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Permafrost Destabilization and Thermokarst following Snow Fence Installation, Barrow, Alaska, U.S.A.

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Abstract

In autumn 1997, a 2.2 km-long, 4 m-high snow fence was constructed east of the coastal village of Barrow, Alaska. A large drift develops each winter on the downwind side of the fence, and a smaller drift forms upwind. To monitor the thermal impact on ice-rich permafrost, nine monitoring sites were installed near the fence in 1999 to measure soil temperature at 5, 30, and 50 cm; an additional three sites were located in the undisturbed tundra as a control. Maximum thaw and snow depth were measured annually. The results of the 6-yr study indicates that soil temperatures beneath the drift are 2 to 14°C warmer than the control in winter due to the insulting effects of the snow. Since the drift persists 4 to 8 wk after snow has disappeared from the undisturbed tundra, soil thaw is delayed and soil temperatures in summer are 2 to 3°C cooler than the control. The mean soil temperature over the 6-yr period of record has warmed 2 to 5°C, and the upper permafrost has thawed. The ground surface has experienced 10 to 20 cm of thaw subsidence in many places, and widespread thermokarst is apparent where snow meltwater ponds. Both direct soil warming and the indirect effects of ponding contribute to local permafrost destabilization.

Introduction

Snow is the most widespread and abundant surface mineral in the Arctic. It is an efficient thermal insulator and strongly attenuates heat conduction between the atmosphere and underlying ground. As noted by Walker et al. (1999), the snow cover is probably "the single most important mesoscale variable controlling biological systems in Arctic and alpine ecosystems."

There is a growing body of evidence suggesting an increase of shrub abundance in the Arctic tundra in response to warming (Sturm et al., 2001; Chapin et al., 2005). Drifting snow is captured by shrubs, increasing the snow depth and ground surface temperatures. One method to experimentally evaluate the potential physical, chemical and biological impacts of greater snow depth is to construct a snow fence.

Long-term experiments designed to enhance snow capture have been operating for more than a decade at Niwot Ridge in Colorado and Toolik Lake in northern Alaska. The purpose is to monitor the impact of enhanced snow accumulation on soil physical and chemical properties (D. A. Walker et al., 2001), biogeochemical processes (Williams et al., 1998; Welker et al., 2000), and vegetation response (M. D. Walker et al., 1999; Wahren et al., 2005). These studies report a significant increase in soil temperature beneath the drift, and demonstrate the impact of snow thickness on soil microbial activity, nutrient cycling, gas fluxes and vegetation response.

Increasing local snow-cover thickness can have opposing soil thermal impacts. Since snow is an effective thermal insulator, winter ground heat loss is reduced and soil temperatures are elevated (Seppälä, 1994; Sturm et al., 2005). Since the initial system enthalpy is higher, the energy required to warm and thaw the soil in summer is reduced and there is an overall tendency for the average annual soil temperature to increase. In regions characterized by permafrost, the active layer would become deeper as near-surface permafrost thaws. Melting of ice in supersaturated permafrost results in ground subsidence and thermokarst.

Field evidence supporting the ground warming scenario was reported by Mackay (1978). Two 1.2-m-high snow fences were used to increase snow depth over an ice-wedge trough. The frequency of ice-wedge cracking was greatly reduced due to the insulating effect. In the Schefferville mining area (55°N) on the Nouveau-Quebec–Labrador Peninsula, Nicholson (1978) used snow fences that increased snow-cover thickness 10 to 90 cm. This caused permafrost degradation and increased the thickness of the active layer such that, at the end of a 5-yr period, it was 2.5 times (6.5 m) deeper than the control. The impact was largely attributable to warmer winter soil conditions.

However, there is some evidence (e.g., D. A. Walker et al., 2001) to indicate that drifting can also counteract soil warming in spring and summer since thick accumulations of snow persist several weeks or months after the surrounding tundra is snow free. Soil warming is consequently delayed, average summer soil temperatures are cooler, and the active layer thickness is reduced. Outcalt and Goodwin (1975) conducted a modeling experiment using a snow-soil simulator. The results suggested that an artificial snowdrift exceeding 4.2 m in height would not ablate in summer at Barrow, Alaska.

Resolving these opposing thermal effects requires examination of both the direct and indirect impact of deep snow drifting on the soil thermal and moisture regimes. In poorly drained areas, for example, melting snow promotes water ponding at the frozen ground surface and enhances infiltration of meltwater through pores and contraction microcracks (Hinkel et al., 1997, 2001; Kane et al., 2001). This effect can be detected because near-surface soil warming occurs at rates far exceeding those typical of a purely conductive system. Further, as air is displaced by water in soil pores, the soil thermal conductivity increases and the potential

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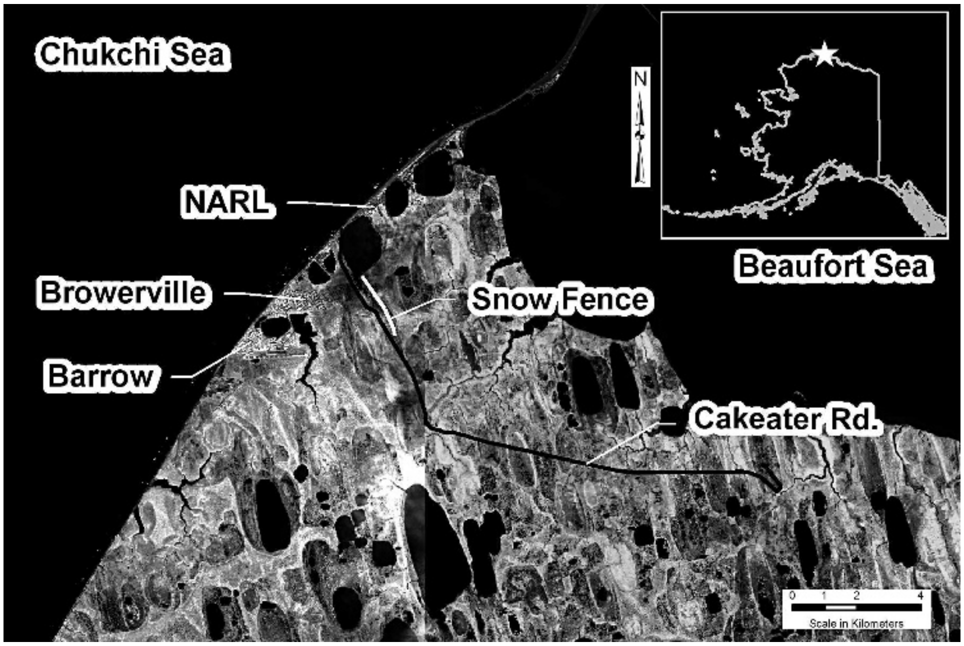

FIGURE 1. Location of 2.2-km-long snow fence east of Barrow, Alaska. Winter winds are primarily easterly.

downward soil heat flux is enhanced (Farouki, 1981). Similarly, thaw beneath ponds tends to be deeper (Hinkel and Nelson, 2003). Thus, indirect effects include localized ponding and thaw, which promotes the development of thermokarst.

The long-term influence of deep snow drifting on the ground thermal regime remains unclear. Our experience at sites in northern Alaska suggests that, in regions where the substrate is ice rich, deep drifting tends to promote destabilization of the upper permafrost and results in ground subsidence and thermokarst. The snow fence at Barrow provided an opportunity to monitor the soil thermal impact from near the inception of the disturbance.

In autumn 1997, a snow fence was constructed east of the village of Barrow (71°20′N, 156°40′W). The purpose was twofold: to alleviate snow drifting in the Browerville neighborhood that often partially buried homes, and to encourage residential development along Cakeater Road near the snow fence.

The snow fence is located 200 to 300 m east of Cakeater Road (Fig. 1) and is designed to capture snow driven by the prevailing easterly winter winds. The wooden plank structure is about 4 m (16 ft) high and extends approximately 2.2 km from Middle Salt Lagoon south to near Footprint Lake. The fence was installed in October following soil refreezing and establishment of a snow cover to prevent damage to the tundra.

Prior to the onset of this study, anecdotal information suggested that a large drift developed downwind (west) of the snow fence in winter 1997–98, extending 100 m from the fence for nearly its entire length and reaching the height of the fence (Fig. 2a). A smaller drift developed upwind (east) of the fence. The lee drift was still present the following June, although the surrounding tundra was snow free. The drift resisted ablation and persisted until mid-July. In August, it appeared that the tundra beneath the melted snowdrift was significantly wetter and greener than the surrounding unaffected tundra.

Background and Study Area

Barrow is a coastal village located on the Chukchi Sea with a cold maritime climate. The mean annual temperature is −12°C, with July the warmest month (4.7°C) and February the coldest (−26.6°C). Mean annual precipitation is 106 mm, with 46% falling as rain in July and August. During the long, dry winters, snow covers the ground to a mean maximum depth of 36 cm (NOAA, 2005). At Barrow, the ground is underlain with continuous permafrost to a depth of around 400 m. The active layer is typically 40 cm deep. The silty substrate is extremely ice rich, exceeding 70% in the upper 2 m (Brown and Johnson, 1965), and is highly susceptible to thaw subsidence.

The study site is located near the southern end of the fence, in a drained thaw lake basin dominated by low-center ice-wedge polygons and wet meadow tundra vegetation. Here, the local slope is less than 0.5° but locally may occasionally reach 1.0°. By contrast, the northern two-thirds of the fence is near ravines draining into Middle Salt Lagoon; slopes here are much steeper and vary from 1 to 32°.

Methods

In August 1999, three transects were established in the primary study area near the southern end of the snow fence. Transects run perpendicular to the fence and are separated by 50 m, as shown in Figure 3. Each transect extends 60 m on the western (leeward) side of the fence and 50 m on the eastern (windward) side. Nine copper pipes were drilled into the permafrost and allowed to refreeze; they serve as frost-defended bench marks near the end of each transect and at the fence. A measuring tape was placed between the bench marks to measure distance. The origin is at the snow fence; distance to the east (windward) is negative and to the west (leeward) is positive. Measurements of thaw depth and snow thickness were often extended beyond the limits of the study area along the transects.

In addition, a control area was established on the open tundra about 300 m south of the snow fence. Here, two transects were laid out along cardinal directions. The north-south transect is 100 m long while the east-west transect is 110 m.

GROUND- AND AIR-TEMPERATURE MEASUREMENTS

In August 1999, temperature recording data loggers were installed at nine sites near the snow fence, and at three sites in the

FIGURE 2. (a) Leeward snowdrift at southern end in May 2001 looking north; (b) aerial view of southern end of snow fence in mid-August 2001; (c) view of ponded study area west of snow fence looking north; (d) fragmented surface organic mat in 2004.

study control. At each site, three thermistors were installed at depths of 5, 30, and 50 cm below the surface. Loggers were encased within a sealed PVC container and positioned several meters from the transects to prevent disruption of the ground surface. The location of the logger containers on the study site are represented by letters in Figure 3.

Thermistors were attached to a thin wooden dowel embedded in the permafrost; this ensured that they would not be displaced by freeze-thaw effects. The lower two thermistors were near the active layer–permafrost interface where summer temperature hovers around the freezing point. For this reason, customized Hobo Stowaway® data loggers (Onset Computer, Inc.) with a temperature range of $-30°$ to $+5°C$ were used. These loggers have a nominal precision of 0.14°C. The near-surface thermistor experienced greater temperature extremes so a standard Hobo Stowaway® data logger was employed with an operating range of $-37°$ to $+46°C$ and resolution of 0.32°C. Loggers were synchronized to record temperature every 2-h. Following a period of several days to allow the thermal disturbance of installation to dissipate, the record begins at midnight on 21 August 1999 and extends through 7 August 2005. In addition, hourly air temperature data was available from a nearby (~1 km) site.

During the study period, several loggers were destroyed by animals or acts of vandalism. In some cases, the thermistor leads broke (usually during soil refreezing) so the record is discontinuous or ended. Temperature traces from six sites are shown in

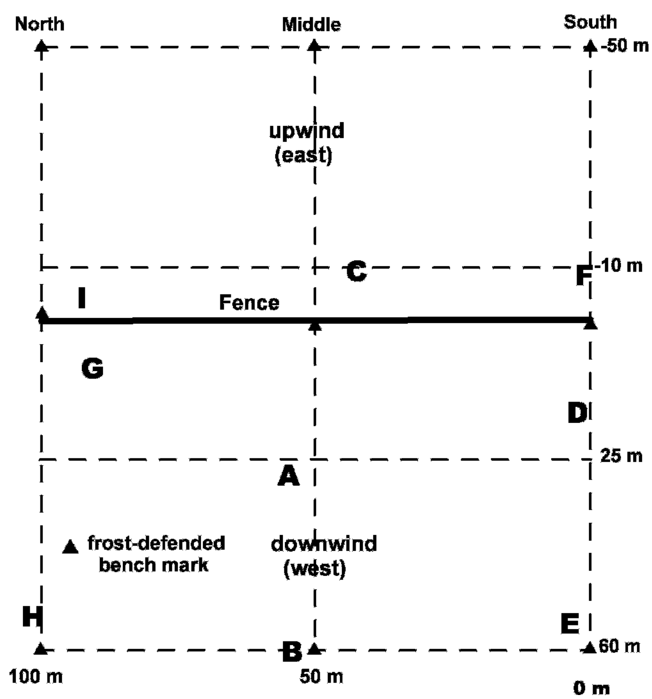

FIGURE 3. Study area near southern end of snow fence showing north, middle and south transects. Letters indicate location of thermistor canisters; triangles show position of frost-defended bench marks.

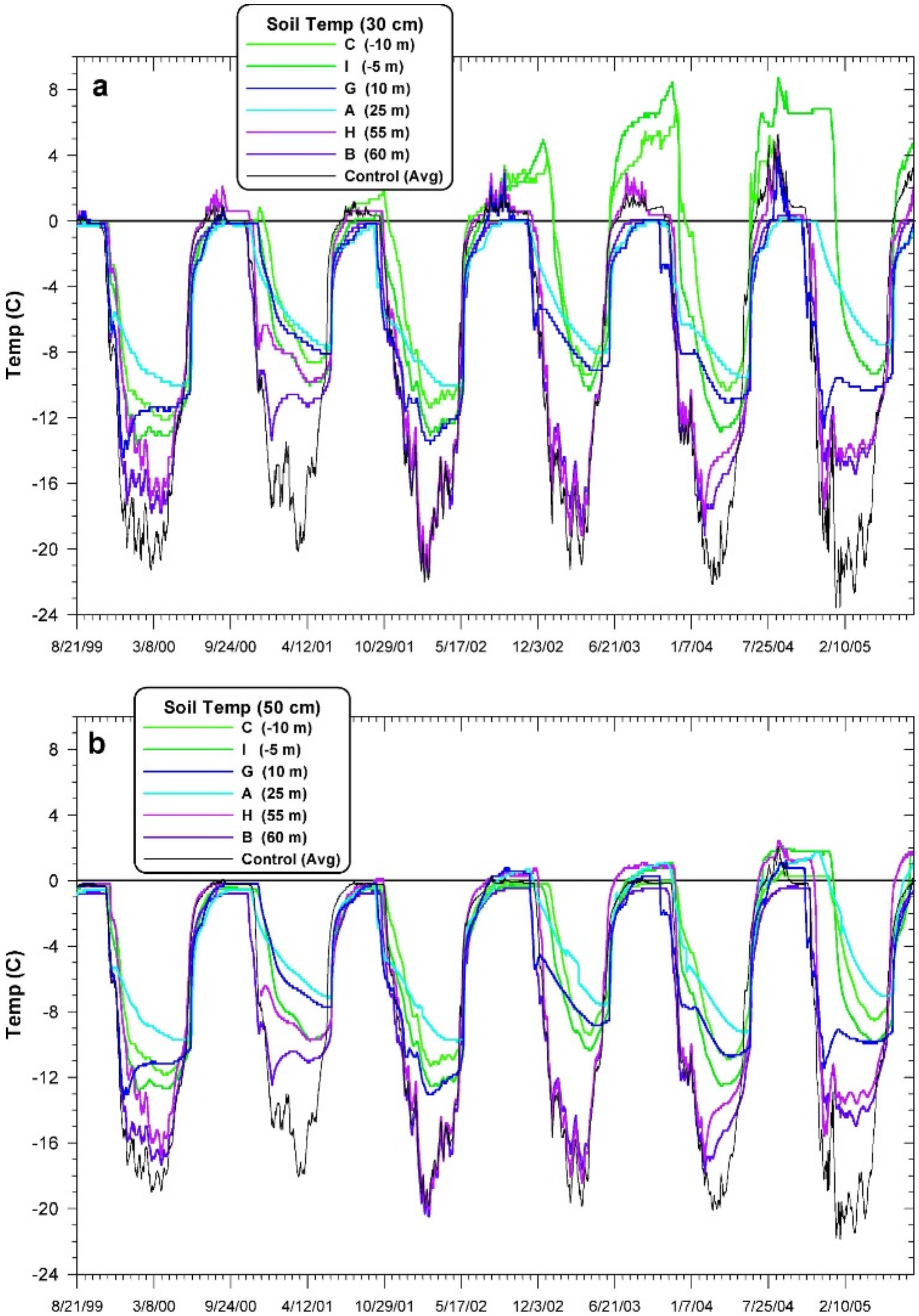

FIGURE 4. Soil temperature time series from (a) 30-cm and (b) 50-cm depth for six sites and average Control over period of record (21 August 1999 to 7 August 2005). Tick marks at 20-d interval. Note that some records are discontinuous or are truncated.

Figure 4a for the 30-cm depth and in Figure 4b for the 50-cm record. Soil temperature at these depths are nearly identical to the temperature at the 5-cm level during winter when diurnal forcing is completely attenuated. In summer, the soil temperature at 30 cm is near the daily thermal damping depth. Thus, the traces in Figure 4 show the effects of synoptic-scale variation superimposed on seasonal trends.

The records from the three control sites were very similar, so the average Control for that depth is plotted on the figures.

PROBING ALONG TRANSECTS

In April or May of years 1999–2005, with the exception of 2000, snow depth measurements were made along the middle transect. A 1-cm-diameter modular probe was used; it has 1-m sections that can be added as necessary to a maximum length of 4 m. Snow depth was recorded at least every 5 m along the transect, although the sampling frequency was increased near the snow fence where gradients were steep. These data are presented in Figure 5 (upper).

Every mid-August from 1999 through 2004, ground thaw depth was measured at 2.5 m intervals along the transects using a graduated metal probe. An example from the middle transect is shown in Figure 5 (lower) for the years 1999 and 2004. Open symbols indicate the presence of standing water at the surface; these are typically ponds. Note that thaw depth is somewhat shallower beneath the drifts and is enhanced in the scour zones near the fence and to the lee of the large drift. In places, the ponds were too deep to obtain measurements.

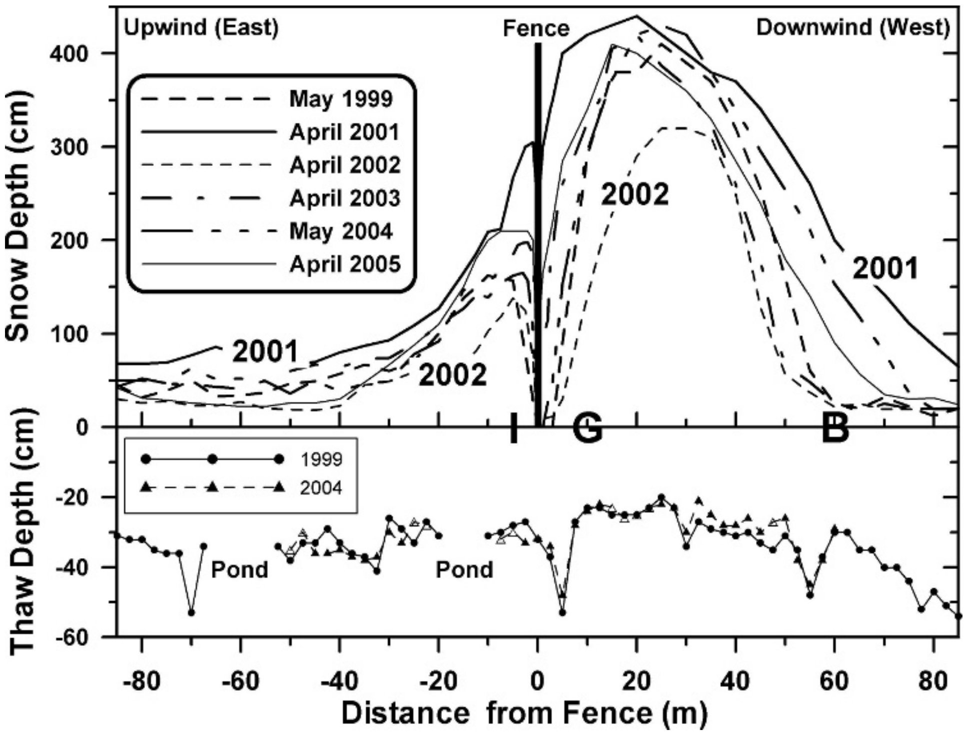

FIGURE 5. (Upper) Snow depth along middle transect with east (upwind) to the left; (lower) ground thaw depth along middle transect for 1999 and 2004. Open symbols indicate the presence of standing water.

MEASURING SUBSIDENCE ALONG THE TRANSECTS

Probing is a reliable method for obtaining ground thaw depth (Brown et al., 2000). However, measurements are made relative to the ground surface. The surface elevation fluctuates seasonally with freezing and thawing, and associated ground heave and subsidence. Locally, the surface may undergo permanent upward displacement due to the formation of ice lenses at depth. Similarly, an increase in heat flow to depth may result in local deepening of the active layer and thawing of the ice-rich upper permafrost. This causes long-term thaw consolidation and subsidence of the ground surface which cannot be detected by probing.

To monitor ground subsidence along the transects, a laser survey was conducted in August 2000. On each transect, the laser was mounted atop the bench mark near the fence and sighted on targets atop the bench mark at the far end of the transect. A stadia rod was used to measure the distance from the laser beam to the ground surface at 2-m intervals. In this way, a *relative* topographic profile was created and serves as the initial surface from which to monitor heave and subsidence effects.

In 2004, a differential global positioning system (DGPS) was made available through the Barrow Arctic Science Consortium. The system consisted of two Trimble® 5700 receivers, one for the base station located at the Naval Arctic Research Laboratory some 5 km from the study site, and one as a roving unit. The roving unit also includes a Trimble Survey Controller® and a Two Meter Stick. Both the base station and the rover were equipped with radio antennas and receivers. A radio relay was set up near the study site to strengthen the signal and to minimize interference or delay in base-to-rover communication. To collect GPS points along the transects, a real-time kinematic survey method was utilized to obtain instantaneous resolution. The DGPS survey posted a horizontal resolution of <1 cm and a vertical resolution of <1.5 cm.

In August 2004, the bench marks on each transect were re-occupied with the DGPS; the exact position of the laser and targets were determined along with the height of each bench mark. A linear interpolation was used to locate the position and elevation of the laser beam at the time of the initial survey in 2000. In this way, the nine bench marks were used to convert the relative laser survey elevations to absolute elevations (m a.s.l.) and create accurate topographic profiles. Replicate measurements were made at every site recorded in 2000, and the two data sets were standardized to sea level as shown in Figure 6 (upper). Changes in the ground surface elevation (Fig. 6, lower) over the 4-yr period are due to ground heave or subsidence.

The DGPS was also used to verify snow depth readings in April 2005. At each location along the middle transect, replicate snow depth measurements were made by probing and using the DGPS. The difference between the elevation of the snowdrift and elevation of the ground surface as measured the previous August was calculated to yield the snow thickness. The results compared extremely well ($\pm$8 cm) with snow depth as measured by probing.

Results

The results from snow-depth measurements reveal significant interannual variation in drift size, although every year a very large drift formed (see Fig. 5). The downwind drift crest was typically at the same height as the snow fence (4 m) with the crest 10 to 30 m downwind. Analysis of the temperature data from the 5-cm level (not shown) shows that this drift usually persisted 6 to 8 wk after snow meltout on the tundra, although the massive drift that formed in winter 2000–01 never completely ablated (Fig. 2b). A smaller drift formed upwind each winter and was typically 1.5 to 2.0 m in height. It crested about 5 to10 m from the fence. In most years, a narrow scour zone developed at the base of the fence (Fig. 5), yielding a steep snow face on the lee side. Another scour zone often developed to the lee of the large drift.

Table 1 presents summary results of snow thickness on the middle transect, and ground thaw depth on the three transects and control. The column labeled "Area Drift" is simply the area (m^2) under the snow thickness curve from -85 m to +85 m—a region that encompasses most of the drift. Maximum drift size occurred

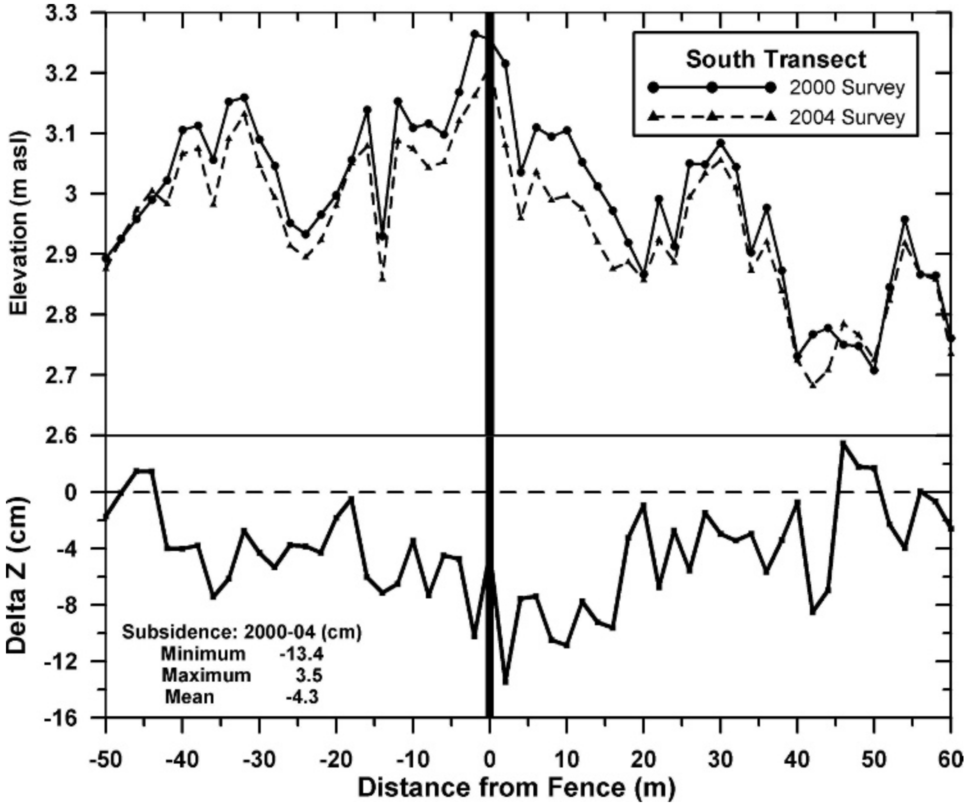

FIGURE 6. (Upper) Topographic profile (m a.s.l.) along south transect from surveys in August 2000 and August 2004; (lower) difference in elevation along transect (Delta Z) over the 4-yr period showing general ground subsidence.

in winter 2000–01 and the smallest drift formed the following winter. To enable interannual comparisons, drift size is standardized to the maximum (2000–01) and the results are presented as a percentage in the table. Thus, the cross-sectional area of the drift in winter 2001–02 was only 43% of the drift that formed the previous winter.

In addition, the average snow depth on the control is recorded in Table 1. Note that there is no correlation between drift size and snow cover thickness on the unaffected tundra; Hinkel et al. (2003) noted that drift aggradation is largely related to the occurrence of blizzards. The impact of the snow fence can be estimated by calculating the area under the curve for a constant average snow depth across the same distance (170 m). The ratio of drift area to control area represents the footprint of the snow fence for that year, and ranges from 3.2 to 7.0 with a mean value of 4.7. In a sense, this is a measure of the snow-fence efficiency; snow depth in the fence area is about five times the thickness of snow on the open tundra.

Average ground thaw depths along the transects demonstrate considerable interannual variation. However, these data must be treated with caution since they were collected at different times in August. Typically, maximum thaw depth occurs in mid-August, so measurements made early in the month do not represent active layer depth. In comparing the average thaw depth on the transects to the control, there is no obvious correlation.

GROUND-TEMPERATURE MEASUREMENTS

Summary mean annual ground temperature (50 cm) statistics for all sites near the fence are shown in Figure 7 as box-and-whisker plots. These are organized in the same arrangement as in the field, with east to the top of the figure. Since Unit F was destroyed by animal activity, the mean Control is plotted in the upper right corner. The median, quartiles, and outliers are shown for each annual cycle (1 Jan–31 Dec) using a common *y*-axis scale. In addition, the mean value is plotted as a dot and labeled. A large disparity between the mean and median values suggests a non-normal distribution.

Data for sites in the middle tier (Sites G, A, and D) reflects the thermal effects of the large downwind drift. The temperature

TABLE 1

Summary statistics of snow depth measurements along middle transect and control, and ground thaw depth on all transects.

Snow Depth-Middle Transect							Ground Thaw Depth					
Winter	Date	Area Drift (m^2)	%Max (00–01)	Control (cm)	Area Cont (m^2)	Area (drift/cont)	Summer	South (cm)	Middle (cm)	North (cm)	Avg (cm)	Control (cm)
1998–1999	14-May	21773	63	40	6800	3.2	1999 (17 Aug)	38	34	36	36	33
1999–2000	no data	—	—	—	—	—	2000 (16 Aug)	29	25	25	26	32
2000–2001	21-Apr	34678	100	29	4964	7.0	2001 (10 Aug)	26	26	22	25	28
2001–2002	11-Apr	15024	43	24	4012	3.7	2002 (11 Aug)	34	25	28	29	31
2002–2003	18-Apr	22798	66	31	5338	4.3	2003 (6 Aug)	23	20	19	21	26
2003–2004	20-May	25958	75	33	5627	4.6	2004 (13 Aug)	37	30	32	33	44
2004–2005	15-Apr	24444	70	27	4505	5.4	no data	—	—	—	—	—

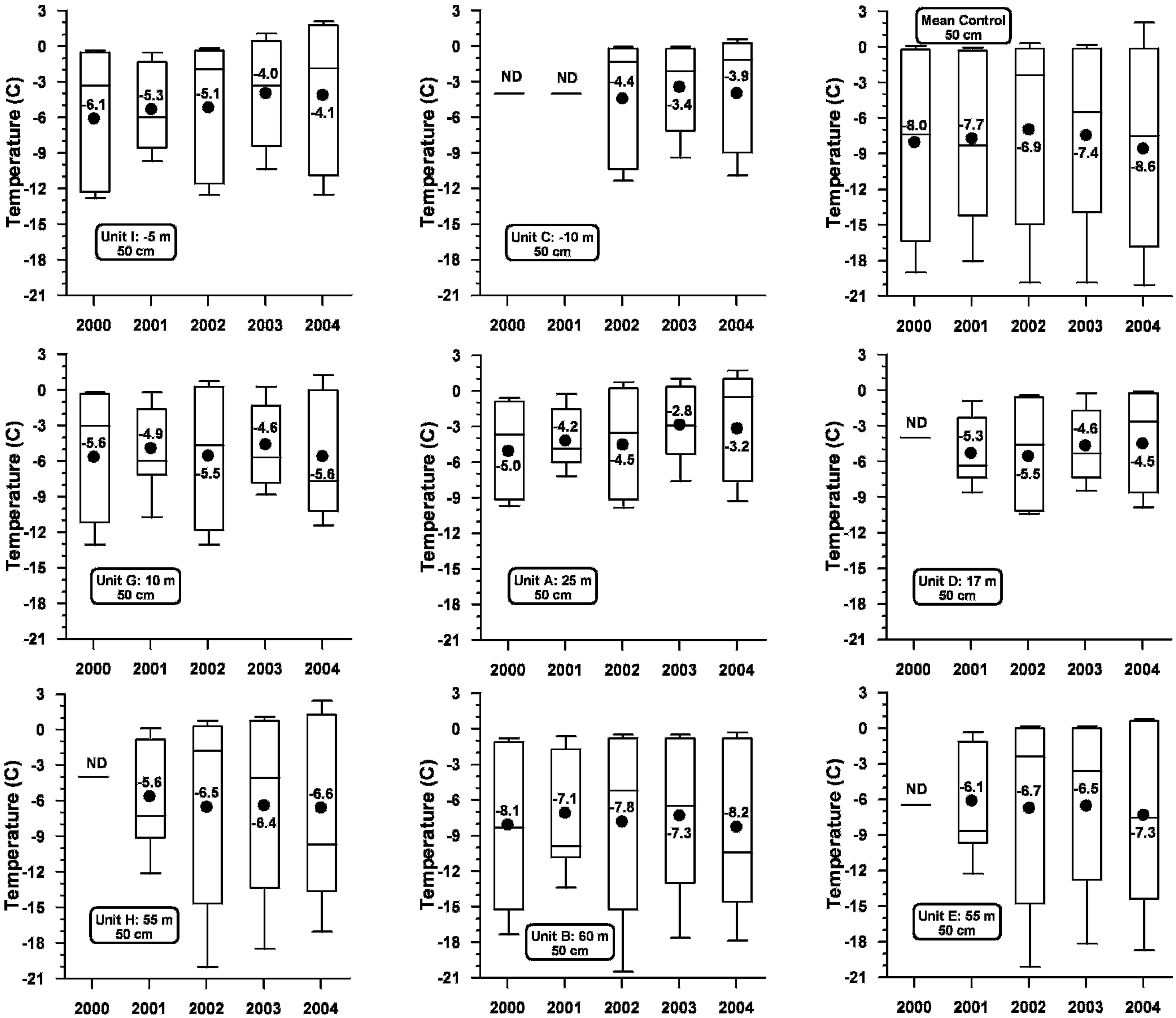

FIGURE 7. Summary descriptive statistics of soil temperature at 50-cm depth showing median, quartiles, outliers, and mean (dot with value) for the calendar year. Figures are identical to placement of canister units in the field, with north to the left. Control (average) shown in upper right panel. Note the common *y*-axis scale.

range is greatly constricted relative to the Control; average and minimum temperatures are substantially warmer. From Figure 4, it is clear that Sites A and G are strongly and consistently impacted. The pattern of gradual soil cooling ends abruptly in late spring with extremely rapid soil warming following drift ablation in early to late July—some 6 to 8 wk after snow has disappeared from the open tundra. Thaw penetration in summer is limited to the shortened snow-free season. However, the magnitude of this summer thaw increases over the period of record and is apparent in the summary statistics and temperature traces.

The magnitude and temporal pattern of subsurface warming can be verified by comparing time series of affected sites to the Control. This is done by subtracting the average Control from the site-specific temperature at 30 cm, and plotting the trace as Delta T. This is shown in Figure 8 for four representative sites; the plot for the 50-cm level (not shown) is very similar but less detailed. Positive values indicate that site-specific soil temperature is warmer than the Control. The degree of winter warming is mitigated by interannual variations in snow thickness and the site location. The traces for Sites A and G are 6 to 14°C warmer than the Control throughout much of the winter, and Delta T increases in magnitude each winter. In spring, the large negative Delta T values reflect subfreezing soils temperatures at sites still covered with drifted snow; the Control temperature is rapidly warming. Slightly negative Delta T values in summer indicate retarded thaw penetration and reduced soil warming by several degrees.

A similar pattern is observed at the two sites (C and I) upwind from the fence (upper tier in Fig. 7). Winter temperatures are initially very similar to those found beneath the larger, deeper downwind drift. Over time, however, two patterns emerge. First, both sites show a general increase in mean, median, and extreme temperatures. Second, the duration of the thaw period increases, causing a significant delay in soil freezing in winter; this is apparent as a shift in the cooling traces for these two sites in Figures 4a and 4b.

Site I has demonstrated a consistent increase in Delta T since summer 2003 (Fig. 8). Soil warming has progressed much faster than other sites, and warming is accelerating. Soil freezing was delayed well into winter 2004–05, causing the large positive spike (27°C). Soil at depth is about 5°C warmer than the Control

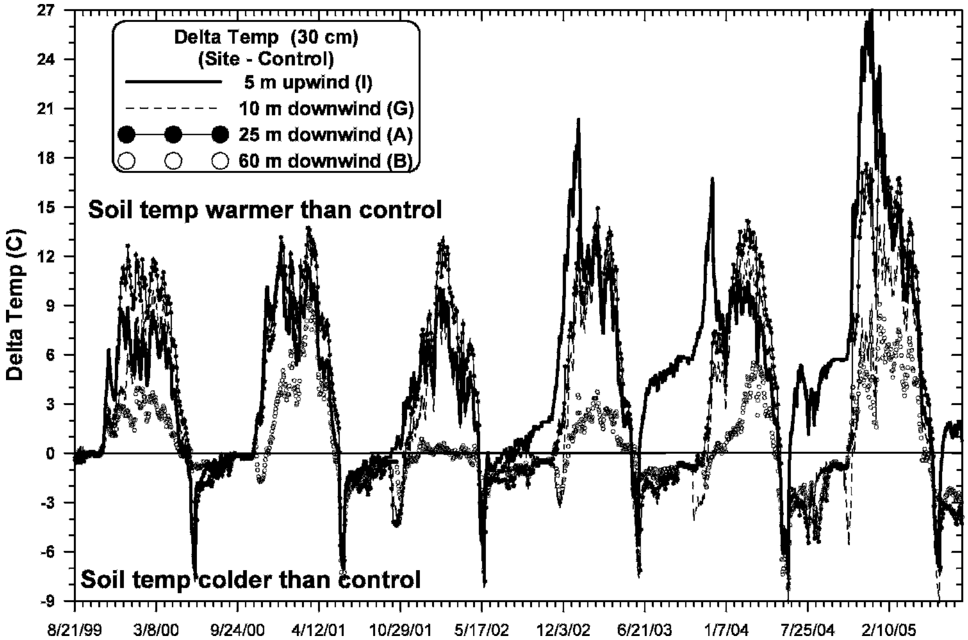

FIGURE 8. Difference in soil temperature (Delta T) between Control (average) and four sites for period of record at a depth of 30 cm. Tick marks at 20-d interval. Traces have been slightly smoothed.

throughout the spring and summer. As is apparent in Figure 4a, the same pattern is true for Site C. These signatures reflect the impact of pond expansion and encroachment during the period, resulting in a permanent shift in the soil thermal regime.

The effect is less pronounced at sites farther downwind (bottom tier in Fig. 7), which can experience deep snow accumulation in some years (2001) and scouring in others (2002) because they are typically located beyond the drift margin. However, when the drift extends downwind as in winter 2001–02, the soil temperature is up to 8°C warmer (Fig. 8). Note, however, that temperatures are 2 to 6°C warmer than the Control in the last three winters, and that soil at the 50-cm depth has consistently thawed each summer since 2002 (Fig. 4b).

Discussion

This study was developed with the assumption that the primary direct impact of the snowdrift would be to attenuate heat flow between the atmosphere and the permafrost. The objective was to determine if permafrost would become destabilized given the conflicting effects of soil warming in winter due to enhanced snow accumulation, and a reduced thaw period in the shorted summer. We assumed that there was little or no surface disturbance induced by snow fence construction and, indeed, no evidence of such was observed.

The overall impact can be assessed by calculating the change in the average temperature over the 6-yr period at depths of 30 and 50 cm, as shown in Table 2. The mean temperature at the Control was about −8°C at both depths. All sites near the fence demonstrated warming, but the magnitude varies with position relative to the fence. Upwind sites C and I have experienced maximum impact (3–5°C) due to the combined effects of reduced winter heat loss and pond encroachment. Sites G, D, and A are beneath the large drift and show average warming of 2.5 to 4.4°C over the 6-yr period. This is largely due to the insulating impact of the long-lasting drift. The three sites located beyond the drift margin (E, H, and B) show moderate to minimal (0.3–2.1°C) warming. Here, the effect is due to the sporadic influence of drift thickness annually, or to the influence of ponding near the site (e.g., Site E). It should be noted that this method provides a conservative estimate since the nearly 6-yr-long record is short by 2 wk in August when warming is maximized.

The impact of soil warming and thaw penetration is apparent in Figure 4b. Initially, all temperature traces at 50-cm depth show subfreezing maximum summer temperatures. Over time, most of the traces demonstrate summer thaw, which becomes progressively more pronounced. These patterns are not related to ambient conditions since the Control soil temperature does not show warming, and indicates permafrost degradation at sites impacted by drifting.

The spatial variability in the response to the ground thermal regime also became apparent over the course of this study. There were several converging lines of evidence to suggest differential thaw subsidence. It was clear that the distance between the ground surface and the base of the vertical wooden fence planks was increasing over time. Similarly, rust lines developed on the vertical steel support posts where they came into contact with the wet soil; over time, this rust line appeared to migrate upward, implying that the ground near the fence was subsiding.

It also became obvious that preexisting ponds were becoming larger and deeper, and new ponds were developing. Fed by snow meltwater, some ponds merged while others became too deep to measure thaw depth (Fig. 2c). Annual refreezing of ponded water

TABLE 2

Average temperature warming (°C) over the ~6-year period as it varies with site location and distance upwind (−) or downwind (+) from the fence. The average temperature for the Control is shown in the first row.

Site	Location (m)	30 cm (°C)	50 cm (°C)
Control	**NA**	**−8.1**	**−8.0**
C	−10	5.1	3.7
I	−5	4.9	3.1
G	10	2.5	2.6
D	17	4.1	3.3
A	25	3.7	4.4
E	55	2.1	1.7
H	55	1.7	1.5
B	60	0.5	0.3

disrupted the structural integrity of the surface organic mat, and freeze-thaw action reduced it to small clumps and fragments (Fig. 2d) that had little insulating value. In many places, the wet meadow tundra vegetation died to leave the surface barren or only sparsely vegetated with the moss *Polytrichum commune*. It appeared that permafrost destabilization was also resulting from indirect effects related to water ponding, and this was borne out by the DGPS ground survey in August 2004.

Along the northern and southern transects, subsidence was clearly evident from the analysis of the surveyed elevation profiles (Fig. 6, lower). Along the southern transect between 2000 to 2004, the average subsidence was 4.3 cm. Maximum subsidence occurred near the fence directly beneath the upwind and downwind drifts, with greater ground subsidence beneath the larger downwind snowdrift. The northern transect demonstrated similar results with a net average subsidence of 6.1 cm; subsidence exceeded 30 cm near pond margins in some places.

Along the middle transect, there was no average change (0.1 cm) in surface elevation. However, near the fence are localized areas of subsidence of about 10 cm, whereas other areas had up to 6 cm of heave (Fig. 5, lower). It therefore appears that thermokarst and thaw subsidence were widespread, though not spatially uniform.

As a final note, a meeting was held in August 2004 with an engineer of UIC Construction, Inc., the contractor for the Cakeater Road snow fence. The purpose was to determine the original design parameters and rationale. The vertical steel support posts are 9.1 to 10 m (30–33 ft) in length. They were drilled 5 m (16 ft) into the ground and anchored in the permafrost for heave protection. The design calls for the base of the vertical wooden planks to be 36 to 46 cm (14–18 in) above the ground surface; this is the optimal spacing to induce drift formation. An upwind drift must first form to decrease wind velocity beneath the fence sufficient for the downwind drift to form. When the gap exceeds 60 cm (24 in), wind velocity is too fast to allow effective development of the downwind drift.

To determine the degree of subsidence along the fence, 30 ground-to-plank distance measurements were collected in the study area in August 2004. The average was 68 cm—significantly greater than the maximum design gap height of 46 cm. A second set of 30 measurements was collected in similar flat ponded terrain just north of the study site; the average gap height was 57 cm.

At three other locations in the middle and northern reaches of the fence, 30 gap heights were measured and averaged 46 cm; these data sets are statistically different from those collected near the study area the 0.05 significance level. Here, the ground slope is sufficiently steep ($>4°$) to transport snow meltwater away from the fence, and there was little ponding and no evidence of widespread subsidence.

Conclusions

This study was concluded eight years after the construction of the snow fence in autumn 1997. Based on nearly six years of data (1999–2005), the impact of drifting is summarized below:

(1) A 4-m high snowdrift forms each year on the downward (western) side of the fence, and a smaller (1.5–2.0 m) drift forms upwind. The width of the drifts varies interannually and is not related to snow-cover thickness on the undisturbed tundra.
(2) The thermal regime beneath the drifts is strongly modulated. Soil temperatures near the top of the permafrost at the 30- to 50-cm depth range from 2 to 14°C warmer than the control in winter, and depends on snow thickness and location of the monitoring site.
(3) The drifts persist 4 to 8 wk after snow has melted from the open tundra. This delays the onset of soil thaw and limits soil warming in summer.
(4) The mean soil temperature near the top of the permafrost beneath the drifts increased an average of 2 to 5°C over the 6-yr period of record, and permafrost is degrading. The effect is widespread and appears progressive over time.
(5) There is a great deal of snow meltwater that ponds in the study area. Vegetation beneath the drift has died in many places, and the organic mat is fragmented. Preexisting ponds have become larger and deeper over time, and new ponds have developed. This is apparent in the thermal records of some affected sites, and through measurements of ground thaw depth and subsidence. These indirect effects appear to have enhanced thermokarsting.
(6) Ground subsidence demonstrates large spatial variability across several scales. There is general subsidence near the fence of at least 10 to 20 cm. Most subsidence appears to be confined to the southern end of the fence where the terrain is flat and the drainage is poor. The middle and northern sections are well drained, lack ponding, and demonstrate no evidence of subsidence.

The Cakeater Road snow fence is one of several around Barrow. Others exist and are being constructed in northern Alaska to protect villages or enhance drinking water supplies. The snow fence contractor has been informed of these findings, and we have recommended that surface drainage lines be installed in the southern end of the snow fence prior to further residential development of the area.

Given the extreme depth of snow accumulation, the drift does not provide a realistic analog of shrub tundra. However, similar thermal and hydrological effects might be associated with human structures, the lee sides of topographic prominences, and in low-lying areas such as stream valleys and drained thaw lake basins.

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