1   **Frost Boils and Soil Ice Content: Field Observations**

2

3   P. P. Overduin, Potsdam Institute for Climate Impact Research, Telegrafenberg C4,

4   14473 Potsdam, Germany. Tel: +49-331-288-2534. Fax: +49-331-288-2137. Email:

5   fsppo@uaf.edu.

6

7   Dr. Douglas L. Kane, Water and Environmental Research Center, University of Alaska

8   Fairbanks, P. O. Box 755860, Fairbanks, AK, USA. Email: ffdlk@uaf.edu.

9   **Ice content and frost boils: Field observations**

10   Overduin, P. P., Kane, D. L.

11

12   **Abstract**

13   Our aim is to measure and explain the seasonal changes in soil ice content in the frost

14   boils of Galbraith Lake, Alaska. Instruments were installed in a frost boil to monitor the

15   ground surface position and soil state over a period of four years. By comparing the

16   subsidence and thaw rates, we calculate the soil ice content as a function of depth.

17   Measured soil temperatures, liquid water contents and bulk apparent thermal

18   conductivities are used to estimate latent heat production and release in the soil. The

19   frost boils heave during freezing and settle during thaw while the surrounding tundra

20   heaves negligibly, but subsides measurably. Despite large changes in freezing rates

21   from year to year, total heave and its distribution across the frost boil are similar

22   between years. Winter air temperature and snow depth influence the freezing rate and

23   ice distribution as a function of depth, but not the overall heave. This suggests that

24   heave is controlled by water availability rather than the rate of heat removal from the

25   soil. Areal ground subsidence rates between 2 and 5 cm/yr are due to the disappearance

26   of ice at the base of the active layer, raising the possibility of ongoing thermokarst

27   expansion around Galbraith Lake.

28

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

55 ice at a variety of scales. Methods that have been used for ice content include:

56 gravimetric; radar, either as ground-based or remote techniques; acoustics; neutron

57 absorption for the determination of total water content; dielectric methods (Bittelli *et al.*

58 2003) and volumetric methods. We use the changes in soil volume, temperature and

59 thermal properties during freezing and thawing to draw conclusions about changes in

60 ice volume. Our objectives are to describe the heave dynamics of a frost-boil, to

61 quantify the role of seasonal ice growth in these features and to evaluate the role of ice

62 formation and segregation in differential frost heave at this site.

63

64 **Methods**

65 *Site*

66 Galbraith Lake is about 3 km long. It is located at the west end of Atigun Gorge on the

67 north side of the Brooks Range, Alaska (68°28.890'N, 149°28.744'W). Water enters the

68 lake through a tributary at its northern end and through an alluvial fan that forms its

69 western shore. It drains into the Saganivirktok River, which flows across the North

70 Slope of Alaska into the Arctic Ocean. Aerial photographs show evidence of a series of

71 old shorelines beyond the northern end of the lake. Within this region of exposed and

72 reworked lake bed, a mixture of lacustrine and fluvial sediments form a nearly

73 horizontal plain between outwash fans of streams into the valley from the west, and

74 bluffs on the east. At present, there is an airstrip located on the west side of the plain,

75 and the Dalton highway and the Trans-Alaska Pipeline run north to south along the east

76 side of the valley. The site is poorly drained and the water table is within 20 cm of the

77 ground surface during the growing season. Land-cover type is classified as moist non-

78 acidic tundra, the soil pedon is classified as a cryaquept and the soil horizons are

79   contorted by cryoturbation (Ping *et al.* 2003). From 1987 until 2001, mean annual

80   ground temperature ranged between -8 and -4 °C at the permafrost table and between -

81   5.7 and -4.7 °C at 20 m (Osterkamp 2003).

82   Visual and near infrared band images of the site were collected using the Bi-camera

83   Observation Blimp (BOB) low-altitude aerial photography system (Harris and

84   Helfferich 2005). Ground features were used to georectify the images, which have a

85   resolution of better than 0.07 m. The site is shown (Figure 1) in a normalized difference

86   image (NDVI) of the reflected near infrared and red bands. NDVI is an index for

87   reflection from photosynthesizing vegetation. It is used here to identify vegetation-free

88   patches of ground, which appear black. The only features devoid of vegetation in the

89   field of view are the frost boils, water surfaces and some humans, which appear darker

90   than the vegetated surfaces (Figure 1).

91   ***Frost boil morphology***

92   Frost boils at Galbraith Lake are identified at the ground surface as roughly circular

93   areas free of vegetation and between 0.5 and 1.5 m in diameter. In some cases, several

94   frost boils appear to have developed close enough to one another to form linear regions

95   of frost boil up to 3 m long. The vegetation cover of the surrounding tundra is complete,

96   and the frost boils thus have well-defined boundaries (Figure 1, inset). Based on the 110

97   frost boils identified in Figure 1 (C), the mean distance between frost boils is 5.6 ± 0.6

98   m. They are not uniformly distributed, as the terrain is also the site of other periglacial

99   features, including ice-cored mounds (A), thermokarst ponds (B) and ice wedges (D),

100 that are found throughout the image. Dark spots within the oval (E) are people. Frost

101 boils are also found in the surrounding tundra, outside of the old lake bed region, but

102   tend to be covered with vegetation to some degree, and are smaller with less distinct

103   boundaries.

104   Galbraith Lake frost boils appear to have undergone sorting, with fine silty-clay mineral

105   material in the frost boil center and sandy soil underlying the vegetated margins and

106   surrounding tundra of the boils. The surfaces of the frost boils are cracked in a columnar

107   structure, and a crust has developed. Major ion analyses of a crust sample showed little

108   difference between the crust and the soil beneath it (Ping *et al*. 2003). They are slightly

109   domed, and their elevation relative to the surrounding tundra varies with the time of

110   year, which is examined in the section on frost heave, below.

111   A soil pit roughly bisecting a frost boil was carefully excavated on August 25, 2001.

112   The soil pit was approximately 1.4 m wide at the ground surface. Soil was removed by

113   horizon with consideration for the three-dimensional structure of the horizons, and

114   replaced and compacted so that the pit was refilled. Beneath the center of the frost boil

115   there is little change in soil composition with depth. All sampled horizons were mineral

116   with organic contents of less than 3% by weight, had bulk densities in the range of 1.62

117   to 1.76 g cm^{-3} and saturated soil-paste pHs of 8 to 8.2 (Table 1). Little difference was

118   observed in soil texture variation with depth, down to 0.73 m. Soils surrounding the

119   frost boil have organic upper horizons 0.05 to 0.20 m thick with lower bulk density,

120   higher carbon content, and support a 100% cover of mosses and sedges. The soil pit was

121   excavated to a depth of 1.3 m and the frost table beneath the center of the frost boil was

122   encountered at 0.73 m. A jackhammer was used to excavate below the frost table. The

123   thaw depth in the frost boil exceeded the thaw depth immediately adjacent to and at a

124   distance of 2.5 m from the boil by over 0.22 m, creating a bowl-shaped depression in

125   the frost table beneath the frost boil. Many ice lenses were encountered between 0.75 to

126   0.83 m, running parallel to the frost table and containing ice lenses up to 7 mm thick.

127   They were angled upwards towards the margin of the frost boil and not found beneath

128   the adjacent tundra. Below 0.83 m, ice layers with a blocky, horizontally-oriented

129   structure and 0.02 to 0.05 m thickness were separated by thinner soil layers. This ice

130   was associated with the transition zone, the layer of soil that is usually permafrost, but is

131   subject to infrequent thaw during warm years. A core was drilled into the bottom of the

132   soil pit to determine the lower extent of the ice-rich deposits: ice layers were

133   encountered down to te termination depth of 1.6 m. Thus, the frost boil was thawed to a

134   depth of 0.73 m, with thin ice lens for 0.10 m below the frost table, and thicker ice

135   layers underlying these lenses for at least another 0.77 m. The estimated volumetric ice

136   content of the soil below 0.83 m was over 0.90 m^3 per cubic meter of soil.

137   The frost boil morphology and material grain-size results in a classification after Zoltai

138   and Tarnocai (1981) as raised-center mudboil, despite the poor drainage of the site.

139   Ping *et al.* (2003) concluded that the soil profile morphology showed upward extrusion

140   of clayey material from a lower soil horizon into an overlying sandy horizon, locally

141   displacing the upper horizon and creating a textural differentiation at the ground

142   surface.

143   ***Instruments***

144   Temperature, volumetric liquid water content and thermal conductivity sensors were

145   installed in the soil pit wall between the surface of the soil and 1.36 m depth (Figure 2).

146   Temperatures were measured hourly using a vertical array of thermistors. The

147   thermistors were calibrated using a de-ionized water-ice mixture, from which a

148   thermistor-specific offset, δ_o, for the Steinhardt-Hart equation was generated:

149     $$T^{-1} = 1.28 \times 10^{-3} + 2.37 \times 10^{-4} \left(\ln\left(R_T - \delta R_0\right) \right) + 9.06 \times 10^{-8} \left(\ln\left(R_T - \delta R_0\right) \right)^3 \qquad (1)$$

150 where R_T is the measured resistance and δR_0 is the resistance offset at 0 °C. The

151 temperature sensor uncertainty is better than ±0.01 °C at 0 °C.

152 Time domain reflectometry sensors (Campbell Scientific CS605) connected to a

153 reflectometer and datalogger (Campbell Scientific TDR100 and CR10X) were used to

154 record the apparent sensor length at one-hour intervals, which is related to soil moisture

155 content. In addition, the entire reflected waveforms for each sensor were measured once

156 per week to verify the TDR100 waveform analysis algorithm, and to provide an

157 alternate time series. Occasional TDR100 waveform analysis algorithm failure occurred

158 as a result of temperature-induced changes in the sensor cable length. This problem was

159 solved by trial and error determination of two cable lengths in the datalogger program,

160 one for thawed soil and one for frozen soils. The sensors had been previously calibrated

161 in the laboratory using water and air as reference materials following the method of

162 Heimovaara *et al.* (1995). The bulk relative dielectric permittivity data was used to

163 calculate liquid water content using Roth *et al.*'s (1990) mixing model for mineral

164 horizons and an empirical relationship for organic horizons similar to the ones under

165 study (Overduin *et al.* 2005). The accuracy of the method is estimated to be better than

166 0.05 m^3 m^{-3} (Heimovaara *et al.* 1995).

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

168 $$D = \frac{k}{\rho C} \qquad (2)$$

169 where D, k, ρ and C are the bulk thermal diffusivity, conductivity, density and

170 volumetric heat capacities of the soil, respectively. Thermal conductivity was measured

171 using Hukseflux TP01 transient heat pulse sensor installed at 0.1 and 0.2 m depth

172 beneath the center of the frost boil. The radial temperature difference following heating

173   for 180 s was measured and analyzed following Overduin *et al.* (2006). All data points

174   were included in the analysis, including peaks in bulk apparent thermal conductivity due

175   to changes in phase induced by the transient heat pulse. The uncertainty in k is

176   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.*

177   2006).

178   ***Thaw and frost penetration depth***

179   We used the decrease in liquid water content at each TDR sensor to identify the time at

180   which the freezing front passed each sensor. Phase change was identified as the point of

181   inflection in the time series of soil liquid water content. To interpolate thaw and frost

182   front positions between sensor depths, the data were fit to a simplified solution of the

183   Stefan problem:

184    $$z = A\sqrt{\sum \left| \overline{T} - T_f \right|} + B \qquad\qquad (3)$$

185   where the sum indicates summation over days of the year, $\overline{T}$ is the mean daily air

186   temperature, T_f is the freezing point and A and B are constants. The term within the

187   square root is referred to as the degree-days of freezing or thawing (Yershov 1990). The

188   coefficients A and B are specific to a freeze or thaw season, and were used to predict

189   thaw based on cumulative degree-days of thawing.

190   ***Frost heave***

191   By comparing soil volumes in the thawed and frozen states, changes in the volumetric

192   fractions of water and ice may be inferred. Heave was measured in two ways. For the

193   excavated site, fiberglass rods 1.22 m long and 3 mm in diameter were anchored in the

194   permafrost at depths between 0.7 and 1 m using 5 cm long cross-pieces fastened using

195   screws. The vertical rods extended above the ground surface. The rods were distributed

196   in a line from the approximate center of the frost boil to a point 2.47 m away (Figure 2).

197 Metal washers were free to slide along the rods and rested on the ground surface. The

198 rod length exposed above the metal washer was measured at irregular intervals

199 throughout the year. At a second frost boil a few meters from the first, five ultrasonic

200 distance sensors (Sonic Ranger SR50, Campbell Scientific Corporation) were mounted

201 on a horizontal rod supported by two vertical members, which were anchored in the

202 permafrost at a depth of over 1.5 m and encased in tubing that extended over the active

203 layer depth. The sensors have a resolution of 0.1 mm and an accuracy of +/-0.01m, after

204 correction for variations in air temperature. The sensors record the signal from the

205 nearest target in their approximately 22° field of view. They are thus affected by the

206 presence of snow and vegetation. The hourly values used here are smoothed using a

207 running average during snow-free periods only.

208 Following suggestions by Burn (1998), we define the active layer depth as the

209 penetration depth of the 0 °C isotherm. To calculate the rate of thawing in spring, the

210 position of the 0 °C isotherm from 0.12 to 1.36 m was interpolated as described in the

211 previous section. We assume that daily subsidence at the surface is due to daily ice

212 melting at the thaw front. Although uncertainty in settling rates at a daily time scale is

213 high and probably depends on flow-rates of excess moisture from the thawing front: the

214 correspondence between the rates of subsidence and thaw suggests that our assumption

215 is reasonable. Measurable subsidence did not continue after the active layer reached its

216 maximum depth at the end of summer.

217 We define the vertical positions of the ground surface $z_0(t)$, and the phase change

218 boundary, $z_{ft}(t)$, relative to the ground surface at the start of thawing or freezing. Settling

219 is ascribed to the melting of ice in the soil to the depth z_{ft}. Accounting for the difference

220   in ice and water densities, the segregated ice content averaged over the entire profile

221   from z_0 to z_{ft} before thawing is then given by:

222    $$\theta_i'' = \frac{\Delta z_0}{\Delta z_{ft}} - \theta_w \frac{\rho_w}{\rho_i} \frac{\Delta z_{ft} - \Delta z_0}{\Delta z_{ft}}$$ (4)

223   where $\rho_i = 917.0$ kg m^{-3} and $\rho_w = 999.8$ kg m^{-3} are the densities of water and ice,

224   respectively, and Δz_0 and Δz_{ft} are the change in the ground surface and the phase change

225   boundary positions.

226   *Volumetric latent heat production*

227   Changes in ice content can also be inferred from latent heat production and

228   consumption in the frozen ground (Roth and Boike 2001). Conservation of energy and

229   Fourier's law lead to the heat diffusion equation in one dimension:

230    $$\frac{\partial}{\partial t}(\rho C T) = r_h + \frac{\partial}{\partial z}\left(k \frac{\partial T}{\partial z}\right)$$ (5)

231   where r_h is the heat production or consumption per unit volume. We use the temperature

232   profile beneath the frost boil center to estimate the heat production or consumption in

233   the soil by integrating over each soil layer between sensors at depths z_j and z_{j+1}, and

234   between observation time steps t_k and t_{k+1}, where j and k are ordination subscripts

235   denoting steps in observation time or depth. Integrals are approximated using the

236   trapezoidal rule and the partial derivative of temperature is estimated using a central

237   finite difference. The approach of Roth and Boike (2001) is followed, except that a time

238   dependent C is calculated using equation 6. The bulk soil heat capacity depends

239   primarily on the soil's volumetric liquid water content and is calculated from measured

240   soil composition. The dependence of the bulk heat capacity on composition is the

241   volumetric fraction-weighted arithmetic mean of the soil components:

242 $$\rho C = \sum_{n=a,i,w,s} \rho_n C_n \theta_n \qquad (6)$$

243 where ρ is the soil bulk density, and ρ_n is the density, C_n the heat capacity, and θ_n the

244 volumetric fraction of soil component n. The subscripts refer to air (a), ice (i), water (w)

245 and soil (s). The volumetric heat capacities of ice, water and soil used were 1.93, 4.18,

246 and 1.88 MJ m^{-3} K^{-1}, respectively, while that of air was neglected. The volumetric liquid

247 water content was taken from the TDR measurements and the volume fraction of soil

248 was taken from the porosity measurements in Table 1. Ice content was estimated as the

249 difference between measured liquid water content and the soil porosity, which is

250 equivalent to the assuming saturation and that moisture migration does not take place.

251 This coarse assumption contradicts the clearly observed effects of water migration but is

252 only used in the estimation of soil heat capacity. Since the relative dielectric

253 permittivity of ice is low and similar to that of the soil matrix, the liquid water content is

254 only linearly and weakly dependent on ice content. A 10% change in the ice content

255 results in a volumetric liquid water content change of less than 0.8% for the entire

256 period of record, well below the estimated uncertainty in water content. The heat

257 capacity is linearly dependent on estimates in soil component fractions. We estimate its

258 uncertainty, δC as the sum of the uncertainties in soil fractional heat capacities:

259 $$\delta C = \sum_{n=i,w,s} \delta C_n \qquad (7)$$

260 where the uncertainty introduced by the exclusion of air is ignored. For the unfrozen

261 soil, δC is equivalent to the sum of the products of the volumetric heat capacities of the

262 soil components and their uncertainties ($\delta\theta_i = \pm0$, $\delta\theta_w = \pm0.05$, $\delta\theta_s = \pm0.05$). At subzero

263 temperatures, however, ice may segregate. Since we do not measure soil ice content

264 directly, our uncertainty for bulk volumetric soil heat capacity in the frozen soil

265   increases by a factor of almost 10 and reaches values comparable to the heat capacity of

266   bulk frozen saturated soils (1.6 to 2 MJ m^{-3} K^{-1}; Yershov 1990), based on uncertainties

267   in soil component volumetric fractions of $\delta\theta_i = \pm1.0$, $\delta\theta_w = \pm0.05$, $\delta\theta_s = \pm0.35$, where

268   the latter value is $(1-\eta)$, where η is the porosity. However, this uncertainty is reached

269   only if ice segregation results in the complete displacement of liquid water and soil

270   particles from the soil. The freezing characteristic curves obtained by the TDR sensors

271   and thermistors for these soils show that liquid water exists throughout the measurement

272   period so that soil particles cannot have been displaced, at least at the scale of

273   measurement of the TDR sensor.

274   Values of bulk apparent soil thermal conductivity are calculated using the least squares

275   fit of measured thermal conductivity values against soil temperature (for temperatures

276   below –1.5 °C) shown in Figure 3, which includes both cooling and warming data over

277   a three year period. Calculations of latent heat are restricted to times and depths for

278   which the soil temperature is below –1.5 °C to avoid the influence of latent heat

279   changes on measured thermal conductivity values. Fuchs *et al.* (1978) showed

280   theoretically that the effects of phase change on the apparent thermal conductivity are

281   limited to a well-defined temperature range (for example, -0.5 to 0 °C for a Palouse

282   loam). The lower limit of this temperature dependency is a function primarily of total

283   soil water content.

284

285   **Results and Discussions**

286   *Rates of freezing*

287   Soil temperature and volumetric liquid water content for four hydrologic years

288   (2001/02, 2002/03, 2003/04 and 2004/05) show differences between the four winter

289 periods (Figure 4). Following, we describe the observed differences in freezing rate

290 between 2002 and the other years in a variety of ways. During 2001, 2003 and 2004 the

291 active layer rapidly changed from positive temperatures to values less than –0.1 °C

292 throughout the soil profile. In these three years the lowering of the zero curtain (*i.e.* the

293 time taken to bring the entire soil profile to temperatures less than or equal to –0.1 °C)

294 from the shallowest sensor at 0.12 m depth took less than a week (September 19-21,

295 2001, September 9-11, 2003 and September 12-15, 2004). Between 0.12 and 0.77 m,

296 the mean freezing rates (calculated using the position of the 0 °C isotherm) were 4.8 x

297 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,

298 respectively. Although the freezing rate was more rapid in 2001 than in 2003 or 2004, it

299 occurred ten days earlier than in 2003. The freezing rate during the fall of 2002 was

300 more than an order of magnitude lower than in the preceding and following years and

301 the active layer at 0.77 m depth remained unfrozen until after November 20. Even if the

302 –0.5 °C isotherm is used to indicate freezing, freezing was delayed by over a month in

303 at this depth in 2002, compared to other years.

304 We attribute the delay in freezing to changes in snow cover and winter air temperatures.

305 On Oct 1, 2002, over 0.16 m of snow accumulated. As a result of subsequent snowfalls,

306 the snow pack grew to almost 0.5 m thickness without diminishing over the course of

307 the winter. During the following winter, snow fell between Sep 6 and Sep 26, 2003, but

308 had disappeared from the frost boil surface by October 1, 2003. Subsequent snow

309 covers that exceeded 0.02 m lasted for less than 20 days. The effect of snow cover on

310 soil temperatures is not straightforward. Sokratov *et al.* (2001) used a 30-year data set to

311 show that the ratio between mean air temperatures and soil temperatures at 0.2 m depth

312 are not correlated with snow depth except during February and March. Taras *et al.*

313 (2002) distinguished snow depths sufficient to decouple air and soil temperatures from

314 one another (> 80 cm) from shallow snow depths (< 25 cm) in which air and soil

315 temperatures are highly coupled. Intermediate snow depths were strongly coupled. They

316 found that a 15 cm increase in snow depth produced a 0.5 to 3.0 °C increase in soil

317 temperature at the snow-soil interface. Under this scheme, the snow depths observed in

318 2002/03 and 2003/04 correspond to intermediate and shallow snow depths, respectively.

319 Air temperatures in 2003/04 were also lower than in other years. Mean monthly air

320 temperatures for September 2001 to August 2005 are presented in Table 2. The winter

321 months from December to April were all colder during 2003/04 than during other years;

322 February and March, in particular, were over 8 °C colder. Monthly mean soil

323 temperatures were colder by 5 to 11 °C for most of the winter of 2003/04 than the

324 previous winter. The mean monthly temperatures during freeze-up were also warmest

325 in 2003/03. While the effect of a change in air temperature on soil temperatures under

326 variable snow covers is not straightforward, Stieglitz *et al.* (2003) demonstrated the

327 equivalent importance of air temperature and snow cover variability in determining

328 ground temperatures. It seems reasonable that air temperature and snow cover suffice to

329 explain differences in freezing rates between years. Since freezing rates in 2001/02,

330 2003/04 and 2004/05 are similar in magnitude, freezing in 2002/03 appears to have

331 been anomalously slow for this location.

332 Despite microclimatic changes from year to year, freeze-back within the frost boil is

333 generally rapid compared to other tundra locations, especially those protected with an

334 organic mat (e.g. Hinzman *et al.* 1998). Mean monthly winter air temperatures for each

335 of the years 2001/02, 2002/03 and 2004/05 fall within a few degrees of each other.

336 Thus, based on soil temperatures during 2002/03, the effect of an intermediate snow

337   cover, under normal winter conditions, is to delay freeze-back by at least one month at

338   most depths. The combined effects of the exposed frost boil mineral soil, shallow snow

339   depth and cold winter temperatures in 2003/04 contribute to the extremely cold soil

340   temperatures in the frost boil.

341   ***Thaw and Subsidence***

342   For the springs of 2003, 2004 and 2005, interpolated thaw depth data are shown in

343   Figure 5, along with air temperature, the timing of phase change at each TDR sensor

344   and the position of the ground surface. The ground surface and interpolated thaw depth

345   positions are shown, the former relative to its pre-thaw position. In all three years, the

346   correlation coefficient between the Stefan solution and the observed phase change was

347   better than $r = 0.99$, with coefficients $\{A, B\}$ in equation 3 of $\{-3.67 \times 10^{-2}, 0.87 \times 10^{-2}\}$,

348   $\{-2.98 \times 10^{-2}, - 4.59 \times 10^{-2}\}$ and $\{-3.09 \times 10^{-2}, -7.27 \times 10^{-2}\}$ for 2003, 2004 and 2005,

349   respectively. During the springs of 2004 and 2005, the profile thawed down to 0.12 m

350   by May 15th. In 2003, however, thaw reaches this depth more than 2 weeks later, a

351   reflection of differences in degree-days of thawing, which by July 18 totalled 687 °C·d

352   in 2004 but only 458 °C·d in 2003, with thaw depths of 0.88 and 0.78 m, respectively

353   (Figure 5). The spring subsidence in each year was similar in magnitude and reached

354   2.09×10^{-2}, 2.24×10^{-2} and 4.17×10^{-2} m by July 18 in 2003, 2004 and 2005,

355   respectively (for 458, 687, 527 degree-days of thawing). Nonetheless, the amount of

356   subsidence occurring over the first 30 cm of thaw represented 68, 19 and 44% of the

357   end of summer total subsidence in 2003, 2004 and 2005, respectively, suggesting

358   significant differences in the distribution of segregated ice from year to year, at least for

359   the upper soil horizons. Egginton (1979) recorded subsidence of two mud boils west of

360   Hudson Bay, and found that over 80% of total summer subsidence (6 to 9 cm) occurred

361   within the first month after thawing began when the active layer was about 40 cm deep,

362   in agreement with MacKay and MacKay's (1976) observed ice lensing distribution in

363   non-sorted circles and with MacKay's (1980) model of formation for non-sorted circles.

364   Time series of surface position measured with both heave rods and ultrasonic distance

365   sensors between the fall 2001 and spring 2005 show annual subsidence. As a group, the

366   surface position measurements show a steady decrease in surface elevation (Figure 6).

367   Ground surface position relative to the heave rods lowered from 5.1 cm/yr at a point

368   2.47 m from the frost boil center to 7.5 cm/yr at the frost boil center. Subsidence and/or

369   frost-jacking of the rods would result in lowering of the ground surface. In addition,

370   soil pit excavation and refilling at the end of the summer of 2001 was probably followed

371   by settling of the ground, leading to higher measured lowering.

372   Results from an undisturbed frost boil corroborate the observed subsidence. The

373   ultrasonic distance sensors were mounted on permafrost anchors sheathed in PVC

374   tubing to prevent frost-jacking, and the frost boil was not excavated. Using ground

375   surface positions at the end of summer from these sensors, mean subsidence is 5.1

376   cm/yr at the frost boil, and ranges from 2.0 to 4.8 cm/yr for the other sensors for the

377   period from fall 2003 to fall 2005. The subsidence observed here affects both the frost

378   boil and the surrounding tundra, and does not directly affect active layer depth. Since

379   winter heave is similar from year to year and mostly due to the segregation of ice in the

380   upper 30 cm of the soil profile, the subsidence observed is attributable to the melting of

381   ice at the base of the active layer. It is more pronounced beneath the frost boil, where

382   summer thaw reaches greater depths, than beneath the surrounding tundra. We

383   observed 0.77 m of ice-rich soil (estimated >90% by volume) beneath the active layer.

384   The potential exists for continued subsidence and the northward expansion of Galbraith

385   Lake through thermokarsting.

386   *Heave*

387   Figure 7 shows three net winter heave profiles, measured between the end of summer

388   (late August - early September) and early spring (mid-May), superimposed on the

389   microtopography of the frost boil. Net winter heave increases towards a point just

390   outside the boil's edge from close to zero at a distance of 1.6 m. The maximum heave of

391   around 0.12 m occurs outside the frost boil boundary in all three years. The unvegetated

392   frost boil heaves between 0.04 to 0.08 m. The three heave profiles are remarkably

393   similar, given the differences observed in freezing rates for each fall season.

394   Usually, a few days after the ground begins to freeze from the surface downward, soil

395   temperatures between the upper and lower frost fronts reach slightly negative

396   temperatures (Osterkamp and Romanovsky 1997). In the high snow year, however, this

397   process is extended over a longer time period. Nonetheless, the depth of the freezing

398   front as a function of lateral position is similar from year to year (Figure 7). The 0 °C

399   isotherms, recorded at 9:00 am on the indicated dates, all show a thicker frozen layer

400   beneath the frost boil (10 – 30 cm) than beneath the tundra (10 – 15 cm). Segregation of

401   ice at the frost front results in heave normal to the frost front. The freezing front beneath

402   the margin of the frost boil is inclined so that a normal vector to the freezing front will

403   have a lateral component pointing towards the center of the boil.

404   We explain the location of maximum net winter heave outside the frost boil in one of

405   two ways: either (i) the shape of the freezing front and the relative freezing rates

406   beneath the boil and its margin lead to greater heave at the frost boil's margin than at its

407   center, or (ii) maximum heave does occur at the frost boil center at some time during the

408   winter but subsequent subsidence beneath the frost boil occurs prior to the end of snow

409   melt (during the winter). A single winter measurement of frost heave made in

410   November 2002 was 0.02 m higher than in the following pre-thaw May, suggesting that

411   sublimation or evaporation processes from the bare soil of the boil's apex could have

412   decreased the ice content over the winter. The microtopography, surficial cracking and

413   heave of the boil would contribute to these processes. The frost boils at Galbraith are

414   elevated above the surrounding tundra and the high winds in the Atigun Valley can

415   blow the snow off the frost boils alone. Under these conditions the surfaces of the boils

416   are observed to be desiccated. When the snow pack is sufficient to cover the frost boils,

417   there is no indication of the location of the boils at the snow surface. The snow pack is

418   necessarily thinner above the frost boils than over the tundra, and moisture migration

419   into the snow pack is exacerbated at the boil by a steeper temperature gradient through

420   the snow pack. Since the soil here is at or near saturation on freezing, there can be no

421   indication for this process using volumetric liquid water content, since it is independent

422   of total water content at cold temperatures. Since thaw begins with the infiltration of

423   melt water, the soil is saturated before all of the ice disappears in spring. Such a loss of

424   ice to the atmosphere is consistent with the shape of the heave curves in Figure 7.

425   The amount of heave depends on the freezing rate, and the dramatic difference in

426   freezing rate between 2003 and the other years lead us to expect differences in heave.

427   For example, O'Neill and Miller (1985) simulated heave and observed an increase in ice

428   segregation with a decrease in freezing rate. To explain the similarity between years, we

429   must either invoke processes that remove ice in late winter after a slow freeze-back, or

430   the possibility that heave is only weakly dependent on the freezing rate. The

431   correspondence between the rate of water flow to the freezing front and the driving

432   gradient (temperature) should enhance heave, so that we find the latter option difficult

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

434   *Ice content as a function of depth*

435   The timing of subsidence and the temperature profile during thawing are related to the

436   distribution of ice content as a function of active layer depth. Figure 8 shows the total

437   volumetric ice content as a function of depth based on cumulative settlement and thaw

438   in the springs of 2003, 2004 and 2005. The years refer to the spring in which ice

439   content was observed; in each case, most of this ice probably formed during fall freeze-

440   up in the previous year. Ice contents throughout the profile lay between 0.5 and 0.8,

441   with lower ice content in 2003 in the upper 0.4 m of the profile than in subsequent

442   years. Ice content exceeded the porosity in 2004 and 2005 over this same interval,

443   suggesting that more segregation occurred in the upper profile during freezing in 2003

444   and 2004 than during the fall of 2002 (corresponding to the graph for 2003). The

445   integrated ice contents between 0.12 and 0.40 m (for which depths data are available in

446   all three years) are 0.16 (2003), 0.19 (2004) and 0.19 m (2005). Heave is the same (+/- 2

447   cm) each year, but subsidence varies as much as 7 cm from year to year, suggesting that

448   ice lost in deep thaw years is not replaced by segregation during freezing.

449   In all three years, freezing occurs via both the descent of a freezing front from the

450   surface and an increase in the height of the frost table, although the latter process is

451   much more pronounced in 2002/2003 (Figure 4). The last depth to freeze in 2003/2004

452   is 0.79 m. Ice content is locally higher between 0.85 and 0.90 m in both years. The

453   qualitative similarity between increases and decreases in ice content with depth suggest

454   that the relative distribution of ice from year to year is stable, except for the upper

455   portion of the active layer, which is subject to repeated diurnal freeze-thaw cycles

456   during freezing (Overduin *et al.* 2003). The spikes in ice content below 0.75 m

457   correspond with the segregated ice observed in the field during excavation. Based on its

458   position, this ice is probably segregated during upward movement of the frost table

459   during freeze-back. Based on position, the inter-annual variability in shallow active

460   layer ice content is determined by snow pack and heat transfer conditions at the surface,

461   which affect post-freeze redistribution of water, rather than freezing rate. Deeper ice

462   content does not seem to be freezing-rate dependent, and is therefore less variable from

463   year to year.

464   ***Volumetric latent heat production***

465   Latent heat changes (positive or negative) for the temperature profile beneath the frost

466   boil for the 2002/03 and 2003/04 winter periods, when the entire profile is at or below

467   -1.5 °C, are presented in Figure 9. Since the soil is at or close to saturation on freezing,

468   changes in latent heat during this period are probably the result of freezing and thawing

469   in the frozen soil rather than vapour transport processes. Although it is probable that

470   other heat releasing or consuming processes are at work, such as sublimation near the

471   ground surface or vapour transport, the scale is given in W m^{-3} and also in the rate of

472   phase change of water (kg m^{-3} d^{-1}), using 333 kJ kg^{-1} as the latent heat of phase change.

473   Lighter colours indicate the consumption of latent heat. Our goal is to identify depths

474   and times of ice formation or disappearance during the winter.

475   In both winters, latent heat release dominates the profile between 0.4 and 0.9 m from

476   freeze-up until the end of April. This stands in stark contrast to the observations of

477   Roth and Boike (2001) for a site on Spitsbergen, who showed strong heat consumption

478   at intermediate depths during the cold winter period. They regarded this heat

479   consumption as evidence for evaporation of water and its diffusion upwards out of the

480 soil into the snow pack or atmosphere. Corroboration of this vapour diffusion using

481 changes in soil water content was not possible, since soil liquid water content does not

482 change appreciably as water is removed. Net changes in saturation over the winter are

483 masked by the infiltration and refreezing of melt water prior to soil thawing in the

484 spring, which results in near-saturation of the soil at thaw. In addition to the fact that

485 their analysis included constant soil thermal properties, ice segregation in the Galbraith

486 Lake frost boils presumably draws moisture laterally which continues to freeze well-

487 after freezing begins. The integrated net latent heat fluxes between Jan 1 and March 31

488 for the years 2002/03 and 2003/04 are 28.2 and 23.6 MJ m^{-2}, corresponding to water

489 amounts of 85 and 71 kg m^{-3}. These estimates account for the depth intervals of 0.31 to

490 1.1 m. Given that the upper and lower portions of this interval tend to lower values, net

491 latent heat consumption probably occurs in the overlying and underlying soil horizons.

492 These estimates can therefore only be regarded as upper bounds on the amount of ice

493 created for the whole profile. In the slow freeze-up year (2003/04) more latent heat is

494 released in the intermediate depths 0.5 to 0.98 m than in the previous year. This is more

495 than balanced out by latent heat consumed in the upper part of the profile, consistent

496 with greater moisture losses through evaporation and sublimation there, so that less

497 latent heat is produced in 2003/04 than in 2002/03. The consumption of latent heat that

498 occurs in the lower part of the horizon is difficult to explain, but may be related to the

499 transport of moisture upward in the profile in response to the temperature gradient and

500 suggests the steady loss of ice at depths where the soil composition is more than 90%.

501 These losses are observable as subsidence and represent a decrease in total permafrost

502 thickness, but will not be observable using active layer probing or by observing ice

503 content profiles that do not extend through the ice-rich layers.

504   Shallower depths are affected by higher amplitudes of latent heat values, with episodes

505   of heat consumption, probably as a result of ice sublimation from the surface. In both

506   years, latent heat consumption deepens towards spring, as the soil is warmed from

507   above. The influence of winter temperature fluctuations is seen throughout the profile,

508   but is muted in the warmer winter (2002/03) compared to the following cold winter.

509   The latent heat release at depths below 0.9 m suggests a steady decrease in ice content

510   of the soil throughout the winter, consistent with heave amounts lower than those

511   predicted by the latent heat release at intermediate depths, and perhaps consistent with

512   the observed annual subsidence.

513

514   **Conclusions**

515   As a result of changes in ice content during freezing, frost boils at Galbraith Lake in the

516   Brooks Range of Alaska heave during freezing. Freezing rates vary greatly

517   interannually, but heave remains similar from year to year, both in terms of magnitude

518   and distribution across the frost boil. Enough latent heat is released in intermediate

519   depths in the frost boil to account for winter heave. The low net winter heave levels at

520   the center of the frost boil are attributed to loss of moisture via evaporation or

521   sublimation through the exposed mineral soil of the frost boil during the winter. We

522   suggest that the winter removal of water from shallow soil horizons at the frost boil's

523   center results in low lateral displacement of soil, leading to extrusion of lower horizons

524   but to insignificant downward movement of soil material at the frost boil margins.

525   Settling of the ground surface and the development of the active layer together permit

526   estimation of the pre-thaw vertical distribution of ice in the active layer. Net annual

527   subsidence suggests that melting ice at the base of the active layer is not replaced by

528   annual winter heave, which results from ice formation in the upper portion of the frost

529   boil.

530   On the basis of surface position measurements during heave and subsidence, soil state

531   and composition measurements, and latent heat flux calculations, we observe small

532   interannual variability in heave magnitude for a frost boil, despite large changes in

533   freezing rate, snow cover and air temperature. More latent heat is released within the

534   soil during the winter than is necessary to account for the observed heave. It is

535   compensated for by processes consuming latent heat in the near-surface soil and below

536   the active layer.

References

Bittelli B, Flury M, Gaylon SC. 2003. A thermodielectric analyzer to measure the freezing and moisture characteristic of porous media. *Water Resources Research* **39**(2). doi:10.1029/2001wr000930.

Burn CR. 1998. The active layer: two contrasting definitions. *Permafrost and Periglacial Processes* **9**: 411-416.

Egginton PA. 1979. Mudboil activity, central District of Keewatin. In: *Current research, Part B: Geological Survey of Canada*, Paper 79-1B; 349-356.

Fuchs M, Campbell GS, Papendick RI. 1978. An analysis of sensible and latent heat flow in a partially frozen unsaturated soil. *Soil Science Society of America Journal* **42**(3): 379-385.

Harris C, Helfferich D. 2005. Who is BOB? *Agroborealis* **36**(2): 29-32.

Heimovaara TJ, Focke AG, Bouten W, Verstraten JM. 1995. Assessing temporal variations in soil water composition with time domain reflectometry. *Soil Science Society of America Journal* **59**: 689-698.

Hinzman LD, Goering DJ, Kane DL. 1998. A distributed thermal model for calculating soil temperature profiles and depth of thaw in permafrost regions. *Journal of Geophysical Research* **103**(D22): 28975-28991.

MacKay JR, MacKay DK. 1976. Cryostatic pressure in nonsorted circles (mud hummocks), Inuvik, Northwest Territories. *Canadian Journal of Earth Sciences*, 13 : 889-897.

MacKay JR. 1979. An equilibrium model for hummocks (nonsorted circles), Garry Island, Northwest Territories. Current research, Part A : Geological Survey of Canada Paper 79-1A, pp. 165-167.

MacKay JR. 1980. The origin of hummocks, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, 17 : 996-1006.

O'Neill K, Miller RD. 1985. Exploration of a rigid ice model of frost heave. *Water Resources Research* **21**: 281–296.

Osterkamp TE. 2003. A thermal history of permafrost in Alaska. In *Proceedings of the 8th International Conference on Permafrost, Zurich*, Phillips M, Springman SM, Arenson LU (eds). Balkema: Lisse, Netherlands; 863–868.

Osterkamp, TE, Romanovsky, VE, 1997, Freezing of the Active Layer on the Coastal Plain of the Alaskan Arctic, *Permafrost and Periglacial Processes*, **8**:23-44.

Overduin PP, Ping C-L, Kane DL. 2003. Frost boils, soil ice content and apparent thermal diffusivity. In *Proceedings of the 8th International Conference on Permafrost, Zurich*, Phillips M, Springman SM, Arenson LU (eds). Balkema: Lisse, Netherlands; 869-874.

Overduin PP, Yoshikawa K, Kane DL, Harden J. 2005. Comparing electronic probes for volumetric water content of low-density feathermoss. *Sensor Review* **25**(3): 215-221.

587 Overduin PP, Kane DL, Loon WKP van. 2006. Measuring thermal conductivity in
588   freezing and thawing soil using the soil temperature response to heating. *Cold*
589   *Regions Science and Technology* **45**(1): 8-22. DOI:
590   10.1016/j.coldregions.2005.12.003.

591 Ping C-L, Michaelson GJ, Overduin PP, Stiles CA. 2003. Morphogenesis of frost boils
592   in the Galbraith Lake area, Arctic Alaska. In *Proceedings of the 8th*
593   *International Conference on Permafrost, Zurich*, Phillips M, Springman SM,
594   Arenson LU (eds). Balkema: Lisse, Netherlands; 897-900.

595 Roth K, Schulin R, Fluehler H, Attinger W. 1990. Calibration of time domain
596   reflectometry for water content measurement using a composite dielectric
597   approach. *Water Resources Research* **26**: 2267-2273.

598 Roth K, Boike J. 2001. Quantifying the thermal dynamics of a permafrost site near Ny-
599   Ålesund, Svalbard. *Water Resources Research* **37**(12): 2901-2914. DOI:
600   10.1029/ 2000WR000163.

601 Sokratov SA, Golubev VN, Barry RG. 2001. The influence of climate variations on the
602   thermoinsulation effect of snow cover and on the temperature regime in the
603   underlying soil (in Russian, with English summary). *Kriosfera Zemli* **2**: 83-91.

604 Stieglitz M, Dery SJ, Romanovsky VE, Osterkamp TE. 2003. The role of snowcover in
605   the warming of Arctic permafrost. *Geophysical Research Letters* **30**(13): 1721-
606   1726.

607 Taras B, Sturm M, Liston GE. 2002 Snow-Ground Interface Temperatures in the
608   Kuparuk River Basin, Arctic Alaska: Measurements and Model. *Journal of*
609   *Hydrometeorology* **3**: 377-394.

610 Vliet-Lanoë van B. 1991. Differential frost heave, load casting and convection:
611   converging mechanisms; a discussion of the origin of cryoturbations. *Permafrost*
612   *and Periglacial Processes* **2**: 123-139.

613 Walker DA, Epstein HE, Gould WA, Kelley AM, Kade AN, Knudson JA, Krantz WB,
614   Michaelson GJ, Peterson RA, Ping C-L, Raynolds MK, Romanovsky VE, Shur
615   Y. 2004. Frost-Boil Ecosystems: Complex Interactions between Landforms,
616   Soils, Vegetation and Climate. *Permafrost and Periglacial Processes* **15**: 171–
617   188. DOI: 10.1002/ppp.487.

618 Washburn AL. 1956. Classification of patterned ground and review of suggested
619   origins. *Geological Society of America Bulletin* **67**: 823-865.

620 Yershov ED. 1990. General Geocryology. Cambridge University Press: Cambridge;
621   580.

622 Zoltai SC, Tarnocai C. 1981. Some nonsorted patterned ground types in Northern
623   Canada. *Arctic and Alpine Research* **13**(2): 139-151.

624 Zhu D-M, Vilches OE, Dash JG, Sing B, Wettlaufer JS. 2000. Frost heave in argon.
625   *Physical Review Letters* **85**: 4908-4911.

List of Figures

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

674 subsequent thaw for the winters of 2001/02, 2002/03 and 2003/04. The vertical scale for
675 the heave data has been exaggerated by a factor of two (scale at right). Frost front
676 penetration profiles recorded at 9:00 am on the dates shown are shown for four years.
677
678 Figure 8. Pre-thaw volumetric soil ice content estimated from spring subsidence rates
679 as a function of the depth of thaw, for 2002/2003, 2003/2004 and 2004/2005. Soil
680 porosity measured on profile soil samples is indicated as grey circles; porosity values
681 are joined with a grey dashed line as a visual aid.
682
683 Figure 9. The latent heat production calculated from the temperature data. A black line
684 and a discontinuity in the grey scale indicates the transition between heat source and
685 sink.
686

687   Table 1. Selected frost boil soil profile properties.*

sample depths	bulk density	clay, silt, sand contents			total carbon	porosity	C/N	CEC
[m]	[g cm^{-3}]	[% wt.]			[% wt]	[-]	[-]	[meq g^{-1}]
0.06	1.62	43.9,	44.4,	11.7	2.21	0.44	10	0.121
0.16	1.74	44.5,	44.2,	11.3	2.21	0.40	9	0.139
0.28	1.76	43.2,	43.7,	13.1	2.32	0.39	8	0.122
0.44	1.75	43.4,	43.1,	13.5	2.25	0.39	9	0.126
0.66	1.73	43.2,	45.1,	11.7	2.18	0.40	11	0.125
0.83	1.75	36.0,	38.7,	25.3	2.62	0.39	9	0.113
1.09	1.72	44.3,	43.3,	12.4	2.19	0.40	9	0.141
from region surrounding boil:								
0.10 (A-horizon)	n. d.	12,	34,	54	6.1	0.69	33	n. d.
0.20 (Bg horizon)	0.5	15,	15,	70	4.0	0.88	8	0.129

688   * data provided by the Soil Survey Staff, 2005. National Soil Survey Characterization
689   Data, Soil Survey Laboratory, National Soil Survey Center, USDA, NRCS , Lincoln,
690   NE., Thursday, February 17, 2005.

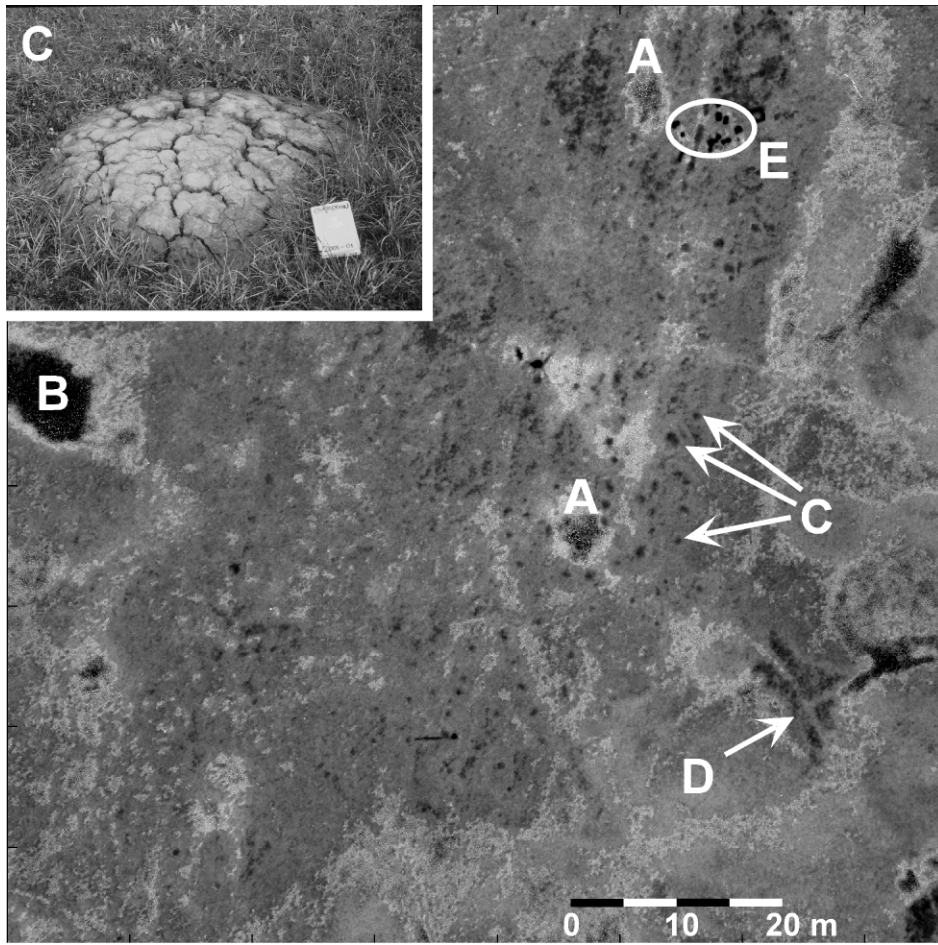691   Table 2. Mean monthly air temperatures for the four-year period of record.

	2001/02	2002/03	2003/04	2004/05
September	2	2	0	-1
October	-3	-1	-3	-3
November	-7	-5	-6	-6
December	-12	-7	-17	-14
January	-13	-16	-22	-14
February	-16	-17	-26	-16
March	-14	-16	-24	-15
April	-13	-12	-16	-15
May	-1	-5	-2	2
June	6	7	11	8
July	8	8	10	7
August	4	6	9	8

692

693   Table 3: List of Symbols

A	constant	--
B	constant	--
C	heat capacity (C_n: of the nth soil component)	$\mathrm{J\ kg^{-1}\ K^{-1}}$
D	thermal diffusivity	$\mathrm{m^2\ s^{-1}}$
R_T	thermistor resistance	Ω
T	soil temperature	°C
$\bar{T}$	mean daily air temperature	°C
T_f	freezing temperature	°C
k	thermal conductivity	$\mathrm{W\ m^{-1}\ K^{-1}}$
r_h	heat production (+ve) or consumption (-ve) in the soil	$\mathrm{W\ m^{-3}}$
t	time	s
z	depth below ground surface	m
Δz_0	change in ground surface position relative to permafrost	m
Δz_{ft}	change in frost table position relative to permafrost	m
δx	uncertainty in x	units of x
θ	volumetric content	$\mathrm{m^3\ m^{-3}}$
θ_i''	volumetric segregated ice content	$\mathrm{m^3\ m^{-3}}$
ρ	density	$\mathrm{kg\ m^{-3}}$
η	soil porosity	--

<div align="center">Subscripts</div>

a	air	
i	ice	
j, k	measurement depth, time	
n	nth soil component	
w	water	
s	soil	

694

695
696

697

698
699

$$k = A - e^{B|T|}$$
A = 1.55, B = -0.20
$R^2 = 0.89$ (N = 858)

701

702

703

ground surface position [cm] = 0.015 * time [days] - 563
$R^2 = 0.95$ (N = 25)