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MEASUREMENTS AND MODELLING OF THE WATER – ICE HEAT FLUX IN NATURAL WATERS

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ABSTRACT

Ice in natural waters grows and decays as forced by the fluxes through the upper and lower boundaries. In particular, the flux at the lower boundary – i.e. the heat flux from the liquid water body into the bottom of the ice sheet – is not very well known quantity. This question is approached by measurements and mathematical modelling. The data are from Saroma-ko lagoon, a saline lake on the northern coast of Hokkaido, and Lake Pääjärvi, a fresh water basin in southern Finland. Three-dimensional current velocity, temperature and salinity were measured at a fixed depth, and the resulting heat flux was normally 5–10 W/m² in both basins, a bit more in Saroma. But even in Lake Pääjärvi, which is a rather quiet water body in wintertime (total ice coverage with very weak currents) the heat flux from the water is important in the heat budget of the ice sheet. A three-layer (snow/snow-ice/congelation ice) model is used to examine the evolution of ice thickness and temperature.

Key words: heat flux from water, floating ice, growth, melting, measurements

1. INTRODUCTION

The thermodynamic evolution of the thickness of ice in lakes and seas is forced by the heat fluxes at its upper and lower boundaries. Much research work has been done on the heat budget at the upper surface by measurement campaigns and analytical and numerical modelling. The lower boundary, i.e. the heat flux from the liquid water body into the bottom of the ice sheet is, however, less understood. This heat flux is of the order of 1–100 W/m², which makes a significant contribution to the ice thickness evolution. It is also as such important for the heat content of the water body itself.

In this work the heat flux from the water body to ice is examined by measurements and mathematical modelling. The data of two experimental sites are utilised: Saroma-ko Lagoon on the northern coast of Hokkaido and Lake Pääjärvi in southern Finland. Both sites have provided very good data for the seasonal evolution of the ice conditions. The modelling work

includes analytical and numerical models, which have been earlier used in both basins (*Leppäranta and Uusikivi, 2002; Shirasawa et al., 2005*). The data collected are illustrated with examples, and the methods used to estimate the heat flux are explained. The final analysis is in progress and will be completed in summer 2006.

2. MATERIAL

Saroma-ko Lagoon is located on the northern coast of Hokkaido. Its surface area is 149.2 km², maximum depth is 19.5 m, and mean depth is 14.5 m. The lagoon has two inlets, which are connected to the Sea of Okhotsk. Freshwater input by two major rivers keeps the salinity to less than 32 PSU (Practical Salinity Unit $\equiv 10^{-3}$). The average freezing date is January 11th. At the beginning of the freeze-up, atmospheric forcing keeps the lagoon open and frazil ice is formed. Ice reaches a maximum thickness of 35–62 cm annually in the southeastern part of the lagoon. There are large spatial variations in ice thickness due to the oceanic heat flux from the inlets.

Lake Pääjärvi is localized in the southern boreal vegetation zone at 61°04' N, 25°08' E in southern Finland. Annual mean air temperature is 3.6°C, and annual mean precipitation is 619 mm. The lake is 13.4 km² in area, 87 m in maximum depth, and 14 m in mean depth, and it is a dimictic, oligo-mesotrophic, brown colour water body, with light attenuation coefficient of about 1 m⁻¹. The average freezing date is November 30th (*Kuusisto, 1994*). The lake becomes fully ice covered every winter, and the ice reaches a maximum thickness of 30–80 cm annually.

Field programme is ongoing in both sites. The objective is to map, analyse and model the evolution of ice thickness and the heat budget. The observations consisted of surface layer meteorology, and temperature and light level in snow, ice and water at various depths. Three-dimensional current velocity, temperature and salinity were measured at a fixed depth, and the resulting heat flux was estimated from the data by the bulk transfer method. Even in Lake Pääjärvi, which is a rather quiet water body in wintertime (full ice cover with very weak currents) the heat flux from the water is important in the heat budget of the ice sheet. The observation periods were from December to April.

3. RESULTS

An example of the data outcome is shown in Fig. 1 for Lake Pääjärvi. The period of observations covered the ice season with freezing and breakup dates included. The deployment was made on December 5th when the lake was open and the water temperature was about 3°C. The freezing date was December 19th, and the breakup date was April 27th, about a week before the end of the data collection. The lake has a solid ice cover, and thus there is not much momentum transfer from the wind to the water body. The main forcing of the circulation comes from small rivers and heating of the water body by the heat flux from the bottom sediments and the solar heating which becomes active as soon as the snow has melted.

At the time of the freezing of the lake the current speed goes down and the water temperature becomes rather stable. There is, however, a weak warming trend in the middle of the winter due to the heat flux from the bottom sediments. In the beginning of April, as the snow has

melted, the water temperature increases fast, and at the time of the ice breakup the water body is already heated to almost 4°C. Also with the solar heating penetrating into the water the dynamics of the water body becomes more active.

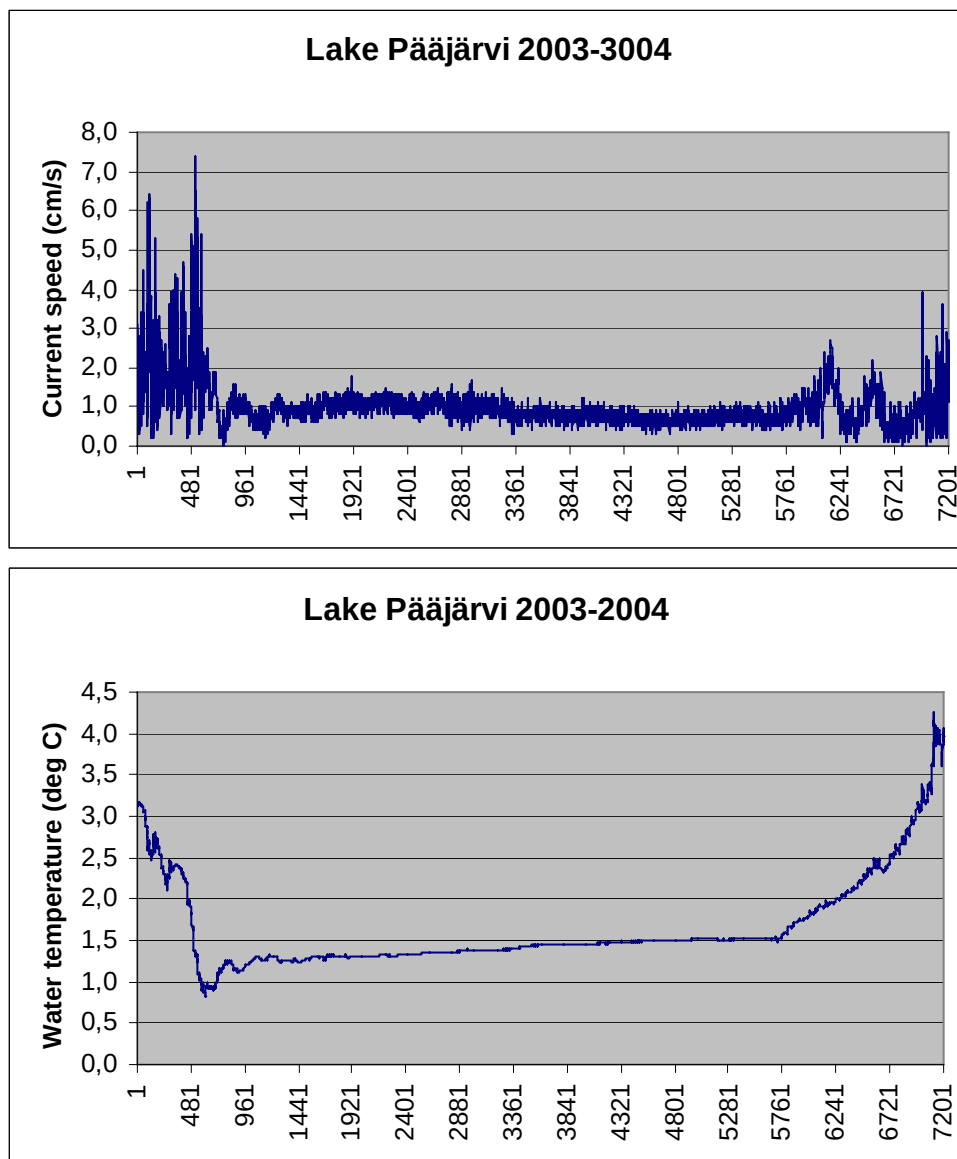


Fig. 1. Current speed and water temperature in Lake Pääjärvi, from December 5, 2003 to May 2, 2004, at the depth of 5 m. The unit in the horizontal axis is 30 minutes, starting point December 5th, 2003 at 00 hrs.

Another example of the current measurements is shown in Fig. 2 for Saroma-ko Lagoon. The current speed was as high as 10 cm/s before ice freeze-up and reduced to 2-3 cm/s during the ice-covered period, which might cause reduction of vertical mixing of the water under the ice. The water temperature and salinity decreased till the temperature reached the freezing point at late January. The salinity increased slightly under the sea ice due to brine extraction from the growing sea ice. The temperature increased rapidly at late March, and correspondingly ice breakup started and the mixing increased due to the increase of the current speed. The water level varied about 0.7 m at maximum under the ice at the tidal cycle.

The heat flux from the water body to the ice bottom can be obtained directly from current and temperature data using the bulk formula method. Indirectly, it is possible to estimate the heat flux as a residual to explain observed evolution of ice thickness.

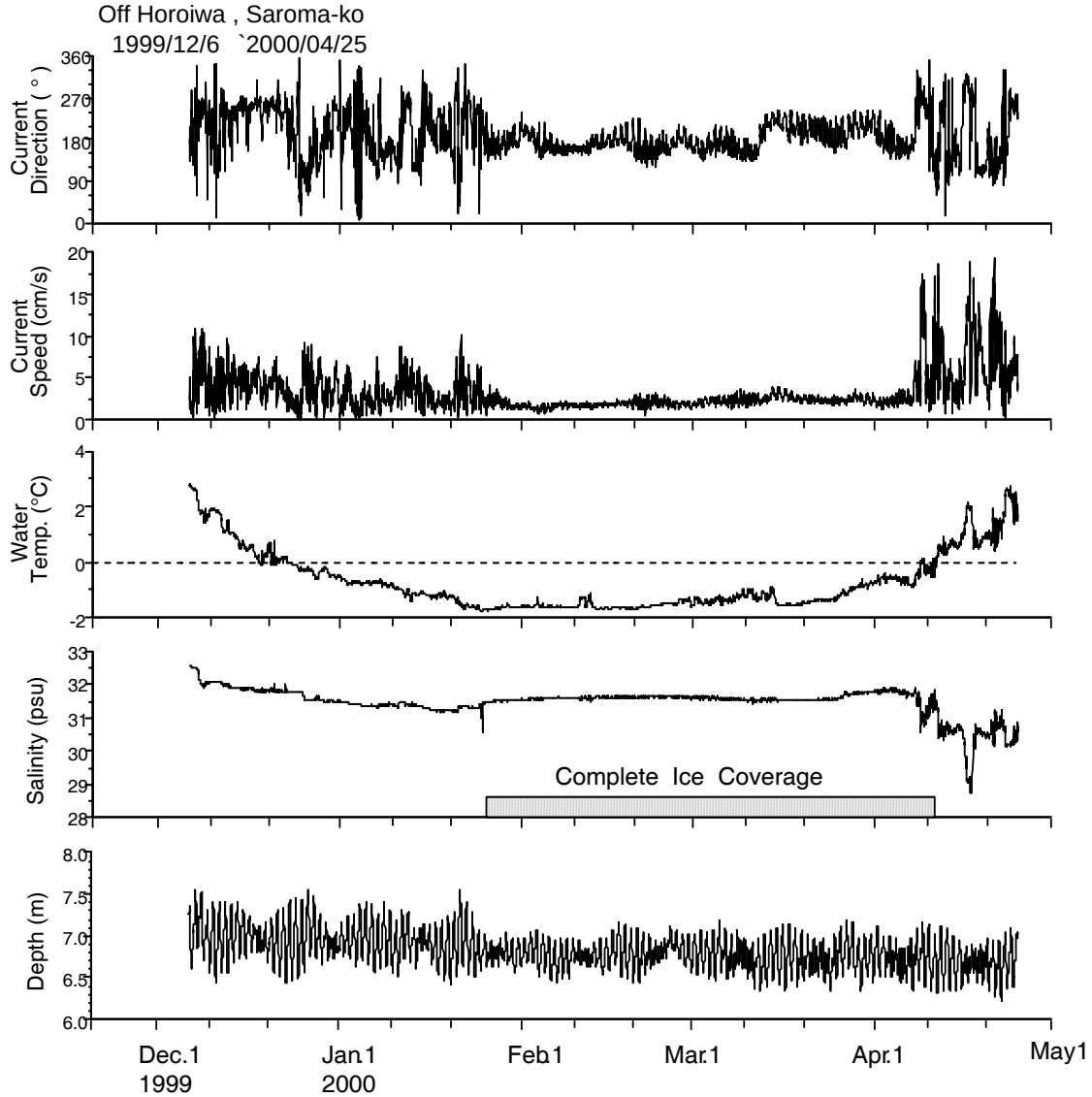


Fig 2. Time series of currents, temperature and salinity obtained from the mooring station at the central area of Saroma-ko lagoon during the period from December 6, 1999 through April 25, 2000 [after Shirasawa and Leppäranta, 2003].

Using the bulk formula the heat flux is expressed as

$$Q_w = \rho_w c_w C_H (T - T_o) U_w \quad (1)$$

where ρ_w is water density, c_w is the specific heat of water, C_H is the heat exchange coefficient, T and T_o are water temperature and ice bottom temperature, and U_w is current speed; C_H , T , and U_w depend on the depth z ($z = 0$ at the bottom of the ice sheet). The heat exchange

coefficient used in Saroma-ko lagoon has been $2 \cdot 10^{-3}$ (Shirasawa *et al.*, 1997). The value of $0.39 \cdot 10^{-3}$ was obtained from Santala Bay, the Gulf of Finland during the HANKO experiment, and from the southern Bothnian Bay during the BALTEX/BASIS experiment (Shirasawa *et al.*, 2003). The value of $0.39 \cdot 10^{-3}$ gives the heat flux magnitude of 5 W/m^2 in Lake Pääjärvi.

Another approach is the profile method:

$$Q_w = \kappa_w \left. \frac{\partial T}{\partial z} \right|_{z=0-} \quad (2)$$

where κ_w is the thermal conductivity. The lower limit of the thermal conductivity is the molecular thermal conductivity, $\kappa_w = 0.6 \text{ W/(m } ^\circ\text{C)}$. Taking the vertical temperature gradient from the Pääjärvi data as $1.5^\circ\text{C}/5\text{m}$ gives this heat flux as 0.2 W/m^2 for the heat flux. This is too low, and therefore the heat conduction beneath the ice cover is much stronger than molecular conduction.

Third method is indirect, providing the heat flux as the residual

$$Q_w = \kappa \left. \frac{\partial T}{\partial z} \right|_{z=0+} - \rho L \frac{dh}{dt} \quad (3)$$

where $\kappa = 2.1 \text{ W/(m } ^\circ\text{C)}$ is the thermal conductivity of ice, ρ is ice density, L is latent heat of freezing, and h is ice thickness, and t is time. The temperature gradient in the ice sheet is typically $1^\circ\text{C}/0.1\text{m}$ in cold periods, and therefore the first term on the right hand side is 20 W/m^2 . The ice grows by 1 cm/day , and therefore the heat released in the ice growth is of the same order of magnitude. Thus this method gives the heat flux as the difference of two large terms, and consequently both terms must be determined to a high degree of accuracy in order to estimate the heat flux from the water.

In spite of their crudeness, simple analytic models provide rather good results for the climatology of sea ice thickness (Leppäranta, 1993). The basic analytic model for congelation ice growth is the Stefan's law (Stefan, 1891). A generalized form including the oceanic heat flux and air – ice interaction reads in differential form

$$\frac{dh}{dt} = a \frac{T_f - T_a}{h + d} - \frac{Q_w}{\rho L} \quad (T_a \leq T_f) \quad (4)$$

where $a = \kappa/(\rho L)$, T_f is the freezing point temperature, T_a is air temperature, and d is effective insulating thickness of the near surface air–snow buffer. The Stefan solution $h^2 = 2aS$, where $S = \int_0^t \max(0, T_f - T_a) ds$, comes from $d = Q_w = 0$ and can be taken as an upper bound for favourable ice growth conditions.

Consider now the role of the heat flux from water. If $Q_w \neq 0$ and $T_a = \text{constant}$, there is a steady-state solution

$$h + d = \kappa(T_f - T_a)/Q_w \quad (5)$$

Since $h \geq 0$ and $d \geq 0$, it is seen that the heat flux of $Q_w = \kappa(T_f - T_a)/d$ is sufficient to prevent ice formation; this works as a simple polygonal equation. If the heat flux is less than that, the steady state ice thickness becomes $h^* = \kappa(T_f - T_a)/Q_w - d$. With $T_a = \text{constant}$, the time evolution is obtained from Eq. (1) as

$$\frac{h}{h^*} + \log\left(1 - \frac{h}{h^*}\right) = -\frac{\kappa(T_f - T_a)}{h^{*2}} t \quad (6)$$

In the beginning ($h \ll h^*$) the ice grows as proportional to the square root of time

For a numerical modelling effort a three-layer (snow/snow-ice/congelation ice) model is used to examine the evolution of ice thickness and temperature, giving good fit with validation data from specific field campaigns.

4. CONCLUDING REMARKS

The heat flux from the water body to ice has been examined by measurements and mathematical modelling. Ice in natural waters grows and decays as forced by the fluxes through the upper and lower boundaries. In particular, the flux at the lower boundary – i.e. the heat flux from the liquid water body into the bottom of the ice sheet – is not very well known quantity. The data of two experimental sites are utilised: Saroma-ko Lagoon on the northern coast of Hokkaido and Lake Pääjärvi in southern Finland. The data collected are illustrated with examples, and the methods used to estimate the heat flux are explained. Three-dimensional current velocity, temperature and salinity were measured at a fixed depth, and the resulting heat flux was normally 5–10 W/m² in both basins. The final analysis is in progress and will be completed in summer 2006.

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