



## Spatial and temporal variations in active layer thawing and their implication on runoff generation in peat-covered permafrost terrain

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[1] The distribution of frost table depths on a peat-covered permafrost slope was examined in a discontinuous permafrost region in northern Canada over 4 consecutive years at a variety of spatial scales, to elucidate the role of active layer development on runoff generation. Frost table depths were highly variable over relatively short distances (0.25–1 m), and the spatial variability was strongly correlated to soil moisture distribution, which was partly influenced by lateral flow converging to frost table depressions. On an interannual basis, thaw rates were temporally correlated to air temperature and the amount of precipitation input. Simple simulations show that lateral subsurface flow is governed by the frost table topography having spatially variable storage that has to be filled before water can spill over to generate flow downslope, in a similar manner that bedrock topography controls subsurface flow. However, unlike the bedrock surface, the frost table is variable with time and strongly influenced by the heat transfer involving water. Therefore, it is important to understand the feedback between thawing and subsurface water flow and to properly represent the feedback in hydrological models of permafrost regions.

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### 1. Introduction

[2] Permafrost (lithospheric material that has a temperature at or below 0°C for a minimum of 2 consecutive years) and seasonally frozen soils play a significant role in slope runoff generation and the basin hydrology of cold regions [e.g., Dingman, 1973; Woo and Winter, 1993; Carey and Woo, 2001; Quinton and Hayashi, 2005]. Permafrost in high-latitude regions is often saturated or oversaturated with ice, and, therefore, acts as a confining layer, limiting the movement and storage of groundwater to a seasonally thawing active layer, below which the frozen ice-saturated soil is relatively impermeable [Dingman, 1975]. Seasonal ice in the active layer decreases the hydraulic conductivity and available storage capacity of the soil, which significantly reduces water infiltration [Kane and Chacho, 1990], potentially increasing surface ponding and the magnitude of slope runoff during snowmelt and spring rainfall events. However, the occurrence of overland flow is rare on permafrost slopes covered by highly permeable, moist peat. On these slopes, the drainage of precipitation and meltwater inputs occurs primarily as subsurface flow through the thawed water-saturated layer perched above the frost table

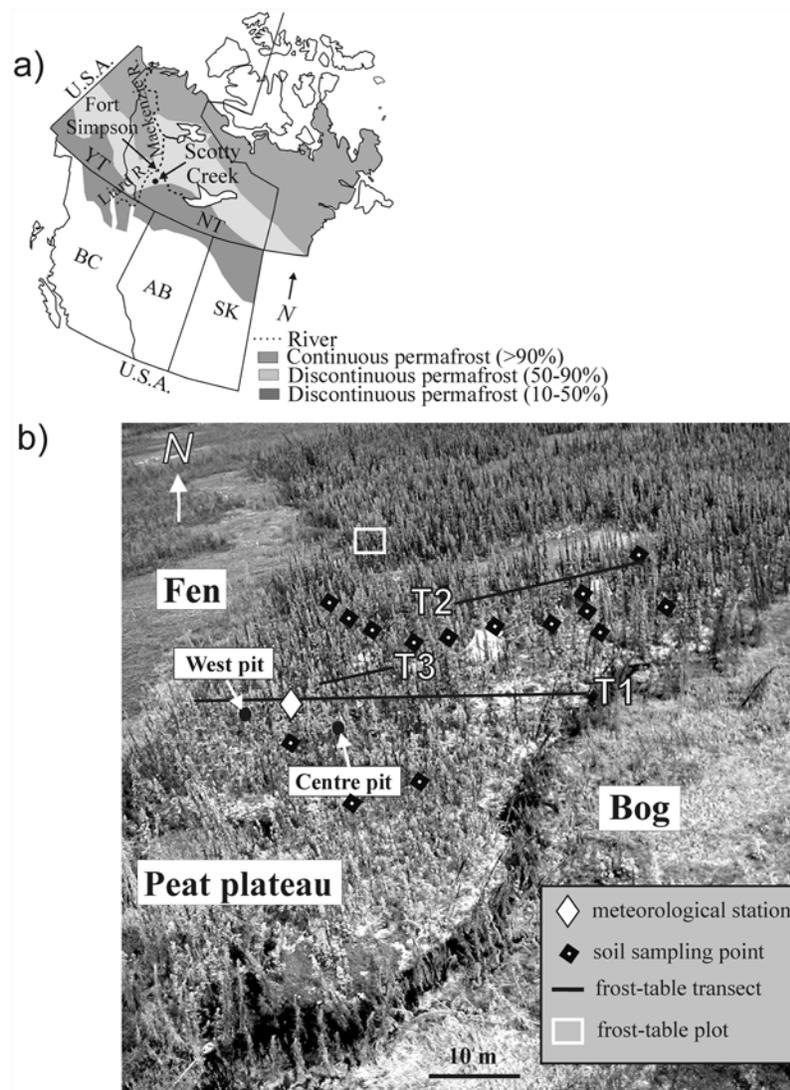
[Slaughter and Kane, 1979; Quinton and Marsh, 1999]. In the context of subsurface flow, the frost table is the relatively impermeable upper surface of the frozen, ice-saturated soil that coincides closely with the 0° isotherm during soil thawing [Carey and Woo, 1998]. The depth and distribution of the frost table indirectly controls the position of the zone of saturated water flow, which changes as the soil thaws [Woo, 1986]. During soil thawing, the frost table and the water-saturated zone above it descend through the soil profile. Because of the large reduction of the saturated hydraulic conductivity of organic soils with depth [Quinton et al., 2000], lateral flow rates decrease as frost table depths increase [Glenn and Woo, 1987; Wright et al., 2008].

[3] Uneven soil thawing on frozen slopes also affects the mode and rate of water flow downslope [e.g., Woo and Steer, 1983; Stadler et al., 1996; Metcalfe and Buttle, 1999; Quinton et al., 2004]. Therefore, knowledge of the spatial and temporal variation of thaw depths (i.e., frost table depths) is essential to understanding how water flows downslope. These often complex variations in frost table topography [Nelson et al., 1998, 1999] must be considered before point measurements are scaled up for use in basin studies [Young et al., 1997]. Furthermore, understanding the factors that control the distribution of frost table depths is critical to hillslope runoff modeling in permafrost basins [Woo, 1986; Metcalfe and Buttle, 2001], particularly in wetland-dominated, discontinuous permafrost basins, where permafrost often exists only as scattered patches (ranging from 0.5 to 2 m in height above the surrounding terrain and commonly extending to several square kilometers in area) that appear as islands among otherwise unfrozen, peat-

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**Figure 1.** (a) The location of Scotty Creek within northwestern Canada. (b) Photo of the peat plateau study site, positioned between two seasonally frozen wetlands (a fen and bog), and the location of the instrumentation and frost table measurement transects (T1, T2, and T3).

covered, saturated terrain [National Wetlands Working Group (NWWG), 1988]. These slightly inclined permafrost slopes, called peat plateaus, are hydrologically important to basin drainage, as they generate a large amount of runoff for streamflow [Wright *et al.*, 2008].

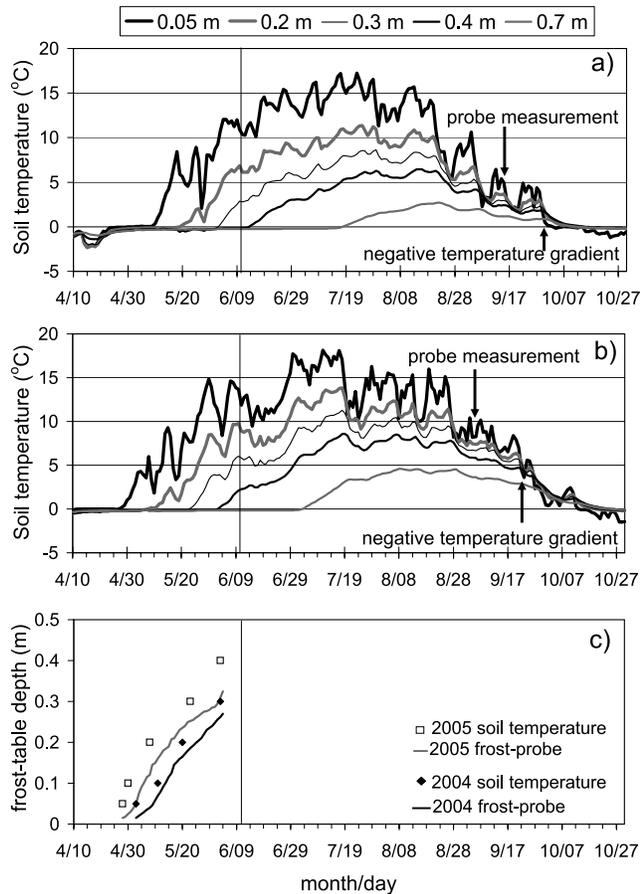
[4] There have been numerous field measurement and modeling studies describing the spatial and temporal variation in frost table depths for different landcover types, and attempts to ascertain the specific factors controlling the distribution [e.g., Smith, 1975; Nelson *et al.*, 1997, 1999; Hinkel and Nelson, 2003]. However, most of these studies are conducted at relatively large spatial scales (often  $\geq 100$ -m measurement intervals) and are limited to end-of-summer or annual frost table depths only. Hinkel and Nelson [2003] noted the need for specific small-scale studies to understand the causes of the significant intrasite variation in thaw depth and moisture conditions found at the larger scale. There is a lack of knowledge on the small-scale spatial and temporal variability of frost table depths on permafrost slopes, including the factors controlling it, and more importantly,

how it influences runoff generation. The overall goal of the present study is to close this knowledge gap and provide new scientific insights into the complex interplay between subsurface flow and active layer thawing on permafrost hillslopes. More specifically, this study will (1) document the spatial variability of frost table depth and its temporal evolution across a peat plateau during the thawing seasons of 2003–2006; (2) examine how the variability of frost table depth is affected by air temperature, rain and snowmelt inputs, soil moisture, vegetation, snow cover, and surface topography; and (3) demonstrate how the frost table variability may result in preferential flow pathways for subsurface runoff.

## 2. Study Area and Methods

### 2.1. Site Description

[5] Field studies were conducted on a peat plateau located in the 152-km<sup>2</sup> Scotty Creek basin (61°18'N, 121°18'W; 285 m above sea level), which lies in the lower Liard River valley, Northwest Territories, Canada (Figure 1a). The study



**Figure 2.** Average daily soil temperature of the Center pit from 10 April to 1 November (a) 2004 and (b) 2005. (c) The arrival of the thawing front at discrete depths, according to soil temperature data at the Center pit (symbols), and the daily frost table depth measured by the frost probe near the Center pit (continuous line) during the spring field seasons of 2004 and 2005.

area is located in the zone of discontinuous permafrost [Heginbottom and Radburn, 1992], and is in the continental high boreal wetland region of Canada, slightly south of the transition to the low subarctic wetland region [NWWG, 1988]. Climate data are available from the nearest Environment Canada weather station at Fort Simpson airport (169 m above sea level), 50 km north of the study site. There, the mean (1964–2006) annual air temperature is  $-3.2^{\circ}\text{C}$ , with a mean total annual precipitation of 363 mm, including the moisture equivalent of 163 mm of snow [Meteorological Service of Canada, 2006]. Snowmelt usually commences in late March, and because of occasional periods of sub- $0^{\circ}\text{C}$  air temperatures and additional snowfall events, generally lasts from two to six weeks. The stratigraphy of the region includes a peat layer of varying thickness (approximately 0.5 m to 8 m) that overlays a mineral silt-sand layer (up to 1 m thick), below which lies a thick (average thickness 6 m) clay to silt-clay deposit of low permeability [Rutter et al., 1973; Aylsworth et al., 1993]. Permafrost thickness under peat-covered terrain in the Fort Simpson area has been reported to be 5–10 m [Burgess and Smith, 2000].

[6] The ground surface of the peat plateau examined in this paper (Figure 1b) rises 0.9 m above the surrounding

permafrost-free terrain, and has a considerable microrelief (Figure 2) (for measurement see section 2.3). The plateau supports a sparse stand ( $1 \text{ stem m}^{-2}$ ) of relatively short (mean height 3.1 m) black spruce trees (*Picea mariana*), and a variety of shrubs, of which Labrador tea (*Ledum groenlandicum*) is the dominant. The ground cover is dominated by lichen (mostly *Cladina* spp.), covering 65% of the forest floor, while the remaining 35% is occupied by moss (mostly *Sphagnum* spp.). The top 0.15–0.2 m of active layer is composed of living vegetation and lightly decomposed fibric peat, below which lies a layer containing denser, more decomposed sylvic peat with dark, woody material, and the remains of lichen and moss, rootlets and needles. Small ponds occupy up to one fifth of the total plateau area during snowmelt, at which time some surface flows have been observed. However, normally the ground surface is relatively dry, and the lateral movement of water occurs as subsurface flow through the thawed, saturated zone between the water table and frost table.

## 2.2. Site Instrumentation and Soil Moisture Measurements

[7] Table 1 provides a summary of the following field measurements in terms of their spatial and temporal resolution and methodology. Two soil pits (see Figure 1b for location) were excavated to the frost table on 20 August 2001 to install PVC observation wells (0.1 m inner diameter), and soil temperature and moisture sensors. The observation wells at the Center and West pits were each instrumented with a pressure transducer (Global Water WL15) that measured the elevation of the water table every minute, and averaged and recorded these measurements every 30 min. The water table depth at each of the wells was also measured daily with a water level sounder and ruler during the spring field campaigns in 2004 and 2005, to ensure the accuracy of the transducer measurements.

[8] Only the temperature and soil moisture data from the Center pit were used in this study (data from the West pit were published by Hayashi et al. [2007]). The depth to the frost table in relation to the peat surface at the Center pit was 0.7 m at the time of installation, and the groundcover was composed of both lichen and moss. The Center pit was instrumented with thermistor probes (Campbell Scientific 107B) at depths of 0.05, 0.1, 0.15, 0.2, 0.25, 0.3, 0.4, 0.5, 0.6, and 0.7 m, and water content reflectometers (Campbell Scientific CS 615) at 0.1-, 0.2-, 0.3-, and 0.4-m depths. The water content reflectometers were calibrated against the water content of 17 peat samples (sample volume =  $92 \text{ cm}^3$ ) that were collected from the face of the soil pit at the time of installation. After the sensors were installed, the pits were backfilled with the excavated material, with care taken to preserve the original layering and groundcover. Using the water content reflectometer data, depth-integrated average water content was calculated for the unfrozen portion of the active layer at a daily interval.

[9] Soil water content was also measured daily during the spring study period (3–31 May) in 2005 at 15 flagged points, indicated as “soil sampling points” in Figure 1b. Seven of the 15 points were under moss and eight under lichen. The exact sampling locations were varied on a daily basis to avoid disturbance, though all samples were taken within 1 m of the flagged point, and at the same ground cover type. Soil moisture was measured gravimetrically at

**Table 1.** Overview of the Field Measurements, Including the Spatial and Temporal Resolution and Methodology

Variables Measured	Spatial Resolution	Period of Measurement	Temporal Resolution	Measurement Method
Water table depth	Center and West pit observation wells	2004–2005	half hourly	pressure transducer
Soil temperature	Center pit, 0.05–0.1-m depth interval	2003–2006	half hourly	thermistor probes
Soil water content	Center pit, 0.1-m depth interval	2003–2006	half hourly	water content reflectometers
	15 points, 0.05-m depth interval	3–31 May 2005	daily	in situ gravimetric measurements
	1 m along transect T1, 0.2-m depth	4 Sep 2006		time domain reflectometry
Air temperature	meteorological station at plateau	2003–2006	half hourly	thermistor in a radiation shield
	Fort Simpson airport	1999, 2002	daily	
Rainfall	meteorological station at plateau	2003–2006	half hourly	tipping bucket rain gauge
	Fort Simpson airport	1999, 2002	daily	
Snow depth	1-m interval, transect T1	1999, 2002–2006	daily	steel ruler
	1-m interval, transect T2	2004, 2005	daily	
Snow water equivalent	5-m interval, transect T1	1999, 2002–2006	daily	Eastern snow conference snow sampler and scale
	5-m interval, transect T2	2004, 2005	daily	
Frost table depth	2-m interval, transect T1	1999, 2002–2003 <sup>a</sup>		graduated steel rod (frost probe)
	1-m interval, transect T1	2004–2006 <sup>a</sup>		
	1-m interval, transects T1, T2, and T3	May–Jun 2004, 2005	daily	
	Center and West soil pits	Apr–Jun 2004, 2005	daily	
	15 soil sampling points	3–31 May 2005	daily	
	0.25-m interval in 5 m × 6.75 m grid	12 Jun 2006		

<sup>a</sup>See Table 2 for measurement dates.

0.05-m depth increments to the water table at each of the 15 points, using 162-cm<sup>3</sup> soil sample tins, and depth-integrated average water content was calculated. Near-surface soil moisture (0–0.2-m depth) was also measured every 1 m along transect T1 (Figure 1b) on 4 September using time domain reflectometry (TDR) (Soil Moisture Equipment Corp., MiniTrase), with a 20-cm probe inserted vertically into the soil.

[10] Air temperature (using a thermistor housed in a Gill radiation shield) and net radiation (using a Kipp and Zonen, CNR1 sensor) were measured 2 m above a moss-covered ground surface at a meteorological station located between the Center and West pits (Figure 1b). Rainfall was measured with a tipping bucket rain gauge (Jarek Manufacturing) that was located next to the meteorological station. All sensors were connected to Campbell Scientific CR10X data loggers, programmed to measure every minute and record hourly or half hourly averaged values. Rainfall data from the Fort Simpson weather station were used for analysis in 1999 and 2002, because the rain gauge was not installed on the peat plateau until 2003.

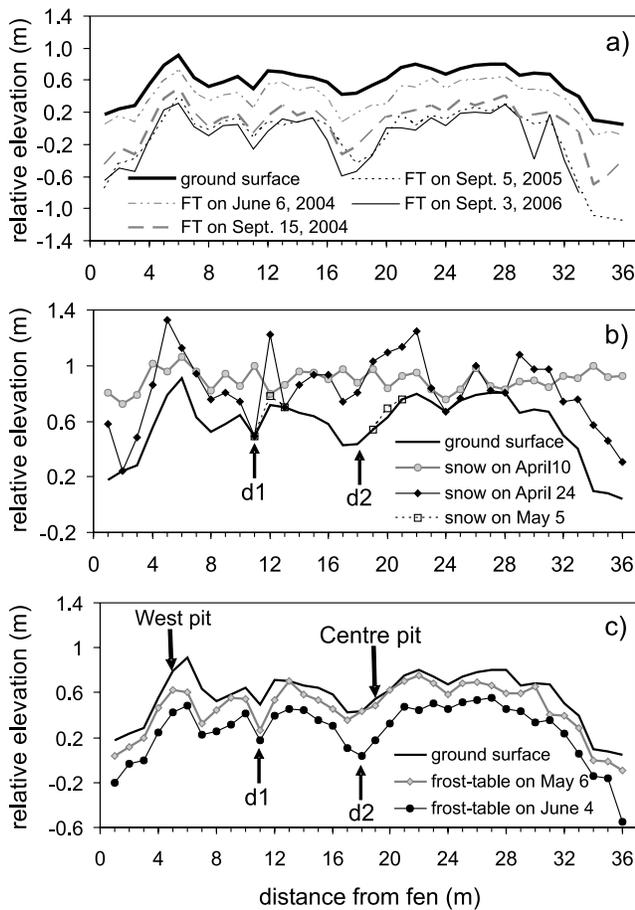
### 2.3. Frost Table Depth and Snow Measurements

[11] Snow depth and snow water equivalent (SWE) measurements commenced at the peak of the snow accumulation season, and were made daily using a steel ruler and an Eastern Snow Conference snow sampler (ESC-30) (B. E. Goodison, Accuracy of snow samplers for measurement of shallow snowpacks: An update, paper presented at the 34th Eastern Snow Conference, Hanover, New Hampshire, 1978) and scale (calibrated to ±3 mm). Snow depth was measured at 1 m, and SWE at 5-m intervals along a 36-m transect (T1) that traversed the width of the peat plateau (Figure 1b) in 2003–2006, and along an additional 28-m-long transect (T2 in Figure 1b) in 2004 and 2005. The average snow depth and SWE of the peat plateau at the peak of snow accumulation in 1999 and 2002 was measured by Environment Canada [Carter and Onclin, 1999; Bastien et al., 2002], in the same fashion as in 2003–2006.

[12] Depth to the frost table in relation to the peat surface was measured every 2 m in 2003 and every 1 m in 2004–2006 along transect T1 at the end of the spring field season (early to mid-June) and at the end of the summer field season (late August to mid-September). Frost table depths were also measured approximately every 2 m along T1 at the end of the summer field season in 1999 and 2002. In 2004 and 2005, the frost table depth was measured along T1, T2, and an additional 10-m-long transect (T3 in Figure 1b), every 2 days during the early soil thaw period (May to early June), at 1-m intervals. During the same time period, the frost table depth was measured at least every 2 days at both soil pits in 2004 and 2005, and daily at each of the 15 soil sampling points in 2005. All frost table measurements were made by probing the ground with a graduated steel rod, which clearly detected the top of the frozen layer. Subsurface temperature data at the Center pit were used to examine the accuracy of the timing of frost probe measurements in relation to the spring and maximum annual thaw depth. The ground cover type (moss or lichen), and any microrelief feature (depression or mound) relative to an arbitrary reference level, were recorded at each measurement point along the frost table transects.

[13] On 12 June 2006, the frost table depth was measured in a 33.8-m<sup>2</sup> plot on the plateau, next to the channel fen (Figure 1b), with a frost probe, every 0.25 m along the 21 transects that ran south to north, in a 5 m × 6.75 m grid with a spatial resolution of 0.25 m. Each of the 21 transects had a taught wire tied from end to end, leveled using a bubble level, from which the distance between the wire and ground surface was measured with a ruler, at the same 0.25-m interval as the frost table depth measurements. The elevation of the frost table,  $Z_{FT}$ , in relation to the elevation of the ground surface,  $Z_{GS}$ , was computed from the frost table depths,  $d_{FT}$ , as:

$$Z_{FT} = Z_{GS} - d_{FT} \quad (1)$$



**Figure 3.** (a) Seasonal and interannual variability of the frost table (FT) elevation along T1 (see Figure 1 for location). (b) Elevation of the ground surface and snow surface on transect T1 in April and May 2005 and (c) elevation of the ground surface and frost table in May and June 2005. Arrows with 1 and 2 indicate the depressions discussed in the text. West pit and Center pit indicate the approximate location of the soil pits.

where  $Z_{GS}$  was computed by subtracting the distance between the taught wire and the ground surface from an arbitrary datum, such that  $Z_{GS} = 0.4$  m at the fen water level. The ground surface and frost table elevation data were interpolated by the kriging method [Davis, 1973] to assign elevation on a 0.1-m grid using commercial software (Golden Software, Surfer).

### 3. Field Results

[14] In this study, “spring” is defined as the period beginning at the onset of soil thaw and ending on the last day of frost table measurements in early/mid-June. The onset of soil thaw is indicated by the rise of soil temperature above  $0^{\circ}\text{C}$  at 0.05-m depth at the Center pit, which was 4 May in 2004 (Figure 2a) and 27 April in 2005 (Figure 2b). Figure 2c shows the daily depth to the frost table measured by the frost probe near the Center pit as continuous lines, and the arrival of the thawing front at discrete depths, according to soil temperature data at the Center pit, as symbols. There was relatively good agreement between

the frost probe data and soil temperature data in 2004 (Figure 2c). In 2005, the two data sets were consistent for the onset of soil thaw, but frost table depth values had a difference of 0.05–0.1 m (Figure 2c). This difference was most likely due to the high variability of thaw depths found on the plateau (as described in section 3.1), as the frost probe measurements were made 0.5 m away from the soil temperature sensor array at the Center pit to avoid disturbance to the instrumented pit. Frost probe measurements were also made in late August to mid-September; these measurements are referred to in the following as “end-of-summer” frost table depths. Romanovsky and Osterkamp [1997] showed that the maximum frost table depth occurs approximately two weeks prior to the start of soil freezing. Figures 2a and 2b show that the reversal of temperature gradient occurs around 26–30 September, therefore, the end-of-summer frost probe measurements may not be maximum, but were fairly close, giving confidence that the frost table measurements are comparable in the different years.

#### 3.1. Frost Table Depth and Spatial and Seasonal Variation

[15] Figure 3a illustrates the variability of frost table elevations along transect T1 over relatively short distances (1 m) and its seasonal and interannual consistency. The points that had lower frost table elevation (i.e., deeper frost table) in June also had lower frost table elevation in September, and the points with lower frost table elevation in 2004 generally had lower frost table elevation in other years. Mean frost table depths measured in June of 2003–2006, were on average, 39% of the end-of-summer frost table depths measured over the same 4 years (Table 2). The locations with a deeper frost table in June tended to maintain higher thawing rates over the summer as well. This is illustrated in the frequency distributions of frost table depths measured on T1, T2 and T3, which were skewed with a tail toward deeper frost table depths; a pattern that persisted and became more pronounced during the thawing season (Figure 4). Taking the logarithm of the frost table depths resulted in a normal distribution and thus, the logarithm of depth was used to calculate the geometric mean and standard deviation, as well as the observed level of significance ( $p$ ) using two-tailed  $t$ -test. Frost table depths were deeper beneath moss ground covers compared to lichen covers ( $p \leq 0.04$ ) (Table 3a); a trend that persisted seasonally and interannually (Table 3a).

#### 3.2. Frost Table Depth and Interannual Variation

[16] The literature suggests that the two primary factors controlling frost table depth are soil temperature [e.g., Gray *et al.*, 1988] and soil moisture [e.g., Kane *et al.*, 2001]. Here we examine the influence of air temperature (used as a surrogate for near-surface soil temperature as it governs interannual variations in ground temperature) and soil moisture on the interannual variation of frost table depths measured at the peat plateau in spring and at the end of summer.

[17] To determine the interannual relationship between air temperature and frost table depth on the peat plateau, the accumulated degree days of thaw,  $ADD_T$  ( $^{\circ}\text{C}$ ) was computed by summing the daily average air temperatures for the period beginning at the onset of soil thaw and ending on the last day of frost table measurements in early to mid-June

**Table 2.** Average Daily Frost Table Depths Measured on the Peat Plateau at the End of the Spring and Summer Field Seasons<sup>a</sup>

Year	Day of Measurement	Spring Thaw Depth (m)	Day of Measurement	End-of-Summer Thaw Depth (m)
1999	-	-	18 Sep	0.57
2002	-	-	26 Aug	0.64
2003	10 Jun	0.26	26 Aug	0.60
2004	6 Jun	0.19	15 Sep	0.55
2005	4 Jun	0.27	5 Sep	0.69
2006	12 Jun	0.29	3 Sep	0.72

<sup>a</sup>Units expressed as meters below the ground surface. No measurements were made in the spring of 1999 and 2002.

of 2003–2006 (see Table 2 for the actual dates). The *ADDT* was also computed for the end-of-summer period ending in late August to early September in 1999, 2002–2006 (Table 2). It is commonly observed that frost table depths are strongly correlated with the square root of *ADDT* over a large scale [e.g., *Nelson et al.*, 1997; *Hinkel and Nelson*, 2003]. While frost table depths in June had a relatively strong correlation with the square root of *ADDT* ( $r^2 = 0.80$ ) (Figure 5), the end-of-summer frost table depths were poorly correlated ( $r^2 = 0.07$ ) to the square root of *ADDT* (data and linear regression not shown), where  $r^2$  is the coefficient of determination. The relatively low values of *ADDT* in the spring of 2004, compared to other years, was due to lower mean daily air temperatures and fewer days with above-freezing temperatures, which delayed the progression of the thawing front. For example, the average daily air temperature during the first 10 days of thaw in 2004 was  $-0.37^\circ\text{C}$ , compared to 2.9, 4.9 and  $5.7^\circ\text{C}$  in 2003, 2005 and 2006, respectively. The data suggest that air temperature (and thus, soil surface temperature) strongly affects thaw rates of the relatively shallow frost table in May–June when there is a greater temperature gradient between the ground surface and the frost table, but it likely has weaker effects on thaw rates in July–August, when the frost table is much deeper, as the effects of air temperature variation is dampened in the soil during downward heat conduction. Other studies have shown that snowpack can have a significant influence on soil freezing and thawing [*Bayard et al.*, 2005; *Iwata et al.*, 2008; *Luetsch et al.*, 2008]. However, no clear relationship was found at this particular site during the 4 years of study.

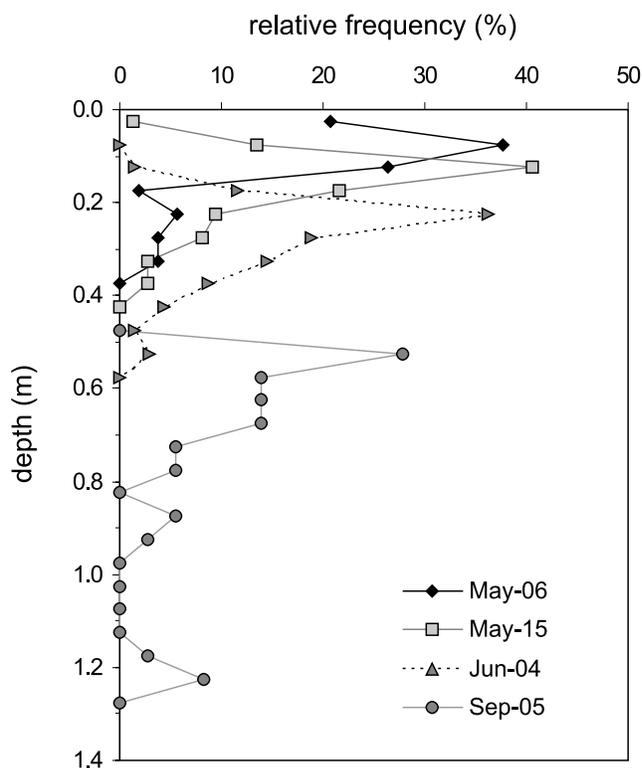
[18] To determine the interannual relationship between frost table depths and moisture conditions on the peat plateau, frost table depths were compared with the total water input (i.e., estimated snowmelt and rainfall) during the period between the peak of snow accumulation and early to mid-June of 2003–2006, and between the peak snow accumulation and late August to early September of 1999, 2002–2006. The total amount of snowmeltwater input was estimated from the average snow water equivalent (SWE) that was measured on the peat plateau at the peak of snow accumulation (late March or April depending on the year), as described in the methods. The average SWE was added to the cumulative rainfall, also from the peak of snow accumulation (to include rain on snow events), to the day when the frost table depths were measured at the end of the spring (early to mid-June) and summer (late August to early September) field seasons (Table 1). The relationship

between average spring frost table depths and total water input to the peat plateau was not strong over the 4-year period of 2003–2006 ( $r^2 = 0.15$ ) (Figure 6). Unlike the spring thaw depths, the average end-of-summer frost table depths were significantly correlated ( $r^2 = 0.978$ ,  $p < 0.01$ ) to the total water input to the plateau over the 7 years of measurement (Figure 6), with deeper frost table in years with greater water inputs, than in years with lower precipitation amounts.

[19] To examine the correlation between total water inputs and the soil moisture condition on the peat plateau, the depth-integrated average liquid water content data at the Center pit was used, which represents the average moisture conditions on the plateau reasonably well [*Wright et al.*, 2008], even though this is only one point. Total water inputs were highly correlated to the seasonal mean of water content at the Center pit over the same time period (to the end of summer) in 2003–2006 ( $r^2 = 0.911$ , data and linear regression not shown).

### 3.3. Spatial Variability of Soil Moisture, Snow Depth, and Frost Table

[20] Having identified a significant effect of water input (and moisture condition) on the interannual variability of frost table depths, the effect of soil moisture on the spatial variability of frost table depths was investigated. Figure 7a illustrates the positive correlation between spring frost table depth and depth-integrated average soil volumetric water content (VWC) at the 15 soil sampling points. Frost table depth at the end of summer (3 September 2006) was also related to average soil water content in the 0–0.2-m depth



**Figure 4.** The distribution of frost table depths measured at 40 to 74 points along T1, T2, and T3 on the peat plateau in 2005.

**Table 3a.** Geometric Mean Frost Table Depths, Standard Deviations, and Observed Level of Significance (*p*) for the Difference Between the Two Vegetation Covers of Moss and Lichen<sup>a</sup>

	June Geometric Mean Thaw Depths (m)			September Geometric Mean Thaw Depths (m)		
	Moss	Lichen	<i>p</i> -value	Moss	Lichen	<i>p</i> -value
2004	0.21 ± 0.05	0.18 ± 0.05	= 0.05	0.62 ± 0.16	0.49 ± 0.07	<0.01
2005	0.36 ± 0.06	0.27 ± 0.04	<0.01	0.69 ± 0.16	0.57 ± 0.07	<0.01
2006	0.30 ± 0.05	0.27 ± 0.04	= 0.02	0.81 ± 0.24	0.67 ± 0.11	<0.02

<sup>a</sup>Units expressed as meters below the ground surface. Note that vegetation cover was not recorded in 2003.

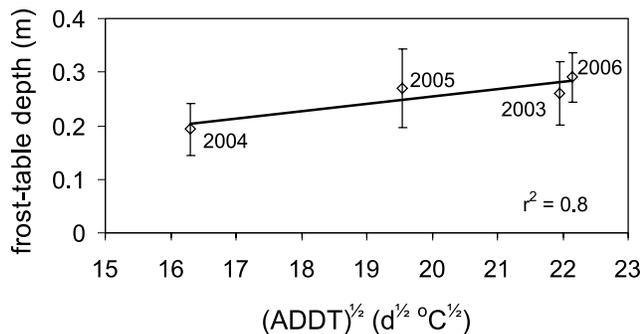
interval measured by TDR on transect T1 (Figure 7b). This correlation was positive as well, though strongly influenced by one data point (Figure 7b). This measurement point was located at the bottom of the slope next to the fen (a wetland), which may have influenced the greater saturation and frost table depth at this point.

[21] Since the frost table depths were correlated to soil moisture (Figure 7) and to ground cover types (Table 3a), a significant correlation is expected between soil moisture and ground cover. This was true during spring 2005 to depths of 0.15 m, but not for end of summer 2006 (Table 3b). In spring, the frost table and water saturated zone are close to the ground cover. *Sphagnum* mosses have hyaline cells which allow them to store water (up to 20 times their dry weight) [Vitt, 2000], and to retain capillary water far above the actual water table [Lafleur et al., 2005]. Lichens on the other hand, have no such water retention capabilities, and thus, act as a mulch layer providing the underlying soil with thermal insulation by decoupling the moist subsurface from the atmosphere [Rouse and Kershaw, 1971]. These differences control soil temperature and the rate of soil thaw in spring. However, at the end of summer the frost table and the water saturated zone are much deeper, and as neither mosses nor lichens have roots and are active in moisture exchange only when moist [Rouse, 2000], the correlation is weaker.

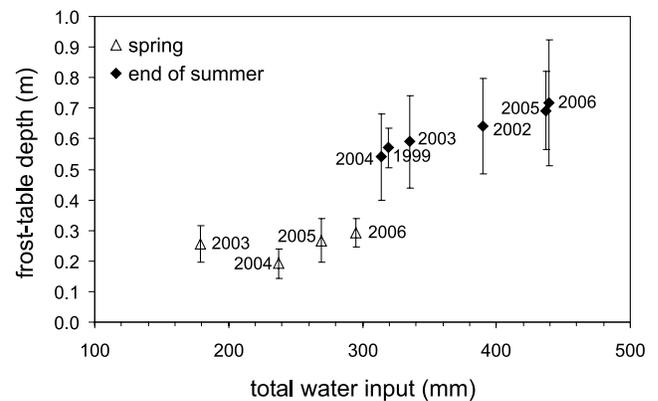
[22] To examine the effect of snow cover and topography, Figures 3b and 3c show the progression of snow cover and frost table depth along transect T1 from April to June 2005. Distances 11 m and 17–19 m are areas of relatively low ground surface elevation, and are hereafter referred to as d1

and d2, respectively. Both depressions had relatively similar snow depths on 10 April 2005 (Figure 3b). However, d1 became snow free on 24 April, and the frost table was among the deepest of all points on the transect on 6 May (Figure 3c); while d2 only became snow free on 5 May, and the frost table was very close to the ground surface on 6 May. The frost table depth at d2 reached a local minimum on 4 June despite the shallow frost table on 6 May (Figure 3c), indicating that the thawing rate at d2 was much greater than at d1, and most other points on T1. The main difference between d2 and the other points was that it was the only location where water was ponding above the ground surface (because of saturation of the soil) during snowmelt and for more than a week afterward. This suggests that the high rate of soil thaw at d2 is related to the wetness of this location.

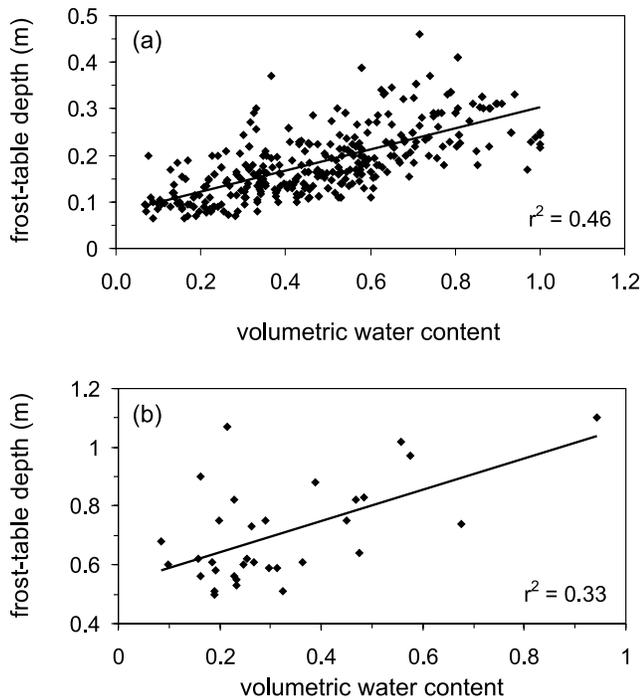
[23] The relatively deep frost table measured near the edges of T1 (at distances of 1–3 m and 34–36 m in Figure 2c) on 4 June, were largely due to the lateral transfer of energy provided by relatively warm water flowing in the ditch that surrounds the peat plateau (visible in Figure 1b on the bog side). This ditch conveys fen water downstream, and bog water around the peat plateau to the fen (during snowmelt). This is consistent with larger-scale studies that have found frost table depths to be greater in wetlands than in drier uplands [Nelson et al., 1997; Hinkel and Nelson, 2003].



**Figure 5.** Mean frost table depth at the end of spring versus the square root of accumulated degree days of thaw ( $ADDT^{1/2}$ ), computed over the spring of 2003–2006. Bars indicate one standard deviation.



**Figure 6.** Mean frost table depth at the end of spring and at the end of summer along T1, versus the total water input from the start of snowmelt to the end of spring in 2003–2006, and to the end of summer in 1999 and 2002–2006. Bars indicate one standard deviation.



**Figure 7.** Frost table depth versus volumetric water content measured across the peat plateau (a) at the 15 soil sampling points throughout May 2005 and (b) along transect T1 on 4 September 2006.

**3.4. Physical Link Between Soil Moisture and Thawing Rates**

[24] The data above have shown that the spatial and temporal variability of frost table depths is related to soil moisture conditions. This is consistent with the results of Hayashi *et al.* [2007], who used a heat conduction model with the soil temperature and moisture data from the Center and West pits to show that the difference in thawing rates between the two pits was due to the difference in thermal conductivity, as the thermal conductivity of peat is strongly dependent on water content. The Center pit is located in an area of relatively low ground surface elevation, while the opposite is the case for the West pit (Figure 3c). The frost table was 0.13 m deeper at the West pit than at the Center pit in this study when the measurements began on 1 May, possibly because the snow cover melted much earlier at the West pit than at the Center pit (Figure 3b). However, during

**Table 3b.** Mean Volumetric Water Content, Standard Deviations, and Observed Level of *p* for the Difference Between the Two Vegetation Covers<sup>a</sup>

Moss	Lichen	<i>p</i> -value
<i>June VWC, 2005</i>		
0.76 ± 0.17	0.49 ± 0.09	<0.01
<i>September VWC, 2006</i>		
0.36 ± 0.23	0.29 ± 0.17	ns

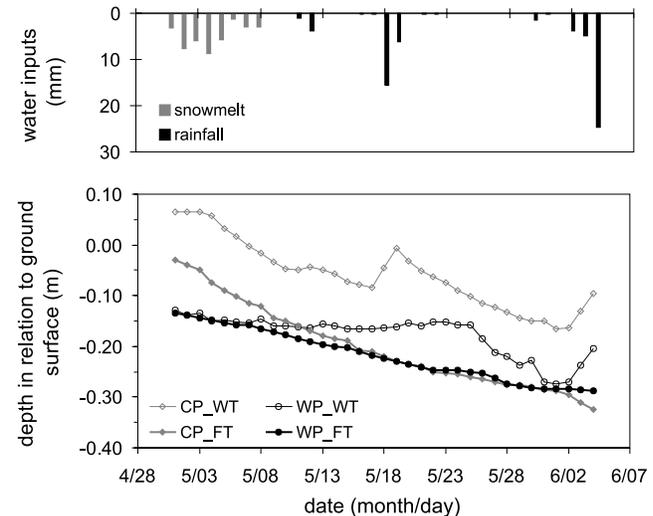
<sup>a</sup>Data were taken at the 15 flagged sampling points during the spring thaw period of 2005 and at T1 on 4 September 2006. VWC, volumetric water content; ns, data were not significantly different.

the 35-day measurement period, the mean soil thaw rate at the Center pit (9 mm d<sup>-1</sup>) was much higher than that at the West pit (4 mm d<sup>-1</sup>); so that by 4 June, the frost table at the Center pit was deeper (Figure 8). The greater thaw rate observed at the Center pit was largely due to the shallower water table measured at the Center pit compared to that measured at the West pit (Figure 8). Given that both pits had similar vertical inputs of snowmelt and rainwater, the difference in water table depths, and thus, soil thaw rates between the two sites, must be due to the convergence of lateral flow to the depression containing the Center pit, as this process (1) maintains a high soil moisture (and thermal conductivity), thereby enhancing vertical heat conduction, and (2) provides additional energy for ground thaw by lateral advection.

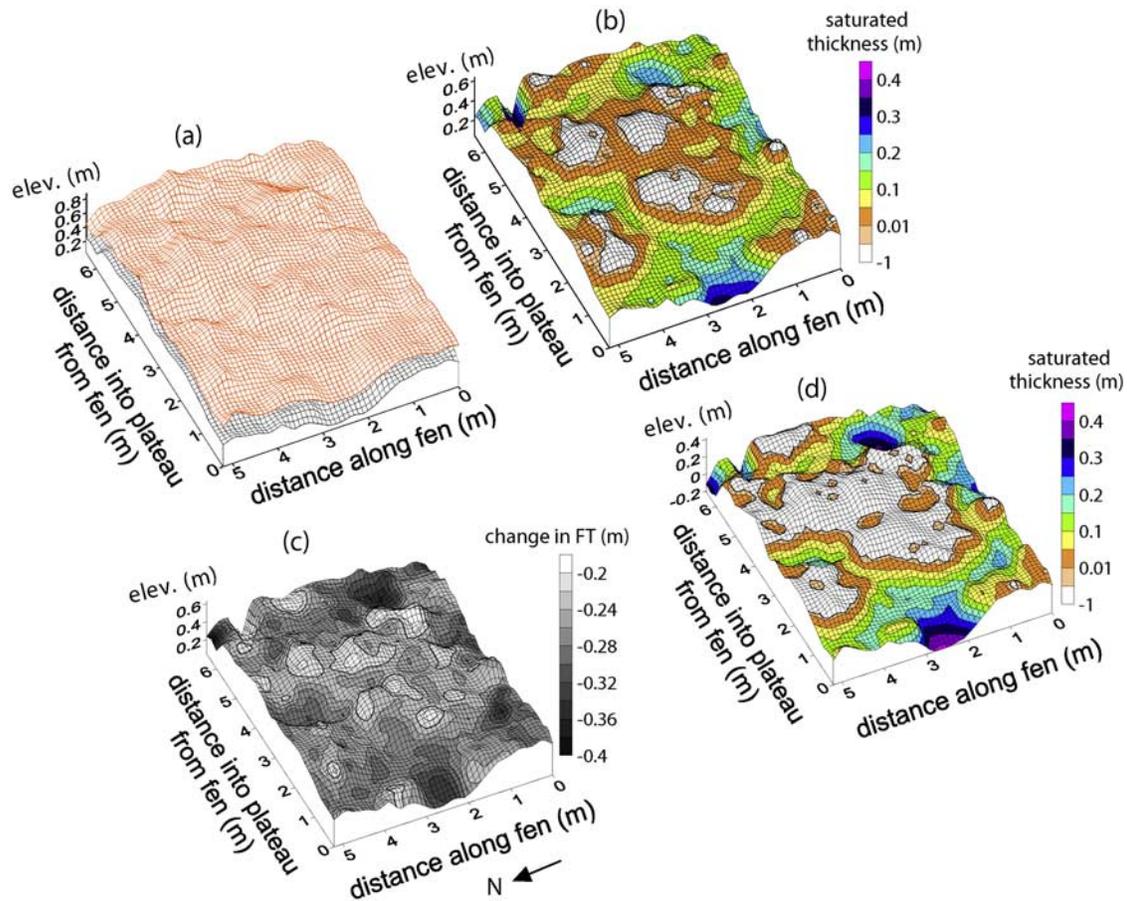
[25] There have been studies suggesting that the correlation between soil moisture and thawing rates is partially explained by the vertical advection of heat by infiltration of relatively warm rainwater [e.g., Hinkel *et al.*, 1997] and snowmeltwater [e.g., Rist and Phillips, 2005]. We used the soil temperature and moisture data from the Center pit to examine the influence of rainwater infiltration on soil thawing rates. The vertical infiltration of water during the spring rain events of 2004 and 2005 had little effect on soil thawing rates; if anything, soil thaw rates decreased during periods of rain, when both the air temperature and near-surface soil temperature decreased. This suggests that the effect of soil moisture on the ground thermal regime at the peat plateau is primarily due to its influence on thermal conductivity, while the advection by rainwater infiltration may play a secondary role.

**3.5. Spatial Variability in Two Dimensions**

[26] While the variability of the surface and frost table topography is shown one dimensionally at 1-m intervals across the plateau in Figure 3, the thickness of the thawed layer as a composite of the surface and frost table topography is shown two dimensionally at smaller measurement intervals of 0.25 m at the frost table plot (Figure 1b) in Figure 9a. The overall slope of the ground surface was very



**Figure 8.** Daily snowmelt and rainfall plotted with the frost table (FT) and water table (WT) depths, measured daily at the Center (CP) and West (WP) pits in 2005.



**Figure 9.** (a) Variability of the ground surface (orange grid) and frost table (black grid) elevation (elev.) on the peat plateau (see Figure 1b for location). Note that both surface and frost table topography were measured from the same datum. (b) Simulated thickness of the saturated zone (i.e., distance between the water table and the frost table) in relation to the frost table elevation measured on 12 June, with a vertical water input of 15 mm; (c) simulated change in the frost table (FT) (grayscale contours) between 12 June and 11 July, overlaid on the original frost table topography; and (d) simulated thickness of the saturated zone in relation to the frost table elevation simulated on 11 July, with a vertical water input of 15 mm.

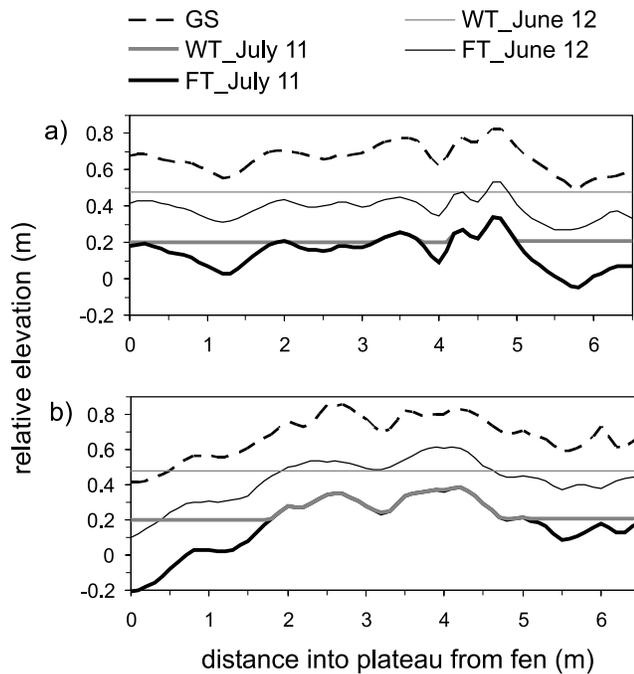
small, 0.003 in the east-west direction (referenced distance into peat plateau) and  $-0.005$  in the north-south direction (referenced to as the distance along the fen). However, the ground surface and frost table had significant relief, with surface elevation ranging from 0.41 to 0.88 m above the survey datum, and frost table elevation ranging from 0.04 to 0.64 m above the datum. The surface and frost table topography were not spatially constant, suggesting large variability concerning the soil moisture and material composition of the frost table plot, which has important implications for runoff generation from the plateau.

#### 4. Simple Flow Model Simulation

[27] As soil hydraulic conductivity increases with increasing water content, water preferentially moves through areas of greater soil moisture [Woo and Steer, 1983; Kane et al., 2001]. This has special implications for permafrost regions. Woo and Steer [1983] suggested that uneven thaw rates, which generates an irregularly shaped frost table (e.g., Figures 3 and 9a), could result in lateral preferential flow paths as described above (section 3.4). This situation is similar to hillslopes studies conducted in temperate regions,

where irregularly shaped bedrock surfaces have a strong influence on flow [e.g., Freer et al., 2002]. The hillslope flow controlled by the topography of a bedrock surface is referred to as “fill and spill” [Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006]. Weiler and McDonnell [2004] demonstrated the usefulness of numerical experiments using a simple numerical flow model, HillVi, to investigate how the complex bedrock topography enhances the variability of subsurface runoff. A simple flow model, similar to HillVi, was developed in this study to demonstrate the effects of the frost table (instead of the bedrock) topography on subsurface flow on the peat plateau.

[28] The model Simple Fill and Spill Hydrology (SFASH) solves the two-dimensional Dupuit-Forschheimer flow equation [Wigmosta and Lettenmaier, 1999, equation 1] using the finite difference spatial discretization and implicit time scheme [Huyakorn and Pinder, 1983, p. 56]. Similar to HillVi, vertical water input is added directly to the water table. However, unlike HillVi, saturated hydraulic conductivity is independent of depth in the current version of SFASH, which is meant to be a simple demonstrative tool. Therefore, transmissivity in the Dupuit-Forschheimer equa-



**Figure 10.** Cross sections of the model domain (see Figure 9a) at (a) 1.0 m and (b) 2.6 m along the channel fen, showing the ground surface (GS), frost table elevation (FT) measured on 12 June and simulated on 11 July, and the water table elevation (WT) simulated by SFASH on 12 June and 11 July with a vertical water input of 15 mm. The frost table on 11 July was generated using SFASH coupled with a simple heat conduction algorithm. Note that the fen-plateau boundary is point 0 on the  $x$  axis.

tion is given by saturated hydraulic conductivity times the thickness of the saturated zone (i.e., distance between the water table and frost table). Transmissivity between two grid nodes is set to zero when the node on the up-gradient side is completely unsaturated. The nonlinear flow equation is solved iteratively using a modified Picard method [Celia *et al.*, 1990, equation 16] and a variable time step size to ensure the convergence of numerical solution.

[29] The frost table in Figure 9a represents the physical boundary of the SFASH model domain on 12 June 2006. The values of saturated hydraulic conductivity ( $10^{-4} \text{ m s}^{-1}$ ) and drainable porosity (0.21) were chosen on the basis of the soil properties measured at the peat plateau by *Quinton and Hayashi* [2005, Figure 3]. For simplicity, it was assumed that (1) the peat was completely drained to field capacity prior to the input of water to the domain; (2) water inputs were directly delivered to the water table, and were not retained in the unsaturated zone, as peat soils are highly permeable when they are wet during infiltration events; and (3) there was no lateral input or output of water across the domain boundary (no-flow condition). The last assumption may not be representative of the field condition, but it was used to demonstrate the effects of fill and spill. In the simulation runs, a set amount of water was instantaneously added over the entire model domain at time = 0, and allowed to flow laterally until the steady state was reached. The purpose of this exercise was to examine how the soil wetness of the domain affects the connectivity of saturated areas.

[30] Figure 9b shows the simulated thickness of the saturated zone overlying the three-dimensional surface of the frost table, with 15 mm of water added to the domain. The simulated distribution of saturated areas suggests that it is largely controlled by the frost table topography, as they followed the topographic lows; with deeper water tables in areas with lower frost table elevation (i.e., deeper frost table). Simulations with different amounts of water input (5–30 mm) showed that the spatially variable frost table topography results in highly convoluted, tortuous subsurface flow paths with small amounts of water input. As water input increases, the maximum extent of the saturated areas increases, as does subsurface channeling in the downslope direction, which increases the area contributing to runoff from the plateau.

[31] In order to elucidate the effects of differential thawing on frost table topography and subsurface flow, SFASH was coupled with a simple heat conduction algorithm of *Hayashi et al.* [2007, equation 3], in which the thermal conduction is driven by the temperature gradient between the ground surface and the frost table, and thermal conductivity is dependent on soil moisture content [Hayashi *et al.*, 2007, equation 8]. The thawing algorithm was applied at each time step and meltwater was treated as water input in SFASH to simulate the simultaneous deepening of the frost table and the lateral redistribution of liquid water. Starting from the 15-mm input scenario presented in Figure 9b, the model simulated the thawing and flow for 30 days, assuming the no-flow boundary condition on all four sides of the model domain. In this simple simulation, surface temperature was assumed constant and equal to the average air temperature ( $18^{\circ}\text{C}$ ) during 12 June to 11 July 2006. After the 30-day simulation was complete, all excess water (above the field capacity) was drained from the model domain to create the initial condition for the second rainfall experiment, in which 15 mm of rain was added and SFASH was run without the heat conduction algorithm until the steady state was reached.

[32] Figure 9c shows the change in the frost table elevation (grayscale contours) resulting from the simulated 30-day thawing, overlaid on the original frost table topography. The largest change in frost table elevation generally occurred in the wetter areas. Figure 9d shows the simulated frost table topography and the thickness of the saturated zone, which resulted from 15-mm rainfall on the simulated frost table. Comparing Figures 9b and 9d, more areas are exposed above the water table in the simulated topography. This is because the valleys and depressions in the frost table, where the meltwater converges, are more deeply incised in the simulated than in the original frost table because of differential thawing of the wet areas. As a result, a larger amount of rainfall would be required to connect the flow pathways on the simulated frost table.

[33] A closer inspection of the individual transect cross sections at 1 m (Figure 10a) and 2.6 m (Figure 10b) along the channel fen, gives further insight into how the spatial and temporal variability of frost table depths affects the lateral flow of water downslope. Frost table depressions located in the middle and upper reaches of the slope (shown at both transects) act as a barrier to the lateral drainage of water downslope to the channel fen. As the soil thaws and the depressions deepen, a larger amount of water is needed

to fill them before the water stored upslope can connect to flow downslope. Thus, the location and the volume of frost table depressions on the plateau (and their evolution with thawing), and the amount of water input, determines the degree of saturated connectivity, and thus, affects the timing and magnitude of runoff. Therefore, differential thawing caused by the variability in soil wetness provides a feedback mechanism between the thermal and hydrological processes.

## 5. Discussion

[34] The SFASH simulation showed that the frozen, saturated layer of the peat plateau affects the movement of water (by creating a barrier or conduit to lateral flow) in the same way (but maybe at a different scale) that the confining subsurface topography (e.g., bedrock) controls the flow from slopes in nonpermafrost locations [e.g., *Buttle and McDonald, 2002; Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006*]. A major difference between the hillslope studies highlighted above, and this study, is that the aquiclude of the peat plateau is ice-saturated, frozen ground (not bedrock or glacial till), which undergoes an annual freeze-thaw process. Thus, the association between the flux and storage of water evolves as the active layer thaws [*Woo and Steer, 1983; Wright et al., 2008*]. As shown in the field and simulation results, when the lateral flow of water (from rain, snowmelt, and ground ice melt) converges to depressions in the frost table, the water table in the depressions is maintained close to the ground surface. This causes the rate of soil thaw at these points to increase, because wet peat has a higher thermal conductivity than dry peat. As depressions in the uneven frost table are able to store water, they are also an important source of runoff when the active layer is not saturated (particularly in summer); a water source that is “rapidly released when part of the frozen sill is breached by continual thawing” [*Woo and Steer, 1983*]. Thus, the depth of water input needed to produce a spill in the downslope direction increases as the frost table deepens and varies in topography.

[35] High-resolution mapping of the frost table provided detailed information on the frost table topography. Combined with a simple hillslope water flow and heat transfer model, this provided new scientific insights into the effects of the spatial and temporal variability of frost table on subsurface flow on the peat plateau. However, the parameterization of water storage and redistribution used in the current model is too simple for realistic flow simulations. To advance our understanding of the flow processes in the active layer, the model needs to incorporate depth-dependent hydraulic conductivity, which is the essential feature of peat-covered terrains in northern Canada [*Quinton et al., 2000*]. Such a model will provide a valuable tool to explore the feedback between soil moisture and thaw, its effect on water transport and runoff from permafrost slopes, and eventually the response of the permafrost slopes to climatic fluctuations.

## 6. Conclusions

[36] The frost table depths measured on a forested peat plateau, located in the discontinuous permafrost region of northwestern Canada, exhibited high spatial variability over

relatively short distances (0.25–1 m). There was a seasonal consistency in the frost table depth distribution, as the locations with a deeper frost table in spring tended to have higher thawing rates over the summer; a pattern that persisted over the 4-year study period. While the thaw rates in spring (late April to early June) were strongly correlated to air temperature, the correlation was weak during the summer months. In contrast, rain and snowmeltwater inputs to the plateau had a significant effect on the interannual variability of frost table depths in both spring and summer, with deeper frost table depths in years with greater water inputs, indicating the importance of the soil moisture condition in influencing the spatial variability of frost table depths. As the thermal conductivity of peat increases with water content, deeper frost table in wetter areas is due to higher rates of thermal conduction providing melt energy to the frost table. The ground covers, lichen and moss, were found to have different controls on soil temperature and development of soil frost, because of their differences in moisture retention capabilities during spring.

[37] Simple subsurface flow simulations illustrated that lateral subsurface flow on the peat plateau is governed by the frost table topography having spatially variable storage that has to be filled before water can spill over to generate subsurface flow. This has important implications for the magnitude and timing of runoff from peat plateaus, as the relief of the frost table acting as aquiclude, and the amount of water input (apart from the soil’s hydraulic conductivity), determine if and when water is exported from the slope. The field data and simulations also suggest that the soil moisture distribution is influenced by lateral flow converging to frost table depressions. Coupling the flow simulations with a simple heat conduction algorithm helped elucidate the role of the evolving frost table and the feedback between thawing and lateral advection.

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