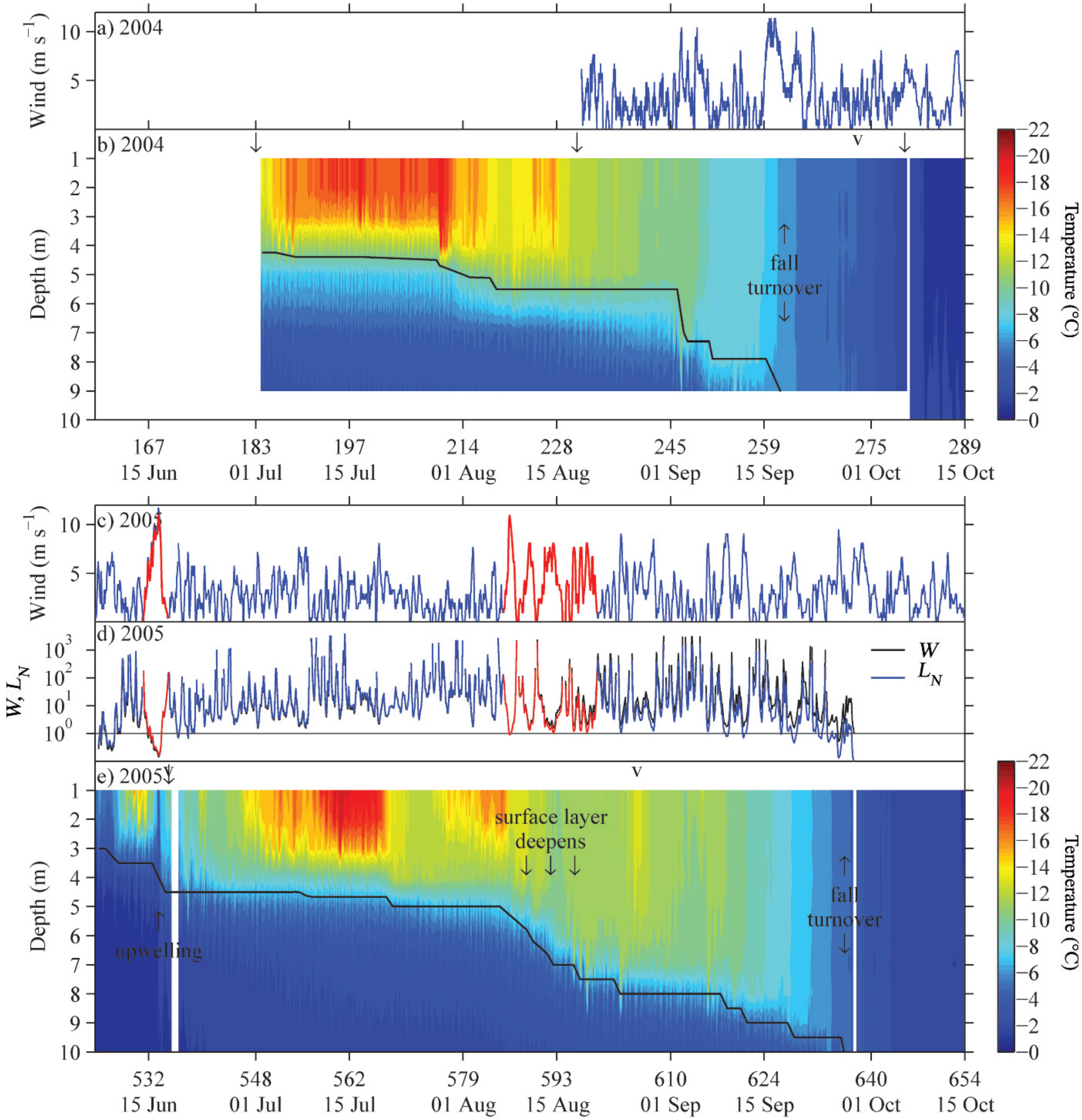

Fig. 4. Open-water (a) wind 2004, (b) water temperature 2004, (c) wind 2005, (d) Wedderburn and Lake number, 2005 and (e) water temperature 2005 for TLK-N. The panels start and end at the approximate time of ice-off and ice-on. The moorings were installed 01 Jul 2004 and the met station on 18 Aug 2004. In the contours, a solid line marks the estimated depth of the pycnocline. Arrows (CTD) and v (oxygen) mark the dates of the data shown in Figs. 2 and 6, respectively. Selected storms are highlighted in red.

water into the hypolimnion, which can be seen in the small jump (0.2°C) in deep temperature (Fig. 3 inset; 16 Jun, day 533).

Oxygen—Dissolved oxygen data were consistent with the absence of spring turnover. There was little or no oxygen recorded in the hypolimnion in spring 2005 (18 Jun; Fig. 6). The absence of oxygen was supported by the smell of H_2S from water collected below the pycnocline. However, had spring turnover occurred, the absence of oxygen a week or two later could also have resulted from the high oxygen demand of this eutrophic lake. In contrast to spring, dissolved oxygen was replenished throughout the water column during fall turnover (29 Sep 2004; Fig. 6).

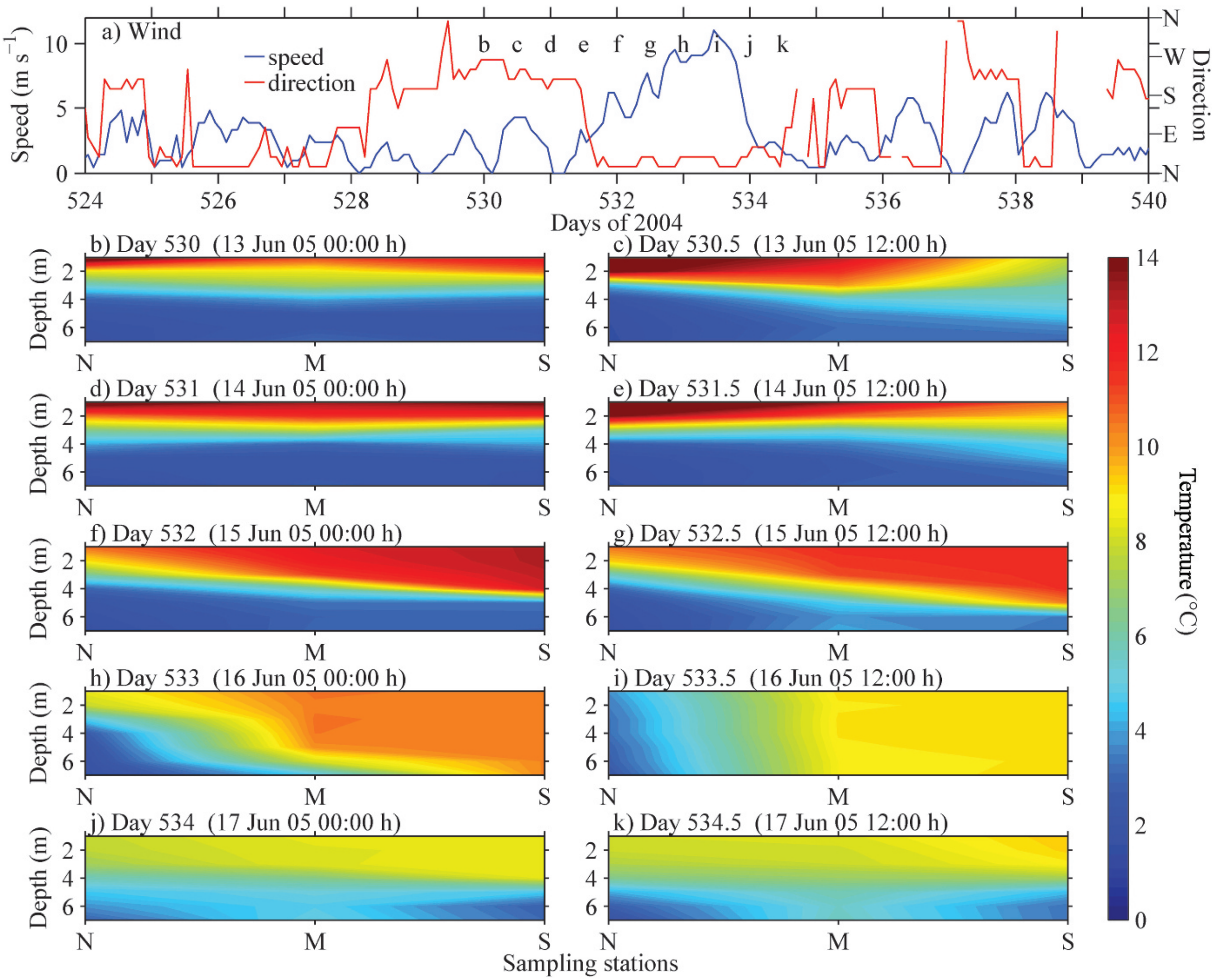

Fig. 5. (a) Wind speed for 08–22 Jun 2005 before and after a large windstorm on 15–16 Jun 2005. (b–k) Along-lake contours of water temperature at selected times. N, M, and S mark TLK-N, TLK-M, and TLK-S, respectively. At the height of the storm on 16 Jun, the wind from the north upwells deep water to the surface at TLK-N.

Discussion

In many respects Tailings Lake behaves as a typical dimictic lake. However, the elevated salinity, along with significant ice cover in winter, results in a freshwater cap with sufficient density contrast to inhibit spring turnover. We will first discuss salt exclusion from ice; then, we will look at when salt exclusion might first affect winter stratification; and finally, we will evaluate the ability of the ice melt and freshet inflow to suppress spring turnover.

Salt exclusion and winter circulation—Black ice is transparent and appears black because it shows the dark water at depth (Wetzel 2001). Black ice grows at the bottom of the ice sheet and salt is excluded into the water below the ice. We observed more than 97% exclusion of salt from black ice in winter 2004–2005 (Table 1). Samples taken at different depths within the black ice had a consistently high exclusion of salt, as did samples from different depths within the black ice on Zone 2 Pit. During the following winter, however, only 87% of the salt was excluded from black ice in Zone 2 Pit, based on only one sample, but during a winter when freeze-up was late (no samples were collected from Tailings Lake). The Faro, Grum, and Vangorda pit lakes in the Yukon showed 93–98% salt exclusion in January 2006. For comparison, Belzile et al. (2002) report vertical profiles of salt exclusion from 6 lakes that ranged from 78% to 99%, similar to that reported here. We would expect that the fraction of excluded salt might vary depending on the rate at which black ice was formed.

White ice forms after the weight of snow overcomes the buoyancy of the ice and lake water floods the ice surface. This creates a layer of water-saturated snow or slush on top of the ice. White ice forms as the slush freezes; it appears white because it is polycrystalline and scatters light. As the slush freezes, the salt in the lake water is at first excluded into the slush (slush with a conductivity greater than that of

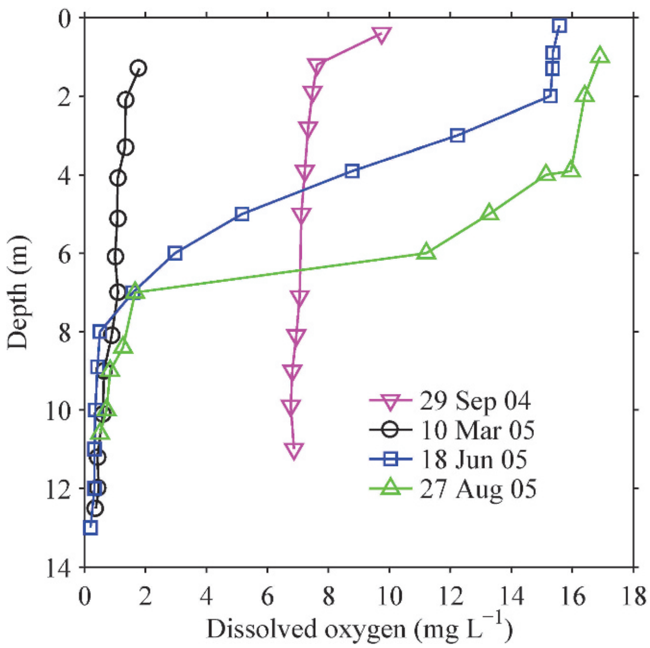

Fig. 6. Selected profiles to illustrate the change in oxygen at TLK-N from fall turnover 2004 through winter ice cover, ice-off, and summer stratification 2005. Readings below 2 mg L^{-1} are inaccurate and usually indicate anoxic conditions.

the lake water was observed on Zone 2 Pit). However, if all the slush freezes, this salt must ultimately be incorporated in the white ice layer. In the absence of slush, the conductivity of the white ice gives the fraction of snow-derived water in the white ice, f_w (Table 1).

Northern lakes are notable for the large proportion of black ice cover (Woo and Heron 1989). Extreme and consistently low air temperatures allow black ice to grow rapidly at the bottom of the ice sheet. Rapid growth favors the formation of black ice because it reduces the probability of accumulating sufficient snow to create white ice. In Tailings Lake, approximately 15% of the lake volume was ice in winter 2004–2005; the fraction for the shallow Spot and Paddle Lakes was considerably higher. Note that the winter conductivity in Spot Lake was 1.7 times that in summer; in Paddle Lake the winter conductivity was 2.7 times the summer value.

To determine when salt exclusion could result in under-ice mixing, we estimate the thickness of ice that would provide sufficient salt to overcome the reverse temperature stratification beneath the ice. We integrate the simplified equation for density (Eq. 1) over an idealized profile of reverse stratification to determine the total salt needed to create a profile of uniform density and neutral stability. Further addition of salt would then overcome the reverse stratification.

The structure of the reverse stratification shortly after freeze-up is illustrated for Tailings Lake and Zone 2 Pit in Fig. 7a. Unfortunately, the shallowest temperature sensor was at 1 m depth, and all that can be said of the reverse stratification is that it occurs in the top 1 m. Additional depth resolution was available for Waterline Pit Lake in central British Columbia (Leung 2003) and for Ryan Lake, Minnesota (Fang et al. 1996). In Ryan Lake the reverse stratification occurs in the top 0.6 m. Note also the range of temperatures below the reverse stratification, from ~1 to 3.5°C (Fig. 7a). For a given water body, this deep temperature will vary from year to year, depending on the exact sequence of meteorological events preceding freeze-up. Increasing the initial deep temperature, T_i, increases the strength of the reverse stratification, though

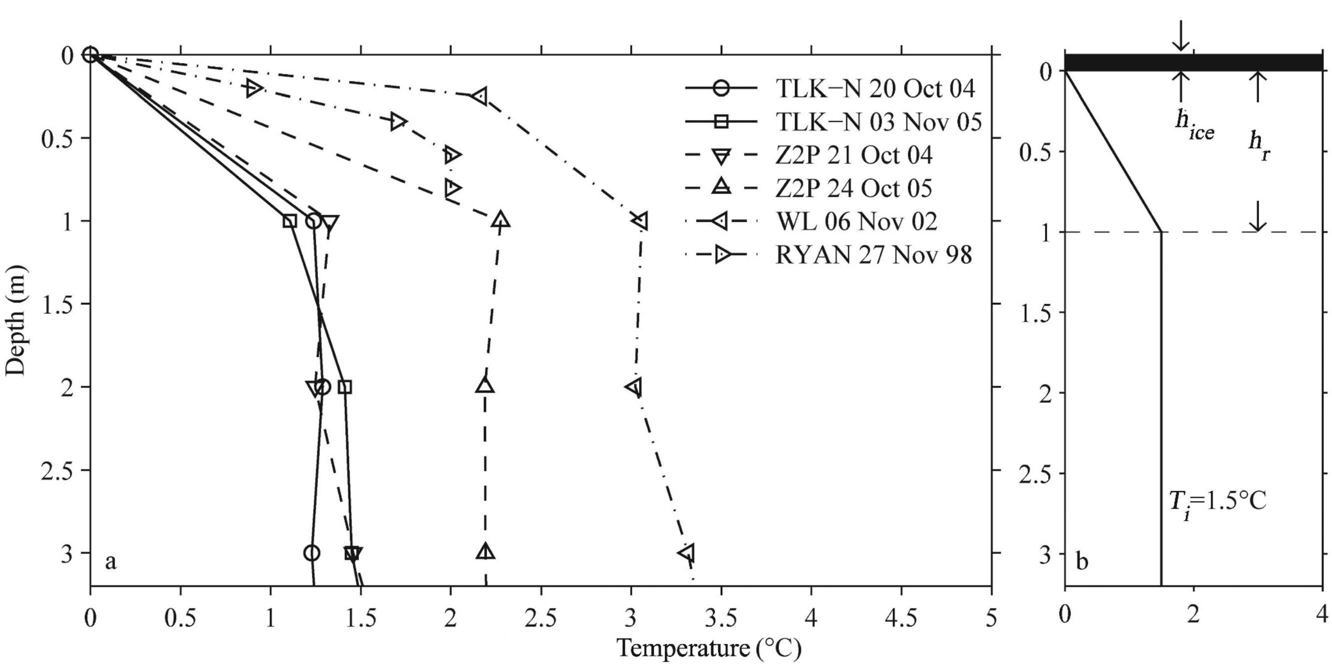

Fig. 7. (a) Temperature profiles from mooring data about 5 d after ice-on showing the reverse stratification for Tailings Lake (TLK), Zone 2 Pit (Z2P), Waterline (WL), and Ryan lakes (RYAN). (b) Idealized profile of reverse stratification and definitions.

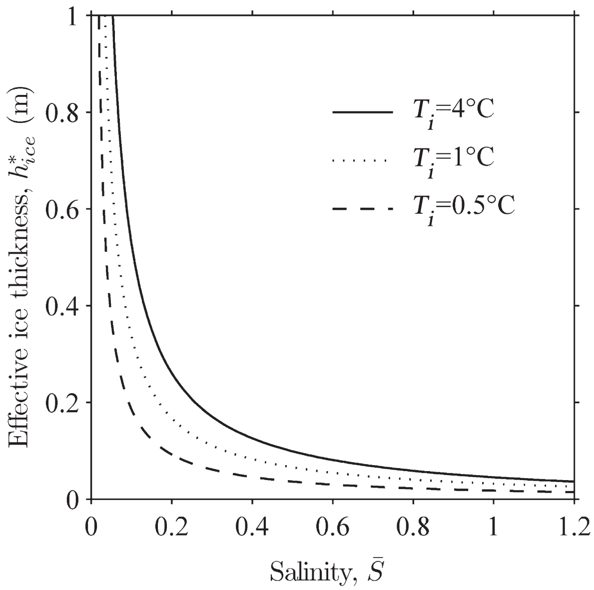

Fig. 8. Effective ice thickness required to destabilize the reverse stratification after ice-on as a function of average lake salinity for $h_r = 1$ m and $T_i = 0.5°C$, 1°C, and 4°C.

the increases are, of course, marginal as T_i approaches T_{MD}.

To simplify the following analysis, we assume that the reverse stratification consists of a linear profile from 0°C at the base of the ice to the initial deep temperature, T_i, at the bottom of the reverse stratification, h_r (Fig. 7b). This provides a slight overestimate of the stability if the actual profiles are curved as suggested by the limited field data.

At neutral stability the density in the region of reverse stratification would be the same as the density below the reverse stratification. This requires additional salinity, $\Delta S(z)$, in the region above h_r to balance the stable temperature stratification, $T(z)$. Using Eq. 1, and equating the density in the region of reverse stratification to that below it, yields

$$\begin{aligned}(\rho_o + bS_i + b\Delta S(z))\left[1 - a(T(z) - T_{MD})^2\right] \\ = \left(\rho_o + b\bar{S}\right)\left[1 - a(T_i - T_{MD})^2\right]\end{aligned} \tag{3}$$

where $\bar{S}$ is the lake average salinity at freeze-up. Assuming linear temperature stratification, $T(z) = (T_i/h_r)z$, solving for $\Delta S(z)$ and integrating through the reverse stratification, from $z = 0$ to $z = h_r$, gives the total salinity needed for constant density:

$$\begin{aligned}\Delta S_{TOT} = \left(\frac{a\, h_r\, T_i}{30}\right)\left(\frac{\rho_o + b\bar{S}}{b}\right) \\ (3T_{MD} - 2T_i)\ \left(\text{g cm}^{-2}\right)\end{aligned} \tag{4}$$

To exclude this salt requires an ice thickness of $h_{ice} = \left(\frac{10\Delta S_{TOT}}{f_b\bar{S}}\right)$, where f_b is the fraction of salt excluded from the black ice (Table 1). The effective ice thickness, $h^*_{ice} = f_b h_{ice}$, is the equivalent thickness of ice excluding 100% of salt. Fig. 8 shows the effective ice thickness, h^*_{ice}, required to create neutral density from a reverse stratification that extends to $h_r = 1$ m, as a function of average salinity, $\bar{S}$, and deep-water temperature, T_i. If the initial salinity is relatively low, thicker ice is required to overcome the reverse stratification, because it takes more ice to accumulate sufficient salt. As the initial salinity increases, relatively less ice is required.

For Tailings Lake in 2005 with $T_i = 1.5°C$, $h_r = 1$ m, $\bar{S} = 0.96$, and $f_b = 0.99$, the ice thickness required is only 0.04 m. Even for a fresh lake with $\bar{S} = 0.1$ and assuming $h_r = 1$, $T_i = 1$, and $f_b = 0.99$, an ice thickness of only 0.3 m is needed before salt exclusion can potentially play a role in under-ice circulation. For h_r other than 1 m, the appropriate effective ice thickness can be obtained by multiplying that of Fig. 8 by h_r. Thicker ice is required to overcome deeper reverse stratification.

For Tailings Lake, the formation of significant black ice, and the resulting flux of salt beneath the ice, was sufficient not only to break down the reverse stratification but also to redistribute this salt throughout the water column (Fig. 2b). However, it should be noted from the temperature data that the lake was never fully mixed, as it was during fall turnover (Fig. 3). Instead the temperature remained stratified throughout the winter, though at times layers of thickness 1–2 m became isothermal, indicating localized mixing. This suggests that the excluded salt drives successive episodes of convective mixing over limited regions of depth.

The precise mechanism by which salt is redistributed through the water column is difficult to determine because at least four other mechanisms are also potentially active: double diffusion, though characteristic steps in temperature and conductivity were not observed in the CTD casts; boundary effects, such as heat release from the sediments (Welch and Bergmann 1985) or differential ice formation, could effect mixing through interleaving; seepage from Dam 1, which is pumped back into the surface of Tailings Lake throughout the year (0.01 $m^3\ s^{-1}$), could cause localized mixing; and, finally, the continuous release of gas bubbles (predominately hydrogen) from the lake sediments, which may be responsible for large holes (0.5-m diameter) observed in the ice at the south end of the lake, could also contribute to mixing.

Suppression of spring turnover—Spring circulation is usually thought to occur in the following sequence of steps: ice-off; warming of the cold (~0°C) surface layer until its temperature matches that of the deep water, T_d, where $0°C < T_d < T_{MD}$; and spring turnover as the entire lake warms from T_d to T_{MD}. Spring turnover can continue, warming the entire lake above T_{MD}, if windstorms are frequent and strong enough to prevent stratification. The duration of spring turnover depends on the weather; during calm, sunny periods, the surface layer quickly develops sufficient buoyancy to initiate permanent summer stratification.

Tailings Lake behaves differently. At the start of freshet, inflowing snowmelt and spring rain form a freshwater lens under the ice cover; this freshwater lens then warms to ≈7°C before ice-off (e.g., Fig. 3a; 07 Jun 2005, day 524). Water from the partly melted ice likely mixes with this freshwater lens. Just after ice-off, this fresh surface layer is about 3 m deep, as inferred from the temperature data. In order for turnover to occur in Tailings Lake, this less-dense surface layer would have to be mixed with the remainder of the water in the lake.

Several processes can contribute to the mixing of a lower-density surface layer with higher-density deep water: wind generated turbulence from the lake surface can erode the pycnocline and mix deep water with the surface layer; wind stress can generate shear and localized mixing at the pycnocline; wind-driven upwelling can drive mixing between the layers; and surface cooling can induce penetrative convection, which may erode the pycnocline. Using the Monin-Obukhov length scale (Imboden and Wuest 1995), penetrative convection was found to be relatively minor except in the fall, when it played an important role in fall turnover. Spring turnover in Tailings Lake, were it to occur, would differ from fall turnover and would be dominated by wind-driven processes.

Deepening of the pycnocline decreases the total stability (Eq. 2) of the lake. We can estimate the rate at which the total stability, St, of the water column decreases by

$$\frac{dSt}{dt} = -\frac{1}{2} g(\rho_2 - \rho_1) h_1 \frac{dh_1}{dt} \tag{5}$$

Experiments of wind induced upwelling by Wu (1973) and Monismith (1986) suggest that the effect of surface turbulence, upwelling, and shear on the depth of the pycnocline, h_1, can be characterized collectively by

$$\frac{dh_1}{dt} = C_m \frac{u_*^3}{g' h_1} \tag{6}$$

for $W > 1$, which we extrapolate to $W \sim 1$. Monismith (1986) found $C_m = 0.076$ and Wu (1973) found $C_m = 0.23$. Combining Eqs. 5 and 6 yields

$$\frac{dSt}{dt} = -\frac{1}{2} C_m \rho_2 u_*^3 \tag{7}$$

Using Eq. 7, the change in total stability for a windstorm can be estimated. The ability of a given storm to mix the water column increases with duration and with the wind velocity cubed. The spring storm of 15–16 Jun 2005 and three other large summer storms yield values of C_m between the results of Monismith (1986) and Wu (1973).

Next, we wish to explore the stability that salt exclusion imparts to the water column after ice-off as a function of lake salinity and ice thickness. To remove the effect of temperature stratification we define the salinity stability as the stability (Eq. 2) computed at a constant temperature (4°C). The salinity stability of Tailings Lake as a function of effective ice thickness is shown for various salinities in Fig. 9. The salinity stability increases with both salinity and effective ice thickness. The salinity stability is shown over a range (grey band) corresponding to an initial surface layer thickness equal to the effective ice thickness (top of band), and an initial surface layer mixed down to three times the depth of the effective ice thickness as observed in 2005 (bottom of band). The estimated salinity stability at ice-off

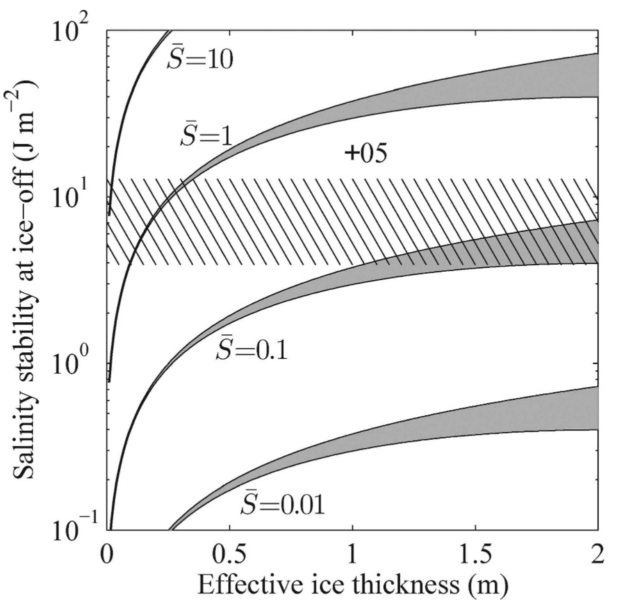

Fig. 9. Salinity stability of Tailings Lake at ice-off as a function of effective ice thickness and average lake salinity $\bar{S}$ = 0.01, 0.1, 1 and 10 g L^{-1}. The grey bands mark the range of likely surface layer depths at ice-off (see text). The hatched region marks the calculated change of total stability as a result of the storm on 15–16 Jun 2005 varying C_m from 0.07 to 0.23. The + marks the salinity stability of Tailings Lake at ice-off in 2005 when $\bar{S}$ = 0.96 g L^{-1}.

on 07 Jun 2005 is in reasonable agreement with the predicted stability, near the lower edge of the $\bar{S} = 1$ band as $\bar{S} = 0.96$ for Tailings Lake.

For the purpose of estimating stability, the effective ice thickness includes not only black ice, but also white ice and freshet runoff that entered Tailings Lake before ice-off. Note that, unlike white ice and freshet inflow, black ice not only contributes to a fresh surface cap but also increases the salinity of the deep water, an effect important in shallow lakes but having only a minor effect in Tailings Lake.

The ability of the storm of 15–16 Jun 2005 to decrease the stability of Tailings Lake was estimated using Eq. 7, and is shown for the range of C_M in Fig. 9 (hatch). By comparing the change in stability for one of the largest storms observed to the salinity stability imparted by the ice-melt, we can predict the following for Tailings Lake. With a salinity of $\bar{S} = 0.01$, spring turnover would be guaranteed, and with $\bar{S} = 10$ spring turnover would not occur for the given storm. For $\bar{S} = 0.1$, turnover would be likely except for thick ice, and for $\bar{S} = 1$ turnover would be unlikely except for thin ice. In this case, the occurrence of spring turnover may vary from year to year. In a year of thick ice, spring turnover would be suppressed, but in another year spring turnover might occur as a result of reduced ice thickness or increased wind velocity, a consequence of either natural variability or climate change. Turnover would require a large windstorm during the short period when salinity alone maintains stability. The procedure above could be used to estimate the likelihood of spring turnover in other lakes given basin geometry, salinity, ice thickness, hydrology, and wind conditions.

Salt exclusion from ice, and the resulting freshwater ice melt, can act to modify the seasonal structure of a brackish lake. Salt exclusion is particularly important in high-latitude lakes, because of the predominant black ice, which excludes a high proportion of salt. The behavior of Tailings Lake is intermediate between freshwater lakes, where spring turnover occurs and saline lakes, which are meromictic. In winter, as a result of excluded salt, the salinity of the entire water column increased. In spring, turnover was suppressed because of the presence of fresh ice melt. In effect, turnover occurred in winter rather than spring, with the result that oxygen was not mixed into the deep water. Estimates of the amount of excluded salt needed to break down the reverse stratification under the ice show that excluded salt can cause under-ice mixing in relatively fresh lakes. Suppression of spring turnover can occur if salinity and ice thickness are high relative to the strength and duration of spring windstorms.

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