Impact of Linear Disturbances on a Discontinuous Permafrost Peatland Environment

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by

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Introduction

Permafrost is ground that remains at or below 0°C for at least two consecutive years (PIWP, 2012). In the continuous permafrost zone its thickness may exceed 300 m, in lower latitudes, permafrost thickness is progressively decreasing, its occurrence becomes discontinuous and eventually sporadic (PIWP, 2012). At the southern fringe of permafrost in northwestern Canada, just about few meters. In subarctic peatlands permafrost is one of the main elements governing the hydrological processes. Being very thin and with temperature just fractions of the degree below 0^{o}C, it is very vulnerable to any changes in soil temperature and water content. Permafrost in peatlands is found exclusively under peat plateaus, which are elevated 1-2 m over fens and bogs that do not contain perennially frozen ground. In this environment the peat plateaus act as natural frozen dams between such water bodies as fens and bogs. If the surface of a peat plateau is disturbed naturally, as by forest fires, or artificially, as by human activity, the function of the peat plateau as a frozen dam is diminished with the permafrost thaw. The effects of ground surface disturbance propagates through the water, ice, soil and vegetation components of peatland ecosystems (Woo, 2011). Permafrost degradation can occur from the top due to the changes in ground surface conditions, from the sides by heat conduction and convection from the surrounding permafrost free terrains and from the bottom due to the geothermal heat flux. The combination of two or more of these scenarios can lead to rapid permafrost thaw and its complete disappearance.

While climate warming is the main natural cause of permafrost degradation in the zones of discontinuous permafrost, economic development of natural resources resulting in considerable ground surface disturbance can also lead to permafrost thaw. Vegetation has a profound effect on the energy balance at ground level. For example, in winter time trees intercept large amounts of snow, creating conditions for deeper freezing of the soil. In summer time tree crowns create a shade and as a result, the amount of short wave radiation, which reaches the ground level decreases, thereby decreasing soil temperature. Peatlands are widespread in regions of permafrost. Peat has physical properties that are different from the physical properties of mineral soil. Undisturbed, it contains a large amount of air and has very low thermal conductivity, thus heat does not penetrate into the deeper soil layers in the summer. Conversely, in winter wet frozen peat has a high thermal conductivity. Consequentially, peat freezing in winter prevails over thawing in summer, creating favorable conditions for permafrost generation.

Vegetation removal and peat layer compaction are the most common and usually unavoidable types of environmental damage. This creates an imbalance in the thermal regime of the active layer as well as of permafrost. The process of permafrost degradation can develop so rapidly that within a few years the landscape can be completely transformed.

The cutting of seismic lines is one of the most widespread forms of surface disturbance in the northern ecosystem. Compared to modern practices, the seismic explorations of the 20th century were less concerned with environmental impact. As a result of the former relaxed regulations, the magnitude of negative effects of seismic geophysical activities in subarctic environment is hard to overestimate. The wide10-15 m seismic lines bulldozed through boreal forest and peatlands caused soil waterlogging and shifting of the water drainage patterns. Soil contamination and erosion are also the direct outcome of the seismic exploration of the past century. The disappearance of permafrost from below the seismic lines leads to surface subsidence and changes to the balance between mass and heat fluxes. The observations show that seismic lines often convey water over their surfaces and as a result are able to connect previously hydrologically isolated landscape elements such as fens and bogs, at the same time collecting drainage water from the adjacent undisturbed peat plateaus.

Knowledge of permafrost distribution below the linear disturbances helps understand the hydrological role of such disturbances in the low Arctic wetland ecological complex.

The thesis consists of two manuscripts which examine the effects of linear disturbances in a Scotty Creek basin through a combination of visual observations, instrumentation measurements, GPR surveys and thermal modeling. The main subject of the study is the analysis of hydrological effects of seismic lines and development of tools and methods, which can be used for quantitative and qualitative evaluation of permafrost degradation and talik development in the area of discontinuous permafrost. The first Manuscript focuses on the development of conceptual model of permafrost degradation under the seismic line based on the field data collected during the 2012-2013 summer and winter research seasons. The manuscript also presents the analysis of active layer and talik temperatures at various locations of the examined seismic line as well as the interpretation of the GPR data.

Based on the conceptual model described in the first manuscript, we developed a mathematical model of the active layer and permafrost thaw which is described in the second manuscript. This model allows us to make a prediction of when the examined seismic line becomes free of permafrost free. We also demonstrated that seismic lines create a permanent link between previously disconnected hydrological units.

A brief summary of the main research findings explaining how linear disturbances of peat plateaus contribute to permafrost degradation concludes the thesis. Furthermore, the final chapter identifies gaps in the existing body of scientific knowledge on the subject of permafrost degradation and presents recommendations on potential areas for future research.

Hydrological impacts of seismic lines in the wetland-dominated zone of thawing, discontinuous permafrost, Northwest Territories, Canada

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Abstract

Intensive seismic exploration in the Northwest Territories began in the late 1960s. Since that time, the legacy of seismic surveys - *i.e.* strait lines cut through boreal forest and tundra remains visible throughout northern Canada and Alaska (ECOG, 2012). The modern practice of seismic exploration is designed to minimize the impact on ecosystems (NLUS, 2003). No summer cuts are aloud anymore, however, 5-8 m wide lines are still created and heavy machinery still used to clear a linear route for geophysical surveys (Williams, 2012). Regions of discontinuous permafrost are especially vulnerable to disturbances, and consequently the impact of seismic surveys is most severe in such regions, where permafrost is relatively warm and thin, and thermally insulated by a continuous, well-drained organic. The removal of trees and compaction of the ground surface alters the thermophysical properties of the active (*i.e.* seasonally thawed) layer to such an extent that the underlying permafrost seriously degrades or even disappears completely (Williams *et al*., 2013). Such a transformation along linear corridors that cut indiscriminately across different terrain types with contrasting hydrological functions (Turetsky and St. Louis, 2006) has potentially serious implications to the redistribution of water and energy within and among landscape units with feedbacks to permafrost thaw, landcover change and runoff generation. This paper characterizes the flow and storage of water and energy along a seismic cut-line in the high-boreal zone of discontinuous permafrost in order to improve the understanding of these processes, their interactions and hydrological implications. As such, this paper lays a conceptual foundation for the development of numerical models needed to predict the hydrological and thermal impact of seismic lines in this sensitive region.

Introduction

Permafrost is ground that remains at or below 0°C for at least two consecutive years (PIWP, 2012). In the Mackenzie River valley of Canada's Northwest Territories (NWT) (Figure 1), permafrost thickness decreases with decreasing latitude from over 300 m at 68° N to approximately 10 m at 62° N (PIWP, 2012) at south border of Northwest Territories. The zone between approximately 58°N and 62°N is considered the southern fringe of permafrost (Kwong and Gan, 1994), where permafrost is discontinuous, relatively thin - less than 50 m and typically occurs below tree-covered peat plateaus that rise 1-2 m above the surrounded permafrost-free and tree-less wetlands (*i.e.* flat bogs and channel fens) (Beilman and Robinson, 2003). Climate warming-induced permafrost thaw occurs at the highest rate in this southern fringe, where insolation is relatively high, and where vertical heat conduction through the active layer is augmented by lateral heat advection from adjacent permafrost-free terrains. The permafrost temperature in this zone is already at or a fraction of a degree below the melting point (Burgess, and Smith,. 2000), thus very little additional energy is required to initiate permafrost thaw. In peat land environment permafrost thaw results in ground surface subsidence, the tree-covered plateaus overlying permafrost are flooded as they subside, and consequently are transformed to wetlands. Thus permafrost thaw in the southern fringe zone results in both the local disappearance of permafrost and a landcover transformation from forest to wetland (Osterkamp, 1983), simultaneous processes that are occurring over a wide geographic region (Rowland *et al*., 2010).

Disturbance of the ground surface often creates positive feedback processes (Williams, 2013) that accelerate thaw and landcover transformations initiated by climate warming. When the ground surface is disturbed, the effects propagate through the water, ice, soil and vegetation components of permafrost ecosystems (Woo, 2011). For example, in the presence of foliage short wave radiation reflects off the crown of the trees. Consequently, the amount of heat that reaches the ground surface beneath the trees is significantly less than in open areas. The removal of the overlying tree increases insolation at the ground surface, resulting in more energy available for conduction into the ground. Compaction of the ground surface increases the soil moisture content of the near-surface peat along the line thus increasing its thermal conductivity. These changes accelerate the rate of seasonal active layer thaw and the rate and depth of thaw penetration, resulting in the thaw of the underlying permafrost and subsidence of the ground surface (Williams and Quinton, 2012).

Degradation of permafrost below seismic lines can alter water flow and storage processes with implications to the water balance at local and regional scales. Being saturated with ice and a small fraction (<15% of total water content) of water, permafrost in this zone of high wetland coverage is relatively impermeable, as is the frozen portion of the active layer. As such, the zone of liquid water storage and flow is limited to the unfrozen portion of the active layer and, if present, to a talik (*i.e.* perennially unfrozen) layer situated between the active layer and permafrost. The fragmentation and eventual disappearance of permafrost below a seismic line results in a linear, permafrost-free corridor that cuts through a permafrost body, connecting adjacent permafrost-free terrains, thus allowing the possibility for heat and mass transfer between them. Because permafrost thaw below the line lowers the ground surface, there is the possibility that subsurface flow through the line can be augmented by surface flow over it. A length of seismic line traversing a peat plateau imports water from upslope wetlands intersected by the line and from the undisturbed forest on both sides of the length. Overland flow along seismic lines is often observed during periods of high moisture supply, such as spring snowmelt and in response to large summer rain events. Knowledge of permafrost distribution below linear disturbances, the rate and pattern of its decay, and the possibility of its regeneration, contributes to the understanding of the overall hydrological impact of seismic disturbance in the southern fringe of permafrost, and ultimately to a new numerical predictive capacity of the same. The objectives of this paper are to: (a) present the results of a field study on ground thaw and water flow and storage processes along a seismic line in the wetland-dominated zone of discontinuous permafrost; (b) Comparing disturbed (seismic line) and undisturbed (areas adjacent to seismic line) examine how these processes change over the thaw season and over time since the initial disturbance; and (c) propose a new conceptual model that describes these changes, and can be used as a basis for numerical simulations.

Study Site

Site characteristics:

The 152 km^2 Scotty Creek drainage basin (61°18' N, 121°18' W) is located about 60 km South East of Fort Simpson, NT, Canada in the zone of discontinues permafrost (Figure 1).

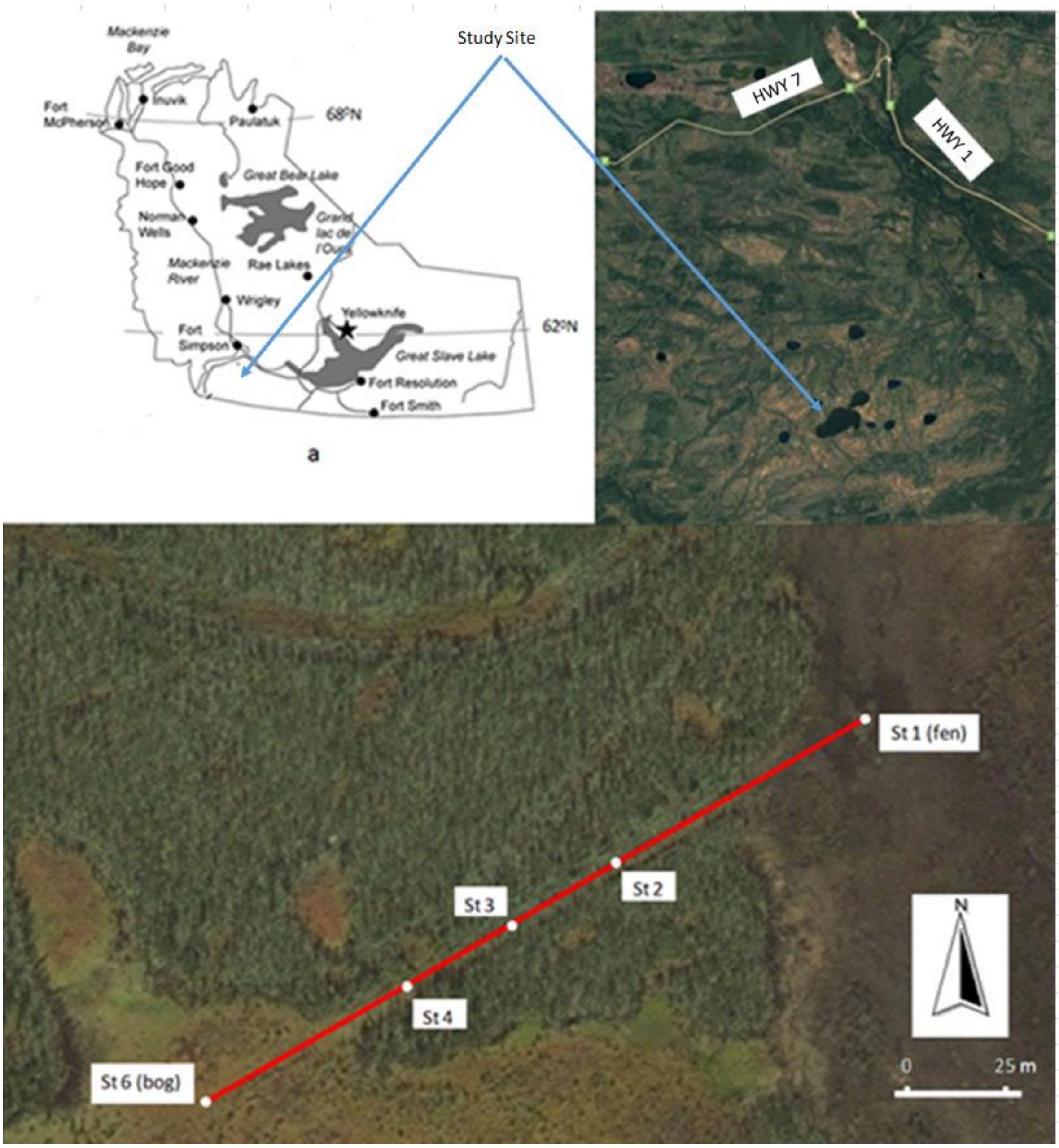

Figure 1: a) the Scotty Creek study site within the NWT, b) the examined seismic line (red) and the locations of the measurement stations (white circles). The forested areas (green) are peat plateaus, underlain by permafrost. The remaining area in the image is wetland, including fen (darker surface on the right of the image) and bog.

The 1964-2013 climate normals indicate that Fort Simpson has a dry continental climate with short, dry summers and long, cold winters. On average, the annual air temperature at Fort Simpson is -3.2^{o} C, and the annual total precipitation is 388 mm, of which 38% is snow (MSC, 2010). The study area is predominantly flat, and gently sloping toward the North-West. As in much of the southern fringe of permafrost, the permafrost at Scotty Creek occurs predominantly below peat plateaus, the forested “islands” within a treeless terrain of flat bogs and channel fens. The contrasting biophysical properties of these three peatland types, give each a specific role in the water cycle. Plateaus have a limited capacity to store water input owing to their relatively thin active layer. Runoff generated from their sloping surfaces, is conveyed into adjacent wetlands (Quinton and Baltzer, 2013). As such, plateaus function primarily as runoff generators, with runoff occurring predominately through the thawed, saturated layer between the water table and the relatively impermeable frost table below it. Bogs receive most of their water from precipitation and they are predominantly water storage features. The fens are broad and hydraulically-rough channels that convey water toward basin outlets. The water table in the wetlands (*i.e.* bogs and fens) remains at or within a few centimeters of the ground surface (Williams *et al.,* 2012). By contrast, the plateaus are mantled by an unsaturated layer that may exceed 50 cm during summer. The high thermal resistivity of this layer is critical to preserving the underlying permafrost. The active layer on the plateaus, containing both the water table and frost table, ranges in thickness between 50 and 120 cm.

Measurement stations

Five measurement stations were installed along a 195 m long segment of a seismic line that extends across a 90 m wide peat plateau that separates a channel fen on the east side of the plateau from a flat bog on the west side (Figure 2). Station 1 (St 1) was located in the fen, approximately 75 m from where the seismic line enters the plateau. St 2 and St 3 were situated about mid-way between the bog and the fen where the elevation of the seismic line is greatest, and St 4 was located on the plateau on a floating peat mat at a distance of 10 m from where the seismic line entered into the bog. The next station, named St 6 (*i.e*. St 5 malfunctioned) was placed in the middle of the bog, 60 m from St 4.

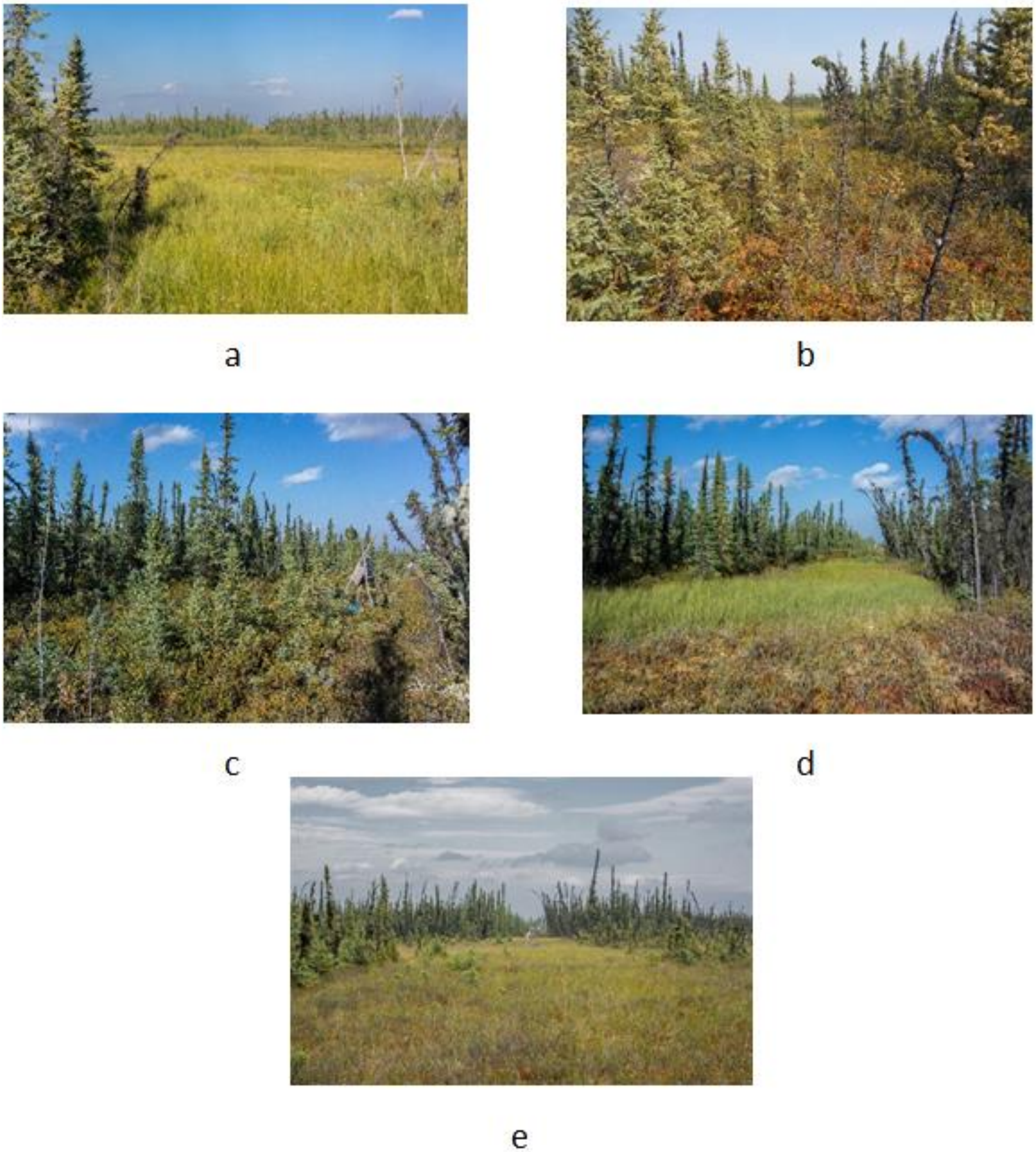

Figure 2: Photo of instrumented sites: a) St 1, b) St 2, c) St 3, d) St 4, e) St 6

The fen contains areas of open water that appear ponded or are slowly moving. Other areas of the fen support vegetation that conceals the water surface. As such, flow over the fen is highly diffuse and tortuous with rills meandering through sedge tussocks (*Carex stricta*) and other aquatic vegetation. A sphagnum (mainly *Sphagnum fallax*) mat is the lowest level of living vegetation on the fen surface. At St 2 on the plateau, the ground surface is saturated and covered with sphagnum moss. Black spruce (*Picea mariana*) 0.5 - 2 m tall sparsely occupy the north (*i.e.* south-facing) side of the seismic line while no trees are present on the opposite side. Labrador tea (*Ledum latifolium*) covers the entire width of the line. The ground surface slopes toward the south side where the forest floor of the adjacent peat plateau is about 1 m above the seismic line. However, the slope of the ground surface between the forest and seismic line is much greater on the south (70^o) than on the north (5^o) side of the line. At St 3, the ground surface is much dryer than at St 2. Black spruce occupy about 80% of the ground surface at this site, and rise to a maximum height of roughly 2 m on the north side and 0.6 m on the south. As at St 2, Labrador tea covers the entire width of the line at St 3. The south side of the line at this site is also characterised by black spruce trees at the forest edge leaning into the seismic line in a way often described as “drunken forest”, a clear indicator of permafrost thaw. By contrast, the forest edge on the northern side appears to be re-generating with numerous black spruce saplings present, as well as willow shrubs. At St 4, about 80% of the seismic line width is occupied by a floating peat mat. As a result, the water table remains close to this unstable peat surface (Fritz *et al.*, 2008) composed of sphagnum moss with sparse sedge tussocks. The surface at the St 6 site is covered by the light brown wet sphagnum moss with sparse sedge tussocks. Black spruce trees 0.5 - 1.0 m are present elsewhere in the bog, but not along the seismic line.

Methods

Ground surface and active layer properties

High-resolution (1 m horizontal, 0.2 m vertical) air-borne LiDar data was collected in July, 2010 by the Applied Geomatics Research Group, NS, Canada. The processing and interpretation was completed and described by Chasmer *et al.* (2008). Measuring points were located at regular intervals of 10 m along the 90 m plateau portion of the seismic line, a transect perpendicular to the line was used to measure the depth of thaw at 1 m intervals for a total of 7 measurements on each transect. Measurements were made using a 2 m long graduated steel frost probe bi-weekly between May 28 and July 25, 2013. Five boreholes ranging in depth between 2.2 m and 6 m were excavated at the stations described above. Permafrost was detected at St 2 (2.2 m), St 3 (1.7 m) and St 4 (2.6 m). St 1 (fen) was equipped with three thermistors (0.2, 0.5, 1.0 m), St 2 with three (0.6, 1.1, 2.3 m), St 3 with five (0.2, 0.7, 1.2, 1.7, 1.8 m), St 4 with five (0.3, 1.3, 1.8, 2.3, 2.4 m) and St 6 with five (0.15, 0.7, 1.15, 1.7, 2.15 m). Each station was also equipped with a pressure transducer at 1.5 m depth for continuous measurement. Water pressure was measured every 30 minutes and averaged and recorded every 6 hours. The sensors were left in the talik (perennially unfrozen) layer through the winter (31 August, 2012 to 28 August, 2014), so that during summer they measured the pressure due to the height of water above them, while during winter they measured the pressure in the talik below the freezing or frozen active layer. Measurements were corrected for atmospheric pressure recorded by a separate barometric sensor housed at a meteorological station approximately 500 m to the north. Ground temperature monitoring began on 26 March 2013 with measurements taken every minute and averaged and recorded every hour. The thermistors cables were attached to 0.05 x 3 m PVC pipe and driven into predrilled boreholes. The bottom of each pipe was sealed and the pipe was perforated throughout its length. Transducers were placed into the same pipe below the expected maximum depth of frost penetration. Weekly average air and ground temperatures were computed from the higher frequency measurements to minimize the effects of diurnal temperature variations and meteorological events. The frost table depth (*i.e.* thaw depth) along the seismic line was measured during the 2013 field season (28 May – 25 July).

GPR survey

The presence of a talik below the seismic line was investigated during the preliminary site investigations in August 2012, and found to be present and ranging in thickness between 0.5 m and 1.4 m according to temperature data. The presence of the talik aids in delineating the upper boundary of permafrost boundary from Ground Penetrating Radar (GPR) surveys given the high contrast in dielectric properties between frozen and unfrozen water. GPR surveys were conducted on 27 January, 2013 and 23 March, 2013. At these times, the active layer was frozen, and as such provided very little attenuation of the electromagnetic pulse. The MALA GPR system with 100 MHz unshielded antenna was used for this purpose. The approximate depth range for this antenna is about 2-15 m and the lower limit of the object target size is 0.1 - 1.0 m. The transmitter and receiver of the antenna were mounted in an in-line configuration and have a fixed distance of one meter. The temperature during the surveys ranged between -10^{o} and -25^{o} C. A GPS unit was used to track the location of the survey transect in January so that it could be repeated for the March survey. Both the transmitter and receiver were towed over the snow surface along the 195 m distance between St 1 and St 6 at an average speed of 2 km/hour. GPR

Depth validation.

Depth of signal penetration in the wetlands environment is the one of the main concerns for the GPR survey. The GPR results presented after processing the data are not directly linked to the depth, but to the time period when the signal travelling to the object of interest and back. This time called two-way travel time (TWT). If the wave velocity is known the conversion of TWT into deth can by expressed by simple formula:

$$d = v \times t/2 \tag{1}$$

Where d – depth in m, v wave propagation velocity (nm/c) and t is TWT in ns.

Propagation velocity depends on the relative dielectric constant εr and can be determined as:

$$v = c/\sqrt{\varepsilon r} \tag{2}$$

Where c is the speed of light and it equals 300mm/ns

Dielectric constant for the Scotty Creak peat was determined with TDR and was assumed as 45 at its maximum. Which gives us minimum wave velocity of 44.7 mm/ns? The maximum TWT determined by GPR was 300 ns or in depth equivalent 6.7 m.

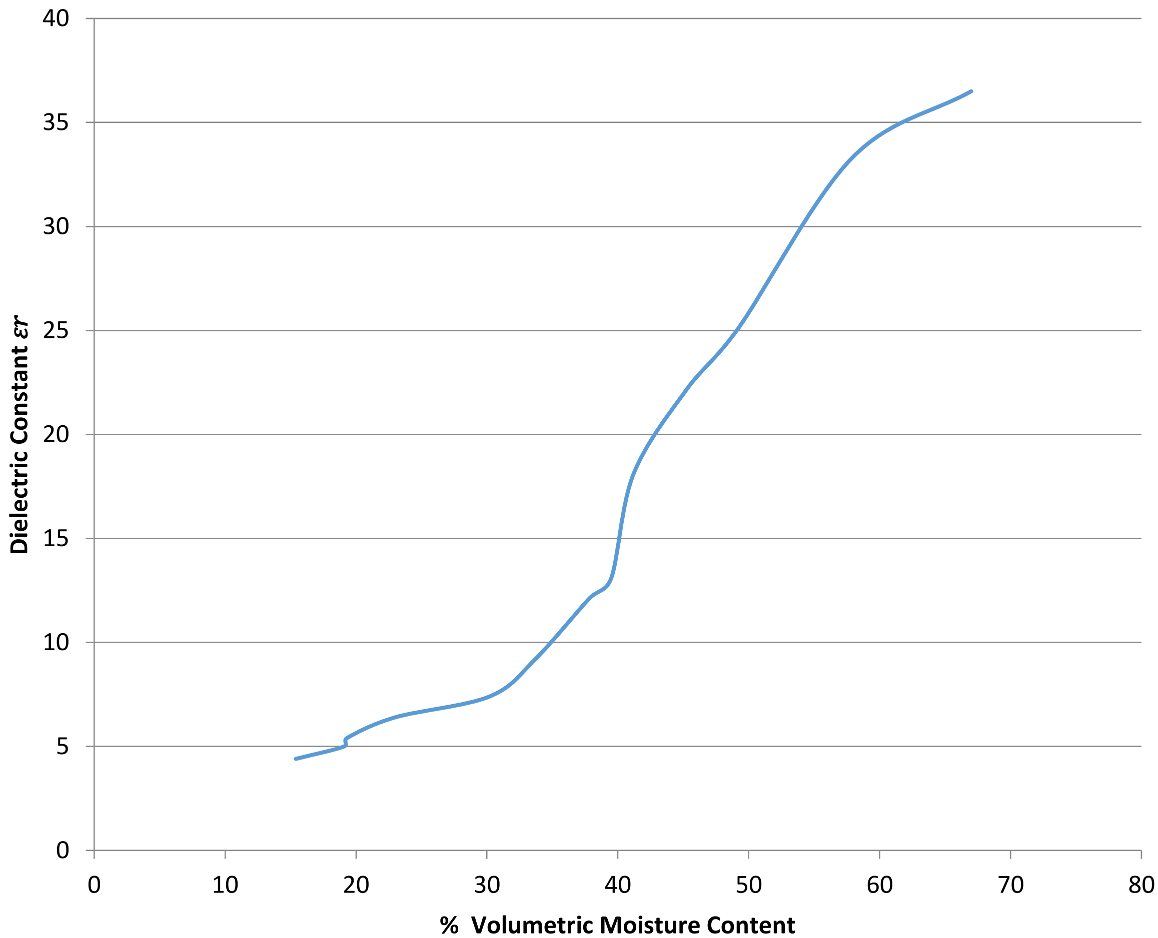

Figure 3: Relationship of dielectric constant ε_r to volumetric water content of Scott Creek peat. The maximum value of ε was used for estimation of rate of GPR wave propagation.

The permafrost at the depth within 3 m was correlated with hand auger drilling in the close proximity of monitoring stations 1, 2, 3 and 4. The depth of mineral soil in the open bog area was assumed as more than 7 m where hand auger drilling at monitoring station 6 has not detect mineral soil at this depth. These correlation points were taken as a base for GPR results interpretation, which gave us most likely overestimated depth of 8m to the mineral soil at western part of the transect.

The raw data produced during the GPR surveys were analysed using REFLEXW software of Sandmeier geophysical research with the Dewow filter to eliminate low frequency signals. The band pass frequency filter was used to remove the frequencies which lay outside the GPR bandwidth. The static correction filter was used to compensate a possible time delay in the snow layer, shifting up all traces towards smaller times. The background removal filter was also so that background noise arising from ringing in the antenna could be removed. The signal from the depth range 5 - 15 m was amplified by 10 to 30 times.

Results and Discussion

3-D reconstruction of the seismic line ground surface

A LiDar derived DEM was collected and processed for Scotty Creek by Applied Geomatics Research Group, NS, Canada. The LiDar data were used to generate a digital elevation model (DEM) of the seismic line and surrounding terrain (Figure 4). The DEM shows a distinct narrowing of the line with distance into the plateau from both the bog and fen edges to the point on the line mid-way between the two wetlands, suggesting thermal erosion from both the fen and bog sides. The DEM also reveals a steeper slope on its south side of the line throughout its 90 m length over the peat plateau, as described above for St 2.

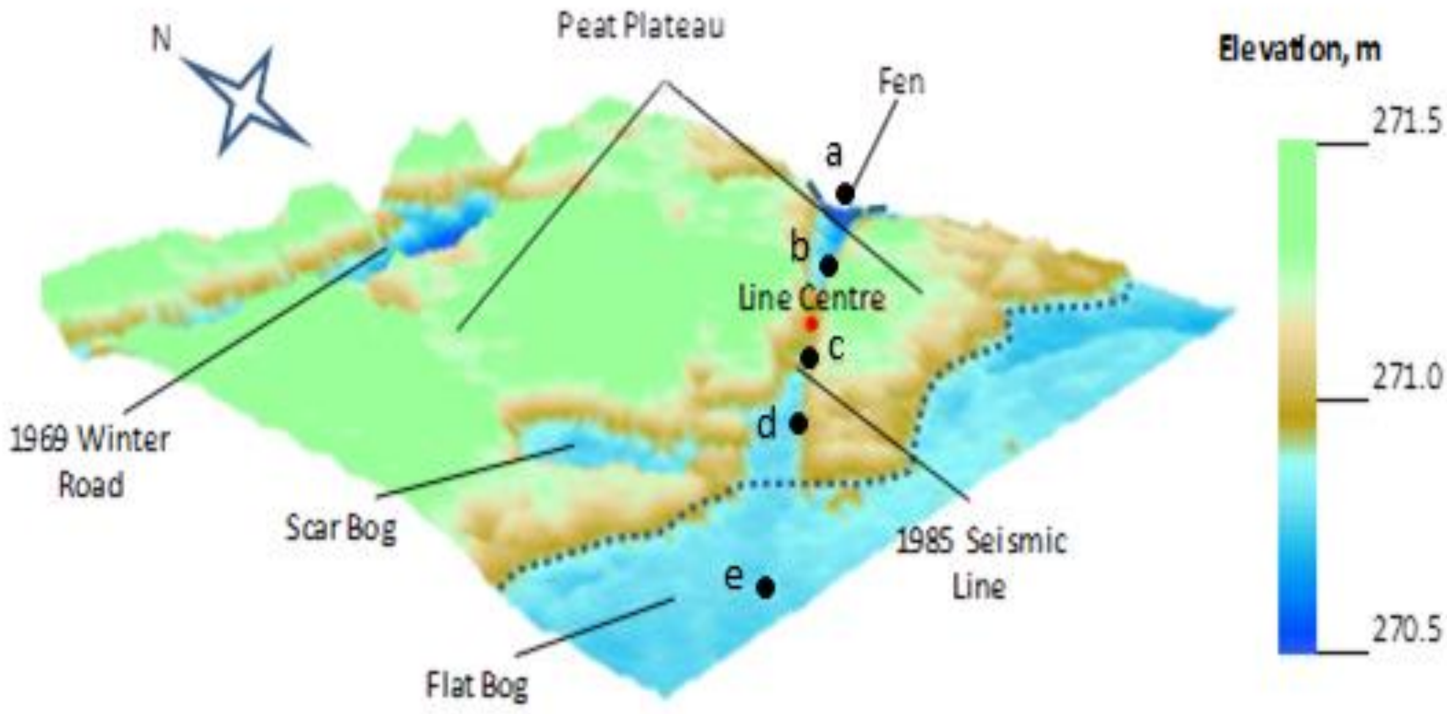

Figure 4: LiDar-derived digital elevation model of the seismic line and adjacent bog and channel fen. Black dots represent locations of measurement stations. a) St 1, b) St 2, c) St 3, d) St 4, e) St 6

Permafrost characteristics below the seismic line

The permafrost body under the peat plateau was readily identified by the intensity of reflections (Figure 5).

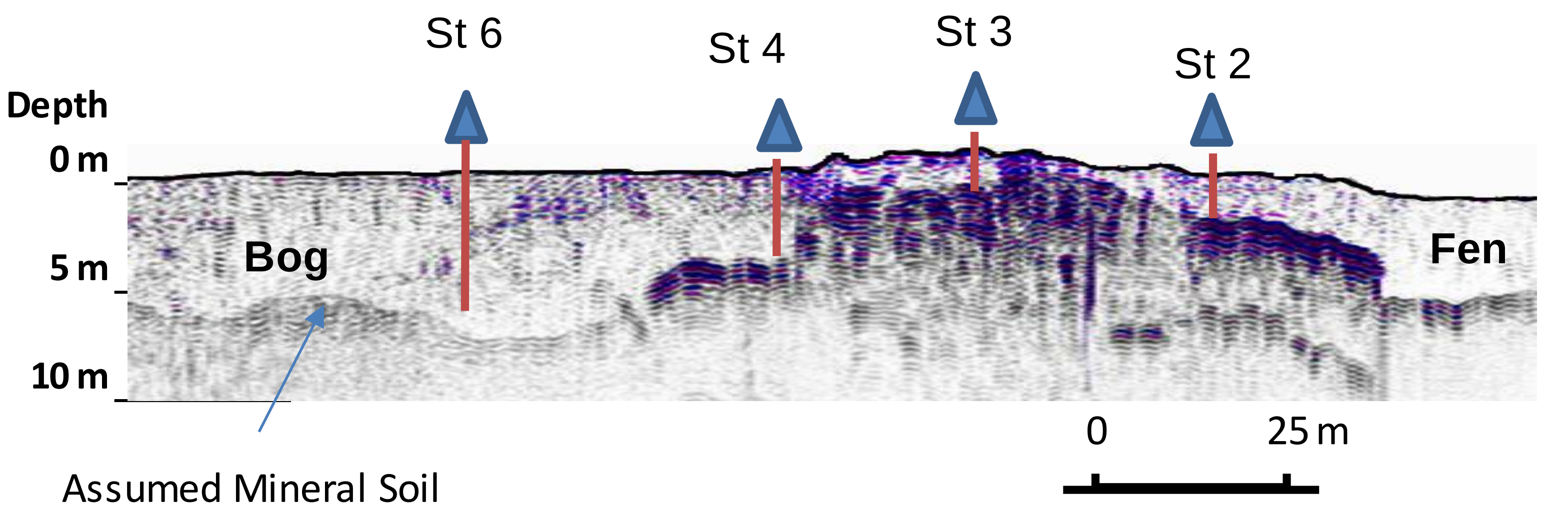

Figure 5: GPR transect between bog and channel fen. The solid vertical lines indicate verification boreholes. Darker area under station 2, 3 and 4 – permafrost. Darker area under station 6 assumed top of mineral soil, not reached with verification borehole.

Four layers with contrasting dielectric properties were distinguishable from the change in dielectric properties at the following boundaries: snow - active layer, active layer - talik, talik - permafrost, and peat - mineral sediment. The bottom of the permafrost however could not be detected due to the limitations of the sensors.

As indicated by the DEM, the geophysical data also shows that permafrost degradation was most pronounced near the plateau's boundaries with the fen and bog, where the permafrost table was about 1 m deeper, and the permafrost thickness (2-4 m) less than half the permafrost thickness mid-way between the two wetlands (about 9 m). The top of the mineral sediment is between 7 and 8 m below the ground surface in the permafrost free parts of the seismic line.. The topography of the permafrost table roughly approximates that of the ground surface. This similarity makes it possible to deduce the areas along the line where permafrost is most degraded or even absent based on the surface relief.

Water flow through the seismic line

Given the relatively high topographic position of the bog, the general flow direction is from the bog, through the seismic line, and into the channel fen. This mean flow direction was verified by the relative pressures recorded at the five stations, which on average for the 2012-2014 period indicated flow toward the fen where pressure was lowest (Figure 6a). This flux direction persisted through the winter months although during this time, flow occurs only through the talik that separates the frozen active layer above and the permafrost below (Figure 6b and 6c). However, there were deviations from this mean flow direction depicted in Figure 6a for short (*i.e.* seasonal) periods. For example, during the summers of 2013 (Figure 6d) and 2014 (Figure

6e), subsurface flow diverged from the middle of the seismic line toward both the bog and the fen. The only time when water entered the seismic line from the fen was during the 2013 spring freshet when the high water pressure in the channel fen temporarily reversed the hydraulic gradient such that flow, including overland flow, was driven into the seismic line.

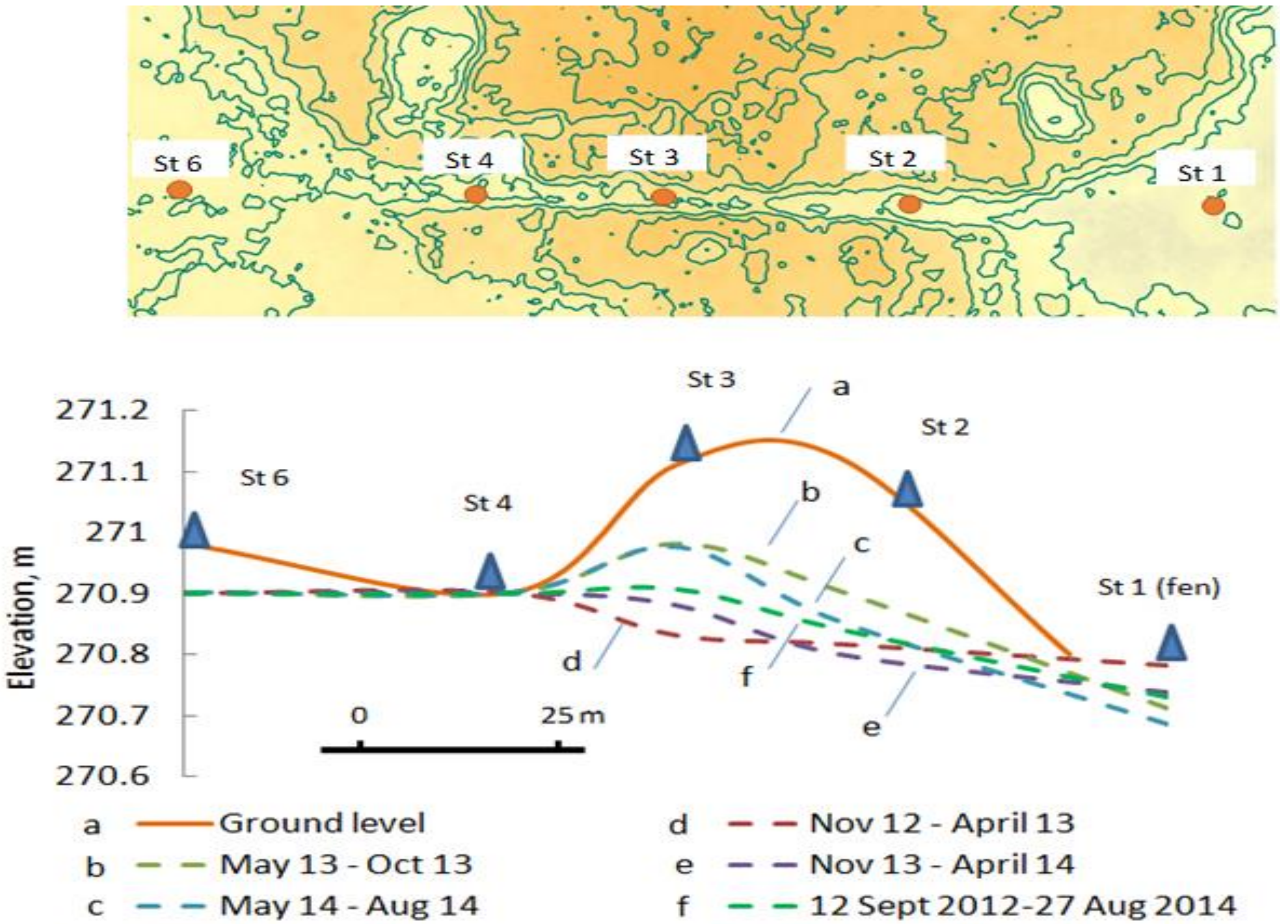

Figure 6: The arithmetic mean hydraulic head (m) recorded by the transducer at each station (dashed lines) with respect to ground surface elevation (solid line) for the following periods: a) 12 Sept 2012 – 27 Aug 2014; b) 6 Nov 2012 – 30 Apr 2013; c) 5 Nov 2013 – 29 Apr 2014; d) 7 May 2013 – 29 Oct 2013; e) 6 May 2014 – 27 Aug 2014 .

The minimum, mean and maximum annual soil temperatures for selected sites indicate a relatively thin (< 60 cm) active layer in the bog (Figure 7a), and at St 3(Figure 7b), fen (Figure 7d), the monitoring site on the plateau portion of the seismic line closest to the fen. By contrast, the maximum active layer thickness at St 4 fen (Figure 7c), (the plateau station closest to the bog) is 2.4 m.

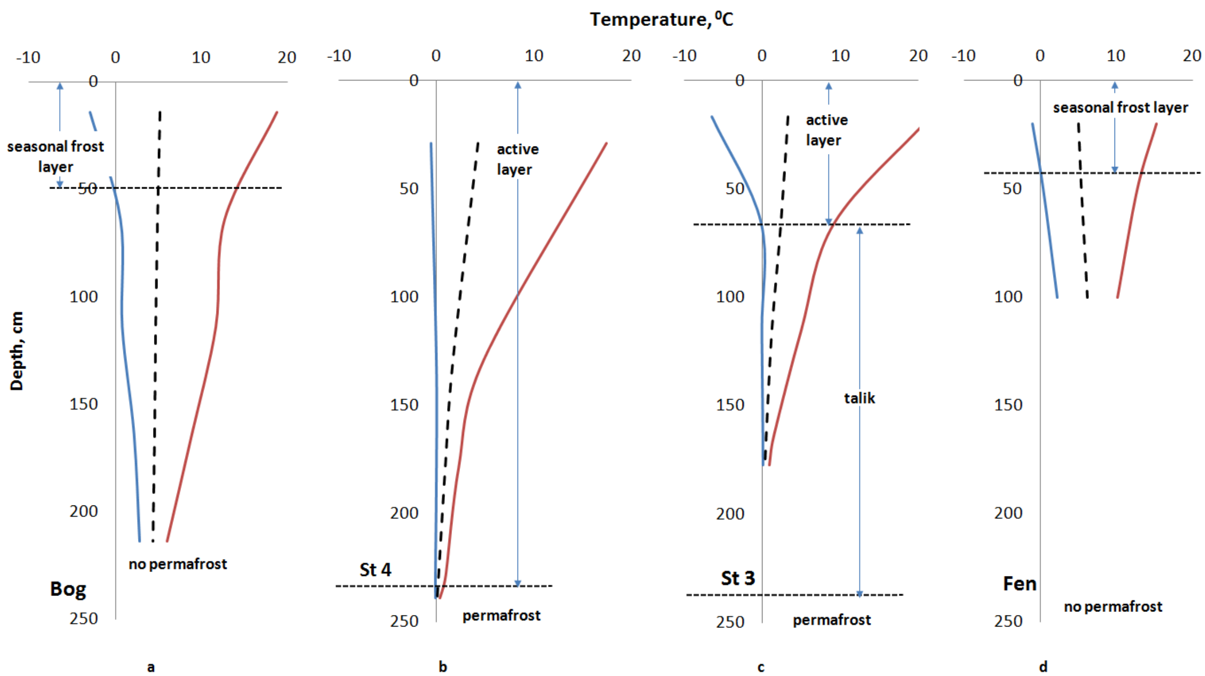

Figure 7: Minimum, maximum and mean annual soil temperature variations for the period 17 July, 2013 to 17 July, 2014 at the a) bog (St 6), b) St 4, c) St 3 and d) fen (St 1) stations. (St 2 not presented due to failure of logger).

As noted above, the talik layer enables the seismic line to continue conducting subsurface flow throughout the winter. During the summer months, the cross-sectional area for subsurface flow through the line is the product of the line's width (*i.e.* the distance between the permafrost "walls" on either side of the line), and the depth of the flow zone, *i.e.* the depth from the water table to the impermeable frost table, or in the case of full active layer development, the depth to the permafrost. Both the frost table and permafrost table vary with distance along the seismic line as depicted in Figure 4. The flux of the subsurface flow along the line can be described by Darcy's law:

$$q = K(\frac{dh}{dx}) \quad [1]$$

where q is Darcy flux [VT^{-1}] , K is the saturated hydraulic conductivity of the peat in the horizontal direction [LT^{-1}], and dh/dx is the dimensionless hydraulic gradient along the line. Field and laboratory measurements indicate that the hydraulic conductivity of peat at Scotty Creek and other similar environments, is uniformly high (200-300 m d^{-1}) in the upper ~20 cm of active layer, uniformly low (~1 m d^{-1}) in depths greater than ~40 cm (Quinton *et al.*, 2008). Based on these findings Wright *et al.*, 2008 proposed the following function, which represents changes of peat hydraulic conductivity K with depth z [L] :

$$\log K_{sat}(z) = 0.15 + 2.41/[1 + (\frac{z}{0.15})^{4.3}] \quad [2]$$

The average depth of the water table in May - Oct 2013 and May - Aug 2014 was 0.06 m between bog and St 3 and 0.11 m between St 3 and fen. The geometric mean hydraulic conductivity for the saturated zone of the active layer between St 3 and the St 6 (bog) was computed as 1.8 m/day and between St 3 and the St 1 (fen) as 1.6 m/day. Using these values of K, and equation (1) the flow rate through cross-sectional area of the 1 m^2 of the seismic line from St 3 and St 1 (fen) and between St 3 and St 6 (bog) was calculated (Figure 8).

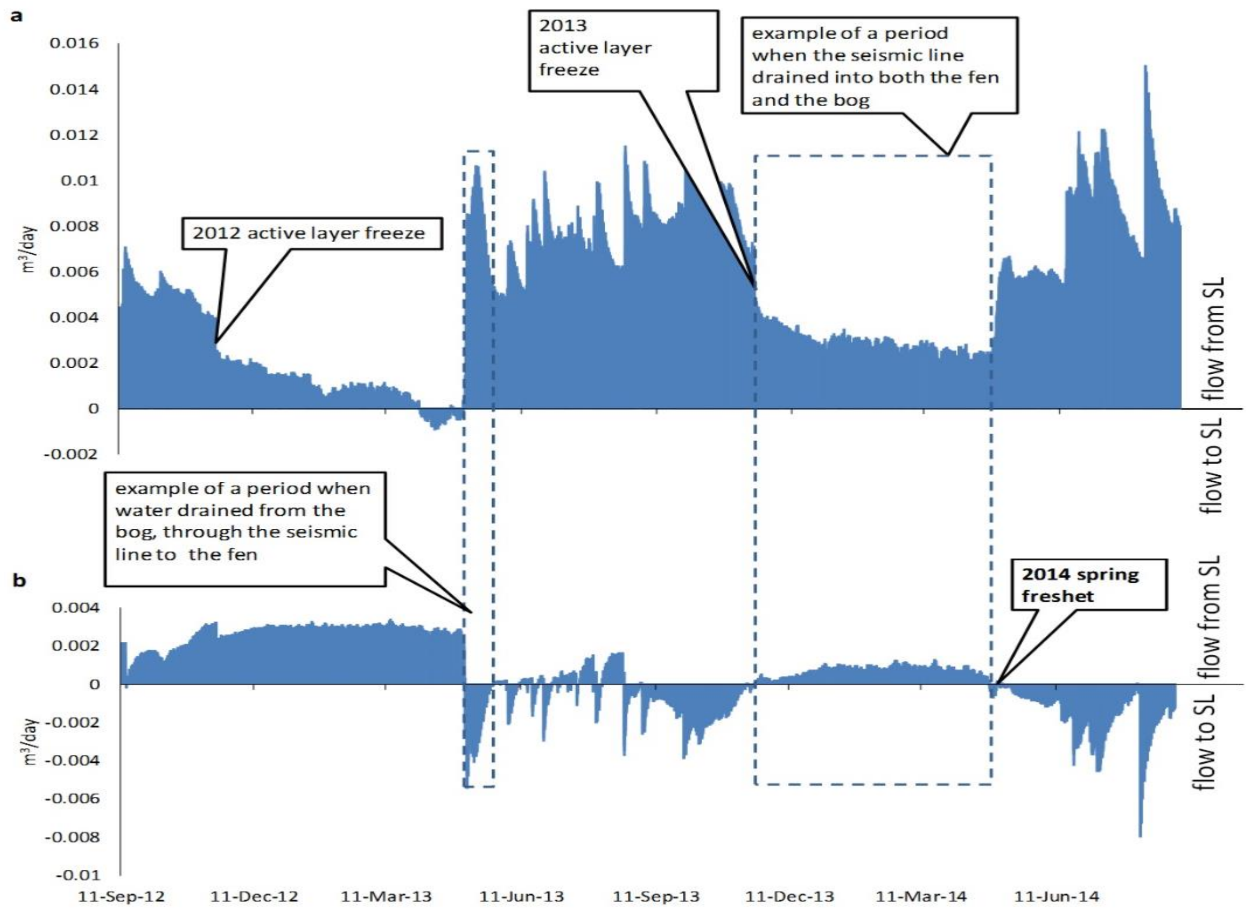

Figure 8: Subsurface flow along the seismic line for the sections between a) St 3 and the fen (St 1), and b) St 3 and the bog (St 6). Positive values indicate water flow from the seismic line (SL), while negative values indicate flow towards the seismic line.

During winter the bog drains into the fen via the talik below the seismic line, while in summer the flow direction on the bog side of St 3, alternates in and out of the seismic line (e.g. see Figure 5d), thereby interrupting bog drainage toward the fen,. The high flow from bog through seismic line towards fen in spring time is attributed to spring freshet, when overland flow may occur. The standard deviations in Figure 5 are relatively high during summer owing to water table fluctuations in response to precipitation, evaporation and drainage, while during winter, the frozen active layer shields the underlying talik from these effects.

Thermal properties and energy flows

The surface energy balance is controlled by the relative magnitudes of incoming and outgoing shortwave and longwave radiation, non-radiative (*i.e*. conduction and convection) energy transport processes, turbulent heat exchanges, and by other sinks and sources of energy at the ground surface, such as phase changes. According to Burgess (1998), ground surface and subsurface temperatures changes can be related directly to temperature changes of the overlying air. We estimated that in Scotty Creek warm permafrost conditions the effect of thermal wave propagation can be observed down to a depth of approximately 10 m.
Surface temperature measured within the period of September 2014 and September 2015 presented at Fig. 9

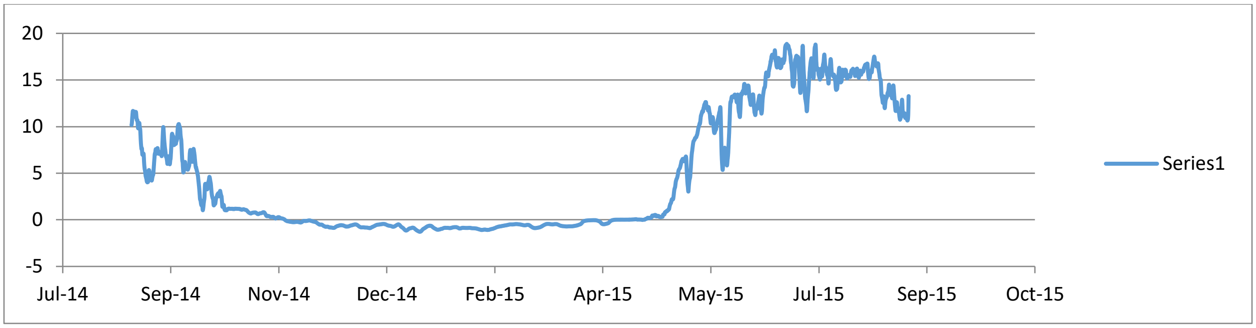

Fig. 9: Surface temperature measured at undisturbed elevated peat plateau in close proximity to the seismic line. With an amplitude of the temperature wave of ~ 10°C.

The amplitude of the annual temperature variations decreases exponentially with depth increase and can be calculated as

$$A_z = A_s \exp(-z\left(\frac{\pi}{\alpha P}\right)^{\frac{1}{2}}) \tag{3}$$

Where A_z is the amplitude of the temperature wave at depth z, A_s is the amplitude of the surface temperature wave, α is thermal diffusivity, and P is the period of the wave – one year. With the assumption of thermal diffusivity α= 5E–6 m2/s the attenuation of the temperature wave is shown in Figure 10.

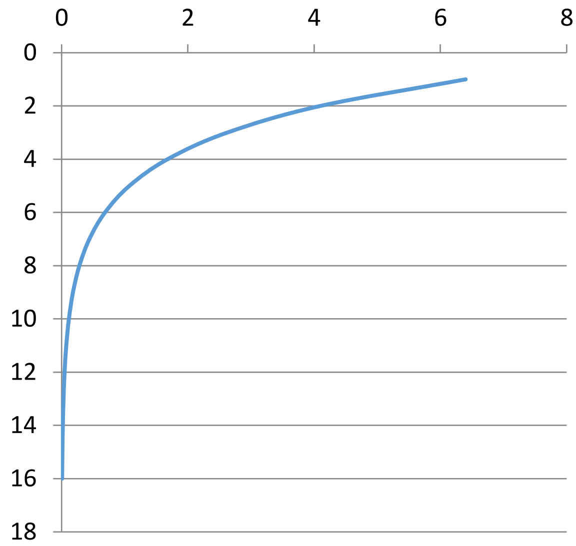

Figure 10: Theoretical attenuation of the temperature wave, generalized for Scotty Creek conditions.

This idealized temperature distribution in Figure 10 is not taking into account the presence of about 10 m of "warm" permafrost in the state of phase change at a temperature of zero point depression. The heat which enters permafrost from above is consumed as latent heat preventing temperature raise below the permafrost body. The effect of latent heat as a filter was described by Riseborough 1989. However, subsurface temperatures also depend on other variables such as precipitation and vegetation. On a daily basis, the magnitude of these other variables can be so large that it can mask the influence of the air temperature. While the active layer is freezing or frozen (October-March) the presence of snow or release of the latent heat during the soil freezing may effectively decouple the active layer from the atmosphere in terms of their temperatures. The thermal properties of a material is described in terms of thermal conductivity (1), specific heat (2) and thermal diffusivity (3).

$$k = \frac{\mathrm{QL}}{\mathrm{A}\Delta T} \tag{4}$$

k is thermal conductivity in W/m K, Q is amount of heat transfer through the material in W, A is the area of the body in m2, ΔT is difference in temperature in K.

$$c_p = \frac{\mathrm{Q}}{\mathrm{m}\Delta T} \tag{5}$$

c_p is specific heat in J/kg K, Q is heat added in J m is mass in kg, ΔT is difference in temperature in K.

$$\alpha = \frac{\mathrm{k}}{\rho c_p} \tag{6}$$

k is thermal conductivity in W/m K, c_p is specific heat in J/kg K, ρ is density in kg/m3

Thermal conductivity is defined as the heat flow through a unit of thickness under the temperature gradient, the specific heat is the amount of heat required to change one unit of mass or volume by one degree C, and the thermal diffusivity is the ratio of the two per unit volume or mass. In other words, the diffusivity indicates how quickly a material responds to heat added or removed from it. The upper layer of peat in Scotty Creek has porosities in excess of 0.8 (Quinton and Baltzer 2013), and consequently the bulk thermal properties of the active layer are strongly influenced by variations in moisture content and changes in phase. With the exception of the upper 5-10 cm, the active layer of the seismic line is saturated and unfrozen throughout its depth during the June - October period. It was for this steady thermal condition that the thermal properties of the active layer were evaluated by Hayashi *et al.* (2007). The amplitude of the temperature wave propagated downward into the active layer diminishes with increasing depth, and with depth its arrival time is increasingly delayed. In unfrozen soil, the amplitude and lag time are directly related to the thermal diffusivity by the following (Woo, 2012):

$$TA_z = TA_{z-1}\exp[-z\sqrt{\pi/LD_T}], \text{ and} \quad (7)$$

$$t_{lag} = 0.5z\left[\frac{L}{\pi D_T}\right], \quad (8)$$

where TA_{z-1} *and* TA_z represent a temperature range at depth z-1 and z respectively, D_T is the thermal diffusivity in the range of depths z-1 and z, L is the time period and t_{lag} is lag time of the temperature wave. Thermal diffusivities for the peat along the seismic line were computed using both the range (7) and lag (8) methods (Table 1).

Station	Range method	Lag method	Depth
St 1 Fen	1.1	3.2	0.5-1.0 m
St 3	0.8	0.4	0.7-1.8 m
St 4	1.2	0.7	1.3-2.4 m
St 6 (bog)	2.2	0.4	0.6-2.1 m

Table 1: The average depth-integrated values of thermal diffusivity (10^{-7} m^2 s^{-1}) computed from the range (equation 4) and lag (equation 5) methods.

The greatest difference between estimates of D_T derived from the two methods occurred at St 1 (fen) and at St 6 (bog), both outside of the plateau portion of the seismic line. The geothermal heat gained from the lower and warmer peat layers in the bog (Figure 11) resulted in relatively high diffusivity values computed from the range method for St 6. Flowing water along the fen at St 1 dampened the downward heat propagation, and therefore reduced the range method estimates of D_T.

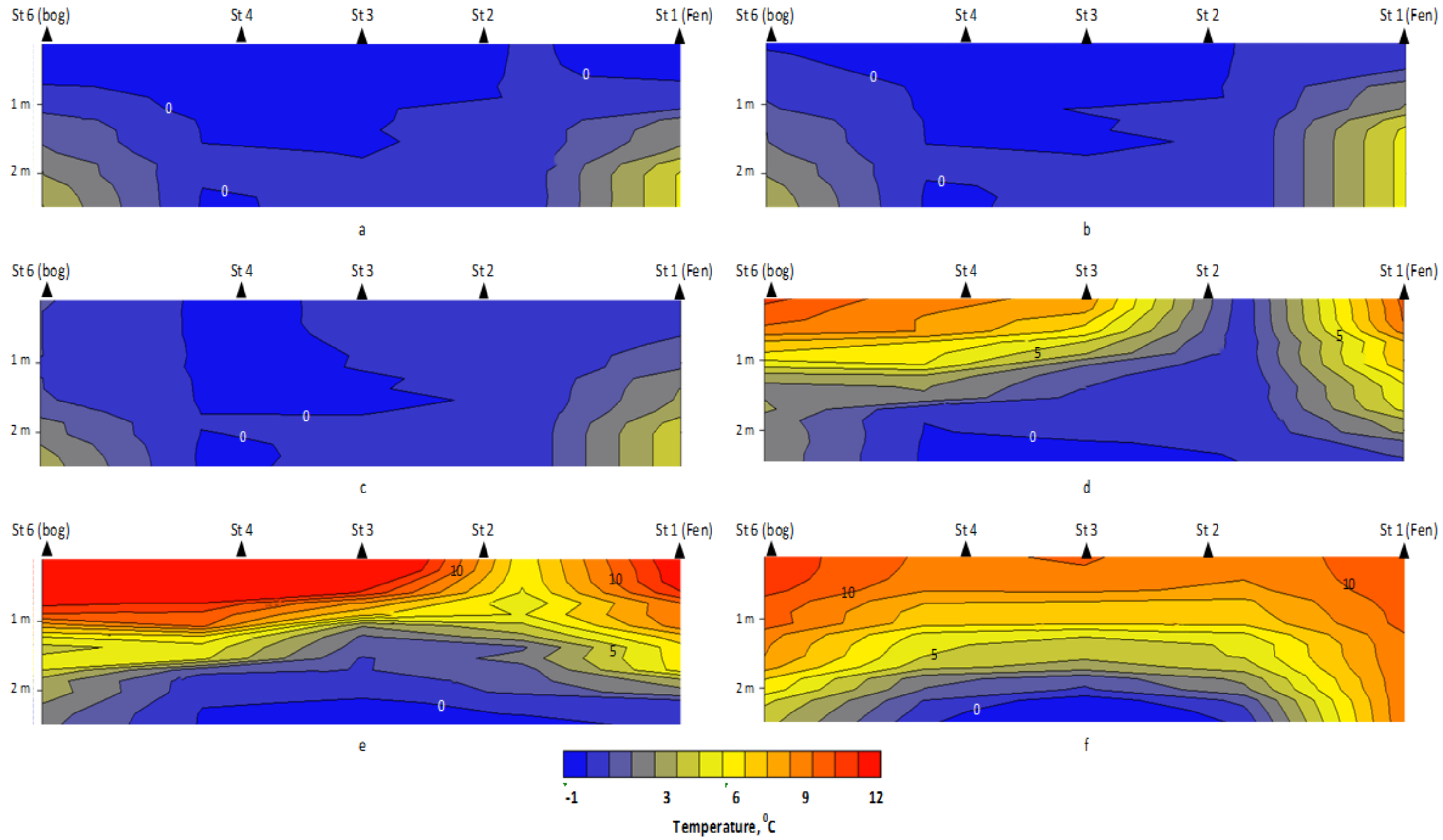

Figure 11: Variations in ground temperature along the seismic line between the fen and bog for six times: a) March 22, 2013; b) 13 April 2013; c) 18 May 2013; d) 15 June 2013; e) 6 July 2013; f) 14 September 2013.

Active layer thaw in the seismic line

An analysis of the rate and pattern of active layer thaw in the seismic line for the period May 28 to July 25, 2013, shows a distinct bi-directional heat flux towards St 3 near the mid-point of the seismic line (Figure 9). Being free of permafrost, the fen and bog are heat sources during winter. After the spring freshet, the ground temperature started rising first in the fen (St 1) and then in the bog (St 6). This was followed by the warming of the stations on the seismic line closest to these wetlands. The station at greatest distance from the wetlands (St 3), warmed and thawed last (Figure 12).

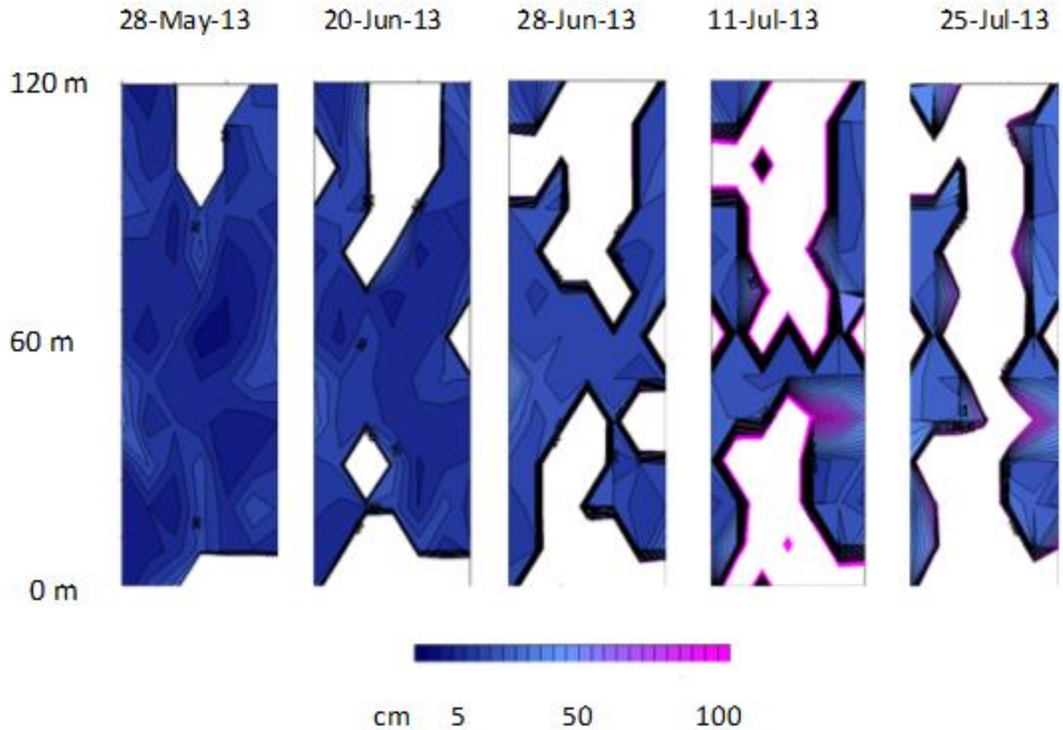

Figure 12: Variation in the degree of seasonal frost thaw along the seismic line over the peat plateau between the plateau-fen edge (0 m) and the plateau-bog edge (120 m). White areas indicate absence of seasonal frost.

The initial temperature rise was most pronounced in the fen (St 1) and on the fen-half of the seismic line. Although the bog temperatures rose later than fen temperatures, they reached higher temperatures by mid-summer, and as a result, the active layer thawed earlier in the summer in the bog than in the fen. Comparative analyses of micrometeorological data from bogs and fens at Scotty Creek (Wright *et al.*, 2008) indicated that the relatively low albedo and absence of ventilation by flowing water resulted in higher ground surface temperatures on bogs than fens.

Conceptual Model

The insights gained from this study have given rise to a new conceptual model describing coupled thermo-hydrological processes within seismic lines in the high-boreal zone of discontinuous permafrost. Linear seismic disturbances in this region appear to initiate a thermal and hydrological response with three distinct and consecutive stages.

Stage 1: Thaw driven by internal processes: The first stage is initiated by the over-winter removal of the tree canopy over the line. In the following spring, ground surface depressions resulting from the heavy machinery used to cut the line, serve as focal points for permafrost degradation in subsequent years. During periods of excess moisture (*e.g.* end-of-winter snowmelt), water accumulates in these depressions so that they become areas of preferentially high soil moisture, which may include ponding. Since wet soils are better thermal conductors than dry soils (Hayashi *et al.*, 2007), preferential wetting of the depressions results in active layer thickening and therefore permafrost thaw below the depressions. This results in the subsidence of the ground surface, and therefore a further deepening of the depression, which increases the local hydraulic gradient, enabling the depression to draw water from further afield. As such, a moisture-thaw feedback is initiated that increases not just the depth of depression but also its width, a process causing individual depressions to coalesce along the line. Over time, preferential thaw and coalescence transform the seismic line into a linear depression that often takes on the appearance of a broad (~10 m wide) ditch through the forest. The processes described above modify the seismic line so that it can conduct water along it length by overland flow and subsurface flow through the active layer.

Stage 2: Development of continuous flow connection: Thaw along the line in the first stage was driven largely by internal processes within the portion of the seismic line traversing the peat plateau. The second stage is initiated once permafrost thaw has lowered the permafrost table below the elevation of the water table of the adjacent fen and/or bog. Once this occurs, the water in the bog and fen is no longer impounded by the intervening raised permafrost, and as a result, water and energy can flow from these adjacent wetlands onto the seismic line. Early in the second stage, water incursions over the ground surface and through the active layer from the bog and fen are separated as depicted in Figure 3, since in the middle of the fen (*i.e.* St 3), the ground surface and the underlying permafrost are elevated above the fen and bog water tables. However, with further permafrost thaw and ground surface subsidence, a continuous hydrological connection is formed throughout the length of the seismic line via the active layer, and with continued thaw, via a talik as well. The latter forms a year-round connection for water and energy flows between the bog and fen. In the extreme case, Stage 2 would lead to the complete removal of permafrost throughout the line. This stage produces a subsurface flowpath through the talik that supplements the overland flow and active layer flowpaths developed in stage 1.

Stage 3: Differential permafrost thaw and recovery. Stage 3 is initiated once the seismic line shows indications of permafrost recovery. Whether or not this stage is reached depends upon the regional climate, and other, local factors affecting the heat flux into the ground (*e.g*. snow accumulation regime, density and height of tree canopy adjacent to the line, orientation of the seismic line, and proximity of the line to wetland water sources). The seismic line in the present study is oriented West-East, and as such, its north side (*i.e.* south-facing) has relatively high insolation. Williams *et al*. (2013) found that this results in a dryer ground surface on the north side of the line, enabling greater regeneration of vegetation, including black spruce. As a result, the ground below the north side is better insulated than on the south side, where water accumulates preferentially, and surfaces are wet and therefore more thermally conductive. Thermal erosion is therefore greater on the south side, where the edge of the seismic line forms a steep embankment which continues to thaw as evidenced by the leaning of trees over the south side of the line. Permafrost regeneration on the north side and permafrost loss on the south side, gives the appearance on aerial imagery that the line is slowly migrating southward. In addition to permafrost regeneration on one side of the line as described here, Zoltai and Tarnocai (1974) found that isolated patches of permafrost can develop throughout seismic lines where permitted by local topographic and drainage conditions. The flow through the talik on the northern side of the line became much less significant than on the southern side due to permafrost regeneration. Preferential permafrost thaw on the south side of the seismic line may also, in part, be driven by the preferential flow through the active layer and talik along that side of the line.

Summary and Conclusions

We characterized thaw and flow processes along a studied seismic line in the southern fringe of permafrost, in northwestern Canada. Ground surface characteristic, including widening of the line near the plateau-wetland boundaries, and variations in topography and vegetation across the ~10 m width of the line were explained by preferential permafrost thaw and the resulting ground surface subsidence. Variations in subsurface properties, such as ground temperatures, ground thaw rates, depth to permafrost, and presence and thickness of a talik, were also explained in these terms, and in relation to the flow of water through the seismic line. Overland flow accounted for a very small fraction (<1%) of the total flow through the line over the study period. It was found that the seismic line is a highly dynamic pathway for subsurface flow in terms of both rate and direction. At some times of the year, it conveyed water from the bog into the channel fen on the other side of the intervening plateau, while at other times, it drained into both the bog and the fen. For the two-year study period, water was conveyed from fen to bog only once and for a relatively brief period during the 2013 freshet. Overland flow and flow through the active layer are reasonably well understood processes in this environment, and these processes are transferable to seismic lines. This study identified the perennially unfrozen (*i.e.* talik) layer and demonstrated its importance as a subsurface flowpath. This is an important finding, since it was also shown that while flow over the surface and through the active layer stops during the winter months, flow through the talik continues throughout the year, and at an appreciable rate — for the annual period starting September 2012, approximately 50% of the total flow through the seismic line was conveyed through the talik. This study has important implications to how we view the potential hydrological impacts of seismic lines in this region.

For example, it was shown that seismic lines disrupt the characteristic storage function of bogs by allowing them to drain into fens and onward to basin outlets. Seismic lines not only have the potential to increase the runoff contributing area and therefore the amount of runoff produced by basins, but by cutting indiscriminately across bogs, fens and plateaus, they also have the potential to short-circuit the natural drainage network, thereby reducing runoff transit times. Over longer periods, these effects have the potential to alter the water of basins, especially where the density of seismic lines (*i.e.* the cumulative length dived by the basin area) is high. At Scotty Creek, the density of seismic lines is approximately seven times larger than the natural drainage density. Further research is needed to investigating these potential hydrological impacts of seismic lines. The new conceptual model describing coupled thermo-hydrological processes within seismic lines provides a foundation for such research.

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Permafrost degradation under a linear disturbance in discontinuous permafrost

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ABSTRACT

The discontinuous permafrost below subarctic peat plateaus is highly sensitive to ground surface disturbances. Seismic lines are the most widely occurring form of linear disturbance in the Northwest Territories, Canada, yet their hydrological impact is poorly understood. Ground temperature and hydraulic head observations, geophysical surveys, and model simulations indicate that permafrost thaw under seismic lines in this region is irreversible and ultimately produces a permafrost-free corridor for subsurface drainage. It has also been found that once the permafrost table descends below the elevation of the water table in the wetlands on either side of the permafrost body, the seismic line provides a year-round subsurface hydrological connection between the two wetlands. During winter, subsurface flow is confined to a talik (*i.e.* perennially-thawed) layer. It has been estimated that winter subsurface flow is about 35% of total annual water flow through the examined seismic line. This study also indicates that ice loss from within thawing and disintegrating permafrost enables water seepage through thawing permafrost bodies, a subsurface flowpath not yet reported in literature.

Key words: permafrost, linear disturbance, subsurface flow, permafrost degradation

Introduction

Permafrost, ground that remains at or below 0° C for at least two consecutive years, covers between 40% and 50% of Canada's landmass (Robinson *et al*., 2003). The thickness of permafrost varies from 1000 m at high latitudes and altitudes (e.g. mountains of Baffin Island, Canada) to several meters in lower latitudes and altitudes (e.g. northern British Columbia, Canada), where the temperature of permafrost is close to 0^{o}C. Warming air temperatures in recent decades have led to increasing permafrost thaw and disappearance (Robinson and Moore, 2002), particularly in areas of discontinuous permafrost (Osterkamp and Romanovsky, 1999). The permafrost of northern peatlands is typically saturated with ice and a fraction (15-20%) of unfrozen water (Quinton and Baltzer, 2013). Peatlands containing permafrost, such as peat plateaus, are therefore far less permeable to water than adjacent permafrost-free terrains, such as channel fens and flat bogs. Since the permafrost table of peat plateaus rises to an elevation above the ground surface of adjacent permafrost-free wetlands, the plateaus impound the water of the latter, a process that favours wetland development and preservation (Zoltai and Tarnocai, 1975).

Peat plateaus rise 1-2 m above the surrounding wetlands. Since plateaus also have a relatively large snowmelt water supply and a limited capacity to store water, they shed water into adjacent wetlands and subsided linear disturbances (Williams *et al*., 2013) primarily as subsurface flow (Wright et al., 2008) through the thawed, saturated layer between the water table and the relatively impermeable frost table below. By contrast, bogs are primarily water storage features because they are surrounded by the raised permafrost that occupies an elevation higher than the water table in the bogs. Bogs are waterlogged through precipitation and water drainage from adjacent peat plateaus. They may recharge local groundwater systems (Gordon *et al*., 2016). However, recent studies by Connon *et al.* (2015) demonstrated that breaches in the permafrost impounding some bogs allow them to become hydrologically connected to one another as a cascade, and possibly to the basin drainage network (*i.e.* channel fens) during periods of high moisture supply. Channel fens convey surface and near surface flows along separate, broad, hydraulically rough channels, which collectively form the basin drainage network (Quinton *et al*. 2003). Water draining into the fens from bog cascades and directly from the slopes of plateaus at the plateau-fen margin, is slowly conveyed as overland flow or subsurface flow toward the basin outlet. While climate warming is the primary driver of permafrost thaw (Solomon *et al*., 2007), resource development can also initiate and/or accelerate this process (Raynolds et al., 2014). During the second half of the 20th century, when the oil and gas industry expanded north of the Arctic Circle (AECOM, 2009), numerous seismic lines were cut through the Boreal region using heavy machinery, a process that has marked the landscape with long (10s to 100s of km) and wide (5 to 15 m) trails that remain clearly visible. Seismic lines are the most widespread type of disturbances related to oil and gas exploration (Kemper and Macdonald, 2009). According to the Department of Environment and Natural Resources of the Northwest Territories (NWT), the Mackenzie River Delta has the highest seismic line density in the NWT of 6 km/km^2, while the southern part of the Taiga Plains ecological region, where the study site is located, has a line density of approximately 1.2 km/km^2 (GWNT, State of the Environment Report, 2015)**.** Seismic cut lines 10-15 m wide were common in the 1950s to 1980s, while 5-8 m wide cuts have been used in more recent years (Williams *et al*., 2013). Seismic surveys involve drilling shallow boreholes for explosive charges, exploding the charges in a specific sequence, and recording the seismic signal for processing and interpretation. The impact of seismic lines on soil and vegetation (Lee and Boutin, 2006), wildlife habitat (Kansas *et al*., 2015), and aquatic ecosystems (Wilson, 2002) is well documented, however their impact on hydrology and permafrost has received little study and is therefore poorly understood.

Seismic lines cut indiscriminately over all terrain types in their path. Since the major terrain types in the study region contrast sharply in terms of their dominant hydrological processes, the impact of a seismic line on local hydrology depends upon the local terrain type. Williams *et al.* (2012) suggested that hydrologically, seismic lines on peat plateaus functioned analogously to channel fens, draining water from adjacent plateau slopes and from upstream wetlands intersected by the line, and conveying this water along ephemeral surface channels to downstream ecosystems.

In the peatland-dominated southern Taiga Plains ecoregion, the discontinuous permafrost is isothermal at the ice nucleation temperature throughout the entire 10-15 m thickness. (Smith *et al.*, 2010) The upper surface of permafrost is thermally insulated by the active (i.e. seasonally thawed) layer. However, the heavy equipment, which was used in the seismic surveys compressed the active layer and therefore degraded its capacity to thermally-insulate the underlying permafrost (Williams et al., 2013).

Seismic lines in permafrost environments cause ground surface subsidence, altering local water flow and storage processes (Woo, 2012). Williams *et al*. (2013) examined ground surface flow and storage processes on seismic lines in the southern Taiga Plains ecoregion at Scotty Creek, NT and found that the cumulative daily overland flow volume from seismic lines during peak runoff periods was approximately three times larger than the total volume of water stored on the ground surfaces of the same seismic lines. In a subsequent study at Scotty Creek, Braverman and Quinton (2015) reported that the permafrost thickness below a 1985 seismic line was 15% to 25% less than the average off-line thickness of 10 m, although permafrost thaw was significantly greater below ground surface depressions. The authors also noted that such depressions impede the continuous overland flow along the seismic line, since continuous flow can occur only once depressions along the line have been filled in a manner consistent with the "*fill and spill*" process described by Spence and Woo (2003).

Williams *et al*. (2013) speculated that seismic lines convey surface runoff according to two seasonally characteristic hydrological regimes. A connected regime occurs during periods of high moisture supply, such as during the end of winter snowmelt event, and in response to large rain events. During this regime, water flows over the seismic line surface along a continuous ephemeral channel, discharging into a fen, bog, lake, or other terrain unit. A disconnected regime occurs when the moisture supply is insufficient to overcome the depression storage capacity along the line and as a result, the continuity of the ephemeral channel is lost, impeding the overland flow from the seismic line to downstream units. Over a period of decades, it is possible that the higher ground separating the depressions will subside. Such a process would diminish the threshold (*i.e.* "fill and spill") response of a seismic line. However, given the ephemeral nature of overland flow, the magnitude of the subsurface water flux is potentially far greater. Although to date, there has been no study of subsurface flow conveyance along seismic lines. This is a major gap in the literature on cold regions hydrology and water resources, given the very high prevalence of seismic lines in the North.

This study seeks to increase the understanding of and ability to predict the subsurface flux along seismic lines in the wetland-dominated southern margin of discontinuous permafrost. This will be accomplished through the following steps: 1) create a simple heat transfer model for organic soils saturated with ice and water; 2) evaluate the performance of the model with field observations; 3) use the validated model to investigate the impact of permafrost thaw on subsurface flow through the seismic line; and 4) estimate the subsurface mass flux from the seismic line as a result of permafrost thaw.

Study site

Scotty Creek is located approximately 60 km southeast of Fort Simpson, NT (Figure 1).

Figure1: Location of the study site in the zone of discontinuous permafrost, approximately 60 km SW of Fort Simpson, Northwest Territories, Canada. The 500 m stretch of seismic line between the channel fen and a 1943 military road, located about 400 m from Goose Lake and positioned perpendicular to the basin drainage, which is located towards the northwest from Goose Lake into First Lake.

The average annual air temperature in Fort Simpson is -2.8^o C (1964-2013), and the average July and January temperatures for the same period are 17.1^o C and -25.9^o C, respectively. The annual total precipitation is 388 mm, with approximately 39% falling as snow. Scotty Creek is located in the continental climate zone, and the zone of discontinuous permafrost. Approximately two-thirds of the Scotty Creek basin is continuous peatland, consisting of peat plateaus, flat bogs (*i.e.* collapse scars) and channel fens. Peat plateaus rise 1.0 to 1.5 m above the surrounding bogs and fens (Quinton *et al*., 2011). Permafrost is a distinctive feature of peat plateaus (Burgess and Smith, 2000) with thicknesses ranging between 10 and 20 m (Hayashi et al., 2007). The ~0.5 to 8 m thick peat deposits are underlain by sandy silt (Figure 2).

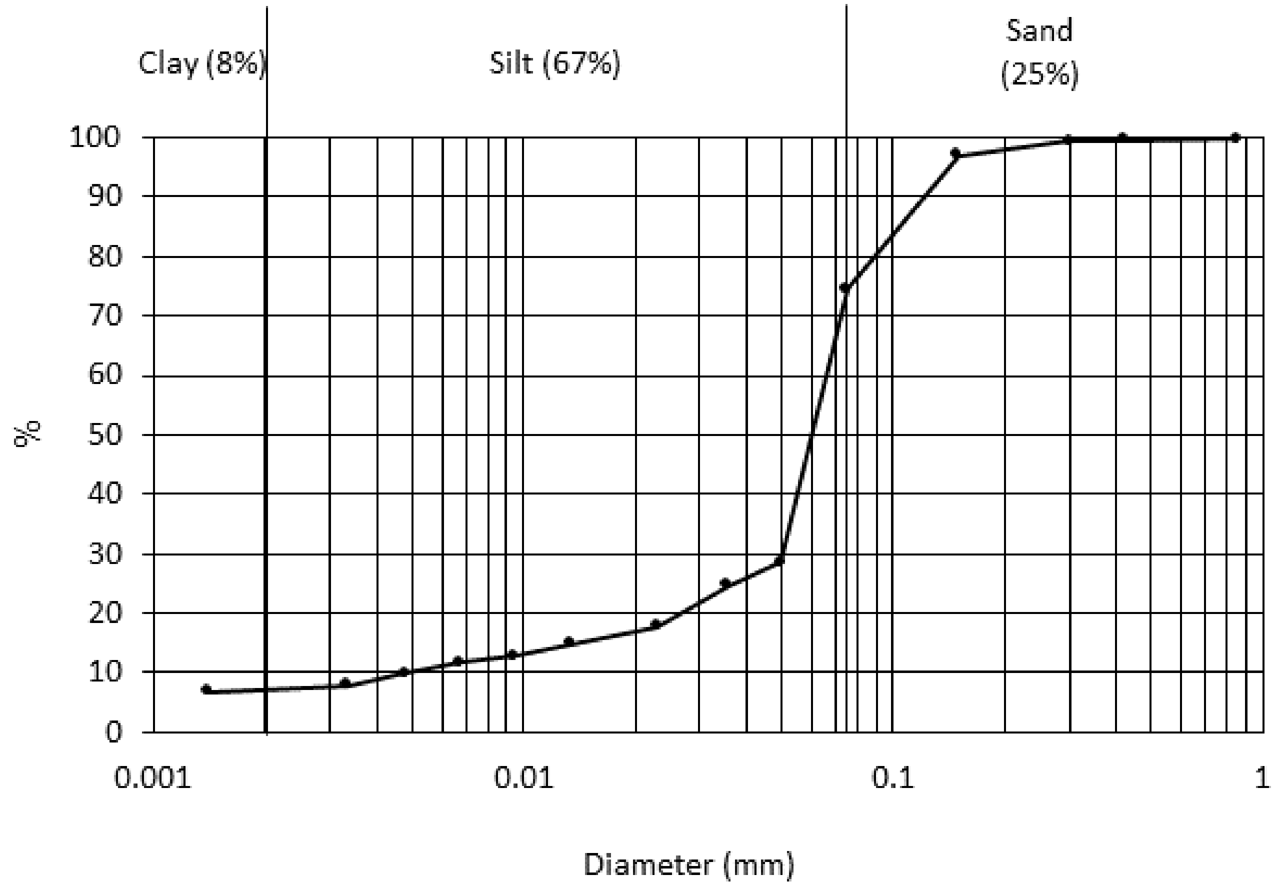


Figure 2: Grain size distribution of mineral soil underlying the peat layer. The silt size particles 2.0-75.0 μm represent the main components of soil: 67%, sand is 25% and clay (< 2.0 μm) is only 8% of the soil.

The seismic line examined in this study is located 0.4 km north of Goose Lake, and was cut in 1985. It is approximately 15 m wide and is oriented NE-SW (Figure 1). This study focusses on a 500 m length of this line, located between a channel fen and a former winter road (locally known as Fort Nelson Trail) not in use since 1970. Over the 500 m length, the seismic line crosses a bog connected to the channel-fen, two small isolated collapse scar bogs, and several peat plateaus. The line has a distinctive slope of approximately 5^{o} NE to SW direction, and is covered with sparse black spruce 0.5 m to 1.5 m tall, except for those sections of the line that cross the channel fen or bogs, or where the seismic line is covered by a floating mat of hydrophilic vegetation. A permafrost table is present throughout the seismic line, ranging in depth from 1.5 m to 2.5 m below the ground surface, with shallower depths being on the north side of the line.

Theory and Methods

Field observations

Soil temperatures and water levels were measured at five sites along a 250 m representative length of the seismic line (Figure 3).

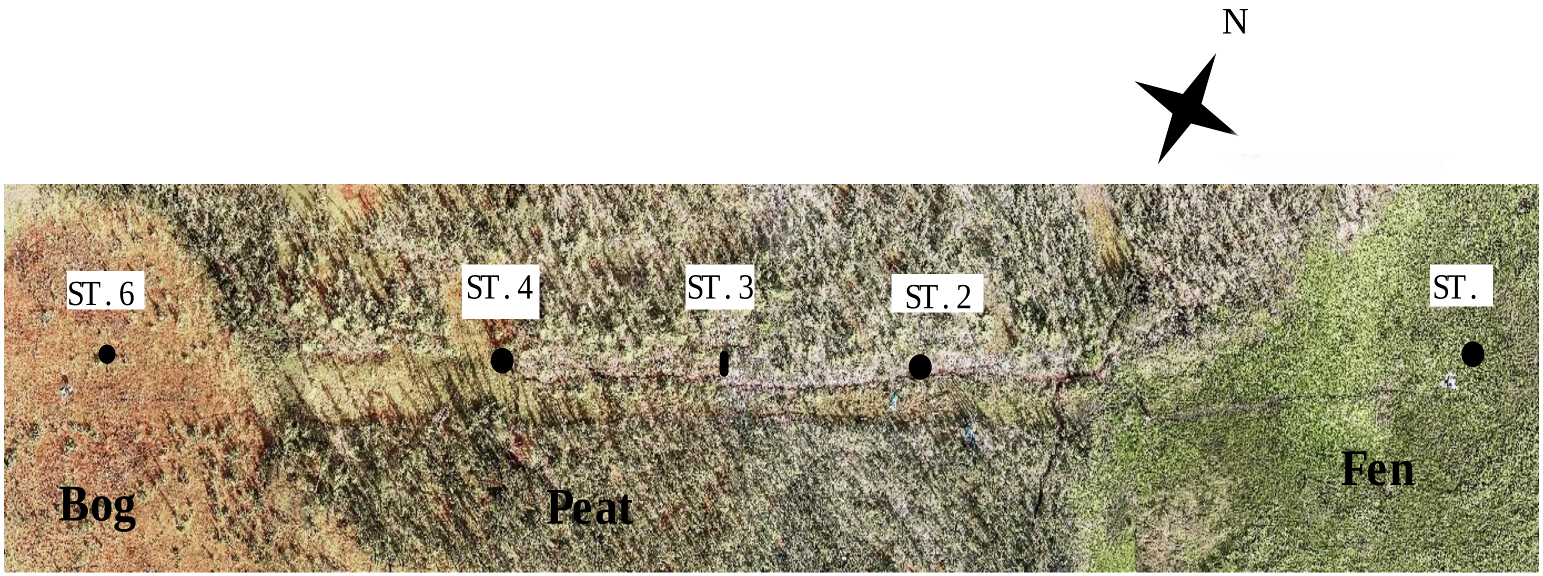


Figure 3: Monitoring stations along the seismic line. Each station was equipped with a HOBO electronic water level transducer and thermistors connected to a Campbell Scientific data-acquisition system at depths of 15 cm to 250 cm. St. 1 is representative of the channel fen, St. 2, St. 3, St. 4 are representative of the peat plateau, and St. 6 is representative of the bog area.

The monitoring stations (St.) were located as follows: St. 1 was in the fen, St. 2, St. 3 and St. 4 were on a portion of the line that traverses a peat plateau. The ground surface at St. 2 and St. 3 was relatively dry, while St. 4 occupied a saturated area with a floating vegetation mat. St. 6 was located in a bog opposite to St. 1 across the peat plateau.

Soil temperatures were measured with CSI 109 soil thermistors installed between 0.15 m and 2.5 m below ground surface (Table 1) and connected to Campbell Scientific data loggers, which measured temperatures every minute, and averaged and recorded temperature values every hour.

Station 1	0.20	0.50	1.00	-	-
Station 2	0.61	1.11	2.29	-	-
Station 3	0.17	0.66	1.15	1.65	1.78
Station 4	0.29	1.29	1.79	2.29	2.39
Station 6	0.14	0.64	1.14	1.64	2.14

Table 1: Depth of thermistor installation, m

At each monitoring station, HOBO pressure transducers were installed in 2 inch inner-diameter slotted PVC pipes to a depth of 1.5 m below ground surface. Water levels were measured every 30 minutes, and averaged and recorded every six hours. The latter were corrected for atmospheric pressure surveyed with DGPS and adjusted for absolute elevation. Prior to equipment installation, depth to permafrost was measured at each monitoring site by drilling holes with a soil auger. The depth to the frost table (*i.e.* the lower boundary of the thawed layer) was measured biweekly over the May-July period in both 2013 and 2014.

Twenty-nine transects crossed the seismic line at right angles, with each transect containing seven measurement points. Starting at the fen and moving into the plateau, the first 17 transects were spaced at 10 m intervals, with the following 12 transects spaced at 30 m intervals. On each transect, the first (south) and last (north) points were located on the undisturbed peat plateau, with all points being 3 m apart from each other, covering a distance of approximately 21 m across the seismic line. On March 27-28, 2015 five boreholes were drilled along a 180 m stretch on the most western side of the 500 m seismic line (Figure 3). The thickness of the active layer and of the talik were measured in each borehole. The low thermal conductivity of the snow cover has a profound effect on ground temperatures. In regions of discontinuous and sporadic permafrost, it can determine whether or not permafrost is present (Mackay and Mackay, 1974). Snow depth, density and water equivalent were all measured along the seismic line and along a parallel transect below the tree canopy of the undisturbed plateau in March 2013 and March 2014. A digital elevation model (DEM) was derived from air-borne LiDAR measurements acquired on August 2, 2010, using an Airborne Laser Terrain Mapper (ALTM 3100) provided by the Applied Geomatics Research Group, NS, Canada, with a resolution of 1.0 m horizontal and 0.2 m vertical. The LiDAR elevations were verified with DGPS in August 2014. An eBee RTK by senseFly fixed-wing unmanned aerial vehicle (UAV) equipped with 3.2 Mpx thermoMAP camera with a ground resolution of 11.57 cm/px at a height of 61 m above the ground was used to acquire thermal imagery of both the seismic line and adjacent undisturbed plateaus. This data was acquired on August 28, 2015 between 17:45 and 18:00 local time, below a partly cloudy sky, with an air temperature of 16^{o} C at ground level and winds from the NW at 3.5 m s^{-1} at ground level. Postflight Terra 3D V3 software was used to produce index maps of ground surface temperatures for qualitative interpretation.

Snow Water Equivalent

The snow water equivalent (SWE) data was collected in March 2013 and March 2014 (Table 2). Snow depth was measured every 10 meters and SWE values were calculated for every 50 meters from west to east along the south forested-plateau of the seismic line, and then from east to west along the central line of seismic line. There is almost no difference in snow depth and SWE between peat plateau and seismic line

	March 2013		March 2014	
	Seismic line	Peat plateau South of seismic line, forest transect	Seismic line	Peat plateau South of seismic line, forest transect
Snow depth, cm (st. dev.)	96.7 (6.9)	94.4 (9.3)	62.3 (4.9)	61.1 (6.1)
SWE, cm (st. dev.)	16.4 (1.8)	16.9 (2.3)	9.8 (1.4)	9.1 (1.3)

Table 2: Snow survey results. Data collected in March 2013 and March 2014.

GPR

A MALA GPR system with a 100 MHz unshielded antenna was used in March and April 2013, when the active layer was at its maximum frozen thickness. The approximate depth range for this antenna is 2-15 m and the lower limit of the object target size is 0.1-1.0 m. Nine transects were measured, three along the seismic line and six across it. The ground conditions at this time of year are ideal for GPR surveys, as the frozen active layer provides little attenuation of the electromagnetic pulse. A handheld GPS unit connected to the GPR was used to determine the location of measurements along each line. The GPR antenna was towed while walking along the seismic line at an average speed of 2 km/hr. Reflex-w software provided by Sandmeir Scientific Software was used for data processing. GPR profiles were verified and correlated with hand augered boreholes and graduated frost probe measurements. Travel times measured by GPR were converted to depth with an average velocity of 0.06 m/ns. Based on the information obtained from the boreholes, we expected the GPS measurements to detect five distinct layers: snow, active layer, talik, perennially frozen material (peat or mineral), and unfrozen mineral material. Two transects were surveyed in March, 2013. Transect 1 was measured along the same undisturbed transect used to measure SWE below the tree canopy (described above). Transect 2 was measured along the middle of the seismic line (Figure 4).

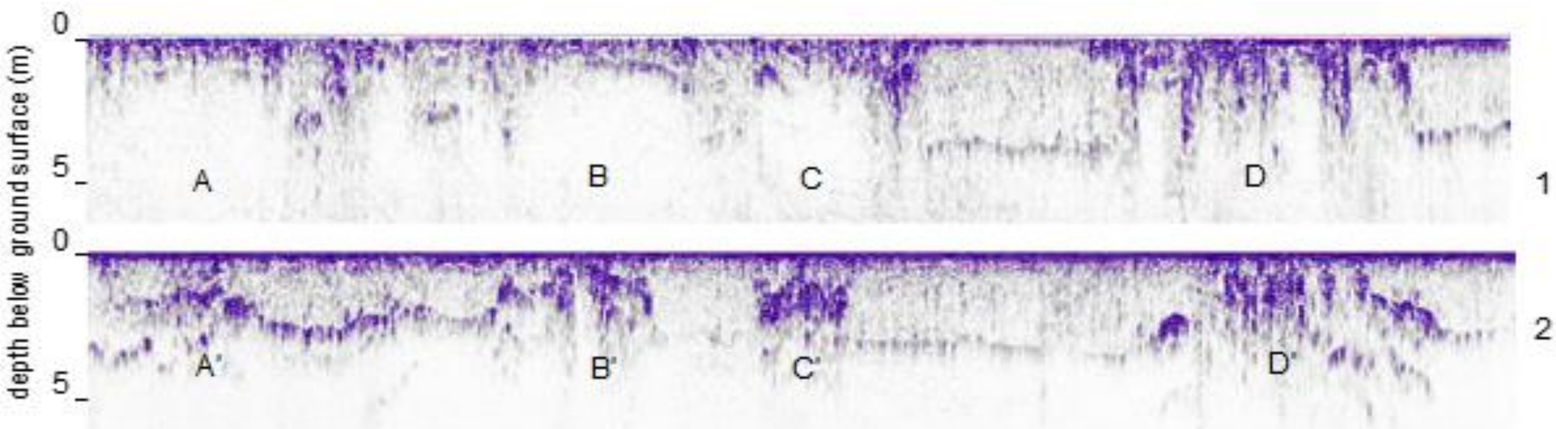


Figure 4: GPR radar profiles through undisturbed (1) and disturbed (seismic line) (2) areas. In this combined double profile, the areas A, B, C, D of undisturbed ice-cored peat plateaus of radargram 1 correspond to areas A′, B′, C′, D′ of disturbed peat plateau under the seismic line. In radargram 2, where the ice is partly thawed, the ice/water interfaces are indicated by the high intensity of signal reflections (deep purple colour).

Active layer thaw and freeze

A simple method for simulating the frost-table depth in hydrological models, described by Hayashi et al. (2007) was used to compute active layer thaw. This method assumes that the majority (>85%) of the ground heat flux into the organic, ice-rich medium is consumed (thawing) or emitted (freezing) as the latent heat required for the phase change. This approach assumes that: 1) the soil consists of an unsaturated layer above a saturated layer; 2) the thermal properties of the peat are water content dependent and vary throughout each layer; and 3) the freeze and thaw of the active layer is one dimensional and governed by the ground surface temperature.

The heat flux into the active layer is described by Fourier's Law:

$$Q = k\,\frac{Ts - T_{fpd}}{Z}, \qquad (1)$$

where Q is the amount of energy transferred [kJ], k is the thermal conductivity [W/m*K], Ts is the temperature [°C] of the ground surface, T_{fpd} is the freezing point depression [°C] , and z is the depth of freeze/thaw [m]. Taking into account that the latent heat of fusion for water is a constant value of L= 334 kJ m^{-3}, we can define:

$$Q_L = mL, \qquad (2)$$

where Q_L is the amount of energy consumed or released during the phase change [kJ] and m is the mass of the water [kg]. Assuming that Ts is constant for the finite period of time Δt, the freezing or thawing will occur to depth dz:

$$Q_L = Lf\frac{dz}{dt}, \tag{3}$$

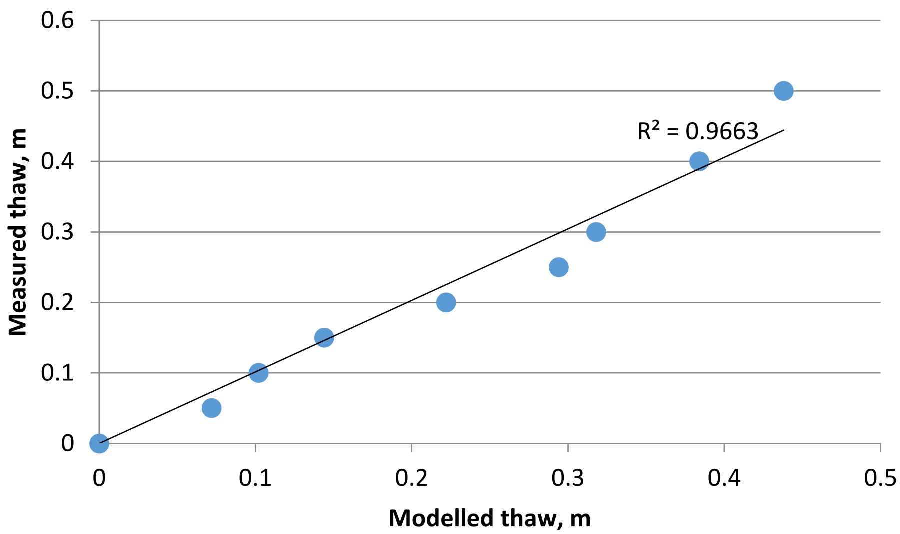
where f is the fraction of water to be frozen or thawed. Combining equations (1) and (3) we can calculate the depth of the freeze/thaw for period dt.

$$\frac{dz}{dt} = k\frac{Ts - T_{fpd}}{ZLf}. \tag{4}$$

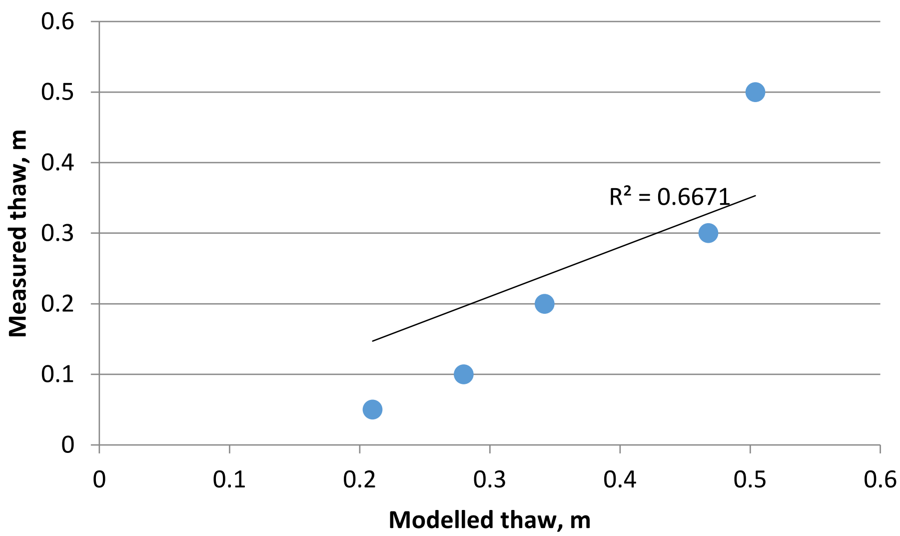
The same approach was used for simulating active layer freeze with the heat flux in the opposite direction.

Model verification and validation.

To verify our model, along with data collected by the author, we used two independent data sets: data collected through a lab experiment at UWO by Aaron Mohammed et al 2014, and field data published by Hayashi et al, 2007. Coefficients of determination (R2 values) for measured versus calculated frost front depths were used to analyse the ability of the proposed model to predict thaw front penetration.



a)



b)

Figure 5: Comparisons between active layer thaw measurement depths published by Hayashi et al, 2007 and model output. a) active layer thaw at the central part of the peat plateau, b) active layer thaw at close proximity from the fen. The solid line is the best fit of the linear regression relationship.

The field data was found to be in relatively strong agreement with the proposed model R2 = 0.95 for data presented by Hayashi et al 2007 in the center of the peat plateau, and R2 = 0.67 for data collected at the edge of the same peat plateau. The lab data showed a considerably slower advance of the thawing front compared to the modeled data (Table 3).

	Days	
Depth, cm	UWO lab experiment	Modelled
0	0	0
5	40	1
15	66	4
25	95	8
35	140	17
45	221	25

Table 3: Data of thaw front propagation collected through UWO experiment and calculated with the proposed model.

The high discrepancy between lab experiment data and modelled data can be explained by a difference in boundary conditions. The proposed model assumes that the bottom of the active layer is at a temperature close to zero point depression, while in the lab experiment the temperature of the bottom layer was maintained within a range of -4ºC and -2.5ºC. The influence of artificial soil refrigeration may explain the slow advance of the thawing front. At the same time we realize that due to the simplicity of the proposed model it cannot be considered as universal and, while it can be applied to the specific field conditions of Scotty creek, it may not work in a different environment.

Sensitivity analysis

The model evaluation was done through the determination of the sensitivity index (SI). The sensitivity index shows output percent difference while changing one input parameter from its minimum value to its maximum. The method was described by Hoffman and Gardner (1983). Results of the sensitivity analysis are presented in Fig.6

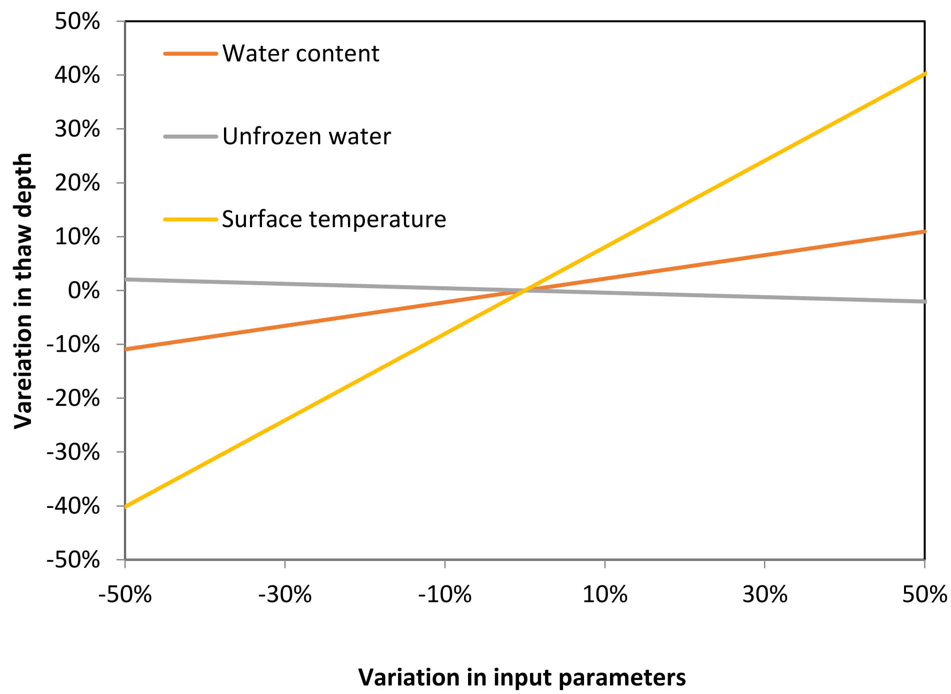


Figure 6: Sensitivity analysis results. The proposed model is most sensitive to surface temperature fluctuations and least sensitive to unfrozen water content.

This analysis was performed by changing one parameter at a time, while keeping the others constant and using their values in model simulation. As seen from the results of the sensitivity analysis, surface temperature had the most profound effect on the rate of thaw, followed by water content and amount of unfrozen water. The total water content of peat is a determining value for thermal conductivity of the soil, which is one of the most important factors for the rate of heat flux.

Permafrost thaw

For disturbed ground, the surface energy balance is shifted towards heat consumption with a reduction in heat loss during the wintertime. This shift causes an increase in energy, which dissipates through conduction to permafrost that is below the active layer. Once the seasonal ice in the active layer has melted, this additional energy is used to warm the active layer and underlying permafrost. Warming continues within a permafrost body until the freezing point depression is reached, after which time energy is consumed as latent heat in support of the phase change from ice to water (Woo, 2012). If the thawed layer does not completely refreeze in the following winter, a perennially-thawed layer (i.e. talik) is formed between the active layer and the permafrost. We suggest that once a talik has formed, the temperature of the permafrost below will slowly rise to the freezing point depression, which is typically between 0^{o} and -0.3^{o} C (Quinton and Baltzer, 2013).

Zero-curtain effect

The 'zero-curtain' refers to the period of several days to several weeks when the warming or cooling of the ground is arrested at the temperature of the freezing point depression owing to the consumption (melting) or release (freezing) of the latent heat of fusion that must occur when passing that temperature threshold. The duration of the zero-curtain is proportional to the amount of water or ice in the ground that must freeze or melt before the temperature threshold is overcome and cooling or warming can resume. Although endothermic, this process is capable of 'heating' or 'cooling' the ambient soil. The rate of temperature change over time in uniform soil can be described by the one-dimensional thermal diffusion equation:

$$\rho c \frac{\partial T}{\partial t} = \frac{dk}{dz}\frac{\partial T}{\partial z} + k\frac{\partial^2 T}{\partial z^2} + \frac{L_f \rho_w}{V_u}\frac{\partial V}{\partial x}, \quad (5)$$

where T is temperature (^{0}C), t is time (s), ρc is the soil heat capacity of frozen soil (J mP^{3} K$^{-'I}$), k is the thermal conductivity of frozen soil (W m^{-1} K^{-1}), z is the depth (m), V_u is the unit volume (m^3), L_f is the latent heat of fusion of ice (J kg^{-1}), ρ_w is the density of water (kg m^{-3}), V is the volume of ice (m^3), and $\frac{\partial T}{\partial z}$ is the thermal gradient. The term $k\left(\frac{\partial^2 T}{\partial z^2}\right)$ represents the heat conducted, while $\frac{L_f \rho_w}{V_u}\frac{\partial V}{\partial x}$ is the latent heat flux. Assuming that permafrost thaw is an isothermal process, we can modify equation 5 as follows:

$$\rho c \frac{\partial T}{\partial t} = \frac{dk}{dz}\frac{\partial T}{\partial z} + k\frac{\partial^2 T}{\partial z^2} + \frac{L_f \rho_w}{V_u}\frac{\partial V}{\partial x} = 0 \quad (6)$$

Since the heat flux is driven by the temperature gradient between the permafrost and the overlying talik, the loss of ice would be most pronounced near the permafrost-talik interface, where the temperature gradient is the largest.

The thickness of organic deposits at the study site ranges from 4.0 m to 8.0 m, which means that about 40% to 80% of the permafrost is ice rich. Due to the large amount of latent heat consumed by melting ice, permafrost remains at temperatures near 0° C, which can be defined as a zero curtain period. The apparent heat capacity depends on the amount of ice melt and represents mostly the latent heat. Theoretically, during the zero curtain period, the temperature remains unchanged while ice is melting. This makes it difficult to estimate the apparent thermal diffusivity during the phase change. To overcome this problem it was assumed that the phase change happens not at a specific temperature, but within the narrow temperature range of -0.2° and -0.5° C. This approach incorporates the effect of latent heat and can be used for modelling of non-conductive forms of heat transfer. Thermal conductivity was calculated as the weighted average of the thermal conductivities of all constituents that form the soil: water, ice, air, and peat fibres.

Subsurface flow along a seismic line

Thawing permafrost can alter water flux and storage processes in the local area of thaw and at larger scales (Wrona *et al*., 2016), ultimately changing the water balance. Water flow is initiated between previously disconnected water bodies following the disappearance of such frozen barriers as permafrost under the peat plateaus. The elevation of the permafrost table was about 1 m deeper than the lowest water level observed during the period of 2012 to 2015. The maximum seasonal frost penetration reached an elevation of 270.5 m, creating artesian conditions and causing a steady increase in hydraulic head during the cold season (Figure 7).

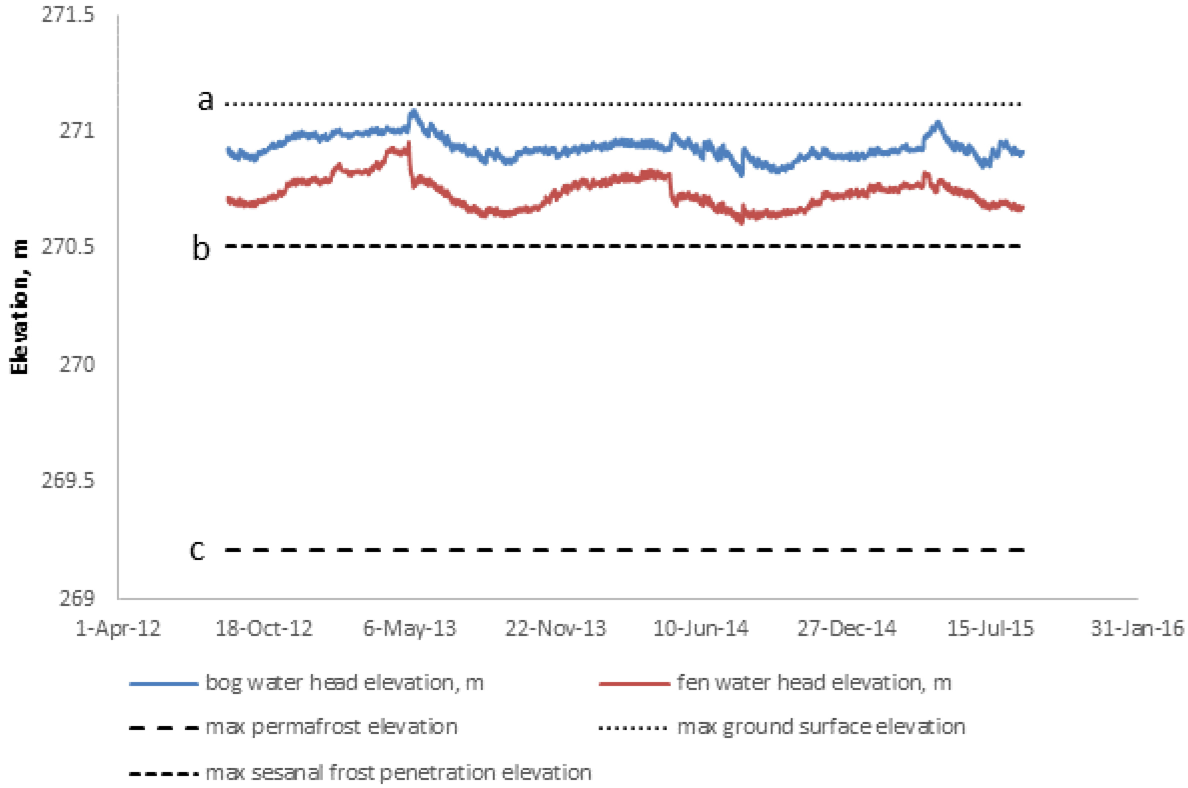


Figure 7: Water levels in the flat bog (blue line) and channel fen (red line) in the period from September, 2012 to September, 2015. Line (a) is the maximum ground surface elevation along the seismic line between the flat bog and the channel fen; line (b) is the elevation of maximum seasonal frost penetration; and line (c) is permafrost elevation at its maximum along the seismic line.

Thus, permafrost thaw under the seismic line creates favourable conditions for year-round water flow between previously disconnected hydrological units. In order to evaluate the hydraulic conductivity of the talik, a slug test was performed in March, 2015. The advantage of conducting this test in wintertime is that the active layer is completely frozen and as such does not exchange water with the unfrozen talik, but confines to the talik, the water introduced to the well. The well was drilled to the top of the permafrost layer and stayed open for the duration of the test. The diameter of the borehole was 0.23m in the active layer, and 0.15m in the underlying permafrost. Measurements of head were collected with a HOBO water level transducer at 1 sec time intervals.

The subsurface water flux through the porous media of a seismic line can be described by Darcy's Law:

$$q = -A_x K_{hx} \frac{dh}{dx}, \tag{7}$$

where q is the rate of flow (m^3 s^{-1}), A_x is the cross-sectional area of flow path (m^2), K_{hx} is the saturated hydraulic conductivity in the x direction (m s^{-1}), h is the total hydraulic head (m), $\frac{dh}{dx}$ is the hydraulic gradient. In saturated soil, the area A_x is given by $A_x\phi$, where ϕ is porosity for which we used a constant value of 80%, the average value reported by Quinton *et al*. (2008) for a peat plateau at Scotty Creek. Darcy's law can be applied to water flow in a confined aquifer, which is the case for winter flow through the talik. During summer, with the absence of frozen active layer the water flow can be described by Dupuit equation for steady unconfined flow:

$$q = \frac{1}{2} K_{hx} \frac{dh^2}{dx} \tag{8}$$

Results and Discussions

Simulation of active layer thaw and freeze

The spatial distribution of ground temperatures within the study area (Figure 7) indicates that the temperatures of the channel fen (5) and the bog (4) were approximately the same on 28 August, 2015 when the thermal image was acquired. It also suggests that they were much warmer than the ground surface temperatures of the winter road (2), narrow bog (3) and peat plateau (1) (Figure 8), which all had similar surface temperatures. The coldest surfaces were along the edges of the peat plateaus, where the relatively cold water draining from the plateau active layers enters the wetlands.

Active layer thaw simulations were conducted for three different ground surface temperature scenarios, distinguished by their mean summer temperature values of 4^{o} C, 6^{o} C, or 8^{o} C (Figure 9).

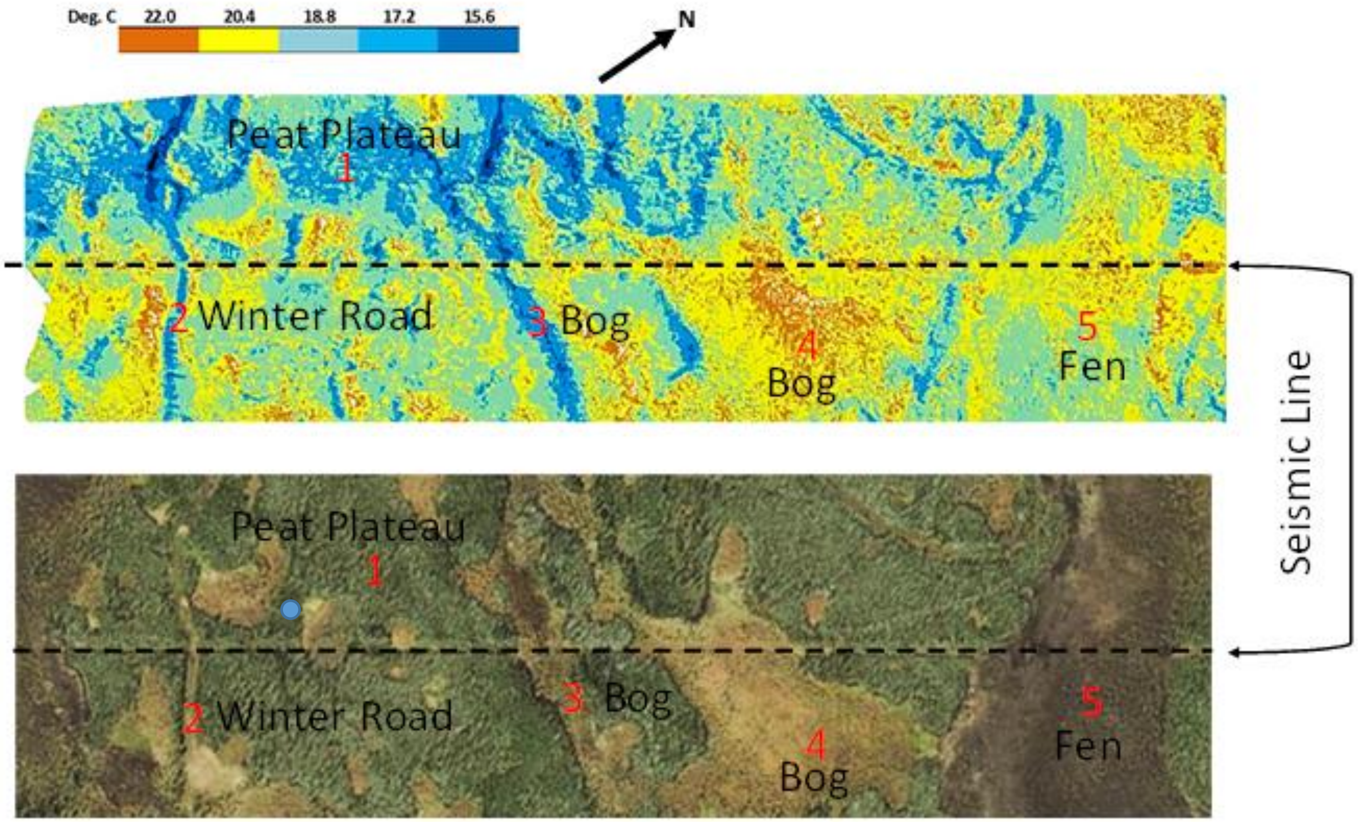


Figure 8: Thermal image of the seismic line, August 28, 2015 5:45 PM, and the corresponding aerial view. The thermal image indicates the ground surface temperatures of the seismic line and adjacent surroundings, where cooler temperatures are represented as blue and warmer temperatures as yellow. The image spans an area of about 0.06 sq. km. The data was obtained with an eBee fixed-wing unmanned aerial vehicle (UAV) equipped with 3.2 Mpx thermoMAP camera.

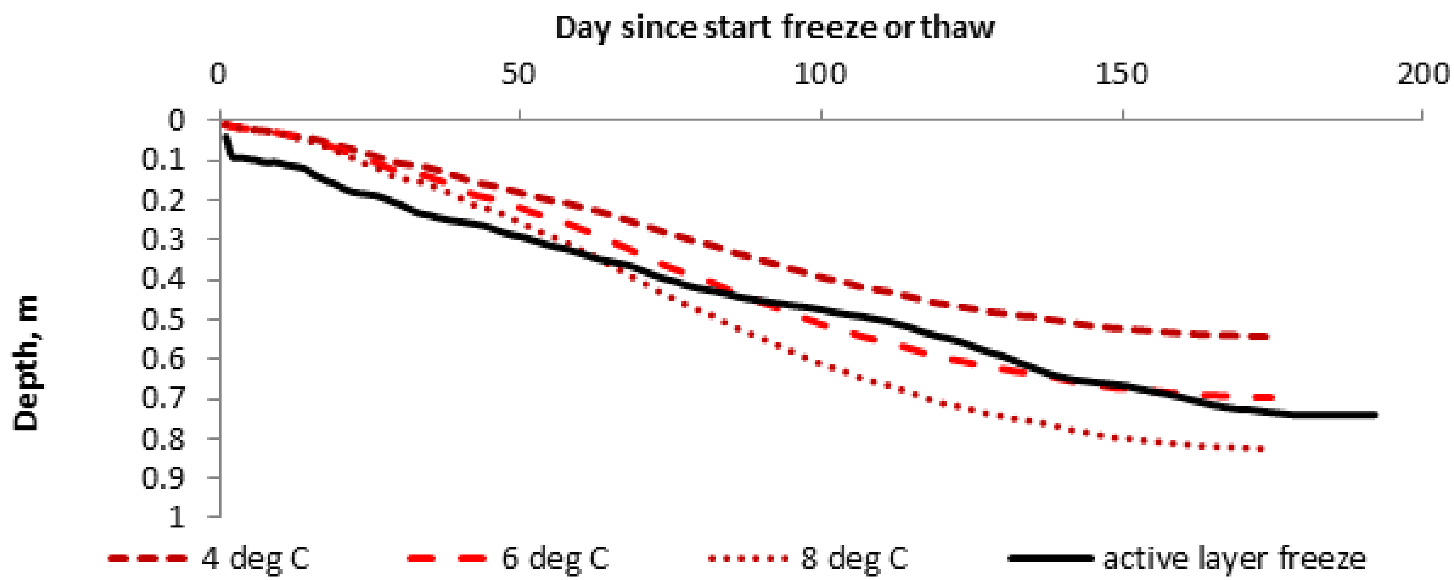


Figure 9: Modelled freeze (black line) and thaw (dashed red line) depth of the active layer for three representative ground surface temperatures simulated for the first year after the seismic line has been cut.

The scenario with the highest surface temperature was assumed to represent the first summer after the seismic line was cut due to the minimum height of vegetation and maximum exposure to short wave radiation. The model showed that the thickness of the active layer in the undisturbed peat plateau with a mean summer temperature of 4^{o} C never exceeded 0.5 m. During wintertime the active layer re-froze to the depth of the permafrost table, thereby maintaining the permafrost in a stable state. However, when the active layer temperature exceeded 6^{o} C it enabled soil to thaw beyond winter frost penetration, creating a perennially unfrozen zone (*i.e.* talik) between the permafrost and the active layer. With a talik present, the complete thaw of the overlying active layer occurred in 4 to 6 weeks, exposing the permafrost to even greater vertical energy conduction from the ground surface, enough to initiate permafrost thaw.

GPR results

The GPR profiles (Figure 4) and borehole observations indicate a peat - mineral interface at approximately 5 m depth. The darker shades indicate the relatively high signal reflection from materials with high liquid moisture content. The absence of colour on transect 1 in areas A and B at a depth of approximately 1.5 m indicates highly porous, ice-rich soils and low liquid water content. Plateau D is located between two large water bodies: an open bog on the west side and a fen on the east side. The high volumetric heat capacity of water allows the talik to act as a large heat storage that affects the thermal regime in areas of the peat plateaus adjacent to the seismic line.

Plateau C was also affected by heat from the open bog, as indicated by the very high reflections on its east side. Owing to the close proximity of plateau C to the larger plateau B, the west side of plateau C has a higher volume of ice, causing lower reflections on the radargram. The disturbance from the seismic line along transect 2 results in permafrost thaw to a depth of about 3 m. The darker shade on transect 2 corresponds to the higher number of point reflectors, *e.g.* ice-water interfaces. It also indicates a higher amount of unfrozen water not only at the border between the permafrost and talik or unfrozen bog and fen, but throughout the whole body of permafrost as well.

Modelling thermal regime of thawing permafrost

The one-dimensional thermal diffusion equation with phase change (5) on page 59 was used to model permafrost thaw.

The following assumptions were made: 1) thawing is initiated at the bottom of the talik and 85% of available heat is used for phase change, while 15% is used for soil warming (Hayashi et al., 2007); 2) the mean annual talik temperature and soil properties control the heat transfer to the top of permafrost; 3) the distribution of ice in the frozen layer determines the nonconductive heat transfer throughout the permafrost body; 4) permafrost consists of peat and mineral layers, both of which are fully saturated; 5) the physical and thermal properties of each soil type (Table 4) are assumed to be constant throughout the whole thickness of the layer .

Porosity	80 %
Initial water content by total volume	25 %
Initial ice content by total volume	55 %
Fraction of peat fibers by total volume	20 %

Table 4: Soil properties used for modelling of the permafrost thaw.

The thickness of the organic layer is 4 m, and the thickness of the underlying permafrost in, mineral soil is 6 m. The simulation was conducted over the period 1985 to 2015 during which the average annual air temperature rose by 0.01° C per year. The model shows that by 2015, the ice content in the permafrost reduced to approximately 70% of what it was in 1985, and the top of the permafrost descended to approximately 2.5 m from the ground surface. Furthermore, the model suggests that by 2216, permafrost below the seismic line will be sporadic, and by 2276, it will disappear completely . However, once the permafrost becomes discontinuous, the one-dimensional approach described here would underestimate the rate of thaw, since the energy would also enter the remaining permafrost fragments horizontally. It is also expected that the rate of thaw would increase with increasing fragmentation of the remaining permafrost (Baltzer et al., 2014). The author realises that any prediction for over 250 years is not feasible due to the very complex processes that will take place over such a lengthy period of time. However, the proposed model reflects the main trend of permafrost degradation, which makes its recovery under current conditions impossible.

Seismic line as a flowpath between water bodies

The slug test indicated a hydraulic conductivity of 3.5 m/day for the talik. Assuming that hydraulic conductivity is constant along the entire length of the seismic line between the bog and the fen, and that on average, the talik thickness is 2 m and width is 10 m, the subsurface discharge volume through the seismic line was computed for the period of October 2012-September 2015 (Figure 10).

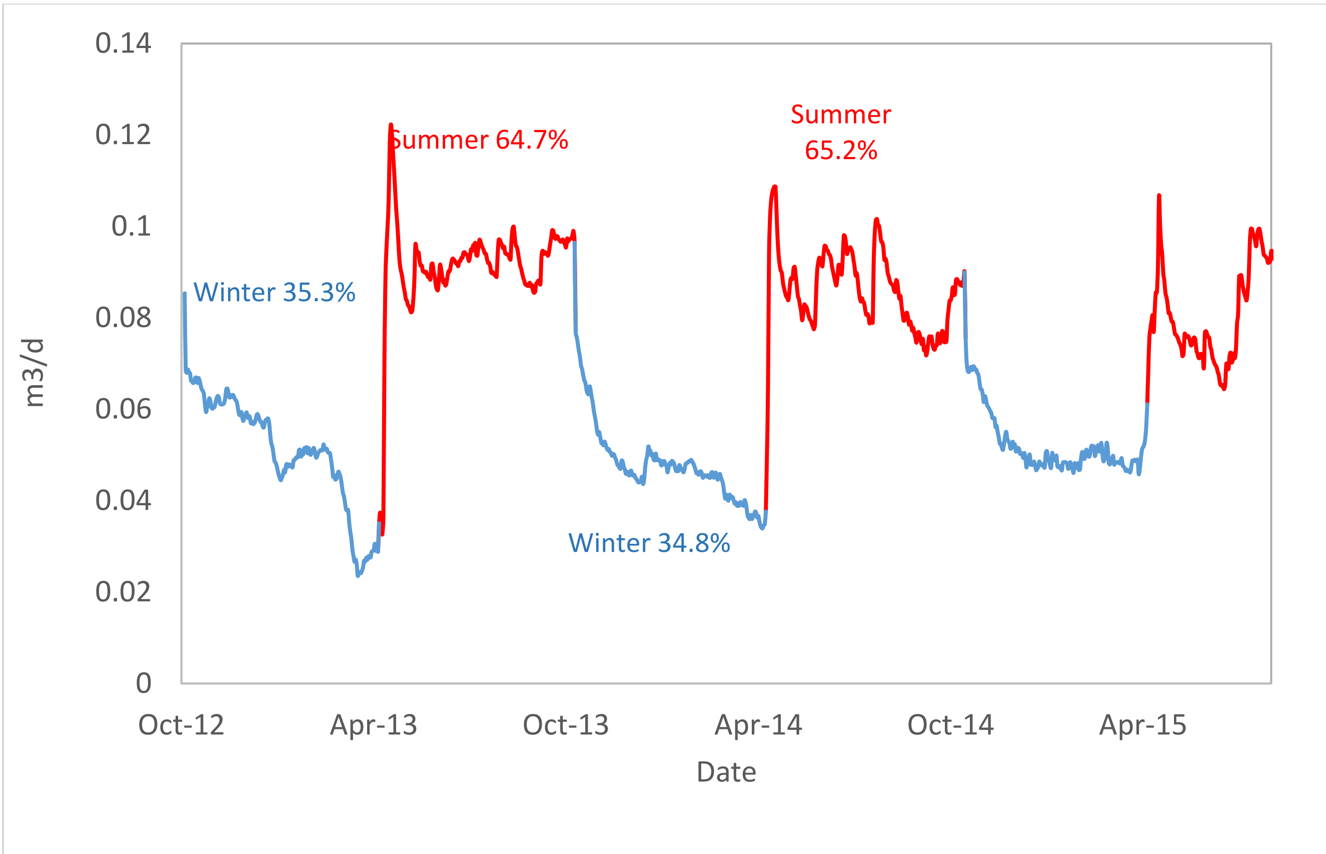


Figure 10: Specific discharge m^3 d^{-1} through 180 m of the seismic line from bog to fen for the period of September 2012-September 2015. Winter of 2012-2013 – 35.3% of annual discharge; summer of 2013 – 64.7% of annual discharge; winter 2013-2014 – 34.8% of annual discharge; summer 2014 – 62.5% of annual discharge. Blue colour represents winter flow and red colour summer flow.

Over this period, the winter flows through the talik (October until May) were about half of the summer flow through the entire seismic line. For the three years, the average total annual discharge was relatively low (28 m^3 yr^{-1}, SD =0.02), owing to the very low hydraulic gradient (0.0011). However, as permafrost proceeds, and the permafrost table lowers, the annual subsurface discharge through the seismic line will increase. Interpretation of the GPR measurements in light of the thaw simulations suggests that although thermally homogeneous at the freezing point depression, thawing permafrost bodies may be heterogeneous with respect to water and ice contents and as a result, heterogeneous in terms of hydraulic conductivity. Subsurface drainage through "*wet permafrost*" has not been reported in the literature, yet may be important as a hydrological flowpath, and as such, warrants further study.

Conclusion

This study analyzed the impact of a seismic line on peatland hydrological regimes in the region of discontinuous permafrost in the Scotty Creek basin. Based on field observations, interpretation of the GPR images, and the results of mathematical modelling, we concluded that linear disturbances, such as seismic lines cut over peatlands in the region of discontinuous permafrost, trigger talik development, and a decoupling of permafrost from freezing winter temperatures. The model shows that the complete refreeze of the active layer to the depth of 0.5 m is possible if the mean ground surface temperature during the thaw period remains below 6^{o} C. When the mean temperature rises above 6^{o} C, a talik inevitably appears between the permafrost and the active layer and as a result, the permafrost table maintains contact with liquid water throughout the year, a condition that enables permafrost thaw. When talik is present, the upward freezing of the active layer in wintertime no longer occurs. The absence of sub-zero temperatures above the permafrost makes thawing irreversible. As soon as the top layer of permafrost subsides below the water level, a hydrological link between previously disconnected water bodies such as flat bogs and channel fens can be established. It was also found that in the presence of a hydraulic gradient along the linear disturbance water flow takes place all year round, with the winter volume being approximately half of the summer volume. Mathematical modelling revealed that the initial stage of permafrost warming to a zero-point depression is followed by a longer stage of permafrost thaw, characterized by latent heat exchange between the center and edges of the permafrost. Ice content gradually decreases throughout the entire body of permafrost, creating a slushy ice water mixture within the soil matrix. In cases where the soil is highly porous, this process can cause a significant increase in hydraulic conductivity, reducing the role of the peat plateau as a natural barrier between the hydrological units of the peatlands. These processes likely have sufficient influence over water distribution in the local water basin increasing total runoff from the basin. By comparing GPR images of both the disturbed and undisturbed areas, it was found that large water bodies adjacent to the peat plateaus have the same effect on the frozen core of peat plateaus as linear disturbances. These results allow us to estimate the approximate time when the connection between bog and fen was established (10-12 years after line was cut) and predict the complete disappearance of permafrost under the linear disturbance within 250 years. This time period can be considerably shorter, however, if advection as well as climate warming scenarios are taken into account. Further studies of hydrological parameters of disturbed as well as undisturbed peat plateaus are recommended.

The main causes of permafrost thaw under linear disturbances are: the increase of the mean annual ground surface temperature due to vegetation removal; and an increase in the density and water content of the unsaturated layer, which results in an increase in the thermal conductivity of the soil. Deeper than usual thawing of the active layer in the first summer following the introduction of a disturbance creates a suprapermafrost zone, or talik, that remains unfrozen all year round.

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Conclusion

The presented study demonstrates that, under warming climate conditions, such local factors as linear surface disturbance in the form of seismic lines can have a profound effect on the ground thermal regime. According to our field observations and based on the results of mathematical modelling, we came to the conclusion that the formation of a perennially unfrozen layer between permafrost and the active layer - talik - is an inevitable result of the cutting of seismic lines over peat plateaus. Once permafrost is decoupled by talik from subzero winter temperatures of the active layer, permafrost thaw becomes irreversible. In peatlands, the frozen core of peat plateaus forms a natural barrier between bogs and fens. However, when the depth of the permafrost table becomes lower than the ground water table, seismic lines create links between fens and bogs, slowly conveying water from a higher level to a lower one. This process may affect water distribution in the local water basin, increasing its total runoff. We have demonstrated that due to the unique geometry of seismic lines, talik occurrence has a significant effect on local hydrology and leads to water drainage occurring all year round, rather than being limited to the summer time or spring freshet. This process may cause a substantial shift in the water balance of the basin by draining previously isolated bogs and affect the thermal regime of the frozen core of peat plateaus, causing an accelerated thaw of the latter. Our study shows that seismic lines have a dual effect on the peatland environment: 1) they increase runoff of contributing areas and consequently the total runoff of the basin; 2) they short-circuit the natural drainage network by cutting indiscriminately across bogs, fens and plateaus, thus reducing runoff transit times. These effects have the potential to alter the water balance of the basin.

Using our mathematical model, we showed that the maximum thickness of the active layer in the peatlands of the Scotty Creek basin cannot exceed 0.7 m, which was confirmed by field observations. Based on this conclusion, we can state that any permafrost deposits, even those occurring on undisturbed peat plateau below this depth, are decoupled from the freezing winter temperatures of the active layer by talik and as such are subject to degradation. The proposed model forms the basis for forecasting thermal effects, not only in response to surface disturbances such as seismic lines, but also in response to warming climate. It should be noted that conduction is the only heat transfer method involved in the proposed model. In case of linear disturbance, the effects of convection can be as significant as those caused by conduction and should be considered as a subject for the further study.

Future Research

It should be noted that the process of talik development and permafrost thaw in peat plateaus discussed in this paper occurs naturally in response to a warming climate. However, this process is expedited by the cutting of seismic lines. Therefore, studying permafrost degradation under such linear disturbance could help predict future transformations of the peatland landscape under the conditions of a warming climate. As was mentioned in this research, the density of seismic lines in the Mackenzie Valley varies from 6 km/km^2 in the delta to 1.2 km/km^2 in the research area. The hydrological impact of linear disturbances that short-circuit the natural drainage network is little understood and has not received the deserved attention in the scientific community. Further research should be conducted in order to gain a better understanding of this impact.

Our study suggests that permafrost thaw under linear disturbances in the Scotty Creek basin is irreversible under natural conditions. However, it provides the necessary ground for future research of possible techniques for artificial permafrost restoration. The seismic industry is currently adopting Low Impact Seismic techniques where mulching is considered as a essential component of the seismic line cut process. Mulch as artificial substitution for removed or compacted peat has low thermal conductivity, preventing active layer from heat penetration during summer time. Together with active ground cooling with thermosyphons it can by successful technique for permafrost restoration.