Nordic Hydrology, 28, 1997, 129-152

Temperature and Salt Content Regimes in Three Shallow Ice-Covered Lakes

2. Heat and Mass Fluxes

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A field study was carried out in three small shallow ice-covered lakes to study heat and mass fluxes and their spatial and temporal variability. During the main part of the winter, the heat flux at the ice-water interface, being of the order 0.5-1 W m^{-2}, was dominated by conduction from water to ice and did not show any significant variations in time or space. The heat flux from sediments to water was the main source for the lake water heating during early and mid-winter, being depth-dependent and 1-4.5 W m^{-2} in early winter, and 0.5-3 W m^{-2} in late winter. A heat transport from shallow regions to deep parts was shown to occur during the winter, being of the same order as the vertical fluxes, and should thus be accounted for in any attempt to predict the temperature evolution in an ice covered lake.

The salt flux from sediments was found to be of the order 1-10 $\times$ 10^{-10} kg m^{-2} s^{-1}. A comparison of this flux with salt content changes indicates that the former is of the same order as the horizontal salt flux which is directed from shallow regions to the deeper parts of a lake during winter.

Introduction

The heat content changes in lakes during the ice-covered period are normally quite small compared to those experienced during spring, summer and autumn, but can in some shallow lakes be about 20% of the annual heat budget (see, *e.g.* Bengtsson *et al.* 1996). The main reason for this is that the heat fluxes are small, of the order of 1

W m^{-2}, during the main part of the winter. However, these heat fluxes have a significant influence on the lake environment, since they are directly or indirectly linked to circulation and mixing characteristics.

During the winter 1994/1995 a field study was made in three small and shallow Karelian lakes (Russia) without significant throughflow, focussing on the temperature and solute concentration regimes. The lakes were covered with ice from the beginning of November 1994 until the middle of May 1995. A layer of snow was present on top of the ice from early winter until mid-April, isolating the water from solar radiation. In mid-April the snow melted rapidly, and solar heating below the ice became apparent. A description of the development of the temperature and salt content distribution has been given in Malm *et al.* (1997), hereafter referred to as paper 1. The present article focusses on heat and mass fluxes at the boundaries and on horizontal heat and mass fluxes within the lake, during the ice-covered period. The heat and mass fluxes are determined on the basis of temperature and conductivity measurements at the sediment-water and ice-water interfaces, as well as on solar radiation measurements. The objectives of the study are,

- to determine the solar radiation heat flux below the ice during late winter and its importance on the thermo- and hydrodynamics of ice-covered lakes;
- to determine the spatial and temporal variability in the conductive heat flux from water to ice;
- to determine the spatial and temporal variability in the sediment heat and mass fluxes;
- to determine the magnitude of horizontal heat and mass fluxes and compare these with the vertical fluxes.

The article begins with a review of previous studies on heat fluxes in ice-covered lakes, followed by a description of the investigated lakes and types of measurements, measurement results, and conclusions.

Review of Previous Studies on Heat Fluxes in Ice-Covered Lakes

During winter, the lake waters are warmed by the 1) heat transferred from bottom sediments and 2) solar radiation penetrating through the ice cover. Heat is lost by 3) conduction from water to ice and by 4) heat advection with river inflow/outflow. The present study focusses on the first three heat exchange mechanisms, since heat advection by river throughflow in the lakes investigated here is negligible. Results of previous studies on these heat fluxes are summarized below.

The *sediment heat flux* has been the focus of several studies (see for instance, Johnsson and Likens 1967; Golosov and Kreiman 1992; Svensson 1987; Likens *et al.* 1985; and Bengtsson *et al.* 1996) and its physics is well understood. The sediments gain heat from the water during spring and summer and lose heat to the water during autumn and winter (Birge *et al.* 1927). The main reason for this heat transfer at the sediment-water interface in shallow lakes is the annual temperature fluctuations in the water. Therefore, the heat flux is higher in shallow areas compared to sediments located below the position of the summer thermocline. During the ice-covered period the heat transfer from sediments to water decreases with time, due to the continuous drainage of heat. Typical sediment heat flux values in winter in shallow lakes, as obtained by several investigators (see, for instance, Likens and Johnson 1969; Thanderz 1973; Welch and Bergmann 1985; Svensson 1986; and Bengtsson and Svensson 1996) are about 1-4 W m^{-2}, where the higher values correspond to shallow sediments and early winter and the smaller values to big depths and late winter. Most earlier investigations have, however, mainly considered the heat flux dynamics at a few points within the lake. Therefore, more information about the spatial variability in sediment heat flux is needed, especially in shallow lakes where the annual temperature amplitude in water is about the same at different depths.

In late winter or early spring when the snow on the ice starts to disappear, *solar radiation* is able to penetrate through the ice cover and warm the water below (Hutchinson 1957). Very few direct measurements of solar radiation beneath the ice with submerged pyranometers have been made. One such field study was made by Svensson (1987) in four Swedish lakes. He found that the solar radiation flux during the last month before ice break-up was about 10-40 W m^{-2}. Similar values was obtained by Bengtsson *et al.* (1996) from heat content change calculations in a small Karelian lake during late winter. During early and mid-winter, the snow cover on top of the ice isolates the water rather effectively from solar radiation, so the solar radiation penetrating the ice is usually neglected, when calculating the heat budget of ice-covered lakes. A rough estimate of the amount of penetrating solar radiation to the water during this period can be made from the expression (Gu and Stefan 1990)

$$R_{wi} = R(1-A_s)(1-\alpha_s)e^{-\eta_s d_s}(1-A_i)(1-\alpha_i)e^{-\eta_i d_i} \tag{1}$$

where R_{wi} is solar radiation penetrating into the water; R is solar radiation reaching the snow surface; A is albedo; α is surface adsorption coefficient; η is extinction coefficient; d is layer thickness; and the subscripts s and i stand for snow and ice, respectively. The following values are adopted (from Gu and Stefan 1990): A_s=0.6, A_i=0.55, α_s=0.25, α_i=0.25, η_s=30 m^{-1}, and η_i=4 m^{-1}. If the following typical thicknesses of snow and ice, d_s=0.1 m, and d_i=0.5 m, are considered, the amount of solar radiation that reaches the water is reduced by a factor 0.7×10^{-3} compared to the amount that reaches the snow surface. If R is set to 100 W m^{-2}, as a upper value on the daily average for the period December to March, a daily average solar radiation flux to the water less than 0.1 W m^{-2} is obtained. This is smaller by one order of magnitude compared to the two other main heat fluxes, from sediments to water and from water to ice.

The conductive *heat flux from water to ice* is the main heat sink in a lake without significant throughflow. Few field studies of this heat flux have been made and little is known about its spatial and temporal variability. In one study by Bengtsson and Svensson (1996) in eight Swedish lakes, heat flux values from water to ice of the order 0-10 W m^{-2} was obtained. A mid-winter estimate by Ellis *et al.* (1991) for a small and shallow Minnesota lake gave a heat flux of 2.8 W m^{-2}. Numerical calculations by Gu and Stefan (1990) for Lake Calhoun, and Sahlberg (1988) for Lake Tulebo showed that heat flux values lie within a range of 1-4 and 1-8 W m^{-2}, respectively. The temporal variation during early and mid-winter for these two cases was essentially different, with a continuous decrease for Lake Calhoun, and a continuous increase for Lake Tulebo.

Description of the Investigated Lakes and Performed Measurements

Lakes Uros, Rindozero, and Vendyurskoe are located in the southern part of the Republic of Karelia, Russia (latitude 62°10'-62°20'N, longitude 33°10'-33°20'E). The three lakes are relatively small (areas of 10 km^2 or less) and shallow (average depths of 5 m or less), see Table 1. The water residence time (time for exchange of the lake water volume) is 3 years for lakes Uros and Vendyurskoe, and in Lake Rindozero ~5 times smaller than in the other two lakes. The influence of river throughflow on the lakes heat budget during the ice covered period is negligible, see paper 1.

From an ecological point of view Lake Uros is different from the other two lakes, being an oligotrophic lake, while Lakes Rindozero and Vendyurskoe are mesotrophic lakes. Consequently, the water transparency is higher, and it is possible to see the bottom at almost any point in Lake Uros, maximum depth 9.5 m. Corresponding typical Secchi disc readings in Lakes Vendyurskoe and Rindozero are 3-4 m and 1.5-2 m, respectively. The bottom sediments in the three lakes consist of sand up to 2 m depth, and silt containing organic material in the deeper parts (Litinskaya and Polyakov 1975).

Measurements were made from the ice along six cross-sections in Lake Vendyurskoe (in total 49 measurement stations), at two stations in Lake Uros, and at three sta-

Table 1 – Main characteristics of the three investigated lakes (Litinskaya and Polyakov, 1975)

	Uros	Rindozero	Vendyurskoe
Area (km^2)	4.3	1.8	10.4
Max depth (m)	9.5	9.5	13.4
Mean depth (m)	2.9	3.9	5.3
Volume (10^6 m^3)	12.2	7.9	54.8
Drainage basin area (km^2)	9.9	43.6	82.8
Water residence time (years)	3	0.6	3

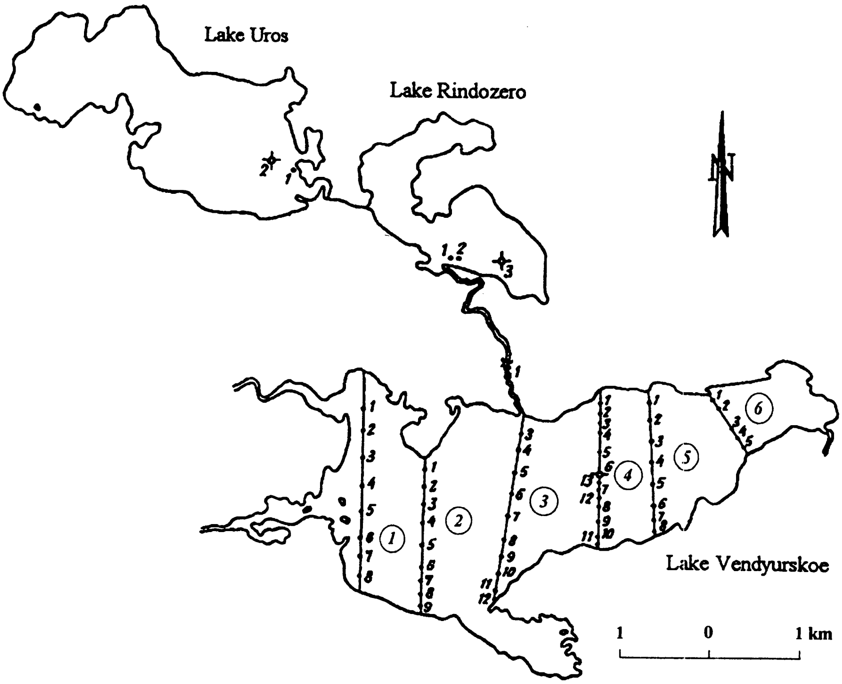


Fig. 1. Location of measurement stations in the three investigated lakes. Stations marked with a cross represent position of thermistor chains.

tions in Lake Rindozero, see Fig. 1. In total four surveys were made during the winter 1994/1995: in December, February, March, and April. The measurements during each survey included (characteristics of the measurement devices are given in paper 1),

1) full vertical profiles of temperature and conductivity registered from a hole in the ice at all stations. The devices used were a TCD (Temperature-Conductivity-Depth) profiler, having a vertical depth resolution of 5 cm, and a digital TC meter (used during the second survey in Lake Vendyurskoe), where readings were taken each half a metre;
2) profiles of temperature and conductivity in the vicinity of the ice at one cross-section (section 4) in Lake Vendyurskoe and at the stations in Lake Uros and Lake Rindozero. The registrations were made with the TCD profiler, mounted on a special construction (described by Bengtsson *et al.* 1995), which allowed readings to be taken at 8 mm intervals for a 1 m layer below the ice and at a distance of 1 m away from the hole in the ice;
3) registrations of temperature gradients in the upper 10 cm of the sediments at the stations in Lakes Uros and Rindozero, and at two cross-sections (section 4 – all

surveys; section 5 – March and April surveys) in Lake Vendyurskoe. The specially designed measurement probe, see Bengtsson *et al.* 1995 for a thorough description, includes two thermistors, spaced 10 cm apart, and a circular disc for detection of the sediment-water interface;

4) profiles of temperature and conductivity at the sediment-water interface (April survey) at the stations along three cross-sections (sections 4, 5, and 6) in Lake Vendyurskoe. The measurement probe included an optical sensor for detection of the sediment-water interface and was lowered using a micro-winch, allowing for a spatial resolution of 8 mm between registrations;
5) determinations of ice and snow thickness at all stations;
6) continuous registrations of incident and reflected solar radiation and of the amount of solar radiation that penetrated through the ice in the middle of Lake Vendyurskoe (April survey, station 4-3).

The water and sediments of the lakes were characterized by measuring the extinction coefficient for water and determining sediment type, porosity, and organic content. Furthermore, thermistor chains were installed in the deep parts of Lake Vendyurskoe (station 4-6, bottom depth 11.3 m), Lake Rindozero (station 3, bottom depth 9.0 m), and Lake Uros (station 2, bottom depth 5.5 m). The dates and locations for each kind of measurements and type of equipment used have been summarized in paper 1.

To obtain values on salt content from the conductivity readings, the relation

$$S = 0.821\,\kappa_{18} \tag{2}$$

was used, where c (mg l^{-1}) is salt content, and κ_{18} (μS cm^{-1}) is the measured water conductivity related to 18°C. This relation was determined by the Northern Water Problem Institute in Petrozavodsk, Russia, (unpublished report) for the considered lakes. The density as a function of water temperature and salt content was then calculated using the formula given by Chen and Millero (1986).

Measurement Results

Conductive Heat Flux From Water to Ice

To estimate the heat flux from water to ice during the surveys, measurements of the temperature structure beneath the ice cover were made at the stations along cross-section 4 in Lake Vendyurskoe, and at the stations in Lakes Rindozero and Uros. One typical example is shown in Fig. 2. Here, it can be seen that the temperature gradient does not change much in the upper metre, and is almost constant in the upper 10-20 cm during each of the three first surveys in all three lakes. During the April survey, solar radiation that penetrated through the ice had warmed the water

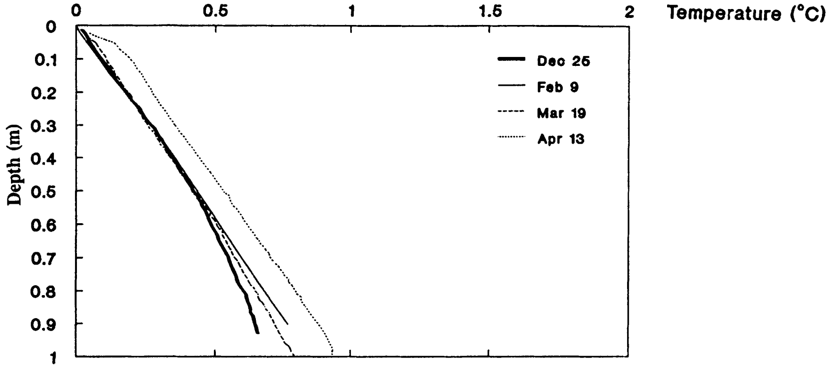


Fig. 2. Vertical temperature structure in the vicinity of the ice recorded during four surveys in Lake Vendyurskoe at station 4-2.

close to the ice compared to the previous three surveys, leading to an increased temperature gradient very close to the ice.

The heat transfer from water to ice in a lake without significant water motions and mixing is mainly diffusive. In the vicinity of the ice there is a laminar boundary layer, where the temperature gradient is constant and the molecular properties of water determine the heat transfer (Malmberg and Nilsson 1985). An approximate estimate of the thickness of this layer, δ_v, assuming a fully developed flow regime in the vicinity of the ice, is given by, *e.g.* Vennard and Street (1982) as

$$\delta_v = \frac{3.5\,u}{u_*} \tag{3}$$

where ν is molecular kinematic viscosity ($\sim 1.8\times 10^{-6}$ m s^{-1} at 0°C), and u_* is friction velocity, which using a drag coefficient of 2×10^{-3} for the ice cover (Larsen 1973) and a typical velocity of 1×10^{-3} m s^{-1} for Lake Vendyurskoe in winter, see Bengtsson *et al.* (1996), is $\sim 4.5\times 10^{-5}$ m s^{-1}. Then, the thickness of the viscous sublayer is about 15 cm which is of the same order of magnitude as the thickness of the observed layer in Fig. 2 with constant temperature gradient below the ice, being ~ 0.5 m. This laminar boundary layer thickness is also in good agreement with the estimates of Ellis *et al.* (1991) for Ryan Lake, USA, that gave a vertical temperature diffusivity approximately equal to the molecular value down to a depth of 10-15 cm below the ice cover.

If the temperature gradient at the ice water interface and the conductivity, λ, are known, the conductive heat flux from water to ice, Q_{wi}, may be estimated from the gradient method

$$Q_{wi} = -\lambda\frac{\delta T}{\delta z} \tag{4}$$

where T is temperature, and z is vertical coordinate. The average temperature gradient was defined within a 5 cm layer in the vicinity of the ice to assure that laminar

Table 2 – Temperature gradient ($\partial T/\partial z$) in the vicinity of the ice and heat flux from water to ice (Q_{wi}) for the stations along cross section 4 in Lake Vendyurskoe during the four surveys in the winter 1994/1995.

Station	Survey 1 (23-25 Dec, 1994)		Survey 2 (7-9 Feb, 1995)		Survey 3 (19 Mar, 1995)		Survey 4 (13 Apr, 1995)	
	$\partial T/\_z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})
4-1	1.03	0.6	1.00	0.6	1.48	0.8	1.61	0.9
4-2	0.72	0.4	0.88	0.5	1.27	0.7	2.41	1.4
4-3	0.75	0.4	1.00	0.6	1.26	0.7	2.07	1.2
4-4	1.16	0.7	1.40	0.8	1.18	0.7	1.66	0.9
4-5	-	-	0.94	0.5	0.76	0.4	1.26	0.7
4-6	1.49	0.8	1.31	0.7	1.20	0.7	2.41	1.4
4-7	-	-	1.33	0.7	1.45	0.8	1.91	1.1
4-8	2.59	1.5	1.64	0.9	1.33	0.7	1.45	0.8
4-9	1.44	0.8	1.31	0.7	0.82	0.5	1.47	0.8
4-10	1.16	0.7	0.90	0.5	0.79	0.4	1.05	0.6
Average (1-10)		0.7		0.7		0.6		1.0
Standard deviation (1-10)		0.3		0.1		0.2		0.3

Table 3 – Temperature gradient ($\partial T/\partial z$) in the vicinity of the ice and heat flux from water to ice (Q_{wi}) for the stations in Lake Uros during the four surveys in the winter 1994/1995.

Lake	Station	Survey 1 (24-25 Dec, 1994)		Survey 2 (10 Feb, 1995)		Survey 3 (17 Mar, 1995)		Survey 4 (19 Apr, 1995)	
		$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_{wi} (W m^{-2})
Uros	1	1.19	0.7	1.10	0.6	1.43	0.8	5.39	3.1
	2	1.50	0.9	0.98	0.6	1.38	0.8	2.67	1.5
	Average (1-2)		0.8		0.6		0.8		2.3
Rindo-zero	1	1.38	0.8	0.69	0.4	1.82	1.0	8.47	4.8
	2	1.25	0.7	1.35	0.8	1.45	0.8	13.02	7.4
	3	3.55	2.0	1.24	0.7	2.25	1.3	5.49	3.1
	Average (1-3)		1.2		0.6		1.0		5.1

conditions prevailed and that the gradient method could be used with the molecular value of conductivity for water at 0°C, λ =0.569 W m^{-1} $°C^{-1}$, when estimating the heat flux to the ice.

The calculated heat fluxes from water to ice at the investigated stations in the three lakes during the four surveys are given in Tables 2-3. The average heat flux from water to ice is rather consistent for the three lakes during the December, February, and March surveys, being in the range 0.6-0.8 W m^{-2}, except for the December and March surveys in Lake Rindozero, with estimated heat fluxes of 1.2 W m^{-2} and 1.0 W m^{-2}, respectively. The heat flux values do not show any dependence on station location or time during the winter. The spatial variation of heat flux is relatively small for the three first surveys, although a few estimates differ considerably from the averages. It seems reasonable to consider the heat flux from water to ice to be location independent in a lake during the main part of the winter. During the April survey, the heat fluxes from water to ice are higher or considerably higher than during the rest of the surveys. In Lake Vendyurskoe, the average heat flux was estimated to 1.0 W m^{-2}, which can be compared to 0.6 W m^{-2} during the March survey. The April survey in this lake (on April 13) was made when spring heating became apparent, causing the snow to melt and allowing penetration of solar radiation to the water. This heat input led to increased temperatures in the vicinity of the ice, see Fig. 2, and thus higher heat transfer from water to ice. The fourth survey in Lake Rindozero and Lake Uros was made six days later (on April 19), when a significant temperature increase in the vicinity of the ice due to penetrating solar radiation can be seen in the temperature profiles, see paper 1. This is the reason for the considerably higher values for the heat flux from water to ice, being in average 5.1 W m^{-2} in Lake Rindozero and 2.3 W m^{-2} in Lake Uros.

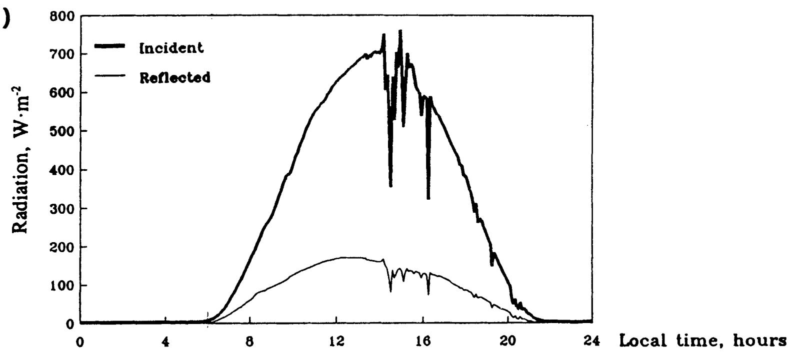
Solar Radiation Heat Flux

The three investigated lakes were covered with snow during the main part of the winter. The snow layer, varying in thickness in the range 0.1-0.4 m, and the ice, increasing to 65 cm in the course of the winter, isolated the water body effectively from solar radiation. This is also confirmed by the recorded temperature profiles during the three first surveys in the three lakes, which did not show any influence of a heat input due to penetrated solar radiation. No measurements of penetrating solar radiation were made during the December to March surveys. However, some measurements were made during the winter 1995/1996 to get an idea of the magnitude of the penetrating solar radiation flux to the water. During the periods November 26-30 and December 27-28, the registered daily average radiation flux penetrating through the ice and snow cover was 0.3 W m^{-2} and not recorded, respectively. Corresponding ice/snow thicknesses were 17/14 cm and 38/15 cm. Consequently, as was also shown by a hypothetical example in the review section, the solar radiation heat flux in early and mid-winter can be expected to be one order of magnitude smaller than the sediment heat flux and heat flux from water to ice.

Table 4 – Daily averages of measured incident and reflected solar radiation (R_d, R_r), radiation at the ice-water interface (R_{wi}), ice and snow thickness (h_i, h_s), calculated extinction coefficient for ice and surface albedo value (η_i, A), during the period April 13-23 at station 4-3 (measurements started at 12:15 on April 13, and ended at 18:40 on April 23, why the presented data from these days are not really daily averages). The superscript * stands for values estimated from measurements at other stations.

Date	R_d (W m^{-2})	R_r (W m^{-2})	R_{wi} (W m^{-2})	h_i (cm)	h_s (cm)	η_i (m^{-1})	A
April 13	299.8	197.4	1.6	68	10	-	0.66
April 14	151.2	90.7	1.9	65*	8*	-	0.60
April 15	100.9	49.8	3.0	63*	6*	-	0.49
April 16	194.1	64.7	9.4	61*	3*	-	0.33
April 17	91.0	26.9	6.6	59*	0*	-	0.30
April 18	53.9	13.5	4.7	57*	0*	3.8	0.25
April 19	125.0	34.2	9.6	54	0	4.2	0.27
April 20	151.4	48.4	11.0	55	0	4.1	0.32
April 21	204.9	52.0	17.5	50	0	4.3	0.25
April 22	245.7	61.8	24.1	46	0	4.4	0.25
April 23	222.4	50.1	28.6	45	0	4.0	0.23

a)



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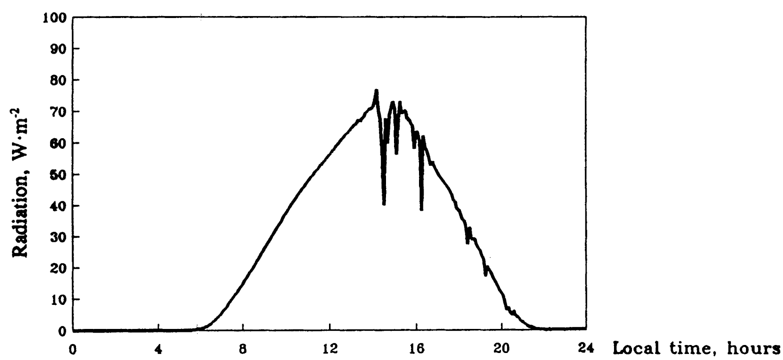

Fig. 3. Measured incident and reflected solar radiation (**a**), and radiation beneath the ice (**b**) during April 22 at station 4-3. The ice cover thickness was ~46 cm.

During the April survey, solar radiation (incident, reflected, and the amount penetrating through the ice) was registered with 5-minute intervals during the period April 13-23 at station 4-3 in Lake Vendyurskoe. Most of the ice cover was snow ice, formed from the top of the ice. The daily averages of incident and reflected solar radiation at the snow/ice surface (from April 17, the snow cover was no longer continuous over the lake area, but remained as bright and dark (snow saturated with water) spots), the amount of radiation penetrating through the ice, calculated values on albedo and extinction coefficients are given in Table 4. An example of variation of the radiation components during the day is given in Fig. 3. The daily average of the solar radiation penetrating into the water is about 2 W m^{-2} on April 14, which is the same order of magnitude as the heat fluxes from sediments to water and from water to ice. The penetrating radiative flux increases during the period to ~30 W m^{-2} on April 23. The increase is mainly due to melting of the snow cover which disappeared on April 17 at the location of measurements, and a continuous decrease in ice thickness (from 65 to 45 cm during the period). Also, the snow surface albedo decreased caused mainly by the disappearance of the snow cover during the period, from ~0.65 which is typical for old snow prior to melting (Grey and Male 1981) to ~0.25, which is typical for ice (Patterson and Hamblin 1988). The albedo variation during daytime was characterized by a continuous decrease, and at the end of the day a slight increase was observed when the solar angle was small.

The extinction coefficient for the ice which was mainly snow ice was estimated to ~4 m^{-1} for the snow free period. The extinction coefficient for ice is strongly dependent on its character (and also radiation wavelength), with reported comparative values ranging 0.2-10 m^{-1} (Ashton 1986).

Sediment Heat Flux

Heat is transferred from water to lake sediments during the summer and is released back into the water during the winter period (Hutchinson 1957). This sediment heat flux is normally the main heat source in early and mid-winter in an ice-covered lake. No direct measurements of the heat flux from bottom sediments to water were made in the three Karelian lakes. Instead, the temperature gradients in the upper 10 cm of the sediments were measured for the stations along cross-section 4 (and also along cross-section 5 during the third and fourth survey) in Lake Vendyurskoe, and for the stations in Lake Rindozero and Lake Uros. The temperature gradients in the upper part of the sediments were approximately constant, see Fig. 4.

Provided that the sediment conductivity is known, the heat flux from water to sediments can be estimated using the gradient method (see Eq. (4) above). The conductivity of the sediments, λ, was estimated from values of porosity and quartz content using a method proposed by Johansen (1975)

$$\lambda = \lambda_w \rho \lambda_s^{(1-p)} \qquad (5)$$
$$\lambda_s = 7.7^q \times 2.0^{(1-q)}$$

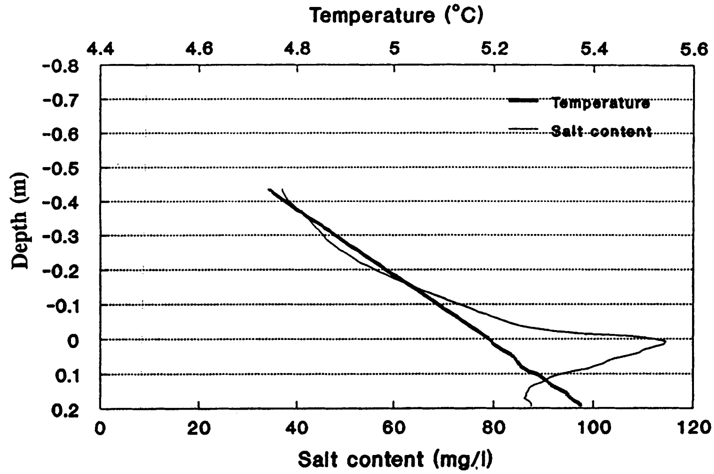

Fig. 4. Temperature and salt content profiles at the sediment-water interface at station 4-13 on April 16. Positive depth values in sediments, negative in water.

where p is porosity, q is quartz content (quartz has a considerably higher conductivity than other common minerals, Sundberg (1986)), and the subscripts w and s denote water and sediment grains, respectively. This approach was tested by Sundberg (1986) for various types of soils and was found to give good agreement with measured values of conductivity.

The type of sediment, porosity and content of organic material of the upper 10 cm sediment layer were determined at the stations along cross-section 4 in Lake Vendyurskoe and at the stations in Lakes Rindozero and Uros. For cross-section 5 in Lake Vendyurskoe, the sediments were characterized using a map over the sediment type (unpublished report, Karelian Scientific Center, Petrozavodsk, Russia). At all stations (except station 1 in Lake Uros, and stations 1 and 2 along cross-section 5 in Lake Vendyurskoe) in the three lakes, the solid material type in the upper 10 cm of the sediments was silt. According to Sundberg (1986), the quartz content in silt ranges from 30 to 70%, with an average value of 50%, which will be adopted in the calculations of the sediment conductivity (ranges will also be given). The organic material was assumed to have the same characteristics as the mineral grains, except that the quartz content was set to zero.

The porosity, calcination losses, and calculated sediment conductivities for the stations are given in Table 5. According to Lozovik (1991), 70% of the calcination losses are related to the content of organic material and 30% to the decomposition of mineral matter. Therefore, it is assumed that the organic content corresponds to 70% of the calcination losses.

All samples from the upper 10 cm of the sediments had very high water contents. The corresponding porosities for most samples were higher than 95%. The average calcination losses were about 30%, corresponding to an average content of organic

Table 5 – Porosity (P), calcination losses, and calculated conductivities (mean, λ, and range due to variations in quartz content for silt, λ_{range}) for the upper 10 cm sediment layer at the stations along cross-section 4 in Lake Vendyurskoe and the stations in Lake Rindozero and Lake Uros.

Lake	Station	P (%)	Calcination losses (weight-% of dry substance)	λ (W m^{-1} °C^{-1})	λ_{range} (W m^{-1} °C^{-1})
Vendyurskoe	4-1	91.1	12.6	0.67	0.66-0.68
	4-2	96.4	27.2	0.61	0.60-0.61
	4-3	97.1	31.4	0.60	0.59-0.60
	4-4	97.5	32.9	0.59	0.59-0.60
	4-5	96.4	28.2	0.61	0.60-0.61
	4-6	97.0	32.1	0.60	0.60-0.60
	4-7	96.9	32.7	0.60	0.60-0.60
	4-8	97.0	30.3	0.60	0.60-0.60
	4-9	94.5	30.9	0.63	0.62-0.63
	4-10	97.6	32.2	0.59	0.59-0.60
Uros	10-1	-	-	1.75[1)]	1.40-2.20[1)]
	10-2	97.8	48.2	0.59	0.59-0.59
Rindozero	20-1	97.4	32.6	0.59	0.59-0.60
	20-2	97.9	30.0	0.59	0.59-0.59
	20-3	97.3	31.4	0.59	0.59-0.60

1) As the solid phase in the sediments consisted of sand, no sediment sample could be taken. Instead, the conductivity and its range for sand sediments was evaluated from Sundberg (1986), who determined the conductivity for a large set of sand samples.

material in the solid phase of approximately 20%. This leads to sediment conductivities that are slightly larger to that of water, on average about 5% higher, *i.e.* ~0.60 W m^{-1} °C^{-1}. This value of conductivity will also be assumed to be valid for the silt sediments at stations 3 to 8 along cross-section 5 in Lake Vendyurskoe. The highest sediment conductivity compared to water, about 15% higher, was obtained close to the shore on cross-section 4 in Lake Vendyurskoe. The variations in conductivity due to possible variations in quartz content for the silt sediment grains are negligible. For station 1 in Lake Uros and stations 1 and 2 along cross-section 5 in Lake Vendyurskoe, the bottom sediment consisted of sand, which should have a considerably smaller water content and also higher quartz content than silt. Therefore, the sand sediments should have a higher conductivity than silt sediments. As no samples were taken at these stations, the conductivities were evaluated from conductivities determined by Sundberg (1986) for a large set of sand samples. The average value of 1.75 W m^{-1} °C^{-1} is approximately 3 times higher than that of water, which indicates that heat conduction is far more efficient in sandy sediments than in the upper layer of silt sediments.

Table 6 – Temperature gradients in the upper layer of sediments (0-10 cm), and estimated heat fluxes from sediments to water, Q_b, for the stations along cross-section 4 and 5 in Lake Vendyurskoe during the four surveys.

Station	Depth (m)	Survey 1 (23-25 Dec, 1994)		Survey 2 (7-9 Feb, 1995)		Survey 3 (17-19 Mar, 1995)		Survey 4 (13-14 Apr, 1995)	
		$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})
4-1	3.60	6.32	4.6	4.59	3.1	3.42	2.3	3.37	2.3
4-2	5.70	4.69	2.9	3.47	2.1	3.19	1.9	2.75	1.7
4-3	7.85	3.88	2.3	2.61	1.6	2.13	1.3	1.98	1.2
4-4	7.70	5.00	3.0	2.66	1.6	2.10	1.2	2.12	1.2
4-5	8.60	-	-	3.06	1.9	1.98	1.2	2.00	1.2
4-6	11.35	2.73	1.6	2.24	1.3	1.28	0.8	1.13	0.7
4-7	9.40	-	-	3.16	1.9	1.94	1.2	2.16	1.3
4-8	6.30	6.02	3.6	4.23	2.5	2.88	1.7	2.64	1.6
4-9	6.25	5.81	3.7	3.67	2.3	2.63	1.7	2.70	1.7
4-10	5.50	5.30	3.1	3.93	2.3	2.82	1.7	3.08	1.8
5-1	2.80	-	-	-	-	2.75	4.8	1.64	2.9
5-2	3.80	-	-	-	-	1.75	3.1	1.53	2.7
5-3	5.90	-	-	-	-	2.49	1.5	2.25	1.4
5-4	8.45	-	-	-	-	2.49	1.5	2.25	1.4
5-5	10.40	-	-	-	-	1.32	0.8	1.34	1.2
5-6	8.60	-	-	-	-	2.09	1.3	1.94	1.2
5-7	7.00	-	-	-	-	2.43	1.5	2.37	1.4
5-8	4.70	-	-	-	-	3.06	1.8	2.57	1.5

Table 7 – Temperature gradients in the upper layer of sediments (0-10 cm), and estimates of the heat flux from the sediments to water, Q_b, for the stations in Lake Uros and Lake Rindozero during the four surveys.

Lake	Station	Depth (m)	Survey 1 (24-25 Dec, 1994)		Survey 2 (10 Feb, 1995)		Survey 3 (17 Mar, 1995)		Survey 4 (19 Apr, 1995)	
			$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})	$\partial T/\partial z$ (°C m^{-1})	Q_b (W m^{-2})
Uros	1	2.75	2.55	4.5	1.62	2.8	1.20	2.1	1.74	3.0
	2	4.95	3.11	1.8	2.10	1.2	1.80	1.1	1.09	0.6
Rindozero	1	3.40	3.06	1.8	1.92	1.1	1.56	0.9	1.41	0.8
	2	6.30	3.72	2.2	2.70	1.6	2.00	1.2	1.79	1.1
	3	8.85	1.92	1.1	1.56	0.9	1.08	0.6	1.07	0.6

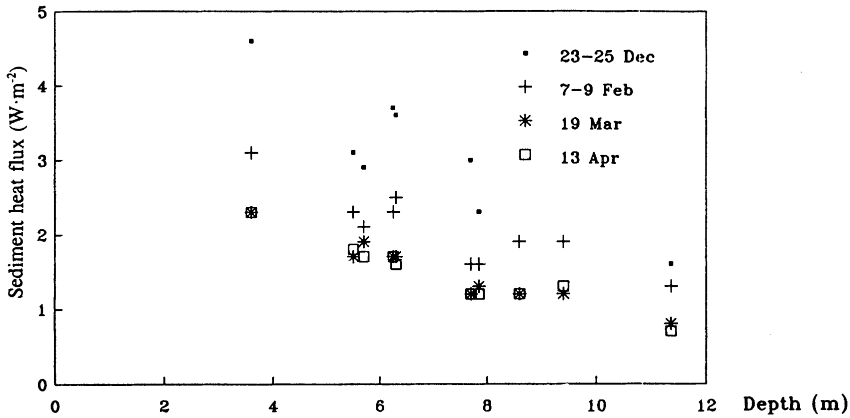

Fig. 5. Sediment heat flux registered at cross-section 4 in Lake Vendyurskoe during the four surveys versus bottom depth.

In Tables 6-7, the measured temperature gradients in the upper 10 cm layer of the sediments and the calculated sediment heat fluxes are given for the stations along cross-section 4 and 5 in Lake Vendyurskoe and for the stations in Lake Uros and Lake Rindozero during all four surveys (measurements of temperature gradients along cross-section 5 were, however, just made during the third and fourth survey).

The sediment heat flux magnitudes for all stations between December and April are within a range of 0.5-4.8 W m^{-2}. This is quite consistent with reported values from other investigators. The sediment heat flux decreases with time during the ice covered period, as illustrated in Fig. 5. This decrease with time is due to the continuous decrease of available heat in the sediments for warming of the lake waters. The sediment heat flux also varies with bottom depth, see Fig. 5, with higher values in shallow regions than in deeper parts. The smaller heat flux values in the deeper parts may presumably be explained by the smaller annual temperature variations at the sediment-water interface, which is the cause of the heat transfer, compared to shallow regions.

Sediment Salt Flux

The exchange of soluble mineral matter through the water-sediment interface is caused mainly by mineralization of organic matter in the upper layer of the sediments and diffusion of different ions from deeper layer of the sediments. The magnitude of these fluxes is constrained by processes like sedimentation, destruction of the organic matter, redox potential, and diffusion. Here, however, no attempt was made to quantify these processes. Instead, only the total transport of solutes, S_b, was estimated using the gradient method

$$S_{bi} = - D_i \frac{dc_i}{dz} \tag{6}$$

Table 8 – Bottom salt concentration gradients (dc/dz) and salt fluxes (S_b) along cross-section 4, 5, and 6 in Lake Vendyurskoe during April 14-16.

Station	Depth (m)	dc/dz (kg m^{-4})	S_b (10^{-10} kg m^{-2} s^{-1})	Type of sediments
4-1	3.50	0.84	4.2	silt
4-2	5.65	1.03	5.2	silt
4-3	7.70	0.26	1.3	silt
4-4	7.65	1.94	9.7	silt
4-5	8.50	1.42	7.1	silt
4-6	11.30	-0.01	-0.0	silt
4-13	11.15	0.70	3.5	silt
4-7	9.35	1.19	6.0	silt
4-12	7.90	1.28	6.4	silt
4-8	6.25	0.27	1.4	silt
4-9	6.25	0.90	4.5	silt
4-10	5.45	1.06	5.3	silt
4-11	4.05	0.59	3.0	silt with sand
5-1	2.90	0.70	3.5	sand
5-2	3.75	0.32	1.6	sand
5-3	5.80	0.53	2.7	silt
5-4	8.35	1.58	7.9	silt
5-5	10.35	1.38	6.9	silt
5-6	8.55	1.44	7.2	silt
5-7	6.85	1.67	8.4	silt
5-8	4.50	0.44	2.2	silt
6-1	2.35	0.13	0.7	sand
6-2	4.10	1.48	7.4	silt
6-3	5.00	1.22	6.1	silt
6-4	4.50	0.39	1.9	silt with sand
6-5	4.10	1.65	8.2	silt with sand
Average		0.94	4.7	
Standard deviation		0.51	2.7	

where D_i is the diffusion coefficient for the i^{th} considered substance, and dc_i/dz is the vertical concentration gradient.

Measurements of conductivity were made at the sediment-water interface (space resolution 8 mm) along cross-sections 4, 5 and 6 in Lake Vendyurskoe during the April survey to get information about the salt content distribution in the bottom layers, see Malm *et al.* (1996), and to estimate the bottom flux of salts. As the processes at the sediment-water interface are expected to be the same during early and late winter, the data from the April survey is assumed to be representative for the whole ice-covered period. The concentration gradients were determined for a 1.5 cm layer positioned just above the sediment-water interface. Molecular diffusion coefficients for different soluble ions are dependent upon their concentrations and the temperature of medium, as well as, if it is below the sediment-water interface, its porosity (Manheim 1970); they are to some extent also influenced by the content of organic matter. In this study, the diffusion coefficient was taken as 5×10^{-10} m^2 s^{-1}, which is representative for the main ions (calcium, magnesium, *etc.*) at a temperature of 5°C in water (Li and Gregory 1974).

Table 8 presents calculated values of diffusive fluxes of salts from sediments to water for cross-sections 4, 5, and 6. The obtained salt fluxes are of the order $1\text{-}10 \times 10^{-10}$ kg m^{-2} s^{-1}, with an average value of 4.7×10^{-10} kg m^{-2} s^{-1}. This average salt flux would correspond to an average salt content increase of ~1 mg l^{-1} during a month in a one metre layer above the bottom, which is quite small compared to the mean salt concentration in the lake water, being ~20 mg l^{-1}. The salt release from the bottom sediments during a 6-month period is comparable to the salt release by formation of a 40 cm ice layer. Opposite the case for the heat fluxes, no clear dependence of salt flux magnitude on bottom depth was found. However, in general the salt flux values are lower when, as is typical for the shallow regions, the upper silt layer, lying above sand sediments, is small or absent and has a low content of organic matter. The salt flux values are also lower at stations located at isolated deep subbasins. For stations 4-6 and 4-13 this may to some extent be linked with changed reduction-oxidation conditions. During sampling, it was found that the upper layer of sediments was of dark greyish-green color and, as distinct from the other stations, the surface oxidized layer of brown color (about 0.5-1.0 cm of thickness) was absent.

Heat Content Change

The temperature structure in the three investigated ice-covered lakes during winter was characterized by horizontal isotherms, see Malm *et al.* (1996). The continuous temperature increase that occurred during the winter corresponds therefore to a larger heat content increase in deep verticals than in shallower ones, in spite of the larger sediment heat fluxes in shallow regions. It is therefore clear that a horizontal heat transport from shallow regions to deeper parts must occur during the winter. This horizontal heat transport can be estimated by calculating the changes of heat content in water columns between two surveys. The change of heat content in a water column of unit area is, provided solar radiation can be neglected, caused by heat conduction from sediments to water, Q_b, from water to ice, Q_{wi}, and horizontal heat transport

$$\frac{\Delta H}{\Delta t} = Q_b - Q_{wi} + \text{horizontal net heat transport} \tag{7}$$

Here

$$H = \int_0^D \rho_w c_w T(z,t)\, dz \tag{8}$$

Table 9 – Sediment heat flux (Q_b), heat flux from water to ice (Q_{wi}), the rate of heat content change for the water column ($\Delta H/\Delta t$) between the first and third survey, and net horizontal heat transport *to* a water column (NHHT) for the stations along cross-section 4 in Lake Vendyurskoe and the stations in Lakes Uros and Rindozero. The sediment heat flux, and heat flux from water to ice are averages for the first three surveys.

Lake	Station	Depth (m)	Q_b (W m^{-2})	Q_{wi} (W m^{-2})	$\Delta H/\Delta t$ (W m^{-2})	NHHT (W m^{-2})
Vendyurskoe	4-1	3.50	3.3	0.7	0.0	-2.6
	4-2	5.65	2.3	0.5	0.6	-1.2
	4-3	7.70	1.7	0.6	1.3	+0.2
	4-4	7.65	1.9	0.7	1.5	+0.3
	4-5	8.50	1.9	0.5	2.2	+0.8
	4-6	11.30	1.2	0.7	3.9	+3.4
	4-7	9.35	1.9	0.7	2.9	+1.7
	4-8	6.25	2.6	1.0	1.2	-0.4
	4-9	6.25	2.6	0.7	1.3	-0.6
	4-10	5.45	2.4	0.5	0.8	-1.1
	Average		2.2	0.7	1.6	+0.1
Uros	1	2.55	3.1	0.7	0.5	-1.9
	2	4.95	1.4	0.8	2.1	+1.5
Rindozero	1	3.10	1.3	0.7	0.4	-0.2
	2	6.25	1.7	0.8	1.8	+0.9
	3	8.85	0.9	1.3	3.1	+3.5

is the heat content of a water column with unit area and a depth equal to D, and Δt is the time interval between the surveys.

The rate of heat content change is expressed in terms of a heat flux. If there is no horizontal heat transport (indicating that only vertical heat transport processes are of importance), the rate of heat content change in a water column should be equal to the heat flux from sediments, Q_b, minus the heat flux from water to ice, Q_{wi}. By comparing the rate of heat content change computed from observed temperature distributions with the sediment heat flux and the heat flux from water to ice, the net horizontal heat transport within the lake can be estimated.

The rate of heat content change in a water column was calculated between the December and March survey for all stations in the three lakes, to cover most of the winter period. The April survey was excluded as a significant amount of solar radiation had started to penetrate through the ice during this survey. The results from the stations along cross-section 4 in Lake Vendyurskoe, and the stations in Lake Rindozero and Lake Uros are given in Table 9.

The horizontal heat flux values in all three lakes are of the order of 0-3.5 W m^{-2}, *i.e.* the same magnitude as the sediment heat flux and the heat flux from water to ice. Therefore, any attempt to predict the temperature distribution and its development in a shallow ice-covered lake has to include this transport. The results for cross-section 4 in Lake Vendyurskoe, show that heat is transported from the shallow regions to the deep parts. This type of transport is also indicated in the two other lakes, although the number of stations there are too few for any certain conclusions. One possible mechanism causing this horizontal heat transfer is the transport of warmer heavier water along the bottom from shallow regions to deeper parts of a lake, as suggested by several investigators (Mortimer and Mackereth 1958; Likens and Ragotzkie 1965; 1966; Welch and Bergmann 1985; Rahm 1985).

Since the length-width ratio for Lake Vendyurskoe is about four, it seems reasonable to assume that most of the horizontal heat transport occurs along the investigated cross-sections. This means that there is very little lateral heat transfer across the section and that the average rate of heat content change in cross-section 4 should be almost equal to the sum of the vertical boundary heat fluxes for the section. This is also the case, as seen in Table 9 where the net horizontal heat transport of section 4 is almost zero, confirming the validity of the estimated heat fluxes from sediment to water and from water to ice.

In Table 10, the rate of heat content change is also given for the verticals in the deep part of the three lakes where thermistor chain recordings were made. The calculations are made on a monthly basis during the winter, excluding the solar heating period in late winter. The temperature increases quite rapidly in the deep parts in early winter (November-December) in both Lake Vendyurskoe and in Lake Rindozero (see paper 1), and the results in Table 10 show that the rate of heat content change is several times higher than in mid-winter (January-March). It is therefore reasonable to assume that the horizontal heat transport toward the deeper parts is highest (and presumably also the sediment heat flux) during this period.

Table 10 — Rate of heat content change during the winter for the three verticals (one in each lake) where thermistor chain recordings were made.

	Rate of heat content change (W m^{-2})		
	Vendyurskoe	Rindozero	Uros
November (8-30)	13.5	13.1	-
December	7.2	6.2	-
January	5.3	3.5	2.7
February	3.4	2.3	1.9
March	3.4	2.3	1.1

Salt Content Change

Changes in salt content in the waters of Lake Vendyurskoe during the ice-covered period are due to, 1) release of salts into water during ice formation; 2) diffusion of salts from sediments; 3) destruction of organic matter within a water column (negligible, however, as the ice covered period is associated with very few biological activity); 4) salt gain/loss with in/outflows. As the inflows and outflows to the lake were very small, their influence on the salt content changes in the lake are negligible, except at locations near river inlets/outlets.

Analogously to the heat content change calculations, it is assumed that the salt content change can be evaluated as

$$\frac{\Delta C}{\Delta t} = S_b + S_i + S_h \qquad (9)$$

where

$$C = \int_0^D c(z,t)\,dz \qquad (10)$$

is the content of salts within a water column of depth D; Δt is time between surveys; S_b is salt flux from/to sediments; S_i is salt release due to ice crystallization; and S_a is the horizontal net salt transport.

The rate of salt content change in a water column was calculated between the December and March survey at cross-section 4. Measurements of ice thickness showed that new ice was mainly formed as white ice on top of the ice layer between December and March. However, this type of ice formation does not lead to a salt release into lake waters. The thickness of black ice remained almost constant, increasing at most with 1-2 cm. This corresponds to a maximum salt release of $2\text{-}5 \times 10^{-11}$ kg m^{-2} s^{-1} due to the ice formation between December and March which is one order of magnitude smaller than the salt flux from sediments. Therefore, the salt release from ice to water was assumed to be negligible and S_i was consequently set to zero. Further, the salt flux from sediments to water was assumed time constant and thus the values obtained during the fourth survey were assumed to be representative for the considered period. The results from the calculations are given in Table 11. The calculated net salt transport, being of the order $0\text{-}20 \times 10^{-10}$ kg m^{-2} s^{-1}, shows that a salt transport occurs from the shallow regions to local hollows (station 4-3) and the deep parts of the section (station 4-6). As the main salt content changes occurred in the vicinity of the bottom (Malm *et al.* 1996), it seems probable that the salt transport was localized to the bottom layer. Bearing in mind the accuracy of the conductivity measurements, there is a quite good agreement between the total salt input from sediments along section 4 of Lake Vendyurskoe and the salt concentration change, which indicates that the adopted assumptions included that of no freeze-out of salt during the ice production in the calculation method are reasonable. The average change in salt concentration between the first and third survey for cross-section 4 is

Table 11 – Rate of salt content change for a water column ($\Delta C/\Delta t$), sediment salt flux (S_b), and net salt transport to a water column (S_a) for the stations along cross-section 4 in Lake Vendyurskoe. All values are multiplied by a factor of 10^{10} and expressed in kg m^{-2} s^{-1}.

Station	Depth, m	$\Delta C/\Delta t$	S_b	S_a
4-1	3.50	1.5	4.2	-5.7
4-2	5.65	5.3	5.2	0.1
4-3	7.70	4.5	1.3	3.2
4-4	7.65	5.0	9.7	-4.7
4-6	11.30	21.4	-0.0	21.4
4-8	6.25	0.1	1.4	-1.3
4-9	6.25	0.9	4.5	-3.6
4-10	5.45	1.2	5.3	-4.1
Average	6.7	4.6	3.9	0.7

0.5 mg l^{-1}, which is an increase of only a few per cent compared to the average salt concentration for the section. Thus, the general salt content changes are relatively small compared to heat content changes.

Conclusions

During the main part of the winter, the heat flux at the ice-water interface in the three investigated lakes was dominated by conduction from water to ice. This heat flux was typically in the range 0.5-1 W m^{-2} for all lakes and did not show any significant variations, neither in time nor space. From middle of April until ice break-up, heating due to solar radiation penetrating through the ice dominated the development of the thermal regime. Typical daily averages in the second half of April were about 10-30 W m^{-2}.

The extinction coefficient for ice was determined from solar radiation measurements above and below the ice, and was found to be ~4 m^{-1}. Albedo values were determined from registrations of incident and reflected solar radiation, being typically 0.6 for snow and 0.2 for ice.

The heat flux from sediments to water was the main source for the lake water heating during early and mid-winter and had similar magnitudes in all three lakes, being 2-4.5 W m^{-2} for shallow stations and 1-1.5 W m^{-2} for deep stations in early winter, and 1-3 W m^{-2} for shallow stations and 0.5-1 W m^{-2} for deep stations in late winter. A heat transport from shallow regions to deep parts was shown to occur during the winter. The horizontal heat fluxes are of the same order as the vertical fluxes, and should thus be accounted for in any attempt to predict the temperature evolution in an ice covered lake.

The salt transfer from sediments was found to be of the order $1\text{-}10 \times 10^{-10}$ kg m^{-2} s^{-1}. A comparison of this flux with salt content changes in water columns during the winter indicates that a horizontal salt transport of the same order as the sediment salt flux occurs from shallow regions to the deeper parts of a lake during winter. This horizontal flux is the most likely cause for the observed continuous salt content increase in local hollows and in the deepest parts of the lake.

Acknowledgements

The present study is a part of a cooperative project between the Department of Water Resources Engineering, Lund University, and the Northern Water Problems Research Institute, Karelian Scientific Center, Russian Academy of Sciences, Petrozavodsk, Russia. The project is sponsored by the Swedish Natural Science Research Council, the Royal Swedish Academy of Sciences, and the Russian Academy of Sciences. Additional funding for measurement equipment have been provided by Åke och Greta Lisshed's Foundation. The support is gratefully acknowledged. We would also like to express our gratitude to Professor Nikolaj Filatov for his continuous support during the work. Furthermore, we would also like to thank Mr. Andrej Mitrokhov, hydrologist, for his help during the field measurements.

References

Ashton, G.D. (ed.) (1986) River and lake ice engineering, Water Res. Publ., 485 pp.

Bengtsson, L. (1996) Mixing in ice-covered lakes, *Hydrobiologia, Vol. 322*, pp. 91-97.

Bengtsson, L., Malm, J., Terzhevik, A., Petrov, M., Boyarinov, P., Glinsky, A., and Palshin, N. (1995) A Field Study of Thermo- and Hydrodynamics in a Small Karelian Lake during Late Winter 1994. Report 3185, Department of Water Resources Engineering, Lund Institute of Technology, Lund University, Sweden, 72 pp.

Bengtsson, L., Malm, J., Terzhevik, A., Petrov, M., Boyarinov, P., Glinsky, A., and Palshin, N. (1996) Field Investigation of Winter Thermo- and Hydrodynamics in a Small Karelian Lake, *Limnol. Oceanogr., Vol. 41*, pp. 1502-1513.

Bengtsson, L., and Svensson, T. (1996) Thermal regime of ice covered Swedish lakes, *Nordic Hydrol., Vol. 27*, pp. 39-56.

Birge, E.A., Juday, C., and March, H.W. (1927) The temperature of the bottom deposits in Lake Mendota; A chapter in the heat exchanges of the lake, *Trans. Wisconsin Acad. Sci., Vol. 23*, pp. 187-232.

Chen, C.T., and Millero, F.J. (1986) Precise thermodynamic properties for natural waters covering only limnological range, *Limnol. Oceanogr., Vol. 31*, pp. 657-662.

Ellis, C.R., Stefan, H.G., and Gu, R. (1991) Water temperature dynamics and heat transfer beneath the ice cover of a lake, *Limnol. Oceanogr.*, Vol. 36, pp. 324-334.

Golosov, S., and Kreiman, K. (1992) Heat exchange and thermal structure of water-sediment system, Vodnye resursy (Water resources), *No. 6*, pp. 12-18 (in Russian).

Grey, D.M., and Male D.H. (ed.) (1981) *Handbook of Snow. Principles, Processes, Management & Use*, Pergamon Press, 776 p.

Gu, R., and Stefan, H.G. (1990) Year-round temperature simulation of cold climate lakes, *Cold Regions Sci. Technol., No. 18*, pp. 147-160.

Hutchinson, G.E. (1957) *A treatise on limnology*, Vol. 1. Geography, physics, and chemistry, J. Wiley & Sons, Inc., 1015 p.

Johansen, Ö. (1975) Heat transfer properties in soils. Report, Norway Institute of Technology, Trondheim (in Norwegian).

Johnson, N.M., and Likens, G.E. (1967) Steady-state thermal gradient in the sediments of a meromictic lake, *J. Geophys. Res., Vol. 72, No. 12*, pp. 3049-3052.

Larsen, P. (1973) Hydraulic roughness of ice covers, *J. Hydraul. Div.*, HY1.

Li, G.H., and Gregory, S. (1974) Diffusion of ions in the seawater and in deep sea sediments, *Geochim. et Cosmochim. Acta, Vol. 38*, pp. 703-714.

Likens, G.E., Eaton, J.S., and Johnson, N.M. (1985) B. Physical and chemical environment, pp. 89-108. In: G.E. Likens (ed.). *An ecosystem approach to aquatic ecology: Mirror lake and its environment*, Springer-Verlag New York, Inc.

Likens, G.E., and Johnson, N.M. (1969) Measurements and analysis of the annual heat budgets for the sediments in two Wisconsin lakes, *Limnol. Oceanogr., Vol. 14*, pp. 115-135.

Likens, G.E., and Ragotzkie, R.A. (1965) Vertical water motions in a small, ice-covered lake, *J. Geophys. Res., Vol. 70*, pp. 2333-2344.

Likens, G.E., and Ragotzkie, R.A. (1966) Rotary circulation of water in an ice-covered lake, *Int. Ver. Theor. Angew. Limnol. Verh., Vol. 16*, pp. 126-133.

Litinskaya, K.D., and Polyakov, Yu.K. (1975) Lakes of Vendyury group – Uros, Rindozero, and Vendyurskoe, pp. 57-66. In *Water resources of Karelia and their use*. Petrozavodsk, Karelian branch of USSR Acad. Sci. (Russ.).

Lozovik, P.A. (1991) Chemical content of suspenses, solid and fluid factions of sediments and exchange processes on the interface "water-sediments", pp. 114-118. In *Surface waters of the Shuya lake-river system under conditions of anthropogenic influence*, Petrozavodsk.

Malm, J., Terzhevik, A., Bengtsson, L., Petrov, M., Boyarinov, P., Glinsky, A., Palshin, N. (1997) Temperature and salt content regime in three shallow ice-covered lakes: 1. Temperature, salt content, and density structure, *Nordic Hydrology, Vol. 28*, pp. 99-128.

Malmberg, A., and Nilsson, T. (1985) Heat transport from water to ice in lakes. M.Sci. thesis, Department of hydraulics, Chalmers University of Technology, Gothenburg, Sweden, 78 pp.

Manheim, F.T. (1970) The diffusion of ions in unconsolidated sediments, *Earth and Planet. Sci. Lett., Vol. 9*, pp. 307-309.

Mortimer, C.H., and Mackereth, F.J.H. (1958) Convection and its consequences in ice-covered lakes, *Verh. Internat. Verein. Limnol., Vol. 13*, pp. 923-932.

Patterson, J.C., and Hamblin, P.F. (1988) Thermal simulation of a lake with winter ice cover, *Limnol. Ocenogr., Vol. 33*, pp. 323-338.

Rahm, L. (1985) The thermally forced circulation in a small, ice-covered lake, *Limnol. Oceanogr., Vol. 30*, pp. 1122-1128.

Sahlberg, J. (1988) Modelling the thermal regime of a lake during the winter season, *Cold regions Science and Technology, Vol. 15*, pp. 151-159.

Sundberg, J. (1986) Heat transfer properties in Swedish soils. Report R104:1986, Swedish Building Research Council (in Swedish).

Svensson, T. (1986) Heat Budgets in ice covered lakes. Measurements in Lake Tulebosjön, pp. 56-70. In Physical processes in lakes (NHP-Report No.16).

Svensson, T. (1987) Temperature and heat exchange in lakes during winter. Swedish Council for Building Research, R 97:1987, 145 pp (in Swedish).

Thanderz, L. (1973) Heat budget studies, pp. 51-78. In Dynamic studies in Lake Velen, Report 31, Swedish Natural Science Research Council, Ed. M. Falkenmark.

Vennard, J.K., and Street, R.L. (1982) *Elementary fluid mechanics*, John Wiley & Sons, New York, 685 pp.

Welch, H.E., and Bergmann, M.A. (1985) Water circulation in small arctic lakes in winter, *Can. J. Fish. Aquat. Sci., Vol. 42*, pp. 506-520.

Received: 19 July, 1996
Accepted: 22 November, 1996

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