Resilience and vulnerability of permafrost to climate change [1]

M. Torre Jorgenson, Vladimir Romanovsky, Jennifer Harden, Yuri Shur, Jonathan O'Donnell, Edward A. G. Schuur, Mikhail Kanevskiy, and Sergei Marchenko

Abstract: The resilience and vulnerability of permafrost to climate change depends on complex interactions among topography, water, soil, vegetation, and snow, which allow permafrost to persist at mean annual air temperatures (MAATs) as high as +2 °C and degrade at MAATs as low as -20°C. To assess these interactions, we compiled existing data and tested effects of varying conditions on mean annual surface temperatures (MASTs) and 2 m deep temperatures (MADTs) through modeling, Surface water had the largest effect, with water sediment temperatures being ~10 °C above MAAT. A 50% reduction in snow depth reduces MADT by 2 DC.Elevation changes between 200 and 800 m increases MAAT by up to 2,3 DC and snow depths by ~40%. Aspect caused only a ~1°C difference in MAST. Covarying vegetation structure, organic matter thickness, soil moisture. and snow depth of terrestrial ecosystems, ranging from barren silt to white spruce *(Picea glauca* (Moench) Voss) forest to tussock shrub, affect MASTs by ~6°C and MADTs by ~7°C. Groundwater at 2-7 °C greatly affects lateral and internal permafrost thawing. Analyses show that vegetation succession provides strong negative feedbacks that make permafrost resilient to even large increases in air temperatures. Surface water, which is affected by topography and ground ice, provides even stronger negative feedbacks that make permafrost vulnerable to thawing even under cold temperatures.

Résumé : La résilience et la vulnérabilité du pergélisol face aux changements climatiques dépendent d'interactions complexes entre la topographie, l'eau, le sol, la végétation et la neige qui permettent au pergélisol de se maintenir à des températures moyennes annuelles de l'air (TMAA) aussi élevées que +2 °C et de se dégrader à des TMAA aussi basses que –20 °C. Pour évaluer ces interactions, nous avons compilé des données existantes et testé les effets de diverses conditions de température moyenne annuelle en surface (TMAS) et à une profondeur de 2 m (TMAP) par l'entremise de la modélisation. L'eau de surface avait l'effet le plus prononcé alors que la température à l'interface entre l'eau et les sédiments était ~10 °C plus élevée que la TMAA. Une diminution de l'épaisseur de la couche de neige de 50 % réduit la TMAP de 2 °C. Un changement d'altitude entre 200 et 800 m augmente la TMAA de 2,3 °C et l'épaisseur de la couche de neige de ~40 %. L'orientation cause une différence de seulement ~1 °C de la TMAS. La structure de la végétation, l'épaisseur de la matière organique, l'humidité du sol et l'épaisseur de la couche de neige qui covarient dans les écosystèmes terrestres, allant de la toundra limoneuse à la forêt d'épinette blanche (*Picea glauca* (Moench) Voss) et à la zone de transition entre les buttes de gazon et les arbustes, affectent la TMAS de ~6 °C et la TMAP de ~7 °C. À une température de 2–7 °C, l'eau souterraine influence grandement la fonte latérale et interne du pergélisol. Les analyses montrent que la succession de la végétation engendre de fortes rétroactions négatives qui rendent le pergélisol résistant à une augmentation même importante de la température de l'air. L'eau de surface qui est influencée par la topographie et la glace au sol engendre des rétroactions négatives encore plus prononcées qui rendent le pergélisol vulnérable à la fonte même si la température est froide.

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Introduction

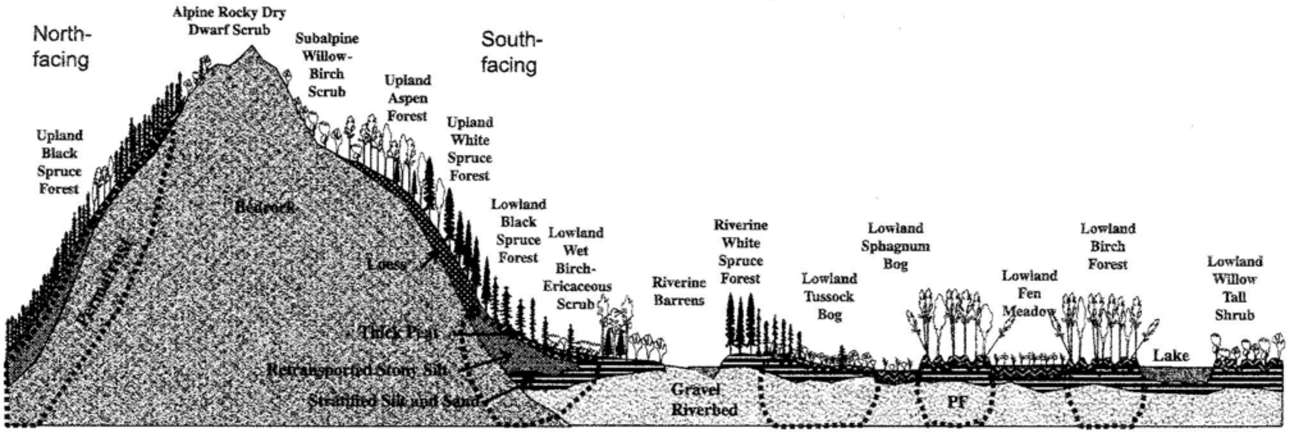
Perennially frozen ground (permafrost) is a unique characteristic of polar regions and high mountains and is fundamental to geomorphic processes and ecological development in tundra and boreal forests. Because permafrost impedes drainage and ice-rich permafrost settles upon thawing, the degradation of the permafrost in response to climate change will have large consequences to boreal ecosystems (Jorgenson et al. 2001; Camill 2005: Jorgenson and Osterkamp 2005). Thawing permafrost affects surface hydrology by impounding water in subsiding areas and enhances drainage of upland areas (Woo 1990). Consequently, changes in soil drainage alter the degradation and accumulation of soil carbon (Schuur et al. 2008), habitats for vegetation and wildlife (Jorgenson and Osterkamp 2005), and emissions of greenhouse gases (Christensen et al. 2004; Turetsky et al. 2007). The consequences will range from microsite changes in hydrology and vegetation to possible global contribution of greenhouse gases (Walter et al. 2006; Schuur et al. 2009). Yet, we still have an incomplete understanding of the complex interactions among factors and feedbacks that allow permafrost to persist at mean annual air temperatures (MAATs) up to +2º under late-successional ecosystems and degrade into thermokarst lakes and thaw slumps at MAATs as low as -20 ºC in the presence of surface water (Grom and Pollard 2008). To predict the responses of permafrost to climate change, it is imperative, then, to better quantify how negative feedbacks can contribute to the ability of permafrost, and its affected ecosystems, to make minor adjustments and recover from surface perturbations and thus be resilient to changes. Equally important is analyzing how positive feedbacks make permafrost more vulnerable to change and lead to radical shifts in physical structure and long-term ecological reorganization.

Permafrost-affected regions cover -80% of Alaska (Jorgenson et al. *2008a, 2008b)* and -22% of the exposed land in the Northern Hemisphere (Brown et al. 1997), and its stability is particularly important to the fate of boreal ecosystems in Russia, Canada, and the United States. A substantial portion of the boreal biome is within the discontinuous (50%-90% of area) and sporadic (10%-50%) permafrost zones where MAATs are approximately -6 to -2°C and -2 to 0 ºC, respectively. The complex interaction of climatic and ecological processes in permafrost formation and degradation is closely associated with these permafrost zones (Shur and Jorgenson 2007). *Climate-driven* permafrost occurs in cold climates regardless of vegetation and soil conditions. *Climate-driven, ecosystem-protected* permafrost, particularly extremely ice-rich Pleistocene permafrost. which formed initially as climate-driven permafrost, can persist in the discontinuous zone during warming climates for long periods because of the protection of organic soil horizons. *Ecosystem-driven* permafrost forms in the discontinuous zone in poorly drained, low-lying, and north-facing landscape conditions under the strong influence of ecological processes (Fig. 1). Finally, *ecosystem-protected* permafrost persists as sporadic patches under warmer climates but cannot be reestablished after disturbance.

Permafrost has long been responding to climatic change and most permafrost is of Pleistocene age (Osterkamp 2007). Within the discontinuous zone, Holocene age permafrost is abundant. including permafrost formed during the Little Ice Age (Jorgenson et al. 2001). The -20 ºC increase in air temperature at the Pleistocene-Holocene transition, as indicated by analysis of oxygen isotopes in a Greenland ice core (Arctic Climate Impact Assessment 2005), led to a large reduction in the permafrost region; ice-wedge casts have been found as far south as Iowa in North America. Permafrost degradation during this transition period caused the formation of the contemporary discontinuous zone where permafrost previously was continuous, the formation of taliks (thawed zones between the seasonally frozen surface and permafrost) under lakes and rivers. and the reduction of permafrost thickness by thawing of permafrost from its base. Permafrost stability thus depends on climate changes over differing periods and magnitudes, which can cause permafrost to persist over hundreds to thousands of years or to be newly formed during cold, snow-poor winters and persist only for years to decades. Temperature monitoring of deep boreholes in northern Alaska since the 1950s (Lachenbruch and Marshall 1986; Osterkamp and Romanovsky 1999; Osterkamp 2007) indicates that permafrost warmed 2-4 °C in the early to mid-20th century coming out of the Little Ice Age and warmed by up to 3 °C during the last 20 years of the 20th century. In contrast, borehole monitoring in the discontinuous zone of Alaska since the 19805 indicates that permafrost generally has warmed only 0.3-1 "C, but temperatures in this zone already are near thawing (Osterkamp and Romanovsky 1999: Osterkamp 2007).

Permafrost is not connected directly to the atmosphere because its thermal regime is mediated by topography, surface water, groundwater, soil properties, vegetation, and snow, and there are numerous interactions among these ecological components that can lead to both positive and negative feedbacks to permafrost stability (Fig. 2). Topography affects the amount of solar radiation to the soil surface, causing permafrost in the discontinuous zone to occur generally on north-facing slopes that receive less direct radiation and in flat low-lying areas where vegetation has a greater insulating affeet and where air temperatures tend to be colder during winter inversions (Viereck et al. 1986). Surface water provides an important positive feedback that enhances degradation when water is impounded in subsiding depressions (Hinzman et al. 1997). Groundwater in the active layer or within permafrost delivers heat and is often surrounded by thawed zones (Woo 1990). Soil texture affects soil moisture and thermal properties with the result that gravelly soils tend to be well drained with little difference between thermal conductivities when frozen or thawed (Shur and Jorgenson 2007). In contrast, surface organic soils, and to a lesser extent clayey and silty soils, tend to be poorly drained and have much higher thermal conductivities when frozen in winter than when unfrozen in summer. This difference leads to rapid heat loss in winter and slower heat penetration in summer and accounts for the "thermal offset" between the ground surface and the permafrost table (Burn and Smith 1988; Romanovsky and Osterkamp 1995). Vegetation has a particularly important effect in flat areas through interception of solar radiation, growth of mosses and accumulation of organic matter, and interception of snow by trees and shrubs (Viereck 1970; Jorgenson et al, 2003). Snow slows

Fig. 1. Schematic of common covarying ecosystem components across boreal ecosystems in central Alaska.



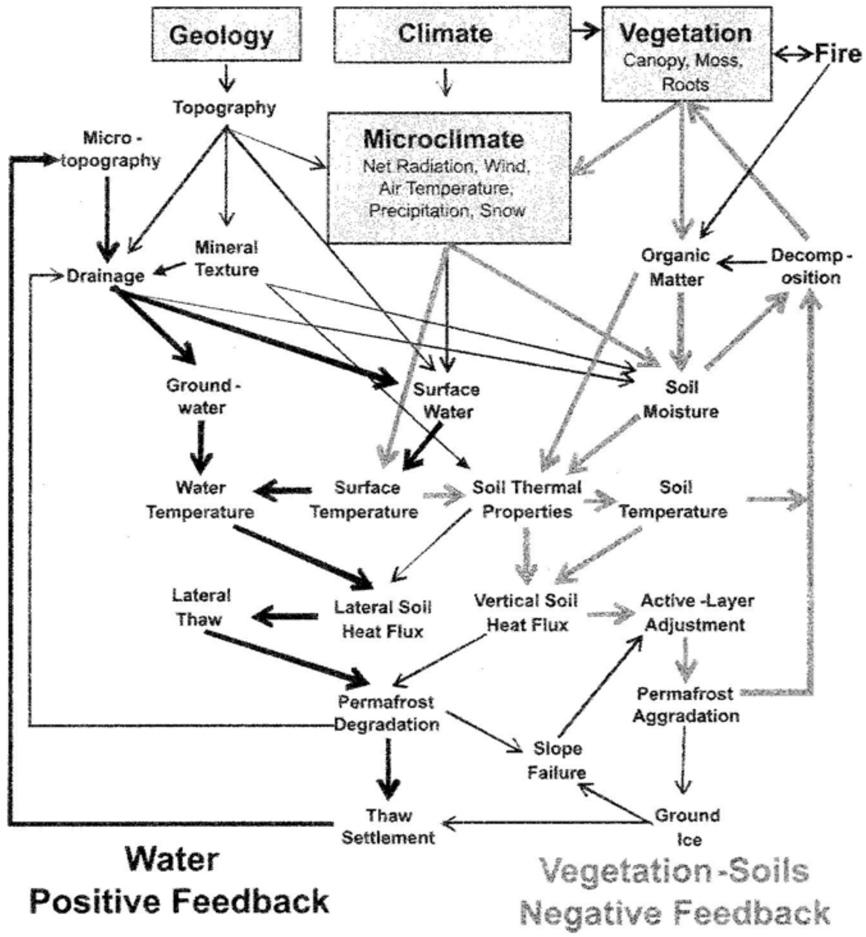
the rate of cooling in winter; thus, the seasonality (e.g., deep snow in early winter) and depth of snow are important. Because permafrost is greatly affected by these ecological components, permafrost properties evolve along with the successional patterns of ecosystem development (Viereck 1970). In turn, the patterns of ice aggradation and degradation influence the patterns of vegetation and soil development. This coevolution of permafrost and ecological characteristics at the ground surface is most evident after disturbance, such as river channel migration, lake drainage, or fire (Shur and Jorgenson 2007). The dependence of permafrost stability on both air temperatures and ecological factors greatly complicates the prediction of the consequences of climate change.

The thawing of ice-rich soils causes the surface to subside, creating depressions in the ground surface termed "thermokarst terrain". The types of thermokarst terrain and the ecological implications are extremely variable depending on climate, topography, soil texture, hydrology, and amounts and types of ground ice (Jorgenson and Osterkamp 2005). Thawing can occur downward from expansion of the active layer (seasonally thawed soil above permafrost), laterally from heat flow from surface and groundwater, internally from groundwater intrusion, and upward from the bottom due to geothermal heat flux (Fig. 3). Spatially differentiating the stages of thawing helps characterize the severity or extent of the thawing. For example, thawing of ice wedges under cold temperatures in northern Alaska typically only causes surface degradation and rarely leads to intermediate degradation that is characterized by the formation of a talik. At the next level, lateral degradation from an expanding deep thermokarst lake in northern Alaska leads to talik formation and intermediate degradation, but complete degradation of permafrost is rare because of the thickness of the permafrost In contrast, lateral degradation from a thermokarst lake in central Alaska quickly leads to complete degradation because the permafrost is thinner. These different phases of degradation can even be closely interspersed; permafrost near the shoreline of a thermokarst lake can be undergoing only transient degradation, while the permafrost under the lake has completely degraded, Common landforms include thermokarst lakes from lateral thermomechanical erosion, thaw sinks from subsurface drainage of lakes initially surrounded by permafrost, linear collapse-scar fens associated with shallow groundwater movement, round, isolated collapse-scar bogs from slow lateral degradation, and irregular thermokarst mounds from thawing of ice-poor silty soils (Jorgenson and Osterkamp 2005). Thawing of rocky, ice-poor permafrost does not create distinctive landforms and is difficult to detect.

Disturbances of vegetation and soil by human activity and wildfire are important factors contributing to permafrost degradation. The removal of vegetation by fire or human activities leads to an increase in the active layer and degradation of permafrost, which in many cases can be irreversible. Mining, oil development, settlements, logging, and farming that have become widespread since the beginning of the 20th century have been important drivers of localized degradation of permafrost (Brown and Grave 1979). Even more widespread are the effects of fire on eliminating the vegetation canopy and moss layer, removing a portion of the surface peat, and reducing the albedo (Van Cleve and Viereck 1983; Yoshikawa et al. 2003; Randerson et al. 2006). Large wildfires in recent years have been attributed to warming and drying climates in boreal forest regions (Kasischke et al. 2010). This combination of increased wildfires and climate change greatly increases the potential for permafrost degradation in the discontinuous zone (Yoshikawa et al. 2003; Shur and Jorgenson 2007; Myers-Smith et al. 2008).

The resilience and vulnerability of permafrost to climate change is controlled to a large degree by the positive and negative feedbacks affecting the surface energy balance and ground temperatures and by the volume and structure of ground ice within the soil. Resilience is the capacity of a system to sustain its fundamental function, structure, and feedbacks when confronted with perturbations, such as unprecedented warming (Chapin et al. 2009). For permafrost, this is the capacity to maintain frozen temperatures and similar ground ice contents and morphologies. Vulnerability is the degree to which a system is likely to experience harm due to exposure and sensitivity to a specified hazard or stress and its adaptive capacity to respond to the stress (Chapin et al. 2009). For permafrost, this is the extent to which permafrost thaws vertically and laterally and how much thaw settlement occurs during thawing of ground ice. Changes in the structure and functioning of permafrost in turn affect the services that permafrost provides to boreal

Fig. 2. Conceptual model of factor interactions and feedbacks affecting the resilience of vulnerability of permafrost to climate change.



ecosystems and society, including a foundation for vegetation, human facilities, and off-road travel, regulation of surface and groundwater movement, and sequestration of soil carbon. Loss of these services to ecosystems and societies, however, can be viewed as either detrimental, as in the case of collapse of human structures, or favorable, as in the case of the creation of new waterbird habitats after the collapse of forests, depending on human values.

In this paper, we evaluate the relationships among factors affecting ground thermal regimes through empirical relationships developed from field measurements, numerical modeling, and literature review. To assess the relative importance of the various factors, we compare changes in mean annual temperatures at the surface (MASTs) and at 2 m depth (MADTs) as simple, universally comparable metrics that are directly related to permafrost thawing, We then evaluate the effects of varying terrain and ecological effects on ground temperatures to assess positive and negative effects that increase the vulnerability and resilience of permafrost to thawing, Finally, we highlight some important ecological consequences of thawing permafrost.

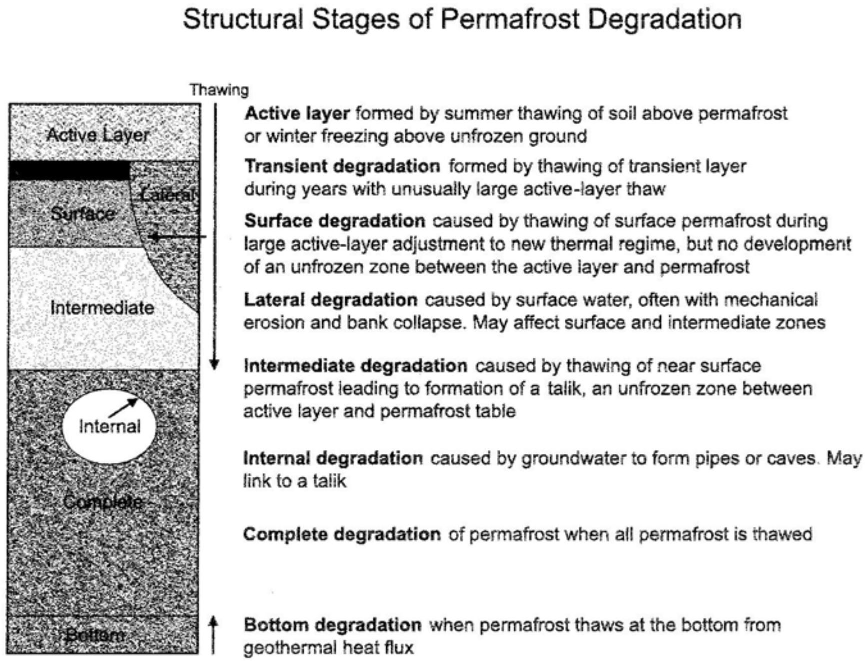
Methods

To assess the range of effects of ecosystem components on ground temperatures, we compiled and analyzed field measurements from our previous related studies, conducted numerical modeling to test effects of varying conditions on soil temperature, and summarized literature for surface and groundwater effects to allow us to integrate these factors into our analysis. For evaluation, we chose 13 ecosystem types that dominate the lowlands of the boreal region of central Alaska that are associated with a successional sequence after disturbance, such as a severe fire, from barren silt to white spruce (*Picea glauca* (Moench) Voss) forest, a paludification sequence from black spruce (*Picea mariana* (Mill.) BSP) to tussock shrub, and a permafrost degradation sequence from open water to sphagnum bog (Fig. 1: Table 1).

Field measurements

We compiled data on texture, bulk density, soil moisture, and thickness of major soil horizons from sites in interior Alaska including Delta (Harden et al. 2006; Manics et al. 2006), White Mountains near Eielson Air Force Base (Jorgenson et al. 2003), Hess Creek (O'Donnell et al. 2009a, 2009b, and unpublished data), the Bonanza Creek LTER database (Hollingsworth et al. 2008), and Healy (Schuur et al. 2009). Samples for bulk density were obtained using set-diameter soil cores for frozen and unfrozen soils or by cutting blocks and measuring their sides for very low-bulk-density

Fig. 3. Permafrost degradation is a three-dimensional process that occurs through various stages and through a variety of processes. It is useful to classify the stages to be able to use more precise terminology when referring to the extent, depth, and stage of degrading permafrost. Degradation need not include all stages.



organic samples. Organic soils were categorized into two parts: near-surface moss and fibrous horizons (live moss, Oi-fibric horizons) or deeper and more decomposed horizons (generally Oe-hemic and in some cases Oa-sapric horizons). Means and standard deviations were calculated for percent carbon, bulk density, and volumetric moisture content of each soil horizon type. While volumetric moisture content is dependent on antecedent moisture at the time of sampling, we assume that the hundreds of observations are a reasonable approximation of moisture content of a particular soil horizon. These data were used for model input described below.

Soil temperature data were collected at 39 sites in the Yukon- Tanana Uplands near Eielson Air Force Base during 1998-2001 (Jorgenson et al. 2003). Air temperatures, shallow soil (-5 cm depth). and deep soil (-1.1 m) were measured with small dataloggers (H8 and Hobo Pro, Onset Corp.). Freezing and thawing degree-days were based on degrees Celsius. At each site, topographic, soil, and vegetation data were also collected. Snow depths also were measured one to three times a winter by sampling five depths at 1 m increments along a transect across each datalogger site. Water temperatures were monitored in a shallow (0.8 m) and a deep (2 m) lake in Denali National Park by Chris Arp, US Geologica1 Survey, Anchorage, Alaska, using small waterproof temperature recorders (Onset Corp.).

Thermal conductivity of surface organic horizons relative to soil moisture content was measured on samples of organic soils taken from feather-moss- and sphagnum-dominated black spruce ecosystems. Samples were taken from three distinct horizons (live, fibrous, and amorphous) that differ with respect to physical properties and extent of decornposition. Volumetric samples were obtained by cutting soil blocks with a knife and measuring dimensions with a ruler to avoid compaction of the very low-density material, Samples were transferred into small Tupperware containers and stored in a cooler until returned to the laboratory for further analysis. In the laboratory, all samples (n = 5 per site per horizon, total = 90 samples) were saturated and allowed to equilibrate for 24 h to reach field capacity. Once samples reached field capacity. we measured thawed thermal conductivity (K_t) using a KD2 Pro Thermal Properties Analyzer (Decagon Devices, Inc., Pullman, Washinton) based on the transient line heat source method. Samples were allowed to air dry, and we took daily measurement to generate a *Kcmoisture* content relationship for each sample from each horizon type,

Ground ice data were collected near Gosling Lake, northwest Denali National Park and Preserve, and Billy Lake, northwest Wrangells National Park and Preserve (Shur and Jorgenson 2007), and Eight-mile Lake near Healy (Osterkamp et al. 2009). Samples for volumetric moisture contents were obtained with a SIPRE corer to depths of 2-3 m, measured for frozen volume, and weighed wet and after drying to calculate bulk density and moisture contents.

Modeling

To assess the effects of snow, surface, and subsurface soil properties on ground temperatures, we used the Geophysical Institute Permafrost Laboratory model, which simulates soil temperature dynamics and the depth of seasonal freezing and thawing by solving a one-dimensional nonlinear heat equation with phase change (Marchenko et al. 2008). In this model, the phase change (water to ice or ice to water) occurs within a range of temperatures below 0^{o}C. This phase

Table 1. Characteristics of boreal ecosystems including soil properties for thermal modeling.

	Vegetation structure					
	Barren	Deciduous forest	Mixed forest	White spruce forest	Black spruce forest	Burned
Dominant species		*Betula papyrifera*	*Picea glauca, Betula papyrifera*	*Picea glauca*	*Picea mariana*	*Epilobium angustifolium, Calamagrostis canadensis*
% of lowlands	1	4	5	1	48	5
% overall	9	9	6	6	26	3
Slope	Flat	Flat	Flat	Flat	Flat	Flat
Snow depth (% deviation from average)	100	80	75	70	75	100
Water table (cm) (+ above, – below)	–100	–45	–65	–70	–50	–110
Layer 1 texture	Silt	Litter, live	Litter, live	Litter, live	Litter, live	Litter-live
Layer 1 thickness (cm)	50	9	3	6	2	2
Layer 1 density ($g{\cdot}cm^{-3}$)	1.31	0.03	0.07	0.02	0.02	0.04
Layer 1 moisture (% volume)	28	2	5	3	9	16
Layer 2 texture	Silt	Fibric	Fibric	Fibric	Fibric	Fibric
Layer 2 thickness (cm)	100+	7	7	6	2	3
Layer 2 density ($g{\cdot}cm^{-3}$)	1.24	0.07	0.07	0.05	0.05	0.07
Layer 2 moisture (% volume)	41	8	5	14	19	28
Layer 3 texture		Hemic	Hemic	Hemic	Hemic	Hemic
Layer 3 thickness (cm)		5	17	17	12	9
Layer 3 density ($g{\cdot}cm^{-3}$)		0.21	0.21	0.3	0.19	0.18
Layer 3 moisture (% volume)		25	21	25	72	69
Layer 4 texture		Silt	Silt	Silt	Silt	Silt
Layer 4 thickness (cm)		39	70	52	48	92
Layer 4 density ($g{\cdot}cm^{-3}$)		1.08	1.64	1.47	0.68	0.94
Layer 4 moisture (% volume)		31	21	44	63	55
Layer 5 texture		Silt	Silt	Silt	Silt	Silt
Layer 5 thickness (cm)		100+	100+	100+	100+	100+
Layer 5 density ($g{\cdot}cm^{-3}$)		1.18	1.49	1.52	0.72	0.49
Layer 5 moisture (% volume)		44	35	46	62	56
Mean thaw depth (cm), Fairbanks area	na	110	77	75	68	110
N factor for thawing (Nt)	1.1	0.45	0.5	0.6	0.5	nd
N factor for freezing (Nf)	0.2	0.25		0.3	0.3	nd

Note: na, not applicable. *N* factor is the ratio of soil thawing degree-days to air thawing degree-days.

Tall and low willow	Low birch–ericaceous shrub	Tussock bog	Sphagnum bog	Herbaceous fen	Shallow lake	Deep lake
Salix bebbiana, Salix planifolia	*Betula nana, Ledum groenlandicum*	*Eriophorum vaginatum*	*Sphagnum* spp., *Andromeda polifolia*	*Menyanthes trifoliata, Equisetum fluviatale*	na	na
6	15	4	4	5	1	1
4	22	4	2	2	2	1
Flat	Flat	Flat	Flat	Flat	Flat	Flat
100	100	95	100	100	100	100
–35	–30	–20	–5	0	50	200
Litter, live	Litter, live	Litter, live	Litter, live	Litter, live	Water	Water
2	2	3	4	3	100	200
0.03	0.03	0.07	0.01	0.05	1	1
25	25	38	34	55	100	100
Fibric	Fibric	Fibric	Fibric	Fibric	Silt	Silt
18	13	29	5	41	100+	100+
0.08	0.37	0.11	0.05	0.07	1.24	1.24
30	55	49	51	42	41	41
Hemic	Hemic	Hemic	Fibric	Hemic		
11	12	20	97	50		
0.21	0.66	0.25	0.05	0.18		
60	53	63	80	76		
Silt	Silt	Silt	Silt	Silt		
98	37	16	100+	100+		
0.75	1.45	0.65	0.6	0.51		
50	45	55	45	45		
Silt	Silt	Silt				
100+	100+	100+				
0.4	0.23	0.8				
50	34	56				
120	62	55	na	na	na	na
0.7	0.5	0.45	0.95			
0.4	0.4	0.4	0.1			

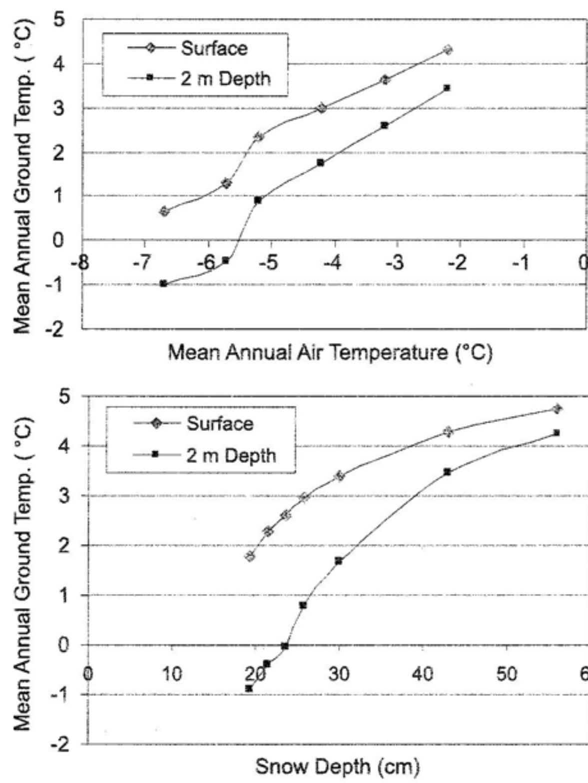
change process is characterized by an unfrozen water content curve, which is specific for each soil layer. A detailed description of this model including the explanation of the numerical solution and comparison between analytical and numerical solutions with some discussion on the stability of this solution can be found in Nicolsky et al. (2007). The model simulates soil temperature and liquid water content fields for the entire depth of consideration with daily time resolution. The depth resolution is 1 cm for the upper 1 m of soil with lower resolution for the deeper layers. This model does not include advective heat transfer processes related to water or water vapor movement in soils and this simplification may impose some limitations on the model results in non saturated coarse-grained soils, but our modeling involved cases with completely saturated fine-grained soils where advective heat transport is negligible. This model also has been successfully validated using ground temperature measurements of a very high quality (precision generally at 0.01 °C) in shallow boreholes across Alaska (Romanovsky and Osterkamp 2000; Nicolsky et al. 2009).

The model was used to test the effects of varying climatic and ecological conditions on ground temperatures. First, the model was run with air temperatures associated with varying MAATs of 0-5 °C below the recent (1997-2007) MAAT (-2.°C) for Fairbanks and varying snow depth of 50% 100% of recent mean annual snow depths (43 cm). The test used only barren moist silt, which does not support permafrost under present-day climatic conditions (air temperature and snow), to assess climatic effects without varying ecosystem properties. Second, we tested effects of covarying snow, vegetation, and soil properties of 11 terrestrial ecosystems (excluding open water) for a constant MAAT -2.2 °C). We used covarying properties because the ecological factors are not independent. Input data for these simulations were compiled from field measurements as described above (Table 1). Finally, we tested the effects of varying organic layer thickness and soil moisture for black spruce under one temperature regime to approximate the effects of differing levels of burn severity on organic removal. In these simulations, once soil moisture is specilied, it is held constant throughout the year. Although we recognize that this is not realistic, we do not have sufficient data on each ecosystem type to allow more specific prescriptions.

Results

Climate effects were evaluated by modeling the effects of varying air temperature and snow depth on bare ground to avoid the influence of vegetation and organic matter accumulation (Fig. 4L Modeling of varying air temperatures, when snow depth (recent 10-year average for Fairbanks of 43 cm) was held constant over bare soil, indicate that there is a large offset (6.6-7.4 °C) between MAAT and MAST, The offset (5.7-6.1 °C) was slightly smaller between MAST and 2 m deep temperatures (MADT). The offsets decreased slightly with warming temperatures, and there was a small (up to 0.3 °C) nonlinear response in offsets as permafrost develops near the -6 °C MAAT due to effects of partial unfrozen water contents in the soil. These results indicate that permafrost is formed by climate conditions alone at MAAT

Fig. 4. Modeled effects of mean annual air temperatures (snow depth constant at 43 cm) (top panel) and snow depth (air temperature constant at 2.2 °C) (bottom panel) on surface and deep (2 m) soil temperatures for bare soil conditions where biological effects are absent.



near -6 °C. This is the value that is frequently used (sometimes down to –8°C) to delineate the southern boundary of continuous, or "climate-driven", permafrost (Brown and Pewe 1973).

Snow depth had a large effect on ground temperatures. A 30% (from 43 to 30 cm) and 50% (to 22.5 cm) reduction in snow resulted in a decrease in MAST of 0.8 and 2.0°C, respectively, and a decrease in MADT of 1.8 and 3.8 °C, respectively (Fig. 4). The effects of snow, however, are nonlinear, with soil temperatures decreasing more rapidly as snow cover thins. A 30% increase in snow (to 56 cm) increased MADT by 0.8°C.

Elevation has a strong influence on summer and winter air temperatures, snow depth, and surface temperatures, but north-south aspect has only minor affects. Data compiled from the monitoring of air and soil temperatures in the Yukon-Tanana Uplands near Eielson Air Force Base by Jorgenson et al. (2003) revealed that thawing degree-days during summer were ~10% higher at 400 m elevation than at 200 m in the valley bottom but then decreased rapidly with additional elevation (Fig. 5). Freezing degree-days during winter were ~40% lower (warmer) at 400 m elevation in comparison with the valley bottom and were still ~ 20% lower at 800 m. On an annual basis, this equates to MAATs at 400 and 800 m being ~2.3 and 0°C warmer, respectively, than MAATS at 200 m. Snow was also strongly related to elevation, with snow depths nearly double at 800 m

Fig. 5. Dependence of thawing and freezing degree-days and snow depths on elevation based on 3 years of monitoring at 39 sites in the Yukon-Tanana Uplands near Fairbanks, Alaska.

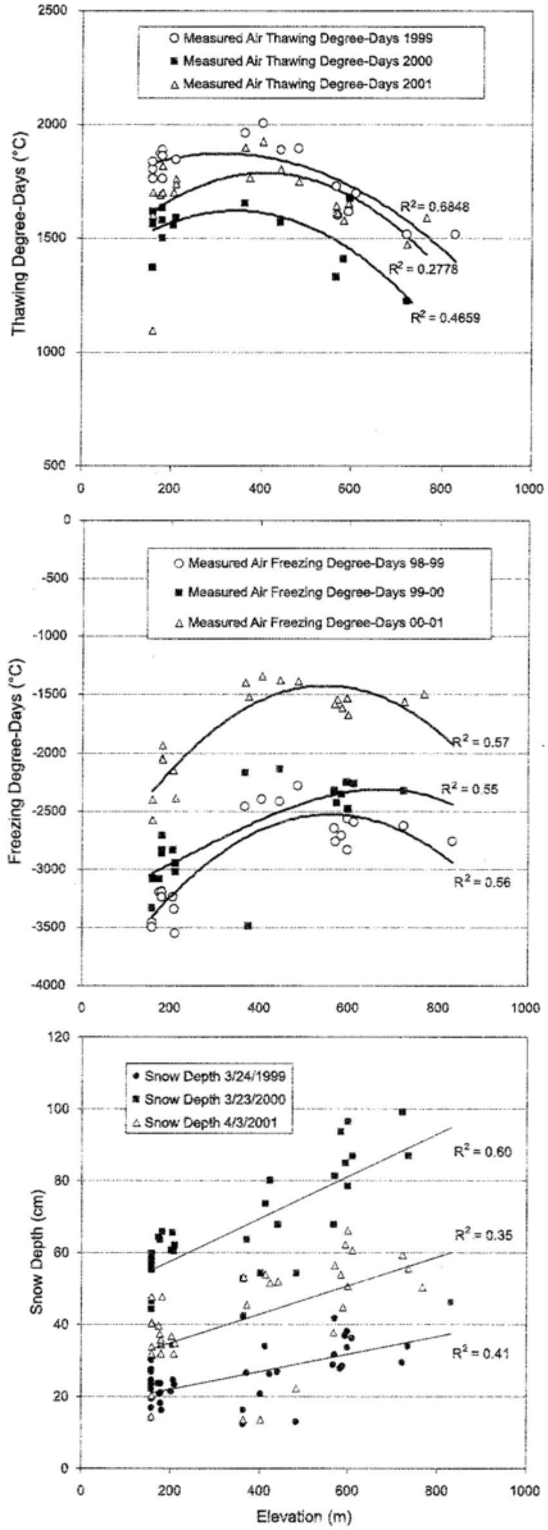
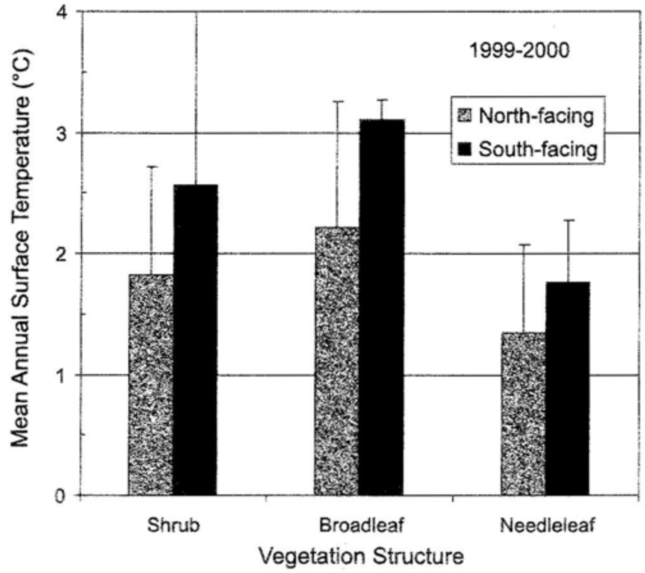


Fig. 6. Comparison of mean (±SD) (n = 3) annual surface (−5 cm depth) temperatures between north- and south-facing slopes when grouped within three vegetation structural classes.

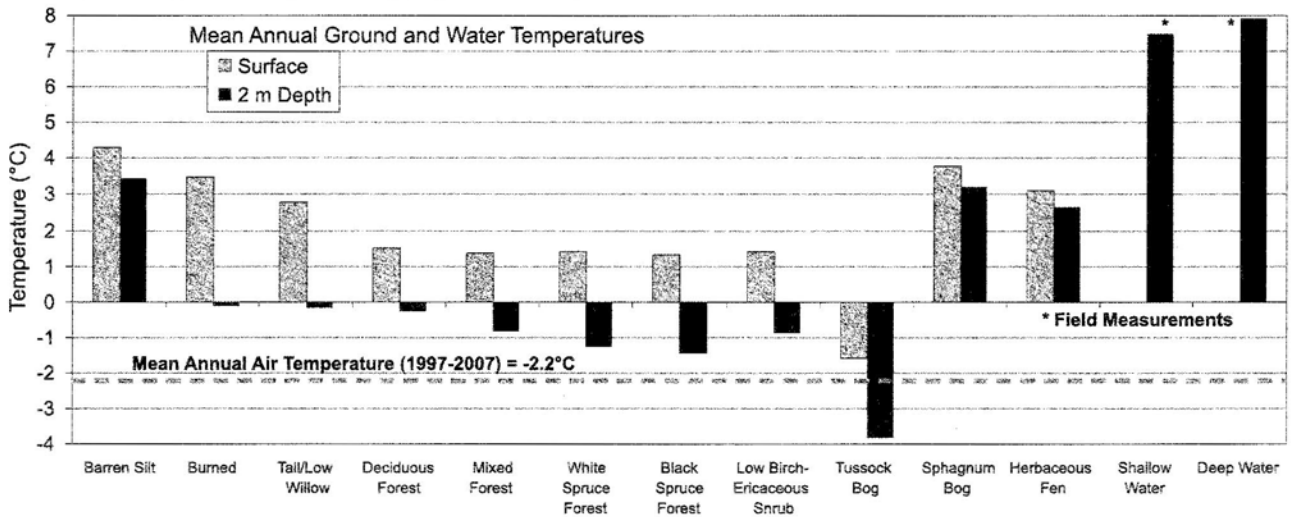


compared with 200 m (Fig. 5). In contrast, aspect had less of an effect MASTs were 0.4-0.9 °C warmer on south-facing slopes than on north-facing slopes when compared within similar vegetation structures (Fig. 6). Topography also affects surface water by allowing surface water to run off on slopes and to impound on flats. Lakes, fens, and bogs created by thawing permafrost occupy ~4% of the boreal forest region of Alaska and these occur almost exclusively in lowland and riverine landscapes (Jorgenson and Osterkamp 2005).

Boreal ecosystems in subarctic Alaska have diverse characteristics in response to varying conditions across the landscape (Table 1). In the boreal biome, regional physiography is dominated by lowlands (43% of area) and uplands (25%), with alpine (18%), riverine (8%), subalpine (4%), and lacustrine (1%) systems occupying the remaining area (Jorgenson et al. 2007). In this paper, we focus on lowland ecosystems for the purpose of assessing ecosystem effects to avoid the complication of varying topography and mineral soil textures. Within lowland areas, ecosystems are dominated by black spruce forest (48%) and low birch - ericaceous shrub (15%), with nine other terrestrial (34%) and two aquatic ecosystems (3%) occupying the remaining area. The terrestrial ecosystems in Table 1 are ordered by their successional relationships ranging from barren ground to midsuccessional birch and white spruce forests along a fire disturbance recovery sequence. which are common on silty lowland soils. In better-drained gravelly lowland soils, aspen is a common midsuccessional deciduous component. These are later replaced by black spruce, dwarf birch shrub, and tussock bogs that progress along a paludification sequence over very long time periods. After thawing of ice-rich terrain, which typically occurs only in late-successional ecosystems, sphagnum bogs and herbaceous fens develop. These terrestrial and aquatic ecosystems have highly variable water, soil, and thermal properties.

Ecosystem characteristics, with their covarying vegetation structure, organic matter thickness, soil moisture, and snow

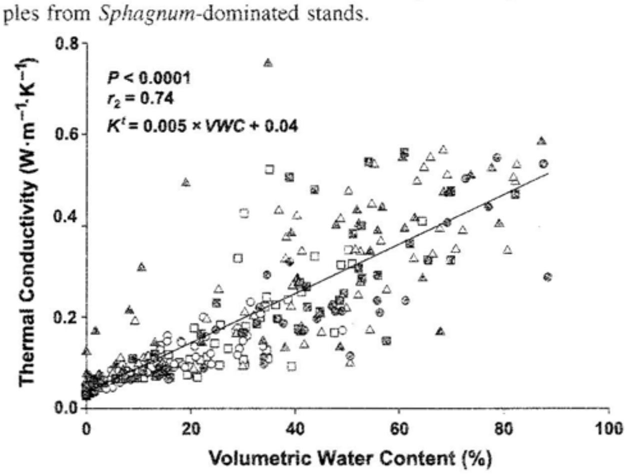
Fig. 7. Modeled mean annual temperatures at the surface and at 2 m depth for common boreal ecosystems in central Alaska. Mean annual temperatures for shallow (<1.5 m) and deep (>1.5 m) waterbodies were measured in waterbodies (n = 1 each) in Denali National Park and Preserve (Chris Arp, US Geological Survey, unpublished data).



depths, had large effects on ground temperatures (Fig. 7). For terrestrial ecosystems, modeled MASTs ranged from 4.3 °C for barren silt to -1.6 °C for tussock shrub and MADTs ranged from 3.5 to -3.°C, respectively. At early-successional stages (barren to deciduous forest), ground temperatures are close to those necessary to form permafrost when air temperatures are near -2 to -3 °C. At later successional stages, the additive effects of increasing organic matter thickness and moisture and decreasing snow depth contribute to permafrost formation. After thawing of permafrost, collapse of the ground surface, and the impoundment of water near the surface, MASTs in bogs and fens were ~5-6 °C above MAATs. Measured temperatures at the bottom of lakes (surface of sediments) were near 8°C, nearly 10°C above MAATs. This variation in surface conditions, which can occur over distances of only a few kilometres, resulted in a total range in near-surface temperatures of ~2 °C across the boreal landscape under similar climate conditions.

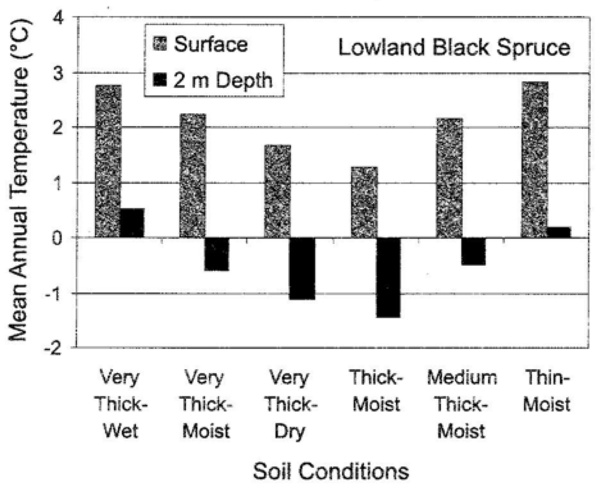
Soil thermal properties, which are affected by organic matter thickness and moisture, contribute substantially to the modeled differences in ground temperatures. Organic matter thickness, subdivided into fibric and hemic layers, ranged from 0 cm on barren ecosystems to >100 cm in sphagnum bog ecosystems. Thermal conductivity was significantly (P < 0.0001) related to volumetric water content in surface organic horizons from black spruce ecosystems (Fig. 8). This relationship was consistent across different moss types (feather moss and *Sphagnum* spp.) and organic horizon type (live moss, fibrous organic matter, and hemic organic matter). Live *Sphagnum* spp. have a higher field capacity than live feather moss (O'Donnell et al. *2009b),* which accounts for the higher range of thermal conductivity values observed in live *Sphagnum.* This difference between moss types was less pronounced in fibrous and hemic samples where we observed a similar range of thermal conductivity values in organic horizons derived from feather moss and *Sphagnum* spp,

Fig. 8. Linear regression between thermal conductivity and volumetric water content of surface organic horizons from black spruce (*Picea mariana*) ecosystems. The different symbols represent live moss (circles), fibrous organic matter (squares), and hemic organic matter (triangles). The open symbols represent samples from feather-moss-dominated stands and shaded symbols represent samples from *Sphagnum*-dominated stands.



Organic matter thickness and moisture effects were examined separately for lowland black spruce on flat terrain while holding air temperature and snow constant (Fig. 9). When comparing the effect of organic matter thickness at constant moisture conditions, both MASTs and MADTs varied by 1.5 °C. Deep soil was thawed only under thin, moist surface organic soil conditions. The lowest modeled MADTs (-1.5 °C) occurred under thick (30 cm cumulative), moist soil conditions. These temperature changes with varying organic matter thickness approximate what one could expect from burning and recovery of black spruce forest. We also compared modeled temperatures for wet conditions, which can occur at later stages of paludification or when thaw settlement leads to wetting conditions, and for dry conditions,

Fig. 9. Modeled mean annual surface and 2 m depth temperatures in relation to thickness of surface organics and moisture content at the surface. Thickness ranged from very thick (6 cm moss, 12 cm fibric, and 82 cm hemic) to thick (4, 13/, and 3 cm), medium thick (3, 9, and 9 cm), and thin (2, 5, and 5 cm).



Which can occur when adjacent areas thaw and collapse, causing higher areas to become drier. Under wet conditions, MADT was above freezing at 0.4 °C even though the organic layer was very thick. In contrast, under very thick-moist and very thick-dry conditions, MADTs were reduced to -1.1 and 0.6°C, respectively. This indicates the effects of dry years on lowering permafrost temperatures.

Ground ice morphology and distribution in permafrost-affected soils are highly complex at both the microsite and site levels. Examples of the vertical distribution of ground ice near Healy, northwest Denali National Park and Preserve, and northwest Wrangells - St. Elias National Park and Preserve show that ice contents frequently are 60%-85% by volume in the top 3 m (Fig. 10). In some profiles, ice contents are highest at 50-150 cm depth and decrease rapidly at greater depth, while at some sites, ice contents were maintained throughout the entire profile. These data illustrate the difficulties in trying to predict the distribution of ground ice across the landscape and the need for a large data set to adequately partition the variation in ground ice by terrain unit and ecosystem type. Based on limited data and our experience of sampling ground ice across central Alaska, we developed conceptual models of vertical ground ice distribution that include six main types of ground ice profiles (Fig. 11). Ice-poor epigenetic permafrost (formed after sediment is deposited) occurs in all soil textures, and permafrost formation is either climate or ecosystem driven. Ice-rich parasyngenetic permafrost develops in closed taliks in lacustrine and glaciolacustrine deposits. and formation is usually climate driven. Ice-poor epigenetic permafrost with an ice-rich intermediate layer near the surface occurs in all soil textures, and formation is ecosystem driven or climate driven, ecosystem modified. Ice-rich syngenetic permafrost (formed during continuous sedimentation) with numerous thick ice-rich layers to substantial depths usually occurs in silty soils, and formation is climate-driven. Ice-rich syngenetic permafrost with an ice-rich intermediate layer occurs in silty soils (intermediate layer forms after surface sedimentation has stopped), and formation is climate driven, ecosystem modified. Ice-rich syngenetic permafrost with an ice-poor layer near the surface caused by thawing and refreezing of soils near the surface occurs in silty soils and formation is climate driven, ecosystem protected (ecological recovery and refreezing after disturbance occurs faster than permafrost can completely degrade).

Ice volume and vertical distribution affect the response of permafrost to climate change in three ways. First, the distribution of ice affects how much newly thawed soil can be incorporated into the active layer as it thickens in response to warming. If the degrading permafrost is mostly ice, little soil is added to the active layer, and every year, thawing continues into new ice. Second. the total amount of ice throughout the profile determines how much settlement will occur upon thawing. Ice-poor soils will have little total and differential thaw settlement and low likelihood of impounding water. Third, latent heat of fusion is directly proportional to ice content and the high latent heat content of ice-rich soil impedes the rate of downward thawing, especially below 3 m. This often allows vegetation succession to reestablish freezing conditions before a significant amount of permafrost is thawed.

Surface water had the largest effect on ground temperatures (Fig. 7). Mean annual temperatures at the sediment surface at the bottom of two lakes in Denali National Park and Preserve were near 8°C, about 10 °C above the MAAT (Chris Arp, US Geological Survey, unpublished data). We relied on measured temperatures for these ecosystems because our soil thermal modeling does not apply to these conditions.

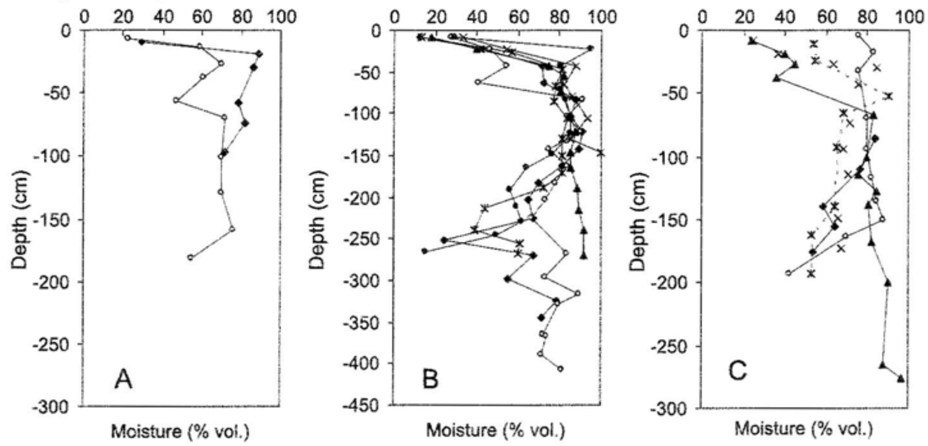
Groundwater effects on permafrost stability have long been noted (Williams and van Everdingen 1973), but data are site specific and information on groundwater patterns across boreal landscapes is inadequately synthesized because of sparse data. Unfrozen zones in permafrost due to groundwater movement are frequently related to braided floodplains, alluvial fans, and retransported deposits on lower slopes in valleys. Groundwater, which has temperatures that range from 2 to 7 °C near Fairbanks, Alaska, is capable of delivering substantial heat to subsurface flow zones. Collapse-scar fens and sink holes are prominent manifestations of permafrost degradation caused by groundwater movement. Collapse-scar fens, which occur over 2% of the boreal forest region of Alaska, are the most abundant thermokarst landform (Jorgenson et al. 2007).

Discussion

Relative effects of biophysical factors

Field observations and modeling show that MAST (including surface of lake sediments) can vary by ~12 °C across the boreal landscape within a similar climatic region due to the complex interaction of topography, hydrology, snow, soil properties, and vegetation. As yet, there is no single model that can incorporate these disparate effects and we were forced to isolate and partition these factors to assess their relative effects. We found that standing surface water, with its very low albedo, had the greatest effect by increasing near-surface temperatures by ~10°C above MAATs. This is similar to the range of mean annual temper-

Fig. 10. Vertical distribution of soil moisture for three sites in central Alaska: (A) Twelve-mile Lake near Healy has a thin layer of peat and silt over rocky till, (B) Gosling Lake in northwestern Denali National Park and Preserve has thick peat over alluvial silt, and (C) Billy Lake near Glennallen has thick peat over glaciolacustrine silt.



atures near the surface on the coastal plain of northern Alaska where temperatures under deep lakes are ~1°C and under vegetated tundra are mostly -7 to -10 °C (Romanovsky et al. 2003). This large effect of surface water is responsible for extensive permafrost degradation of both cold and warm permafrost and can thaw permafrost even under extremely cold conditions (MAAT of -19.7 at Ellesmere Island) in the high Arctic (Jorgenson et al. 2006; Grom and Pollard 2008). Standing water in turn is primarily controlled by topography, with water able to impound on flat terrain or after surface settlement resulting from melting of ground ice. Thus, the vertical distribution of ground ice in relation to the surface is important in affecting thaw settlement when the active layer adjusts to climate change or disturbance.

Terrestrial ecosystem properties, with covarying vegetation structure, organic matter to °C across a flat landscape, with MADTs ranging from -3 °C under barren silt to -4°C under tussock bogs modeled for recent MAATs or the Fairbanks area of -2.2 °C. Without vegetation, modeling indicates that permafrost forms on bare ground at approximately –6 °C, consistent with the common attribution of -6°C as the southern limit of the boundary of continuous permafrost. Although we did not examine the relative contribution of each covarying property because of the large number of interacting effects, we can infer several major interactions. The shift from deciduous trees to coniferous trees during succession after disturbance is a major turning point in MADTs due to the interception of snow on coniferous branches, the reduction in litter fall that allow mosses to become more prominent under coniferous forests, and the accumulation of thicker organic horizons (Viereck 1970). **In** addition, increase in soil moisture during succession affects the ratio of frozen to unfrozen thermal properties, thus reducing heat input during summer and increasing heat loss during winter. This causes a thermal offset of the active layer whereby deep soil temperatures can be colder than surface temperatures. Our modeled temperature offsets (0.9-1.8 °C) between MAST and MADT were similar to those (0.4-1.7 °C) measured by Burn and Smith (1988) in northwest Canada. This large ecological effect provides strong evidence for the concept of *ecosystem-driven* permafrost in the discontinuous permafrost zone. At late stages of succession and paludification (e.g., tussock bog), ecosystem properties can reduce MADTs to nearly 2 °C below MAATs. These late-successional ecosystems are responsible for the persistence of *ecosystem-protected* permafrost at MAATs of up to *+2°C* in southcentral Alaska.

Snow had a substantial effect on ground temperatures by lowering MADT by 2°C in barren silt when snow thickness was reduced by 30%. Thus, variation in snow depth over time alone is sufficient to affect the aggradation and degradation of permafrost for some conditions in central Alaska. Snow has been shown to be a major contributor to observations of recent ground warming in centra! Alaska (Romanovsky et al. 2003; Osterkamp et al. 2009). Vegetation, however, also has a large effect on snow depth by capturing blowing snow (Sturm et al. 2001) or by intercepting snow on branches, which can later fall in denser clumps. There is little change in modeled MASTs between lowland deciduous forest and black spruce forests, however, indicating that changes in MADTs during late-successional. ecosystems are more attributable to organic thickness and moisture than to snow depth differences.

Feedbacks

The resilience of permafrost and permafrost-affected boreal ecosystems to climate change is greatly promoted by negative feedbacks from ecosystem properties that develop during vegetation succession, while permafrost is made more vulnerable to climate change by the positive feedbacks caused by impounding water in thermokarst terrain. These feedbacks interact with many ecosystem components (Fig. 2), particularly microtopography, which in turn is affected by ground ice and thaw settlement properties,

Negative feedbacks associated with biological processes that promote permafrost resilience involve canopy affects on shading and snow interception, litterfall quality and quantity, moss growth, increasing moisture and decreasing temperatures that reduce decomposition, and development of permafrost that impedes drainage (Fig. 2). Based on suecessional stages from barren silt to white spruce forest to tussock bog (Table 1), the covarying properties can reduce MADTs by 7°C, and at the last stage, ecosystem properties can reduce MADTs to nearly 2°C below MAATs. The formation of permafrost at intermediate- to late-successional stages blocks subsurface drainage that further increases soil moisture, reduces decomposition, and helps shift vegetation to low-nutrient-tolerant species, which further reduces permafrost temperatures. Mosses play an important role in the resilience of permafrost through their substantial growth rates, slow decomposition and buildup of organic matter, and combustion patterns that affect fire severity and organic matter loss (Turetsky et al. 2010). The transition of deciduous to coniferous trees, and accompanying moss buildup, is a major turning point in soil temperatures and permafrost formation in lowland soils. Recent trends in postfire successional trajectory indicating a shift from coniferous to deciduous stands from more frequent fires (Euskirchen et al. 2010), however, may reduce the ability of permafrost to recover after fire because the important negative feedbacks that accompany the coniferous stage are reduced.

Positive feedbacks that increase the vulnerability of permafrost involve the interaction of surface water and ground ice. Surface water provides a powerful positive feedback because it greatly affects albedo and energy budgets (Boudreau and Rouse 1995). The albedo of surface water can be as low as 5% at high sun angles, whereas the albedos for dry herbaceous and lichen-rich vegetation and for mixed hardwoods can be as high as 30% and 20%, respectively (Lee 1978). Chapin et al. (2005) reported albedo values of 17% for tundra, 15% for shrub, and 11% for forest vegetation. Wind-driven mixing of water in lakes also substantially increases the effective conductivity of water and thus promotes heating of the lake bottom. As a result, surface water can raise near-surface temperatures by 10°C over MAATs in central Alaska, while in arctic Alaska, surface water can raise near-surface sediment temperatures by 12-14°C (Chris Arp, US Geological Survey, unpublished data). During winter, heat loss is reduced by ice formation and snow cover. Because surface water provides such a strong positive feedback, most permafrost degradation is associated with waterbodies. Under cold climates, the increased heat gain from surface water may lead only to surface permafrost degradation, unless ice volumes are sufficient to facilitate development of thermokarst lakes deeper than 2 m. Under warmer climates, a talik may develop and thaw can easily proceed through the entire permafrost layer. The relatively warm mean annual temperatures of surface water in lakes and at the margins of bogs and fens lead to lateral thawing of permafrost. The widespread occurrence of thermokarst lakes, collapse-scar bogs, and collapse-scar fens is attributable to this type of thawing. Thermokarst lakes, however, become susceptible to drainage from lateral breaching or subsurface drainage (Yoshikawa and Hinzman 2003).

Ground ice volume and its vertical distribution influence the potential for positive feedbacks from surface water by affecting microtopography and depth of water impoundment (Fig, 11). Ice-poor soils will have little thaw settlement and the likelihood of impounding water is low, while thaw settlement of as little as 10-20 cm can cause water to impound. Where permafrost is extremely ice rich, thaw settlement can lead to impoundment of water deeper than 2 m, which no longer freezes to the bottom during the winter. Ground ice also affects the rate of thawing due to the high latent heat content of ice. If ground ice is mostly at the surface, this often can be quickly eliminated, whereas if the ground ice is high at greater depths, this will require greater heat tlux. Soil thawing at depth tends to slow down over time, especially at depths >3 m. The rate of thawing is important because slow thawing can allow vegetation succession to reestablish freezing conditions before thawing has penetrated very deep (Fig. lID).

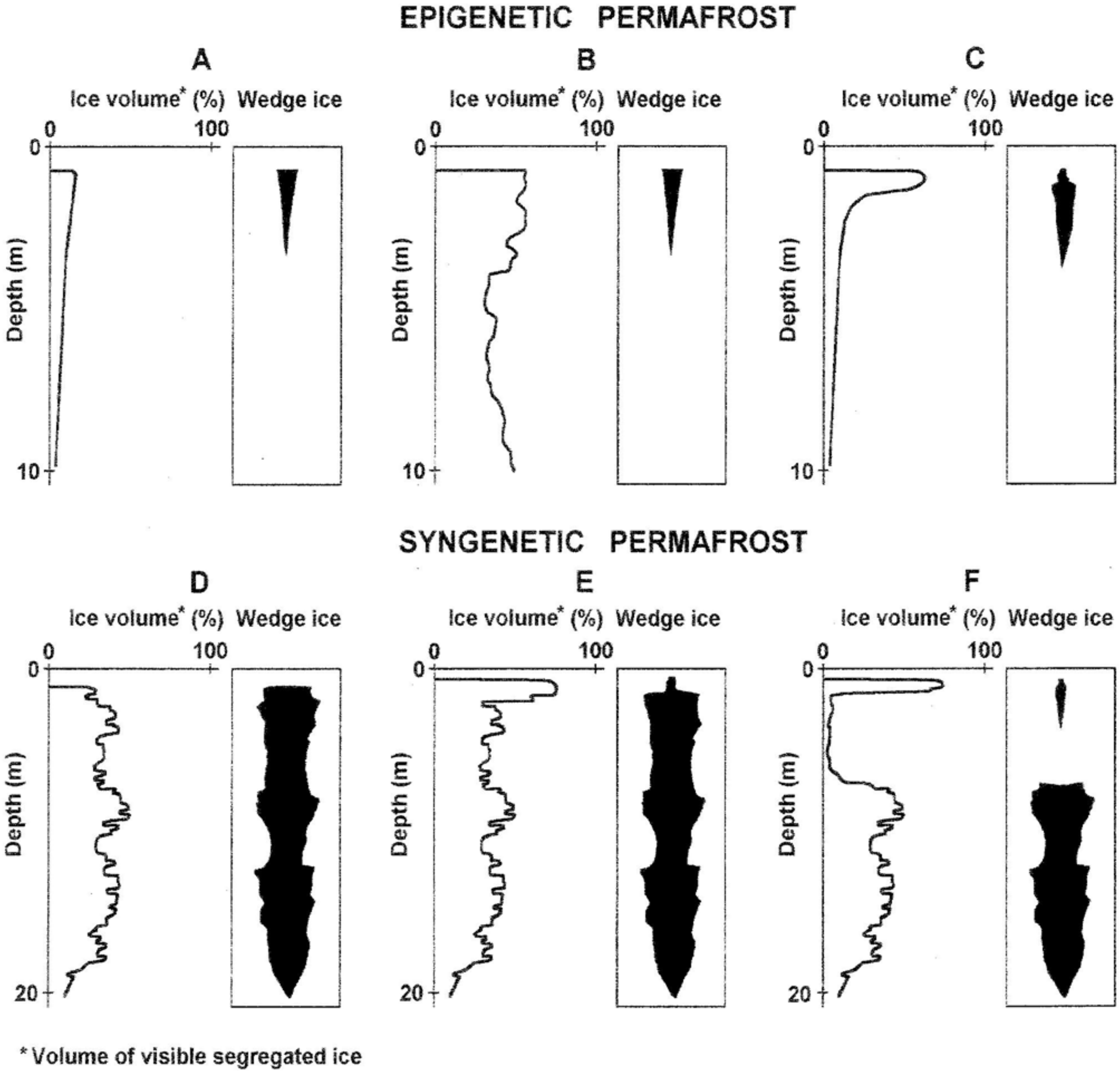
Topography strongly influences whether positive or negative feedbacks to permafrost stability will dominate. Sloping terrain improves runoff and substantially reduces the potential for the positive feedback from impounded water. The resilience of extremely ice-rich permafrost of Pleistocene age (yedoma) on loess and colluvium mantled slopes in central Alaska is a good example of how topography interacts with extremely ice-rich terrain. On these ice-rich slopes, surface impoundment is reduced, the latent heat of fusion in massive ice slows thawing at depth, and vegetation succession allows freezing conditions to reestablish after disturbance. In contrast, extremely ice-rich loess on flat terrain is almost entirely absent in Alaska regardless of climatic conditions.

These strong negative and positive feedbacks are larger than the predicted climate warming of -5 °C in central Alaska and complicate our ability to predict the response of permafrost to climate warming. With the negative feedbacks from ecological succession, and lacking severe fire disturbance that creates bare soil, permafrost in some environments could persist to MAATs as high as +2 °C, about the temperature predicted for central Alaska under intermediate climate scenarios (Scenarios Network for Alaska Planning 2008). In contrast, with the even stronger positive feedback from surface water, the degradation of extremely ice-rich permafrost to form thermokarst lakes and thermo karst pits over degrading ice wedges is common even under the cold arctic climatic conditions.

Disturbance

Fires dramatically change the energy balance of the landscape by (i) decreasing the thickness of the insulating organic soil (Yoshikawa et al. 2003), (ii) increasing the reflectance, or albedo (Randerson et al. 2006; Euskirchen et al. 2010), and thereby reducing the solar insolation during both winter and summer, and (iii) decreasing snow interception and contributing to a thicker snow cover (Jorgenson et al. 2003). Our model results for the combined effects of organic matter thickness and moisture indicate that MADTs are near 0 °C for burned black spruce and for black spruce with thin moist organics. This indicates that after a severe burn that combusts most of the organic matter, the thawing or persistence of permafrost can go either way depending on small differences in site conditions. How much permafrost will degrade after fire, however, will depend on ground ice contents and distribution. If the ice, and thus latent heat, contents are low and permafrost is thin, the permafrost could totally degrade within decades, the soils become well drained, and the recovery of thick organics slowed. If ice and latent heat contents are high, degradation can be slow enough that the decrease in soil temperatures over decades

Fig. 11. Generalized profiles of vertical ice distribution for common types of permafrost. The left diagram of each pair shows the vertical distribution of visible segregated ice, while the right diagram shows a typical cross section of ice wedges, which usually form at mean annual ground temperatures less than –4 °C). (A) Ice-poor epigenetic permafrost (formed after soil material have been deposited); ice-rich horizons in epigenetic permafrost can be observed at different depths where downward freezing reached aquifers. (B) Ice-rich parasyngenetic permafrost formed in closed taliks in lacustrine and glaciolacustrine deposits. (C) Ice-poor epigenetic permafrost overlaid by ice-rich intermediate layer formed near the surface when organic matter accumulation during late-successional stages causes the active layer to thin. (D) Ice-rich syngenetic permafrost formed during ongoing surface sedimentation, such as in fluvial and eolian deposits. (E) Ice-rich syngenetic permafrost later modified by an ice-rich intermediate layer formed in association with organic matter. (F) Ice-rich syngenetic permafrost with the layer of ice-poor thawed and refrozen soils near the surface overlaid by ice-rich intermediate layer.



of successional development can limit talik development and allow permafrost to recover.

While in recent centuries, permafrost has recovered after fire as the forest vegetation and organic soils have built back up after burning (Harden et al. 2006; Yi et al, 2009), shorter fire return intervals and higher fire severities in North American boreal forests have been noted since the 1980s (Kasischke et al. 2010), With shorter fire return times, and particularly with severe fires, conifer overstories and thin moss and organic matter layers may not develop and permafrost may not be able to recover to predisturbance depths, Thus, with drying soils, black spruce and larch forests might be replaced by birch, aspen, white spruce, and lodgepole pine, species that currently occupy well-drained, permafrost-free soils in the boreal zone. Fire severity and the removal of organic matter, however, are affected by moss species composition, topography, soil temperature, and thaw depth above permafrost (Kasischke et al. 2010: Turetsky et al. 2010). These biophysical factors can interact to reduce fire severity and thus increase the resilience of permafrost.

With additional regional atmospheric warming of ~5 °C in central Alaska during the next century under intermediate climate change scenarios (Scenarios Network for Alaska Planning 2008), our modeling indicates that permafrost will not be able to recover from disturbances, such as fire, despite the negative feedbacks that accompany succession and promote recovery. Moreover, if snow depths increase in the interior with a changing climate, permafrost is unlikely to persist after fire even with only a 1 °C increase above recent Fairbanks MAATs of -2.2 °C because of the additional strong insulating effect of snow that reduces heat loss in winter. However, there is substantial uncertainty as to whether winter precipitation wiil increase in the future (Scenarios Network for Alaska Planning 2008) and there has actually been a decreasing trend since 1989 (Euskirchen et al. 2010).

Consequences for ecosystem services

Permafrost thaw and ground collapse have numerous biological and physical effects on both natural ecosystems and human infrastructure. In turn, the radical changes in ecosystem characteristics after thawing alter how the landscape provides ecosystem services in terms of global regulation of climate and in provisions of ecosystems to higher trophic levels and subsistence consumers of fish and wildlife.

Permafrost persistence or degradation will affect climate regulation services by altering the overall heat gain of the landscape through changes in albedo, affecting the sequestration and release of soil carbon, and potentially increasing the emissions of CH_4 from expanding aquatic systems. Just as the albedo feedback from sea ice melting has the potential to accelerate the trend towards an ice-free Arctic Ocean during the summer and to affect global temperatures (Perovich et al. 2007), a similar potential exists for ice-rich lowland terrestrial ecosystems in Arctic and Subarctic Alaska. When forested ecosystems with relatively high albedos are converted to waterbodies with low albedos, there is a large increase in heat flux into the water. Much of this heat is used to evaporate water and the heat is delivered to the atmosphere through vapor. Thermokarst lakes already occupy a substantial portion of lowland ecoregions in Alaska and their abundance is likely to increase. The albedo effect from conversion of coniferous forests to earlier suscessonial deciduous scrub and forest (Euskirchen et al. 2010), as well as lake drainage, will counteract to some extent the albedo affect of thermokarst lakes.

Permafrost thaw has the potential to affect the exchange of CO_2 and CH_4 between ecosystems and the atmosphere as plant and soil processes are altered by ground subsidence (Harden et al, 2008). Permafrost soil contains large pools of organic carbon accumulated over thousands of years (Harden et al. 2006; Ping et al. 2008; Tarnocai et al. 2009) and this carbon would become vulnerable to decomposition upon thaw (Schuur et al. 2007, 2008). Increased thaw and ground surface subsidence can increase plant growth and change species composition in upland tundra ecosystems (Schuur et al. 2008). Initially, increased plant carbon uptake can offset carbon losses from respiration, but over decades of thaw, upland tundra areas appear to have become net sources of carbon to the atmosphere due to decomposition of older carbon deep in the soil (Schuur et al. 2009). The overall carbon balance resulting from the radical changes in soil properties as forests are converted to thermokarst bogs and fens, however, remains uncertain. Lowlands have been shown to accumulate surface organic carbon (Turetsky et al. 2007), consistent with increased plant growth, but there is a lack of information on the overall balance between carbon released by thawing of forest peat that has become submerged and carbon gained by newly developing bog and fen peat.

Methane emission rates are drastically altered by thawing of ice-rich permafrost (Christensen et al. 2004; Myers-Smith et al. 2008; Turetsky et al. 2007). This is of global concern because CH_4 has 25 times greater heat trapping capacity than CO_2 on a century time scale (Arctic Climate Impact Assessment 2005) and because permafrost-affected soils in the Arctic region are a globally significant source of CH_4, with estimates of net methane emissions ranging from 31 to 100 Tg $CH_4 \cdot year^{-1}$, and after accounting for CH_4 consumption (McGuire et a1. 2009). Walter et al. (2006) raised the possibility that CH_4 emissions from thawed Pleistocene-aged carbon released into thaw lakes can affect global atmospheric concentrations of CH_4. The magnitude of CH_4 from diverse thermokarst landforms in boreal ecosystems, however, remains poorly quantified.

Ecosystem provisioning services related to surface and groundwater abundance and habitat use by fish, wildlife, and people also will be radically altered by permafrost degradation. The hydrologic redistribution of surface water caused by degrading permafrost can be dramatic. Permafrost prevents subsurface drainage and limits the developmen of hydrologic networks in permafrost regions (Woo 1990). Permafrost thaw can increase subsurface drainage in uplands and cause lake drainage in some lowland situations where thawing of highly permeable subsurface grave] layers can drain water laterally (Yoshikawa and Hinzman 2003; Walvoord and Striegl 2007). The observed increase in freshwater discharge, particularly during winter, may be related to increased infiltration of water into the groundwater system (Walvoord and Striegl 2007). Changes in surface runoff versus groundwater input are likely to increase dissolved cations while reducing dissolved organic carbon (Striegl et al. 2005). In other lowland areas, permafrost degradation can create isolated thermokarst lakes that alter patterns of evaporation and runoff.

Plant and animal populations are directly affected by the changing hydrologic regime as permafrost thaws, most significantly by locally increasing or decreasing wetland extent (Jorgenson and Osterkamp 2005). With drying soils, black spruce and larch forests might be replaced by faster growing birch, aspen, white spruce, and lodgepole pine, species that currently occupy well-drained, permafrost-free soils in the boreal zone. Birds and mammals that occupy mature forests may be at a disadvantage, as ecosystems on permafrost ground are disturbed by thaw processes and undergo succession. Thermokarst lakes in lowland boreal and tundra regions are already important habitats to a wide range of waterbirds; these habitats will expand or shrink depending on local permafrost conditions. If ground ice content is insufficient to allow thermokarst lake development, or where accumulation of sphagnum mosses is rapid, nutrient-poor bogs develop. Bogs support a diverse moss flora, but they are typically poor habitat for many vascular plants, birds, and mammals because of acidification and low primary production (Turetsky et al. 2010). High-nutrient fens that develop in degrading permafrost areas with groundwater movement support more productive herbaceous plant species that provide forage for herbivores. Interestingly, all of the National Wildlife Refuges in central Alaska that were established to protect waterbird habitats, including Tetlin, Yukon Flats, Kanuti, Koyukuk, and Innoko, are areas that have abundant lakes and wetlands created by thawing permafrost. The importance of permafrost in ecosystem processes is highlighted in a recent summary of wildlife responses to environmental change in the Arctic by the US Fish and Wildlife Service (Martin et al. 2009).

Conclusion

Assessing the response of permafrost to future climate warming is complicated by the complex interaction of biophysical factors, which create strong negative feedbacks from vegetation and soil processes that make permafrost more resilient to climate warming and disturbance and positive feedbacks from thaw settlement and water impoundment that make permafrost more vulnerable to warming, Vegetation and soil feedbacks can reduce deep soil temperatures by 7 ºC and help permafrost to persist at MAATs of up to +2 'C, while water impounded by thaw settlement can increase ground temperatures by 10 ~C and make permafrost vulnerable to thawing at MAATs as low as -20ºC. These feedbacks are larger than the predicted climate warming of 5ºC for central Alaska, greatly complicating the prediction of the responses of permafrost to climate warming. Severe disturbance from fire adds to the dynamics by creating bare ground in which surface soil temperatures can be as much as 6 ºC higher than MAATs. Yet, rapid changes in albedo with canopy development during early succession. establishment of mosses, accumulation of organic matter. and interception of snow during the transition to a coniferous tree canopy help permafrost recover. If MAATs increase by only 1-2ºC, however, permafrost may not be able to recover after fire due to the combined effects of ongoing warming, the possibility of increased snow cover, and the many decades needed for vegetation succession to reestablish conditions favorable for permafrost development. While our results provide a better understanding of how surface properties and feedbacks affect the resilience and vulnerability of permafrost to climate change and disturbance, we still lack sufficient data and models to predict the distribution of ground ice that controls the magnitude of thaw settlement and the potential for positive feedbacks. Further progress in assessing the response of permafrost to climate change will require advanced integration of geomorphic, hydrologic, thermal, ground ice, and ecosystem succession models to address the full range of complex biophysical interactions

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