The influence of ice season on the physical and ecological conditions in Lake Vanajanselkä, southern Finland

Matti Leppäranta, Anniina Heini, Elina Jaatinen and Lauri Arvola

ABSTRACT

The results from a winter field study in 2009–2010 in Lake Vanajanselkä are presented. This lake is shallow, eutrophic, and annually ice-covered on average for 5 months. The ice sheet was 41–55 cm thick at the annual maximum, consisting of snow-ice and congelation ice. The e-folding depth of light intensity was 50–100 cm for congelation ice and 5–10 cm for snow. The water body had a 4-m thick upper mixed layer and lower continuously stratified layer. Fall cooling process was crucial to determine the temperature of the lower layer at freeze-up, anything within 0–4 °C. Oxygen concentration decreased in winter, especially close to the bottom sediments, and carbon dioxide concentration increased due to respiration activity. Phytoplankton production and biomass level were low or very low and, therefore, heterotrophic and mixotrophic species were abundant. Oxygen depletion in the hypolimnium had several chemical and ecological consequences, such as release of phosphorus from the bottom sediments. In spring, just before the ice-out, photosynthesis was at a high level beneath the ice due to improved light conditions and started to elevate the oxygen concentration in the topmost water layer.

Key words | ice thickness, lake ice, light transfer, under-ice chemical conditions, under-ice thermal conditions, under-ice phytoplankton

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INTRODUCTION

In the boreal zone, lakes are normally covered by ice in winter. Freezing of the surface water changes a lake into a different physical and ecological state as compared with the open water season (Bengtsson *et al.* 1996; Leppäranta 2009). Due to the ice cover, interaction between the atmosphere and lake water body is strongly reduced. The water temperature is inversely stratified with the bottom temperature between 0 and 4 °C depending on autumn mixing and heat flow from the lake sediments (Bilello 1968; Malm *et al.* 1997). The ice cover is static, apart from very large lakes and, in consequence, the circulation in the water body is thermohaline, forced by the bottom heat flux and in spring also by the solar radiation (Malm *et al.* 1998; Jonas *et al.* 2003; Huttula *et al.* 2010).

Heat loss from the lake to the atmosphere is very slow because of the insulation capacity of snow and ice (Hamblin & Carmack 1990; Jakkila *et al.* 2009). In shallow lakes, the bottom sediment accumulates significant heat storage in summer (Fang & Stefan 1996; Terzhevik *et al.* 2009). In winter, the water temperature is fairly stable under the ice, and the light intensity is low (Arst *et al.* 2006; Lepistö & Arvola unpublished). Light transfer through the ice cover is largely restricted by snow on ice. This results in very low and vanishing levels of light beneath the ice in mid-winter as long as dry snow is present, bringing 'polar night' also to sub-polar lakes. Full ice cover also prevents exchange of gases and particles between the atmosphere and the water body. After snowmelt in spring, solar radiation penetrates the ice and initiates convection.

Life in ice-covered lakes is limited by the temperature, light and oxygen. The low temperature slows down biological processes but does not stop them. However, primary production ceases due to absence of light in mid-winter when the snow cover is thick enough. Renewal of dissolved

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oxygen is weak under ice and stops with the primary production closing down (e.g., Golosov et al. 2006). Therefore, oxygen level decreases under the ice cover and, especially in eutrophic lakes, anoxic conditions may result. Oxygen depletion may have serious chemical and ecological consequences, such as release of phosphorus from bottom sediments and fish kills with subsequent food web impacts. In spring, after snowmelt, photosynthesis can be high beneath ice due to improved light conditions. In very shallow lakes, volumetric changes associated with the formation of ice may significantly reduce the volume of liquid water.

Lake Vanajanselkä is a large, shallow and eutrophic boreal lake in southern Finland. It belongs to the drainage basin of the River Kokemäenjoki with outflow to the Gulf of Bothnia, Baltic Sea on the Finnish west coast. It is regionally important for fishery, traffic and recreation, and the water quality is of major concern to the local community. The lake has a long eutrophication history with very poor water quality in the 1960s and 1970s (Kansanen 1981). Vanajanselkä freezes over annually, and the average length of the ice season is 150 days, with a range of 86–171 days (Korhonen 2005). Almost no studies have been made on wintertime processes in Lake Vanajanselkä. Water quality, circulation, oxygen and primary production are the major winter questions to be asked, and also the full annual cycle of the lake with linkages between seasons needs to be better understood.

A 3-year research project, YMPANA, was carried out in Lake Vanajanselkä during 2008–2010. The objectives were to examine the all-year functioning of the physics, chemistry and ecosystem in the lake and to develop an automatic near real-time lake monitoring system for the local communities. Field experiments were performed in all seasons, and special attention was paid on winter investigations because of the limited knowledge of this season (Jaatinen et al. 2010). Also, modelling work has been done for the winter season, for ice growth and decay (Yang et al. 2012) and water circulation dynamics. This paper presents results of YMPANA wintertime field investigations, focusing on the physical and geochemical conditions and the ecological state. The work is based on field campaigns performed in winters 2009 and 2010.

METHODS

Lake Vanajanselkä is located in southwestern Finland, in the rectangle $61°00'–61°20'$ N $\times$ $24°00'–24°30'$ E (Figure 1). It is a large, shallow and eutrophic lake. The surface area is 103 km^2, the mean and maximum depths are 7.7 and 24 m, respectively, and the catchment area is 2,774 km^2. The water balance is dominated by the inflow of the River Lepaanvirta in the southeast and outflow to Lake Pyhäjärvi in the northwest. The theoretical residence time is 450 days. Along the lake mid-line in a southeast–northwest orientation there is a long underwater esker, which divides the lake into two basins.

The average freezing and ice breakup dates were November 30 and April 30, respectively, in 1971–2000, and the average thickness of ice at annual maximum is about 50 cm in the region (Korhonen 2005). The ice sheet consists of snow-ice and congelation ice, and the growth is largely determined by the air temperature and snow accumulation. Ice grows on average by 0.5 cm per day, while snow cover on top is 10–20 cm in mid-winter (Korhonen 2005; Yang et al. 2012). With progress of snow accumulation, flooding of the ice surface takes place and slush and snow-ice is formed; on average the snow-ice contribution is about 10 cm or 20% of the ice sheet. Ice decay begins in the latter half of March when the radiation balance upcrosses zero. The rate of ice decay is 1–2 cm per day.

Ice and especially snow are good insulators. As soon as ice has grown to more than 10 cm thickness, only weak heat loss can be observed from the water body through the ice. The heat flux from the water to the ice bottom was estimated as 0.5 W m^{-2} in a modelling study (Yang et al. 2012), showing that the ice cover strongly protects the heat content of the water body. Only in spring after snowmelt does the heat flux from the water to ice bottom increase. This is because a fraction of solar radiation penetrates the ice and heats the near-surface water layer, from where some of the heat is further conducted to the ice bottom.

The winter field campaigns were performed in January–March 2009 and 2010 in the YMPANA project (Jaatinen et al. 2010; Lei et al. 2011). Hydrographical surveys were made regularly, an automated ice station was set up, and optical investigations were performed (Jaatinen et al. 2010;

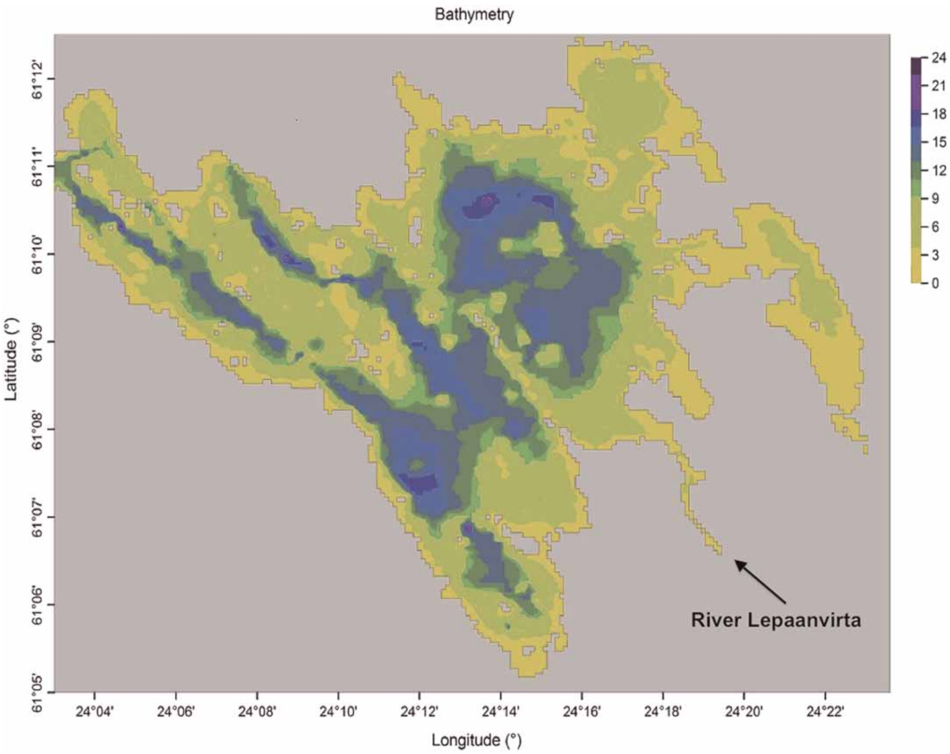

Figure 1 | Topography of Lake Vanajanselkä.

Lei *et al.* 2011). The collected data contain ice samples, soundings for temperature, conductivity, pH and oxygen, light measurements, and water samples for nutrients, chlorophyll *a* and phytoplankton across the lake. The grid contained 28 points quite evenly spread throughout the lake. The freezing dates were later than normal (January 1, 2009 and December 14, 2009), and the field trips were made in mid-winter (January 26, 2009 and February 3–15, 2010) and at the end of the ice growth season (March 18, 2009 and March 25, 2010). Because of weather problems, sampling in February 2010 was spread over 12 days, and we can refer these data dated nominally for February 10.

Ice samples were taken for the stratigraphy and crystal structure. The samples were sawed from the ice sheet and taken to a cold room for further analyses. Thickness, stratigraphy and quality of ice and snow were recorded at the site. Hydrographic soundings were made for temperature and conductivity with a CTD (conductivity–temperature–depth (pressure)) device (Falmouth Scientific micro CTD) and additionally for oxygen and pH with a YSI multichannel sounding instrument. Water samples were also taken and processed in the laboratory for nutrients and chlorophyll *a* and for phytoplankton studies at the Lammi Biological Station.

Light measurements were made manually and with recording sensors. Small PAR (photosynthetically active radiation) sensors (diameter 18 mm and length 115 mm) were deployed into the ice sheet at three depths for time series recording of quantum PAR irradiance in the 400–700-nm band (MDS-L, Alec Electronics Co. Ltd, Japan). Spectral measurements of downwelling and upwelling irradiance were made using Ramses ACC-VIS hyper-spectral radiometers (TriOS, Inc., Germany) in the wavelength window 320–950 nm with spectral resolution of 3.3 nm. The sensor is equipped with a cosine receptor and the field of view is a full hemisphere. The sensitivity goes down to 4×10^{-5} W m^{-2} nm^{-1}, and spectral accuracy is 0.3 nm. Ramses was employed for the measurements of surface spectral reflectance and light attenuation due to the ice cover. Downwelling irradiance was first recorded, and immediately after that the sensor head was inverted and upwelling irradiance was recorded. The sensor was kept at about 40 cm above the surface during measurements of upwelling irradiance ensuring that the sensor collected more than 96% of the signal (Nicolaus *et al.* 2010). Irradiance under ice was measured with an aluminium frame, which was lowered through a 50 × 40 cm hole. The frame was equipped with a mechanism to push the sensor southwards (i.e., towards the sun) beneath the ice by 1 m (see Lei *et al.* 2011).

Phytoplankton samples with a volume of 200 mL were collected from the depth of 1 m using the Limnos water sampler and preserved immediately by acid Lugol's solution. In the laboratory the samples were settled for more than 24 hours in a 50-cm^3 (occasionally 25-cm^3) chamber. The algae and cyanobacteria were determined with Wild M40 and Leica 090–135 inverted microscopes and counted with Wild M40 (modified method of Utermöhl (1958)). For larger plankton species and colonies the whole cuvette bottom (Ø 25 mm) was calculated with a magnification of 150 ×, or half a bottom with 300×. For smaller species ≥30 fields of view were calculated with a magnification of 600×. Plankton biomass values were obtained as mg m^{-3} (wet mass).

RESULTS

Ice cover and under-ice physics

Winter 2008–2009 was mild. The freezing and ice breakup dates were January 1 and April 27, respectively (Table 1).

Table 1 | Characteristics of the study ice seasons and the field days. Also shown is the estimated discharge from the River Lepaanjoki January 1–March 31

	2008–2009	2009–2010
Freezing date	Jan 1	Dec 14
Lepaanjoki discharge (m^3 s^{-1})	16	9
1st field day	Jan 26	Feb 10
– ice thickness (cm)	25	35
– snow thickness (cm)	15	20
2nd field day	Mar 18	Mar 25
– ice thickness (cm)	38	55
– snow thickness (cm)	18	30
Ice break-up	Apr 27	Apr 21

November and December were warm, with an average air temperature of 0.6 °C, and freeze-up took place 1 month later than normal. From late December to the end of March, the air temperature was permanently below the freezing point, on several occasions approaching –20 °C. The thickness of ice reached 41 cm in late March. At the beginning of April the air temperature increased gradually to 0 °C and higher, and melting commenced at the top surface. The average growth rate was 0.48 cm per day while the average melt rate was 1.24 cm per day.

Winter 2009–2010 was exceptionally cold and with much snow. The lake froze over on December 14 (Table 1). The temperature was very low but snow accumulation was large which is not a common combination. Consequently, there was much slush and snow-ice but the total ice thickness was not very great. Slush, or a mixture of snow and liquid water, is accepted as a part of the snow layer when it is above the ice sheet; but when slush freezes to form snow-ice or is locked inside the ice by snow-ice formation on top of the slush, it becomes a part of the ice layer. In 2010, a slush sub-layer persisted in the snow layer from February to the melting season. The maximum ice thickness reached 55 cm in March. April was warmer than normal, so that in spite of the large volume of snow and ice the lake was already ice-free on April 21. The average growth rate was 0.54 cm per day while the average melt rate was 2.04 cm per day.

In the mid-winter field trips, the ice thickness was 25 and 35 cm in 2009 and 2010, respectively, with already 15–20 cm snow on top. On March 18, 2009, the total thickness of ice was 38 cm, of which 1 cm was snow-ice, and snow thickness was 18 cm including a slush layer next to the ice surface (Figure 2). From then there was a semi-persistent thick slush layer, so that moving on ice by foot, skis or snow mobile was very difficult. Snow-ice grew, as usual, down from the top of the slush layer, and an internal slush layer remained within the ice sheet. In mid-March 2010, the total ice thickness was 55 cm consisting of 16 cm snow-ice, 9 cm slush and 30 cm congelation ice.

There was a large difference in river water discharge between the study seasons (Table 1). In 2009 the range was 9–30 m^3 s^{-1} in January–March, at its greatest in the beginning and at a minimum at the end of this period. In 2010 the discharge was on average smaller than in the

Figure 2 | Ice stratigraphy in March 2009 and 2010.

year before, especially in January–February, and the range was 4–12 $m^3 s^{-1}$. The total inflow in January–March 2009 corresponded to 16% of the lake volume.

At the end of January 2009, a distinct two-layer thermal structure in the water-column was recognised (Figure 3(a)). The temperature was close to 0 °C in the top 4-m layer. The lower layer was continuously stratified by 0.1–0.5 °C m^{-1}, at its strongest in the 4-m thick sub-layer at the bottom, and the bottom temperature was 1.5 °C (Figure 3(a)). Thus the water-column was already at that time, 25 days after ice-on, inversely stratified. On March 18, the water temperature at 17 m depth, close to the bottom of the lake, had reached more than 3.5 °C, having increased more than 2 °C during 2 months. However, the warming of the upper water layer was much lower, about 0.3 °C. Because the temperature of the inflowing water was less than 0.5 °C, it could not explain the warming of the lower layer. The cold inflowing water spread into the upper layer of the lake and kept it cold over the whole winter season.

Light conditions

Ice cover has a major influence on the solar radiation transfer into a lake water body, especially in the presence of snow on ice. The principal optically active impurities in ice and snow are gas bubbles, which scatter light and thus weaken the light penetration. The attenuation length (or e-folding distance) of downwelling light irradiance is 10–15 cm in snow and snow-ice and 50–100 cm in congelation ice (Roulet & Adams 1986; Arst *et al.* 2006). Since congelation ice growth rejects most of the dissolved and suspended optically active substances out of the ice sheet, the transparency of the ice may be better than the transparency of the lake water (Leppäranta 2009). Light transfer into a lake water body is crucial for thermal conditions, circulation and biological processes (Salonen *et al.* 2009). Also, optical properties of water can be considered as indicators of the ecological state of the water body (Arst 2003).

In the present study, measurements on the optics of Lake Vanajanselkä were carried out in March–April 2009 (Figure 4(a)). Water quality observations were also made in connection with the optical measurements. On March 18, 2009, there was an 18-cm thick snow cover, with the lower 7 cm forming a slush sub-layer, and the ice sheet consisted of a 1-cm snow-ice layer and a 37-cm congelation ice layer. The net surface irradiance was 50–100 mW m^{-2} nm^{-1}. The vertically averaged spectral attenuation coefficient was 6–8 m^{-1} in the PAR band (400–700 nm), reaching higher values toward ultraviolet and near infrared wavelengths (Figure 4(b)). The strong increase in short wavelengths was due to lake water flooded onto the ice: coloured dissolved organic matter (CDOM) in the lake water absorbs short wavelengths. Water itself (whether liquid or solid phase) strongly absorbs longer wavelengths. The minimum attenuation was at about 575 nm wavelength, and also this is due to CDOM in the slush. The 'average' PAR attenuation coefficient was 6.8 m^{-1}. The euphotic depth in natural ice-free waters is usually defined as the depth where the irradiance is 1% of the surface level that results in the depth of 4.6

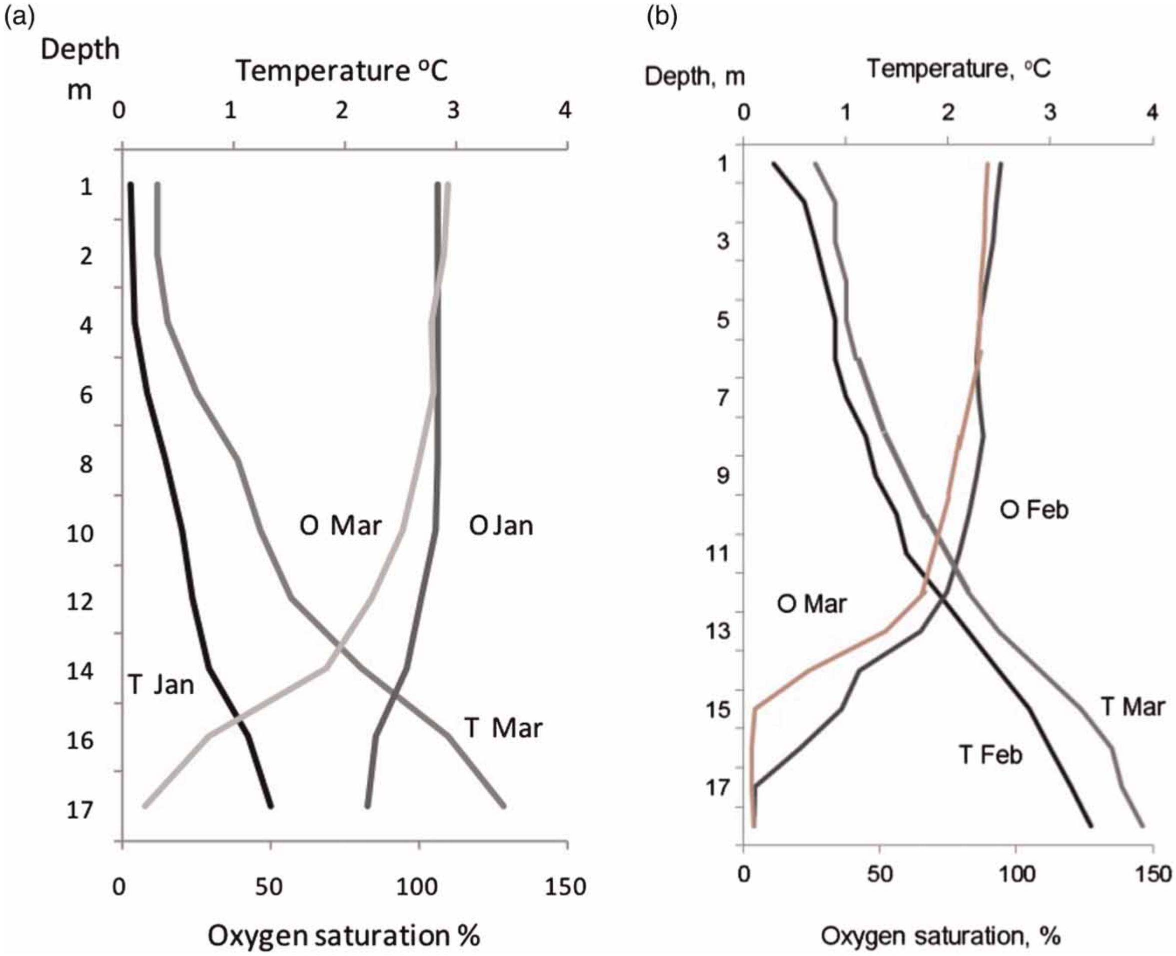

Figure 3 | (a) Vertical distribution of water temperature (*T*) and oxygen saturation (*O*) in the field visits of January 26 and March 18, 2008. (b) Vertical distribution of water temperature (*T*) and oxygen saturation (*O*) in the field visits of February 10 and March 25, 2009.

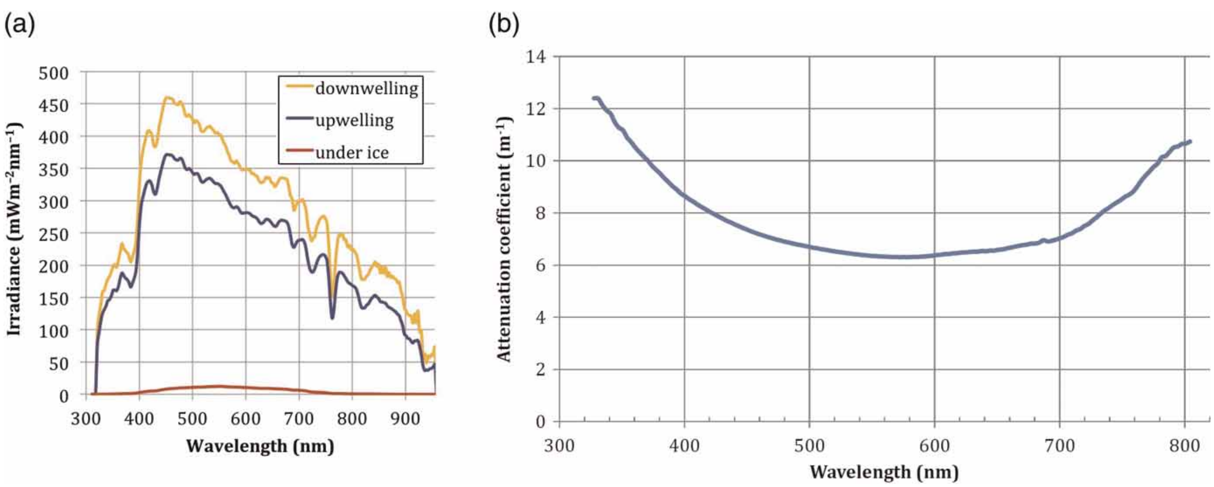

Figure 4 | Solar radiation measurements on March 18, 2009. (a) Measured spectral downwelling and upwelling irradiance above ice and downwelling irradiance beneath ice; (b) vertically averaged spectral diffuse attenuation coefficient.

times the mean attenuation depth (Arst 2003). Here the resulting euphotic depth was 68 cm, which is more than the actual total thickness of the ice and snow (56 cm). According to the direct measurements, beneath the ice the downwelling irradiance was less than 10 mW m^{-2} nm^{-1}. This level of irradiance is just enough for primary production.

Figure 5 shows a 1-week time series for the attenuation coefficient of snow and ice in the PAR band obtained with the recording sensors deployed in the ice and snow (Lei *et al.* 2011). In the snow cover, the attenuation coefficient changed from 30 to 8–10 m^{-1} on March 20–22 (Figure 5(a)). The high level was in cold, dry snow, but would be too high to fit with the spectral measurements (15 m^{-1} would give a reasonable fit). The reason for the bias in the PAR results is likely due to local variations in snow properties. Then the weather warmed and the lower value was reached in wet melting snow. Scattering of light was reduced after the appearance of meltwater in the snow pack.

The attenuation coefficient was more stable in congelation ice (Figure 5(b)). It was 2 m^{-1} on April 2 and then decreased to 0.8 m^{-1} at its minimum on April 5. The variation was likely due to the dynamics of gas- and water-filled voids in the ice sheet (see Leppäranta *et al.* 2010). According to the PAR sensor measurements, the total attenuation coefficient of the ice and snow cover was 8.8 m^{-1}. The RAMSES data produced an average attenuation coefficient of 6.8 m^{-1} over the PAR band and 7.8 m^{-1} over the whole measured spectrum of 300–1,000 nm (Figure 4).

Under-ice water chemistry

Beneath the uppermost mixed water-layer, oxygen saturation (and concentration) decreased from January/February to March during both winter seasons, and the decrease was especially clear below 10 m depth (Figure 3). On January 26, 2009, 25 days after ice-on, the upper layer was at the saturation level but in the bottom layer the level was 80%. Two months later, the surface layer was still saturated but in the lower layer the level was less than 50% and the bottom water was almost anoxic. At the end of March 2010, the relative oxygen content was less than in the previous year throughout the water-column, and deeper than at 15 m the water was anoxic. By assuming that the whole water-column was 100% saturated with oxygen at the time of ice-on, the oxygen consumption rate near the bottom layers was around 3% per day during the first few weeks after the ice-on while near the lake surface the rate was only 0.16% per day. From January/February to March the oxygen consumption rate decreased, however, and the oxygen depletion in the whole water-column (from surface to bottom) was 0.036 g m^{-2} per day. According to our observations in this lake, a few weeks before the ice-out intense photosynthesis may start to elevate the oxygen

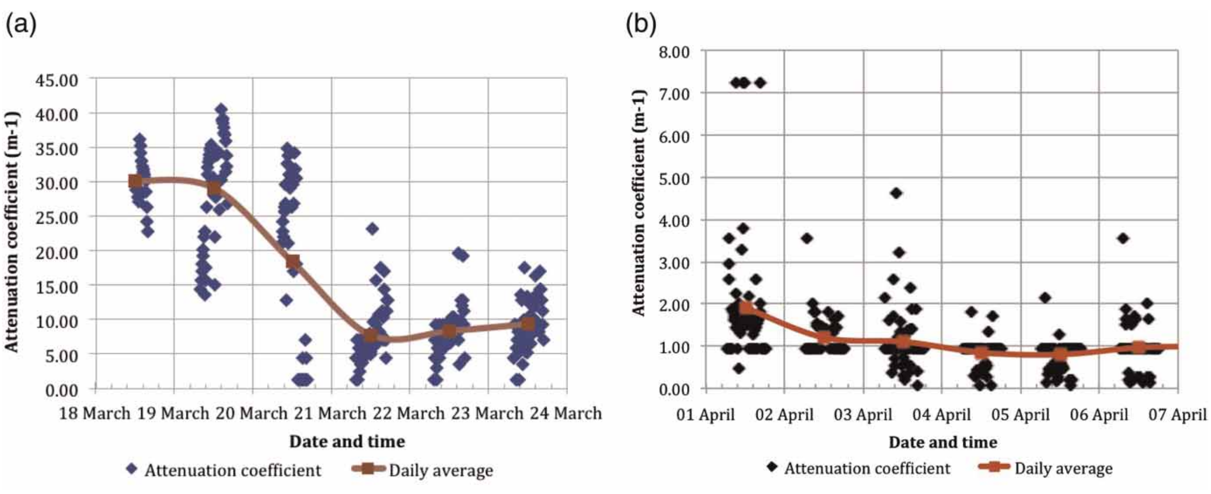

Figure 5 | PAR-band attenuation coefficient in Lake Vanajanselkä, March–April 2009. (a) Snow cover and (b) congelation ice. The line shows daily averages and the dots provide the range.

concentration under the ice-cover however (data not shown; Heini *et al*. unpublished). In deeper layers, the oxygen concentration increases only during full mixing of the water-column, a phenomenon which takes place at the time of ice-out and thereafter.

The conductivity of the lake water was around 150 mS m^{-1} (20 °C) during both winter seasons (Figure 6(a)). The values clearly increased close to the bottom sediments, particularly in March, reaching 200 mS m^{-1} at the bottom. Also the mean conductivity was a little higher in March than in January. The pH also changed in the water-column as well as in the course of the ice season (Figure 6(b)). There was a slight tendency to be lower in deeper layers, and also the whole profile shifted by about 0.3 towards lower in level from January to March.

Together with the conductivity, total phosphorus concentration was elevated close to the bottom sediments in winter but the concentrations also slightly increased across the water-column during the winter season (Figure 7). To some extent, a similar concentration increase in deeper water layers was also found in ammonium while for phosphate and nitrate no clear changes were found.

Under-ice water biology

Biological activity is usually low in the ice season (see Arvola & Kankaala 1989; Tulonen *et al*. 1994). This is due to the low water temperature. Ice-cover isolates the lake water body from atmosphere and hence oxygen concentration decreases in winter, especially close to the bottom sediments (Figure 3). At the same time, carbon dioxide concentration in lake water increases due to the respiration activity.

Phytoplankton biomass level was one order of magnitude lower in winter months than in summer months in 2009 (Figure 8). Due to the limited light conditions in midwinter, phytoplankton production is typically low or very low. Therefore, heterotrophic and mixotrophic species often dominate over autotrophic ones, and the phytoplankton community is also sparse (Arvola & Kankaala 1989;

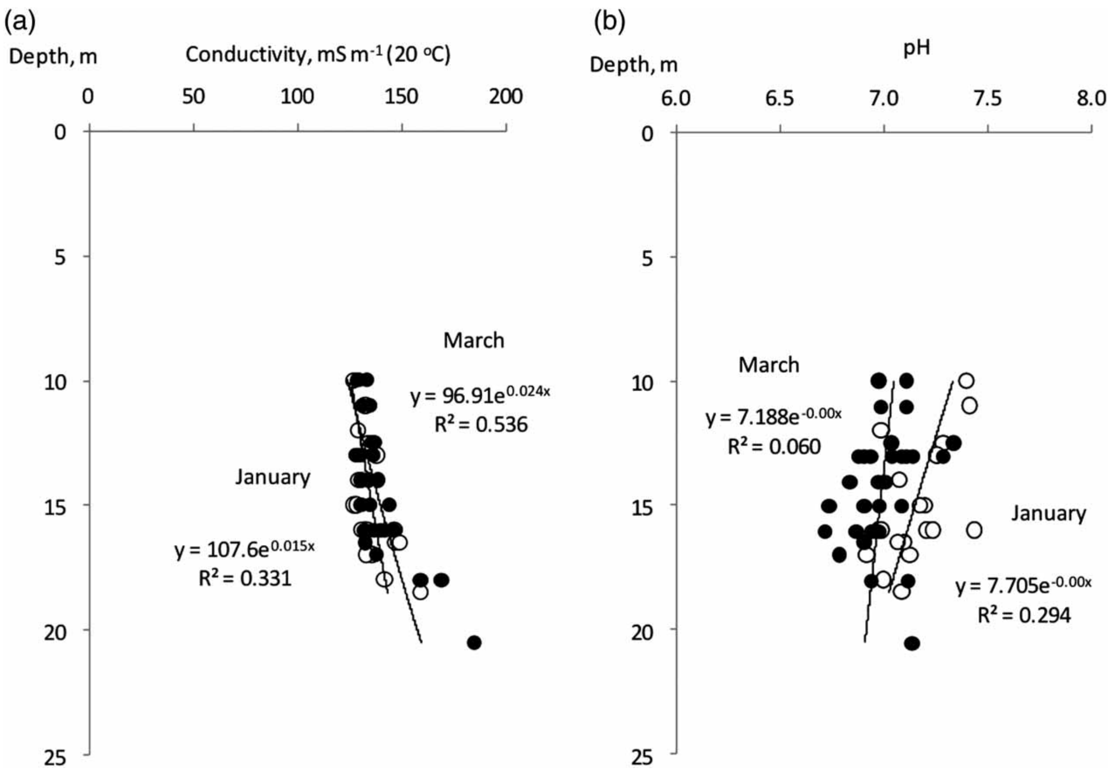

Figure 6 | Vertical structure of conductivity (a) and pH (b) profiles in Lake Vanajanselkä. White dots represent January 26, 2009 and February 10, 2010, and black dots represent March 18, 2009 and March 25, 2010; there are nine sampling points in 2009 and 14 in 2010 over the whole lake area.

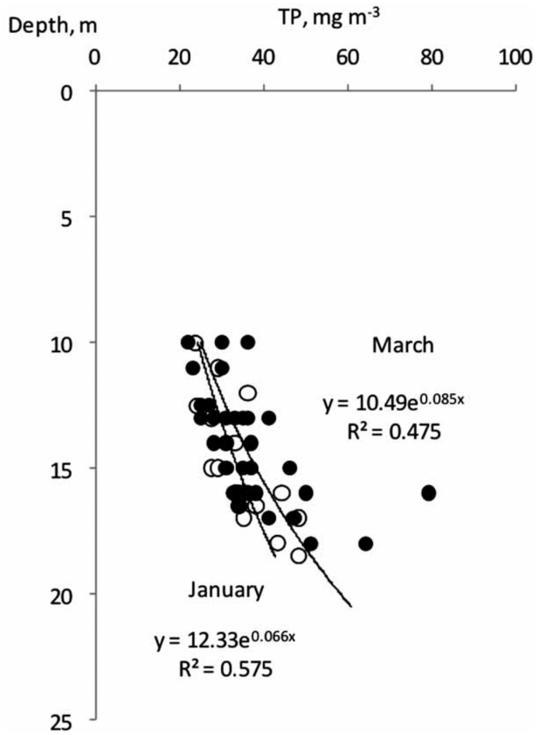

Figure 7 | The vertical profile of total phosphorus concentration in Lake Vanajanselkä. White dots represent January 26, 2009 and February 10, 2010, and black dots represent March 18, 2009 and March 25, 2010; there are nine sampling points in 2009 and 14 in 2010 over the whole lake area.

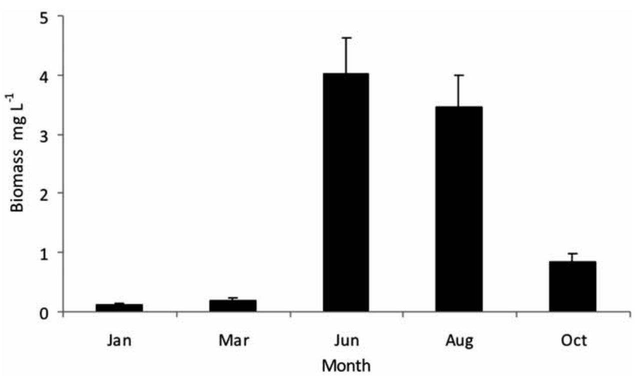

Figure 8 | Phytoplankton biomass (wet weight) in Lake Vanajanselkä in 2009. The bars indicate standard deviations between the different sampling sites.

Lepistö & Arvola unpublished). However, in the absence of snow, especially later in spring, photosynthesis can be high beneath ice due to improved light conditions.

In January 2009, the three most abundant taxonomical groups of phytoplankton were cryptophytes, chrysophytes and chlorophytes, which shared 44, 23 and 6%, respectively, of the total biomass. In March, cyanobacteria, cryptophytes and diatoms were the three dominant groups, and they contributed 39, 29 and 13%, respectively, of the total biomass. Based on these biomass results, it can be concluded that the heterotrophic species comprised 4–14% of the total biomass under the ice. However, there were several potentially mixotrophic species, and altogether mixotrophic and heterotrophic species made a significant contribution to the total biomass. The reduction of the snow is visible in radiation results during the days March 20–21 and the phytoplankton biomass measured 10 days later shows a clear increase in amount (Figures 4 and 8).

Mixing processes are usually very limited under ice which has an effect on the physical and chemical conditions and the biota. Conditions during fall cooling are therefore crucial, because the temperature of the water body at freeze-over depends on them and may reach anything within 0–4 °C. If the ice season is long enough and water temperature has been relatively high at freezing, the risk for oxygen depletion increases. Oxygen depletion may have serious chemical and ecological consequences, such as release of phosphorus from bottom sediments and fish kills with subsequent food web impacts.

DISCUSSION

Characteristic to the water body of an ice-covered lake is the inverse stratification and thermohaline circulation with low velocity. This produces a long time scale L/U, where L is the lake size and U is the velocity scale, 1 cm/s (Huttula *et al.* 2010). The water renewal also slows down because the inflow is largely stored in the snow cover in the drainage basin. The long time scale also stretches the memory of the lake over the ice season. In medium-sized and small lakes, the ice cover largely decouples the water body from the atmosphere that helps to stabilise the conditions in the lake.

The evolution of the heat storage of Lake Vanajanselkä (U) can be evaluated using a simple box model:

$$\frac{dU}{dt} = Q_b - Q_i + Q_s + Q_r \qquad (1)$$

where Q_b is the heat flux from the bottom sediments, Q_i is the heat flux from water body to the ice, Q_s is the heat flux from solar radiation, and Q_r is the heat flux from river inflow. In the first approximation, we can ignore solar radiation in midwinter in our cases. Prior to field trip 1, solar radiation was very low (December and January), and between field trips 1 and 2 there was thick snow cover and the penetration of sunlight into the water was weak. In the absence of mixing, the cold inflowing water spread into the upper surface layer of the lake and kept it cold over the whole winter season. Heat losses through the ice cover to the atmosphere were small. As shown by the model simulations (Yang *et al.* 2012), the ice lid efficiently restricted heat to 0.5 W m^{-2} to escape from the lake.

Significant warming of the lower layer was observed during the ice season, and the heat flux from the bottom sediments was the only possible explanation for that. It should be kept in mind that the bottom sediments of Lake Vanajanselkä warm up to 13–15 °C during the summer season (Heini *et al.* unpublished), and this is why the sediments are capable of releasing plenty of heat during the winter season. The average depth of the lower layer was 4 m, and the change in the heat content corresponded to heat flux of 2.4 W m^{-2} between the field trips.

Assuming the temperature difference of 0.25 °C between the river inflow water and surface layer water, the heat flux from the river water becomes 0.1 W m^{-2}. Thus the surface layer must receive about 0.5 W m^{-2} from the lower layer to conserve its heat storage, and we can estimate the heat flux from the bottom sediments as 3.0 W m^{-2}. The stratification of the water body kept most of the heat coming from the bottom in the lower water layers. In winter 2010, the water temperature was higher (by 0.5–1.0 °C) than in 2009 (Figure 3(b)). The temperature difference was mainly due to the difference in the ice-on date; the lake froze over on December 14, i.e., 2.5 weeks earlier than in the previous winter. In February and March the temperature showed continuous stratification with temperature increase by 0.2–0.6 °C between the field trips (45 days), corresponding to heating by 1.7 W m^{-2}. Because of the presence of a persistent slush layer in the ice sheet, congelation ice was isothermal and heat could not be conducted away through the ice.

Ice and snow cover strongly reduced the light penetration, in particular snow. Irradiance beneath ice, $E(H)$ can be estimated as

$$E(H) = (1-\alpha)\exp\left[-(\kappa_s h_s + \kappa_{si} h_{si} + \kappa_i h_i)\right]E(0) \qquad (2)$$

where α is albedo, κ is attenuation coefficient and h is layer thickness with subscript s, si and i for snow, snow-ice and ice, respectively, $E(0)$ is the downwelling irradiance at the top of the ice, and $H = h_s + h_{si} + h_i$. When $h_s + h_{si} \sim h_i$, congelation ice has a minor role in cutting the light level. In 2009 in Lake Vanajanselkä, the light attenuation coefficient was about 1 m^{-1} for congelation ice and 10 m^{-1} for snow; for snow-ice it could not be estimated because the snow-ice layer was only 1 cm thick, but based on earlier results the value should be 5–10 m^{-1}, close to that of snow (Jakkila *et al.* 2009; Leppäranta *et al.* 2010). The total attenuation in the congelation ice layer (37 cm) was 30% while 15 cm of snow attenuated as much as 80%. This is similar to what has been observed before. The influence of snow on underwater light conditions is strengthened by the albedo, which is 0.8–0.9 for snow and about 0.5 for ice.

The chemical conditions in northern temperate lakes are clearly modified by ice cover (Salonen *et al.* 2009). In Lake Vanajanselkä, the ice-cover isolated lake water from the atmosphere and hence oxygen concentration decreased. Although no direct measurements on the microbial activity of the bottom sediments were performed, oxygen consumption in water layers close to the lake bottom clearly indicated that the microbes were active. Losses of oxygen were due to respiration of organisms and chemical oxidation of reduced materials mainly in the bottom sediments. Therefore, rates of oxygen decay depend on the concentration of substrate available for decomposition (Greenbank 1945) and other environmental factors supporting microbial activity such as temperature (Meding & Jackson 2001). Oxygen depletion may have serious chemical and ecological consequences, such as release of phosphorus from bottom sediments and fish kills with subsequent food web impacts (Tonn & Magnuson 1982; Ruuhijärvi *et al.* 2010).

Compared to the oxygen consumption, the respiration of micro-organisms produced carbon dioxide which may explain the slightly lower pH in the lower layer in March during both winter seasons. According to Charlton (1980), hypolimnion oxygen represents hypolimnion thickness and temperature as well as the productivity of the lake. Therefore, we assume that any changes in water temperature during the ice-on may influence oxygen depletion during the ice-cover season. Meding & Jackson (2001), having studied 23 North American temperate lakes, pointed out that oxygen decay rates correlated with morphometry in shallow lakes, and in deep lakes correlation was with chlorophyll *a*, Secchi depth and the sediment surface area to volume ratio. Our result of an oxygen decay rate of $0.036\ g\ m^{-2}\ day^{-1}$ concurs with the model results of Meding & Jackson (2001) of $0.035\ g\ m^{-2}\ day^{-1}$ for their lakes, excluding the deepest ones. Because the ice-cover period can last several months, the chemical conditions of a lake can be influenced by the biological processes in spite of low temperature.

Based on the chemistry results, the mid-winter circulation patterns and element transport did not markedly differ in the 2 years despite distinct hydrological conditions. The chemical 'fingerprint' of the river inflow was stronger in March compared to January/February although the volume of inflow dramatically decreased at the same time. The stronger impact was presumably because of steeper thermal stratification in March. The magnitudes of mixing processes in ice-covered lakes are discussed, e.g., by Bengtsson (1996).

Due to the low water temperature, biological activity is usually low in the ice season (see Arvola & Kankaala 1989; Tulonen *et al.* 1994). Because of low irradiance beneath the ice, heterotrophic and mixotrophic species often dominate over autotrophic ones under ice conditions in winter (Rodhe 1955; Arvola & Kankaala 1989; Roberts & Laybourn-Parry 1999). In accordance with that, in Lake Vanajanselkä the mid-winter phytoplankton consisted mainly of cryptophytes and chrysophytes, both known to have potentially mixotrophic and heterotrophic species and also capable of surviving in winter conditions under ice (e.g., Jones 1994; McKnight *et al.* 2000), as well as chlorophytes. Later in winter, cyanobacteria, cryptophytes and diatoms dominated the phytoplankton assemblage. The later measurements occurred 10 days after the snow reduction was recognised by radiation measurements. In addition to changes in the dominating species, also the amount of phytoplankton increased. However, in the absence of snow, especially later in spring, photosynthesis can be high beneath ice due to improved light conditions (Tulonen *et al.* 1994; Mundy *et al.* 2009).

CONCLUSIONS

Only a few investigations have been made of the ice season in boreal lakes. Although the first papers appeared about 100 years ago, our knowledge of understanding the ice season is still quite limited. The main reason is the low biological productivity and consequent low interest, and on the other hand, there have been major technical difficulties in performing winter observation programmes. Lake Vanajanselkä is a shallow (mean depth 7.7 m) and eutrophic boreal lake, ice-covered on average for 5 months of the year.

The results from a winter field study in 2009–2010 in Lake Vanajanselkä, southern Finland have been presented. In the first winter, the ice thickness was less than normal, and snow-ice was exceptionally thin due to low snow accumulation. The ice sheet was 41–55 cm thick at the annual maximum, and consisted of snow-ice and congelation ice. The water body had a 4-m thick upper mixed layer and lower continuously stratified layer. The e-folding depth of light intensity was 50–100 cm for congelation ice and 5–10 cm for snow; for dry snow it was 5 and 10 cm for wet snow. In congelation ice this depth was 50–100 cm. Consequently, the light transmittance through the ice was largely controlled by the snow cover in mid-winter resulting in very low levels of light beneath the ice as long as dry snow was present. The heat flux from the lake bottom forced thermohaline circulation in the lake throughout the entire ice season. After snowmelt in spring, solar radiation penetrated the ice and initiated convection.

Oxygen concentration decreased in winter, especially close to the bottom sediments, and carbon dioxide concentration increased due to respiration activity. Phytoplankton production and biomass level were low or very low, and therefore heterotrophic and mixotrophic species were abundant. Oxygen depletion in the hypolimnium had several chemical and ecological consequences, such as release of phosphorus from the bottom sediments. In spring, just before the ice-out, photosynthesis was at a high level beneath the ice due to improved light conditions and started to elevate the oxygen concentration in the topmost water layer.

Further investigations are needed in ice-covered lakes to understand their physics and ecology and the impact of environmental and climate changes on them (Bengtsson 2011).

This work has demonstrated the characteristics of a wintertime boreal lake resulting from the presence of a solid ice cover. The key open question is the thermohaline winter circulation, which re-distributes water with its gases and other impurities during winter. Work is in progress to solve this using spatial data of our sampling grid and a 3-D numerical circulation model. Also, detailed work is ongoing concerning the variations of the River Lepaanvirta's inflow and the consequent influence on the water body.

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