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THE EFFECTS OF CLIMATIC CHANGE ON LAKE ICE AND WATER TEMPERATURE
INTRODUCTION

Changes in the composition of the atmosphere and consequent global climate change will have a profound impact on ice cover, water balance and the stratification of lakes. These in turn may change lacustrine ecosystems.

The effects of climatic change on terrestrial and lacustrine ecosystems are studied widely within the Finnish Global Climatic Change Program (SILMU).

The aim of this paper is to discuss the effects of climatic change on lake ice and water temperature. The work is a part of SILMU. A one-dimensional vertical model for lakes has been applied in three lakes of different sizes. Using the predictions from a climatological model and from a watershed model the input data for the lake model was modified. The results are discussed in this paper.

THE MAIN EQUATIONS IN THE MODEL

The PROBE-model (Svensson 1978), is used as the hydrodynamical lake model of the project. A simple water quality model has also recently been incorporated into PROBE (Malve et al. 1991).

PROBE is a one-dimensional vertical model, in which the water is assumed to be horizontally homogenous. The model uses meteorological data, water flow and material fluxes as input data.

The basic PROBE-model calculates the development of the seasonal thermocline, water balance and the vertical distribution of eddy diffusivity. It is possible to include 14 state variables for water quality modelling.

The basic assumptions in the model are that the lake is horizontally homogenous and that gravitational effects are assumed to obey the Boussinesq approximation. The effects of the...
earth’s rotation are included in the mean flow equations, and the vertical exchange coefficient is calculated using a two-equation model for turbulence. A complete description of the model and the numerical scheme was completed by Svensson (1986).

The temperature equation in the model is:

\[
\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left( \nu_T \frac{\partial T}{\partial z} \right) + S_T
\]  

(1)

where 
- \( \tau \) = time,
- \( T \) = water temperature,
- \( z \) = the vertical coordinate (positive upwards),
- \( \nu_T \) = kinematic eddy viscosity,
- \( \sigma_T \) = turbulent Prandtl number and 
- \( S_T \) = source/sink term.

The incoming short wave radiation is included as a source term, as it penetrates the water surface and decays exponentially with depth. The source term may also include the vertical advection term. The surface boundary condition for the heat during the open water period is:

\[
\nu_T \frac{\partial T}{\partial z} = \frac{F_N}{\rho C_p}
\]

(2)

where 
- \( F_N \) = the sum of the net long wave radiation, sensible heat flux and latent heat flux,
- \( \rho \) = the water density and 
- \( C_p \) = the specific heat of water (Table 1).

At the bottom, a zero flux condition is used for the temperature equation. For the momentum equations on the surface quadratic wind stress components are used with a drag coefficient of 1.3 \( \times 10^{-3} \). At the lower boundary a zero velocity is used for the momentum equations. The vertical mixing is modelled using a two-equation turbulence model, the so-called \( k-\varepsilon \) model. A detailed description of the derivation and application of this model is given by Rodi (1980). The density of the water is approximated by a quadratic relation where it depends on the water temperature.

The equations are solved in the finite difference form as integrated forward in time using a implicit scheme and a standard tridiagonal matrix algorithm (Svensson 1978).

**THE NET HEAT FLUX, \( F_N \)**

In the present work, the heat flux calculations have been carried out as described by Sahlberg (1983) and the following is mostly a quotation by him. The fluxes controlling the heat content of a lake through the air/water interface are (Fig. 1):

| Table 1. Constant values in the heat flux calculation PROBE-model. Sahlberg (1984). |
|------------------|------------------|------------------|------------------|
| Constant         | Value            | Unit            |
| \( g \)          | 9.81             | \( m^2 s^{-2} \) |
| \( a \)          | 8.25 \( \times 10^{-6} \) | \( C \)          |
| \( \rho_0 \)      | 999.975          | \( \rho \)       |
| \( C_p \)        | 4200             | \( W m^{-2} \)  |
| \( T_m \)        | 3.98             | \( ^\circ C \)  |
| \( \sigma_0 \)   | 1396             | \( W m^{-2} \)  |
| \( e \)          | 5.67 \( \times 10^{-8} \) | \( W m^{-2}k^{-4} \) |
| \( c \)          | 0.67             | \( mb^{-1/2} \)  |
| \( b \)          | 0.05             | \( mb^{-1/2} \)  |
| \( d \)          | 0.25             | \( mb^{-1/2} \)  |
| \( \rho \)        | 1.3              | \( kg m^{-3} \) |
| \( C_s \)        | 1.41 \( \times 10^{-3} \) | \( \rho \)       |
| \( L_e \)        | 2.5 \( \times 10^3 \) | \( J kg^{-1} \) |
| \( C_t \)        | 1.32 \( \times 10^{-3} \) | \( \rho \)       |
| \( P \)          | 4370             | \( K m^{-3} \)  |
| \( K \)          | 5418             | \( \rho \)       |
| \( L \)          | 3.35 \( \times 10^3 \) | \( J kg^{-1} \) |
| \( T_k \)        | 273.15           | \( K \)         |

The equations are solved in the finite difference form as integrated forward in time using a implicit scheme and a standard tridiagonal matrix algorithm (Svensson 1978).
The amount of clouds and their height also affects short wave radiation. The “cloud function” \( T_i \) is

\[
T_{\text{LOW}} = 0.35 - 0.015 \sec z \\
T_{\text{MIDDLE}} = 0.45 - 0.01 \sec z \\
T_{\text{HIGH}} = 0.9 - 0.04 \sec z
\]

\( N_i \) is the amount of clouds of the different categories (low, middle, high). However, in this study \( N \) will represent the total amount of clouds, and, as an average, they are assumed to be middle high clouds. According to this assumption, the only “cloud function” used here is the \( T_{\text{MIDDLE}} \) function. The magnitude of \( F_s \) varies a lot during the year. During the autumn and in the beginning of the winter \( F_s \) is much smaller than, for example, the net long wave radiation.

**Net long wave radiation, \( F_1 \)**

The net long wave radiation consists of two parts, one from the water surface to the atmosphere (\( F_1 \uparrow \)) and the other from the atmosphere to the water surface (\( F_1 \downarrow \)).

\[
F_1 = F_1 \uparrow - F_1 \downarrow \quad (\text{W} \text{m}^{-2})
\]

where \( \varepsilon' \) is the emissivity of a water surface, \( \sigma \) is Stefan–Boltzmann’s constant and \( T_s \) is the water surface temperature (K).

The major problem in determining \( F_1 \) is to get a proper estimation of \( F_1 \downarrow \). Sahlberg (1983) recommends a formulation which follows from Brunt’s formula modified with a cloud factor of

\[
F_1 \downarrow = \sigma T_s^4 (c + b/\varepsilon_a) (1 + dN)
\]

where \( T_s \) is the air temperature, \( c, b, \) and \( d \) are constants, \( \varepsilon \) is the atmospheric water vapour pressure and \( N \) is the cloud coverage.

The net absorbed atmospheric radiation is computed based on a 97% absorptivity of the water surface.

The net long wave radiation is one of the major heat fluxes during cooling of a water mass. \( F_1 \) is of the order \( 10^2 \text{ W m}^{-2} \).
Sensible heat flux, $F_c$

The formulation and the values of constants are taken from Friehe & Schmitt (1976). They used a bulk aerodynamic formula:

$$F_c = \rho_a C_p \bar{U} (C_{c1} - C_{c2} (T_s - T_a)) (W m^{-2})$$  

where $\rho_a$ is the air density, $C_p$ the specific heat of water, $C_{c1}$ and $C_{c2}$ are sensible heat transfer coefficients and $\bar{U}$ is the wind velocity at 10 m.

The values of sensible heat transfer coefficients depend on air stability, $S$:

$$S_t = \bar{U} (T_s - T_a)$$

In stable conditions ($S_t < 0$) $C_{c1}=0.0026$ and $C_{c2}=0.86E^{-3}$; In unstable conditions ($0 < S_t < 0$) $C_{c1}=0.002$ and $C_{c2}=0.97E^{-3}$ and in very unstable conditions ($S_t > 25$) $C_{c1}=0.0$ and $C_{c2}=1.46E^{-3}$.

During cooling of a water mass, $F_c$ is one of the major heat fluxes and varies between 0-100 W m$^{-2}$.

Latent heat flux, $F_e$

The latent heat flux is also taken from Friehe and Schmitt (1976) using a bulk aerodynamic formula:

$$F_e = L C_e \bar{U} (Q_w - Q_a) (W m^{-2})$$

where $L$ is the latent heat of evaporation, $C_e$ is the moisture transfer coefficient and $Q_w$ and $Q_a$ are the water vapour densities close to the water surface and in the atmosphere respectively.

Heat fluxes from rivers and groundwater, $F_{in}$, are of minor importance in the cases when the retention time of the lake water is in the order of years. The sediment heat flux, $F_{in}$, has to be included in very shallow lakes in the winter time simulations.

Thus the boundary condition for the model $F_N$ can be written

$$F_N = F_1 + F_c + F_e$$

$F_1$ is so dominant during the spring and summer that it is added as a source term in the temperature equation (1). During cooling in the autumn $F_1$ has minor importance and it is incorporated in the boundary condition.

The short wave radiation through ice and snow

In the winter time, sun radiation penetration to lake water is strongly dependent on the albedo of the ice and the extinction coefficient of the ice, $K_i$. In the model, the results of Grenfell and Maygut (1977) are applied (Sahlberg 1988a, Sahlberg 1988b). They found that in the first 0.1 m of the ice, long wave radiation is absorbed and short wave radiation decays exponentially from a 0.1 m depth of ice to the water surface:

$$F_s^* = F_s (1 - \alpha \bar{i}_0 e^{-K_i h_i}) \text{ if } h_i > 0.1 m$$

where $F_s = \text{insolation falling on the upper layer of the ice}$, $F_s^* = \text{the amount of insolation reaching the ice-water interface}$, $\alpha = \text{the albedo of the ice}$, $K_i = \text{mean extinction coefficient of the ice}$ and $h_i = \text{ice thickness}$.

The variable $i_0$ is the measure for the sun radiation which penetrates the upper 0.1 m of the ice. The effect of snow cover is modelled with same the penetration formulation as used for the ice (Sahlberg 1988a and Sahlberg 1988b). The penetration of the radiation through the snow is dependent on the albedo and extinction coefficient of snow.

In the model, ice is formed when the temperature of the upper 0.2 m top layer is less than zero degrees. The growth of the ice is calculated with a degree day method following Bengtsson and Eneris (1977). For the melting of the ice, the formulation from Ashton (1983) is followed.

The decreasing of the ice thickness ($A h_i$) is a linear function of the air temperature. The final destruction of the ice happens when the thickness of the ice is less than 10 cm and the wind speed is greater than 6 m s$^{-1}$.

Ice growth

$$h_i = K_g (\sum T_a)^{1/2} (m) \text{ if } T_a < 0$$

Where $K_g$ has a value of 0.02 and $T_a$ is the daily mean air temperature.

Ice melting
\[ \Delta h_i = K_m T_a (m) \text{ if } T_a > 0 \]  \tag{14}

Where \( K_m = 4.3 \times 10^{-3} \).

The model calculates ice growth and melting based on the above equations and the direction of the net surface heat flux, \( F_q \). Growth occurs when the heat flux is directed from the ice surface towards the atmosphere and melting occurs otherwise.

In the case of ice, the boundary conditions for the momentum and temperature equations are put to zero value at the upper boundary.

**RESEARCH LAKES AND CALIBRATION DATA**

**Lakes**

The PROBE-model has, as now, been applied in six lakes with different catchment characteristics: Lappajarvi and the last basin in the Langelmavesi watercourse, influenced by agriculture as well as wastewater loadings; Villikkalanjarvi and Pyhajarvi, which are mainly affected by agriculture; Kalliojarvi, which is impacted by forestry measures; and the pristine Hietajarvi. These lakes are located in southern and eastern parts of the country.

Predictions of climate change effects have up till now been conducted on lakes Lappajarvi, Langelmavesi and Kalliojarvi (Fig. 2). Some physical characteristics of the lake basins are shown in Table 2. The lakes have quite different sizes. Lake Kalliojarvi is the smallest and also quite shallow. It is sheltered by the hills on the eastern shore. Lappajarvi is a very big lake and has a theoretical retention time of 2.5 years, whereas the retention time for Lake Langelmavesi is only 60 days. The retention time for Lake Kalliojarvi is 13 months.

**Available calibration data**

The available data is best for Lake Lappajarvi. Surface temperature and water level in the lake are observed daily during the open water period by the Hydrological Office. The ice and snow thickness is measured three times each winter month. Ice formation and the destruction of the ice was also observed. In the summer of 1988, an automatic thermistor chain recorded water temperature in the lake for about two months.

For lake Lappajarvi, synoptic meteorological data from Kauhava airport (about 30 km west) was used. For Langelmavesi, the data was obtained from Tampere airport, which is about 20 km to the west of Lake Langelmavesi. For Kalliojarvi, weather data was obtained from Kuorevesi airport, which is situated about 16 km east from the lake. The model read the wind data in three hours intervals, humidity and cloud data were updated every 6 hours and daily mean values were used for air temperature.

The snow and ice thickness for Kalliojarvi were observed five times during the winter and, for Langelmavesi, the data was obtained from the nearby outlet of the lake in Kaivanto.

The daily inflows and outflows were available for each lake.

When water samples were also taken, water temperature was measured on each lake every week during the summer.

**Hydrological conditions during calibration years**

In the following, the weather and runoff conditions during the period 1986-91 is described by quoting the monthly reports of the Hydrological Office. The monthly values of temperature and wind speed in Kauhava are in Fig. 3. The outflow of Lake Lappajarvi is in Fig. 4.

In 1986, spring began rather early, snow began to melt about two weeks earlier than normal, as did the ice cover over the watercourses. The summer was warm and dry. The autumn was exceptionally wet and during the first weeks also unusually cold. The end of the autumn season was warmer than normal, with the mean November temperature as much as 3°C above the normal mean. Due both to the late onset of winter and a November rainfall about double the monthly mean, very high runoff values to the watercourses were recorded in many areas. Lake water levels were generally 50-100 cm above the seasonal mean in late November and early December and flow rates were as high as two- or threefold compared with the seasonal mean flow rates. Water resources were abundant in November-December.

In January 1987 the weather was dominated by extreme frosts, which in southern Finland were the heaviest recorded in the twentieth century. Ice covers grew considerably. The spring runoff was lighter than normal. The summer season was one of the coldest and and wettest of
Fig. 2. The study lakes.

Table 2. Physical characteristics of the lake basins

<table>
<thead>
<tr>
<th>Lake</th>
<th>Area $\text{km}^2$</th>
<th>Volume $10^9\text{m}^3$</th>
<th>Mean depth $\text{m}$</th>
<th>Max depth $\text{m}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kalliojärvi</td>
<td>0.3</td>
<td>1.1</td>
<td>4.4</td>
<td>12.0</td>
</tr>
<tr>
<td>Längelmävesi</td>
<td>11.2</td>
<td>95.0</td>
<td>8.3</td>
<td>41.0</td>
</tr>
<tr>
<td>Lappajärvi</td>
<td>181</td>
<td>1125</td>
<td>7.4</td>
<td>38.0</td>
</tr>
</tbody>
</table>
Fig. 3. Wind speed and air temperature at Kauhava airport 1986-91 and temperature in a normal year.
the century; precipitation levels were high, evaporation was unusually low and water temperatures remained well below average. At the beginning of August, heavy rainfall caused record-breaking runoff volumes into watercourses. October and November were drier than normal. By the end of the year water levels were generally decreasing rapidly.

In 1988, melting of the thick snow cover coupled with heavy rainfall raised water levels in spring and early summer to record heights for the current century. June and July were exceptionally warm. Surface temperatures generally reached 26°C and more in late June and July, setting new all-time records for Finland. The autumn season began with near average conditions, but November was colder and frostier than usual and winter came about a month earlier than normal and water levels began to fall.

In the beginning of 1989 water levels were high but very low by the end of the year. The mean annual temperature was exceptionally high. The winter was the warmest on record. Snow began to melt early and the spring high water occurred about two months earlier than normal. The summer was long, warm and dry. As a result, water levels decreased rapidly and were very low in late summer and autumn and still below normal level at the end of the year.

Both the weather and the hydrological conditions in 1990 were similar to those of 1989. High levels of precipitation in the first months of the year occurred both as snow and as rain. Melting of snow began several weeks earlier than normally. Both spring floods and the break-up of lake ice were earlier than ever before on record. The summer was long and dry, although not quite as warm as the two preceding summer seasons. Water resources decreased during 1990 and were at a very low level by the end of the year.

The recent trend of unusually warm years continued in 1991. Water reserves were low at the beginning of year. The snow cover was thinner than normal and melting of snow occurred early. Sudden high meltwater runoff peaks occurred in small watercourses but in large rivers the spring flood was lower than normal. Heavy rainfall occurred during spring and again during the autumn, as a result of which water resources were abundant by the end of the year. Water temperatures in August-September were exceptionally high, and freezing of the watercourses was several weeks late towards the end of the year.

PREDICTIONS

The calibration of the model

The models were first calibrated against the lake water temperatures during the open water period. The main parameters for the calibration were the extinction coefficient and the wind. In the
Table 3. The values of calibration coefficient in reducing wind at Lake Kalliojärvi.

<table>
<thead>
<tr>
<th>Wind direction</th>
<th>Calibration coefficient, when wind velocity</th>
<th>&lt;4 ms(^{-1})</th>
<th>&gt;4 ms(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>310°–20°</td>
<td></td>
<td>1.0</td>
<td>0.5</td>
</tr>
<tr>
<td>20°–130°</td>
<td></td>
<td>0.8</td>
<td>0.4</td>
</tr>
<tr>
<td>130°–200°</td>
<td></td>
<td>1.0</td>
<td>0.7</td>
</tr>
<tr>
<td>200°–310°</td>
<td></td>
<td>0.8</td>
<td>0.4</td>
</tr>
</tbody>
</table>

In the case of Lappajarvi the wind velocity was somewhat increased from the velocities observed at the airport (Malve et al., 1991). Because Lake Kalliojärvi is small and narrow (Fig. 2), the wind calibration was used reducing the wind velocities especially in the direction of the transversal axis of the lake (Table 3). The need to reduce wind stress and solar radiation, because of hills on the shoreline of the small lakes has been discussed by Sahlberg (1988b). He suggested an area dependent wind stress reduction on small lakes. This method has not been tested in Finnish applications, since the calibration method has proved to be acceptable.

In Fig. 5, the calculated and observed surface temperatures of Lake Lappajarvi are seen. The warming and cooling of the lake surface temperature is simulated well. The maximum values differ due to the local heating of surface water at the shallow observation site near the shore, whereas the model calculates the mean value of the surface temperature in the whole lake. The vertical temperature profiles were also calculated well. Fig. 6 is an example from Lake Lappajarvi in summer 1988. The greatest deviations between calculated and observed temperature occur in late August when the model calculates the stratification to be about 10 days longer than was observed.

The calibration was done further using ice data. The model was very accurate in calculating the formation and destruction dates of the ice on Lake Lappajarvi. The mean error in Lappajarvi was 1.5 days (Malve et al., 1991). The ice thickness calculation deviated maximally about 10 cm and this was during the mild winter of 1988-89, when the snow melted several times on the ice (Fig. 7). The snow observations turned out to be too few for the model to calculate the heat flux correctly during the period.

Simulation with a changed climate

The climatological predictions in the study are based on the results from the GISS (Goddard Institute for Space Studies) model. The predictions of this model were the best choice of The European Workshop on Interrelated Bioclimatic and Land Use Changes in 1987 for simulations of climatic change (Bach 1989).

Vehviläinen and Lohvansuu (1991) have used a watershed model to predict the effects of climatic change on river discharges and snow cover in Finland. Their work was based on the seasonal temperature and precipitation predictions obtained from the GISS model when a doubling of carbon dioxide relative to present day values was assumed (the 2xCO\(_2\) scenario). The temperature and precipitation predictions of the GISS-model for Finland are in Tables 4 and 5. For the lake model, the input data of snow thickness and water flow were obtained from their calculations.

In the present lake model applications meteorological data (air temperature, humidity, cloudiness and wind speed) from the years 1986-1991 were used. Air temperature was changed according to the prediction of the GISS-model. Predictions of changes in cloudiness, humidity
Fig. 6. The water temperature simulation in Lake Lappajärvi in summer 1988.

Fig. 7. Simulated and observed ice thickness on Lake Lappajärvi (01 Oct 1986 – 30 Apr 1989).
and wind were not available. For the cloudiness and humidity it was assumed that their relative seasonal change is the same as the the change in precipitation data. For the wind measured values were used. In the earlier work (Kauppi et al. 1992) PROBE predictions were done with no climatic change in humidity and cloudiness. In the future the climate generator developed within the SILMU-program will be used for producing the meteorological input data.

**RESULTS**

In Lake Lappajarvi, which is the greatest of all the study lakes, simulations were done in a period extending over three winters (Fig. 8). The ice cover period in Lappajarvi will be significantly shorter in a 2xCO₂ than in the present winter situation. With all three winter data the formation of the ice cover was delayed about two weeks. The date of ice melting occurred 1-2 months earlier. The greatest change occurred in the winter of 1988-89. During the winters of 1987-88 and 1988-89 several periods with open water occurred.

Also the simulations of the PROBE-model for the other two lakes indicated that the ice-covered period will be about two months shorter than in the present (1xCO₂) winter situation. The first freezing of the lakes will be delayed (Fig. 8). During the winter time there will be ice-free periods, especially on large lakes, and the final ice break-up will occur one or two months earlier than today.

Because of the earlier ice-melting in the spring, the wind will mix the lake, and hence the temperature of the hypolimnion will slightly decrease compared with the present situation. On the other hand, there will be no oxygen depletion in the hypolimnion in spring, due to reaeration and mixing. In small sheltered lakes the turnover conditions may change entirely. E.g. in Lake Kalliojärvi which normally has no spring turnover, because the heat obtained due to short-wave radiation penetration through the snowfree ice rather quickly causes a weak summer stratification still when the ice is present. (Figs. 8 and 10). In the new situation (the 2xCO₂ scenario) with shorter ice cover period the lake would also mix in spring.

In summer the thermal stratification of the lakes will be steeper and the temperature of the epilimnion will increase about 5-6 °C (Fig. 9). The thermocline will be about 5-8 m higher than today and the hypolimnion will be about 2-3 °C colder. The longer stagnation period means, in practice, more serious problems with oxygen depletion in the hypolimnion during the late summer.

The results do not differ significantly from the earlier results (Kauppi et al 1992), where the humidity and cloud data were not changed. In the present work, this change was done assu-
Fig. 8. The ice cover in research lakes in present climate and in the 2 x CO₂ climate.
The surface water temperature of Lake Lappajärvi in the present situation and in the 2 x CO₂ climate.

Fig. 9. The surface water temperature of Lake Kalliojärvi in the present situation and in the 2 x CO₂ climate.

Fig. 10. The surface water temperature of Lake Langelmavesi in the present situation and in the 2 x CO₂ climate.

The water temperature of a lake is sensitive to wind speed and solar radiation. Thus, the results of the ice-cover period must be considered to be more reliable than the summer results.

The effect of wind is the main driving force for the vertical mixing in lakes. In the present work the wind data was not changed. The last few years have been exceptionally warm in Finland and also more windy than before. If this means that winds will also increase in the 2xCO₂ climate.

The PROBE-model, with the heat flux calculations presented above, has proved to be very good tool in calculating the vertical temperature structure in a lake with a fairly open surface. There is not much calibration needed and the necessary data can be obtained from synoptic weather stations and water temperature measurements from the water authorities.

Kuusisto (1989) presented results of the change of the ice cover period in Finland in the GISS 2xCO₂-scenario. According to his results the ice cover length will be 40-60 days shorter than in the present climate. This corresponds well with present results. Kuusisto based his calculations on a statistical dependency between mean annual air temperature and the length of the ice cover.

The simulation of clouds is a problem in global climatic models as Dickinson (1986) and Bach (1989) have pointed out. The wide variety of cloud types and the fact that most cloud properties are under the grid scale of the present climatic models make cloud prediction uncertain.

The sensitivity of the model to the change of air temperature was studied in the application of Lake Langelmavesi. Air temperature was assumed to increase one degree less compared to the estimations of the GISS-model. In this case the ice-cover period was only about one month shorter than in the present situation. Ice-free periods did not occur in winter. Also the results of in summer time became more comparable to those of in present situation.

The water temperature of a lake is sensitive to wind speed and solar radiation. Thus, the results of the ice-cover period must be considered to be more reliable than the summer results.
te, the stratification will be affected. This may lead to a more sharper thermocline than predicted now and great variations in summer surface temperatures in some lakes, when the cold hypolimnetic water is mixed with surface waters.

TIIVISTELMÄ


Tuuliarvoja ei muutettu.

Tulokset osoittavat, että järven ensiääräisenä tapahtuman kaksi viikkoa yli sataarmeistarkka ja jääpeitekasvun kaksikymmentä viikkoa eli. Tuulilaskennan ostalta.

Järven lämpötilakerrostuneisuus tulisi olemaan erityisesti huomattavasti jyrkempi ja harppauskerroksia jää noin 5-8 m ylemmäksi kuin nyky-ilmastossa. Pyällysveden lämpötila nosee keskimäärin 5-6 °C. Alusveden lämpötila jää noin 2-3 °C yllemmäksi kuin nykyvaa. Kesän kerrostuneisuuskauden pituuden kasvu liisa alusveden happiongelmiin.

Mallin antamat tulokset ovat luotettavimpia talven jääpeite- ja jään vuodenaikaa laskennallista. Tuulien- ja tuulienlaskennan pumittuissa vuosissa kesän kerrostuneisuuskauden kasvu liisa alusveden havainnoista.

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