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Modeling hydrothermal regimes and potential impacts of climate change on permafrost within the South Mackenzie Plain, Northwest Territories, Canada [1]

Marie-France JONES, Mark CASTONGUAY, Mina NASR, Jae OGILVIE & Paul A. ARP [2],
Forest Watershed Research Centre, Faculty of Forestry and Environmental Management,
University of New Brunswick, PO Box 44555, Fredericton, New Brunswick
E3B 6C2, Canada, e-mail: arp1@unb.ca
Jagtar BHATTI, Northern Forestry Centre, 5320-122nd Street, Edmonton, Alberta T6H 3S5, Canada.

Abstract: Hydrothermal processes are key components of permafrost dynamics and hydrological and carbon cycles in northern forest ecosystems. A forest hydrology model, ForHyM, was used to evaluate these processes by determining how the depth and duration of frost penetration into the soil would vary daily over the course of several decades. This was done for chosen upland/wetland conditions within the Mackenzie Plain south of Fort Simpson, where permafrost is currently sporadic to discontinuous. The model calculations were done using daily weather records from November 1963 to 2010, starting with a hypothetical no-frost condition within the soil and subsoil. Model performance was evaluated by comparing modeled and measured temperatures at different soil depths (upland and peat plateau modelling, R^2 = 0.95 and 0.94, respectively). It was found that well-drained upland forests within the general area would experience deep and complete freeze–thaw cycles each year. In contrast, poorly drained wetlands would develop gradually deepening permafrost that would at first stabilize in depth over the course of 10 to 20 y, with thaw depth limited to <1 m each year. However, recent increases in recorded air temperature (more so in winter than in summer) would destabilize the permafrost layer, and this would especially occur in areas with insufficient surface insulation by local peat, moss, forest litter, and snow accumulations. These estimates are consistent with (*i*) reported thawing depths and (*ii*) the widening encroachment of collapse scars towards the poorly drained portions of the South Mackenzie Plain.
Keywords: climate change, hydrothermal modeling, permafrost.

Résumé : Les processus hydrothermaux sont des composants clés de la dynamique du pergélisol et des cycles hydrologiques et du carbone dans les écosystèmes forestiers nordiques. Un modèle d'hydrologie forestière, ForHyM, a été utilisé pour étudier ces processus en déterminant comment la profondeur et la durée de pénétration du gel dans le sol variaient quotidiennement sur plusieurs décennies. Ceci a été fait pour des conditions sèches et humides dans la plaine du Mackenzie au sud de Fort Simpson, où le pergélisol est actuellement sporadique à discontinu. Les calculs des modèles ont été effectués en utilisant les données météorologiques quotidiennes de novembre 1963 à 2010, avec au départ une condition hypothétique d'un sol et d'un sous-sol non gelés. La performance du modèle a été évaluée en comparant les températures modélisées et mesurées pour différentes profondeurs de sol (modélisation pour les hautes terres et les plateaux tourbeux, R^2 = 0,95 et 0,94, respectivement). Nous avons trouvé que les forêts bien drainées des hautes terres de l'aire étudiée subissaient des cycles de gel et de dégel complets et profonds chaque année. Au contraire, les milieux humides mal drainés développaient un pergélisol qui s'approfondissait graduellement et se stabilisaient en profondeur après 10 à 20 ans; la profondeur de dégel étant limitée à <1 m chaque année. Cependant, les récentes augmentations de la température de l'air (plus marquées en hiver qu'en été) ont déstabilisé la couche de pergélisol, particulièrement dans les zones où l'isolation superficielle est insuffisante; une telle isolation pouvant être procurée par de la tourbe, de la mousse, de la litière forestière ou des accumulations de neige. Les estimations obtenues sont compatibles avec (*i*) les profondeurs de dégel rapportées et (*ii*) l'empiétement de plus en plus important des cicatrices d'affaissement dans les secteurs mal drainés du sud de la plaine du Mackenzie.
Mots-clés : changement climatique, modélisation hydrothermale, pergélisol.

Nomenclature: Flora of North America Editorial Committee, eds, 1993+.

Introduction

Permafrost, defined as subsurface soil remaining below 0 °C for 2 consecutive years, is widespread in the Arctic and boreal regions of the Northern Hemisphere, where it occupies 22% of the exposed land surface area (Zhang *et al.*, 1999). In general, it is influenced directly by climate through changes in air temperature (Osterkamp, 2007; Romanovsky, Smith & Christiansen, 2010) and snow (Stieglitz *et al.*, 2003), or indirectly through disturbance (*e.g.*, wildfire [Yoshikawa *et al.*, 2003]) or local changes in hydrology (Jorgenson *et al.*, 2010), soil moisture (Lawson, 1986; Hinzman *et al.*, 1991; Chapin *et al.*, 2000; Woo & Marsh, 2000; Yoshikawa *et al.*, 2003; Shur & Jorgenson, 2007), and soil thermal regime (Tchebakova, Parfenova & Soja, 2009; Jorgenson *et al.*, 2010). The permafrost

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[2] Author for correspondence.
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in the southern boreal forest region of the southern Mackenzie Plain is relatively "warm" (mean annual temperature 0 to –3 °C), thin, and discontinuous (Jorgenson *et al.*, 2001), and is therefore particularly susceptible to thawing and disappearance (Camill, 2005; Tarnocai, 2006; Quinton, Hayashi & Chasmer, 2011). Permafrost in this area mostly occurs on flat areas covered by 0.5- to 1.5-m elevated peat plateaus (Quinton, Hayashi & Chasmer, 2011). Here, annual thawing tends to be limited to a depth of 50 cm, and is less underneath thick moss and lichens cover (Vitt, Halsey & Zoltai, 1994). In general, increasing air temperatures coupled with increased precipitation would stimulate permafrost thawing, while a persistent decrease in snowfall would reduce the insulating effect of the snowpack, thereby intensifying the extent of soil frost penetration each winter (Stieglitz *et al.*, 2003). Disturbing the forest floor through, *e.g.*, forest harvesting could reduce surface insulation during winter, but this could also lead to a deepening of thawing during summer. Poorly drained peat soils likely remain frozen for longer periods of time than well-drained soils in areas with long winters and short summers, because soils with high moisture content require higher amounts of thermal energy to thaw than soils with low moisture content (Balland *et al.*, 2006). In contrast, well-drained soils would have deep freezing and thawing cycles. Additionally, geothermal gradients below the soil affect the depth of local permafrost formations as well, with steeper gradients occurring along fault lines.

Currently, the peat plateaus are increasingly subject to permafrost collapse, leading to an increased frequency and extent of collapse scars (thermokarst) on peat plateaus. These scars are filled with water from thawing permafrost, and support aquatic herbaceous vegetation (Jorgenson *et al.*, 2001; Jones *et al.*, 2013). The increased presence of collapse scars is therefore an example of the drastic changes in hydrology, vegetation, and carbon cycling caused by permafrost degradation (Jorgenson, Shur & Pullman 2006; Osterkamp *et al.*, 2009; Moskalenko, 2013). Hence, discerning how hydrothermal processes affect the freezing and thawing cycles has become important (McGuire *et al.*, 2002; Walker *et al.*, 2003; Trettin *et al.*, 2006).

The approach used in the present study tracked the flow of moisture and heat through the soil under varying insulating surface cover from barren soils to peat and forest. There are several initiatives coordinated by Natural Resources Canada to monitor and document the increases in ground temperature currently observed throughout northern Canada (Burgess, Desrochers & Saunders, 2000; Smith *et al.*, 2005).

Methods

The approach used in this study

(*i*) modeled soil moisture, heat, and frost with the Forest Hydrology Model (ForHyM; Arp & Yin, 1992; Balland *et al.*, 2006), based on the daily weather records for precipitation and air temperature from 1963 to 2010;

(*ii*) started these simulations with a hypothetical no-permafrost condition everywhere at the beginning of these records; and

(*iii*) simulated the onset of permafrost conditions based on climate conditions at select upland and wetland sites within the South Mackenzie Plain (Northwest Territories, Canada; Figure 1), with reference to temperature documentation by Smith *et al.* (2009).

STUDY AREA

The study area lies within the Mid-Boreal Ecoregion of the Mackenzie and Slave Lowlands. This particular region has the mildest climatic regime of the Taiga Plains, receiving an average of 21.7 cm of rainfall and 14.2 cm of snowfall per year. Mean high and low temperatures vary from 23.1 to 10.9 °C in July and from –23.3 to –31.7 °C in January. The terrain varies from level to gently rolling on low-relief Paleozoic strata. Extensive till, alluvial deposits, and clayey lacustrine plains interspersed by peat blankets are the dominant landforms. Organic cryosols, organic soils, and gleysols are dominant on the poorly drained plateaus and lowlands. Luvisols dominate on the better drained upland positions. Mixed-wood, deciduous, and coniferous stands, including jack pine, occur on imperfectly to rapidly drained mineral soils along glacio-fluvial deposits, notably along the Mackenzie, Liard, and Jean Marie rivers. Upland forests occur on undulating or sloping terrain. Elsewhere, black spruce and mixed-wood stands intermingle with fens, bogs, peat plateaus, and collapse scars. Peat plateaus with stunted black spruce–lichen forest, collapse scars with sedges and mosses, and fens with black spruce, larch, dwarf birch, rap sedges, and mosses are widespread on level terrain, where water tables are usually high and organic materials have accumulated to varying depths (Ecosystem Classification Group, 2007; Figure 1). Four areas were selected for modeling purposes (Figure 2), and 2 of these areas (C and D) were selected for soil temperature measurements. Data collection was done from 2008 to 2009.

HYDROTHERMAL MODELING

GENERAL

ForHyM is an aspatial model designed to simulate water and heat fluxes through all major ecosystem compartments (canopy, snowpack if present, forest floor, rooted portion of the mineral soil, subsoil; Figure 3). ForHyM requires daily weather (mean daily air temperature, daily rain, and snow) and basic site descriptors (slope, aspect, elevation, soil depth, texture, organic matter, coarse fragment content, and forest cover type) as input (Figure 3). The model calculates daily canopy interception, snowpack water equivalents, frost depth (upper boundary of ice formation), lower boundary of ice formation, soil moisture (frozen and unfrozen), and soil temperature at any depth. Soil calculations extend from top of the snowpack to a default depth of 12 m, where soil temperatures tend to remain constant year-round. This depth can be extended downward towards non-frozen conditions where needed.

Throughout the simulations, the principles of mass and heat conservation are strictly obeyed, as moisture and heat are simulated to pass from the atmosphere to the forest canopy, from the canopy to the snow pack when present, from the snowpack to the moss and/or lichens layer (when present), from there to the forest floor and mineral soil

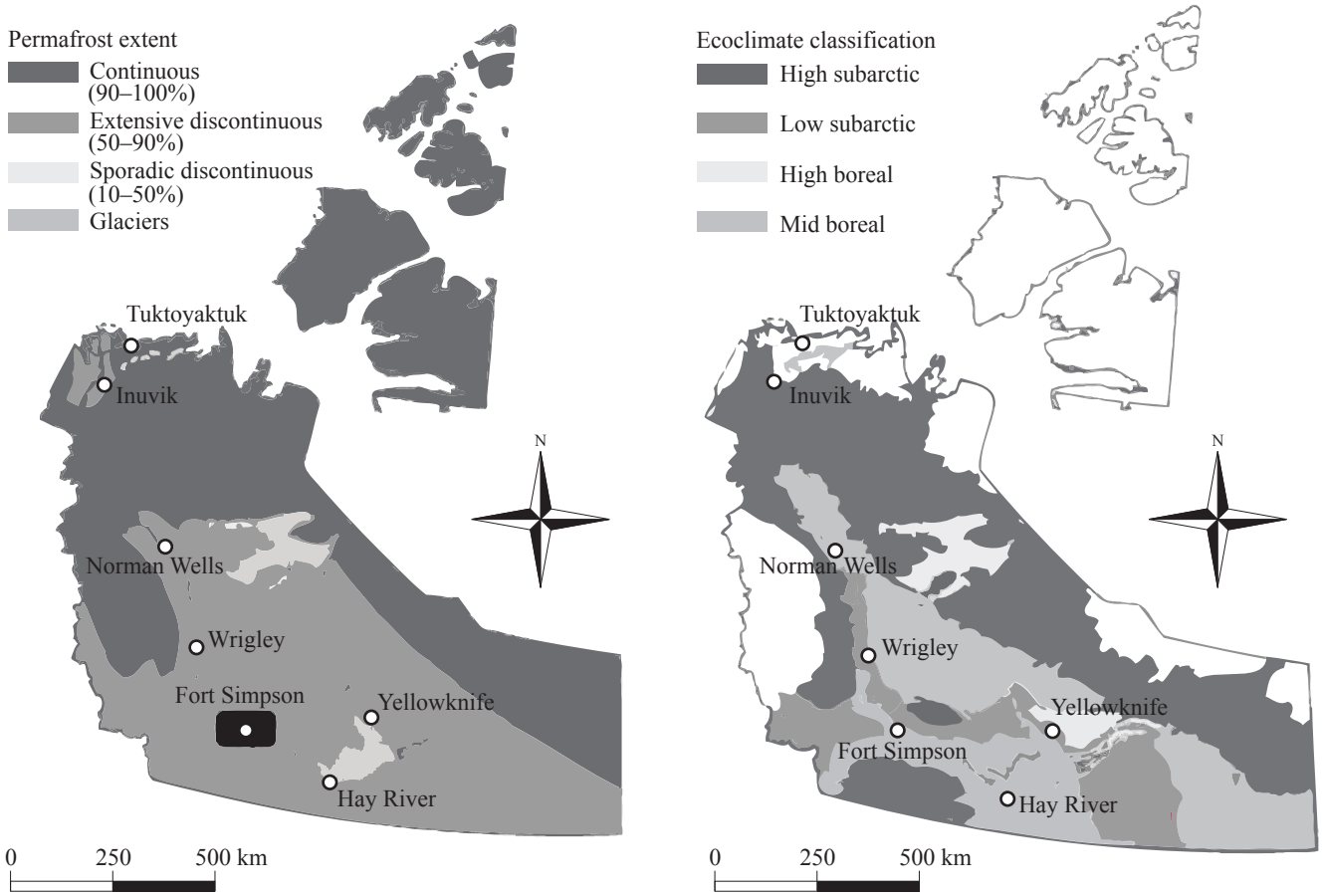


FIGURE 1. Locator map for the study area (black) within the South Mackenzie Plain of the Northwest Territories, Canada.

and subsoil layers below and beyond, including the bedrock. The heat flow calculations for the snowpack and the underlying layers are based on deriving the temperature at the ground/snow surface from the surficial energy balance. This balance is obtained by explicitly addressing all major incoming and outgoing heat fluxes as modified daily by the changing conditions in the atmosphere (set to be cloudy during precipitation events) and albedo (varied depending on the canopy leaf area index and snow cover). The transmission of the heat from one soil layer to the next is based on an implicit difference formulation of the heat flow equation, accounting for changes in heat capacity and thermal conductivity as these parameters change with texture, soil bulk density, organic matter and coarse fragment content, soil moisture, and latent heat change of water to ice and vice versa (Arp & Yin, 1992; Yin & Arp, 1993). Basic to the hydrothermal calculations dealing with soil freezing and thawing is determining how the heat and water retention characteristics of soils vary from frozen to unfrozen, mineral to organic, compact to loose, wet to dry, and coarse- to fine-textured conditions. The process of doing this and evaluation of the results has been described by Balland and Arp (2005), Balland, Pollacco, and Arp (2008), and Jutras and Arp (2010; 2013). The depth of the lower boundary of the frozen ground was set based on available bore-hole data delineated for the general Fort Simpson area. Within the model, the lower heat flow boundary condition was evaluated at a depth where the ground would be non-frozen, using nearest geothermal gradient determinations. In doing this, the ForHyM formulation is similar to the one-dimensional vertical soil freeze–thaw formulations of

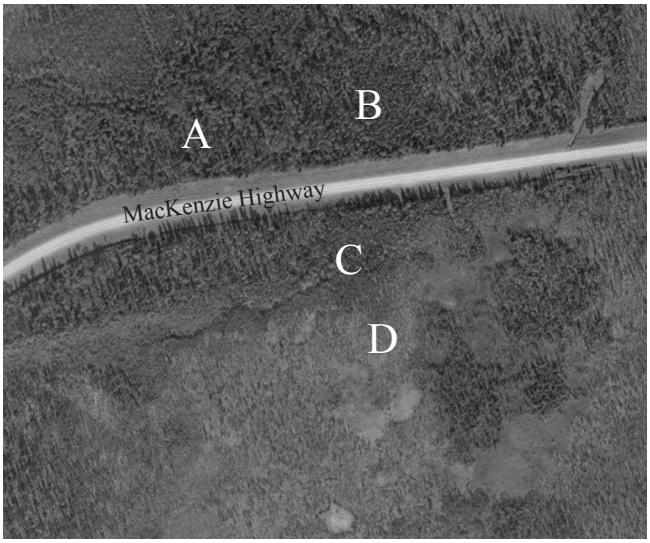


Figure 2. Study area plot locator map; A = aspen upland, B = jack pine upland, C = black spruce upland, D = black spruce wetland (peat plateau). Soil temperature measurements were obtained for sites C and D at 5, 25, and 50 cm depths.

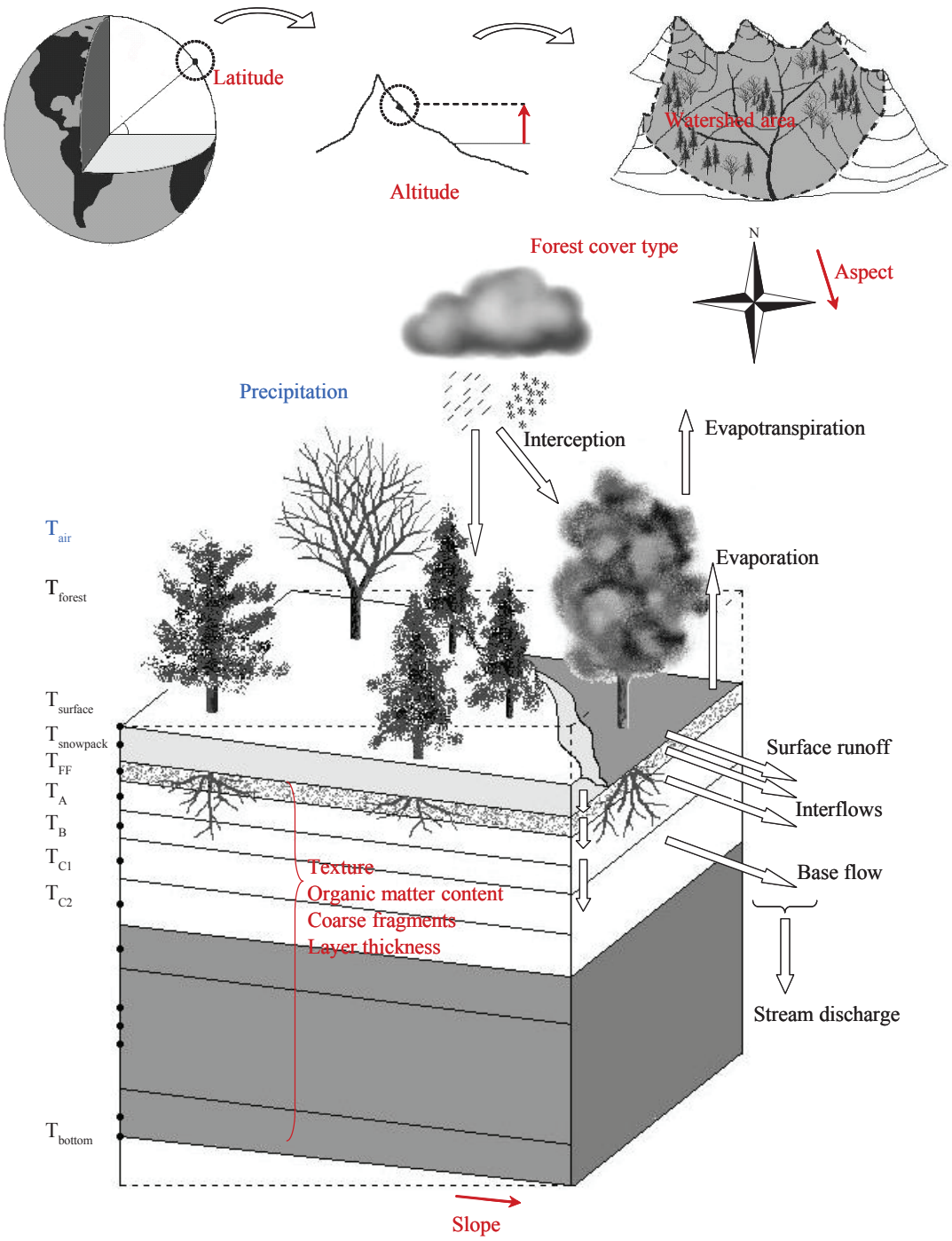


FIGURE 3. Overview of processes and initialization requirements (in red and blue) of the Forest Hydrology Model (ForHyM) (Balland *et al.*, 2006).

Webb (1997), Peck and O'Neill (1997), Albert, Koenig, and Mason (2000), and Hansson and Lundin (2006). The innovative aspect of ForHyM is that it can be calibrated to conform with location-specific data for soil moisture, temperature, snowpack depth, and stream discharge by adjusting its parameters for snow density, actual evapotranspiration, and heat conductivity once basic soil (texture, organic matter content, coarse fragment content) and watershed conditions (slope, aspect, elevation, leaf area index) are specified (Jutras & Arp, 2010; 2013).

APPLICATION

ForHyM was calibrated with daily measurements at the BOREAS old-growth jack pine, aspen, and black spruce forest sites (Balland *et al.*, 2006), and the calibrations so obtained for the snowpack, soil moisture, and soil temperature calculations, including soil freezing and thawing, were then applied to the prevailing site conditions within the South Mackenzie Plain (Table I). Daily air temperature, rain and snow fall, and snow-on-ground (SOG) records were obtained from the Atmospheric Environment Service (AES)

TABLE I. Site location and ForHyM calibration adjustments for jack pine, aspen, black spruce, and peatland sites.

Site	Location coordinates	Canopy transparency adjustments	Snowfall adjustments	Air temperature adjustments	Soil layers	Thickness (cm)	Texture [a]	Organic matter fraction	K_{sat} cm·h^{-1}	Permeability adjustments		Dry thermal conductivity (W·m^{-1}·°C^{-1})	Thermal conductivity adjustments [b]	Maximum frost action [c]
										Downward	Lateral			
Jack pine upland	61°38'20.86"N 121°23'48.44"W	0.27	1	0 °C	Forest floor	3–9 [d]	O	0.515	1.99	1	0.500	0.03	1	1
					A	30	S	0.003	63.46	1	0.005	0.26	1.09	1
					B	30	S	0.001	20.12	1	0.005	0.33	1.07	1
					C1	45	S	0.001	6.80	1	20	0.40	1.09	1
					C2	46	S	0.001	0.29	1	20	0.03	1.09	1
					Subsoil+	100	S	0	0.89	1	20	0.57	0.50	1
Aspen upland	61°38'19.54"N 121°24'2.46"W	0.11	1.25	0 °C	Forest floor	8	O	0.770	0.72	1	0.100	0.03	1	0.75
					A	10	SL	0.015	37.66	0.500	0.005	0.17	1	0.8
					B	45	SCL	0.011	25.95	0.500	0.005	0.25	1.12	0.4
					C1	20	SCL	0.041	11.02	0.002	0.200	0.19	1.35	0.4
					C2	20	L	0.053	10.58	0.002	0.200	0.18	1.42	–
					Subsoil 1	100	L	0	0.11	0.002	0.200	0.54	0.98	–
					Subsoil 2	101	L	0	0.02	0.002	0.200	1.30	0.85	–
					Subsoil+	102	L	0	0.01	0.002	0.200	1.20	0.74	–
Black spruce upland	61°38'15.10"N 121°23'51.20"W	0.17	1	–1.6 °C	Forest floor	10–15 [e]	O	0.880	0.47	1	0.500	0.03	1	1
					A	2	LS	0.028	72.29	1	0.005	0.26	1	1
					B	15	SL	0.010	73.07	1	0.005	0.33	1	1
					C1	25	SL	0.002	7.92	0.200	1	0.4	1.07	1
					C2	30	SL	0.001	2.20	0.200	1	0.03	1.06	1
					Subsoil+	100	S	0	2.63	0.200	1	0.57	1	1
Black spruce peatland	61°38'13.20"N 121°23'51.00"W	0.17	1	–1.6 °C	Forest floor	10	O	1	0.30	1	0.500	0.03	1	1
					Oh	10	O	1	0.30	1	0.005	0.03	1	1
					Ohz	40	O	1	0.30	1	0.005	0.03	1	1
					C1	80	O	1	0.29	0.200	1	0.03	1.07	1
					C2	100	Si	0	0.00	0.200	1	0.89	1.06	1
					Subsoil+	150	Si	0	0.00	0.200	1	1.38	1	1

[a] S = sand, LS = loamy sand, SL = sandy loam, SCL = sandy clay loam, L = loam, Si = Silt, O = organic.

[b] adjustments to default modeled thermal conductivities (multipliers).

[c] limit of the ice/(water + ice) fraction for each soil layer.

[d] forest floor thickness = 3 cm; reindeer moss thickness = 9 cm in summer, permeated by snow in winter.

[e] forest floor thickness = 10 cm; moss layer on top = 5 cm.

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weather station at Fort Simpson (61°27'N, 121°01'W), from November 1963 to December 2010, to drive the model calculations. The SOG data served to further calibrate and verify the snowpack and snowmelt calculations. The model calibrations involved the following considerations and adjustments (Table I; Balland, Pollacco & Arp, 2008):

- Hydro-thermal parameters: The default values for soil bulk density, field capacity, permanent wilting point, thermal conductivity, heat capacity, and soil permeability at saturation as described by Balland & Arp (2005) and Jutras and Arp (2011) remained untouched except for conductivity adjustments noted in Table I. The largest adjustments needed to be made for downward flow (infiltration) for the aspen site (a reduction by 500), lateral flow (interflow) for the A and B layers of the jack pine and black spruce sites (a reduction by 200), and lateral flow for the subsoil of the jack pine site (an increase by 20). The reasons for these changes vary. For example, in the more porous top soil, lateral flow would only occur when the subsoil is saturated. The actual amount of "net lateral flow loss" is subject to calibration, to account for changes in terrain, slopes, and depression along the lateral flow path. All flows are assumed to follow Darcy's law, using average length and overall elevational drops along vertical and lateral flow-paths for determining the associated hydraulic gradients.
- Canopy transparency: In northern forests, foliage is generally clustered in narrow black spruce crowns, in bunches along jack pine branches, and individually along aspen twigs. As a result, there is light reaching the ground. Effective light interception was therefore adjusted to improve the model calculations for soil temperature.
- Fraction of light reflected from the ground surface (albedo): Moss carpets tend be more reflective (albedo = 0.25) than forest floors without moss cover (albedo = 0.12).
- Reducing the near-ground air temperatures for depressed areas: Wet ground tends to be cooler than dry ground for at least 2 reasons: evaporative cooling during summer and influx of cool air into depressed areas. For the poorly drained sites of this study, this effect was simulated by reducing the air temperature by 1.6 °C.
- Adjusting snowfall amounts for deciduous (aspen) sites: The snowpack simulations required that the snow input for aspen be increased by a factor of 1.25 due to reduced snow interception and sublimation of deciduous canopy.
- Adjusting the thermal conductivities: For sandy subsoils, thermal conduction in the topsoil was adjusted to be slightly higher than the default value. For subsoil, thermal conductivities were estimated to drop to one half of the default value. This could be due to a difference in sand mineralogy (*e.g.*, less crystalline, less quartz) or a difference in moisture (drier than calculated). The last possibility would be more plausible than the former, because the upper soil horizons would likely receive upward capillary flow during summer and winter, thereby not only reducing the moisture content of the subsoil (see below), but also lowering the ability of that subsoil to conduct heat.
- Extent of soil freezing: The extent of soil moisture freezing was restricted in proportion to the clay and organic matter fraction of the soil, and freezing would not occur once the unfrozen soil moisture content had fallen below the permanent wilting point.
- Upward capillary flow: An accommodation was made to allow for upward soil moisture flow when the upper soil layers become dry during summer and icy during winter, especially when the subsoil is saturated, as would be the case for the depressed black spruce site.

For each site, the calibrated model calculations started in September 1963 with a hypothetical non-frozen ground condition, to determine whether and how fast a permafrost layer would develop from this condition, when it would stabilize, and how sensitive it would be to changing temperature, snow, and site conditions as affected by vegetation type, soil drainage, and snowpack depth. Each particular soil layer was assumed to be isotropic with respect to heat and water conduction and retention, and to not change in density, regardless of freezing and thawing. The snowpack layer was also considered to be isotropic, but its density would gradually increase towards spring melt, and its thermal conductivity would change correspondingly.

Results

At Fort Simpson, mean January temperatures have increased since 1963 at a more rapid rate than mean July and annual temperatures based on the trend equations in Table II (Figure 4). The equations in Table II imply that the January temperature increased on average by 7.0 °C over the course of 50 y, with corresponding mean July and annual temperature increases amounting to 1.9 and 2.6 °C, respectively. According to Table II and Figure 4, annual amounts of rain, snow, and total precipitation received also increased, by an average by 77, 93, and 170 mm, respectively.

There are good agreements between actual and modeled maximum daily, weekly, monthly snowpack depth from 1982 to 2014 (Table II; Figure 5). By year, there is no significant trend for increasing snowpack depth. Actual snowpack depth recording started in 1981. The time series

TABLE II. Precipitation, air temperature and maximum snowpack trends at Fort Simpson, 1964 to 2014, and actual *versus* ForHyM-modelled maximum snowpack depth.

Trends [a]	R^2	*p*-value	*t*-value
$T_{january}$ (°C) = 0.139*t* – 28.9	0.17	0.003	3.18
T_{july} (°C) = 0.037*t* + 16.2	0.17	0.003	3.14
$T_{average}$ (°C) = 0.052*t* – 4.5	0.39	<0.0001	5.58
Rain (mm) = 1.54*t* + 181.9	0.12	0.016	2.50
Snow (mm) = 1.85*t* + 128.4	0.19	0.001	3.35
Annual precipitation (mm) = 3.39*t* + 310.3	0.26	0.000	4.10
Maximum snowpack depths	R^2	*p*-value	*t*-value
Yearly = 0.94*x*	0.09	<0.0001	21.09
Monthly = 0.86*x*	0.77	<0.0001	52.08
Weekly = 0.85*x*	0.75	<0.0001	91.07
Daily = 0.86*x*	0.74	<0.0001	226.63

[a] *t*: years since 1964; *x*: modelled maximum snow depth (cm).

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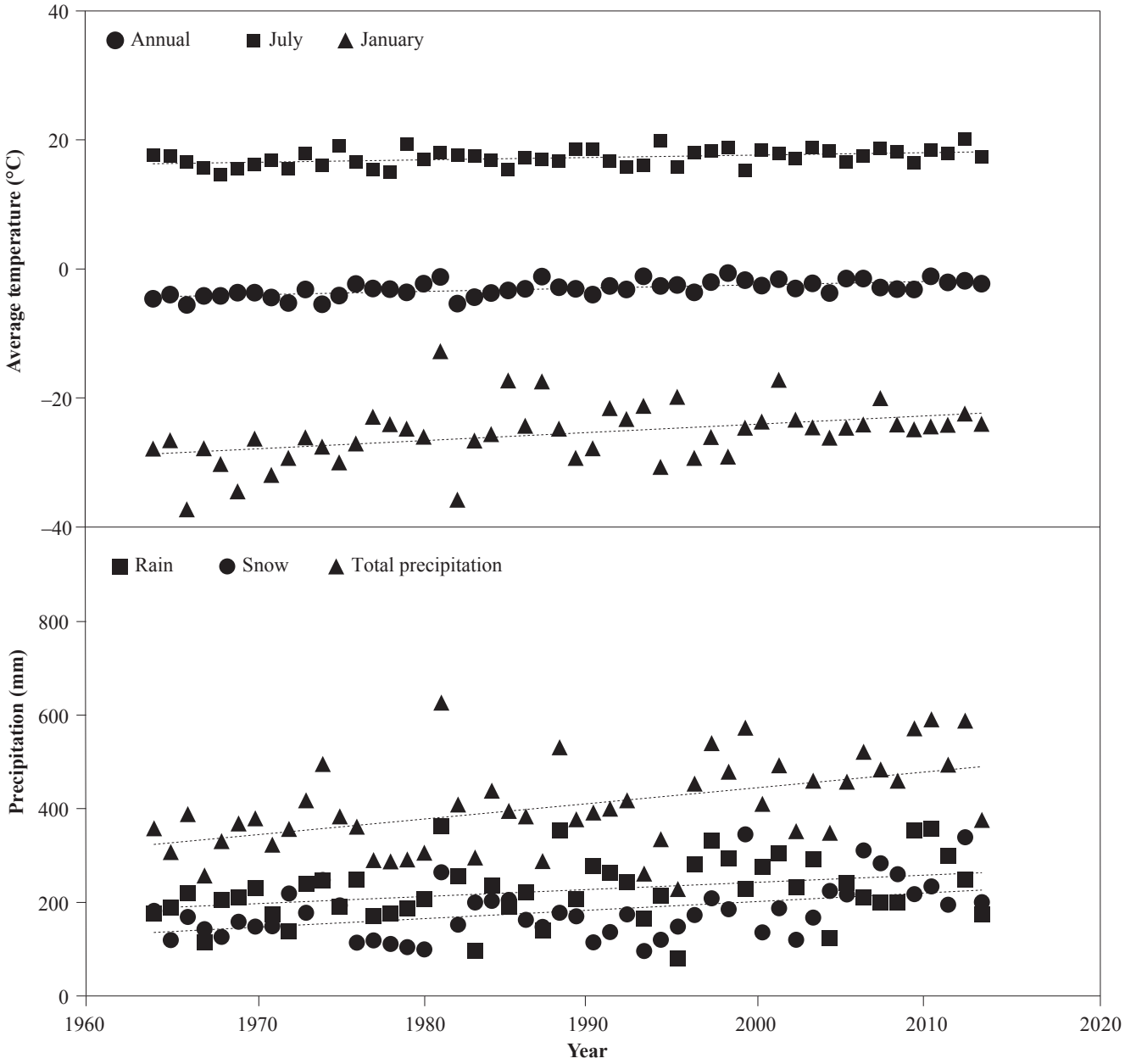


FIGURE 4. Fort Simpson weather records. Top: mean January, July, and annual air temperature. Bottom: Annual rain, snow, and total precipitation amounts, 1964–2013.

for daily air temperature and daily modeled snowpack depth are overlaid in Figure 6 on the daily snowpack depth records as modeled from 1964 forward and in comparison with the actual records from 1984 forward. These tracks are in good agreement in some years and not so good in other years. The reasons for this likely relates to variations in the snowpack data due to variations in wind, as these determine how much snow accumulates or is blown away at any particular location. Other wind-related factors include the extent of snow sublimation and of snow cohesion, as affected by rainfall and the duration of above-zero surface temperatures during the course of each winter.

Soil temperature: Figure 7 compares the modeled soil temperature track with the actual soil temperature data for the upland black spruce and the peat plateau sites at 5, 25, and 50 cm depth. In general, there is good agreement, but the simulated tracks at 5 cm depth are (*i*) somewhat more variable, (*ii*) somewhat cooler for the peat plateau, and (*iii)* somewhat warmer during winter at the upland black spruce site. The greater variability of the modeled 5-cm temperature tracks can be reduced by smoothing the daily air temperature input. The cooler or warmer estimates for that depth likely arise from the propagation of errors that accrue from assuming that (*i*) the snowpack and soil layers are uniform, (*ii*) snow and soil layer thickness, density, texture, organic matter, and coarse fragment content are all properly represented, and (*iii*) the hydrological processes as formulated are sufficient to correctly model the flow and retention of water and heat through the snow and soil layers.

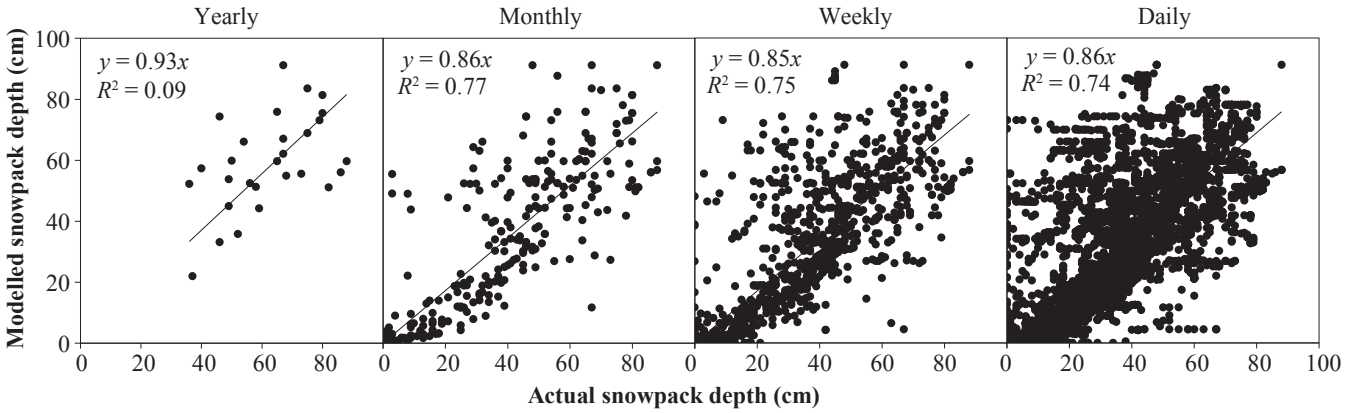


FIGURE 5. Plots of actual *versus* ForHyM-modeled maximum snowpack depth by year, month, week, and day, since 1981/82.

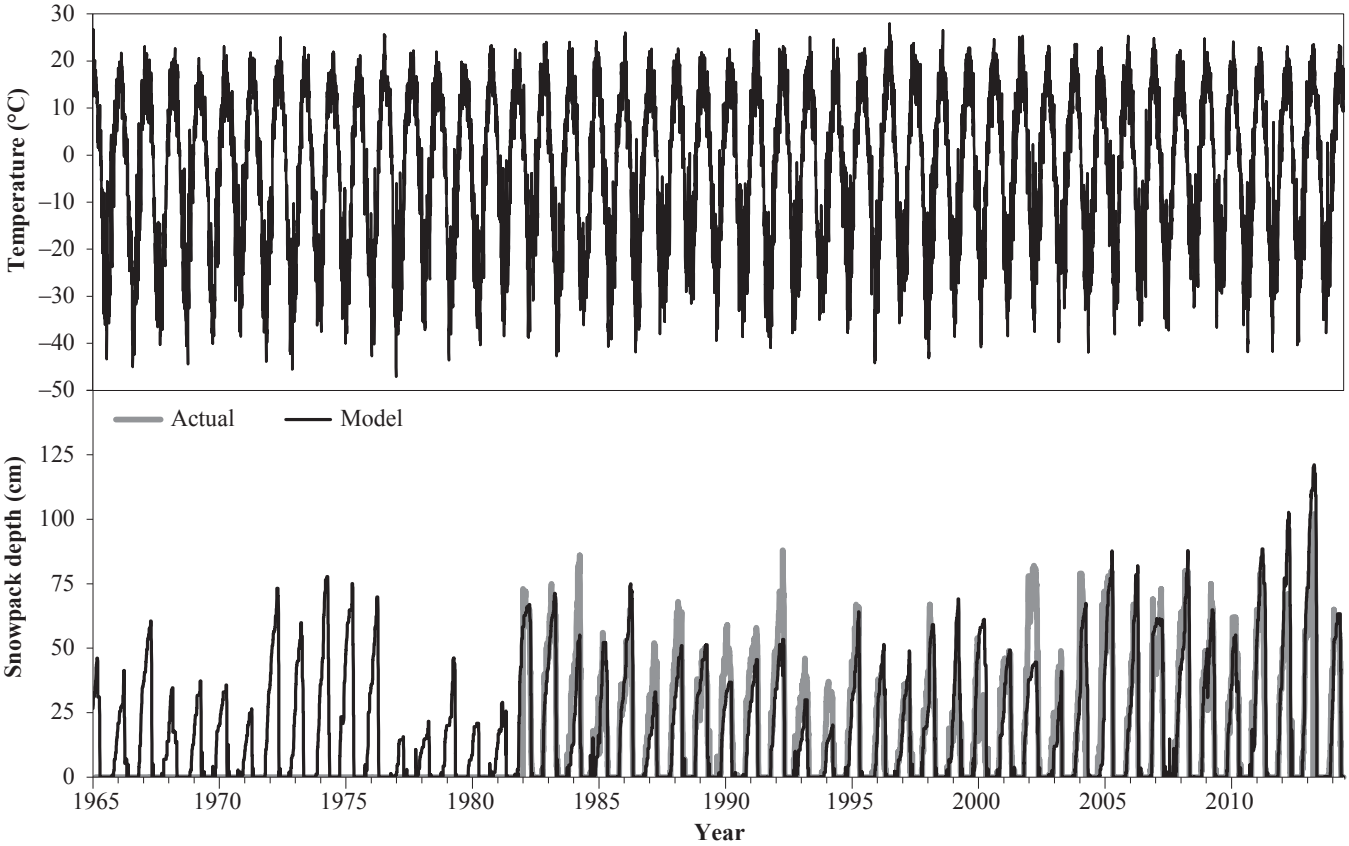


FIGURE 6. Daily air temperature (top) and snowpack depth (actual and ForHyM-simulated; bottom) for Fort Simpson from 1965 to 2013.

Frozen *versus* unfrozen soil moisture: Following the snow and soil temperature calibrations for the 5 vegetation conditions (upland jack pine, aspen, black spruce, wet peatlands, and poorly drained peatlands), ForHyM produced the annual patterns of frozen *versus* non-frozen soil moisture in Figure 8. On the upland sites, the simulation showed that freeze and thaw cycles would be discontinuous. For the peatlands, continuous frost would persist longer with decreasing drainage. For all site conditions, the annual extent of frozen soil moisture would follow the annual air temperature pattern, and the extent of soil frost would decrease if the current trend towards higher air temperatures were to persist.

Upper and lower boundaries of the frozen soil: The temperature derivations for the snowpack, forest floor, and mineral soil layers were used to determine the changes in the depth of the upper and lower boundaries of the frozen soil. The example for the poorly drained black spruce wetland in Figure 9 indicates how these depths would vary from year to year: The lower frost border would have stabilized at a depth of about 130 cm over a period of 35 y, while the upper border would have been subject to annual thawing down to about 50 to 60 cm, *i.e.*, not too different from reported thawing depths for the general areas of the South Mackenzie Plain.

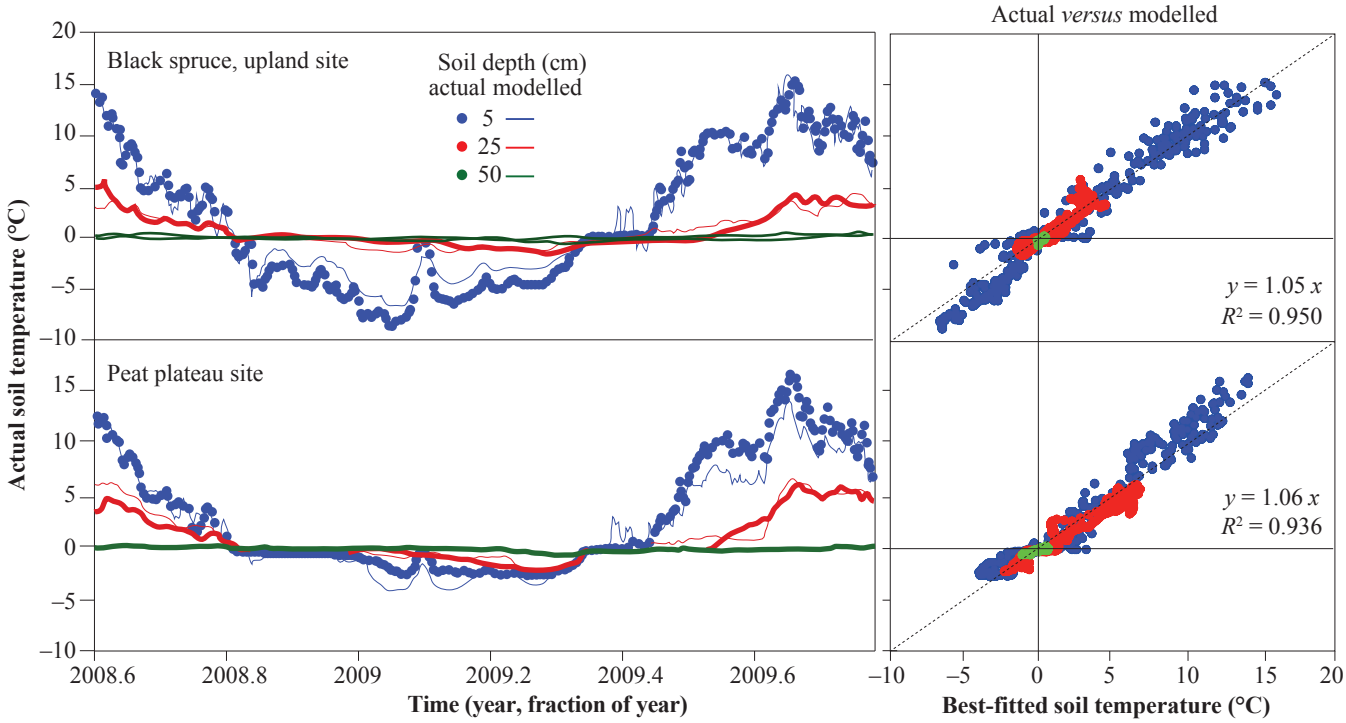


FIGURE 7. Actual best-fitted soil temperature tracks and scatter plots over the course of a year for a black spruce upland (top) and a peat plateau site (bottom).

Freeze–thaw cycles: Figure 10 (left) displays the depth of soil freezing and thawing by upland (jack pine, aspen, black spruce) and wetland (black spruce) site conditions. Here, the upland features display annually discontinuous freezing and thawing cycles. These cycles would be particularly quick and deep for the jack pine site due to high heat conduction through the sandy substrate, while the extent of water retention and freezing within this substrate would be low. For the upland aspen and black spruce sites, the annual freezing and thawing cycles would also be complete, but not as deep in comparison with the jack pine site on account of higher moisture retention and lower rates of heat conduction due to the higher clay content. On the water-saturated black spruce site, however, permafrost would have developed as modeled on account of the ice-producing latent heat loss and the time it takes to compensate for this loss through vertical heat inputs alone.

Figure 10 reveals a strong impact of the snowpack depth on soil freezing and thawing. Typically, the deeper the snowpack, the more insulated the soil is against permafrost development. On the upland sites, the freeze–thaw cycles deepen by a factor of about 3 as the snowpack depth is reduced by half, with 2 episodes of continued soil frost prior to 1980. The absence of snowpack on, *e.g.*, wind-swept upland barrens or open roads (ice roads) would freeze the subsoil permanently to a depth of 18 m, while the upper soil layers would experience fairly deep thawing cycles of about 2 m year after year (Figure 10, middle). On poorly drained wetlands, the deepening of the permafrost layer and of the thawing cycle due to decreasing snow depth would be narrower and more gradual but also be more persistent due to the latent heat requirements for freezing and thawing (Figure 10, right). The amount and duration of snow cover therefore affects the extent of soil heat gain and loss and hence the resulting temperature profile, the extent of soil freezing and thawing, and the development or abatement of permafrost in a major way.

The combined thickness of the forest floor and the moss and lichens layer on top affords additional insulation against soil freezing, thawing, and permafrost development. As a result, increasing the forest floor depth would lead to a gradual decrease in depth and duration of the permafrost layer, while the depth of thawing would increase until the soil frost becomes discontinuous, as demonstrated in Figure 10 (bottom right) for the poorly drained black spruce wetland site.

Mean annual air temperature effects on permafrost conditions: Also entered on Figure 10 are the changes in mean annual air temperature required to create conditions that would not have led to permafrost development from the hypothetical unfrozen soil condition prior to 1964. For example, the mean annual temperatures would have had to be 5 to 6 °C higher than the actual values to have no upland or wetland permafrost condition with and without snow cover within the general Fort Simpson area. With the recorded or modeled maximum snowpack conditions, permafrost would not have developed on the wetlands if the mean annual temperature had been 2 to 3 °C higher than recorded. For the uplands, the freezing and thawing cycles in Figure 10 would still be present, but not as deep. The extra insulation provided by a 30- rather than 10-cm deep organic layer is equivalent to raising the annual air temperature by 1 to 2 °C.

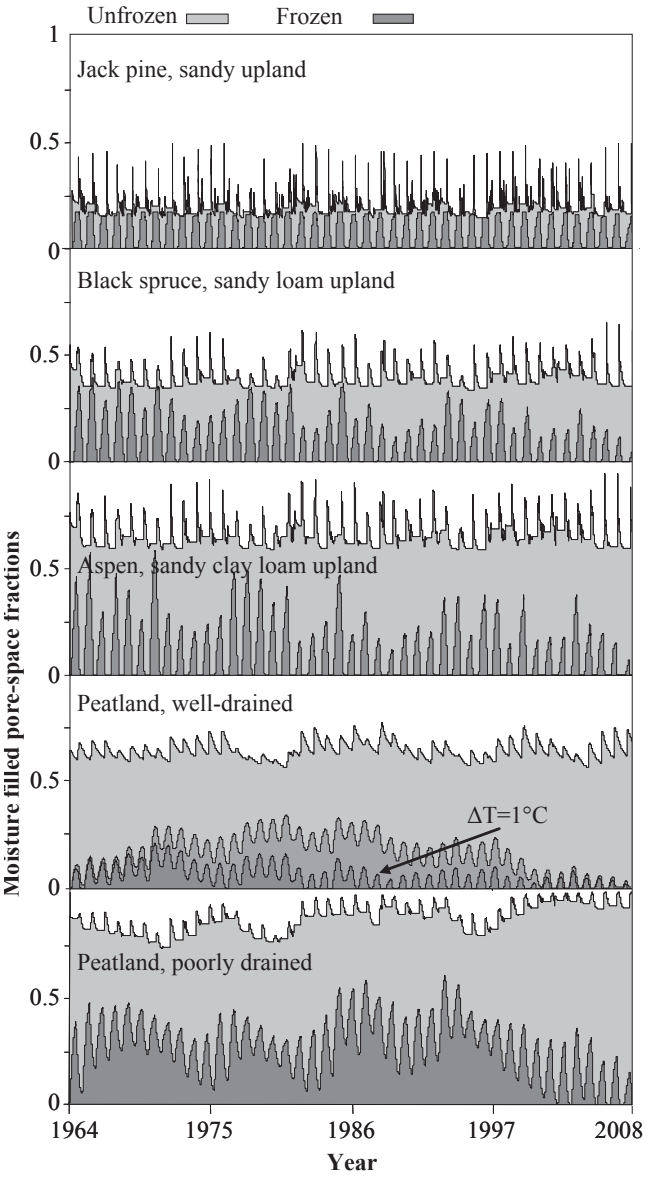


FIGURE 8. Modeled frozen and unfrozen water content (1964–2008) by site type for the Fort Simpson area, starting with the hypothetical no-permafrost condition in September 1963.

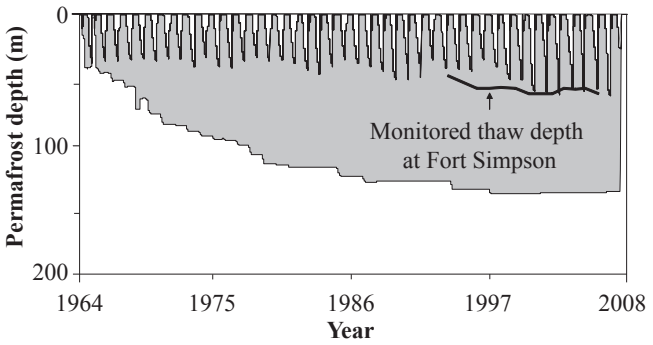


FIGURE 9. Modeled permafrost depth (upper and lower boundaries; black spruce wetland; 1964–2008, starting with the hypothetical no-permafrost condition in September 1963) with actual thaw depth observations for the Fort Simpson area (similar but deeper thaw depths have been reported and discussed by Smith *et al.*, 2009).

Discussion

The ForHyM-generated output for the tracks regarding frozen and unfrozen soil moisture reveals how the variations in surface vegetation affect the hydrothermal soil regime and hence affect permafrost development and duration (Figure 10). These modeled variations are due to differences in soil insulation: this insulation would be largest during winter where the forest canopies enable deep snowpack accumulations, and where the combination of moss cover and accumulated forest litter would enable further insulation during summer and winter. Also, where the ground is well drained, permafrost would not develop, but the annual freezing and thawing cycles could be very deep, especially on sandy sites underneath jack pine cover (Figure 10). In contrast, soils that are poorly drained would develop permafrost, but, depending on location and according to the ForHyM calculation, that permafrost layer, once formed, would become increasingly susceptible to collapse, and permanently so, on account of steadily increasing air temperatures, as recorded since 1964 and as quantified in Table II. Areas where the permafrost is most persistent would be where the ground is barren and windswept, as demonstrated in Figure 10. For both open upland and wetland conditions, annual air temperature would have to rise by about 5 to 6 °C to prevent the formation of permafrost pockets. Since the hydrothermal regimes are very sensitive to local variations in forest type and snow and soil conditions, it is important to know how these differences may affect the preservation of the permafrost layer under the currently changing weather and climate conditions. As calculated, snow, rain, and light interception at canopy and at ground level, extent of evaporative soil moisture losses, soil drainage, texture, density, organic matter and coarse fragment content, snowpack depth, and albedo all affect the water and heat flow and retention through soil (see also Peck & O'Neill, 1995; Stieglitz *et al.*, 2003; Hansson & Lundin, 2006).

Due to the combination of site and soil properties, it would appear that jack pine sites would be the least thermally buffered of the vegetated areas, being warmest in summer and coldest during winter, but still without permafrost. In contrast, poorly drained wetland locations would have the least summer to winter variations, but would also be the most frozen, and would maintain their permafrost condition the longest. In all cases, vegetation and ground cover have a stabilizing influence on the hydrothermal soil conditions, with mature evergreen conifer forests having particularly deep perennial canopies and thick moss carpets (Beringer *et al.*, 2001; Chambers & Chapin, 2003).

One aspect of the above hydrothermal calculations is the overall interdependency of soil permeability and thermal conductivity on the empirical formulations that functionally relate these variables to snow and soil density, texture, organic matter content, coarse fragment content, and the continuous variations in soil moisture by soil layer type and soil depth. This functionality would constrain much of the inter-parametric correlations that would arise by setting all hydrothermally affecting parameters

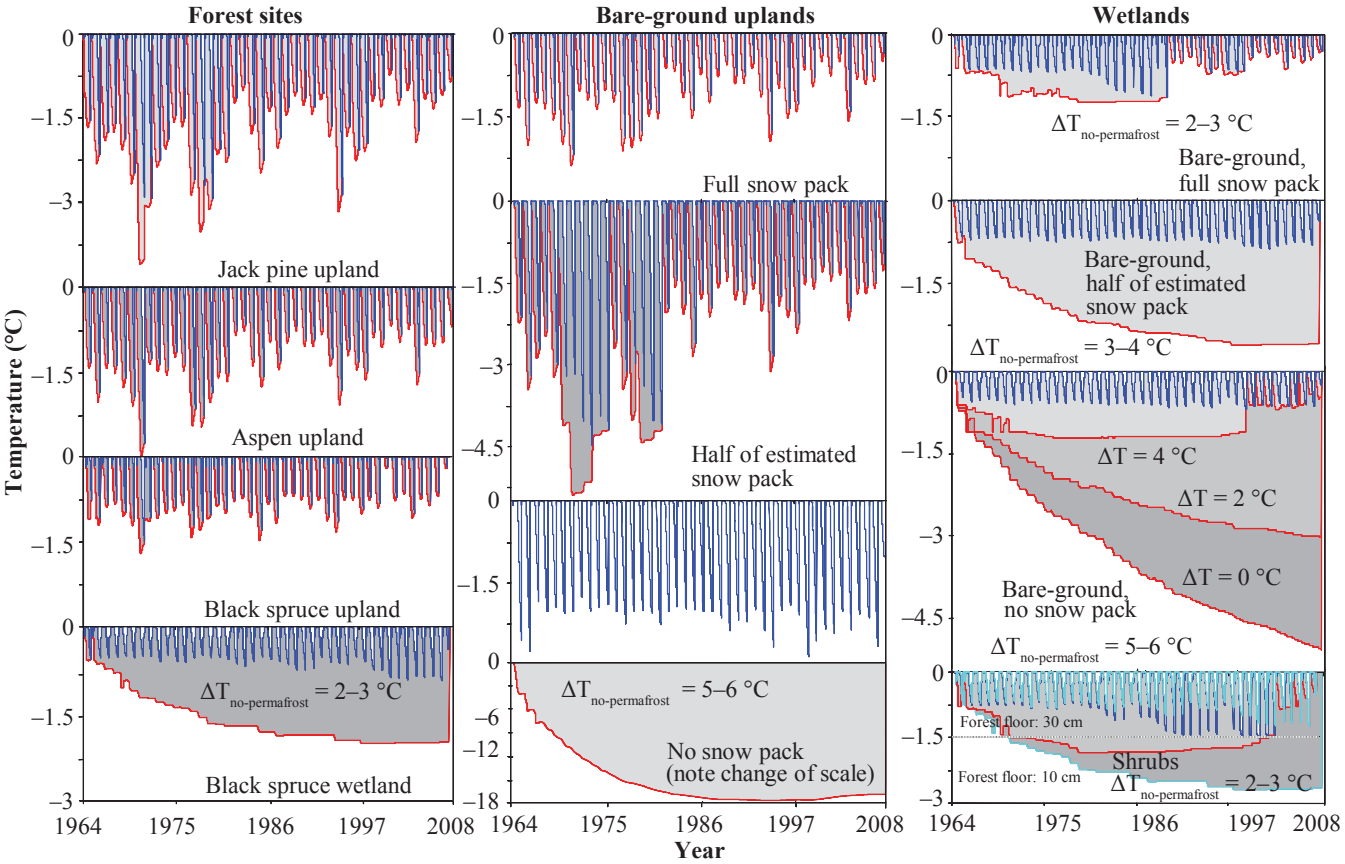


FIGURE 10. Upper (blue) and lower (red) frost boundaries within the soil for 4 forest site conditions (left), bare-ground uplands (middle), and bare-ground wetlands (right), with full, half, and zero ForHyM-estimated snow pack accumulations for the bare-ground simulations, starting from a hypothetical no-permafrost condition in September 1963. The plots on the bottom right illustrate the changes of the permafrost layer and frost and thawing cycles with varying forest floor depth (30 cm, red and dark blue; 10 cm, light blue). The ΔT entries refer to the required rise of mean annual air temperatures to lead to a full collapse of the ForHyM-simulated permafrost condition, where applicable.

independently of one another (Hansson & Lundin, 2006). The default values as derived by Balland, Pollacco, and Arp (2008) for the BOREAS jack pine, aspen, and black spruce locations should therefore provide a reasonable starting point for the calculations, based on the generally good agreement between the actual data and modeled output for snowpack depth, soil temperature, and thawing depth. To further improve on the model formulation and calibrations, additional on-site data would need to be obtained, such as, *e.g.*, actual daily heat fluxes into, through, and from the soil as affected by weather and season, frozen and unfrozen soil moisture content by layer, snow density, soil density, on-site thawing-freezing depths, and local stream discharge.

While the above-generated model output assumes uniform conditions within and across each site by snow and soil layer, such uniformity does not exist as a rule on account of many within-site variations, including topography, which varies from mounds to pits and across landscapes from flat to rolling and hummocky. Nevertheless, the assumption of uniformity, which includes even slopes per site, still leads to reasonable agreements between the calibrated model output and field-determined tracks and profiles (Figures 6–8) for snow pack depth, soil temperature, and thawing depth. Similar results were obtained for the jack pine, black spruce, and aspen sites at the BOREAS locations, where the site-calibrated ForHyM calculations agreed well with the data tracks for soil temperature and soil moisture (Balland, Pollacco & Arp, 2008). Complications, however, would arise where, *e.g.*, the influence of weather and variations in surface and subsoil conditions induced strong variations within particular sites and stands. Examples include stands influenced by seepage, contrasting forest cover (including the lack thereof), the presence of dissecting roads or seismic lines, the formation of ice lenses and related frost heaving, and irregular soil inclusions such as boulders. For such cases, it would be best to consider adding additional constraints on the calculations, *e.g.*, adding heat and water flows on account of seepage, and re-formulating the boundary conditions and the related calculations for the three-dimensional context. In particular, seepage would add water and heat to subsoils, thereby reducing or eliminating the incidence of continuous frost. Ice lenses, once formed, would grow, thereby forcing the surrounding soil to become drier, leading to development of a three-dimensional water and heat flow envelope around the ice lenses. Open corridors through the stand would reduce the insulation along these corridors, thereby promoting deep frost penetration into the ground along the corridors, where the lateral flow across the corridors would subsequently be blocked. This would increase the rate and

deepening of ice formation upslope of the corridors. In all cases, thawing of soil and permafrost would occur more quickly from the top downward than from the bottom upward, in direct reflection of the generally steep temperature gradient on top and the gradual but steady geothermal temperature gradient below.

The model results in Figures 6 to 10 are in general agreement with measured and modeled freeze–thaw results in the literature. For example, Peck and O'Neill (1997) observed and discussed the role of increasing soil moisture content and of sand inclusions on modeled freezing and thawing depths in silty soil. Stieglitz *et al.* (2003) modeled and discussed the role of increasing air temperature and precipitation, including snowpack depth, and related data variations on arctic permafrost conditions. Koenig *et al.* (1997) modeled and discussed snowpack depth and depth of soil frost based on detailed records for precipitation, air temperature, and heat and water fluxes.

Conclusion

The hydrothermal calculations presented above reveal that the permafrost conditions within the general Fort Simpson area are highly sensitive to changing air temperatures and changing amounts of on-the-ground snow: more snow means less freezing and more thawing, especially on sites underlain by permafrost. From soil to soil, the depth of frost penetration and thawing would be further affected by the extent of forest vegetation, the thickness of low-density moss, lichens, and forest floor layers, and the extent of soil drainage of the organic and mineral soil matrices. Low moisture content within the soil would enhance the depth of the annual freezing and thawing cycles, while high moisture levels on, *e.g.*, the peat plateaus and poorly drained wetlands retard this, thereby sustaining the permafrost condition at these locations for longer periods of time. According to the ForHyM calculations, the recent trend of increased air temperatures will accelerate permafrost degradation throughout the area, especially where collapse scars are now spreading into the most poorly drained wetlands. The above computational methods are therefore useful in (*i*) quantifying already existing and changing permafrost conditions, (*ii*) projecting how these conditions may be affected by changes in climate and by vegetative disturbance regimes such as forest fires and land-use practices associated with forestry, mineral exploitation, and road and corridor construction. Additional work is required in (*i*) further calibrating and verifying the above calculations based on local soil moisture and temperature measurements and actual soil profiles, as was done for the BOREAS sites in Saskatchewan (Balland *et al.*, 2006), (*ii*) using heat flux determination for additional model calibrations, and (*iii*) extending the above modeling effort to other permafrost areas of interest.

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