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Instantaneous daytime conductive heat flow through snow on lake ice in Alaska

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Abstract:

The instantaneous daytime conductive heat flow through the snow on lake ice was derived from snow depth, temperature and density measurements made during the course of six winters at MST Pond in central Alaska. The MST Pond data for winter 2003–04 are compared with results for the same period at six other sites (Barrow, Nome, Amos Lakes, Fairbanks, Wasilla, Seward) in Alaska. The maximum heat flow at MST Pond has varied between $-19{\cdot}5$ and $-8{\cdot}8$ W m^{-2}. Each winter, the heat flow decreases as the total thickness of snow and ice increases with time. Superimposed on this trend are variations due to fluctuating air temperatures. The comparison of the MST Pond data with the other locations in Alaska reveals heat flow differences that reflect different weather conditions, particularly air temperature and wind, and the latter's effect on snow depth and density. Notwithstanding the regional differences, the heat flow values are of the same order of magnitude as those obtained for sea ice in the Arctic and Antarctica. The implications for the total winter conductive heat loss at large lakes and for regions where many small lakes cover a large proportion of the land are discussed. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS snow; lake ice; conductive heat flow; Alaska

INTRODUCTION

Two primary processes contribute to the thermodynamic thickening of lake ice: freezing of water at the bottom of the ice cover creates congelation ice, and freezing of slush at the top of the ice cover creates snow ice (Adams, 1981). Slush occurs at the top of the ice cover when the mass of accumulating snow is sufficient to depress the ice surface below the waterline and water flows up through fractures (or gaps between ice and shore) to the ice surface and soaks the base of the snow cover. Congelation ice is sometimes referred to as black ice because its high optical depth allows light to pass right through to the underlying water. Snow ice is sometimes referred to as white ice because its high bubble content causes strong light scattering.

Water freezes at the bottom of the ice and slush freezes at the top of the ice because of negative temperature gradients in the ice and snow (for congelation ice) and the snow alone (for snow ice) that induce conductive heat flow to the atmosphere. In ice that grows only by freezing (thermodynamic thickening), as opposed to dynamic thickening by deformation, the conductive heat flow determines the ice growth rate and the ice thickness.

The conductive heat flow through sea ice and its snow cover has been well documented in the Arctic (Sturm *et al.*, 2001, 2002) and the Antarctic (Massom *et al.*, 1997, 1998; Sturm *et al.*, 1998). It is a major source of heat transfer from the ocean through sea ice and snow to the atmosphere, and a major component of the winter surface energy budget of sea ice (Maykut and Untersteiner, 1971; Maykut, 1978). The same must apply to lake ice, but much less is known about the magnitude and variability of the conductive heat flow through lake ice and its snow cover.

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Jeffries *et al*. (1999) simulated the conductive heat flow through snow on lake ice on the North Slope of Alaska; they found daily heat flow fluctuations (due to daily air temperature variability) superimposed on a trend of decreasing heat flow as the winter progressed (due to the increasing combined thickness of snow and ice). Heat flow values derived from field measurements in mid April 1997 validated the simulated values for the same period, but left the veracity of the magnitude of the values simulated during the previous 7–8 months in question.

The rationale for the Jeffries *et al*. (1999) conductive heat flow study was that understanding climate change and variability at high latitudes requires knowledge of the surface heat budget, including the spatial and temporal variability of heat sources. The same rationale applies to this paper, which builds on Jeffries *et al*. (1999) by using the same empirical approach to derive the instantaneous conductive heat flow from daytime measurements of snow properties over the course of six entire winters (autumn 1999 through to spring 2005) at a frozen pond in central Alaska, and at six other frozen lakes/ponds during a single winter (2003–04) in other parts of Alaska.

STUDY SITES AND METHODS

The primary study site is MST Pond (unofficial name), a natural water body on the University of Alaska Fairbanks Poker Flat Research Range in the Chatanika River valley, 50 km northeast of Fairbanks, Alaska. We will also present some data from three artificial ponds in the same valley: 31·6 Mile, 33·5 Mile and 34 Mile ponds are gravel pits, used for construction of the Steese Highway, that filled with water after being abandoned. MST Pond is at Mile 30. All the ponds are small (90–110 m across) and shallow (maximum depth 3–5 m). The geographic coordinates for the Poker Flat ponds are given in Table I.

The MST Pond study site is set up in mid to late October, when the ice is thick enough (10–15 cm) for two or three people to walk on safely. In winters 1999–2000 and 2000–01, single 80 m long and 100 m long transects respectively were marked with wooden stakes frozen into the ice. Each winter since 2001–02, the study site has comprised a 100 m long transect and two shorter (55–60 and 70 m) transects oriented perpendicular to the long line at distances of 35 and 65 m.

Using Sturm–Holmgren probes (e.g. Sturm *et al*., 1998), snow depth is measured to an accuracy of 1 mm. The 1 m long probes have a thermistor in the tip and a connector at the other end. The temperature at the bottom of the snow cover is measured to an accuracy of 0·1 °C with a digital temperature reader connected to the probe. In the first winter, snow depth was measured at 1, 5 and 10 m intervals along the transect, depending on the weather conditions, the number of people and the amount of time available. In subsequent winters, it was measured at 5 m intervals.

Table I. Geographic coordinates of the study sites

Location	Latitude (°N)	Longitude (°W)
Poker Flat sites		
MST Pond	65·131	147·455
31·6 Mile Pond	65·137	147·442
33·5 Mile Pond	65·148	147·387
34·0 Mile Pond	65·151	147·369
ALISON sites		
Imikpuk Lake, Barrow	71·323	156·651
ACSA Pond, Nome	64·521	165·421
Amos Lakes (Mystic)	62·863	152·620
Aurora Pond, Fairbanks	64·853	147·756
Lucille Lake, Wasilla	61·571	149·155
Bear Lake, Seward	60·192	149·361

The bulk density of the entire depth of snow is measured to an accuracy of 10 kg m^{-3} by weighing a sample of known volume obtained using a steel snow tube (length 0·3 or 0·5 m, depending on snow depth) with a cross-sectional area of 21·7 cm^3. Snow samples were obtained at nine locations along the sampling transect in 1999–2000, 11 locations in the 2000–01, and six to eight locations in all subsequent winters. A mean density value for each visit to the pond is calculated from the multiple snow samples. No systematic snow crystal size and morphology measurements were made, but the gradual metamorphosis of the snow cover, primarily due to depth hoar development, was obvious as snow density samples were obtained during the course of the winter.

The conductive heat flux F_a (W m^{-2}) through the snow cover at each point along the transect is obtained using the Fourier heat conduction equation (Sturm *et al*., 1998, 2001):

$$F_a = k_{eff}(dT/dz_{snow}) \approx k_{eff}[(T_s - T_b)/z_{snow}] \quad (1)$$

where k_{eff} (W m^{-1} K^{-1}) is the bulk thermal conductivity of the snow cover, $(T_s - T_b)/z_{snow}$ (°C m^{-1}) is the temperature gradient, T_s is the snow surface temperature, T_b is the temperature on the ice surface at the bottom of the snow cover, and z_{snow} is the snow depth. T_s is the average of measurements made in the shade with a snow probe at the beginning (0 m) and end (100 m) of each transect. The latter measurements are made within 15 min of each other. In winter 1999–2000 we measured snow surface temperatures at 1 m intervals and found that the mean of the first and last measurements is representative of 101 measurements made over a period of 1 h. A mean F_a value for each visit to the pond is calculated from the multiple individual calculations. The assumptions underlying this method, including the conductive heat flow through the snow being representative of that through the ice and snow, are addressed at length elsewhere (Sturm *et al*., 1998, 2001; Jeffries *et al*., 1999).

k_{eff} is estimated as a function of density ρ (g cm^{-3}) based on experimentally determined parameterizations derived by Sturm *et al*. (1997):

$$k_{eff} = 0{\cdot}138 - 1{\cdot}01\rho + 3{\cdot}233\rho^2 \quad [0{\cdot}156 \leq \rho \leq 0{\cdot}6] \quad (2)$$

$$k_{eff} = 0{\cdot}023 + 0{\cdot}234\rho \quad [\rho < 0{\cdot}156] \quad (3)$$

To place the MST Pond results in a broader context, we have developed ALISON (the Alaska Lake Ice and Snow Observatory Network; www.gi.alaska.edu/alison), a science education project in which participating teachers and students make snow measurements and derive the conductive heat flow at other locations in Alaska. In this paper we present results from six locations (Table I): Barrow and Nome in northwest Alaska, Amos Lakes and Fairbanks in central Alaska, and Wasilla and Seward in south-central Alaska. The teachers and students use the same types of equipment and follow the same procedures that we use at MST Pond: snow depth and snow base temperatures are measured at 5 m intervals along a 100 m transect; snow surface temperature is measured in the shade at the beginning and end of the transect; a mean snow density value is based on the density of three samples of known mass and volume.

Almost without fail the Poker Flat measurements have been made once per week at much the same time of day (10:00–11:00 a.m.) each winter. The timing of the ALISON measurements is much more variable and is often determined by factors beyond the teachers' control, e.g. class schedules, examinations, holiday and vacation closures, and temperature-based rules governing outdoor activities in cold weather. Thus, the ALISON data have not been obtained on as regular a basis as the Poker Flat data.

RESULTS

MST Pond, Poker Flat

MST Pond was visited in all weather, ranging from mild to very cold, as indicated by the snow surface temperatures that ranged from 0 °C to as low as −45 °C (Figure 1a–f). The snow buffered the ice surface from the cold air above; consequently, temperatures at the bottom of the snow were higher and less variable than the air temperatures, and were rarely lower than −10 °C (Figure 1g–l). Because of the relatively thin snow (Figure 1s–x) and the large temperature differences between the top and bottom of the snow, most temperature gradients were very steep (Figure 1m–r) and often exceeded the minimum (−25 °C m^{-1}) needed for temperature-gradient metamorphism and the development of skeleton-type depth hoar (Akitaya, 1975). This has been the dominant crystal type in the snow on the ice by the end of each winter.

The depth of snow on the ice was similar each winter and only occasionally exceeded 0·3 m (Figure 1s–x). Brief snow depth peaks and subsequent snow depth decreases (e.g. mid to late February 2000, late November 2003, early January 2004, early January 2005) were primarily due to significant precipitation events that pushed the snow mass over the threshold for flooding of the ice surface. Subsequent snow ice formation reduced the snow depth as it increased the ice thickness. While flooding and snow ice formation caused relatively large reductions in snow depth, smaller reductions also occurred due to occasional strong winds that eroded the snow surface. Spatial variation of wind erosion and snow ice formation in any given year and interannual differences in wind erosion and snow ice formation account for the variations in snow depth along the transects (Table II).

Although occasional wind erosion occurs, none of the sites displayed any evidence of wind slab, which, coupled with temperature-gradient metamorphism and depth hoar formation meant that, over the course of the six winters, all density values were low (relative to those at the ALISON locations; see next section), ranging from 70 kg m^{-3} to 250 kg m^{-3} (Figure 2a–f), which are typical for this region (Sturm and Holmgren, 1998). Moderate differences in the range of density values each winter can be explained by variations in the frequency and magnitude of precipitation events and by the occurrence of flooding and snow ice formation. For example, the relatively low density of the snow during 2001–02 can be explained by the complete absence of flooding and the retention of larger amounts of low-density depth hoar than in those winters when flooding occurred and removed some depth hoar from the snow cover. The snow density–time curves at MST Pond (Figure 2a–f) have slopes and intercepts similar to those reported for snow on land in this region (Sturm and Holmgren, 1998). This suggests, for the moment, that the compaction of snow on lake ice is determined by the same factors as for snow on land, i.e. grain, bond and stratigraphic characteristics of the snow.

Table II. Summary of snow depth variation (standard deviation expressed as a percentage of the mean of measurements along the transects) at MST Pond and ALISON sites

Location	Year	Min. (%)	Max. (%)	Mean (%)
MST Pond	1999–2000	7·0	37·2	16·0
	2000–01	6·5	56·5	13·7
	2001–02	4·1	13·3	9·3
	2002–03	1·2	50·8	10·5
	2003–04	2·2	134·1	16·5
	2004–05	5·9	38·6	13·5
Imikpuk Lake, Barrow	2003–04	21·9	91·1	52·1
ACSA Pond, Nome	2003–04	17·9	350·0	117·0
Amos Lakes (Mystic)	2003–04	4·0	300·0	43·8
Aurora Pond, Fairbanks	2003–04	3·0	15·5	5·3
Lucille Lake, Wasilla	2003–04	2·3	16·3	9·2
Bear Lake, Seward	2003–04	5·6	360·0	65·4

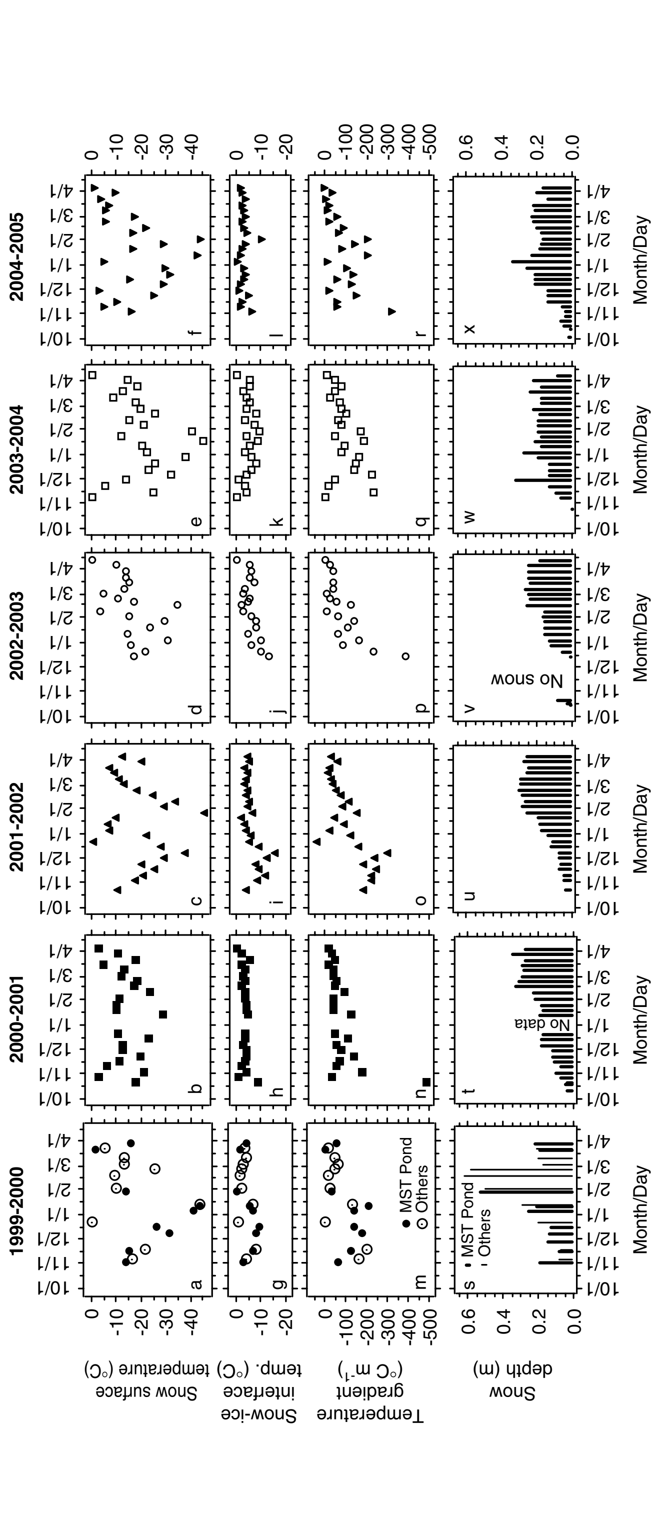

Figure 1. Snow surface temperature (a–f), snow–ice interface temperature (g–l), snow temperature gradient (m–r) and snow depth (s–x) at MST Pond during six consecutive winters, 1999–2000 through to 2004–05. The 1999–2000 data in this figure and in Figures 2 and 3 include measurements made using the same methods at a number of other ponds located within 2–5 km of MST Pond

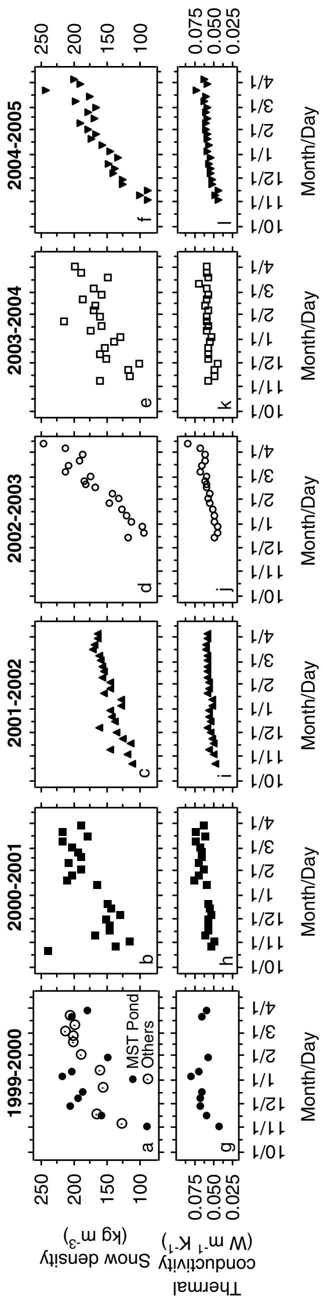

Figure 2. Snow density (a–f) and derived snow thermal conductivity (g–l) at MST Pond during six consecutive winters, 1999–2000 through to 2004–05

Because the snow density was low, the thermal conductivity of the snow was also low (Figure 2g–l). Over the course of the six winters, thermal conductivity values were in the somewhat narrow range of 0·0375 to 0·0875 W m^{-1} K^{-1}. Consequently, and in spite of the high temperature gradients, most of the conductive heat flow through the snow was in the range of 0 to −15 W m^{-2} over the course of the six winters (Figure 3). In any given winter, the heat flow values fluctuated in response primarily to air temperature differences, and heat flow differences between winters primarily reflected interannual differences in the combined influence of air temperature, snow depth and snow density (thermal conductivity). Trends to lower conductive heat flow as the winter progressed are evident in some winters (Figure 3b–d) as the combined thickness of snow and ice increased. Such trends are less obvious in the other winters (Figure 3a, e and f), probably due to greater variation in precipitation rates and air temperature.

On one occasion during three different winters there were sufficient numbers of people available to make simultaneous measurements at MST Pond and two or three nearby ponds. Since this involved K-12 educators and students, it was an interesting scientific exercise for them that also allowed the MST Pond results to be placed in a broader local context and determine whether the heat flow was similar at different locations at the same moment in time. The results (Table III) indicate that neither the snow properties nor the conductive heat flow at MST Pond were significantly different from the other locations on those occasions.

ALISON locations, winter 2003–04

The low snow surface temperatures at Fairbanks and Amos Lakes (Figure 4b) reflect the fact that, like MST Pond, they are in the cold interior of Alaska. In the northwest quadrant of the state, Nome is located on the Bering Sea coast, where the maritime influence gives higher temperatures than Barrow (Figure 4a), 830 km to the north-northwest on the Arctic Ocean coast. Winter temperatures at Barrow are similar to those in the interior (Figures 1a–f and 4b), whereas temperatures in Nome are intermediate between those of Barrow/interior and the south-central region. The latter is relatively warm (Figure 4c) because of the strong maritime influence of the North Pacific Ocean.

Barrow, Nome, Amos Lakes, Seward and Wasilla are windy locations compared with Fairbanks and MST Pond. This results in thinner, more dense snow covers than at MST Pond (Figure 4j–l, p–r). Wind erosion is particularly severe at Barrow, Nome and Wasilla; hence the greater variation of snow depth than at Fairbanks and MST Pond. The thin, high-density snow at Barrow is consistent with previous observations of snow properties on lake ice on the North Slope of Alaska (Jeffries *et al*., 1999; Sturm and Liston, 2003).

Table III. Results of simultaneous snow measurements and derived heat flow at Poker Flat ponds

	Snow depth (m)	Snow surface temperature (°C)	Snow base temperature (°C)	Snow temperature gradient (°C m^{-1})	Snow density (kg m^{-3})	Snow thermal conductivity (W m^{-1} K^{-1})	Conductive heat flow (W m^{-2})
20 March 2002							
MST Pond	0·258	−17·2	−4·2	−50·7	172	0·059	−3·0
31·6 Mile Pond	0·275	−17·9	−4·7	−50·1	171	0·059	−3·0
33·5 Mile Pond	0·256	−17·1	−5·6	−47·5	171	0·059	−2·8
34 Mile Pond	0·259	−17·4	−4·6	−52·4	177	0·060	−3·1
18 February 2003							
MST Pond	0·257	−16·5	−4·3	−47·7	183	0·061	−2·9
31·6 Mile Pond	0·322	−20·9	−3·6	−54·1	138	0·055	−3·0
33·5 Mile Pond	0·280	−17·5	−3·5	−50·1	120	0·051	−2·6
23 February 2005							
MST Pond	0·217	−6·1	−2·3	−17·2	177	0·061	−1·0
31·6 Mile Pond	0·223	−5·9	−2·3	−16·5	202	0·066	−1·1
33·5 Mile Pond	0·239	−5·6	−2·2	−14·2	177	0·061	−0·8

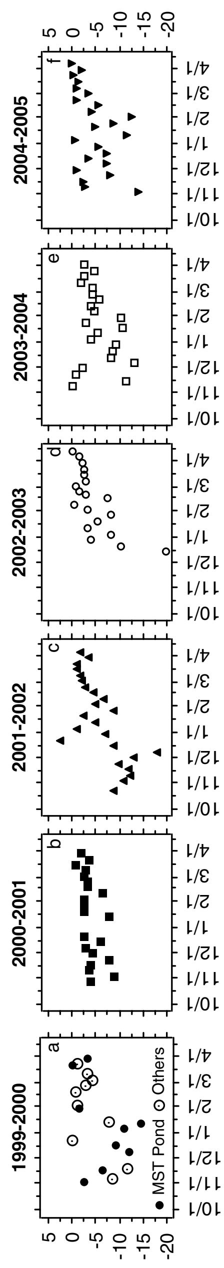

Figure 3. Derived conductive heat flow at MST Pond during six consecutive winters, 1999–2000 through to 2004–05

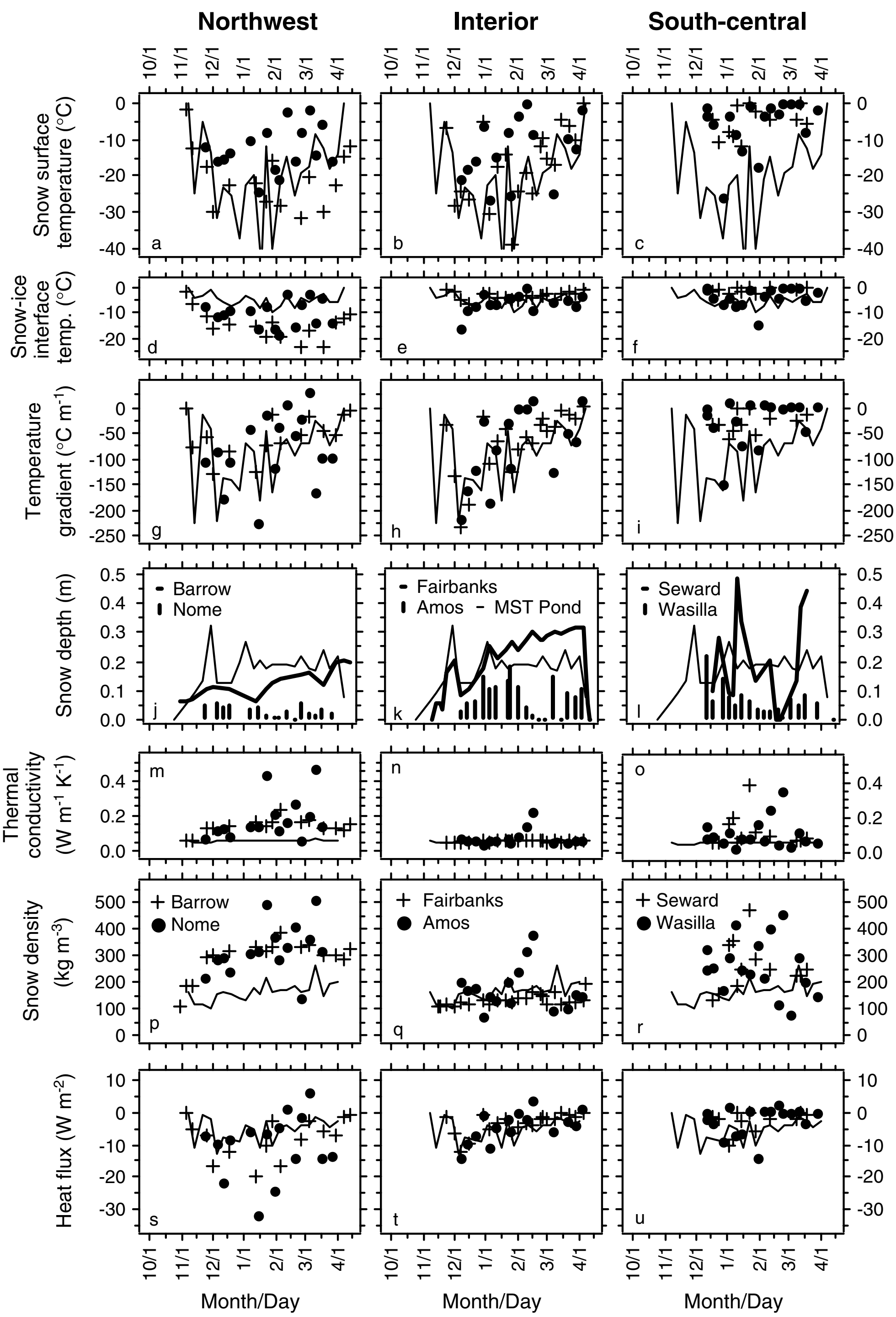

Figure 4. Snow surface temperature (a–c), snow–ice interface temperature (d–f), snow temperature gradient (g–i), snow depth (j–l), derived thermal conductivity (m–o), snow density (p–r) and derived conductive heat flow (s–u) at six ALISON locations in winter 2003–04. For comparative purposes, the data for MST Pond that winter are shown as narrow, solid lines

Snow density at Fairbanks was slightly lower than at MST Pond (Figure 4q). On the other hand, beginning in early January, Fairbanks had more snow than MST Pond, where there was more flooding and snow ice formation; thus, more snow was entrained into the ice cover. Snow depth is less variable at Fairbanks than at MST Pond (Table II) primarily because Aurora Pond is well protected from any winds.

Located close to the Alaska Range, Amos Lakes experiences chinook winds. Consequently, snow depth along the transect can be highly variable on any given day (Table II) and during the course of the winter due to a combination of wind erosion and sometimes complete melting, e.g. late February (Figure 4k and q). Seward, located on the coast, is particularly prone to large snowfall events followed by flooding and snow ice formation, melting, or both; hence the large variations in snow depth and density over the course of the winter (Figure 4l and r). Widespread flooding, snow ice formation and a consequent reduction in snow depth occurred at Wasilla in mid December, when the deepest snow of the entire winter was recorded (Figure 4l).

The high snow density values at Barrow, Nome, Seward and Wasilla mean that the thermal conductivity of the snow was two to three times higher than at Fairbanks and MST Pond (Figure 4m and o). At Barrow and Nome, the higher thermal conductivity of the thinner snow means that the ice surface was less well insulated from the cold air; consequently, snow–ice interface temperatures were lower than at Fairbanks and MST Pond (Figure 4d). At Seward and Wasilla, snow–ice interface temperatures were higher than at MST Pond (Figure 4f) because the air temperatures were higher; in this case, the higher thermal conductivity of the thinner snow means that the ice surface was less well insulated from relatively warm air.

Snow temperature gradients at Seward and Wasilla were consistently lower than those at MST Pond (Figure 4f) due to the higher snow surface and snow–ice interface temperatures. There was much greater overlap between the MST Pond snow temperature gradients and those at Barrow, Nome, Fairbanks and Amos Lakes (Figure 4g and h).

At Barrow and Nome, the higher snow density, and thus, thermal conductivity, compared with MST Pond, coupled with occasionally steeper temperature gradients, resulted in heat flow that was generally equal to and greater than that at MST Pond (Figure 4s). Although snow density and thermal conductivity were higher at Seward and Wasilla than at MST Pond, the lower temperature gradients resulted in generally lower heat flow (Figure 4u). The conductive heat flow at each of the interior locations, Fairbanks, Amos Lakes and MST Pond, were similar (Figure 4t).

DISCUSSION AND CONCLUSIONS

We have presented a comprehensive overview of the instantaneous daytime conductive heat flow through the snow cover on lake ice during six consecutive winters at a single pond, MST Pond, in central Alaska, and compared the conductive heat flow there with the conductive heat flow at six other ponds/lakes in other parts of Alaska during a single winter. Derived from measurements of the depth, temperature and density of the snow on the ice, the results are a proxy measure of the conductive heat flow due to ice formation and add substantially to our knowledge of the conductive heat flow from frozen freshwater bodies.

Each individual derived conductive heat flow value is a snapshot. That is, each value represents a brief moment on a single day, and the best temporal resolution is 7 days, the measurement frequency at MST Pond and Fairbanks. Nevertheless, for the first time there are conductive heat flow data covering the course of the entire winter at seven different locations in Alaska, unlike the Jeffries *et al.* (1999) results, which covered a 7-day period in mid April 1997 at Barrow, Alaska. Most of the conductive heat flow values are also conservative, in that few measurements coincided with the minimum daily temperature, when the heat flow would have been the daily maximum. A preliminary thermal resistance analysis of snow temperature profile data obtained every hour in winter 2003–04 at MST Pond (Gould and Jeffries, 2005)indicates that the instantaneous daytime conductive heat flow values are, on average, 78% of the daily maximum conductive heat flow (M. Gould, personal communication, September 2005).

The data reveal regional differences in conductive heat flow that reflect regional differences in the weather and its effects on the depth, temperature and density of the snow cover. There is a trend from higher to lower conductive heat flow values from the northwest to the interior to the south-central region that should have been mirrored by a trend from thicker to thinner ice in the same direction. Between the northwest and the interior regions this was certainly the case: the late winter ice thicknesses were 1·54 m and 0·95 m at Barrow and Nome respectively; at Fairbanks, Amos Lakes and MST Pond the late winter ice thicknesses were 0·63 m, 0·86 m and 0·76 m respectively. On the other hand, in south-central Alaska, Seward and Wasilla had interior-like ice thickness values of 0·87 m and 0·74 m respectively. The cause of this reverse trend is flooding and snow ice formation, which played a large role in the thickening of the ice cover in a relatively warm climate.

The snow measurements do not have the temporal resolution necessary to distinguish adequately between the conductive heat flow due to snow ice formation and that due to congelation ice formation. A more complete knowledge of freshwater ice growth as a heat source will require a better understanding of how the conductive heat flow is partitioned between the two primary ice formation processes. The need for this is exemplified by computer simulations of the possible effects of climate change on lake ice thickness and composition in central Alaska (Morris *et al*., 2005). They showed that increasing precipitation will increase the total ice thickness as the proportions of snow ice and congelation ice increase and decrease, respectively. In terms of heat transfer, this means that the total conductive heat loss due to ice formation will increase and that the largest proportion of the total conductive heat loss will be due to flooding and snow ice formation.

The temporal resolution of the snow measurements and derived conductive heat flow values are also not sufficient either to calculate the total conductive heat loss due to ice formation over the course of an entire winter, or to partition the total conductive heat loss between flooding plus snow ice formation and congelation ice formation. A thermal resistance analysis of ice temperature profile data obtained in winter 2003–04 (Gould and Jeffries, 2005) provides preliminary information on the magnitude of the total conductive heat loss (conductive heat flow integrated over time) through the combined snow and ice thickness at MST Pond. Between 24 October 2003, when measurements began in 0·15 m thick ice, and 8 April 2004, when the maximum ice thickness (0·72 m) occurred, the total conductive heat loss was 110 MJ m^{-2} due to 0·57 m of ice formation (M. Gould, personal communication, May 2005).

Total conductive heat loss values exceeding 110 MJ m^{-2} were at the high end of the range reported for a variety of sea-ice sites selected according to topographic characteristics at the Arctic Ocean drift station SHEBA (Sturm *et al*., 2001). The similarity between the lake ice and sea ice total conductive heat loss values suggests that the instantaneous conductive heat flow values would be similar. This is indeed the case, as shown by a qualitative comparison of the lake ice values with those illustrated for sea ice in the Arctic and Antarctic (Massom *et al*., 1997, 1998; Sturm *et al*., 1998, 2001, 2002). Data tabulated in Massom *et al*. (1998) allow us to show their mean and range for a more quantitative comparison with the lake ice results (Table IV). The Massom *et al*. data are for August, the Southern Hemisphere equivalent of February in Alaska. The ALISON data cover the entire winter, and it is to be expected that the conductive heat flow values would be a little lower than the Antarctic values.

The heat budget of sea ice has been the subject of intense study for over 40 years because of its vital role in heat and mass exchange between the polar atmosphere and ocean (e.g. Untersteiner, 1961). The results of this study of the conductive heat flow through snow on lake ice in Alaska suggest that the heat budget of lake ice deserves more attention that it currently attracts. This is particularly true of regions such as the North Slope of Alaska and its equivalents in Siberia, e.g. the Kolyma Lowland, where thousands of lakes cover as much as 40% of the land, and the Canadian Arctic and sub-Arctic with their great northern lakes (Great Bear Lake, Great Slave Lake, Lake Athabasca and Lake Winnipeg).

A rough calculation of the total, lake-wide conductive heat loss over the course of a winter at Great Slave Lake, for example, indicates the magnitude of the issue. The lake has an area of 28 400 km^2 and between 1960 and 1991 the mean maximum ice thickness at Back Bay (62·475 °N, 114·35 °W) was $1·33 \pm 0·19$ m

Table IV. Descriptive statistics for conductive heat flow through snow on lake ice in winter 2003–04 in Alaska and through snow on the East Antarctic pack ice (Massom *et al*., 1998)

Location	Heat flow (W m^{-2})			n
	Minimum	Maximum	Mean ± 1 SD	
East Antarctic pack ice	8·8	−51·1	−10·0 ± 10·6	39
MST Pond, 1999–2000	0·1	−14·2	−6·6 ± 5·1	9
MST Pond, 2000–01	−0·9	−8·8	−3·8 ± 2·2	21
MST Pond, 2001–02	0·0	−17·7	−5·9 ± 4·9	25
MST Pond, 2002–03	0·0	−19·5	−4·4 ± 4·6	19
MST Pond, 2003–04	−0·7	−11·15	−5·6 ± 3·5	21
MST Pond, 2004–05	0·2	−14·0	−4·7 ± 4·1	23
Barrow, 2003–04	0·0	−20·1	−7·7 ± 6·2	15
Nome, 2003–04	6·2	−31·9	−10·4 ± 10·0	15
Amos Lakes, 2003–04	3·6	−14·2	−4·3 ± 4·7	15
Fairbanks, 2003–04	0·1	−12·2	−3·5 ± 3·2	20
Seward, 2003–04	0·0	−10·3	−3·0 ± 3·4	12
Wasilla, 2003–04	1·5	−9·0	−2·6 ± 4·3	17

(C. R. Duguay, personal communication, May 2005). This corresponds to a total conductive heat loss through the ice of $\sim 257 \pm 36$ MJ m^{-2} and a lake-wide total conductive heat loss of $(7 \pm 1) \times 10^{11}$ MJ.

It has long been known that lakes affect the weather and climate of their basins and adjacent regions. Lake-effects modelling now involves the interactive coupling of a lake model to a regional climate model, with moderately successful simulation of ice thickness and the spatial evolution of the ice cover on large lakes (e.g. the Laurentian Great Lakes; Hostetler *et al*., 1993; Goyette *et al*., 2000). In view of the large conductive heat flow and total conductive heat loss through lake ice and its snow cover, as demonstrated by this paper, interactive coupled lake–regional climate models offer a means to obtain improved estimates of the total conductive heat loss from large frozen lakes, and their contribution to regional winter heat budgets. In time, as models improve and as computational cost decreases, it also ought to be possible to do this for those areas with many small lakes that cover a large proportion of the land, e.g. the North Slope of Alaska and the Kolyma Lowland.

Finally, we have presented data for Barrow, Nome, Fairbanks, Wasilla and Seward, where measurements were made by educators and students in grades ranging from 5 to 12, and for Amos Lakes, where the measurements were made by a citizen scientist. Their participation has greatly enhanced our investigation of the conductive heat flow and heat loss due to lake ice growth by allowing us to obtain data at a larger number of locations in different climate zones. The data also demonstrate that, with appropriate training and encouragement, K-12 educators and students can be scientists and participants in polar science. In doing so, they are learning about science by doing science, in this case with familiar and abundant materials, i.e. snow and ice.

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