The dependence of the consumption of dissolved oxygen on lake morphology in ice covered lakes

Lars Bengtsson and Osama Ali-Maher

ABSTRACT

The consumption of oxygen in ice-covered lakes is analyzed and related to biological oxygen demand and sediment oxygen demand. An approach for computing dissolved oxygen concentration is suggested assuming horizontally mixed waters and negligable vertical dispersion. It is found that the depletion of dissolved oxygen is mainly due to the transfer of oxygen at the water/sediment interface. The morphology of a lake is very important for how fast the dissolved oxygen concentration is reduced during winter.

Key words | convection, dissolved oxygen, ice covered lakes, lake morphology, sediment oxygen demand

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INTRODUCTION

The water quality in lakes is much dependent on the dissolved oxygen (DO) in the water. In lakes, which are ice covered during long winters, the dissolved oxygen concentration may decrease to very low values. In shallow lakes, anaerobic conditions may develop, as was already discussed by Greenbank (1945). The ice cover isolates the water from the atmosphere and thus there is no gas exchange between air and water. Solar radiation is weak and does not penetrate through the ice and the snow on the ice. Therefore, there is no photosynthesis and no primary production contributing to oxygen production. Instead, DO decreases, when oxygen is consumed by bacterioplankton in the water through organic matter mineralization and by diffusion of dissolved oxygen. The oxygen concentration is reduced faster in the deep parts of a lake than near the ice. In a state-of-the-art paper, Bengtsson (2011) points to the importance of increased knowledge about dissolved oxygen in winter lakes. A short review on DO in ice covered lakes is given by Terzhevik & Golosov (2012). Kirillin *et al.* (2012) in their review on ice-covered lakes also discussed DO in ice-covered lakes, as did Golosov *et al.* (2012) with emphasis on climate change. Primary production can proceed some time after ice formation prior to when there is snow on the ice. At high altitudes, at modest latitudes, and snow free ice cover, primary production can be due to rather intense solar radiation continuing throughout the winter, as shown in a study from Mongolia by Song *et al.* (2019). The DO is prevented from decreasing to low values.

In regions where the ice covered period extends into April or even May (Northern Globe), the solar radiation is intense, when the lakes are still ice covered. After the snow has melted, some solar radiation can penetrate through the ice into the water. Radiative-driven convection is initiated under the ice (Heaney & Matthews 1987; Bengtsson 1996). Maher (2002) and Maher *et al.* (2004) measured how the DO profile changed during the convection period.

There are many reported observations of DO concentrations in ice-covered lakes. When describing the depletion of oxygen, the lakes are mostly treated as well-mixed units (e.g., Hargrave 1972), but Golosov *et al.* (2007) accounted for vertical distribution of DO. In studies of North American Prairie lakes, Barcia & Mathias (1979)

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found that the consumption of DO was related to the depth of a lake and thus to the morphology; the shallower lake, the more oxygen consumption and the lower DO concentration. Mathias & Barcia (1980) stressed the importance of the extension of littoral zones and the oxygen depletion during winter. Chambers *et al.* (2008) pointed to the importance of the relation lake area to lake volume on the DO reduction, which is an indication of the importance of diffusion of DO into the sediments. This was confirmed by Leppi *et al.* (2016) from a study of ice-covered Alaskan lakes.

The DO concentration decreases with depth in ice covered lakes (e.g., Falkenmark 1973; Maher 2002; Terzhevik *et al.* 2010). The relation bottom area, where oxygen can be adsorbed into the sediments, relative to the water volume at a certain depth, increases with depth. Thus, the uptake of DO into the sediments should have a much larger effect on the DO consumption at large depths than near the ice.

The primary objective of the present study is to find out how the reduction of DO and its vertical distribution is related to lake morphology. Attempts are made to simulate the development of DO concentration, especially at the end of the winter. A second objective is to investigate how the solar radiation-induced convection under ice in the early spring changes the DO concentration. It is assumed that the lake water is horizontally well mixed, as is found in most studies of not very large lakes (Bengtsson 1986; Bengtsson *et al.* 1996; Terzhevik *et al.* 2010). Malm (1999) described three types of seiche currents contributing to complete horizontal mixing in small lakes: longitudinal and transverse seiche currents and high-frequency currents. It is further assumed that the DO is consumed within the water through bacterial plankton consumption and by adsorption into bottom sediments. The approach is applied to observations from Lake Velen in Sweden and Lake Vendyurskoe, Russia. The observations are from Falkenmark (1973), Maher (2002), Maher *et al.* (2004) and from Terzhevik *et al.* (2010). The data from Lake Velen have never been analyzed. The Lake Vendyurskoe data from Maher have been analyzed by Terzhevik *et al.* (2010), but not with respect to the influence of the sediments.

The paper is structured so that after the introduction with objectives, the two observation sites are described, followed by a section reporting literature data of observed DO consumption in different ice-covered lakes. The theory on which this paper is based is given, directly followed by applications on the two lakes. The applications are divided into separate sub-sections for each lake, and a sub-section dealing with convection under ice. The results of the applications are summarized followed by discussion.

STUDY SITES

Lake Velen

Lake Velen is situated in south-western Sweden in a boreal forest-type basin. The lake has a glacial origin. The lake basin is elongated, 6.3 km long, and rather narrow, with a maximum width of 1.1 km. The lake area is 2.8 km^2 and the mean depth 6.5 m. The maximum depth is 17 m. The western shoreline is rather steep. The river inflow is minor and almost zero during the ice-covered period. Ice usually forms in December and disappears in April. Detailed morphometric properties of the lake are given in Falkenmark (1973). The bottom sediments at deep water consist of a thick mud layer. The sediments on shallow water have less organic content but are still silty. There is a sand bottom only very close along the eastern shoreline. The distribution of area versus depth is shown in Figure 1 (Falkenmark 1973). The topology of the bottom is regular with no depressions. Continuous temperature measurements were made in four verticals during 1969–1972. The DO was measured by taking samples (from near the surface, at 1 m, 2 m, and then every 2 m downwards) once a month in a vertical in the central part of the lake.

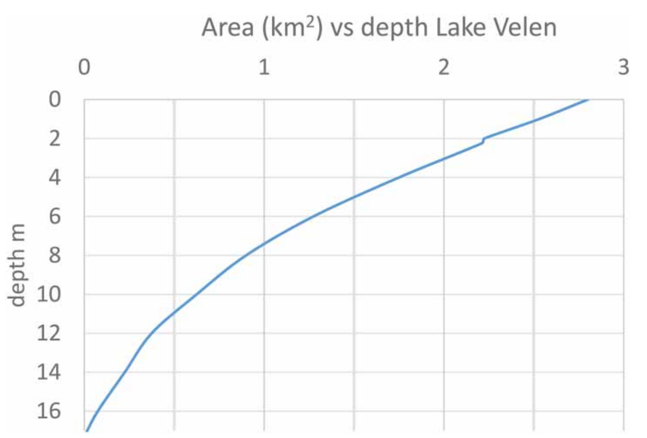

Figure 1 | Hypsographic curve, Lake Velen, Sweden.

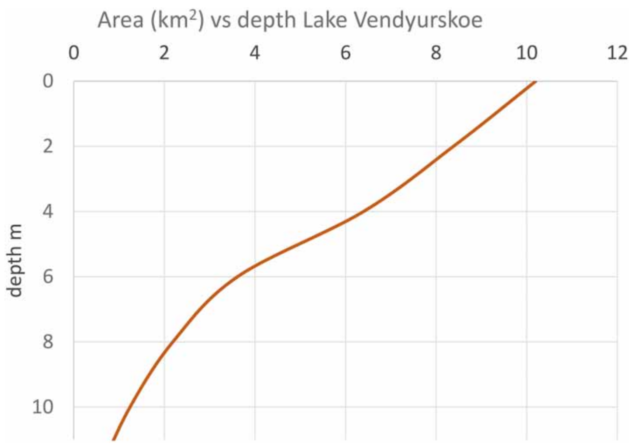

Figure 2 | Hypsographic curve, Lake Vendyurskoe, Russia.

Lake Vendyurskoe

Lake Vendyurskoe is situated in Russian Karelia. The lake is similar to Lake Velen. It has a glacial origin. There is almost no inflow in winter. The lake is situated in the same type of landscape as Lake Velen. The lake is shallow, with a mean depth of 5.3 m and maximum depth 13.5 m. The surface area is 10.5 km^2, and thus larger than the area of Lake Velen. The morphology is given in detail in Maher *et al.* (1999). A hypsographic curve is shown as Figure 2. There are shallow parts and several deep hollows with depth exceeding 8 m distributed over the lake. The areas of the deep hollows away from the central part of the lake are not included in the hypsographic curve. The curve is only extended to 11 m. The sediment of organic matter is less than 0.5 m near shores but exceeds 1 m at the deep bottoms. The ice-covered period is usually from early or mid-November until late April. Temperature and DO were measured *in situ* using an oxygen sensor in many verticals during several week-long campaigns mainly in late winter.

OBSERVATIONS OF OXYGEN CONSUMPTION FROM THE LITERATURE

Barcia & Mathias (1979) measured DO in ten ice-covered lakes less than 4 m deep. They found the consumption rate to be about 0.3 g/d and m^2 surface area. Ellis & Stefan (1989) estimated the oxygen demand per unit surface area to be about 0.2 g/m^2/d. These values correspond to a volume value of the order 0.1 g/m^3/d. From Russian sources, Kirillin *et al.* (2012) suggested a span 0.01–1 g/m^3/d. It is difficult to separate the consumption within the water and the loss to the bottom sediments. The consumption relative surface area increases the deeper a lake is, but the consumption per unit volume decreases.

In ice-covered Lake Vendyurskoe, Maher *et al.* (2004) found the total oxygen reduction including the consumption by sediments to be 0.03–0.09 g/m^3/d. For a very small lake in a Finnish forest, Bai *et al.* (2016) found the oxygen consumption as mean over a five-month period to be 0.15 g/m^3/d near the bottom, 0.06 g/m^3/d at mid depth, which is only about 2 m, and only 0.03 g/m^3/d near the ice. The latter depletion is assumed to be due to diffusion from lower levels. Pulkanen & Salonen (2013) found for the first year in deep Lake Pääjärvi a decrease rate of DO measured over two months in bottled water samples to be about 0.007 g/m^3/d for the mid-depth water and 0.025 g/m^3/d for water at greater depth. Measurements for a second year over three months showed the same value for mid-depth water but a higher value 0.048 g/m^3/d for deep water.

The consumption rate within the water in ice-covered lakes seems to change rather slowly with depth except near the sediments (Pulkanen & Salonen 2013). However, very near bottom sediments, also at shallow water, the DO concentration drops to low values, as observed during measurements in Lake Vendyurskoe (Maher *et al.* 2004), and from the referred Finnish studies.

METHODS – THEORY

Oxygen is consumed within the water and in the sediments. The consumption of oxygen within the water is, when there is no aeration, directly related to the consumption of bacterioplankton, the biological oxygen demand (BOD), as:

$$\frac{dc}{dt} = consumption = -k_L * L \qquad (1)$$

L is biological oxygen demand. The decay coefficient k_L has unit 1/day, t is time. Also L reduces, so that dL/dt is proportional to $k_L * L$:

$$\frac{dL}{dt} = -k_L * L \qquad (2)$$

Thus,

$$\frac{dc}{dt} = -k_L * L_0\, e^{-k_L t} \tag{3}$$

The concentration reduces exponentially. L_0 is the initial value of L. The solution to Equation (3) is:

$$c\,(t) = c_0 - L_0\,(1 - e^{-k_L t}) \tag{4}$$

with index 0 for initial value. This is the classical Streeter–Phelps analysis (Streeter & Phelps 1925).

Since there is no turbulence in the ice-covered lake water, much of the phytoplankton sinks towards the bottom. Therefore, the biological oxygen demand near the ice and at mid-depth reduces at a rate that is less than what can be expected from Equation (2). The settling velocity of phytoplankton varies depending on the type of phytoplankton but is in still water about 0.4 m/d (Burns & Rosa 1980). Burns & Rosa (1980) found this value from experiments with settling chambers placed in a lake. Their literature review shows that this settling velocity is consistent with other investigations although the variations can be within more than 0.2 m/d. Still, it means that the phytoplankton at the same time as being reduced while sinking, reaches 10 m depth in ice-covered lakes within a month after ice has formed on the lake. This indicates that, at least after a month, the reduction of DO in ice covered lakes may be more dependent on processes near the bottom than the consumption within the lake water. This is also in line with the findings by Barcia & Mathias (1979, 1980) that the DO reduction is faster the shallower a lake is.

As seen from Equation (3), the DO concentration reduces at a slower rate as time goes on. Thus, the consumption rate reduces when the concentration reduces. This led Terzhevik *et al.* (2010) to suggest that the DO consumption could be given as, now with notation k_w instead of k_L:

$$\frac{dc}{dt} = consumption = -k_w * c \tag{5}$$

with a decay coefficient k_w. The solution to Equation (5) is of course:

$$c\,(t) = c_0\, e^{-k_w t} \tag{6}$$

The sediments consume oxygen, sediment oxygen demand (SOD). The absorption into the sediments is a transport process over a diffusive boundary layer, in principle:

$$transport\ into\ sediment = k/d(c - c_B) \tag{7}$$

where k is diffusivity, d thickness of boundary layer and c_B is oxygen concentration at the sediment surface. Instead of using k/d, a parameter, transfer coefficient, unit velocity, $k_b = k/d$, can be introduced as was suggested already by Lewis & Whitman (1924). The sediment uptake of oxygen is much reduced when the DO concentration in the water drops below about 3 mg/L (Hargrave 1972; Mathias & Barcia 1980). The value of k_b depends on the character of the sediment and on the slow water movements near the bottom. For the conditions in ice-covered lakes with very slow movements, the parameter k_b must be empirically estimated. Gentle movements increase the sediment uptake, as for example, reported by Gantzer & Heinz (2003) from laboratory experiments. Maher (2002) estimated the sediment uptake in Lake Vendyurske to be about 0.15 g/m^2/d.

As already stated, small winter lakes are horizontally homothermal. Also with respect to DO, the concentration is almost constant in horizontal layers, although it drops to low values near bottoms. However, there are horizontal variations of the DO concentration near the underside of the ice downwards 1–2 m, as shown by Maher *et al.* (2004) and Terzhevik *et al.* (2010). These measured horizontal variations may be due to entrainment of air into the bore holes (although the holes were covered as soon as possible), or it could be the effect of non-uniform freeze-outs of gases and solutes during ice formation.

When the DO concentration is 10 g/m^3, which is about 70% of the saturated value in winter lakes, the consumption 0.1 g/m^3/d corresponds to a decay coefficient of 0.01 day^{-1}. Regarding the sediments, if SOD is 0.15 g/m^2/d, and the DO concentration is 10 g/m^3, then the transfer velocity is $k_b = 0.015$ m/d; if the concentration is 5 g/m^3, the transfer velocity is 0.03 m/d.

Assuming horizontally well-mixed water and neglecting advective terms, the decrease of dissolved oxygen is described by the diffusion equation:

$$\mathrm{A(z)}\frac{\delta c}{\delta t} = A(z)D\frac{\delta^2 c}{\delta z^2} - consumption \tag{8}$$

where c is DO concentration, t is time, z is vertical direction, and D is diffusion coefficient. The consumption term includes consumption within the water as well as loss to the sediments. Malm (1999) found in a study of Lake Vendyurskoe that the vertical mixing is determined solely by molecular processes. As already discussed, the consumption rate per unit volume is of the order 0.1 g/m^3/d. This consumption rate can be compared with the first term on the right-hand side of Equation (8). The scaling is, excluding the area, Dc/H^2, where H is a depth scale. The oxygen diffusion coefficient in cold water is about 10^{-4} m^2/d, so when c is 10 g/m^3 and the depth scale is 5 m, this scaled diffusion term is only 4×10^{-5} g/m^3/d, and thus much smaller than the consumption rate. This was discussed already by Golosov *et al.* (2007). Therefore, what is left of the diffusion equation is only:

$$\frac{\delta c}{\delta t} = -consumption \tag{9}$$

The consumption within the water is following Terzhevik *et al.* (2010) taken as k_w*c. It will be shown later that this consumption is minor compared to the oxygen uptake by sediments. The consumption due to transport of oxygen into the bottom sediments is per unit exposed bottom area, SOD, or $k_b * (c - c_B)$. The volume of an increment dz at level z is $A(z) * dz$, $A(z)$ being the lake area at distance z from the deep part of the lake. The exposed bottom area at that level is the *wetted perimeter* * dz, which is $A(z) - A(z - dz)$. The wetted perimeter is $\delta A/\delta z$. Integrating over the area at a given level z gives:

$$A(z)\frac{\delta c(z)}{\delta t} = -k_w c(z)A(z) - SOD\ \frac{\delta A}{\delta z} \tag{10}$$

with z directed upwards. When SOD is restricted by the transfer through the boundary layer, the equation is after division by $A(z)$:

$$\frac{\delta c}{\delta t} = -k_w c - \mathrm{k}_B\ (\mathrm{c} - \mathrm{c}_B)\frac{\delta A}{\delta z}\frac{1}{A(z)} \tag{11}$$

Since there is no direct mixing between different z-layers, $c(z)$ is simply replaced by c in the formulas below.

The consumption, right-hand side of Equation (11), at a certain level can be described as:

$$consumption(z) = \left(k_w + k_B\frac{\delta A}{\delta z}\frac{1}{A(z)}\right)c - k_B\frac{\delta A}{\delta z}\frac{1}{A(z)}c_B \tag{12a}$$

or after introducing a parameter $K(z)$

$$K(z) = k_w + k_B\frac{\delta A}{\delta z}\frac{1}{A(z)} \tag{12b}$$

$$consumption(z) = K(z) - k_B\frac{\delta A}{\delta z}\frac{1}{A(z)}c_B \tag{13}$$

The introduced parameter K, unit 1/time, is dependent on the depth distribution and since the area reduces with depth, the coefficient K increases with depth.

Lake area is a function of depth, see the hypsographic curves in Figures 1 and 2. Often a simple expression can be used:

$$A(z) = A_s(z/H)^b \tag{14}$$

with A_s, as the surface area and H as the maximum depth and z distance from deep bottom. Then:

$$\frac{\delta A}{\delta z}\frac{1}{A(z)} = \frac{b}{z} \tag{15}$$

and K is

$$K(z) = k_w + bk_B 1/z = function(z) \tag{16}$$

When the oxygen concentration in the sediment drops to zero, the solution of the consumption equation, is simply

$$c = c_0 exp(-Kt) \tag{17}$$

with c_0 as the initial DO with K being dependent on the considered depth (level), either from Equation (16) or from the complete expression Equation (12b). This is consistent with the approach by Terzhevik *et al.* (2010) but with a depth-dependent coefficient. For Lake Velen, the hypsographic curve (Figure 1), is well described by Equation (14) with exponent $b = 1.7$ and Equation (16) could be used for estimating K. For Lake Vendyurskoe, the more complete expression Equation (12) had to be used. Golosov et al. (2007) showed the same exponential equation as Equation (17), but adjusted for different columns within a lake.

The character of sediments at deep water are, in most lakes, different from those near the shores. While the sediments at shallow water are more of mineral character, the deep water sediments are muddy and organic. The organic content increases with depth. Therefore, the sediment oxygen transfer coefficient can be assumed to increase from a surface value to higher value at large depth. A linear increase is assumed:

$$k_B(z) = k_{bsurface} + (k_{bdeep} - k_{bsurface}) * y/H_s \quad (18)$$

with constant value k_{bdeep} below the depth H_s; $y = H - z$ is measured from the surface.

Pulkanen & Salonen (2013), when measuring the oxygen consumption in bottles, found that the DO decrease was faster in the bottles with deep water compared to in the bottles with water from mid-depth. Therefore, it can be expected that also the decay coefficient k_w increases with depth. However, as will be shown, the oxygen consumption within the water is small compared to the flux to the sediments.

RESULTS – APPLICATIONS

The approach was first tested on Lake Velen for two years and then for Lake Vendyurskoe using data from three years.

Lake Velen

The DO measurements from Lake Velen are from the 1970s. Two processes are considered, the biological oxygen demand within the water, and the diffusion of dissolved oxygen into the sediments. There are two parameters in the suggested theory, k_w and k_b, but they can be depth dependent. It is assumed that the oxygen content in the sediments is very low. The oxygen content in mg/L was calculated from values given in mole (Falkenmark 1973).

Prior to freeze-up in late December 1970 (winter 1970–1971) the DO concentration varied in the vertical between 12.8 and 12.4 mg/L, but with lower value at 12 and 14 m. 12.8 mg/L was set as initial value through the entire water column.

The measurements under ice were carried out about 80 days after ice formation (March 1971). The total content of oxygen in the water had decreased by about 9,000 kmol, which corresponds to an average decrease of DO concentration by 1.7 mg/L.

Originally, it was thought to account for the consumption of bacterioplankton within the water. However, already during the first computation, it was found that to get a good fit k_w had to be given very low values. A good fit was obtained when k_w was given an increasing depth value from 0 at the surface to 0.001 1/d at the bottom, while the k_B value was 0.004 m/d near the ice and 0.018 m/d near the bottom. Just as good fit was obtained disregarding k_w but instead slightly increasing k_B at the bottom to 0.02 m/d. Therefore, the contribution of kw was omitted and only the oxygen uptake by the sediments was considered. It seems that the DO consumption within the water is small compared to the consumption by the sediments. The fit is shown in Figure 3. There are three parameters, $k_{bdeep} = 0.020$ m/d, $k_{bsurface} = 0.004$ m/d, and the depth from which the transfer velocity ($k_b = k_{bdeep}$) is constant, $H_s = 8$ m.

The previous winter, 1969–1970, was colder than the winter of 1970–1971. The lake froze over in late November. The dissolved oxygen in the water was measured prior to freeze-up. The dissolved oxygen was then measured from

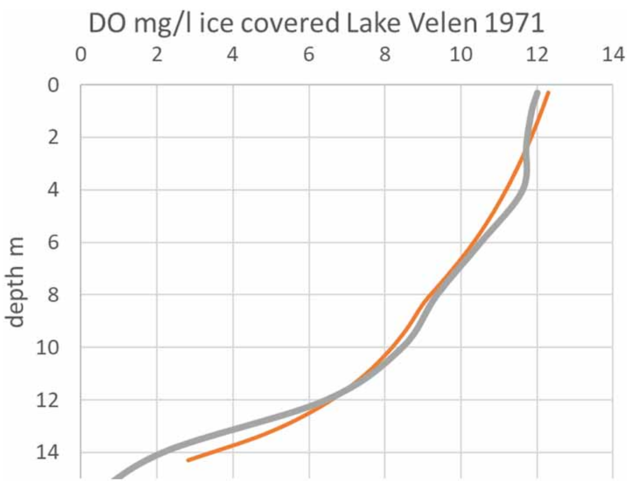

Figure 3 | Computed (thin line) and observed (thick line) DO profiles 80 days after ice formation.

the ice on 1 April, about 130 days after the lake was ice covered. In this year the oxygen reduction in the entire lake water was more than the other year due to a longer ice covered period. The reduction of oxygen was about 1,300 kmol for the entire lake water.

The Lake Velen simulations were repeated for the winter of 1969–1970 using exactly the same three parameters as for the winter 1970–1971. The initial dissolved oxygen concentration was about 12.4 mg/L down to 12 m, and decreased from this depth towards the deep bottom at 17 m. 12.4 mg/L was used down to 12 m as initial value in the computations, and below that decreasing value towards 9 mg/L. The comparison between simulated DO and observations at the end of the winter is shown in Figure 4. Again, the fit is good, except between 6 m and 8 m, where the dissolved oxygen is computed to decrease too much. This gives some indication that the same parameters can be used in consecutive years although the winters are of different length.

Lake Vendyurskoe

The suggested way of computing DO profiles was applied also to Lake Vendyurskoe in Russian Karelia. From this lake, there are measurements in many different profiles. In this paper the observations are from early spring 2000 (Maher 2002), winters 2001–2002 and 2003–2004 (Maher *et al.* 2004), and the same 2004 data from Terzhevik *et al.* (2010). The 2000 year investigations were performed in the spring to study convection under the ice.

Simulations of how the concentration of DO reduced under the ice cover were first carried out for the winter 2003–2004. DO profiles from late winter 2004 can be found in the report by Maher *et al.* (2004). In Maher *et al.* (2004), the DO concentration at the time of freeze-over is given as about 11.5 mg/L, which is chosen as initial value. The oxygen consumption was first computed using the same parameters as for Lake Velen. However, the DO consumption was computed to be too intense when those parameters were used. By decreasing the k_{bdeep} from 0.02 to 0.017 m/d a good fit was obtained, as seen in Figure 5, where the simulated DO profile is compared with the observed one. Thus, the used coefficients are $k_{surface}$ 0.004 m/d, k_{deep} 0.017 m/d, and H_s 8 m. As for Lake Velen, the DO profile can be computed only considering the sediment oxygen demand.

Also the DO consumptions for the winter 2001–2002 were investigated. Maher *et al.* (2004) and Terzhevik *et al.* (2010) reported DO concentrations of about 11.5 mg/L in early December 2001, 2 weeks after ice formation. This was taken as the initial value in the computations and 1 December was taken as the starting date. The same parameters as for the winter 2003–2004 were used for the simulations. Observed and simulated profiles at the end of the winter are shown in Figure 6. The consumption at deep water is somewhat underestimated.

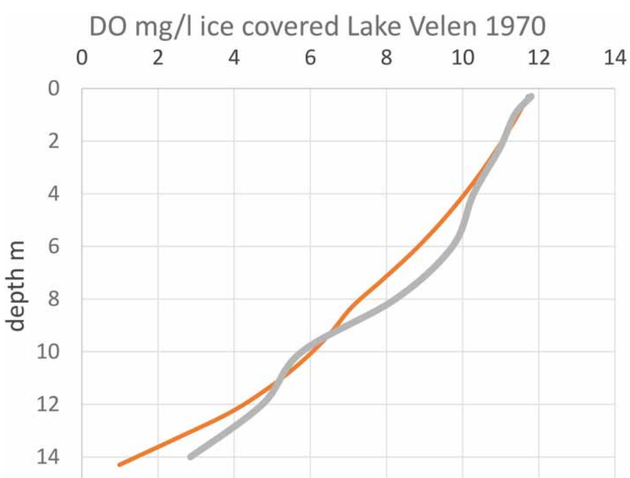

Figure 4 | Computed (thin line) and observed (thick line) DO profiles 130 days after ice formation.

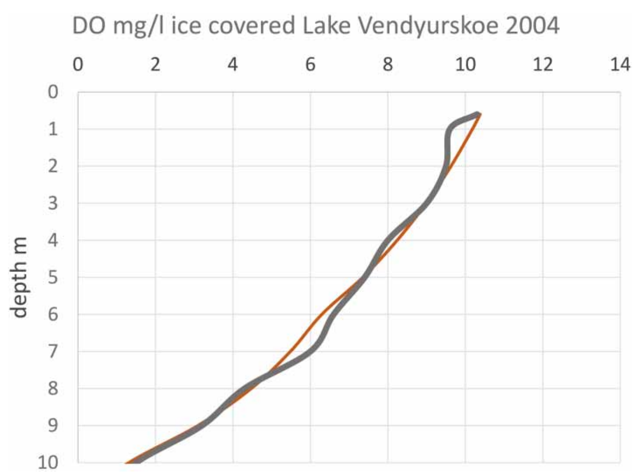

Figure 5 | Computed (thin line) and observed (thick line) DO profiles in mid-April 135 days after ice formation.

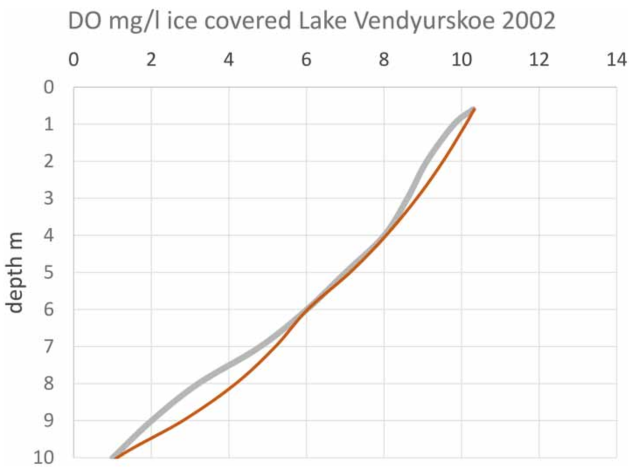

Figure 6 | Computed (thin line) and observed (thick line) DO profiles in mid-April, 145 days after ice formation.

In the study of 2001–2002, the DO was measured several times during the winter (Maher *et al.* 2004), and further analyzed by Terzhevik *et al.* (2010). The DO concentration reduced slowly over time close to the ice, 1–2 mg/L over the whole winter, as is clear from Figure 6. The development over time at many depths is shown in Figure 7, with dots as markers. In the figure the simulated decrease of DO at different depths is also shown (solid lines). The simulated decrease of DO fits well with the measurements, which suggest that the used parameters are not time dependent through an ice-covered season. However, the number of observations during the winter is sparse, so it is difficult to draw conclusions.

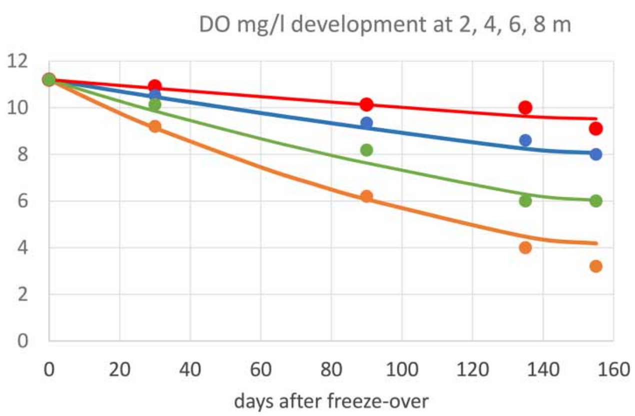

Figure 7 | DO concentration development winter 2001–2002 in Lake Vendyurskoe at, from top and down, depth 2, 4, 6, and 8 m. Dots indicate observed values and solid lines simulated values.

Convection in Lake Vendyurskoe

In early spring, the snow disappears from the ice on winter lakes and solar radiation can penetrate the ice. Convection occurs. A homothermal layer develops near the ice. An example from Lake Vendyurskoe is shown in Figure 8, showing the temperature prior to convection (or when convection had just begun) and during convection. The homothermal layer extends about 4 m down from the underside of the ice. There are 4 days in between the observations. Some convection may have occurred also before the first observations. When solar radiation penetrates into the water, primary production can start, which means that there is a source for oxygen. However, when convection is initiated, it also means that due to the motions the transfer of dissolved oxygen into the shallow sediments increases.

When studying DO profiles in Lake Vendyurskoe prior to convection and 4 days after convection began in April 2000, using data from Maher (2002) (Figure 9), it is seen that the DO concentration is homogeneous over the homothermal layer, but also that the DO concentration in the water close to the ice has decreased.

The oxygen consumption considering only the bottoms above 4 m, where there are considerable water movements due to convection, is about 12 g/m^2, which is a loss rate of

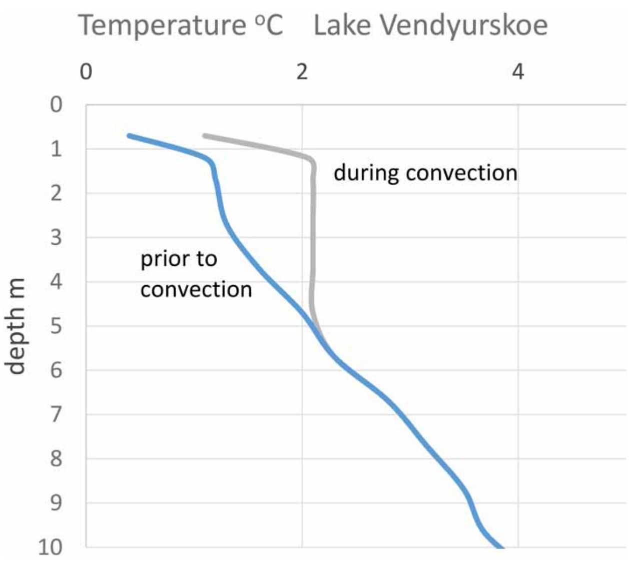

Figure 8 | Temperature profiles 17 April (prior to convection) and 21 April 2000 (during convection).

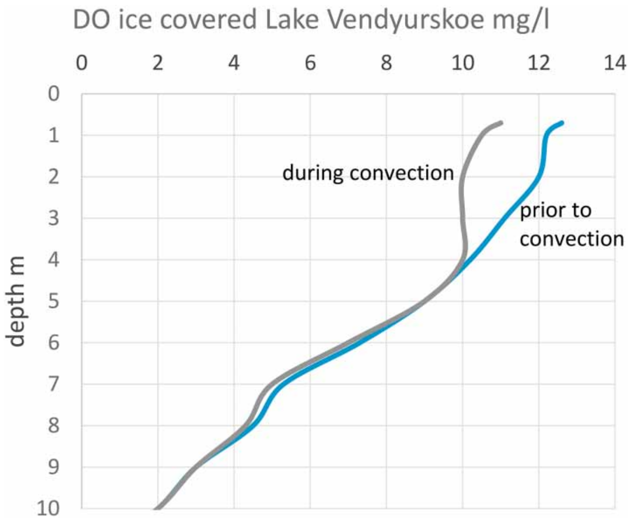

Figure 9 | Dissolved oxygen profiles 17 April (prior to convection) and 21 April 2000 (during convection).

3 g/m^2/d. This means that the coefficient k_b for the exposed bottoms above 4 m should have been about 0.3 m/d. This is at least one order of magnitude higher than when there is no convection. Although the calculations are rough, it shows the importance of water movement on the transfer of oxygen to the sediments. Later, as the solar radiation through the ice increases, the primary production will start and a source for oxygen production is available.

However, the more intense observations in Lake Vendyurskoe from April 2002 are not consistent with the observations from April 2000 discussed above. The measurements presented by Maher *et al.* (2004) show that a homothermal layer of 2.5 °C developed under the ice down to about 5 m during the period between 21 and 27 April. While the temperature profiles were almost the same all over the lake, the DO profiles were not. In some columns there was a complete redistribution due to the mixing, so that the concentration at 2.5–5 m increased at the expense of the concentration near the ice, but with no loss of oxygen. In other verticals, the total DO decreased in the same way as the observations in April 2000 showed. The observations were performed from early morning until night. This could have had an effect on the data, since the solar radiation through the ice varied significantly. Biological activity may have been a source for oxygen and compensated the loss to the sediments.

DISCUSSION AND SUMMARIZING RESULTS

The results from the DO studies in Lake Velen and Lake Vendyurskoe show that the reduction of DO in these ice-covered lakes can be estimated accounting only for the sediment oxygen demand. The same parameters, although slightly different for the two lakes, can be used for different years. Biological oxygen consumption within the water may probably still occur early after ice formation and at deep parts of the lakes, but seems to be minor. Even when it is included in the computations, the sediment uptake parameter is only reduced by 10% to get the same fit as when not accounting for the biological oxygen consumption.

The morphology is important for how DO develops throughout the winter. When the sediment uptake of DO dominates, the ratio exposed bottom area/water volume at given depths determines how fast the water becomes depleted in oxygen. Shallow lakes may be depleted in oxygen, and DO concentration can decrease to very low levels in depressions within a lake.

The initial DO concentration prior to ice formation determines how the DO profile develops during the winter. The initial concentration must be known for the simulations.

The two investigated lakes are similar in size and situated in similar environments. There are no major inflows. The approach of determining DO assuming that the decrease of DO is mainly due to the loss to the sediments can probably be applied also to very shallow lakes. The minor aeration at open water at inlets may have some influence if the lake volume is very small. Reeds along the shores may have some effect on DO.

It ought to be possible to use the suggested approach also in lakes larger than Lake Velen and Lake Vendyurskoe. At very large depths there may remain oxidizable organic matter for rather a long time, but this can be accounted for in the approach, although it would be difficult to put a value on the coefficient. In a very large lake, the water may not be horizontally homogeneous, especially if there are large inflows from rivers. Ice may form at different times over such lakes. When solar radiation penetrates the ice in early spring, convection is initiated. During a few days before biological activity commences, water with high DO content mixes with water at lower level with lower DO content at the same time as the motions of the water masses give rise to increased oxygen loss to shallow sediments. In very shallow lakes, where the convective motions reach close to the bottoms, there is a risk of low DO concentrations also near the ice.

CONCLUSIONS

The important finding of the simulations performed in this study is that the major part of oxygen consumption in ice-covered lakes is the uptake and consumption of DO in the bottom sediments. This explains the much faster decrease of DO concentration at deep water than at near-surface water. The lake morphology is important for how the DO concentration develops in winter lakes. The development of DO at different depth in a winter lake can be simulated accounting only for the DO loss to the sediments. For individual lakes, the same parameters can be used for consecutive years. Since there is no source of oxygen once a lake is ice covered, the initial conditions very much determine the DO concentration throughout the ice-covered period. What is new in the study is that the DO concentration profile can be determined from the hypsographic curve of the lake, once the initital conditions are known.

Another finding is that in early spring, when solar radiation penetrates the ice and induces convection resulting in homothermal conditions near the ice, the induced motion apart from creating homogenous DO concentration over the homothermal layer also results in a faster transfer of oxygen to the shallow bottom sediments. Until biological activity starts under the ice, the DO concentration near the ice reduces.

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