The heat budget of Lake Kilpisjärvi in the Arctic tundra

Matti Leppäranta, Elisa Lindgren and Kunio Shirasawa

ABSTRACT

An extensive research programme has been carried through in 2007–2009 on the ice cover geophysics in Lake Kilpisjärvi, located at 69 °03′N 20 °50′E, 473 m above sea level, and about 60 km from the shore of the Norwegian Sea. The surface area of the lake is 37.1 km^2, and the maximum depth is 57 m. Data were collected of ice, snow and weather conditions with an automatic ice station in the lake. The heat budget together with ice structure, growth and melting was analysed. It was dominated by the radiation balance, down to −50 W m^{-2} in fall and up to 100 W m^{-2} in summer. Turbulent heat fluxes were significant before freeze-up in fall (absolute values up to 30 W m^{-2}), but in the ice season they were small except for an occasional sensible heat flux which was large due to warm air advection. The evolution of ice thickness served as a very good condition for the total surface heat flux, and was found to be consistent with the ice station time series of air–ice heat fluxes.

Key words | Arctic lake, field data, ice formation, ice melting, ice thickness

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INTRODUCTION

The Fennoscandian Arctic tundra zone extends over high altitudes in the Scandinavian mountain chain. The northwest corner of Finnish Lapland is located in this zone at 68–69 °N. There the mean air temperature is below freezing point from October to April, and lakes are frozen from November to June, with growth from November to April and melting during May–June. The ice cover consists of congelation ice and snow-ice with snow cover on top, and the annual maximum thickness of ice is close to one meter. As in frozen lakes in general, the ice cover largely decouples the water body from the atmosphere, and, consequently, weakens turbulence, mixing, and sunlight in the water body. Furthermore, circulation becomes thermohaline, the temperature structure is stabilized, and renewal of oxygen is ceased. Up in the Finnish tundra lakes, the polar night lasts two months and the snow cover is thick, and therefore light is turned off for 3–4 months each winter.

Frozen lakes have been examined for more than 100 years, largely for technical applications. Tundra lakes have not been studied as extensively. Recently, ecology and climatology of polar lakes has gained a great deal of new interest, and research is now progressing well (e.g., Salonen *et al.* 2009; Leppäranta 2015). The primary motivation has been to better understand the lake–atmosphere interactions (Kirillin *et al.* 2012), the role of the ice cover in polar lake ecosystems (Lizotte 2008; Keskitalo *et al.* 2013), and fishery (Shuter *et al.* 2012). For the past decades, in most mid-latitude lakes, freeze-up has come later and breakup earlier year-by-year (e.g., Magnuson *et al.* 2000) but for Arctic lakes there is less data and trends are not as clear (e.g., Lei *et al.* 2012). Climate scenarios for the present century predict shorter ice season and thinner ice cover in Arctic lakes (e.g., Brown & Duguay 2010), which raises the question about the consequences to the physical and ecological lake processes if the warming climate predictions become true.

In this work the heat budget of Lake Kilpisjärvi (Gilbbesjávri in Sámi language), located in the Finnish Arctic tundra at about 69 °N 21 °E, is examined. This lake is a medium-size clear water lake on the top of the Tornionjoki river drainage basin. Kilpisjärvi Biological Station of the University of

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Helsinki, founded in 1964, is located on the lakeshore. Consequently, the lake has been much investigated, especially for its ecology. Physics research has taken place mainly after year 2000 (Leppäranta *et al.* 2012; Yang *et al.* 2013). In late summer the surface temperature reaches 10–15 °C, in the course of a seven-month ice season the mean annual ice thickness has become 89 cm (Lei *et al.* 2012).

Long-term ice records of Lake Kilpisjärvi (from 1952 onwards) have been used in ice phenology studies in Finland and in Scandinavia (e.g. Blenckner *et al.* 2004; Korhonen 2006; Lei *et al.* 2012). Lei *et al.* (2012) also included maximum annual ice thickness in their ice climatology research. More detailed physics research based on an automated ice station was commenced in 2007 (Leppäranta *et al.* 2012), and a few years later a field programme on ice melting and under-ice hydrodynamics was completed (Kirillin *et al.* 2014; Lindgren 2015). In mathematical modelling, the coupling between atmosphere and Lake Kilpisjärvi has been examined and found to be significant in weather forecasting (Yang *et al.* 2013) extending most likely to long time-scales.

This paper gives the results of a heat budget analysis in Lake Kilpisjärvi based on fieldwork in 2007–2009. The data of the second ice season 2008–2009 is much longer and better for climatology investigations and therefore mainly used here. Monitoring data of the weather conditions and the lake hydrology by Finnish Meteorological Institute (FMI) and Finnish Environment Institute (FEI) are also utilized. The net surface heat flux was compared with the ice growth and melting and surface temperature changes, which provided an excellent control to the incoming heat flux. According to modelling experiments (Yang *et al.* 2013), the surface heat balance is dominated by longwave radiation all year, solar radiation in summer, and turbulent fluxes in fall. This is confirmed here and quantified using the field data. The specific ice cover deterioration results will provide new insight into the physics during the melt season, which has been the least known part of the ice season of this lake.

FIELD SITE AND OBSERVATIONS

Lake Kilpisjärvi is located in the Scandinavian Mountain Chain in Finland at 69 °03′N 20 °50′E, 473 m above sea level (Figure 1). The distance to the Norwegian Sea in the North Atlantic Ocean is about 60 km. The lake is at the northwest end of the long River Könkämäeno valley, fells rising up to 1,000 m elevation (500 m from the lake surface) on both sides. The surface area is 37.1 km^2, the average and maximum depths are 19.5 m and 57 m, respectively, and the maximum fetch is 6.2 km. The inflow comes from small mountain brooks, and outflow is in the southeast corner to River Könkämäeno and further to Tornionjoki River and to the Baltic Sea. There are no industrial or agricultural activities in the drainage basin, apart from reindeer herding. In Kilpisjärvi village there are less than 100 inhabitants but annually close to 40,000 tourists visit the region.

Automated ice station *Lotus* was deployed in the lake for measurements during the ice seasons (Table 1). The data include atmospheric surface layer (temperature, humidity, wind speed and direction, solar radiation), snow and ice cover (radiative surface temperature, snow depth, vertical temperature distribution, attenuation of sunlight), and lake water body beneath the ice (temperature, electric conductivity, sunlight, current velocity).

Incoming and outgoing solar radiative fluxes were measured by pyranometers (CM3, PCM-03H, Kipp & Zonen, Holland). Radiative surface temperature was detected by an infrared sensor (THI-303N, Tasco Ltd, Japan). Wind, temperature and relative humidity were measured at about 3 m height above the surface. Turbulent heat fluxes were estimated from these data with bulk formulae with constant transfer coefficients corresponding to neutral stratification (e.g., Leppäranta, 2015). Therefore the estimated autumn values may be a little biased down the way but that does not influence the main conclusions. Snow depth was measured using the sonic ranging sensor (SR50M-45, Campbell Scientific, Canada). Light measurements were made using PAR (Photosynthetically Active Radiation, 400–700 nm) sensors, and snow depth was measured from above using an ultrasonic sensor.

The time interval was 10 minutes in the original data. For the analyses in this study, the original sensor data were averaged into 1-hour time intervals using a binomial 5-point filter. The final results in the tables and figures later in this paper are shown in daily averages. Since 1999 *Lotus* has been used for data collection in several lakes and coastal sites in Finland (e.g., Wang *et al.* 2005).

The first extensive field programme on the physics of Lake Kilpisjärvi ice season was performed in

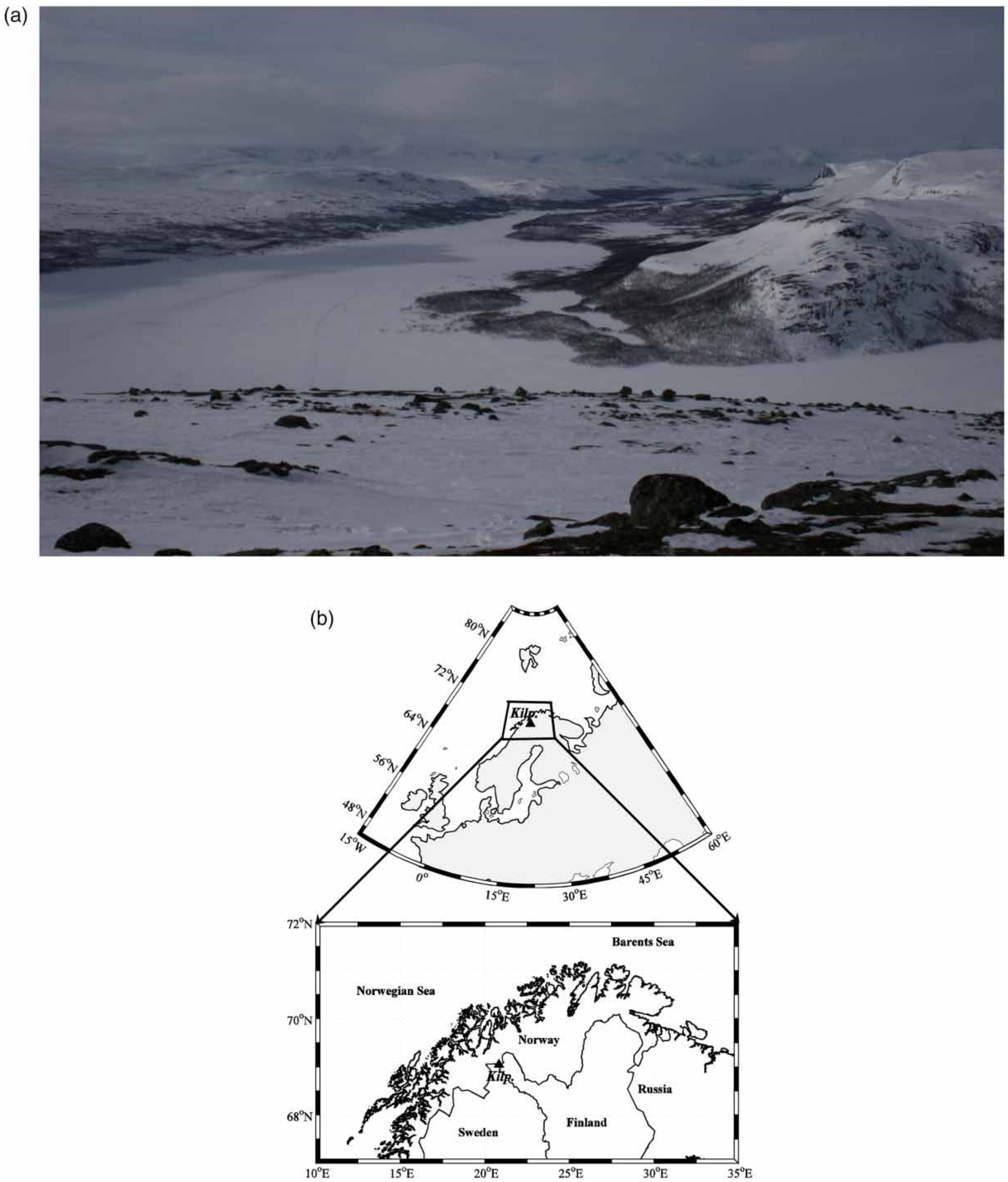

Figure 1 | Lake Kilpisjärvi in tundra mountain area of Lapland at 69 °03′N 20 °50′E, 473 m above sea level. (a) Photograph of the lake taken on April 13, 2009 from Saana fell. (b) The location shown on a map.

2007–2009, over two ice seasons (Leppäranta *et al.* 2012). This was a joint Finnish-Japanese research project. The research work was based on ice station *Lotus* (Figure 2), manual field measurements, and utilization of routine observations. The ice station data was complemented by manual ice observations a few times in an ice season.

In the northwest bay of Lake Kilpisjärvi there is a routine ice and snow monitoring station of the FEI (http://www.ymparisto.fi/; see also Korhonen 2005). Observations are made every 5–15 days from November to May. Ice phenology records are available since 1952. Ice thickness measurements were commenced in 1964, and in 1977, snow depth on ice was added to the observation procedure.

Table 1 | Instrumentation of the ice station in the study winters

	2007–2008	2008–2009
Location	69 °03.344′N 20 °46.445′E	69 °03.360′N 20 °46.378′E
Period	14 Dec–12 Jun	9 Sep–21 May
Radiation	Solar in & out, net	Solar in & out, net
Temperature	Surface, 3.1 m	Surface, 2.9 m
Humidity	3.1 m	2.9 m
Wind	3.2 m	3.2 m
Other	Snow depth, air pressure	Snow depth, air pressure

Snow-ice observations were made in 1981–1990. The mean freezing and breakup dates were November 9 and June 18 in the period 1952–2015, the earliest freezing date was October 21 and the latest breakup date was July 1. The average maximum annual ice thickness was 89 cm (the range was 77–114 cm), reached in April. In the open water season, FEI also provides the surface water temperature measured with a contact thermometer in the surface layer.

Weather station Enontekiö Kilpisjärvi Kyläkeskus (EKK) was founded in Kilpisjärvi village in 1962 by the FMI. Routine weather data are available throughout the year, including wind speed and direction, air temperature, relative humidity, cloudiness and precipitation. Statistics based on this weather data are shown in the website of Kilpisjärvi Biological Station (http://www.helsinki.fi/kilpis/english/index.htm). In 1962–2015 the annual mean air temperature was between –4 °C and –1 °C, with mean monthly air temperatures below 0 °C from October to April. Snow thickness on the ground was on average 90 cm in March–April at the lake. According to the modelling work by Yang *et al.* (2013), the *Lotus* Ice Station data and EKK weather data fit fairly well together when they constructed the atmospheric forcing data for the model simulations.

ALL-YEAR HEAT BUDGET

The heat content of Lake Kilpisjärvi consists of sensible heat and latent heat. A convenient reference is the heat content of the isothermal liquid state of water at 0 °C (E_0). The actual heat content, per unit area, is then $E = E_0 + \Delta E$, where

$$\Delta E = \frac{1}{A}\left(\int_V \rho cTdV - \int_A \rho_i LhdA\right) \quad (1)$$

Figure 2 | Automated ice station *Lotus* in Lake Kilpisjärvi.

Here A is lake surface area (km^2), V is lake volume (m^3), ρ and ρ_i are water and ice densities (kg m^{-3}), respectively, c is specific heat (J kg^{-1} °C^{-1}) of liquid or solid water, T is temperature in °C, L is latent heat of freezing (J kg^{-1}), and h is ice thickness (m); note that in the volume integral ρ and c may be for liquid water ($T \geq 0$ °C) or ice ($T \leq 0$ °C). The mass balance is taken as zero, i.e. inflow, evaporation/sublimation, and precipitation are assumed to compensate for their net mass flux in the outflow to the River Könkämäeno. Therefore the lake mass $\rho_i V_i + \rho_w V_w$ is constant, where the lower index w refers to liquid water.

It is further assumed that the inflow and outflow do not significantly influence the heat budget. Thus the heat content of the lake changes only by the solar radiation, atmospheric fluxes at the top surface, and fluxes at the lake bottom sediment:

$$\frac{d}{dt}\Delta E = Q_n = (1-\alpha)\gamma Q_s + Q_0 + Q_b \tag{2}$$

where t is time (s), Q_n is the net heat flux (W m^{-2}) to the lake body with its ice cover, Q_s is incoming solar radiation (W m^{-2}), α is albedo, γ is the proportion of solar radiation penetrating into the lake (about the same as the proportion of the optical band), Q_0 is surface heat balance (W m^{-2}), and Q_b is the heat flux (W m^{-2}) from the bottom. The surface heat balance is

$$Q_0 = (1-\alpha)(1-\gamma)Q_s + Q_{La} - Q_{L0} + Q_c + Q_e + Q_P \tag{3}$$

where Q_{La} and Q_{L0} are the terrestrial radiation fluxes (W m^{-2}) from the atmosphere and from lake surface, respectively, Q_c and Q_e are atmospheric sensible and latent heat fluxes, respectively (W m^{-2}), and Q_P is the heat flux (W m^{-2}) from precipitation. It is seen that solar radiation distributes into internal source (fraction $\gamma \sim 1/2$, the first term in the right-hand side of Equation (2)) and a term in the surface balance (fraction $1-\gamma$, the first term in the right-hand side of Equation (3)). The bulk formulae for the turbulent fluxes are

$$Q_c = \rho_a c_a C_H (T_a - T_0)U, \;\; Q_e = \rho_a L{*} C_E (q_a - q_0)U, \tag{4}$$

where ρ_a is air density (kg m^{-3}), c_a is specific heat of air (J kg^{-1} °C^{-1}), $L*$ is latent heat of evaporation or sublimation (J kg^{-1}), C_H and C_E are turbulent transfer coefficients of sensible and latent heat, respectively, T is temperature (°C) and q is specific humidity with subscripts a for air and 0 for surface, and U is wind speed (m s^{-1}).

RESULTS

Statistics of the ice seasons

The ice season lasts about eight months in Lake Kilpisjärvi, from November to June (Table 2). The ice cover consists of three principal layers: congelation ice, snow-ice and snow. Occasionally there may be slush sublayers in the snow or snow-ice layer. In slush formation, snow is compacted so that the thickness of slush is less than the original snow thickness. In the ice seasons in 2007–2009, the mean snow and snow-ice thickness corresponded to snow accumulation by about 70 cm, which is less than snow accumulation on the ground (on average 90 cm, from FEI data archives), the difference is largely due to drifting of snow from the lake. The snow in snow-ice corresponded to 30 cm snow accumulation due to snow compression in snow-ice formation (see Leppäranta & Kosloff 2000), and when the snow cover on ice is added to this, we have altogether a representative snow accumulation of 70 cm.

Table 2 | Seasonal and average ice characteristics in the study winters. Comparison to climatology is also provided (Lei et al. 2012): freeze-up and breakup from 1952–2010, ice thickness from 1981–1990, and snow from 1977–2010

	2007–2008	2008–2009	Mean	SD
Freeze-up date	14 Nov	10 Nov	8 Nov	8.2 d
Maximum ice thickness (cm)	85	88	90	7.5
- congelation ice	51	78	70	17.3
- snow-ice	34	10	20	17.1
- date of occurrence	20 Apr	30 Apr	14 Apr	14.9 d
Maximum snow thickness[a]	36	43	37	9.1
- date of occurrence	20 Apr	20 Mar	4 Mar	34.4 d
Ice breakup date	21 Jun	8 Jun	18 Jun	6.8 d

[a]Snow on lake ice.

External heat fluxes

Lotus station, complemented by EKK weather station data, provided the incoming solar radiation, albedo, air and surface temperature, humidity, and wind speed during the ice seasons. In the open water period we utilized the weather station data, and also in the Lotus records there were a few blackouts, which were filled from the EKK source. Cloudiness observations were obtained from EKK so that the radiation balance could be evaluated. In addition, the surface temperature was available in the open water season at the FEI Kilpisjärvi lake station.

The open water heat balance is well illustrated by the surface temperature evolution (Figure 3). The ice breaks up in June, and very soon the surface temperature is above 4 °C since the warm up of water begins when there is still an ice cover on top. The surface temperature site is in a bay, and therefore at complete ice breakup over the lake, the surface temperature may be well above 0 °C in the bay as was the case in summer 2007 (Figure 3). The breakup must have been then a slow process. A year later there was a 5-day lag from breakup to surface layer warming to 5 °C. A characteristic of Arctic lakes is that the heat budget may be continuously positive in summer due to the presence of polar day.

In summer, after ice breakup, the surface temperature increased to 10–15 °C on about August 10. The rate of warming was thus 2.5–5 °C per month, and there were large variations due to different weather patterns as well as due to thermocline erosion and setup. The cooling period was more stable, and in these two ice seasons with data collection, the curves were quite similar. The rate was close to 5 °C per month, caused by the cooling weather and thermocline deepening by convective and mechanical mixing. Warming came from solar radiation and cooling from radiational and turbulent heat losses. Assuming a representative mixed layer depth as 5 m, based on occasional soundings, the mean cooling rate corresponded to heat loss at the rate of 40 W m^{-2} while in the warm-up period, energy absorption was 20–40 W m^{-2}.

Ice seasons 2007–2008 and 2008–2009 were close to normal for ice growth and snow accumulation. However, in 2007–2008 there was much more snow-ice than normal but the opposite was observed a year later. Otherwise the heat fluxes were similar between these two winters.

The heat flux from lake water to ice was not directly available and it became a part of the residual term in the

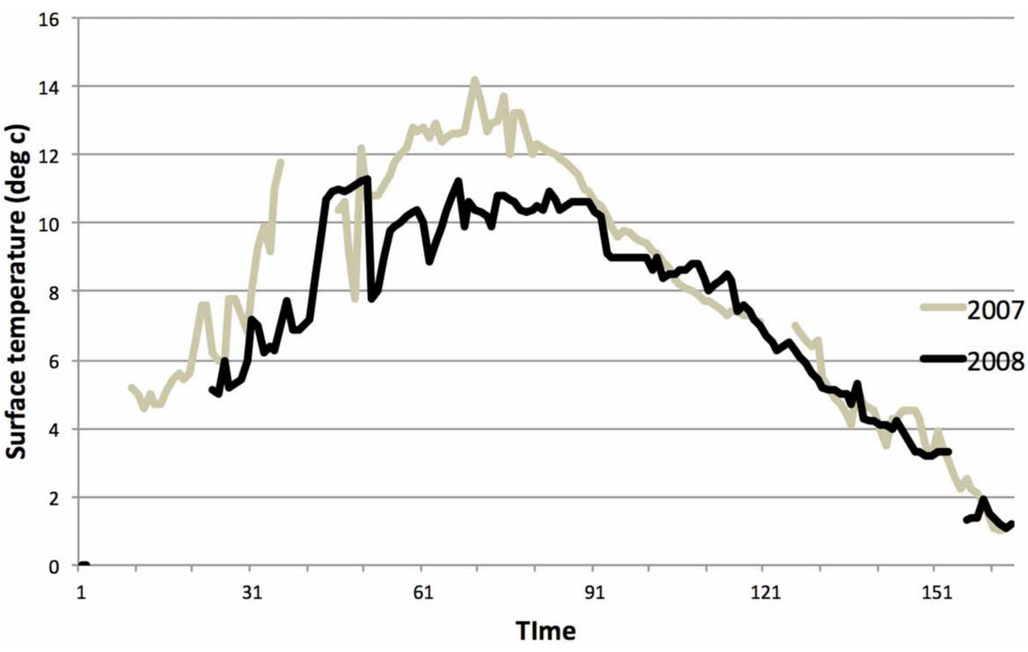

Figure 3 | Daily surface temperature in the open water seasons in 2007 and 2008 (the data are from the FEI archives). The arrows show the ice breakup dates.

heat budget. In any case, this flux is expected to be fairly small, at most around 5 W m^{-2} when compared with results from similar lakes in Finland (Leppäranta 2015). Since the surface heat fluxes were order of tens of W m^{-2}, the heat flux below 5 W m^{-2} from bottom sediments, as assumed, was quite a small factor. In shallow lakes, the heat storage in bottom sediments can be an important factor in the ice season (e.g., Bengtsson 1996; Malm *et al.* 1997).

Surface heat balance in ice season 2008–2009

In the second ice season, *Lotus* data collection was commenced on September 9, 2008, two months before freezing. It was possible to examine then how the ice coverage influenced the surface heat balance and how ice growth was determined. Further on, it was seen how ice and snow melting was triggered in springtime. The outcome is illustrated by the daily mean surface heat fluxes in Figure 4.

The surface water temperature was 8.6 °C at the beginning of the *Lotus* data series, on 9th September. A month later the surface temperature was 5.2 °C, and in October it dropped down to 3.3 °C. Thereafter, with the development of stable stratification in the water body, freezing took place on 10th November. The mean cooling rate was thus 0.14 °C day^{-1} from 9th September until freeze-up. The *Lotus* data shows that the cooling was mainly due to radiational and latent heat (evaporation) losses.

In September 2008, solar radiation was still strong, and the net surface heat flux was positive until 25th September, on average at the rate of 20 W m^{-2} (Figure 4(a)). Net terrestrial radiation was between −60 and −30 W m^{-2}. Then, with the loss of solar power, the net surface flux turned strongly negative, daily values reaching −100 W m^{-2} at minimum. At the time of freeze-up, the net radiation was mainly terrestrial with losses in the range of 20–40 W m^{-2}. Turbulent heat fluxes were between −40 and 20 W m^{-2}. Latent heat flux was always negative and less than −10 W m^{-2}; from definition, a flux of −10 W m^{-2} corresponds to sublimation by 0.3 mm per day. Sensible heat flux had a smaller role and it could be either a minor gain or a loss term, less than 20 W m^{-2}. Latent heat losses were thus close to the radiational losses, and together these terms determined the cooling rate. In the open water season, assuming a representative mixed layer depth as 5 m, the mean cooling rate corresponded to the heat loss by the rate of 34 W m^{-2} (see Figure 4(a)).

After the freeze-up, negative net heat flux is seen as an increasing in ice thickness, i.e. the latent heat released in freezing corresponds to the net flux. Growth of ice by 1 cm per day releases latent heat by the rate of 32 W m^{-2}. In the presence of growing, compact ice cover, the turbulent fluxes were decreasing. This was because the temperature and humidity gradient decreased in the atmospheric surface layer due to the ice cover isolating the warm (0 °C) water from the cold air (Figure 4(a)). But radiative heat losses were not much changed, and the ice could grow by 23 cm during November 10–30. In December the net surface flux was just slightly negative, and occasionally sensible heat flux reached 20 W m^{-2} due to warm air advection from the northeast North Atlantic Ocean.

During the polar night, there was a blackout in the automatic station through December 20–February 5. Then in winter the net surface flux was within about −20 to 0 W m^{-2} (Figure 4(b)). In February–March the net terrestrial radiation was from −50 to −30 W m^{-2}, partly balanced by solar radiation and sensible heat flux. The latter also was consistently positive, with westerly winds bringing Atlantic air to the region. Latent heat flux remained small through the whole winter. From December to April the ice was growing, and the growth rate slowly decreased with time, as was also well reflected in the surface heat balance.

On April 23 the heat flux turned positive and snow melting commenced. This development was due to increasing of solar radiation day-by-day and also it was seen that the albedo started to decrease already before melting started, likely due to metamorphic processes in the snow layer. The albedo of clean, dry snow is around 0.9, and it needs to be pushed down a bit for the surface heat balance to turn positive. Once there, albedo further decreases due to positive feedbacks such as liquid water formation in the surface snow layer.

In spring/summer 2009 the net surface heat flux reached levels of 50–100 W m^{-2} in the first half of May, then increased up to 200 W m^{-2} in late May, in agreement with the very fast rate of melting of ice. Ice melt rate actually corresponded to the heat flux of 130 W m^{-2} but this value is not real since a part of solar radiation was used for warming the water beneath the ice. In the melting season the

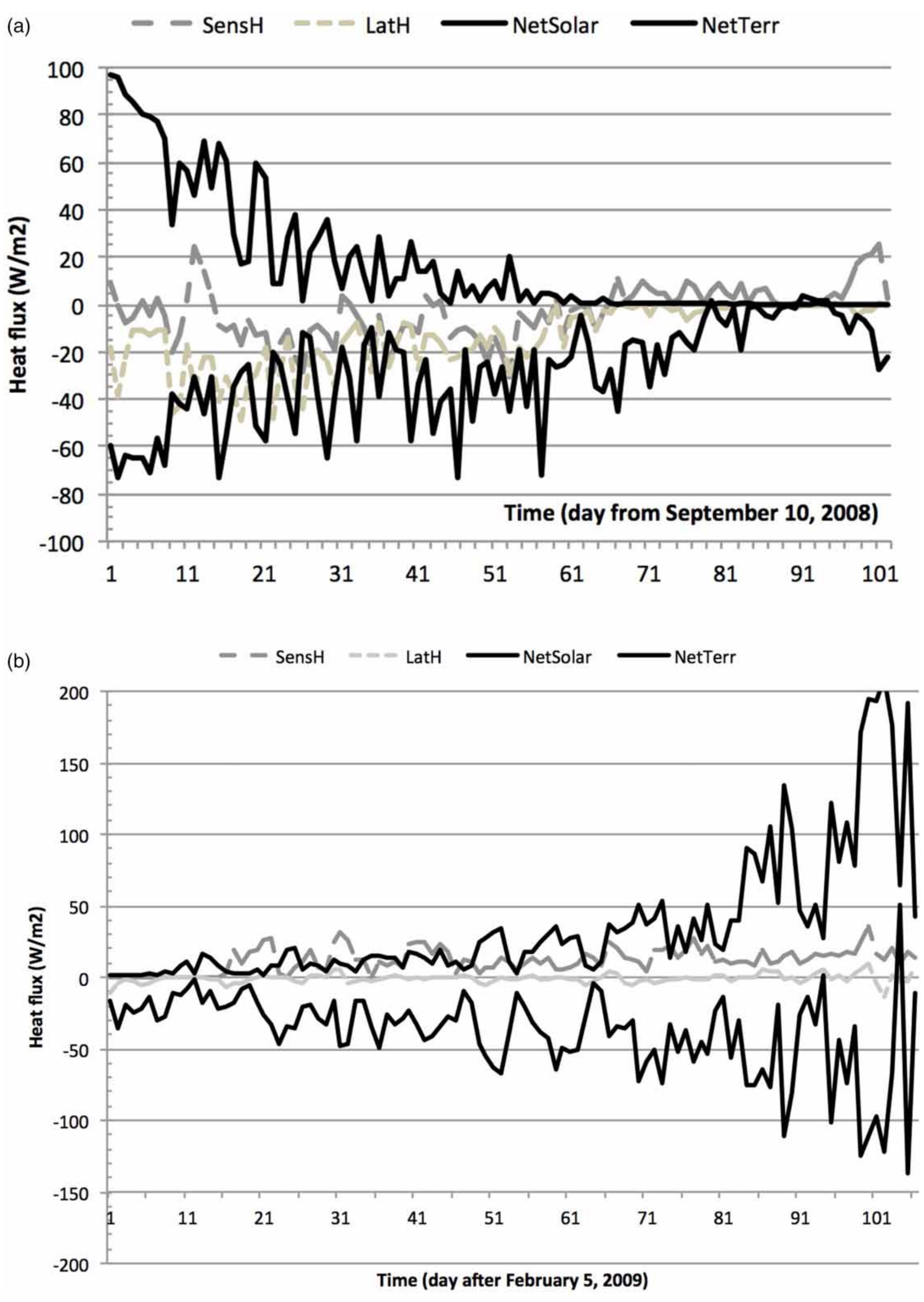

Figure 4 | Daily surface heat balance of Lake Kilpisjärvi. (a) September 10–December 20, 2008; (b) February 5–May 20, 2009. The time of freeze-up was on day 62 in (a).

magnitude of turbulent fluxes was less than 20 W m^{-2}, sensible heat flux was positive, and latent heat flux was negative. The first recording of surface temperature was 5.1 °C on June 24, 16 days after ice breakup. The rate of warming looks like being twice as fast as later in summer, but this is likely not true since the initial conditions of water temperature could be well above 0 °C at ice breakup.

Ice growth and melting

During the ice season, negative and positive net external heat fluxes result in growth and melting of ice, respectively. In fact, we need to consider three layers, congelation ice, snow-ice and snow, separately. The changes in the thicknesses of these layers are illustrated in Figure 5 at 10-day intervals in the ice season 2008–2009. Ice growth rate was on average 0.5 cm per day, while melting was very fast, at the rate of 1.5–3 cm per day.

The solar, atmospheric and bottom heat fluxes modify the temperature or ice thickness in the lake. In the ice seasons, the thicknesses of the three layers in the ice cover were manually measured, and by far most of the external heat fluxes were seen in changes in the thicknesses of these layers. Heat was lost in ice growth and gained in ice and snow melting. Thus a lake ice cover is a good sensor or a boundary condition for the total heat budget, a simple fact that has not been actually utilized before for field data analyses. Define

$$S = [\rho_i(h_{ci} + h_{si}) + \rho_s h_s - \rho_w h_{sw}]L \tag{5a}$$

where h_{ci}, h_{si} and h_s are the thicknesses (cm) of congelation ice, snow-ice and snow layers, h_{sw} is the equivalent thickness (cm) of liquid water in the ice, ρ_i, ρ_s, and ρ_w are ice, snow and water densities (kg m^{-3}), and L is the latent heat of freezing (J kg^{-1}). The heat storage S tells how much heat is needed for ice and snow melting. But in the growth of ice cover, latent heat is released to form congelation ice but the half of the snow-ice layer comes from snow accumulation and is in the solid state originally. As well, snow falls in solid state. Thus, to grow the ice, the necessary heat loss is

$$S' = \rho_i\left(h_{ci} + \frac{1}{2}h_{si}\right)L \tag{5b}$$

The heat storages (Figure 4(a) and 4(b)) can be changed to fluxes by taking the time derivative; the layer thicknesses change but the physical parameters are kept constant.

Consequently, considering the thicknesses of the layers in the ice cover, the mean rate of heat loss from the lake to form congelation ice and snow-ice becomes 18.5 W m^{-2} during the 5-month growth season. Taking the snow density as 400 kg m^{-3} (Järvinen & Leppäranta 2011), we can estimate the heat storage in the snow cover. Then, to melt the

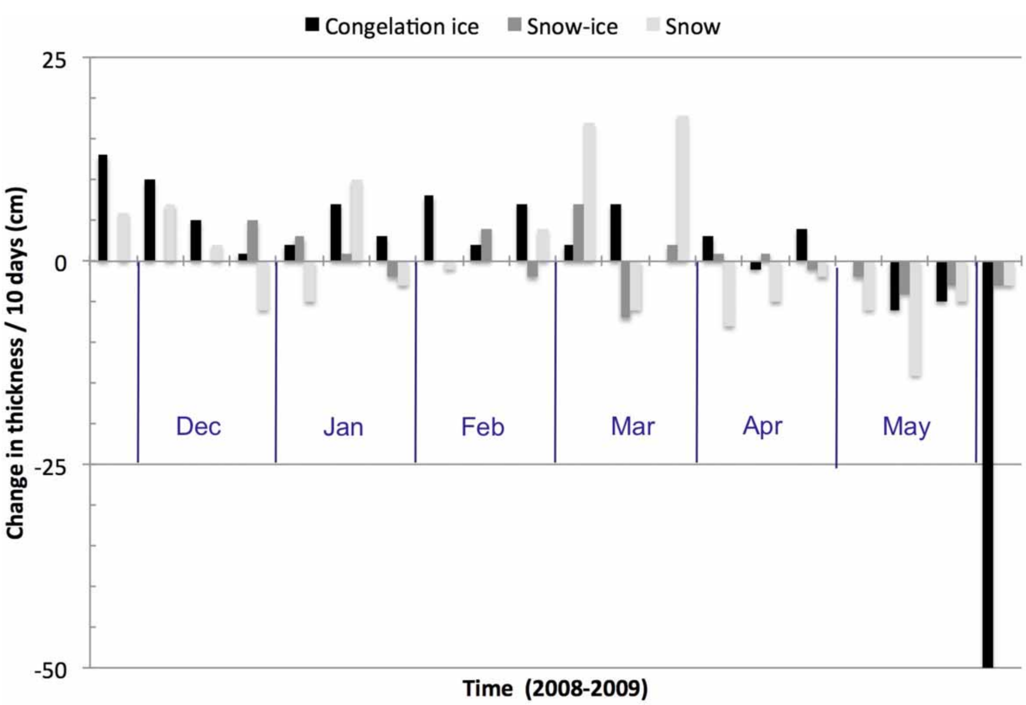

Figure 5 | Changes of congelation ice, snow-ice, and snow thickness in Lake Kilpisjärvi, ice season 2008–2009. The changes are shown at 10-day intervals.

snow and ice in 50 days time, the daily average net heat flux must be 75.2 W m^{-2}, which corresponds to 2.1 cm loss of ice per day. Melting takes place at the boundaries and in the interior, so that ice becomes porous and fragile, breaks due to its own weight into brash ice, and thereafter disappears very quickly. After this breakage event, the heat balance changes since albedo becomes close to open water albedo and turbulence in the water brings more heat from below.

Since the heat flux from the lake bottom sediments was small, at least apart from late melting period, the evolution of ice thickness should correspond to the total surface heat flux. This was, in fact, observed (Figure 6). Consequently, ice growth and melting served as an excellent external condition. This is a key issue especially when standard routine weather or climate data are used to estimate the individual heat fluxes that contain large uncertainties.

CONCLUSIONS

The heat budget of Lake Kilpisjärvi in the northwestern Finnish Lapland has been examined based on field surveys. The focus is on the ice season, with the mean freezing and breakup dates of November 8 and June 18, respectively. A two-year field study was performed there in 2007–2009 with data collected for ice, snow and weather conditions based on an automatic ice station *Lotus* in the lake and routine weather and hydrological monitoring data.

The heat budget was dominated by the radiation balance. Terrestrial radiation losses created the ice sheet, while solar radiation melted the ice. Growth rate was 0.5–1 cm per day, while melt rate averaged to 2.1 cm per day. Turbulent fluxes were large during the cooling period in September–October when the lake was open, but after the freeze-up they became small, magnitude below 20 W m^{-2}. Occasionally sensible heat flux could be large but it remained small on the monthly average level. Latent heat flux was small during the whole ice season. The heat balance in the melting period is quite similar in southern Finland, e.g. in Lake Pääjärvi at 61 °N (Jakkila *et al.* 2009).

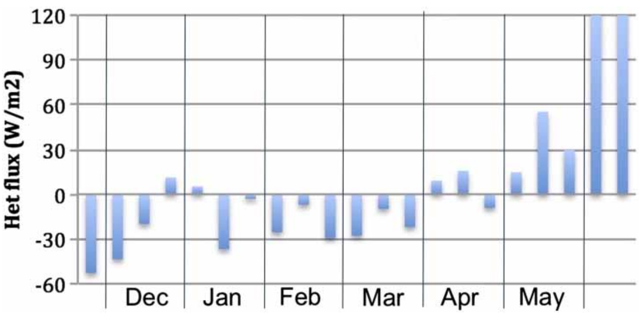

Figure 6 | The net surface heat flux due to ice growth and melting of ice and snow in the ice season 2008–2009.

After ice melting, the surface temperature had interannual variability due to radiation and mixing conditions, with annual maximum in 10–15 °C.

The growth and melting of ice served as a very good control condition for the surface heat balance, especially when the heat flux from the water body to ice was small or known. The latent heat released due to ice growth and needed for ice melting agreed well with the observed surface heat balance. Both congelation ice and snow-ice formed, with a larger fraction of the former type. Loss of snow or ice due to sublimation was small, in the order of 0.3 mm per day in ice equivalent. The net surface heat flux turned negative close to the autumn equinox, and the total radiation balance was negative from October 10 to April 15. Due to positive feedback mechanisms in ice melting, the rate of melting was increasing after start-up. Polar day begins on May 22 and lasts two months. The period of ice melting is strongly tied to solar radiation. In fact, the ice breakup date takes place in a narrow range, between June 3 and July 1 (Lei *et al.* 2012).

The sensible heat flux was negative in fall, typically between −10 and −30 W m^{-2}. Occasionally it was positive in winter, up to 30 W m^{-2}. In the polar night it was the only positive heat flux, and there was a case in December when sensible heat flux could turn the surface heat balance to the positive side. The reason was a flow of warm air from the North Atlantic Ocean. In spring and summer, turbulent heat fluxes were small. In May the absolute value was less than 5 W m^{-2}. This means also that sublimation makes only a minor contribution to the ice cover mass balance. The evolution of ice thickness corresponded to the total surface heat flux, and thus the ice growth and melting served as an external condition for the surface heat balance. This is a key issue since often estimates of the individual heat fluxes contain large uncertainties.

It has been questioned on what conditions perennial ice could occur in Lake Kilpisjärvi. Considering the strong role of solar radiation, July and August would also belong to the melting period, and consequently at least 120 cm more ice would be needed for the survival of ice. To grow this much ice, the mean winter air temperature would need to be about 10 °C less than presently. Another possibility is to have more snow timed with accumulation so as not to favour snow-ice formation (e.g., Leppäranta 2015). By this, the start of ice melting would be shifted forward, and the necessary climate cooling would be less. This question needs to be further examined using paleolimnological data and mathematical models. Also key research topics are now ice–water body interaction (Graves *et al.* 2014; Kirillin *et al.* 2014), ice melting (Lindgren 2015), and modelling the annual cycle of lake ice and temperature (Yang *et al.* 2013). Comparison with other tundra lakes in northern Finland and Scandinavia is in progress.

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