Limnol. Oceanogr., 59(6), 2014, 2121–2132

doi:10.4319/lo.2014.59.6.2121

Interannual variability of circulation under spring ice in a boreal lake

Kalevi Salonen,[1,2,*] Merja Pulkkanen,[2] Pauliina Salmi,[1,2] and Ross W. Griffiths[3]

[1] Lammi Biological Station, University of Helsinki, Finland
[2] Department of Biological and Environmental Science, University of Jyväskylä, Finland
[3] Research School of Earth Sciences, Australian National University, Canberra, Australia

Abstract

A small range (~ 1°C) of under-ice water temperature is shown to result in remarkably different circulation regimes under spring ice in a deep, oligotrophic boreal lake. With the water column at < 4°C, melting of snow led to deepening vertical convection before ice break and a final depth of convection inversely correlated with earlier deep-water temperature. We attribute that to the nonlinear dependence of water density on temperature, albeit further affected by stochastic weather factors. In four of nine study years, convection led to complete under-ice overturn of the lake, indicating that this may not be uncommon in similar lakes with steep topography. River inflow and more intensive warming of water in the littoral zone also created a horizontal density differential, convection that involved flow down the sloping bottom and a lateral intrusion of this sinking water at a depth between the vertical convection and the quiescent deep-water layers. The vertical and horizontal convection together produced a profile of temperature slightly increasing from the surface to the bottom of the convection layer. The contribution of horizontal convection to under-ice mixing was interannually variable, and in one of the study years it eventually dominated under-ice mixing. A thermal bar circulation regime developed occasionally and only in the open water between ice and shoreline. We identified five different under-ice mixing regimes that form an interannually variable continuum of behavior during the ice melting period. The dependence on a narrow temperature range likely makes the circulation regime sensitive to a warming climate.

In dimictic lakes, the decreasing air temperature in autumn leads to overturn of the water column. When the maximum density of water (T_{md}) is reached, convective mixing stops. Hence, small lakes regularly freeze over with deep-water temperature close to 4°C. In contrast, mixing in large lakes is dominated by wind and overturn generally continues to further decrease the deep temperatures (Bengtsson 2012). Very low water temperatures can be reached, depending on stochastic weather conditions. In lakes having water volumes so large that heat flux from the sediment does not significantly increase temperature after freezing, winter temperatures as small as 1°C can be occasionally found in ice-covered lakes until spring (Pettersson 1902; Kiili et al. 2009; Twiss et al. 2012). Regardless of lake size, apart from some heat flux to the ice from the uppermost water layers, snow and ice largely isolate the water from atmospheric influences after freezing, and an inverse distribution of temperature develops in the water column (Bengtsson 2012). White snow cover on the lake ice reflects most incident solar radiation so that, during most of winter, heat losses from the surface layers to the atmosphere and from the sediment to water dominate the heat budget (Bengtsson 1996; Jakkila et al. 2009). The gradually weakening heat flux from the sediment warms the overlying water, which, with temperatures < 4°C, slowly flows toward the deepest parts of the lake (Bengtsson et al. 1996). The compensating return flow field increases the temperature of the most of the water column during the dark winter period (Bengtsson et al. 1996; Bengtsson 2012).

Over a relatively short period after the melting of snow cover, solar radiation leads to much faster warming than that due to the heat flux from the sediment (Bengtsson et al. 1996; Petrov et al. 2005; Jakkila et al. 2009). If water temperature is below 4°C, the warming increases water density and leads to vertical (penetrative) convection in which warmer water parcels sink to the depth at which density is similar (Farmer 1975; Bengtsson 1996; Jonas et al. 2003). With time, the depth of convection and the temperature of the convecting zone gradually increase. Convection is generally described as having a four-layered thermal structure (Farmer 1975; Forrest et al. 2008), including the thin topmost boundary layer (maintained by heat loss to the ice and, in the ice melting phase, by low-density meltwater), the main convection layer, an entrainment layer below the convection, and a warmer quiescent layer at the bottom.

The amount of solar radiation penetrating through the ice can be relatively uniform over the lake area, whereas the heat supplied in shallow water is mixed through a smaller water depth than in the offshore regions. Because horizontal distances between water masses with different temperatures are orders of magnitude larger than the depth of the convection layer, a significant temperature gradient can develop between onshore and offshore waters. The warmer (more dense) water from shallow areas, as well as the (at times) denser water from brooks and rivers (Carmack 1979), then flows down the bottom slope until it reaches its own density level and intrudes into the existing stratification. We refer to this largely buoyancy-driven circulation as horizontal convection. Melting of ice limits warming and the development of horizontal convection, but once there is no ice left, the warming rate and circulation are greatly accelerated. In contrast to vertical convection, where circulation is a complex mixture of

* Corresponding author: kalevi.salonen@jyu.fi

small-scale water parcels sinking and rising through the water column (Wells and Sherman 2001), horizontal convection is a coherent and more long-lived flow of water over the top of the sediment, basically similar to the flow created by heat flux from the sediment in midwinter (Malm 1998). While vertical convection has a local nature, horizontal convection may take on a basinwide scale. Horizontal convection is expected to be three-dimensional and also likely to be influenced by planetary rotation: modeling has suggested that, in midwinter, a lake can therefore have a horizontal rotation driven by the heat flux from the sediment (Huttula et al. 2010). Field evidence from an autonomous underwater vehicle of Forrest et al. (2013) in a Canadian lake showed a gyre of smaller scale.

While the convection layer in the central region (of a large and deep lake) deepens and warms, the temperature gradient between the littoral and offshore waters also increases. Eventually, T_{md} is reached near the shoreline, where convection ceases and the stratification spreads toward the lake center. This represents transition from the weakly stratified winter conditions to strongly stratified summer conditions. There are then two water masses with temperatures on either side of T_{md}. A convergence zone, called a thermal bar (e.g., Holland and Kay 2003; Demchenko et al. 2012), develops between these waters with opposite inshore and offshore circulations meeting at the bar parallel to the shoreline. The offshore rotation corresponds to that predicted by the modelling of midwinter under-ice circulation (Huttula et al. 2010). The zone near to T_{md} may effectively isolate warm, shallow water from cooler pelagic water (Holland and Kay 2003). Thermal bars are a complicated phenomenon with their own temporal evolution (Chubarenko and Demchenko 2008).

The water column under the protection of ice cover in winter has traditionally been considered to be quiescent and invariable from year to year. However, an increasing number of physical observations, particularly in spring, have shown that this is not the case (Jonas et al. 2003). Although many basic phenomena influencing under-ice hydrodynamics in lakes are well documented, the detailed flow structure and its effects can be complicated (Forrest et al. 2008). Virtually no attempt has been made to synthesize how different circulation mechanisms are related to each other and how their interactions evolve during winter under variable weather and climate conditions.

We hypothesized that interannual differences in snow and ice conditions are reflected in the evolution of under-ice hydrodynamics during the few weeks preceding ice break. This period is difficult and even dangerous to access and has received little attention. Here we report the results of a multiyear study of a boreal seasonally ice-covered lake, which in the autumn cools below 4°C, and hence its hydrodynamics may be sensitive to interannually different melting phases of lake snow and ice.

Methods

Our study site was an oligotrophic lake, Pääjärvi (61°04′N, 25°08′E), having length 10 km, area 13.4 km^2, and catchment area 229 km^2 (of which 68% is forest, 21% is fields and 10% is bogs; Ruuhijärvi 1974, with data further updated from the database of the Finnish Environment Institute). The lake has rather consistent slopes (average about 2%) downward toward its deepest point (85 m) near the middle of the lake (Fig. 1). Approximately 40% of the lake area is less than 5 m deep, representing 25% of the lake volume. River inflow is modest; hence, water retention time is 3.3 yr. Of the total catchment area, 73% represents the catchment for the three largest rivers (Fig. 1). The lake is regulated for hydropower purposes, which means that in winter the outflow is higher than natural in order to make space for meltwaters. Details of the chemistry of the lake can be found in Ruuhijärvi (1974). Pääjärvi is generally covered by ice for about 5 months of the year.

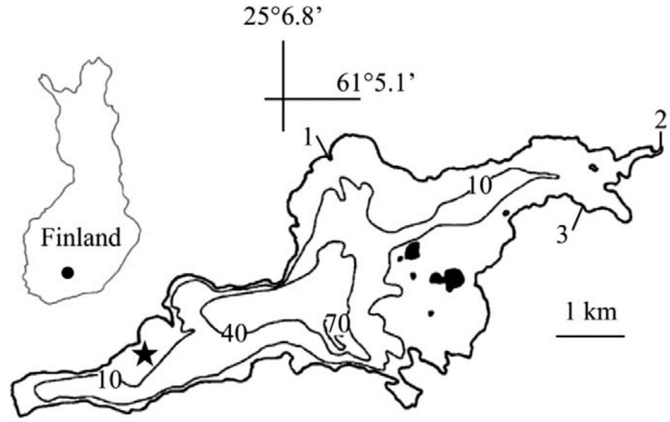

Fig. 1. Bathymetric map of Pääjärvi (13.5 km^2). Major inflow rivers: 1, Haarajoki; 2, Mustajoki; 3, Luhdanjoki.

Temperatures in the inflowing rivers (measured only in 2004 and 2007) and in the lake were measured with Starmon Mini temperature recorders (Star-Oddi, Iceland; accuracy 0.05°C, precision 0.013°C) at either 1 h (in 2004) or half-hour intervals. The recorders were installed (generally before the freezing in each autumn) in the middle of the lake with 5 m (except at 1 m below the surface) depth increments down to 75 m and were recovered after ice break each spring. The data indicated some small systematic differences between recorders, but maximum conformity between recorders was achieved by intercalibrating the results of the five deepest recorders in 2010 and one at 5 m in 2008–2010 against the median temperatures of five factory calibrated recorders at slowly changing temperatures in a circulated water bath over the range of 0–4°C. Temperature measurements for the detection of thermal bars were made with a Yellow Springs Instruments CastAway conductivity, temperature, and depth probe, which also provided the coordinates of the measurement points. Isotherm graphics were produced with MatLab. Mapro animation tool kit (Hemsoft) was used to identify the depth of convection in the construction of figures showing vertical distribution of temperature and for visual verification of ice-break dates. Specific conductivities and spectrophotometrically determined water color from late March to early April, and flow rates of the river Mustajoki were obtained from the database of the Finnish

Environment Institute. Global radiation was measured by the Finnish Meteorological Institute using a pyranometer approximately 90 km to the southwest of Pääjärvi.

Water column stabilities (Schmidt 1928; Idso 1973) were calculated from the temperature recorder data, and heat contents were calculated by multiplying the density of each water layer by its volume, temperature, and specific heat capacity (Johnson et al. 1978).

Results

In winter, approximately 1°C interannual variation of temperature was generally observed at all depths (Fig. 2). Marked warming of water below the shallow surface boundary layer beneath the ice began after the snow cover vanished between late March and early April (Fig. 3). Because water temperatures were consistently below T_{md} of freshwater, the increase in solar radiation reaching the lake surface in spring (Fig. 4) led to warming and increased density of the water and consequent deepening layer of vertical convection (Figs. 2, 3). During the last 4 weeks of ice cover, the water temperature at 5 m depth increased by 0.5–1.5°C which was, with one exception, up to over three times more than the temperature increase at 75 m depth due to the heat flux from the sediment during the preceding 4–5 mo of ice cover. The temperature in the convection layer just before ice break varied between 2.1°C and 3.6°C and was in the early stages vertically uniform (*see*, e.g., 12 d before the ice break in 2010; Fig. 3). In temperature plots with time (Fig. 5), the onset of convection is seen as a departure from originally nearly horizontal lines (such as at 10 m), which successively merged at increasing depths as the temperature of the convection layer increased. In accord with the observations of Farmer (1975), a sudden small drop of temperature was often detected shortly before the water at a given depth joined the convection layer.

Before ice break, the temperature at 75 m depth varied between 2.2°C and 3.45°C, and the depth of the convection layer tended to be greater at the lower end of this temperature range (Pearson r^2 = 0.45, p = 0.047; Fig. 2, summary panel). The correlation with water density calculated from temperature was similar (r^2 = 0.48, p = 0.039). The correlations with thermal stability (r^2 = 0.25, p = 0.17), temperature gradient between 10 and 75 m depths (r^2 = 0.05, p = 0.56), and heat content 1 month before ice-break (r^2 = 0.24, p = 0.18) were insignificant (Fig. 6). Convection generally developed to a maximum of 40–55 m depth, but in the years of coldest water (2004, 2007, 2008, and 2012, when temperatures were 2.2–2.7°C at 75 m before convection), mixing reached the bottom (Fig. 2, summary panel) and increased deep-water oxygen concentration before ice break (Salonen, unpubl.). In these coldest years, the overturn started 1–8 d before ice break. It occurred for both minimum and maximum radiation heat fluxes in the convection period (Fig. 4), and there was no correlation (r^2 = 0.008) between the increase of heat content in the lake (an integrative measure of absorbed solar radiation minus heat consumed in ice melting and evaluated where water depth is greater than 1 m) and maximum mixing depth during the last 4 weeks of ice. In 2006, the maximum energy input from solar radiation during the final 2 weeks of ice cover was 60% higher than the lowest value in 2008, whereas ice break occurred 11 d earlier in the latter year. Thus, water temperature likely affected convective mixing more than increase in heat content. During the last week of ice, including the under-ice overturn years, warming rate of the uppermost 5–20 m water layer was interannually variable (0.04–0.10°C d^{-1}).

The stability of the water column in this oligotrophic lake is dictated mainly by temperature. However, during winter, electrolytes are leaching from the sediment, and the flow of warmer and denser water toward the deepest parts of the lake basin (Pulkkanen and Salonen 2013) confines and concentrates the electrolytes in the limited deepest volume of the lake. Consequently, in late March to early April, before the beginning of convection, the specific conductivity of the water was slightly but consistently higher at 78–79 m depth than at 40 m (mean ratio in the study years 1.031, standard deviation 0.012; Table 1). Total P concentration behaved in a similar manner to specific conductivity, whereas no change could be detected in the very high concentration of total N. Although spring overturn generally reached the bottom of the deepest part of the lake during the first day after ice break, in the calm weather of 2006 (Fig. 7), deepening of the mixed layer from less than 60 m to 75 m took 6 d. In that year, the temperature of the deepest water was one of the highest in the range detected in this study (Fig. 2). The small electrolyte density gradient between 40 m and 78 m (corresponding to a salinity difference of 0.0016, calculated according to McDougall and Barker 2011) was destroyed by wind only when the surface water temperature reached 4°C.

When vertical convection reached roughly 20–25 m depth, 5–19 d before the ice break, the temperature in the convection layer ceased to be uniform but gradually increased downward. The largest gradient of ~ 0.5°C between 1 m and 75 m depths (Figs. 2, 3) was found in 2007. This can be explained by a coherent sinking down the slopes of warm water from the shallow areas of the lake and inflowing rivers and the horizontal intrusion of this water into the interior at its own density level. In contrast to the evolution of the upper convecting layer, temperatures in the horizontal intrusion layer did not trend toward the temperature of the upper layer but instead paralleled it, sometimes remaining up to 0.3°C warmer until the end of the ice cover (Fig. 5) or beyond. In 2007, when late winter temperature at 75 m depth was 2.4°C and the thermal stability was the smallest over the study years (Fig. 6), the horizontal convection was particularly strong and reached the bottom in the deepest parts of the lake about 1 week before the ice break. In that year, snow on the top of the ice melted early, advancing ice break by approximately 2 weeks compared to the other years. Paradoxically, horizontal convection led to increased water column stability during the last week of ice cover (Fig. 5), even though the lake was then overturning. Horizontal convection affected water temperature in the bottom 30 m or more (20–30% of the lake volume) until the end of the ice cover. Thus, we infer

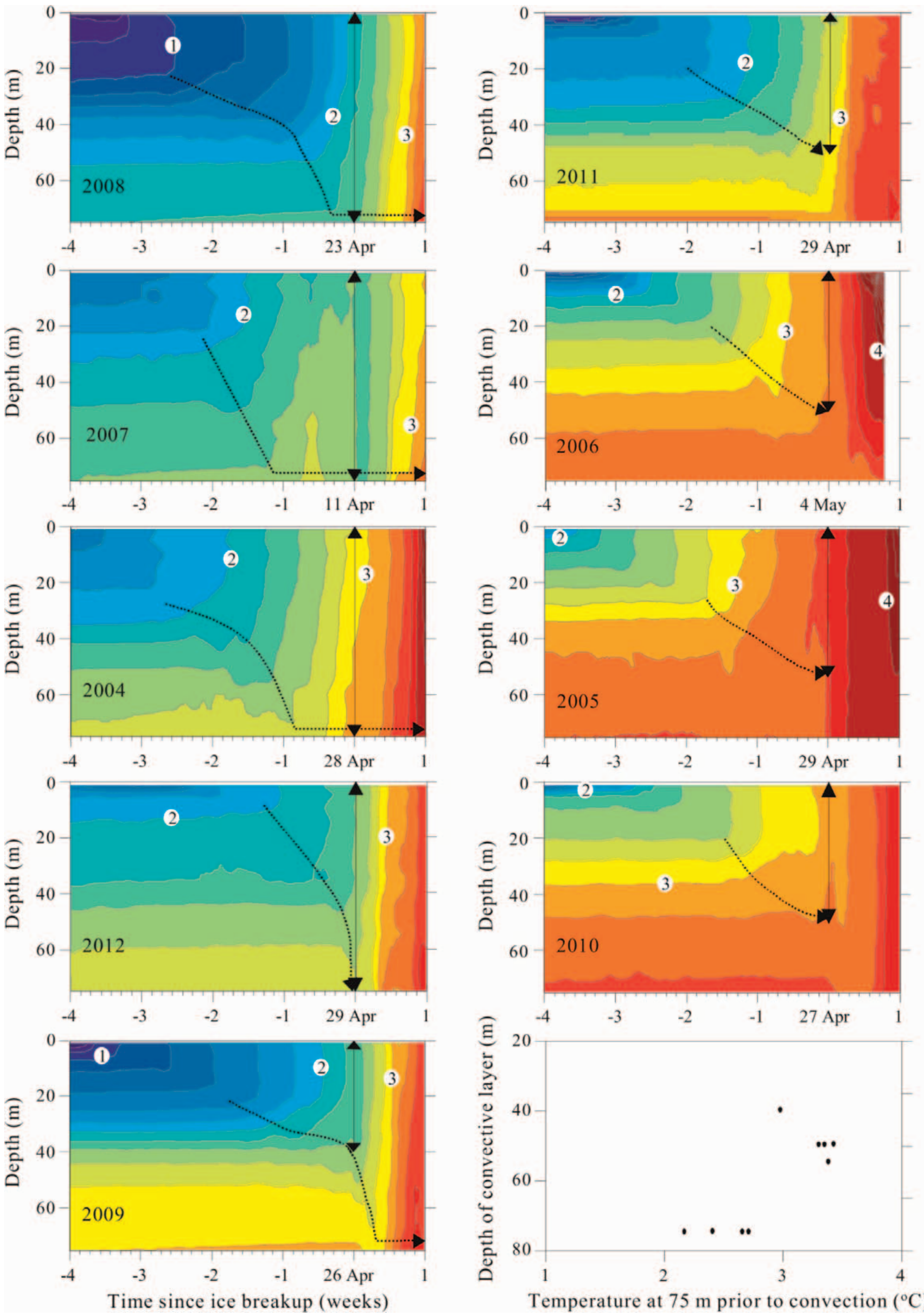

Fig. 2. Water temperature and convective mixing during 4 weeks before and 1 week after the ice break (date given at time 0) in 2004–2012 in the deepest part of Pääjärvi. The isotherm panels with 0.2°C intervals of 24 h moving averages are arranged in the order of increasing water temperature at 75 m 1 month before the ice break. Vertical double arrows, depth of convection at the end of ice cover; dotted arrows, approximate penetration of horizontal convection (for clarity, after reaching 75 m depth, the arrows are drawn with slight upward offset). Lower right panel summarizes the dependence of the depth of convective mixing at ice break on the temperature at 75 m depth before it was affected by convection.

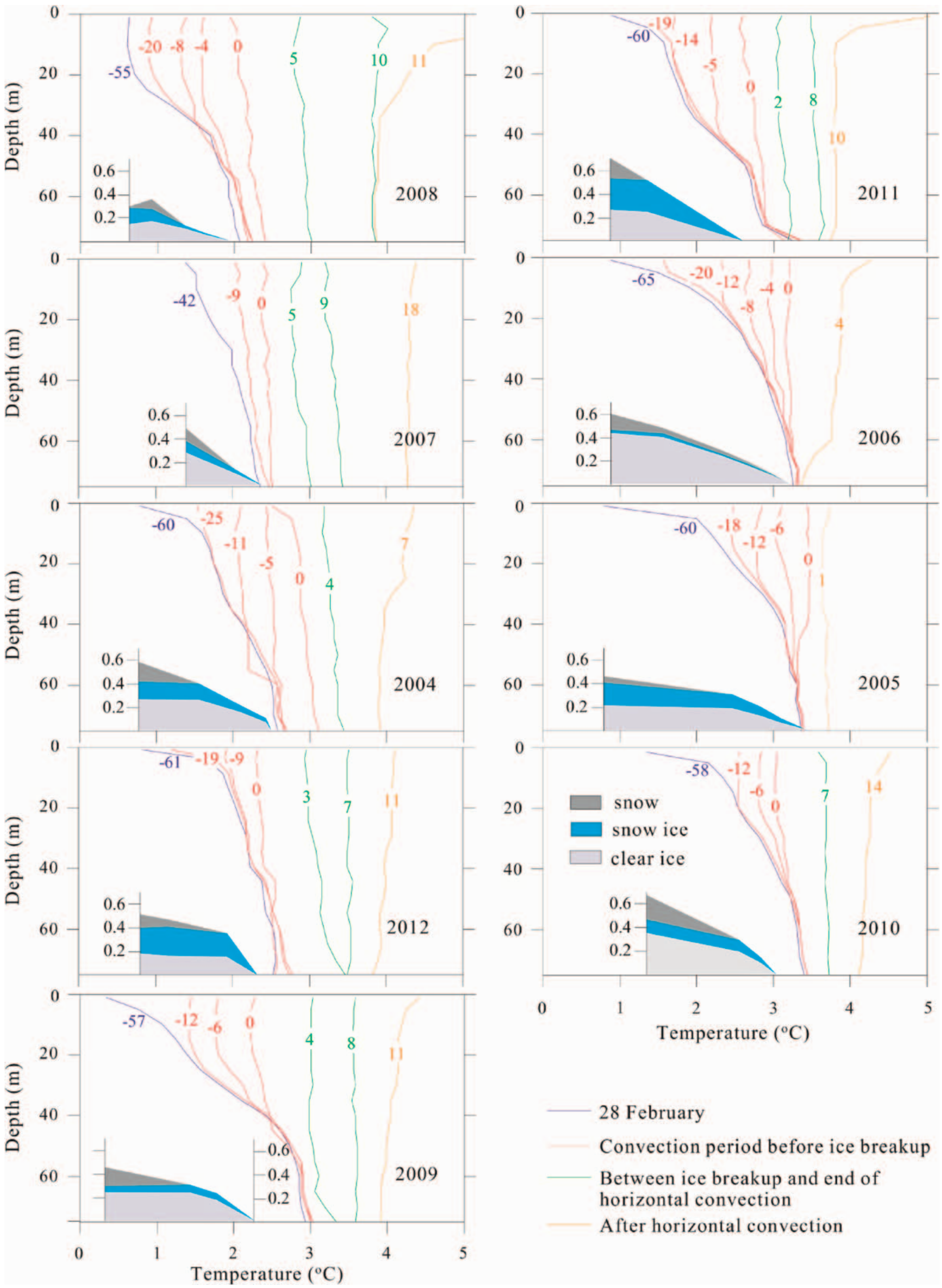

Fig. 3. Evolution of temperature in Pääjärvi between 4 weeks before and 1 week after the ice break. The graphs are generally based on measurements made at noon. Due to the poorly defined nature of ice break, results at 08:00 h (i.e., before the warming during the daytime) were used. Depths of snow and ice layers (in m) are shown with times fixed to the upper ends of each graph. Numbers on profiles indicate days before or after ice break.

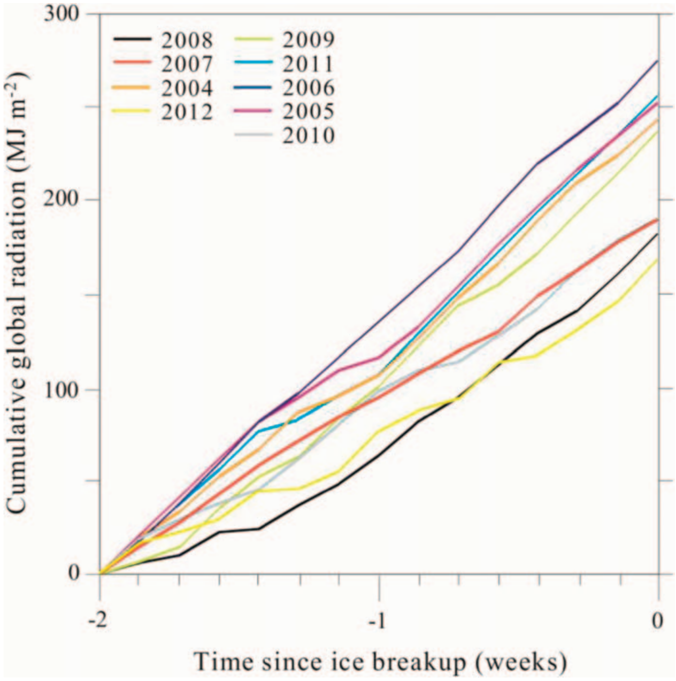

Fig. 4. Cumulative solar radiation during the last 2 weeks of ice. The colors in the key are shown in increasing order of temperature at 75 m depth before it was affected by convection.

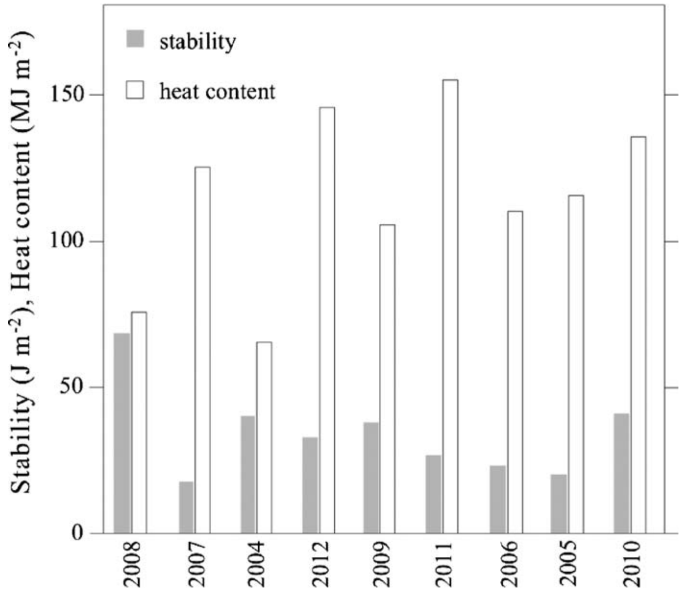

Fig. 6. Schmidt-Idso stability and heat content of the water column of Pääjärvi based on temperature measurements at the deepest part of the lake 1 month before ice break. The years are shown in increasing order of temperature at 75 m depth before it was affected by convection.

that larger water masses were involved in horizontal convection than in any other years. We have no information on the development and extent of the shallow open water zone before the ice break in 2007.

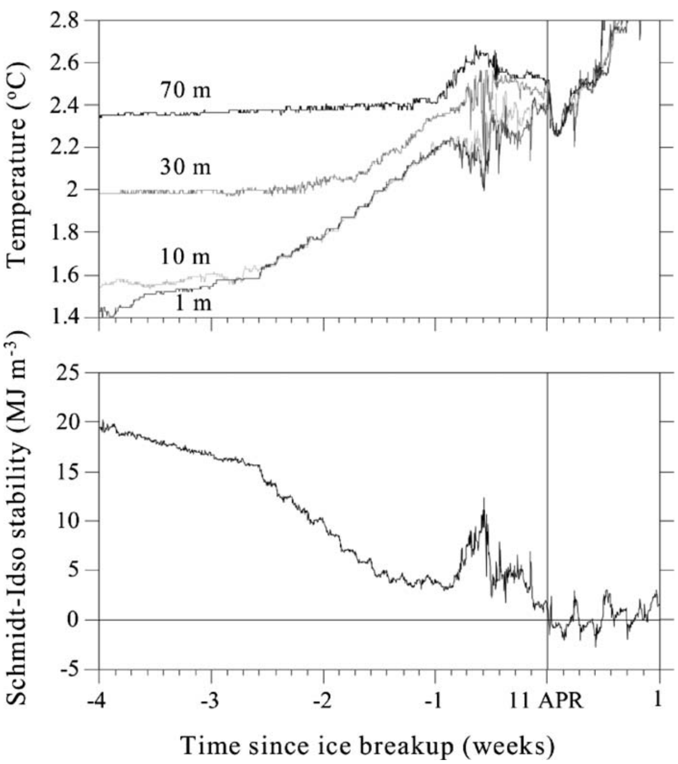

Fig. 5. The development of temperature at selected depths and water column stability under the ice in spring 2007.

Along with faster warming in shallow water, the development of horizontal convection is influenced by inflow to the lake. Inflow data from the Mustajoki river, which carries roughly 40% of the total annual water discharge into the lake, showed an order-of-magnitude interannual variation in transport (Fig. 8). The temperature of this river varied diurnally, but in a smaller river, Luhdanjoki, the amplitude was roughly twice as large (up to 4°C; Fig. 9). More extensive agriculture on the catchment of Luhdanjoki, compared to that for the rivers Mustajoki and Haarajoki (both mostly forested), exposes its catchment to greater effects of solar radiation and thus greater diurnal temperature fluctuations. The agriculture also supports earlier melting of snow compared to forested catchments, explaining the differences in the timing of temperature development (Fig. 9). During the end of ice cover in 2004 and 2007, daytime temperature of the study rivers exceeded 3°C 1–2 weeks before the ice break (Fig. 9), when the penetration of convection reached 40–50 m depth. However, in 2007, the river inflow was only a quarter of that in 2004, yet the effects of the lateral warm intrusion layer at the base of the convecting layer were the most distinct of the four under-ice overturn years. Furthermore, 1–2 weeks before the ice break in 2008, the inflow was 3–10 times larger than in 2007, and this did not result in such a distinct appearance of the horizontal intrusion as seen in 2007. On the other hand, about 2 weeks before the ice break in 2007, there was a temporary temperature maximum in river water that may have contributed to horizontal intrusion.

Due to the high water color, the secchi depth of the lake was small in late March to early April, varying interannually between 1.8 m and 2.4 m (mean 2.1, SD 0.2). Warm groundwater (near 6°C) entering Pääjärvi, on the

Table 1. Mean water color, specific conductivity (25°C), and total nutrients (with standard deviations) in late March to early April 2004–2012. Due to large interannual variation of water color, the values were also standardized according to the annual mean values in the water column, facilitating resolution between the results at different depths.

Depth (m)	Color (mg Pt L^{-1})	Standardized color	Conductivity (mS m^{-1})	Total P (mg m^{-3})	Total N (g m^{-3})
1	74	0.97 (0.09)	9.8 (0.2)	10.0 (1.8)	1.50 (0.05)
40	75	0.96 (0.12)	9.5 (0.1)	9.2 (1.8)	1.42 (0.07)
78–79	81	1.06 (0.10)	9.8 (0.1)	17.9 (2.0)	1.21 (0.09)

other hand, is almost colorless. As there was no sign of dilution of the intense water color in the deepest water during 5 months of winter (Table 1), we assume the significance of groundwater discharge around the end of the ice cover was negligible.

Temperature measurements by recorder chains between the shallow and deep regions made in 2004, 2006, and 2012 provided no indication of a thermal bar. In particular, in 2004, only a negligibly small area of open water developed before the ice break, and as a consequence, the development of a thermal bar was impossible. In contrast, in 2011, two transects made by a conductivity, temperature, and depth probe before the ice break revealed 50-m- and 200-m-wide thermal bars between the ice and littoral water of $> 4°C$ (Fig. 10). At the same time, a horizontal warm intrusion beneath the region of vertical convection was present in the middle of the lake both before and shortly after the ice break, although its effects on the temperature profile (and, by inference, the strength of the circulation) were not different from the other years and did not approach those observed in 2007. A thermal bar had also been observed in 1974 using vertical thermistor measurements. The size of the boat needed to move within the ice prohibited measurements in shallow water, but a 4°C zone

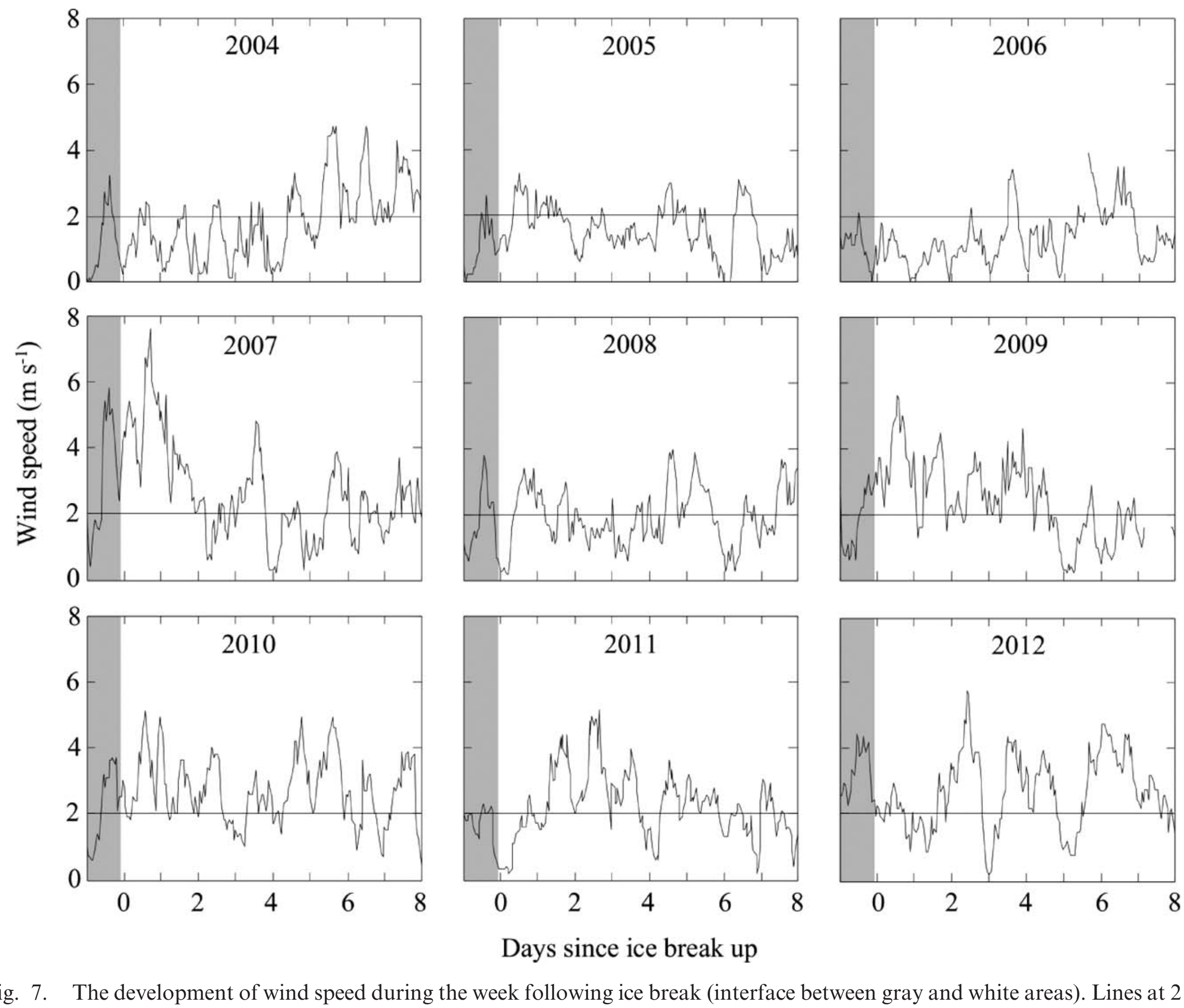

Fig. 7. The development of wind speed during the week following ice break (interface between gray and white areas). Lines at 2 m s^{-1} are shown to facilitate comparison. Data from Lahti District Environmental Services.

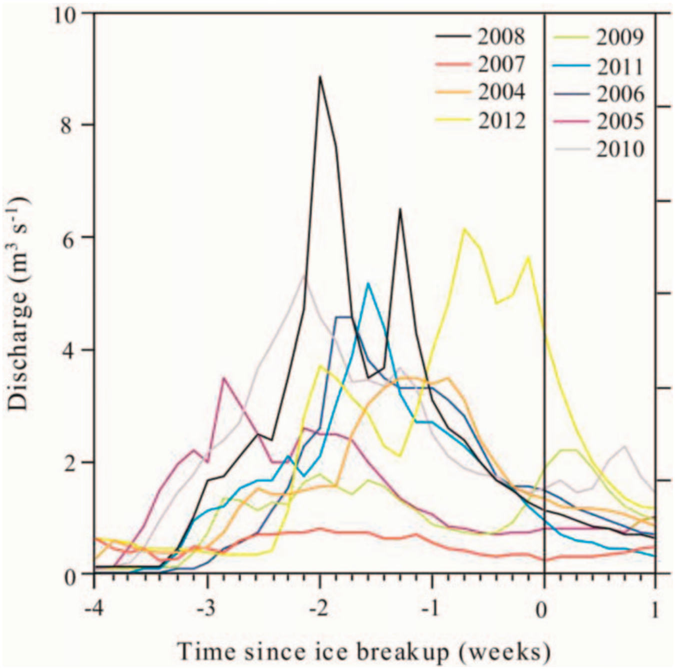

Fig. 8. Discharge of river Mustajoki to Pääjärvi during 4 weeks before and 1 week after the ice break (which is shown as time 0) in 2004–2012. The colors in the key are shown in increasing order of temperature at 75 m depth before it was affected by convection.

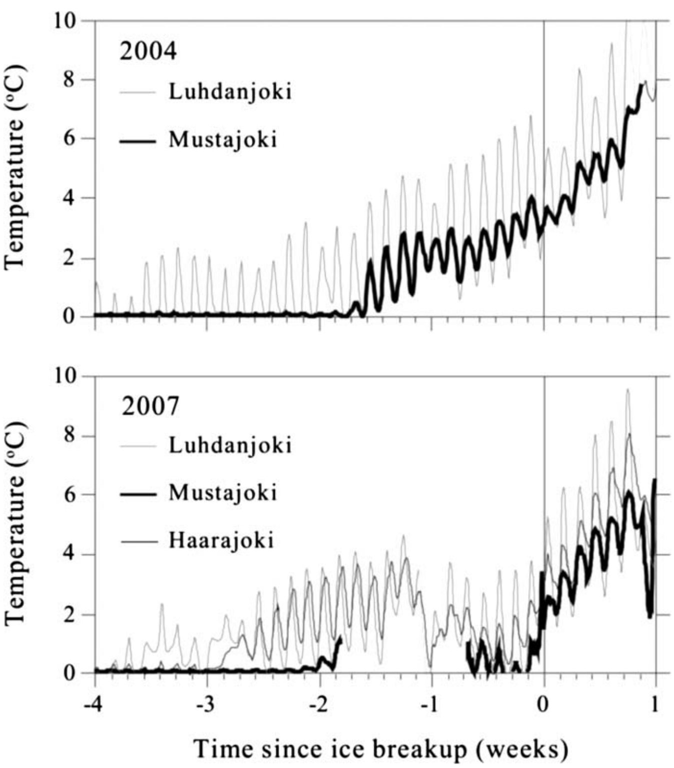

Fig. 9. Variation of water temperatures in the Luhdanjoki, Haarajoki, and Mustajoki rivers before the ice break (which is shown as time 0) in 2004 and 2007. The gap in the graph of Mustajoki is due to a decrease of the water level, exposing the temperature recorder to the atmosphere.

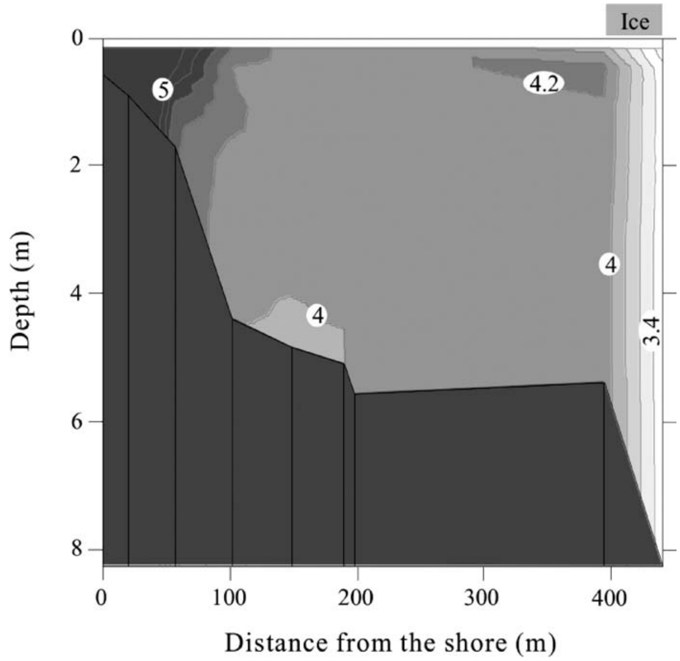

Fig. 10. A thermal bar in Pääjärvi, 28 April 2011. The measurement points are shown by black vertical lines below the isotherms.

of convergence between cold and warm water extending from the surface to the bottom was detected near the margin of the ice at several locations, while the temperature in the under-ice convection layer was 3.3°C.

Discussion

Temperature conditions and hydrodynamics near the end of ice cover in Pääjärvi, a medium-size but deep oligotrophic lake, proved to be far from invariant between years. This was due to stochastic factors such as air temperature in late autumn, snow cover, and solar radiation and snowfall episodes near the end of ice cover. In addition, the variability in the melting of ice cover in shallow areas compared to the deeper central area of the lake complicates the development of lateral water movements and a thermal bar. In some years, the littoral areas can be ice covered almost until the ice break, while in other years, they may be largely open for a couple of weeks before the complete disappearance of ice. In spite of the many factors involved, our results provide strong evidence that the water temperature reached during the previous autumnal overturn is the single dominant factor affecting the amplitude and type of under-ice spring convection regime in Pääjärvi.

Variability of vertical convection—In Lake Pääjärvi, the relatively large deep-water volume reduces the warming of the water column by heat flux from the sediment during the winter small (generally $< 0.5°C$). Hence, the deep-water temperature before the melting of snow in late winter is largely set at freeze-over. Due to nonlinear dependence of water density on temperature, similar changes in temperature result in different changes in density. A 0.5°C increase from 2°C changes water density approximately 2.5 times more than the respective increase from 3°C. While the lake is under ice, it is isolated from wind forcing, and small differences in water density can result in gravity currents between shallow- and deep-water regions and between river mouths and the bulk of the water column. Interannual differences in the resulting density gradients and their timing can mask under-ice mixing depth seemingly independent from the level of radiative flux (Fig. 4). The narrow range of the temperatures explains why there was no significant relationship between increases in lake heat content and the depth of convection during the last weeks of ice cover. Our observations are in line with those from Antarctica (Matthews and Heaney 1987), Alaskan (Vincent et al. 2008) lakes, and the Ristinselkä Basin of nearby Lake Päijänne. The latter lake is deeper (94 m) and an order of magnitude larger than Pääjärvi, and in a year, when Kiili et al. (2009) observed a long under-ice overturn period, its deep-water temperatures approached 1°C. Lower temperature implies a still larger thermal expansivity of the water that further increased the density differences. According to data of the Finnish Environment Institute, temperatures lower than 2°C occured in the deep water of the Ristinselkä Basin of Lake Päijänne 16 times in 50 yr, while in Pääjärvi, it was only five times. Hence, the large size may favor the occurrence of low temperatures and large depth of convection in lakes with steep slopes. In general, cold monomictic polar lakes tend to have lower winter temperature than boreal lakes (Vincent et al. 2008), and their ice break also occurs close to midsummer, when daily heat input from solar radiation is strongest (and, at the lake surface, not much different from daily solar fluxes at lower latitudes), thus increasing the effects of lake warming. In contrast, deep-water temperatures in smaller boreal and temperate lakes approach 4°C, where the expansivity is smallest and under-ice overturn hardly occurs.

The delay of overturn after the ice break in 2006 suggests that very small vertical differences in salinity can significantly limit vertical mixing under ice. Therefore, river waters with higher temperature and electrolyte concentration possibly also contributed to the development of large-scale horizontal convection observed in 2007. However, low river discharges in that year compared to the other years suggest that warming of water in the littoral region was probably responsible for the bulk of horizontal convection. Significant leaching of salts from the sediment into overlying water flowing from the littoral regions during the short period of under-ice convection is unlikely. There were marked differences in specific conductivities of water (L. Arvola, unpubl.) between the main rivers discharging to Pääjärvi (9.3, 10, and 14.8 mS m^{-1} in early April 2013) and between different phases of snow melting (decrease of 32–44% during the first 3 weeks of April). The latter is in line with the observations of Johannessen and Henriksen (1978) that the first meltwaters have higher concentrations of electrolytes than the latest ones. Hence, along with water temperature, the variation in the progress of snow melting on the catchment area can affect the density of river waters and further add to the variation of under-ice spring circulations, particularly in the intermediate stage of ice melting. Although the volume of water supplied by the main rivers to Pääjärvi 2 weeks before the overturn in 2007, when horizontal convection was most distinct, was relatively small ($\sim$ 0.7% of the total volume of the lake), its proportion in water volume below 50 m depth was much higher ($\sim$ 22%). Therefore, it is possible that under suitable weather conditions in late winter and with low water column temperature, river inflow may have affected chemical stratification and triggered a large-scale horizontal convection regime.

In oligotrophic Pääjärvi, the dissolved oxygen concentration in the deep water was never depleted, implying only small leaching of electrolytes from the sediment (Table 1) and hence a small effect of electrolytes on water density and convection. The larger accumulation of electrolytes in eutrophic lakes during winter tends to inhibit deep penetration of convection, as shown by Matthews and Heaney (1987) and Mironov et al. (2002). In oligotrophic lakes, the convection can increase deep-water oxygen concentration and bring deep-water nutrients into the photic zone. Instead, in eutrophic lakes, partial erosion of the anoxic deep-water layer can significantly reduce oxygen concentrations in the convection layer, with consequences for fish and other biota.

The frequent (4 out of 9 yr) occurrence of convection reaching the bottom of Pääjärvi before ice break indicates that this behavior might not be rare in oligotrophic boreal lakes. In large but relatively shallow lakes, where small water volume results in high autumn cooling rates, the variation in water column temperature at the time of freezing can be large, and occasional very low water column temperatures are expected. However, due to the small ratio of water volume to area in those lakes, the low autumn temperatures can be largely offset by heat flux from the sediment during winter (Terzhevik et al. 2009). The smaller bottom slope of shallow lakes also minimizes horizontal convection. Long-term studies similar to that reported here but of large lakes with variable depth are needed to build a broader understanding of under-ice circulations and possible under-ice overturn.

Horizontal convection and thermal bar—The primary indication of horizontal convection in Pääjärvi that temperatures smoothly increased toward the bottom of the convection layer (Fig. 3) was similar to that found in Lake Päijänne (Kiili et al. 2009). It is expected that the interior intrusion will be modified through both shear-driven entrainment of the overlying layer and, during the daytime, erosion by penetrative convection from above. Hence, the full details of horizontal convective circulation could be revealed only by a larger number of diurnal observations made laterally across the gradient of water depth, including river mouths.

The lake-scale horizontal convection as detected in 2007 is in many ways similar to the lateral circulation driven by a uniform surface cooling and wind-induced mixing in lakes of variable water depth (Wells and Sherman 2001) and on the Mediterranean shelf (Gulf of Lions; Durrieu de Madron et al. 2013). In both of these cases, the water is above 4°C (or is seawater for which the maximum density occurs at the freezing point), and the circulation tends to take dense water down the slopes, stratifying the water column in opposition to the action of the vertical convective mixing from the surface, which tends to destroy stratification (Wells et al. 1999). In those cases, as well as in the present freshwater case at low temperature, the circulation process can be seen as a class of developing, transient horizontal convection (Hughes and Griffiths 2008; Griffiths et al. 2013). The sloping bottom plays no substantive dynamical role apart from giving rise to a variable water depth that, in turn, leads to horizontal density differences under a uniform surface buoyancy flux. Horizontal convection will be enhanced by the increased spatial contrast in heat input following the disappearance of ice in shallower water (which decreases the albedo, reduces the latent heat loss to melting, and avoids the possibility of a new light-scattering snow layer on the remaining ice). The circulation is then expected to approach the case of horizontal convection driven primarily by differential surface heat fluxes (Mullarney et al. 2004). Horizontal convection involves a mean circulation, implying effective long-range transport, and has also been shown to be extremely efficient at mixing (Gayen et al. 2013) when compared to the relatively disorganized motions of vertical convection (Hughes et al. 2013). The role of local vertical mixing will always remain an essential one, but it is expected to decrease in a comparative sense as more of the heat added to the surface water is carried to the relatively localized sinking regions in a large-scale horizontal overturning.

Our results indicate that thermal bars are not necessarily associated with either horizontal convection or under-ice overturn. This is consistent with the fact that density differences driving lateral movements of water are smaller, between 3°C and 4°C, than they are at the lower temperatures. We observed the presence of a spring thermal bar in Pääjärvi in 2 yr and its absence in 3 yr, and therefore a thermal bar probably forms in this lake only occasionally. Due to the large variability of daily wind speed (Fig. 7), air temperature, and solar radiation (Fig. 4), taken together with the small water mass in the ice-free shallow water, thermal bars are also likely ephemeral, a conclusion consistent with the absence of previous reports of a thermal bar in lakes of similar size. In contrast, in a very large lake, Ladoga, a distinct thermal bar regularly persists after ice break until midsummer (Malm et al. 1994). Because of a much smaller water mass (by orders of magnitude) in Pääjärvi, wind can more easily mix water masses, and it is likely that only horizontal convection can develop after ice break (Fig. 3).

A continuum of under-ice circulation regimes—Apart from the narrow boundary layer next to ice, water temperatures in small lakes are near 4°C at the beginning of winter. If throughflow is insignificant, only a very shallow convection layer develops in late winter, horizontal convection does not exist, and interannual variation is negligible (Fig. 11A). Thus, the under-ice hydrodynamics of small lakes are relatively predictable. In contrast, larger

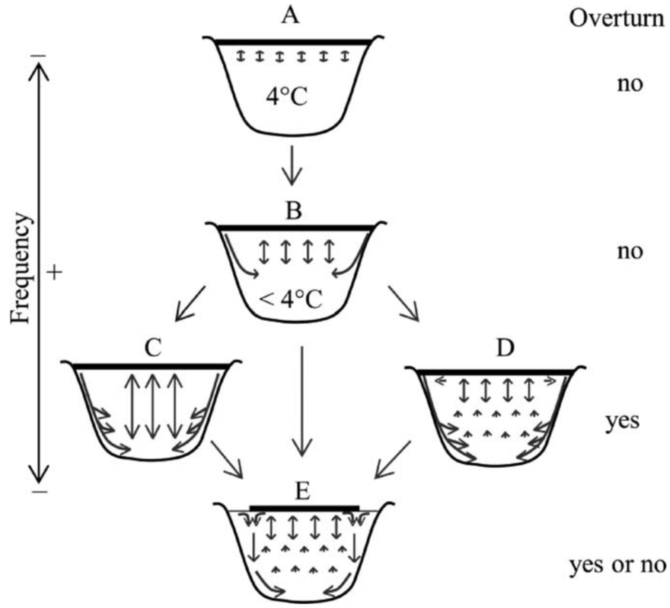

Fig. 11. Continuum of under-ice circulation in lakes evolving from (A–C) a dominance of vertical convection to (D) a dominance of horizontal convection and (in some years) finally to (E) a thermal bar regime. Black bar at the top shows the ice cover. An indicative illustration of the frequency of the circulation regimes for Lake Pääjärvi (A–E) is shown by a double arrow.

lakes cooling below 4°C exhibit high inter- and intra-annual variation. Although detailed temperature results from such lakes in late winter are limited to only a few examples, they include a sufficient range of conditions and behaviors that we attempt to place the dynamics in a generalized continuum in which important governing variables are the difference of water temperature from that of maximum density and the vertical distribution of salt concentration. Weeks before the ice break, the increasing penetration of solar radiation into the water increases the depth of vertical convection, which may dominate until the end of the ice cover in lakes with sufficiently high deep-water temperature, salt concentration, or low bottom slope (Fig. 11B). In some years, the lower temperature near the bottom in oligotrophic lakes leads to stronger vertical convection and completes overturn before the ice break (Fig. 11C). In lakes with sufficient bottom slopes, more rapid warming in shallow regions and warmer river inflow can start basin-scale horizontal convection that works in parallel with vertical convection and can become the primary mechanism to create under-ice overturn (Fig. 11D), as we observed in 2007 in Pääjärvi. Although eutrophic lakes typically have significant vertical salinity gradients that inhibit deep-water convection, in the upper part of their water column they are likely to have similar circulation regimes as in Pääjärvi. In the final stage of spring circulation, some years show extensive loss of ice cover in shallow areas before the central parts of the lake, causing a thermal bar to develop (Fig. 11E), and in really large lakes, this becomes the dominating circulation regime in spring.

Our results have two significant wider implications. First, the observed sensitivity of hydrodynamics near the end of ice cover to small differences in water column temperature imply that climate warming is likely to affect under-ice mixing regimes and frequency of under-ice overturn in large lakes. Indeed, the simulations of Saloranta et al. (2009) predicted that delay in the cooling and freezing of Pääjärvi closer to the winter solstice favors freezing at lower water column temperatures. Second, the large interannual variation of mixing and circulation regime is likely to cause substantial variability in the spring growth of phytoplankton in boreal lakes, where high spring biomass often develops before ice break (Pechlaner 1970; Vehmaa and Salonen 2009). Vertical convection brings more nutrients from deep water, while horizontal circulation leads to a bulk transport of nutrients, oxygen, and biota between littoral and deep zones. A thermal bar, which we suggest is ephemeral in lakes of medium size, separates shallow- and deep-water masses and may maintain more favorable light and nutrient conditions for the early development of phytoplankton in the shallow zone or at the thermal bar (Holland and Kay 2003). Hence, different circulation regimes under the ice cover are expected to have different effects on the development of the spring phytoplankton maximum.

Acknowledgments

This research was supported by the Academy of Finland (grant 104409), the Kone Foundation, and the Australian Research Council (grant DP120102744). We thank the anonymous reviewers for their constructive comments.

References

Bengtsson, L. 1996. Mixing in ice-covered lakes. Hydrobiologia **322:** 91–97, doi:10.1007/BF00031811

———. 2012. Ice covered lakes, p. 357–360. *In* L. Bengtsson, R. W. Herschy, and R. W. Fairbridge [eds.], Encyclopedia of lakes and reservoirs. Springer.

———, J. Malm, A. Terzhevik, M. Petrov, P. Boyarinov, A. Glinsky, and N. Palshin. 1996. Field investigation of winter thermo- and hydrodynamics in a small Karelian lake. Limnol. Oceanogr. **41:** 1502–1513, doi:10.4319/lo.1996.41.7.1502

Carmack, E. C. 1979. Combined influence of inflow and lake temperatures on spring circulation in a riverine lake. J. Phys. Oceanogr. **9:** 422–434, doi:10.1175/1520-0485(1979)009<0422:CIOIAL>2.0.CO;2

Chubarenko, I. P., and N. Yu. Demchenko. 2008. Laboratory modeling of the structure of a thermal bar and related circulation in a basin with a sloping bottom. Oceanology **48:** 327–339, doi:10.1134/S0001437008030041

Demchenko, N., I. Chubarenko, and G. van Heijst. 2012. On the fine structure of the thermal bar front. Environ. Fluid Mech. **12:** 161–183, doi:10.1007/s10652-011-9223-2

Durrieu de Madron, X., and others. 2013. Interaction of dense shelf water cascading and open-water convection in the northwestern Mediterranean during winter 2012. Geophys. Res. Lett. **40:** 1379–1385, doi:10.1002/grl.50331

Farmer, D. M. 1975. Penetrative convection in the absence of mean shear. Quart. J. R. Meteorol. Soc. **101:** 869–891, doi:10.1002/qj.49710143011

Forrest, A. L., B. E. Laval, R. Pieters, and D. S. S. Lim. 2008. Convectively driven transport in temperate lakes. Limnol. Oceanogr. **53:** 2321–2332, doi:10.4319/lo.2008.53.5_part_2.2321

———, ———, ———, and ———. 2013. A cyclonic gyre in an ice-covered lake. Limnol. Oceanogr. **58:** 363–375, doi:10.4319/lo.2013.58.1.0363

Gayen, B., R. W. Griffiths, G. O. Hughes, and J. A. Saenz. 2013. Energetics of horizontal convection. J. Fluid Mech. Rapids **716:** R10, doi:10.1017/jfm.2012.592

Griffiths, R. W., G. O. Hughes, and B. Gayen. 2013. Horizontal convection dynamics: Insights from transient adjustment. J. Fluid Mech. **726:** 559–595, doi:10.1017/jfm.2013.244

Holland, P. R., and A. Kay. 2003. A review of the physics and ecological implications of the thermal bar circulation. Limnologica **33:** 153–162, doi:10.1016/S0075-9511(03)80011-7

Hughes, G. O., B. Gayen, and R. W. Griffiths. 2013. Available potential energy in Rayleigh-Benard convection. J. Fluid Mech. **729:** R3, doi:10.1017/jfm.2013.353

———, and R. W. Griffiths. 2008. Horizontal convection. Annu. Rev. Fluid Mech. **40:** 185–208, doi:10.1146/annurev.fluid.40.111406.102148

Huttula, T., M. Pulkkanen, B. Arkhipov, M. Leppäranta, V. Solbakov, K. Shirasawa, and K. Salonen. 2010. Modelling circulation in an ice-covered lake. Est. J. Earth. Sci. **59:** 298–309, doi:10.3176/earth.2010.4.06

Idso, S. B. 1973. On the concept of lake stability. Limnol. Oceanogr. **18:** 681–683, doi:10.4319/lo.1973.18.4.0681

Jakkila, J., M. Leppäranta, T. Kawamura, K. Shirasawa, and K. Salonen. 2009. Radiation transfer and heat budget during the ice season in Lake Pääjärvi, Finland. Aquat. Ecol. **43:** 681–692, doi:10.1007/s10452-009-9275-2

Johannessen, M., and A. Henriksen. 1978. Chemistry of snow meltwater: Changes in concentration during melting. Water Resour. Res. **14:** 615–619, doi:10.1029/WR014i004p00615

Johnson, N. M., J. S. Eaton, and J. E. Richey. 1978. Analysis of five North American lake ecosystems. II Thermal energy and mechanical stability. Verh. Int. Verein. Limnol. **20:** 562–567.

Jonas, T., A. Y. Terzhevik, D. V. Mironov, and A. Wüest. 2003. Radiatively driven convection in an ice-covered lake investigated by using temperature microstructure technique. J. Geophys. Res. **108:** 3183, doi:10.1029/2002JC001316

Kiili, M., M. Pulkkanen, and K. Salonen. 2009. Distribution and development of under-ice phytoplankton in 90-m deep water column of Lake Päijänne (Finland) during spring convection. Aquat. Ecol. **43:** 707–713, doi:10.1007/s10452-009-9262-7

Malm, J. 1998. Bottom buoyance layer in an ice-covered lake. Water Resour. Res. **34:** 2981–2993, doi:10.1029/98WR01904

———, D. Mironov, A. Terzhevik, and L. Jönsson. 1994. Investigation of the spring thermal regime in Lake Ladoga using field and satellite data. Limnol. Oceanogr. **39:** 1333–1348, doi:10.4319/lo.1994.39.6.1333

Matthews, P. C., and S. I. Heaney. 1987. Solar heating and its influence on mixing in ice-covered lakes. Freshw. Biol. **18:** 135–149, doi:10.1111/j.1365-2427.1987.tb01302.x

McDougall, T. J., and P. M. Barker. 2011. Getting started with TEOS-10 and the Gibbs Seawater (GSW) Oceanographic Toolbox. SCOR/IAPSO WG127.

Mironov, D., A. Terzhevik, G. Kirillin, T. Jonas, J. Malm, and D. Farmer. 2002. Radiatively driven convection in ice-covered lakes: Observations, scaling and a mixed layer model. J. Geophys. Res. **107:** 1–16, doi:10.1029/2001JC000892

Mullarney, J. C., R. W. Griffiths, and G. O. Hughes. 2004. Convection driven by differential heating at a horizontal boundary. J. Fluid Mech. **516:** 181–209, doi:10.1017/S0022112004000485

Pechlaner, R. 1970. The phytoplankton spring outburst and its conditions in Lake Erken (Sweden). Limnol. Oceanogr. **15:** 113–130.

Petrov, M. P., A. Yu. Terzhevik, N. I. Palshin, R. E. Zdorovennov, and G. E. Zdorovennova. 2005. Absorption of solar radiation by snow-and-ice cover of lakes. Water Resour. **32:** 496–504, doi:10.1007/s11268-005-0063-7

Pettersson, O. 1902. Resultaten af den internationella undersökningen af norra Europas djupa sjöar och innanhaf år 1900 [Results of the international research of European deep lakes in 1900]. Bihang till Kungliga Svenska Vetenskaps-Akademiens Handlingar **28**(Afd. II 2):1–18.

Pulkkanen, M., and K. Salonen. 2013. Accumulation of low oxygen water in deep waters of ice-covered lakes cooled below 4°C. Inland Waters **3:** 15–24, doi:10.5268/IW-3.1.514

Ruuhijärvi, R. 1974. A general description of the oligotrophic lake Pääjärvi, southern Finland, and the ecological studies on it. Ann. Bot. Fennici **11:** 95–104.

Saloranta, T. M., M. Forsius, M. Järvinen, and L. Arvola. 2009. Impacts of projected climate change on the thermodynamics of a shallow and a deep lake in Finland: Model simulations and Bayesian uncertainty analysis. Hydrol. Res. **40:** 234–248, doi:10.2166/nh.2009.030

Schmidt, W. 1928. Űber Temperatur und Stabilitätsverhältnisse von Seen [Temperature and stability conditions in lakes]. Geogr. Ann. **10:** 145–177.

Terzhevik, A., and others. 2009. Some features of the thermal and dissolved oxygen structure in boreal, shallow ice-covered Lake Vendyurskoe, Russia. Aquat. Ecol. **43:** 617–627, doi:10.1007/s10452-009-9288-x

Twiss, M. R., and others. 2012. Diatoms abound in ice-covered Lake Erie: An investigation of offshore winter limnology in Lake Erie over the period 2007 to 2010. J. Gt. Lakes Res. **38:** 18–30, doi:10.1016/j.jglr.2011.12.008

Vehmaa, A., and K. Salonen. 2009. Development of phytoplankton in Lake Pääjärvi (Finland) during under-ice convective mixing period. Aquat. Ecol. **43:** 693–705, doi:10.1007/s10452-009-9273-4

Vincent, W. F., S. MacIntyre, R. H. Spigel, and I. Laurion. 2008. The physical limnology of high-latitude lakes, p. 65–81. *In* W. F. Vincent and J. Laubourn-Parry [eds.], Polar lakes and rivers. Limnology of Arctic and Antarctic aquatic ecosystems. Oxford University Press, doi:10.1093/acprof:oso/9780199213887.003.0004

Wells, M. G., R. W. Griffiths, and J. S. Turner. 1999. Competition between distributed and localised buoyancy fluxes in a confined volume. J. Fluid Mech. **391:** 319–336, doi:10.1017/S0022112099005248

———, and B. Sherman. 2001. Stratification produced by surface cooling in lakes with significant shallow regions. Limnol. Oceanogr. **46:** 1747–1759, doi:10.4319/lo.2001.46.7.1747

Associate editor: Craig L. Stevens

Received: 30 November 2013
Accepted: 01 August 2014
Amended: 09 September 2014