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'Warm' Tundra: Atmospheric and Near-Surface Ground Temperature Inversions Across an Alpine Treeline in Continuous Permafrost, Western Arctic, Canada

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ABSTRACT

Permafrost conditions were examined between 2010 and 2014 at four sites across an alpine treeline on Peel Plateau, Northwest Territories, Canada. Ground and air temperature sensors were installed in forest and tundra between 30 and 500 m asl. Annual mean air temperatures increased and the number of freezing degree days declined with elevation, due to persistent winter air temperature inversions. The annual mean temperature at the permafrost surface (T_{ps}) in mineral soils increased with elevation from about −2.5 °C in lowland forest to about −1.5 °C in dwarf shrub tundra. The increase in T_{ps} coincided with higher air temperatures and earlier snow accumulation at tundra sites. The higher T_{ps} in alpine tundra compared to lowland forest in Peel Plain contrasts with the northward decrease in T_{ps} across latitudinal treeline elsewhere in the western Arctic. An increase in T_{ps} with elevation may be common in Arctic mountain environments due to the prevalence of atmospheric temperature inversions in winter. In such contexts, although vegetation characteristics are governed by summer climate, permafrost conditions are critically influenced by the winter regime. The tundra permafrost on Peel Plateau is considerably warmer and, hence, more sensitive to disturbance than perennially frozen ground north of treeline in other parts of the western Arctic. Copyright © 2015 John Wiley & Sons, Ltd.

KEY WORDS: permafrost; thermal regime; treeline; atmospheric inversions

INTRODUCTION

The forest-tundra ecotone presents significant variations in surface conditions that control the thermal regime of permafrost (Lewkowicz *et al*., 2012; Palmer *et al*., 2012; Roy-Léveillée *et al*., 2014). Changes to the ecotone are becoming apparent, as increases in shrub abundance have been observed near treeline (Sturm *et al*., 2001a; Mackay and Burn, 2011; Lantz *et al*., 2012). The interactions between physical and ecological conditions across treeline are likely to influence future changes in ground temperatures, which may in turn affect hydrology, terrain stability, ecosystem function and northern infrastructure.

The Dempster Highway in Canada's Northwest Territories (NWT) traverses the forest-tundra ecotone over 77 km from Peel Plain at about 30 m asl near Fort McPherson, to around 930 m asl at the Yukon/NWT border in Richardson Mountains (Figure 1). For 40 km, the road traverses Peel Plateau, reaching about 500 m asl. The road crosses an alpine treeline, as vegetation changes with elevation from open-canopy forest on Peel Plain to tall shrub tundra and dwarf shrub tundra near the top of Peel Plateau. The vegetation gradient is similar to that at latitudinal treeline, which occurs in the western Arctic uplands to the northeast of Peel Plain (Mackay, 1963; Lantz *et al*., 2010). The entire region lies in the continuous permafrost zone (Figure 1).

Permafrost temperatures decline significantly across *latitudinal* treeline (e.g. Burn and Kokelj, 2009, Figure 11) due to reduced snow cover, higher snow density and lower air temperatures over the tundra (Palmer *et al*., 2012). In a similar fashion, Lewkowicz *et al*. (2012) observed that permafrost temperatures decline with elevation across an *alpine* treeline in mountainous terrain of the Wolf Creek drainage, southern Yukon. Near Dawson City in central Yukon, the relations between permafrost temperatures and elevation are less clear due to the influence of more persistent atmospheric temperature inversions in winter (Lewkowicz *et al*., 2012, Figure 4), which are common at high latitudes (e.g. Burns,

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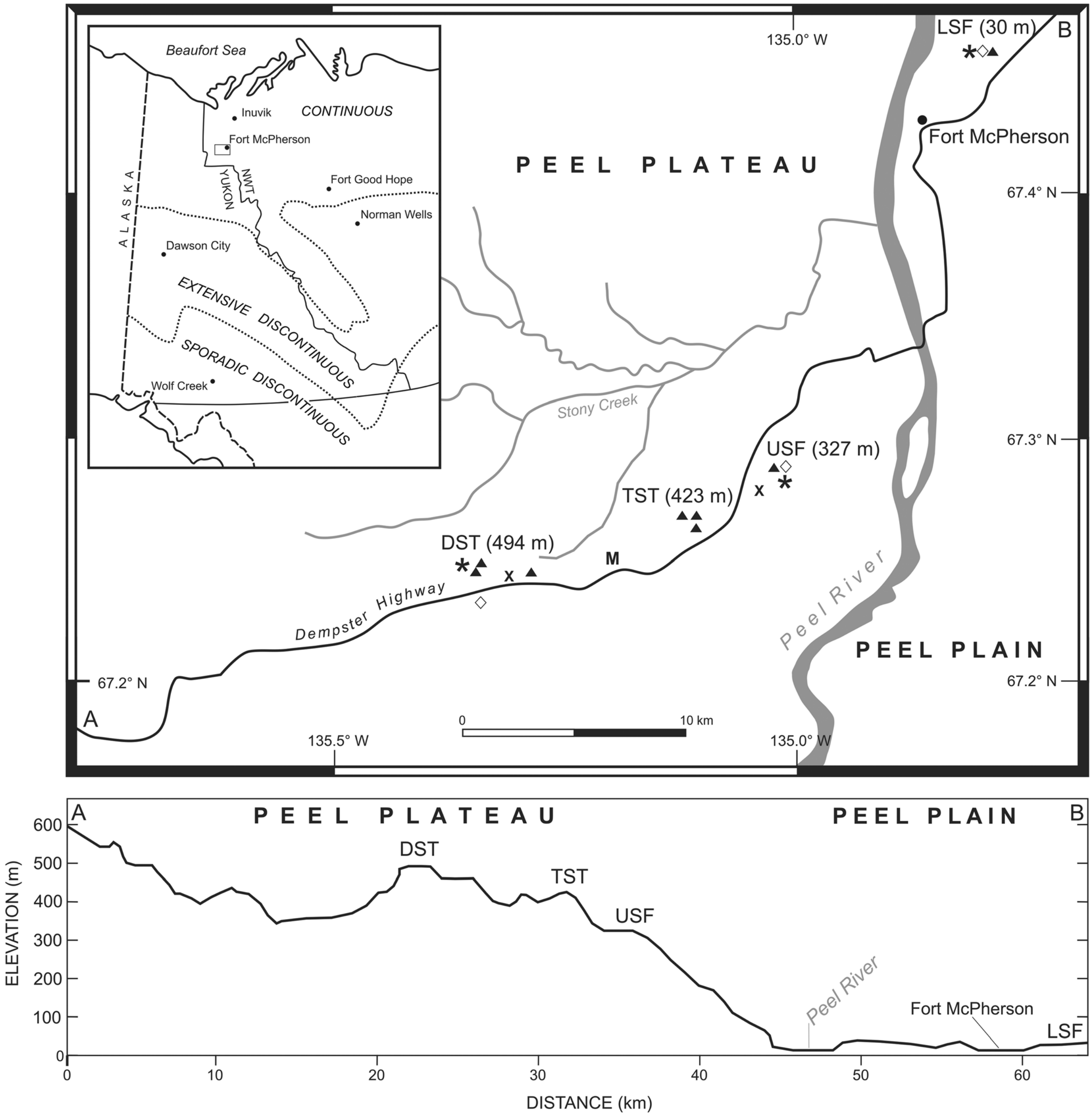

Figure 1 Location of study sites, and elevation profile between points A and B along the Dempster Highway (elevation profile modified from Google Earth V 7.1, 2013). Symbols are as follows: ▲ shallow (1 m) thermistor array; ◊ deep (4–7 m) thermistor array; * air temperature sensor; X iButton array; M meteorological station. The locations of instruments are approximate. The inset is modified from Burn (1994, Figure 1), and permafrost zones are after Heginbottom *et al*. (1995). The Dempster Highway begins near Dawson City and runs through Fort McPherson to Inuvik. See text for abbreviations.

1973; Wahl *et al*., 1987; Burn, 1993; Taylor *et al*., 1998; Bonnaventure and Lewkowicz, 2013). To our knowledge, no studies have examined permafrost conditions across an alpine treeline in the continuous permafrost zone.

This paper investigates ground temperatures at four sites across the ecological transition between the forest in the Peel Plain lowlands and DST on Peel Plateau. We assess the influence of air temperature, vegetation and snow conditions on the thermal regime of permafrost. We determine whether the pattern of declining ground temperatures across treeline observed by Palmer *et al*. (2012) and Lewkowicz *et al*. (2012) applies to the alpine treeline in the Peel Plateau region. Since the study region spans a considerable elevation range, we predict that permafrost temperatures will

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decrease less drastically from forest to tundra than at latitudinal treeline due to winter air temperature inversions. To test our prediction, we: (1) characterise environmental conditions at the study sites across an Arctic alpine treeline; (2) describe air and ground temperatures along this vegetation-elevation gradient; and (3) assess relations between snow properties, vegetation, air temperature and permafrost conditions at the sites.

PERMAFROST TEMPERATURES ACROSS TREELINE

In the continental continuous permafrost zone, snow cover is the most important surficial factor causing local ground temperature variations (Mackay and MacKay, 1974; Smith, 1975; Zhang, 2005; Burn and Zhang, 2009; Morse *et al.*, 2012). The thermal resistance (R_s, m^2 K W^{-1}) of the snow cover limits heat flow out of the ground in winter, and is controlled by the thermal conductivity (λ_s, W m^{-1} K^{-1}) and the snow thickness (ΔH, m):

$$R_s = \frac{\Delta H}{\lambda_s} \quad \text{(Lunardini, 1981, p. 43, Equation 3.9)} \quad (1)$$

The snow cover is thinner and denser in tundra than in forest due to sublimation and packing by strong winter winds (Mackay and MacKay, 1974; Sturm *et al.*, 2001b). It therefore offers low thermal resistance, facilitating ground heat loss. Palmer *et al.* (2012) reported thermal resistances from 2.4 to 9.5 m^2 K W^{-1} in tundra and 11.0 m^2 K W^{-1} in boreal forest north of Inuvik. In the forest, wind distribution of snow is limited, so snow depths are more uniform and the snow cover density is lower (Burn, 2004; Palmer *et al.*, 2012). The marked changes in snow conditions across the ecotone cause ground temperatures to be significantly lower at tundra sites than in forests (Palmer *et al.*, 2012). For example, mean annual permafrost temperatures in upland forests near Inuvik are −1 to −3 °C, but decline to around −6 °C in tundra north of latitudinal treeline (Burn and Kokelj, 2009, Figure 11). For infrastructure management and engineering purposes, tundra in the continuous permafrost zone is generally associated with cold (< −5 °C) permafrost temperatures (e.g. Johnston, 1981, p. 34; MacGregor *et al.*, 2010, p. 3).

Air temperatures also influence permafrost conditions across treeline. The northern limit of trees is predominantly determined by summer climate, but is also influenced by the availability of moisture (Elliott-Fisk, 2000). Treeline coincides with the 12 °C mean July isotherm in continental areas of North America, and is close to the 10 °C mean July isotherm in maritime locations (Walker, 2000). At latitudinal treeline near Inuvik, autumn and winter air temperatures are similar across the ecotone (Palmer *et al.*, 2012). Summer air temperatures decline across the vegetation gradient due to differences in net radiation (Palmer *et al.*, 2012) and onshore breezes blowing off the Beaufort Sea (Burn, 1997). Summer air temperatures also theoretically decline across an alpine treeline, as temperatures decrease with elevation at the environmental lapse rate (e.g. −6.5 °C km^{-1}). However, air temperature inversions are common in winter at high latitudes and, in mountainous terrain, strong inversions may develop during calm periods with a clear sky (Burns, 1973; Wahl *et al.*, 1987; Burn, 1993; Taylor *et al.*, 1998). For short periods in winter, valley temperatures in the Yukon may be −40 to −50 °C, and rise to −15 °C at 1000 m (Wahl *et al.*, 1987). As a result, the number of freezing degree days may decline with elevation, reducing winter ground cooling in higher terrain (Taylor *et al.*, 1998; Lewkowicz and Bonnaventure, 2011).

PEEL PLATEAU REGION

The study area lies along the Dempster Highway on Peel Plateau and Peel Plain near Fort McPherson (Figure 1). The region is within the Taiga Plains ecozone (Smith *et al.*, 2004). Peel Plateau consists of gently sloping terrain with terraces and rounded plateaus of low relief, incised by steep V-shaped stream valleys draining eastward from Richardson Mountains to Peel River. Peel Plain to the east is characterised by low relief, poor drainage and numerous peatlands (Smith *et al.*, 2004).

Peel Plateau is within the limits of the Late-Wisconsin advance of the Laurentide Ice Sheet (Fulton, 1995; Smith *et al.*, 2004). The ground surface was ice-free by about 11 800 ^{14}C years ago (Zazula *et al.*, 2009), and is covered by moraine, glaciolacustrine and glaciofluvial deposits overlying Lower Cretaceous marine shale and siltstone bedrock (Norris, 1984). The sediments on Peel Plateau are characteristically ice-rich (Heginbottom *et al.*, 1995), and may host thick layers of massive ground ice, which is apparent from the relatively high density of large retrogressive thaw slumps on slopes in the region (Lacelle *et al.*, 2010; Kokelj *et al.*, 2013). The substrate on Peel Plain near Fort McPherson also consists of ice-rich, fine-grained glacial deposits, underlain by fluvial silt, sand and gravel over shale bedrock (Heginbottom *et al.*, 1995; Norris, 1984).

The study region has a subarctic, continental climate with long, cold winters and short, cool summers. The mean annual air temperature at Fort McPherson (1987–2006) is −7.0 °C. Total annual precipitation averages 295 mm, with 148 mm falling as rain (Environment Canada, 2012). Precipitation is typically heaviest in late summer and early fall (Burn and Kokelj, 2009, Figure 3).

Mackay (1967) estimated that permafrost at a site on Peel Plateau 22 km due west of Fort McPherson was approximately 120 m thick. Permafrost thickness of 625 m has been inferred from geophysical data collected in unglaciated terrain near the Yukon/NWT border, but at lower elevations on the plateau the thickness is likely closer to 300 m (Smith *et al.*, 2004, p. 79). Near-surface ground temperatures reported on national-scale maps range from −2 to −5 °C (Heginbottom *et al.*, 1995; Henry and Smith, 2001).

The study area was selected because sites along the forest-tundra transition are easily accessed from the Dempster Highway. In addition, significant road maintenance has been

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necessary on Peel Plateau to address problems related to the thaw of near-surface ground ice near the road embankment. Characterisation of permafrost conditions may also inform planning of road maintenance.

STUDY SITES

Instrumented sites (Figure 1) were established in four vegetation units (Figure 2) to describe conditions across the transition from subarctic boreal forest on Peel Plain to dwarf shrub tundra at higher elevation on Peel Plateau. Active layer (n = 42) and organic layer thickness (n = 14) were measured along transects at each site. These data are summarised in this section to provide the environmental context and are discussed later. Most of the instruments were installed in fine-grained mineral soil, though two were located in peatland on Peel Plateau. The substrate at mineral sites was predominantly ice-rich silt and clay, though stones, fine sand and thin gravel layers were sometimes encountered during site installation. In peatland, the active layer and top of the permafrost were composed of organic matter.

Lowland Spruce Forest (LSF)

The first study site was located in flat ground at low elevation (~30 m asl) on Peel Plain. Vegetation consisted of an open canopy of black spruce (*Picea mariana*) and tamarack (*Larix laricina*) trees, with an understory of willow (*Salix* spp.) and dwarf birch (*Betula glandulosa*) shrubs and lower-lying ericaceous shrubs of Labrador tea (*Rhododendron subarcticum*), cloudberry (*Rubus chamaemorus*), blueberry (*Vaccinium uliginosum*) and crowberry (*Empetrum nigrum*) (Figure 2a). The hummocky ground at the site was covered by a continuous (10–30 cm; median 18 cm) layer of mosses and lichens. Active layer thickness ranged between 28 and 55 cm (median 40 cm).

Upland Spruce Forest (USF)

The USF site was located in flat ground at an intermediate elevation (~330 m asl), in a broad valley on the eastern slopes of Peel Plateau, near the maximum elevation of forest growth. The forest at USF was composed exclusively of black spruce trees and the stand density was less than that at low elevation on Peel Plain (Figure 2b). The understory and ground cover were similar to that at LSF, consisting of willows, dwarf birch, Labrador tea, blueberry, cloudberry and crowberry. The ground surface was hummocky and covered in a thinner, more variable layer of mosses and lichens (5–35 cm; median 9 cm) than that at LSF. Active layer thickness ranged between 30 and 88 cm (median 50 cm).

Tall shrub Tundra (TST)

The TST site was in a transition zone of rolling topography above USF, at around 425 m asl. The vegetation consisted of tall (typically > 40 cm) willows, dwarf birch and shrubs of Labrador tea, cranberry (*V. vitis-idaea*) and blueberry (Figure 2c). Some stunted spruce trees were present, and the ground was covered with a thin (< 10 cm) surface organic layer of mosses and lichens, except in the troughs of poorly defined hummocks, where the organic layer was thicker (up to 15 cm). The median organic layer thickness was 7 cm. Active layer thickness ranged between 32 and 140 cm (median 69 cm).

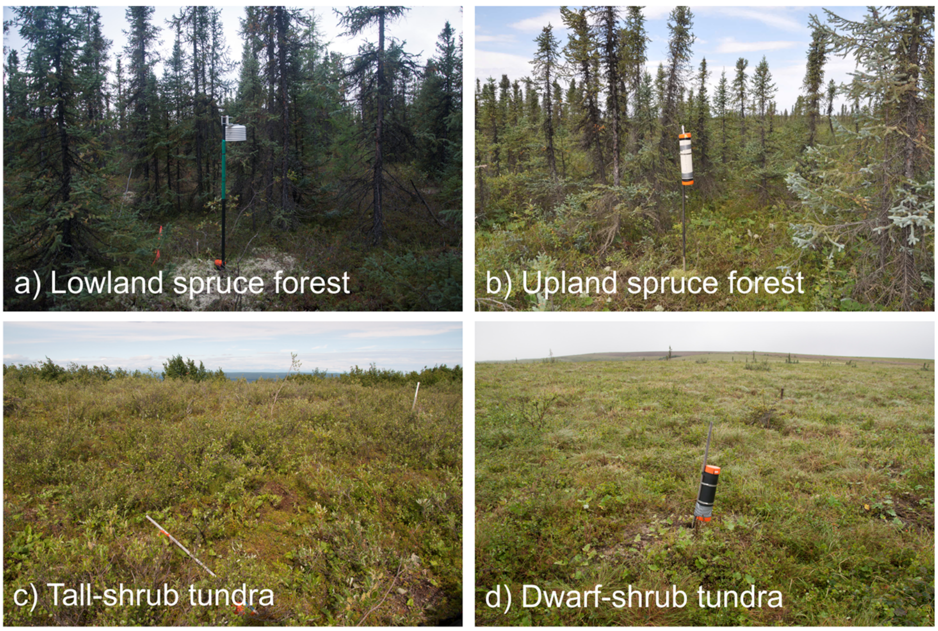

Figure 2 Study sites across the forest-tundra ecotone. This figure is available in colour online at wileyonlinelibrary.com/journal/ppp

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Dwarf shrub Tundra (DST)

The DST site was located at higher elevation on Peel Plateau, between ~ 470 and 495 m asl. The vegetation consisted of stunted willows, dwarf birch and ericaceous shrubs, typically < 40 cm tall (Figure 2d). Well-defined earth hummocks and tussocks covered much of the gently rolling ground. The organic layer was thin (<10 cm) on hummock tops, but greater in hummock troughs (up to ~ 30 cm). The median organic layer thickness was 8 cm. Active layer thickness ranged between 41 and 100 cm (median 67 cm).

Site-Scale Topography and Moisture Contents

All sites were in flat or gently sloping (≤5°) ground, and had similar soil moisture conditions. Active layer soil samples from 30 cm depth were collected from eight soil pits in mineral soil at USF, TST and DST. The samples were carefully extracted to yield a volume of 216 cm^3 (i.e. 6 cm x 6 cm x 6 cm). At LSF, the soil was organic material to below 30 cm depth. Volumetric water content (VWC) was determined by oven drying the samples. VWC was between 0.28 and 0.44 (mean 0.35; sd 0.05). The small variation in VWC suggests that moisture differences between sites did not likely have a profound influence on the ground thermal regime in mineral soil.

METHODS

Air and Ground Temperatures

Air and near-surface ground temperatures across the alpine treeline were measured every 2 h using thermistors (TMC6-HD or -HA, Onset Computer Corporation, Bourne, MA, USA) attached to HOBO data loggers (Onset, U12-006 or H08-006-04). These thermistors have a reported accuracy of ± 0.25 °C over a measurement range of −40 to 50 °C and a resolution of ± 0.03 °C at 20 °C. When used with the U12-006 logger, the precision of measurements is approximately ± 0.02 °C and about ± 0.45 °C with the H08-006-04 logger. Given the range of temperature fluctuations in the air and near-surface ground, the performance of the loggers was sufficient for assessing relative differences between sites.

Air temperature loggers (U12-006) were installed at LSF, USF and DST in radiation shields attached to steel pipes. The sensors were approximately 1.5 m above the ground surface. The air temperature sensors were in operation from August 2011 to August 2013, but the instrument at the USF site failed in autumn 2012. The annual values in Table 1 for 2012–13 were obtained by estimating daily air temperatures at USF using regression equations derived from 2011–12 DST and USF daily air temperatures for both the freezing and thawing seasons (thawing: USF = 0.98(DST) + 1.20; $r^2 = 0.99$; freezing: USF = 1.02(DST) - 0.28; $r^2 = 0.97$). These estimated values are presented in summary form but were not used to quantify the inversions, which were characterised using data from the LSF and DST sensors.

Relations between air and ground temperatures and spatial variation in ground temperatures across the treeline were studied using shallow thermistor arrays, consisting of four thermistor cables mounted on dowels and placed in holes drilled to 1 m. The sensors were typically at 5, 20, 50 and 100 cm depths. The 5 cm thermistor represents the temperature at the ground surface and the 100 cm thermistor approximates the temperature at the permafrost surface. In total, eight shallow thermistor arrays were installed (see Table 2). Three arrays were located at DST, three at TST and one each at LSF and USF. Snow depths were relatively uniform in the forest because the open canopy and stunted trees did not cause substantial tree wells, and wind distribution was minimal, so a single cable was deemed sufficient to describe ground thermal conditions at the LSF and USF sites. The shallow cables were installed in 2010 and 2011 (see Table 2). In TST, two of the shallow thermistor arrays were located in flat ground on a ridge in mineral soil, and another at the base of a north-facing slope below the ridge in a flat peatland (Figure 1). In DST, two of the shallow thermistor arrays were located at around 495 m asl, with one in a flat peatland and one on a gentle (2°) slope in mineral soil. Another was placed at 470 m asl in mineral soil, on a gentle (~5°), north-facing slope. The shallow thermistor cables located in hummocky, mineral soil in TST and DST were placed on hummock tops.

Ground temperatures at depths of 5–8 m were measured in order to obtain values less affected by seasonal variation than those at 1 m. These deeper ground temperatures were measured using thermistor cables with sensors (YSI44033, YSI Incorporated, Yellow Springs, CO, USA) spaced at

Table 1 Annual mean air temperatures (AMAT) and the number of freezing/thawing degree days (FDD/TDD) from three monitoring sites.

		2011–12			2012–13		
Site	Elev. (m)	AMAT (°C)	FDD	TDD	AMAT (°C)	FDD	TDD
LSF (Peel Plain)	30	−7.6	4418	1755	−7.8	4645	1782
USF (Peel Plateau)	326	−6.8	4092	1782	−7.5*	4361*	1619*
DST (Peel Plateau)	492	−6.8	3968	1327	−7.6	4257	1485

*Mean daily temperatures estimated using regressions of DST and USF temperatures from 2011–12. See text for other abbreviations.

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Table 2 Annual mean temperatures at the permafrost surface (T_{ps}) and ground surface (T_s), late-winter snow depths and the estimated dates of snow arrival.

Site	Elev. (m)	2010–11			2011–12				2012–13			
		Snow arr.	T_{ps} (°C)	T_s	Snow arr.	Snow depth (cm)	T_{ps} (°C)	T_s	Snow arr.	Snow depth (cm)	T_{ps} (°C)	T_s
LSF-1	30	Sep. 24	−2.6	0.1	Oct. 9	—	−2.8	0.0	Oct. 17	75	−2.5	−0.4
USF-1	327	Sep. 21	−1.7	0.0	Sep. 24	107[a]	−1.7	0.4	Oct. 17	99[a]	−2.7	−0.8
TST-1	423	Sep. 21	−2.1	−1.9	Sep. 24	62[a]	−2.2	−1.8	Oct. 17	74	−1.7	−1.9
TST-2 (peatland)	415	Sep. 21	−0.5	1.3	-	119[b]	-	-	-	97[2]	-	-
TST-3	423	-	-	-	n/a	62	−1.8	n/a	n/a	73	−1.4	n/a
DST-1	494	Sep. 21	−1.4	−0.5	Sep. 24	68	−1.2	0.2	Oct. 17	67	−1.5	−1.1
DST-2 (peatland)	492	Sep. 21	−2.3	−0.8	Sep. 24	37	−3.7	−2.5	Oct. 17	20	−4.4	−3.4
DST-3	479	-	-	-	Sep. 24	52	−2.0	−0.3	Oct. 17	48	−2.4	n/a

[a]Mean snow depth from an adjacent transect.
[b]Instrument could not be located, snow estimated by taking the average of ten probed locations in a wide circle around the site. See text for other abbreviations.

1 m intervals (1–8 m). The sensors have an accuracy of ± 0.1 °C. The cables were manufactured by M-Squared Instruments, Cochrane, AB, Canada, and were installed in holes drilled by water jet, with a steel pipe used to case the hole. Gravel layers limited the penetration of the water jet drill. The thermistor cables were connected to RBR data loggers that are accurate to ± 0.002 °C and have a resolution of < 0.00005 °C (model XR420 T8, Richard Brancker Research, Ottawa, ON, Canada), with measurements collected every 4 h. A deep cable was installed at LSF to a depth of 4.57 m, at USF to 4.91 m and at DST, near the top of Peel Plateau, to 6.74 m depth. In this paper, we report mean temperatures from the sensor nearest 5 m depth.

Snow Surveys and iButton Arrays

Snow surveys were conducted in February 2012 and March 2013 and 2014. Snow depths were measured in each vegetation unit with a graduated steel probe every 5 m along 70 to 75 m transects marked with PVC pipes to characterise snow cover in the vegetation units. The transects were located on flat or gently sloping ground. It was not possible to measure snow depths at LSF in 2012 due to logistical constraints. Snow depths were also measured at five points within a 1 m radius of shallow ground temperature sensors, and the average was used to estimate the snow depth at the sensor in order to examine the relation between the annual mean temperature at the permafrost surface (T_{ps}) and late-winter snow depth (LWSD). One snow pit was dug in each vegetation unit adjacent to the snow transects in 2013 and near each shallow thermistor cable in 2014 to determine snow densities and the thermal resistance of the snow cover. Snow samples of 100 cm^3 were collected at 10 cm intervals from the snow surface to the base of the snowpack, and the densities were averaged for each pit. The snow cover thermal resistance was calculated using Equation 1, with the thermal conductivity estimated using:

$$\lambda_s = 10^{(2.650\rho - 1.652)} \quad \text{(Sturm } et\ al.\text{, 1997, Equation 7)} \qquad (2)$$

where ρ is the snow cover density in g cm^{-3}.

Snow depth was also measured using arrays of iButton® miniature temperature sensors (model DS1921G, Maxim Integrated, San Jose, CA, USA) to investigate the influence of the timing of deep snow cover development in forest and tundra on ground thermal conditions. These sensors have a resolution of 0.5 °C and an accuracy of ± 1.0 °C between −30 and +70 °C. The iButtons were mounted on dowels: one sensor was located 5 cm above the ground surface and the rest were spaced at 10 cm intervals (10–100 cm), so that snow cover development could be monitored to ≥ 90 cm. The snow level was inferred semi-quantitatively, similar to Lewkowicz's (2008) method: if the temperature difference between two adjacent iButtons deviated persistently by > 1 °C, this suggested that the snow had covered the lower sensor. Plots of the iButton

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measurements from all sensors on the dowel were also compared visually with air temperatures to validate the patterns of temperature deviations with the inferred snow levels.

One iButton array was installed in forest in USF and another was located near the Dempster Highway in DST, adjacent to a snow fence where a deep snowdrift forms. The snow fence iButton site is not representative of average snow conditions in the tundra, but is analogous to locations on the tundra where deep snow accumulates from wind redistribution, such as at the base of slopes, in depressions or in areas with tall shrubs (Sturm *et al*., 2001b). Data from this iButton array were used solely to compare the timing of deep snow accumulation in the tundra with that in the forest. The Water Resources Division (Aboriginal Affairs and Northern Development Canada) established a meteorological (MET) station in 2010 on Peel Plateau, in forest dominated by tall shrubs, to measure air temperature, precipitation, net radiation, wind speed and snow depths (Figure 1). Snow depths inferred from the USF iButton array were compared with depths measured at the MET station. Maximum daily wind speeds at the MET station were used to determine when wind-driven redistribution of snow may have occurred. Snow depths were measured using an acoustic sensor (Sonic Ranger SR50A, Campbell Scientific, Edmonton, AB, Canada) and wind speed was determined using an RM Young wind speed monitor (model 05103, RM Young Company, Traverse City, MI, USA). The data were recorded on a Campbell CRX1000 logger.

Vegetation Heights

The ability of vegetation to trap snow depends on its height and structural complexity, which are closely correlated in shrub tundra (Thompson *et al*., 2004). Vegetation canopy heights were measured at ten randomly selected points within 3 m of the shallow thermistor arrays, as vegetation influences wind-distributed snow over distances of several metres (Sturm *et al*., 2001b). The ten canopy height measurements were averaged to obtain the mean maximum vegetation height, which was used to examine the influence of vegetation on snow depths at the shallow thermistor cables.

Active Layer Thicknesses

The thickness of the active layer was estimated by probing the ground with a graduated steel rod in mid-late August 2011 to 2013 at 5 m intervals along the established transects at each site. Active layer thicknesses were also measured at the two peatland sites in 2011–12, but were only measured at the DST peatland after the site at TST was damaged.

Statistical Analyses

The Wilcoxon rank-sum test was employed to examine differences in snow depths and active layer thicknesses amongst the site transects for each year. This non-parametric test was used because the data violated the normality assumption for parametric tests. Least-squares linear regression was employed to examine relations between (1) snow depth and vegetation height at tundra sites, and (2) T_{ps} and LWSDs for forest and tundra sites, respectively. All statistical tests were conducted at the 0.05 significance level. TST-2 was not included in the regression between T_{ps} and LWSD, due to damage at the site.

RESULTS

Air Temperature Regime

Annual mean air temperatures (AMAT) from LSF, USF and DST are summarised in Table 1. The AMAT at LSF for both years were lower (−7.6 and −7.8 °C) than the historical average at Fort McPherson (−7.0 °C), but the instrument was in undisturbed forest, while the Environment Canada site, which was in operation until 2007, was at the Fort McPherson airport.

Strong air temperature inversions were observed between LSF and DST during the winter, and particularly during the coldest months (Figure 3a, b). Here, an inversion is considered to have occurred when the daily mean air temperature, calculated from measurements at 2 h intervals, was higher in DST than in LSF. During the freezing season (Oct–Apr), when daily mean air temperatures in LSF were below 0 ° C, inversions occurred on 55 per cent of the days. During the thawing season (May–Sep), normal surface lapse rates were persistent (Figure 3b), and inversions between LSF and DST occurred during only 8 per cent of days.

The frequent and persistent temperature inversions caused the number of freezing degree days to decline with elevation from LSF to DST (Figure 3b). The strength of the winter inversions, when positive surface lapse rates dominated, was enough cumulatively to outweigh the weaker normal surface lapse rates in summer, so that the AMAT were lowest at LSF in both years (Table 1). For example, the mean winter air temperature (Dec–Feb) in 2011–12 was −21.1 °C at DST and −25.1 °C at LSF, while the mean summer temperatures in 2012 (Jun–Aug) were 13.4 and 14.5 °C, respectively.

Snow Conditions

The annual start date of a continuous winter snow cover was estimated by examining ground temperatures from the thermistors at 5 cm depth. Snow was assumed to cover the ground when the diurnal variation in surface temperature dropped below 1 °C (Burn and Zhang, 2009). At low elevation on Peel Plain, these conditions were recorded between 24 September and 17 October in 2010–13 (Table 2). On Peel Plateau, the arrival of snow occurred between 21 September and 17 October.

In each year, median LWSD was greater in the forest than on the tundra (Figure 4a). The rank order in the forest has been USF > LSF, but it has not been consistent between TST and DST over the period of study. Variation in LWSD

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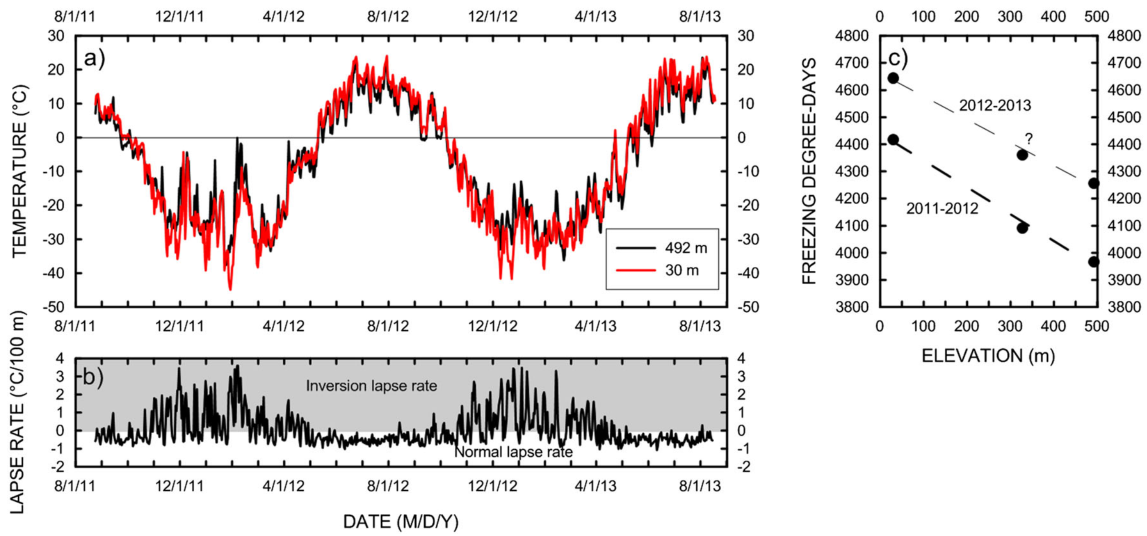

Figure 3 Plots showing (a) air temperatures on Peel Plain (LSF – 30 m) and near the top of Peel Plateau (DST – 492 m); (b) daily surface lapse rates between 30 m and 492 m; and (c) relations between freezing degree days and elevation. The question mark indicates where missing data were estimated for USF using regression (see text). See text for abbreviations. This figure is available in colour online at wileyonlinelibrary.com/journal/ppp

increased with elevation across the ecotone. On Peel Plain in LSF, snow depths were the least variable (47–95 cm). At USF, variation in snow depth was greater (50–132 cm) than at LSF. Snow depth along the transects varied less in TST (32–80 cm) than in USF, but had the greatest variability overall due to deep snow accumulation at the TST-2 thermistor (Figure 4a). Snow depth variation was relatively high along the DST transect (23–111 cm). The Wilcoxon rank-sum tests indicated that snow depths measured along the transects at each site were significantly different from those at all other sites in individual years, except between LSF and DST in 2013 (Table 3a). When the data for all years were pooled, there were significant differences between all vegetation units except DST and TST, and DST and LSF (Table 3).

Snow density increased, and snow thermal resistance decreased with elevation from forest to tundra (Figure 4b). Density was lowest at LSF (172–194 kg/m^3) and increased with elevation to DST (233–311 kg/m^3). The thermal resistance of snow was higher at the forest sites (7.9–11.4 m^2 K W^{-1}) than in tundra (1.4–8.3 m^2 K W^{-1}) (Figure 4b), due to lower densities and deeper cover in the forest. However, at TST-2, the thermal resistance was comparable to forest sites due to snowdrift accumulation at the slope base.

iButton thermistor arrays were employed to characterise the timing of deep snow accumulation in forest and tundra. Snow accumulation inferred from the iButton array in forest at USF suggests that snow accumulates gradually, coincident with precipitation events (Figure 5a). The inferred iButton snow depths closely agree with those measured at the MET station located in forest approximately 6 km to the southwest (Figure 1). The similarity of the plots suggests that snow accumulation is relatively uniform in forest on Peel Plateau and that wind distribution is minimal. In contrast, snow accumulation occurred rapidly at the iButton array located beside the snow fence in DST, presumably due to redistribution by wind (Figure 5b).

Permafrost Conditions Across Treeline

Near-surface ground temperatures in 2010–13 (Table 2) generally increased with elevation across the forest-tundra ecotone (Figure 6). The T_{ps} in LSF on Peel Plain ranged between −2.5 and −2.8 °C in 2010–13 (Table 2). These data are corroborated by a mean temperature of −2.5 °C near 5 m depth from the LSF deep cable in 2012–14 (Figure 7). In forest on Peel Plateau (USF-1), T_{ps} was between −1.7 and −2.7 °C, and was −1.0 °C near 5 m depth at the deep cable in 2012–14. At TST, T_{ps} at sites with mineral soil ranged from −1.4 to −2.2 °C in 2010–13. At DST, T_{ps} was between −1.2 and −2.4 °C at sites with mineral soils in the same period. The mean temperature near 5 m at the deep cable in DST was −2.2 °C in 2012–14. In peatland, T_{ps} was −0.5 °C at TST-2, and between −2.3 and −4.4 °C at DST-2 (Figure 6). The majority of T_{ps} values at TST and DST (12 of 14) were higher than those observed in forest on Peel Plain (Figure 6). The data from deep thermistor cables also indicate that permafrost on Peel Plateau is relatively warm, with higher temperatures at both USF and DST than at LSF (Figure 7).

Ground Temperatures and Snow Cover

Strong associations were registered between LWSD and both T_{ps} and annual mean surface temperature (T_s) at tundra sites (Figure 8). Linear regression indicated that LWSD accounted for the majority of variation in T_{ps} (T_{ps} = 0.04 (LWSD) - 4.79, r^2 = 0.72, n = 19, p < 0.01). The relation for forest sites was not statistically significant (T_{ps} = 0.03

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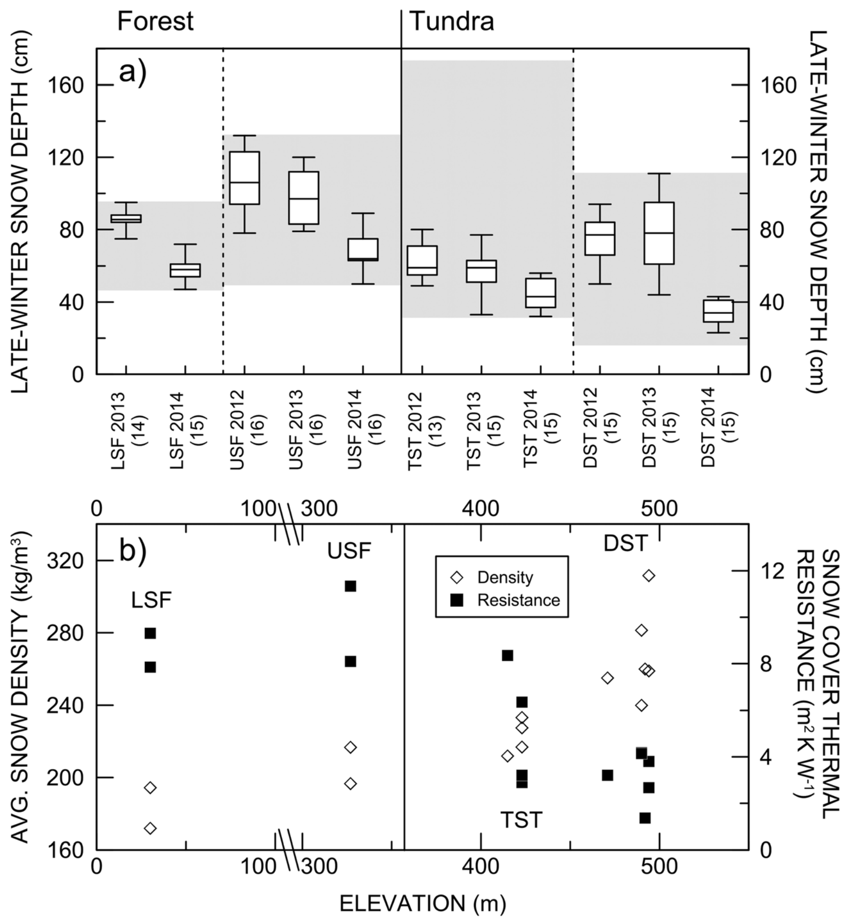

Figure 4 (a) Box plot of snow depths from transects for winters 2012–14. The number of points measured along each transect is in parentheses. The shaded grey areas represent the range of individual snow depth measurements from transects and shallow thermistor sites (Table 2). Snow depths were not measured at LSF in 2012 due to logistical constraints; (b) average snow density and snow cover thermal resistance determined from 2013 and 2014 snow pits. In LSF and USF, the snowpack mainly comprised powder and hoar fractions. In tundra, the depth hoar fraction was smaller, and dense (300–400 kg/m^3) wind slab layers were present at some sites, contributing to the higher densities. See text for abbreviations.

Table 3a Wilcoxon rank-sum results for 2012–14 snow depths from transects.

Site	LSF	USF	TST	DST
LSF	—	20	18	9
USF	X	—	38	29
TST	X	X	—	9
DST	4	X	X	—

(LWSD) - 4.56, $r^2 = 0.43$, $n = 8$, $p = 0.14$), and all the points lay below the tundra regression line (Figure 8).

The thermal effect of snow was also apparent when comparing TST-2 and DST-2. Both sites were in peatland, and had similar late-summer thaw depths (45 cm in 2011–12). Only 77 m of elevation separated the sites, so their air temperatures were also similar. However, TST-2 was at the base of a north-facing slope, where deep snow accumulated annually, whereas DST-2 was exposed and most snow blew away. In 2012 and 2013, LWSDs at TST-2 were 119 and 97 cm, respectively, while there was 37 and 20 cm of snow at DST-2, respectively. At TST-2, the surface temperature was decoupled from the air temperature in autumn, indicating the early accumulation of deep snow (Figure 9). As a result, freezeback was delayed significantly at TST-2. Between 2010 and 2013, T_{ps} at DST-2 was between −2.3 and −4.4 °C, and freezeback in 2010–11 occurred nearly 2 months earlier than at TST-2 (Figure 9). The 2010–11 data suggest that, if all other environmental conditions are similar, snow may account for at least 1.8 °C of the variation in T_{ps} on Peel Plateau (Table 2). This magnitude may only apply at peatland sites, as variation of T_{ps} resulting from snow cover in mineral soil likely differs due to the different thermal properties of the substrates.

Active Layer Thicknesses

Active layer thicknesses and variability generally increased with elevation from forest to tundra (Figure 10), except in peatland. Median active layer thicknesses from 2011 to

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Table 3b Wilcoxon rank-sum results for 2011–13 active layer thicknesses from transects.

Site	LSF	USF	TST	DST	Peatland
LSF	—	10.5	29	27.5	5.5
USF	*1,2*	—	18.5	17	5
TST	*X*	*2,3*	—	1.5	23.5
DST	*X*	None	None	—	22
Peatland	*X*	None	*X*	*X*	—

Note: *X* indicates significant *p*-values between the sites in all years, while 1, 2, 3 and 4 refer to significant *p*-values in 2011, 2012, 2013 and 2014, respectively. The upper right quadrant is a summary of the data pooled for 2011–13. The numbers indicate the difference in the 3 year median values between the sites (cm). Underlined values indicate significant results in the pooled years. No snow depths were measured at LSF in 2012. See text for abbreviations.

2013 were between 36 and 40 cm in LSF on Peel Plain, and in individual years, the Wilcoxon rank-sum tests indicated that the thicknesses were statistically different than those at all other sites, except USF in 2013 (Table 3b). In USF, the 2011–13 median thicknesses were between 48 and 55 cm. TST had the greatest median active layer thicknesses, between 63 and 87 cm, and in individual years, the thicknesses differed statistically from all sites except DST (Table 3b). In DST, median thicknesses were between 63 and 72 cm, and the active layer thicknesses were statistically different from those in LSF and the peatlands (Table 4). The peatland active layer thicknesses in TST and DST were relatively uniform and varied little from 2011 to 2013, with median values from 43 to 46 cm. Peatland active layer thicknesses were significantly different from those measured along the transects at all sites except USF, which had similar median values (Table 3b).

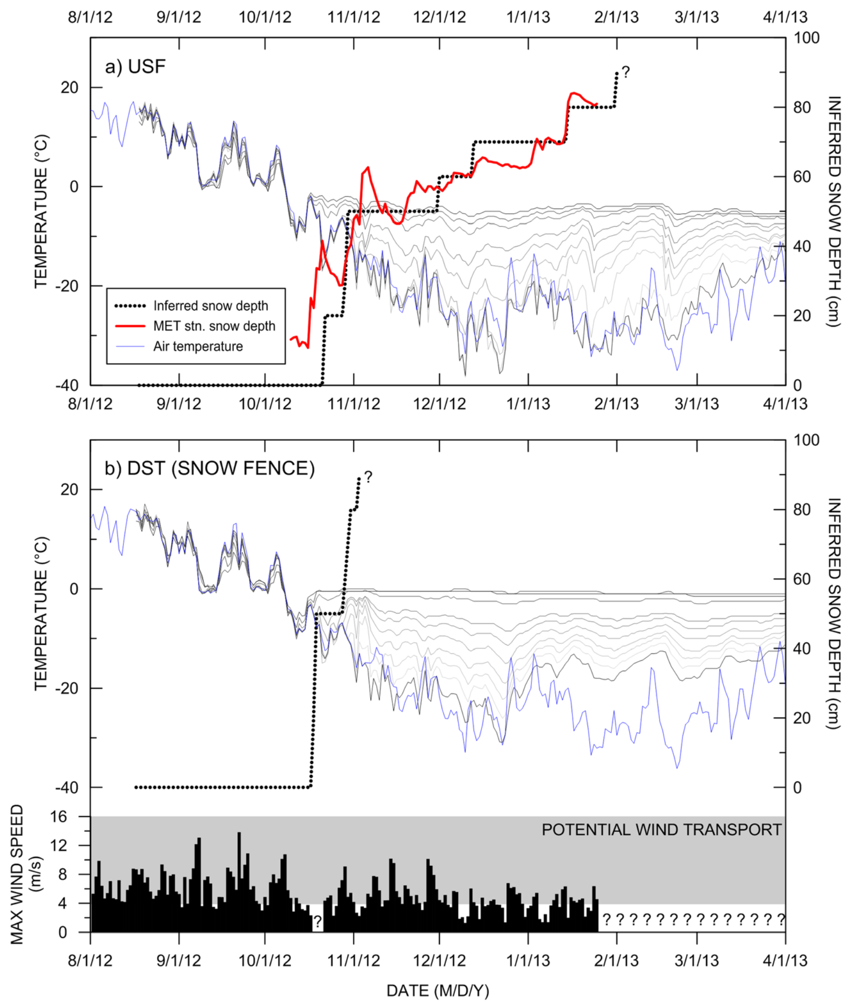

Figure 5 (a) iButton temperatures (grey lines) and inferred snow depth in forest on Peel Plateau; (b) iButton temperatures and inferred snow depth adjacent to a snow fence in tundra. The blue lines in (a) and (b) are air temperatures from the USF and DST air temperature sensors, respectively. Daily maximum wind speeds from the meteorological (MET) station are plotted as black bars, and the grey shading represents wind speeds at which snow transport is possible under cold, dry conditions (Li and Pomeroy, 1997). Question marks indicate the limit of inferred snow depths (when snow is ≥ 90 cm), and missing wind data from the MET station. See text for abbreviations. This figure is available in colour online at wileyonlinelibrary.com/journal/ppp

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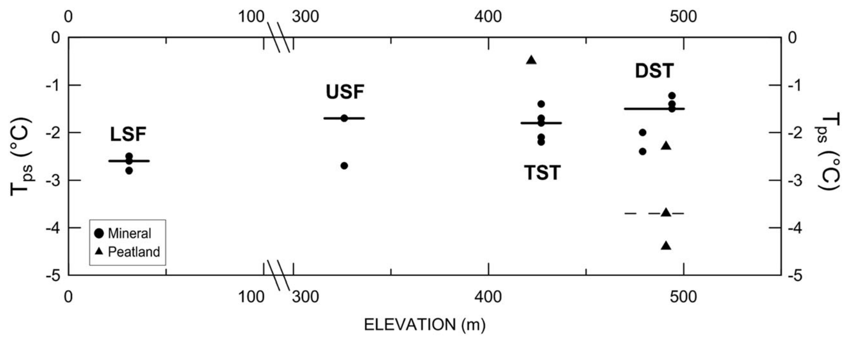

Figure 6 Annual mean temperature near the permafrost surface (T_{ps}) across the forest-tundra ecotone for 2010–13. The solid black lines represent the median T_{ps} from instruments in mineral soil for each vegetation unit, and the dashed line is the median for the DST peatland instrument. See text for abbreviations.

DISCUSSION

Air Temperatures

The air temperature regimes at LSF and DST revealed strong and persistent temperature inversions in winter. The inversion frequencies are similar to results presented by Burns (1973, p. 72), who reported that, between December and February, inversions occurred in Inuvik for 67 per cent of the time at 11 GMT (3 am) and 58 per cent at 23 GMT (3 pm) – the two moments of radiosonde releases. In the Peel Plateau region, daily winter temperature inversions may occur more frequently than the 55 per cent reported, because inversions below 492 m, the elevation of the DST air temperature sensor, may not have been captured. The magnitudes of surface lapse rates observed during inversions in this study were similar in December–February (mean 1.1 °C/100 m) to those reported by Lewkowicz *et al*. (2012, p. 200) in the Yukon, where temperatures increased at about 1.0 °C/100 m to the elevation of treeline in the coldest months. On individual days, increases of up to about 3.5 °C/100 m were recorded between Peel Plain and Peel Plateau.

Figure 11 shows the strong relation (freezing degree days = 223(Latitude) - 10 752; $r^2 = 0.99$) between latitude and average freezing degree days (1981–2010) for seven communities in the NWT and Nunavut. The number of freezing degree days in DST was 450 less than in LSF in 2011–12 and 388 less in 2012–13. The regression equation indicates that these effects of temperature inversions on freezing degree days on Peel Plateau are equivalent to that observed with a southward shift in latitude of nearly 2°. Winter air temperatures on Peel Plateau are similar in aggregate to those observed in lowlands nearly 200 km to the south, between Fort Good Hope and Norman Wells, near the boundary of the continuous and discontinuous permafrost zones.

In summer, air temperature decreased with elevation. The mean July temperature in LSF was 17.7 and 15.4 °C in 2012 and 2013, respectively, while in DST it was 15.0 and 12.7 °C, respectively. The 2012 mean July temperature in LSF (17.7 °C) was higher than the 1981–2010 average mean July temperature (15.2 °C) recorded at the Fort McPherson airport (Environment Canada, 2014), indicating that July 2012 was considerably warmer than normal. The 2013 temperature in LSF (15.4 °C) was similar to the long-term average, and corresponded with a 12.7 °C temperature at DST, close to the 12 °C limit of treeline reported by Walker (2000) for continental areas. Therefore, we interpret that the position of treeline on Peel Plateau is broadly controlled by the prevailing summer air temperature conditions.

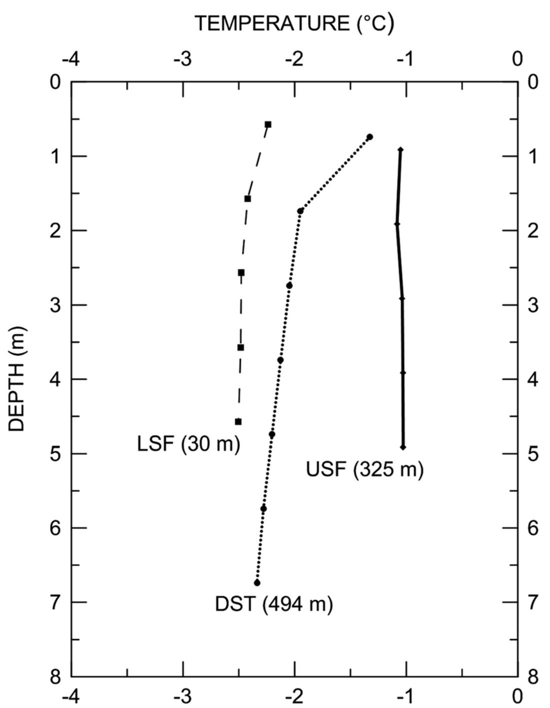

Figure 7 Mean ground temperatures for September 2012 to August 2014. The symbols on each line represent the thermistor locations. See text for abbreviations.

Active Layer Thicknesses

Active layer thicknesses were predominantly associated with the thickness and variation of the surface organic

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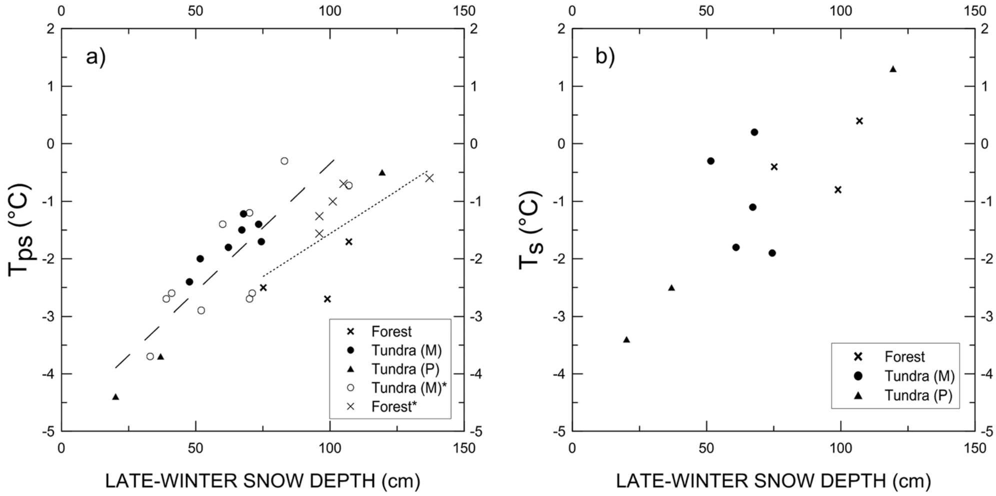

Figure 8 Relations between late-winter snow depth (LWSD) and (a) annual mean permafrost surface temperature (T_{ps}) and (b) ground surface temperature (T_s) for 2012 and 2013. Tundra (M) and Tundra (P) denote sites in mineral and peatland soils, respectively. Asterisks in the legend denote additional data collected on Peel Plateau by Gill *et al.* (2014). The dashed line is the regression relation for tundra sites ($T_{ps} = 0.04(\mathrm{LWSD}) - 4.79$, $r^2 = 0.72$, n = 19, $p < 0.01$), and the dotted line is for forest sites ($T_{ps} = 0.03(\mathrm{LWSD}) - 4.56$, $r^2 = 0.43$, n = 8, $p = 0.14$). There are fewer points for T_s than T_{ps} because subsidence at DST-3 caused the 5 cm thermistor to be exposed after 2011–12, and because TST-3 did not have a surface thermistor. See text for abbreviations.

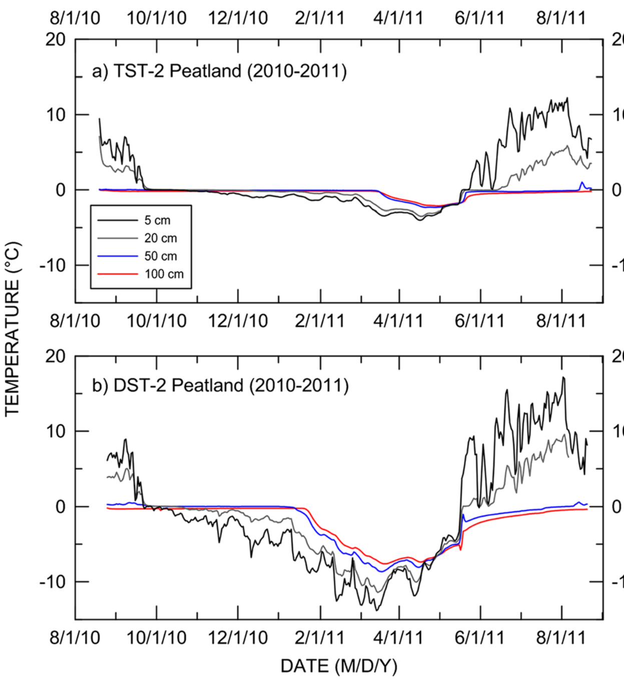

Figure 9 Ground temperatures at (a) peatland at the base of a slope where thick snow accumulates ($T_s = 1.3$ °C; $T_{ps} = -0.5$ °C), and (b) an exposed peatland where little snow accumulates ($T_s = -0.8$ °C; $T_{ps} = -2.3$°C). See text for abbreviations. This figure is available in colour online at wileyonlinelibrary.com/journal/ppp

layer at the four sites, while shading from the forest canopy and substrate composition were also important at some sites. Active layer thickness is generally considered to be predominantly a function of summer temperature, surface conditions and substrate composition (Smith, 1975), though conditions in winter are important in some settings (e.g. Burn and Zhang, 2010). Thawing degree days declined with elevation on Peel Plateau due to the persistent normal surface lapse rates in summer (Table 1), but active layer thicknesses generally increased from the forest up to the tundra (Figure 10).

Active layer thicknesses were relatively shallow and uniform at LSF (Figure 10), likely because a thick, continuous surface organic layer was present, and shading from the canopy decreased incident radiation on the surface (Smith, 1975). The lower permafrost temperatures at LSF may have also contributed to shallower active layer depths, due to the heat required to thaw the colder ground. The median active layer thicknesses at USF were greater than at LSF (Figure 10), and depths were more variable, likely due to the sparser forest canopy and variable surface organic layer thickness at the higher elevation site. TST had the greatest and most variable active layer thicknesses (Figure 10). The deepest measurements were from sandy areas, where it was difficult to ascertain the position of the frost table due to friction along the probe, while the shallow measurements were in troughs of poorly defined hummocks where there was a thicker surface organic layer. DST had less variable active layer thicknesses than TST (Figure 10), due to the absence of sandy substrate, but the differences were not statistically significant (Table 3b). The variation at DST was related to the differences in organic cover on hummock tops and troughs. Peatlands had uniform and thin (~45 cm) active layers, similar to LSF, due to the thick and continuous organic cover. However, the active layer thicknesses in peatland and LSF were statistically different, because LSF had consistently shallower depths, likely due to shading in the forest.

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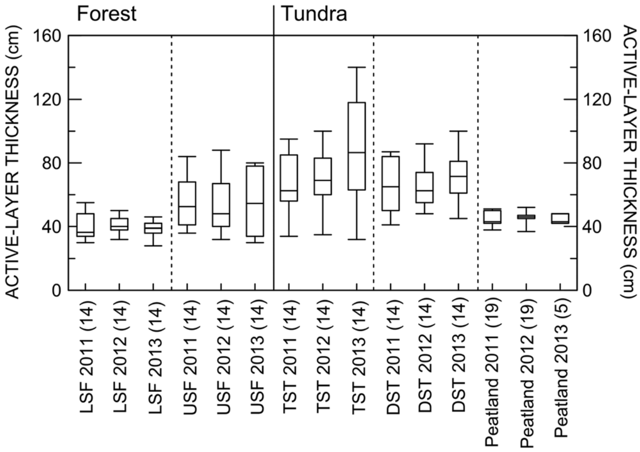

Figure 10 Active layer thicknesses across the forest-tundra ecotone from the four transects in mineral soil and around instrumented sites in peatland (pooled values from the TST and DST peatland). The number of individual points measured is in parentheses. See text for abbreviations.

In summary, the effects of surface and subsurface conditions obscured any relation between thawing degree days and active layer thicknesses across the ecotone.

Snow Conditions

Snow depths generally declined with elevation from forest to tundra and were more variable in tundra (Figure 4a), but the pattern is not as clearly defined as at latitudinal treeline (Palmer *et al*., 2012, Figure 5). Snow depths from the transects in DST exceeded those in TST in 2012 and 2013 (Figure 4a), in contrast with the typically observed relation of increasing snow depth with vegetation height (Thompson *et al*., 2004; Palmer *et al*., 2012). Linear regression showed no apparent relation between mean maximum vegetation height and LWSD at instrumented sites in tundra, suggesting that vegetation is a secondary control of snow accumulation at the site scale. This observation suggests that mesoscale topographic/wind effects may be important in the rolling terrain of Peel Plateau.

Snow cover was thin on Peel Plateau in winter 2013–14 (Figure 4a), due to low snowfall in early winter and strong winds that removed snow from the tundra (S. Tetlichi and C. Firth, personal communication, 2014). Mesoscale topographic effects may explain the deep snow observed in forest at USF (Figure 4a), which is located in a broad valley. Forested valleys in the Peel Plateau region may be sinks for snowfall blown from nearby tundra (e.g. Roy-Léveillée *et al*., 2014), but further investigation is required on this topic, since our study did not control for these effects.

Snow depths in LSF were similar to those observed in forest near Inuvik, but average tundra snow depths were typically deeper on Peel Plateau than has been observed recently north of treeline in the Mackenzie Delta area (~50–60 cm vs 20–50 cm) (Morse *et al*., 2012; Palmer *et al*., 2012). Snow is likely deeper in tundra on Peel Plateau than tundra further north because more precipitation falls in the former than has been measured north of latitudinal treeline (Burn and Kokelj, 2009, Figure 3).

Snow cover density increased with elevation, and tundra snow cover densities were much greater than those in forest. Aggregate snow densities from tundra snow pits were higher than those observed north of latitudinal treeline to the northeast. On Peel Plateau, the snow densities from pits in TST and DST were between 233 and 311 kg/m^3, whereas upland shrub tundra sites in the Mackenzie Delta area have reported average densities of 120–230 kg m^{-3} (Palmer *et al*., 2012) and 238 kg m^{-3} (Morse *et al*., 2012). In LSF and USF, snow cover densities were 172–194 and 197–217 kg/m^3, respectively, similar to the value of 190 kg/m^3 reported by Palmer *et al*. (2012) from forest near Inuvik.

The thermal resistance of the snow covers observed in this study were within the range of those observed at latitudinal treeline, but the average resistance from snow pits in DST (3.2 m^2 K W^{-1}, n = 6) was greater than that reported by Palmer *et al*. (2012) (2.4 m^2 K W^{-1}) for a site north of latitudinal treeline in DST, even though five of the six snow pits were dug in 2014, when snow cover was thinnest on Peel Plateau. The high thermal resistance of the snow cover in tundra on Peel Plateau may contribute to the warm permafrost conditions.

Permafrost Conditions

Ground temperatures did not decline substantially across treeline, in contrast to observations by Lewkowicz *et al*. (2012) and Palmer *et al*. (2012). The majority of tundra sites had higher T_{ps} than those at LSF (Figure 6). Ground temperatures at 5 m depth were also greater on Peel Plateau than those on the lower-lying Peel Plain (Figure 7). Tundra permafrost temperatures were several degrees higher than those typically observed north of latitudinal treeline (e.g. Burn and Kokelj, 2009; Palmer *et al*., 2012), and higher than those reported for the study region on national-scale maps (Heginbottom *et al*., 1995; Henry and Smith, 2001). The combined effects of strong winter temperature inversions and relatively deep snow contribute to the warm permafrost conditions observed at tundra sites. We emphasise that our results apply to the elevation range examined, but that in higher terrain, which may have thinner snow cover and experience the effects of inversions less frequently, ground temperatures may be lower than at LSF.

Our data show that the number of freezing degree days on Peel Plateau may be similar to that of lower latitudes such as the area between Fort Good Hope and Norman Wells, NWT. Burgess and Smith (2000, Figure 3a) reported mean annual ground temperatures at sites near Fort Good Hope between −2 and 0 °C. The average mean annual ground temperature from 17 sites near Norman Wells, reported in the Ground Temperature Database for Northern Canada (Smith and Burgess, 2000) is about −2.7 °C, lower than the median temperature on Peel Plateau, but with a similar range (−0.8 to about −4.5 °C). Therefore, the similar winter air temperature conditions on Peel Plateau and between Fort

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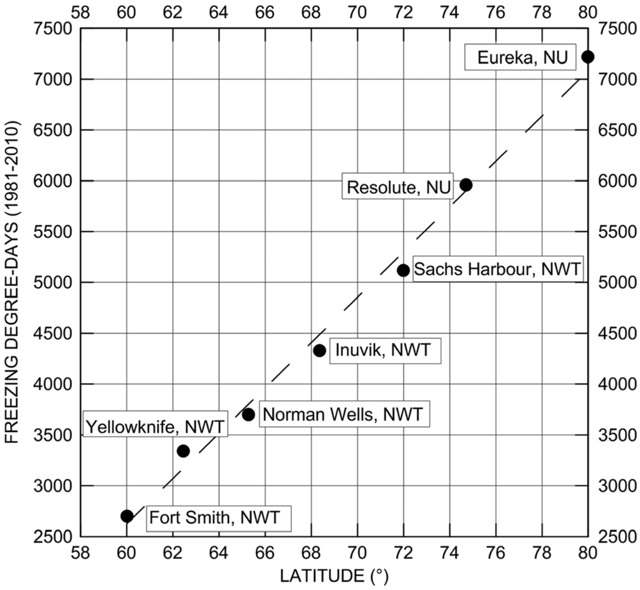

Figure 11 Relation between latitude and average freezing degree days (1981–2010) in seven communities (Environment Canada, 2014). The study area on Peel Plateau is at about 67.2°N, and there were 3968 and 4257 freezing degree days at DST in 2011–12 and 2012–13, respectively. These data are not plotted here, as they are not similar to the long-term averages. See text for abbreviations. NU = Nunavut.

Good Hope and Norman Wells likely lead to similar permafrost temperatures.

Typically, higher permafrost temperatures in forest than tundra are attributed to deep snow covers (e.g. Smith *et al*., 1998; Burn and Kokelj, 2009; Palmer *et al*., 2012). However, on Peel Plateau, where snow in tundra may also be deep, the timing of snow accumulation may be an important factor contributing to lower ground temperatures at forest sites. Data from the iButtons in DST suggested that > 90 cm of snow may accumulate by early November, whereas the snow cover in the forest accumulated more gradually over several months, only surpassing 90 cm depth in early February (Figure 5). The more gradual snow accumulation in forest may partly explain why the relation between T_{ps} and LWSD for forest sites is distinctive from that for tundra sites (Figure 8). This is supported by a shorter duration of freezeback at forest sites. For 2010–13, LSF had a mean freezeback duration of 116 days, while the mean at USF was 118 days; at tundra sites, freezeback averaged 140 days. The shorter freezeback period at forest sites is likely due, in part, to thinner snow cover early in the freezing season compared to tundra sites where wind-blown snow accumulates, and because active layers are thinner in the forest.

In summary, the warm permafrost temperatures observed on Peel Plateau are likely an integrated response to the higher air temperatures in winter caused by atmospheric temperature inversions, relatively deep snow in the tundra and rapid snow accumulation in some areas due to wind redistribution. These conditions may apply over large spatial extents in the continuous and discontinuous permafrost zones, in areas with alpine treelines and air temperature inversions.

The results of this study are also relevant to infrastructure planning, as it is apparent that the presence of tundra in the continuous permafrost zone does not necessarily imply cold (< −5.0 °C) permafrost conditions, in contrast with common understanding (e.g. Johnston, 1981; MacGregor *et al*., 2010). The vegetation gradient at treeline is controlled by summer air temperatures (Walker, 2000), but here we emphasise the role of winter conditions, which do not govern the broad-scale vegetation assemblages across the ecotone, but largely define the ground thermal regime and T_{ps}.

The high T_{ps} values measured on Peel Plateau and their association with snow depth suggest that the ice-rich permafrost is particularly susceptible to degradation following perturbations that increase snow cover. Precipitation in northern regions is likely to increase in the future (Collins *et al*., 2013), which may cause permafrost to warm and degrade if snow cover increases.

The relation between T_{ps} and snow depth may also have important consequences for the ground thermal regime along the Dempster Highway, where the embankment and roadside shrubs that have proliferated since its construction enhance snow depth (Gill *et al*., 2014). In 2013, the Government of the NWT installed two permanent permafrost monitoring sites beside the highway in the study region. Continued research should focus on the thermal regime of permafrost where snow accumulates adjacent to the highway embankment. The compaction or removal of snow should be explored as methods to facilitate freezeback of the active layer given the strong relation between LWSD and T_{ps} observed in this study.

CONCLUSIONS

The results from our study area contrast with previous observations of permafrost temperatures declining markedly across the forest-tundra ecotone. Permafrost in the tundra environment was remarkably warm, contrary to common associations of the vegetation type with cold (< −5 °C) ground temperatures in the continuous permafrost zone. The results of this investigation into air and ground temperatures and snow conditions across this elevational treeline allow the following conclusions to be drawn.

1. Strong and persistent air temperature inversions in winter caused the number of freezing degree days to vary inversely with elevation between Peel Plain and Peel Plateau, so that tundra sites were warmer than forest sites. The difference in freezing degree days due to elevation is equivalent to a southwards shift of nearly 200 km.
2. Snow depths generally declined and snow cover density increased from forest to tundra, suggesting that vegetation controls regional-scale snow conditions across the ecotone. Variation in snow depth was greater at tundra

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sites than in forest. Deep tundra snow is primarily related to topographic setting at the site scale, whereas vegetation height is a secondary factor. Tundra snow depths are greater on Peel Plateau than north of latitudinal treeline, likely due to greater precipitation in the study area.
3. Snow may accumulate rapidly in tundra due to wind redistribution, and gradually in forest, coincident with snowfall events. This contributes to shorter freezeback duration and lower permafrost temperatures at forest sites in comparison with tundra sites with similar late-winter snow accumulation.
4. Permafrost temperatures in the tundra of Peel Plateau are remarkably high as a result of the combined effects of winter air temperature inversions, relatively deep tundra snow depths and rapid accumulation of snow in some tundra settings.

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