Frost Boils and Soil Ice Content: Field Observations

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12 Abstract

Our aim is to measure and explain the seasonal changes in soil ice content in the frost
boils of Galbraith Lake, Alaska. Instruments were installed in a frost boil to monitor the
ground surface position and soil state over a period of four years. By comparing the
subside and thaw rates, we calculate the soil ice content as a function of depth.

Measured soil temperatures, liquid water contents and bulk apparent thermal
conductivities are used to estimate latent heat production and release in the soil. The
frost boils heave during freezing and settle during thaw while the surrounding tundra
heaves negligibly, but subsidizes measurably. Despite large changes in freezing rates
from year to year, total heave and its distribution across the frost boil are similar
between years. Winter air temperature and snow depth influence the freezing rate and
ice distribution as a function of depth, but not the overall heave. This suggests that
heave is controlled by water availability rather than the rate of heat removal from the
soil. Areal ground subsidence rates between 2 and 5 cm/yr are due to the disappearance
of ice at the base of the active layer, raising the possibility of ongoing thermokarst
expansion around Galbraith Lake.

Keywords: permafrost, patterned ground, segregated ice, soil thermal properties, frost
boil, non-sorted circle, freezing and thawing, climate variability, heave, subsidence
Introduction

The formation of ice in soil lies at the heart of many processes unique to permafrost regions, including differential heave, the formation of patterned ground and solifluction (Washburn 1956). This study is directed at frost boils located at Galbraith Lake, in the Brooks Range of Alaska. Walker et al. (2004) mentioned numerous mechanisms at work in the creation of the frost boils they studied in Alaska and Canada, including frost cracking, heave, mass displacement and sorting. Vliet-Lanoë (1991) reviewed the plausible mechanisms of formation of frost-boils. Of the 19 mechanisms originally proposed for patterned ground formation by Washburn (1956), two mechanisms emerged as the most likely: convection-cell like cryoturbation and load casting.

Based on the observed soil horizon morphology at Galbraith Lake, Ping et al. (2003) suggested that their genesis was probably linked to a single extrusion of lower soil horizons through structural weaknesses in the overlying horizons, for example through frost cracks. This is consistent with the lack of organic material in lower horizons at the frost boil boundaries, which would have indicated the convection-cell like movement proposed elsewhere (MacKay 1979). The frost boil center, on the other hand, is free of vegetation, but one may find roots that have been exposed at the surface and stretched by soil movement. Both facts suggest that radially outward spreading of the exposed mineral material is actively occurring. The role of differential frost heave as a result of segregated ice formation in creating and maintaining these features is unclear and what additional processes are involved is also unclear (Vliet-Lanoë 1991).

Measuring changes in soil ice content is not trivial, even by destructive sampling methods. It is further complicated by the fact that ice segregation during freezing of the active layer and upper permafrost can lead to highly variable spatial concentrations of
ice at a variety of scales. Methods that have been used for ice content include: gravimetric; radar, either as ground-based or remote techniques; acoustics; neutron absorption for the determination of total water content; dielectric methods (Bittelli et al. 2003) and volumetric methods. We use the changes in soil volume, temperature and thermal properties during freezing and thawing to draw conclusions about changes in ice volume. Our objectives are to describe the heave dynamics of a frost-boil, to quantify the role of seasonal ice growth in these features and to evaluate the role of ice formation and segregation in differential frost heave at this site.

Methods

Site

Galbraith Lake is about 3 km long. It is located at the west end of Atigun Gorge on the north side of the Brooks Range, Alaska (68°28.890’N, 149°28.744’W). Water enters the lake through a tributary at its northern end and through an alluvial fan that forms its western shore. It drains into the Saganivirktok River, which flows across the North Slope of Alaska into the Arctic Ocean. Aerial photographs show evidence of a series of old shorelines beyond the northern end of the lake. Within this region of exposed and reworked lake bed, a mixture of lacustrine and fluvial sediments form a nearly horizontal plain between outwash fans of streams into the valley from the west, and bluffs on the east. At present, there is an airstrip located on the west side of the plain, and the Dalton highway and the Trans-Alaska Pipeline run north to south along the east side of the valley. The site is poorly drained and the water table is within 20 cm of the ground surface during the growing season. Land-cover type is classified as moist non-acidic tundra, the soil pedon is classified as a cryaquept and the soil horizons are
contorted by cryoturbation (Ping et al. 2003). From 1987 until 2001, mean annual
ground temperature ranged between -8 and -4 °C at the permafrost table and between -
5.7 and -4.7 °C at 20 m (Osterkamp 2003).

Visual and near infrared band images of the site were collected using the Bi-camera
Observation Blimp (BOB) low-altitude aerial photography system (Harris and
Helfferich 2005). Ground features were used to georectify the images, which have a
resolution of better than 0.07 m. The site is shown (Figure 1) in a normalized difference
image (NDVI) of the reflected near infrared and red bands. NDVI is an index for
reflection from photosynthesizing vegetation. It is used here to identify vegetation-free
patches of ground, which appear black. The only features devoid of vegetation in the
field of view are the frost boils, water surfaces and some humans, which appear darker
than the vegetated surfaces (Figure 1).

**Frost boil morphology**

Frost boils at Galbraith Lake are identified at the ground surface as roughly circular
areas free of vegetation and between 0.5 and 1.5 m in diameter. In some cases, several
frost boils appear to have developed close enough to one another to form linear regions
of frost boil up to 3 m long. The vegetation cover of the surrounding tundra is complete,
and the frost boils thus have well-defined boundaries (Figure 1, inset). Based on the 110
frost boils identified in Figure 1 (C), the mean distance between frost boils is 5.6 ± 0.6
m. They are not uniformly distributed, as the terrain is also the site of other periglacial
features, including ice-cored mounds (A), thermokarst ponds (B) and ice wedges (D),
that are found throughout the image. Dark spots within the oval (E) are people. Frost
boils are also found in the surrounding tundra, outside of the old lake bed region, but
tend to be covered with vegetation to some degree, and are smaller with less distinct
boundaries.

Galbraith Lake frost boils appear to have undergone sorting, with fine silty-clay mineral
material in the frost boil center and sandy soil underlying the vegetated margins and
surrounding tundra of the boils. The surfaces of the frost boils are cracked in a columnar
structure, and a crust has developed. Major ion analyses of a crust sample showed little
difference between the crust and the soil beneath it (Ping et al. 2003). They are slightly
domed, and their elevation relative to the surrounding tundra varies with the time of
year, which is examined in the section on frost heave, below.

A soil pit roughly bisecting a frost boil was carefully excavated on August 25, 2001.
The soil pit was approximately 1.4 m wide at the ground surface. Soil was removed by
horizon with consideration for the three-dimensional structure of the horizons, and
replaced and compacted so that the pit was refilled. Beneath the center of the frost boil
there is little change in soil composition with depth. All sampled horizons were mineral
with organic contents of less than 3% by weight, had bulk densities in the range of 1.62
to 1.76 g cm$^{-3}$ and saturated soil-paste pHs of 8 to 8.2 (Table 1). Little difference was
observed in soil texture variation with depth, down to 0.73 m. Soils surrounding the
frost boil have organic upper horizons 0.05 to 0.20 m thick with lower bulk density,
higher carbon content, and support a 100% cover of mosses and sedges. The soil pit was
excavated to a depth of 1.3 m and the frost table beneath the center of the frost boil was
encountered at 0.73 m. A jackhammer was used to excavate below the frost table. The
thaw depth in the frost boil exceeded the thaw depth immediately adjacent to and at a
distance of 2.5 m from the boil by over 0.22 m, creating a bowl-shaped depression in
the frost table beneath the frost boil. Many ice lenses were encountered between 0.75 to
0.83 m, running parallel to the frost table and containing ice lenses up to 7 mm thick. They were angled upwards towards the margin of the frost boil and not found beneath the adjacent tundra. Below 0.83 m, ice layers with a blocky, horizontally-oriented structure and 0.02 to 0.05 m thickness were separated by thinner soil layers. This ice was associated with the transition zone, the layer of soil that is usually permafrost, but is subject to infrequent thaw during warm years. A core was drilled into the bottom of the soil pit to determine the lower extent of the ice-rich deposits: ice layers were encountered down to the termination depth of 1.6 m. Thus, the frost boil was thawed to a depth of 0.73 m, with thin ice lens for 0.10 m below the frost table, and thicker ice layers underlying these lenses for at least another 0.77 m. The estimated volumetric ice content of the soil below 0.83 m was over 0.90 m³ per cubic meter of soil.

The frost boil morphology and material grain-size results in a classification after Zoltai and Tarnocai (1981) as raised-center mudboil, despite the poor drainage of the site. Ping et al. (2003) concluded that the soil profile morphology showed upward extrusion of clayey material from a lower soil horizon into an overlying sandy horizon, locally displacing the upper horizon and creating a textural differentiation at the ground surface.

**Instruments**

Temperature, volumetric liquid water content and thermal conductivity sensors were installed in the soil pit wall between the surface of the soil and 1.36 m depth (Figure 2). Temperatures were measured hourly using a vertical array of thermistors. The thermistors were calibrated using a de-ionized water-ice mixture, from which a thermistor-specific offset, \( \delta_r \), for the Steinhardt-Hart equation was generated:

\[
T^{-1} = 1.28 \times 10^{-3} + 2.37 \times 10^{-4} (\ln(R_r - \delta R)) + 9.06 \times 10^{-8} (\ln(R_r - \delta R))^2
\]
where $R_T$ is the measured resistance and $\delta R_0$ is the resistance offset at 0 °C. The
temperature sensor uncertainty is better than $\pm 0.01$ °C at 0 °C.

Time domain reflectometry sensors (Campbell Scientific CS605) connected to a
reflectometer and datalogger (Campbell Scientific TDR100 and CR10X) were used to
record the apparent sensor length at one-hour intervals, which is related to soil moisture
content. In addition, the entire reflected waveforms for each sensor were measured once
per week to verify the TDR100 waveform analysis algorithm, and to provide an
alternate time series. Occasional TDR100 waveform analysis algorithm failure occurred
as a result of temperature-induced changes in the sensor cable length. This problem was
solved by trial and error determination of two cable lengths in the datalogger program,
one for thawed soil and one for frozen soils. The sensors had been previously calibrated
in the laboratory using water and air as reference materials following the method of
Heimovaara et al. (1995). The bulk relative dielectric permittivity data was used to
calculate liquid water content using Roth et al.’s (1990) mixing model for mineral
horizons and an empirical relationship for organic horizons similar to the ones under
study (Overduin et al. 2005). The accuracy of the method is estimated to be better than
0.05 m$^3$ m$^{-3}$ (Heimovaara et al. 1995).

Thermal diffusivity is calculated as the ratio of thermal conductivity and heat capacity:

$$D = \frac{k}{\rho C}$$  \hspace{1cm} (2)

where $D$, $k$, $\rho$ and $C$ are the bulk thermal diffusivity, conductivity, density and
volumetric heat capacities of the soil, respectively. Thermal conductivity was measured
using Hukseflux TP01 transient heat pulse sensor installed at 0.1 and 0.2 m depth
beneath the center of the frost boil. The radial temperature difference following heating
for 180 s was measured and analyzed following Overduin et al. (2006). All data points were included in the analysis, including peaks in bulk apparent thermal conductivity due to changes in phase induced by the transient heat pulse. The uncertainty in $k$ is estimated to be ±0.01 W m$^{-1}$ K$^{-1}$ over the range of 0.3 to 4.0 W m$^{-1}$ K$^{-1}$ (Overduin et al. 2006).

**Thaw and frost penetration depth**

We used the decrease in liquid water content at each TDR sensor to identify the time at which the freezing front passed each sensor. Phase change was identified as the point of inflection in the time series of soil liquid water content. To interpolate thaw and frost front positions between sensor depths, the data were fit to a simplified solution of the Stefan problem:

$$z = A\sqrt{\sum (\bar{T} - T_f)} + B$$

(3)

where the sum indicates summation over days of the year, $\bar{T}$ is the mean daily air temperature, $T_f$ is the freezing point and $A$ and $B$ are constants. The term within the square root is referred to as the degree-days of freezing or thawing (Yershov 1990). The coefficients $A$ and $B$ are specific to a freeze or thaw season, and were used to predict thaw based on cumulative degree-days of thawing.

**Frost heave**

By comparing soil volumes in the thawed and frozen states, changes in the volumetric fractions of water and ice may be inferred. Heave was measured in two ways. For the excavated site, fiberglass rods 1.22 m long and 3 mm in diameter were anchored in the permafrost at depths between 0.7 and 1 m using 5 cm long cross-pieces fastened using screws. The vertical rods extended above the ground surface. The rods were distributed in a line from the approximate center of the frost boil to a point 2.47 m away (Figure 2).
Metal washers were free to slide along the rods and rested on the ground surface. The rod length exposed above the metal washer was measured at irregular intervals throughout the year. At a second frost boil a few meters from the first, five ultrasonic distance sensors (Sonic Ranger SR50, Campbell Scientific Corporation) were mounted on a horizontal rod supported by two vertical members, which were anchored in the permafrost at a depth of over 1.5 m and encased in tubing that extended over the active layer depth. The sensors have a resolution of 0.1 mm and an accuracy of +/-0.01 m, after correction for variations in air temperature. The sensors record the signal from the nearest target in their approximately 22° field of view. They are thus affected by the presence of snow and vegetation. The hourly values used here are smoothed using a running average during snow-free periods only.

Following suggestions by Burn (1998), we define the active layer depth as the penetration depth of the 0 °C isotherm. To calculate the rate of thawing in spring, the position of the 0 °C isotherm from 0.12 to 1.36 m was interpolated as described in the previous section. We assume that daily subsidence at the surface is due to daily ice melting at the thaw front. Although uncertainty in settling rates at a daily time scale is high and probably depends on flow-rates of excess moisture from the thawing front: the correspondence between the rates of subsidence and thaw suggests that our assumption is reasonable. Measurable subsidence did not continue after the active layer reached its maximum depth at the end of summer.

We define the vertical positions of the ground surface $z_0(t)$, and the phase change boundary, $z_{ft}(t)$, relative to the ground surface at the start of thawing or freezing. Settling is ascribed to the melting of ice in the soil to the depth $z_{ft}$. Accounting for the difference
in ice and water densities, the segregated ice content averaged over the entire profile
from $z_0$ to $z_f$ before thawing is then given by:

$$\theta_i^* = \frac{\Delta z_0}{\Delta z_f} - \theta_w \frac{\rho_w}{\rho_i} \frac{\Delta z_f - \Delta z_0}{\Delta z_f}$$  \hspace{1cm} (4)$$

where $\rho_i = 917.0$ kg m$^{-3}$ and $\rho_w = 999.8$ kg m$^{-3}$ are the densities of water and ice,
respectively, and $\Delta z_0$ and $\Delta z_f$ are the change in the ground surface and the phase change
boundary positions.

**Volumetric latent heat production**

Changes in ice content can also be inferred from latent heat production and
consumption in the frozen ground (Roth and Boike 2001). Conservation of energy and
Fourier’s law lead to the heat diffusion equation in one dimension:

$$\frac{\partial}{\partial t}(\rho CT) = r_h + \frac{\partial}{\partial z} \left( k \frac{\partial T}{\partial z} \right)$$  \hspace{1cm} (5)$$

where $r_h$ is the heat production or consumption per unit volume. We use the temperature
profile beneath the frost boil center to estimate the heat production or consumption in
the soil by integrating over each soil layer between sensors at depths $z_j$ and $z_{j+1}$, and
between observation time steps $t_k$ and $t_{k+1}$, where $j$ and $k$ are ordination subscripts
denoting steps in observation time or depth. Integrals are approximated using the
trapezoidal rule and the partial derivative of temperature is estimated using a central
finite difference. The approach of Roth and Boike (2001) is followed, except that a time
dependent $C$ is calculated using equation 6. The bulk soil heat capacity depends
primarily on the soil’s volumetric liquid water content and is calculated from measured
soil composition. The dependence of the bulk heat capacity on composition is the
volumetric fraction-weighted arithmetic mean of the soil components:
\[ \rho C = \sum_{n=a,i,w,s} \rho_n C_n \theta_n \]  

where \( \rho \) is the soil bulk density, and \( \rho_n \) is the density, \( C_n \) the heat capacity, and \( \theta_n \) the volumetric fraction of soil component \( n \). The subscripts refer to air (\( a \)), ice (\( i \)), water (\( w \)) and soil (\( s \)). The volumetric heat capacities of ice, water and soil used were 1.93, 4.18, and 1.88 MJ m\(^{-3}\) K\(^{-1}\), respectively, while that of air was neglected. The volumetric liquid water content was taken from the TDR measurements and the volume fraction of soil was taken from the porosity measurements in Table 1. Ice content was estimated as the difference between measured liquid water content and the soil porosity, which is equivalent to the assuming saturation and that moisture migration does not take place. This coarse assumption contradicts the clearly observed effects of water migration but is only used in the estimation of soil heat capacity. Since the relative dielectric permittivity of ice is low and similar to that of the soil matrix, the liquid water content is only linearly and weakly dependent on ice content. A 10% change in the ice content results in a volumetric liquid water content change of less than 0.8% for the entire period of record, well below the estimated uncertainty in water content. The heat capacity is linearly dependent on estimates in soil component fractions. We estimate its uncertainty, \( \delta C \) as the sum of the uncertainties in soil fractional heat capacities:

\[ \delta C = \sum_{n=a,i,w,s} \delta C_n \]  

where the uncertainty introduced by the exclusion of air is ignored. For the unfrozen soil, \( \delta C \) is equivalent to the sum of the products of the volumetric heat capacities of the soil components and their uncertainties (\( \delta \theta_i = \pm 0, \delta \theta_w = \pm 0.05, \delta \theta_s = \pm 0.05 \)). At subzero temperatures, however, ice may segregate. Since we do not measure soil ice content directly, our uncertainty for bulk volumetric soil heat capacity in the frozen soil
increases by a factor of almost 10 and reaches values comparable to the heat capacity of bulk frozen saturated soils (1.6 to 2 MJ m$^{-3}K^{-1}$; Yershov 1990), based on uncertainties in soil component volumetric fractions of $\delta \theta_i = \pm 1.0$, $\delta \theta_w = \pm 0.05$, $\delta \theta_s = \pm 0.35$, where the latter value is (1-$\eta$), where $\eta$ is the porosity. However, this uncertainty is reached only if ice segregation results in the complete displacement of liquid water and soil particles from the soil. The freezing characteristic curves obtained by the TDR sensors and thermistors for these soils show that liquid water exists throughout the measurement period so that soil particles cannot have been displaced, at least at the scale of measurement of the TDR sensor.

Values of bulk apparent soil thermal conductivity are calculated using the least squares fit of measured thermal conductivity values against soil temperature (for temperatures below $-1.5$ °C) shown in Figure 3, which includes both cooling and warming data over a three year period. Calculations of latent heat are restricted to times and depths for which the soil temperature is below $-1.5$ °C to avoid the influence of latent heat changes on measured thermal conductivity values. Fuchs et al. (1978) showed theoretically that the effects of phase change on the apparent thermal conductivity are limited to a well-defined temperature range (for example, -0.5 to 0 °C for a Palouse loam). The lower limit of this temperature dependency is a function primarily of total soil water content.

Results and Discussions

Rates of freezing

Soil temperature and volumetric liquid water content for four hydrologic years (2001/02, 2002/03, 2003/04 and 2004/05) show differences between the four winter
periods (Figure 4). Following, we describe the observed differences in freezing rate between 2002 and the other years in a variety of ways. During 2001, 2003 and 2004 the active layer rapidly changed from positive temperatures to values less than −0.1 °C throughout the soil profile. In these three years the lowering of the zero curtain (i.e. the time taken to bring the entire soil profile to temperatures less than or equal to −0.1 °C) from the shallowest sensor at 0.12 m depth took less than a week (September 19-21, 2001, September 9-11, 2003 and September 12-15, 2004). Between 0.12 and 0.77 m, the mean freezing rates (calculated using the position of the 0 °C isotherm) were 4.8 x 10\(^{-1}\), 1.3 x 10\(^{-2}\), 3.5 x 10\(^{-1}\) and 3.1 x 10\(^{-1}\) m d\(^{-1}\) in 2001, 2002, 2003 and 2004, respectively. Although the freezing rate was more rapid in 2001 than in 2003 or 2004, it occurred ten days earlier than in 2003. The freezing rate during the fall of 2002 was more than an order of magnitude lower than in the preceding and following years and the active layer at 0.77 m depth remained unfrozen until after November 20. Even if the −0.5 °C isotherm is used to indicate freezing, freezing was delayed by over a month in at this depth in 2002, compared to other years.

We attribute the delay in freezing to changes in snow cover and winter air temperatures. On Oct 1, 2002, over 0.16 m of snow accumulated. As a result of subsequent snowfalls, the snow pack grew to almost 0.5 m thickness without diminishing over the course of the winter. During the following winter, snow fell between Sep 6 and Sep 26, 2003, but had disappeared from the frost boil surface by October 1, 2003. Subsequent snow covers that exceeded 0.02 m lasted for less than 20 days. The effect of snow cover on soil temperatures is not straightforward. Sokratov et al. (2001) used a 30-year data set to show that the ratio between mean air temperatures and soil temperatures at 0.2 m depth are not correlated with snow depth except during February and March. Taras et al.
(2002) distinguished snow depths sufficient to decouple air and soil temperatures from one another (> 80 cm) from shallow snow depths (< 25 cm) in which air and soil temperatures are highly coupled. Intermediate snow depths were strongly coupled. They found that a 15 cm increase in snow depth produced a 0.5 to 3.0 °C increase in soil temperature at the snow-soil interface. Under this scheme, the snow depths observed in 2002/03 and 2003/04 correspond to intermediate and shallow snow depths, respectively.

Air temperatures in 2003/04 were also lower than in other years. Mean monthly air temperatures for September 2001 to August 2005 are presented in Table 2. The winter months from December to April were all colder during 2003/04 than during other years; February and March, in particular, were over 8 °C colder. Monthly mean soil temperatures were colder by 5 to 11 °C for most of the winter of 2003/04 than the previous winter. The mean monthly temperatures during freeze-up were also warmest in 2003/03. While the effect of a change in air temperature on soil temperatures under variable snow covers is not straightforward, Stieglitz et al. (2003) demonstrated the equivalent importance of air temperature and snow cover variability in determining ground temperatures. It seems reasonable that air temperature and snow cover suffice to explain differences in freezing rates between years. Since freezing rates in 2001/02, 2003/04 and 2004/05 are similar in magnitude, freezing in 2002/03 appears to have been anomalously slow for this location.

Despite microclimatic changes from year to year, freeze-back within the frost boil is generally rapid compared to other tundra locations, especially those protected with an organic mat (e.g. Hinzman et al. 1998). Mean monthly winter air temperatures for each of the years 2001/02, 2002/03 and 2004/05 fall within a few degrees of each other. Thus, based on soil temperatures during 2002/03, the effect of an intermediate snow
cover, under normal winter conditions, is to delay freeze-back by at least one month at most depths. The combined effects of the exposed frost boil mineral soil, shallow snow depth and cold winter temperatures in 2003/04 contribute to the extremely cold soil temperatures in the frost boil.

**Thaw and Subsidence**

For the springs of 2003, 2004 and 2005, interpolated thaw depth data are shown in Figure 5, along with air temperature, the timing of phase change at each TDR sensor and the position of the ground surface. The ground surface and interpolated thaw depth positions are shown, the former relative to its pre-thaw position. In all three years, the correlation coefficient between the Stefan solution and the observed phase change was better than $r = 0.99$, with coefficients $\{A, B\}$ in equation 3 of $\{-3.67 \times 10^{-2}, 0.87 \times 10^{-2}\}$, $\{-2.98 \times 10^{-2}, -4.59 \times 10^{-2}\}$ and $\{-3.09 \times 10^{-2}, -7.27 \times 10^{-2}\}$ for 2003, 2004 and 2005, respectively. During the springs of 2004 and 2005, the profile thawed down to 0.12 m by May 15th. In 2003, however, thaw reaches this depth more than 2 weeks later, a reflection of differences in degree-days of thawing, which by July 18 totalled 687 °C·d in 2004 but only 458 °C·d in 2003, with thaw depths of 0.88 and 0.78 m, respectively (Figure 5). The spring subsidence in each year was similar in magnitude and reached 2.09 $\times$ 10$^{-2}$, 2.24 $\times$ 10$^{-2}$ and 4.17 $\times$ 10$^{-2}$ m by July 18 in 2003, 2004 and 2005, respectively (for 458, 687, 527 degree-days of thawing). Nonetheless, the amount of subsidence occurring over the first 30 cm of thaw represented 68, 19 and 44% of the end of summer total subsidence in 2003, 2004 and 2005, respectively, suggesting significant differences in the distribution of segregated ice from year to year, at least for the upper soil horizons. Egginton (1979) recorded subsidence of two mud boils west of Hudson Bay, and found that over 80% of total summer subsidence (6 to 9 cm) occurred.
within the first month after thawing began when the active layer was about 40 cm deep, in agreement with MacKay and MacKay’s (1976) observed ice lensing distribution in non-sorted circles and with MacKay’s (1980) model of formation for non-sorted circles. Time series of surface position measured with both heave rods and ultrasonic distance sensors between the fall 2001 and spring 2005 show annual subsidence. As a group, the surface position measurements show a steady decrease in surface elevation (Figure 6). Ground surface position relative to the heave rods lowered from 5.1 cm/yr at a point 2.47 m from the frost boil center to 7.5 cm/yr at the frost boil center. Subsidence and/or frost-jacking of the rods would result in lowering of the ground surface. In addition, soil pit excavation and refilling at the end of the summer of 2001 was probably followed by settling of the ground, leading to higher measured lowering. Results from an undisturbed frost boil corroborate the observed subsidence. The ultrasonic distance sensors were mounted on permafrost anchors sheathed in PVC tubing to prevent frost-jacking, and the frost boil was not excavated. Using ground surface positions at the end of summer from these sensors, mean subsidence is 5.1 cm/yr at the frost boil, and ranges from 2.0 to 4.8 cm/yr for the other sensors for the period from fall 2003 to fall 2005. The subsidence observed here affects both the frost boil and the surrounding tundra, and does not directly affect active layer depth. Since winter heave is similar from year to year and mostly due to the segregation of ice in the upper 30 cm of the soil profile, the subsidence observed is attributable to the melting of ice at the base of the active layer. It is more pronounced beneath the frost boil, where summer thaw reaches greater depths, than beneath the surrounding tundra. We observed 0.77 m of ice-rich soil (estimated >90% by volume) beneath the active layer.
The potential exists for continued subsidence and the northward expansion of Galbraith Lake through thermokarsting.

Heave

Figure 7 shows three net winter heave profiles, measured between the end of summer (late August - early September) and early spring (mid-May), superimposed on the microtopography of the frost boil. Net winter heave increases towards a point just outside the boil’s edge from close to zero at a distance of 1.6 m. The maximum heave of around 0.12 m occurs outside the frost boil boundary in all three years. The unvegetated frost boil heaves between 0.04 to 0.08 m. The three heave profiles are remarkably similar, given the differences observed in freezing rates for each fall season.

Usually, a few days after the ground begins to freeze from the surface downward, soil temperatures between the upper and lower frost fronts reach slightly negative temperatures (Osterkamp and Romanovsky 1997). In the high snow year, however, this process is extended over a longer time period. Nonetheless, the depth of the freezing front as a function of lateral position is similar from year to year (Figure 7). The 0 °C isotherms, recorded at 9:00 am on the indicated dates, all show a thicker frozen layer beneath the frost boil (10 – 30 cm) than beneath the tundra (10 – 15 cm). Segregation of ice at the frost front results in heave normal to the frost front. The freezing front beneath the margin of the frost boil is inclined so that a normal vector to the freezing front will have a lateral component pointing towards the center of the boil.

We explain the location of maximum net winter heave outside the frost boil in one of two ways: either (i) the shape of the freezing front and the relative freezing rates beneath the boil and its margin lead to greater heave at the frost boil’s margin than at its center, or (ii) maximum heave does occur at the frost boil center at some time during the
winter but subsequent subsidence beneath the frost boil occurs prior to the end of snow melt (during the winter). A single winter measurement of frost heave made in November 2002 was 0.02 m higher than in the following pre-thaw May, suggesting that sublimation or evaporation processes from the bare soil of the boil’s apex could have decreased the ice content over the winter. The microtopography, surficial cracking and heave of the boil would contribute to these processes. The frost boils at Galbraith are elevated above the surrounding tundra and the high winds in the Atigun Valley can blow the snow off the frost boils alone. Under these conditions the surfaces of the boils are observed to be desiccated. When the snow pack is sufficient to cover the frost boils, there is no indication of the location of the boils at the snow surface. The snow pack is necessarily thinner above the frost boils than over the tundra, and moisture migration into the snow pack is exacerbated at the boil by a steeper temperature gradient through the snow pack. Since the soil here is at or near saturation on freezing, there can be no indication for this process using volumetric liquid water content, since it is independent of total water content at cold temperatures. Since thaw begins with the infiltration of melt water, the soil is saturated before all of the ice disappears in spring. Such a loss of ice to the atmosphere is consistent with the shape of the heave curves in Figure 7. The amount of heave depends on the freezing rate, and the dramatic difference in freezing rate between 2003 and the other years lead us to expect differences in heave. For example, O’Neill and Miller (1985) simulated heave and observed an increase in ice segregation with a decrease in freezing rate. To explain the similarity between years, we must either invoke processes that remove ice in late winter after a slow freeze-back, or the possibility that heave is only weakly dependent on the freezing rate. The correspondence between the rate of water flow to the freezing front and the driving
gradient (temperature) should enhance heave, so that we find the latter option difficult

to accept. We examine differences between the years in ice content below.

Ice content as a function of depth

The timing of subsidence and the temperature profile during thawing are related to the
distribution of ice content as a function of active layer depth. Figure 8 shows the total
volumetric ice content as a function of depth based on cumulative settlement and thaw
in the springs of 2003, 2004 and 2005. The years refer to the spring in which ice
content was observed; in each case, most of this ice probably formed during fall freeze-
up in the previous year. Ice contents throughout the profile lay between 0.5 and 0.8,
with lower ice content in 2003 in the upper 0.4 m of the profile than in subsequent
years. Ice content exceeded the porosity in 2004 and 2005 over this same interval,
suggesting that more segregation occurred in the upper profile during freezing in 2003
and 2004 than during the fall of 2002 (corresponding to the graph for 2003). The
integrated ice contents between 0.12 and 0.40 m (for which depths data are available in
all three years) are 0.16 (2003), 0.19 (2004) and 0.19 m (2005). Heave is the same (+/- 2
cm) each year, but subsidence varies as much as 7 cm from year to year, suggesting that
ice lost in deep thaw years is not replaced by segregation during freezing.

In all three years, freezing occurs via both the descent of a freezing front from the
surface and an increase in the height of the frost table, although the latter process is
much more pronounced in 2002/2003 (Figure 4). The last depth to freeze in 2003/2004
is 0.79 m. Ice content is locally higher between 0.85 and 0.90 m in both years. The
qualitative similarity between increases and decreases in ice content with depth suggest
that the relative distribution of ice from year to year is stable, except for the upper
portion of the active layer, which is subject to repeated diurnal freeze-thaw cycles
during freezing (Overduin et al. 2003). The spikes in ice content below 0.75 m
correspond with the segregated ice observed in the field during excavation. Based on its
position, this ice is probably segregated during upward movement of the frost table
during freeze-back. Based on position, the inter-annual variability in shallow active
layer ice content is determined by snow pack and heat transfer conditions at the surface,
which affect post-freeze redistribution of water, rather than freezing rate. Deeper ice
content does not seem to be freezing-rate dependent, and is therefore less variable from
year to year.

Volumetric latent heat production

Latent heat changes (positive or negative) for the temperature profile beneath the frost
boil for the 2002/03 and 2003/04 winter periods, when the entire profile is at or below
-1.5 °C, are presented in Figure 9. Since the soil is at or close to saturation on freezing,
changes in latent heat during this period are probably the result of freezing and thawing
in the frozen soil rather than vapour transport processes. Although it is probable that
other heat releasing or consuming processes are at work, such as sublimation near the
ground surface or vapour transport, the scale is given in W m$^{-3}$ and also in the rate of
phase change of water (kg m$^{-3}$ d$^{-1}$), using 333 kJ kg$^{-1}$ as the latent heat of phase change.
Lighter colours indicate the consumption of latent heat. Our goal is to identify depths
and times of ice formation or disappearance during the winter.

In both winters, latent heat release dominates the profile between 0.4 and 0.9 m from
freeze-up until the end of April. This stands in stark contrast to the observations of
Roth and Boike (2001) for a site on Spitsbergen, who showed strong heat consumption
at intermediate depths during the cold winter period. They regarded this heat
consumption as evidence for evaporation of water and its diffusion upwards out of the
soil into the snow pack or atmosphere. Corroboration of this vapour diffusion using changes in soil water content was not possible, since soil liquid water content does not change appreciably as water is removed. Net changes in saturation over the winter are masked by the infiltration and refreezing of melt water prior to soil thawing in the spring, which results in near-saturation of the soil at thaw. In addition to the fact that their analysis included constant soil thermal properties, ice segregation in the Galbraith Lake frost boils presumably draws moisture laterally which continues to freeze well-after freezing begins. The integrated net latent heat fluxes between Jan 1 and March 31 for the years 2002/03 and 2003/04 are 28.2 and 23.6 MJ m$^{-2}$, corresponding to water amounts of 85 and 71 kg m$^{-3}$. These estimates account for the depth intervals of 0.31 to 1.1 m. Given that the upper and lower portions of this interval tend to lower values, net latent heat consumption probably occurs in the overlying and underlying soil horizons. These estimates can therefore only be regarded as upper bounds on the amount of ice created for the whole profile. In the slow freeze-up year (2003/04) more latent heat is released in the intermediate depths 0.5 to 0.98 m than in the previous year. This is more than balanced out by latent heat consumed in the upper part of the profile, consistent with greater moisture losses through evaporation and sublimation there, so that less latent heat is produced in 2003/04 than in 2002/03. The consumption of latent heat that occurs in the lower part of the horizon is difficult to explain, but may be related to the transport of moisture upward in the profile in response to the temperature gradient and suggests the steady loss of ice at depths where the soil composition is more than 90%. These losses are observable as subsidence and represent a decrease in total permafrost thickness, but will not be observable using active layer probing or by observing ice content profiles that do not extend through the ice-rich layers.
Shallower depths are affected by higher amplitudes of latent heat values, with episodes of heat consumption, probably as a result of ice sublimation from the surface. In both years, latent heat consumption deepens towards spring, as the soil is warmed from above. The influence of winter temperature fluctuations is seen throughout the profile, but is muted in the warmer winter (2002/03) compared to the following cold winter. The latent heat release at depths below 0.9 m suggests a steady decrease in ice content of the soil throughout the winter, consistent with heave amounts lower than those predicted by the latent heat release at intermediate depths, and perhaps consistent with the observed annual subsidence.

Conclusions

As a result of changes in ice content during freezing, frost boils at Galbraith Lake in the Brooks Range of Alaska heave during freezing. Freezing rates vary greatly interannually, but heave remains similar from year to year, both in terms of magnitude and distribution across the frost boil. Enough latent heat is released in intermediate depths in the frost boil to account for winter heave. The low net winter heave levels at the center of the frost boil are attributed to loss of moisture via evaporation or sublimation through the exposed mineral soil of the frost boil during the winter. We suggest that the winter removal of water from shallow soil horizons at the frost boil’s center results in low lateral displacement of soil, leading to extrusion of lower horizons but to insignificant downward movement of soil material at the frost boil margins. Settling of the ground surface and the development of the active layer together permit estimation of the pre-thaw vertical distribution of ice in the active layer. Net annual subsidence suggests that melting ice at the base of the active layer is not replaced by
annual winter heave, which results from ice formation in the upper portion of the frost boil.

On the basis of surface position measurements during heave and subsidence, soil state and composition measurements, and latent heat flux calculations, we observe small interannual variability in heave magnitude for a frost boil, despite large changes in freezing rate, snow cover and air temperature. More latent heat is released within the soil during the winter than is necessary to account for the observed heave. It is compensated for by processes consuming latent heat in the near-surface soil and below the active layer.
Acknowledgements

Prof. Bernard Hallet and an anonymous reviewer greatly improved this manuscript with their insightful critiques. We are grateful to Jens Ibendorf, Prof. John Kimble, Gary Michaelson, Prof. Chien-Lu Ping, and Joerg Sommer for help in the field. Soil analyses were performed by the Soil Survey Staff at the National Soil Survey Center, USDA. Funding was provided by an Inland Northwest Research Alliance (INRA) fellowship and the National Science Foundation’s Office of Polar Programs (OPP-9814984) through the Office of Polar Programs, Arctic System Science. The Potsdam Institute for Climate Impact Research graciously provided an office for P. P. Overduin during a portion of the writing.
References


Figure 1. A normalized difference image (NDVI) of the study site using near-infrared and visible red wavelengths shows the distribution of periglacial landforms at the study site. NDVI is an index for reflection from photosynthesizing vegetation. It is used here to identify patches of ground free of vegetation, which appear black. A number of periglacial features are visible, including: A. ice, cored mounds; B. thermokarst features; C. frost boils; D. remnant polygonal ice wedges; E. dark spots within the oval are people. The collection of dark spots and square at upper right are people and equipment. With the exception of the frost boils (C), vegetation cover is complete. The field book in the upper left inset image is 12 cm wide (aerial image courtesy of Dr. N. Harris, University of Alaska Fairbanks).

Figure 2. A cross-section of the frost boil (dashed line) and surrounding (thick line) microtopography measured at 0.05 m intervals on August 27, 2001, with the positions of the buried instruments. Subsurface instruments include 0.3 m long TDR sensors for volumetric liquid water content, thermistors for soil temperature, heave rods and transient heat pulse sensors for thermal conductivity and diffusivity. Heave rods extended from anchors in the permafrost to above the ground surface.

Figure 3. Measured thermal conductivity [W m$^{-1}$ K$^{-1}$] at 0.2 m depth beneath the frost boil center as a function of soil temperature [°C]. Three years of data are included in the parameterization, which applies to temperature less than or equal to -1.5 °C only. Around the freezing temperature, latent heat effects inflate the apparent thermal conductivity. In the unfrozen soil, the conductivity is sensitive to changes in soil water content.

Figure 4. Soil temperature (grey scale at right) and soil liquid water content (contour interval of 0.2 m$^3$ m$^{-3}$ for values from 0.1 to 0.5 m$^3$ m$^{-3}$) beneath the centre of the frost boil for four hydrologic years. The shallowest water content sensor was inoperative during freezing in 2001.

Figure 5. The cumulative degree-days of thawing (upper graph), the position of the ground surface and the frost table depth relative to the ground surface (at 0 m, lower graph) are shown for the summers of 2003, 2004 and 2005. Surface position is measured over the frost boil center. Frost table depth is observed as the decrease in liquid water content during freezing and interpolated between sensor depths by fitting thaw depth to degree-days of thawing using the Stefan solution for thaw depth. Observed thaw depths (frost table) are shown as circles for both years. Fitting parameters are given in the text.

Figure 6. The end of summer position of the ground surface as measured using 5 buried heave rods (2001-2003) and 5 ultrasonic distance sensors (2003-2005). Each rod/sensor's data is shown relative to its September 14, 2003 position. There are 25 observations shown, incorporating 2 sensing techniques and 5 locations.

Figure 7. The frost boil topographic cross-section is superimposed on three thin solid lines indicating heave measured between the end of summer and the beginning of
subsequent thaw for the winters of 2001/02, 2002/03 and 2003/04. The vertical scale for
the heave data has been exaggerated by a factor of two (scale at right). Frost front
penetration profiles recorded at 9:00 am on the dates shown are shown for four years.

Figure 8. Pre-thaw volumetric soil ice content estimated from spring subsidence rates
porosity measured on profile soil samples is indicated as grey circles; porosity values
are joined with a grey dashed line as a visual aid.

Figure 9. The latent heat production calculated from the temperature data. A black line
and a discontinuity in the grey scale indicates the transition between heat source and
sink.
Table 1. Selected frost boil soil profile properties.*

<table>
<thead>
<tr>
<th>sample depths [m]</th>
<th>bulk density [g cm$^{-3}$]</th>
<th>clay, silt, sand contents [% wt.]</th>
<th>total carbon [% wt]</th>
<th>porosity [-]</th>
<th>C/N [-]</th>
<th>CEC [meq g$^{-1}$]</th>
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<tr>
<td>0.06</td>
<td>1.62</td>
<td>43.9, 44.4, 11.7</td>
<td>2.21</td>
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<td>0.121</td>
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<td>44.5, 44.2, 11.3</td>
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<td>1.76</td>
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<td>11</td>
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<td>36.0, 38.7, 25.3</td>
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<td>1.09</td>
<td>1.72</td>
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<td>2.19</td>
<td>0.40</td>
<td>9</td>
<td>0.141</td>
</tr>
</tbody>
</table>

from region surrounding boil:

(A-horizon)  
| 0.10 | n. d. | 12, 34, 54 | 6.1 | 0.69 | 33 | n. d. |

(Bg horizon)  
| 0.20 | 0.5   | 15, 15, 70 | 4.0 | 0.88 | 8  | 0.129 |

Table 2. Mean monthly air temperatures for the four-year period of record.

<table>
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<th>2003/04</th>
<th>2004/05</th>
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<td>0</td>
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<tr>
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<td>-3</td>
<td>-1</td>
<td>-3</td>
<td>-3</td>
</tr>
<tr>
<td>November</td>
<td>-7</td>
<td>-5</td>
<td>-6</td>
<td>-6</td>
</tr>
<tr>
<td>December</td>
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<td>-7</td>
<td>-17</td>
<td>-14</td>
</tr>
<tr>
<td>January</td>
<td>-13</td>
<td>-16</td>
<td>-22</td>
<td>-14</td>
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<td>-17</td>
<td>-26</td>
<td>-16</td>
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<td>March</td>
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<td>-15</td>
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<td>July</td>
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<td>10</td>
<td>7</td>
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<tr>
<td>August</td>
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<td>6</td>
<td>9</td>
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### Table 3: List of Symbols

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<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
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<tr>
<td>A</td>
<td>constant</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>constant</td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>heat capacity ($C_{nc}$: of the $n$th soil component)</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>D</td>
<td>thermal diffusivity</td>
<td>m$^2$ s$^{-1}$</td>
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<tr>
<td>$R_T$</td>
<td>thermistor resistance</td>
<td>$\Omega$</td>
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<tr>
<td>$T$</td>
<td>soil temperature</td>
<td>°C</td>
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<tr>
<td>$\overline{T}$</td>
<td>mean daily air temperature</td>
<td>°C</td>
</tr>
<tr>
<td>$T_f$</td>
<td>freezing temperature</td>
<td>°C</td>
</tr>
<tr>
<td>$k$</td>
<td>thermal conductivity</td>
<td>W m$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$r_h$</td>
<td>heat production (+ve) or consumption (-ve) in the soil</td>
<td>W m$^{-3}$</td>
</tr>
<tr>
<td>$t$</td>
<td>time</td>
<td>s</td>
</tr>
<tr>
<td>$z$</td>
<td>depth below ground surface</td>
<td>m</td>
</tr>
<tr>
<td>$\Delta z_0$</td>
<td>change in ground surface position relative to permafrost</td>
<td>m</td>
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<tr>
<td>$\Delta z_f$</td>
<td>change in frost table position relative to permafrost</td>
<td>m</td>
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<tr>
<td>$\delta \chi$</td>
<td>uncertainty in $\chi$</td>
<td>units of $\chi$</td>
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<tr>
<td>$\theta$</td>
<td>volumetric content</td>
<td>m$^3$ m$^{-3}$</td>
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<tr>
<td>$\theta'$</td>
<td>volumetric segregated ice content</td>
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<tr>
<td>$\rho$</td>
<td>density</td>
<td>kg m$^{-3}$</td>
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<tr>
<td>$\eta$</td>
<td>soil porosity</td>
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<table>
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<tr>
<td>i</td>
<td>ice</td>
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<tr>
<td>j, k</td>
<td>measurement depth, time</td>
</tr>
<tr>
<td>n</td>
<td>$n$th soil component</td>
</tr>
<tr>
<td>w</td>
<td>water</td>
</tr>
<tr>
<td>s</td>
<td>soil</td>
</tr>
</tbody>
</table>
$k = A - e^{B|T|}$

$A = 1.55$, $B = -0.20$

$R^2 = 0.89$ (N = 858)
ground surface position [cm] = 0.015 * time [days] - 563

$R^2 = 0.95$ (N = 25)