Vulnerability of high-latitude soil organic carbon in North America to disturbance

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[1] This synthesis addresses the vulnerability of the North American high-latitude soil organic carbon (SOC) pool to climate change. Disturbances caused by climate warming in arctic, subarctic, and boreal environments can result in significant redistribution of C among major reservoirs with potential global impacts. We divide the current northern high-latitude SOC pools into (1) near-surface soils where SOC is affected by seasonal freeze-thaw processes and changes in moisture status, and (2) deeper permafrost and peatland strata down to several tens of meters depth where SOC is usually not affected by short-term changes. We address key factors (permafrost, vegetation, hydrology, paleoenvironmental history) and processes (C input, storage, decomposition, and output) responsible for the formation of the large high-latitude SOC pool in North America and highlight how climate-related disturbances could alter this pool's character and size. Press disturbances of relatively slow but persistent nature such as top-down thawing of permafrost, and changes in hydrology, microbiological communities, pedological processes, and vegetation types, as well as pulse disturbances of relatively rapid and local nature such as wildfires and thermokarst, could substantially impact SOC stocks. Ongoing climate warming in the North American high-latitude region could result in crossing environmental thresholds, thereby accelerating press disturbances and increasingly triggering pulse disturbances and eventually affecting the C source/sink net character of northern high-latitude soils. Finally, we assess postdisturbance feedbacks, models, and predictions for the northern high-latitude SOC pool, and discuss data and research gaps to be addressed by future research.

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1. Introduction

[2] Northern high latitudes, encompassing the arctic, subarctic and boreal zones, have been identified as a region where global climate change will have earlier and stronger impacts than in other regions of the globe [*Chapin et al.*, 2000; *Serreze et al.*, 2000; *Hinzman et al.*, 2005; *Intergovernmental Panel on Climate Change*, 2007]. A wide variety of northern high-latitude climate components such as terrestrial vegetation, permafrost, arctic sea ice, and arctic atmospheric patterns are already readjusting to climate change and are impacting terrestrial ecosystems [*Chapin et al.*, 1992; *Arctic Climate Impact Assessment*, 2004]. Strong and dynamic feedbacks and interactions between northern high-latitude cryosphere, biosphere, atmosphere and geosphere are also an important driving force of climate change with teleconnections to lower latitudes [*Serreze and Francis*, 2006; *McGuire et al.*, 2006]. A critical component in these mechanisms is the carbon (C) cycle with important reservoirs in northern high-latitude regions in terrestrial soils, live terrestrial biomass, the Arctic

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Ocean, and the atmosphere [*Gruber et al.*, 2004; *McGuire et al.*, 2009]. C fluxes among these reservoirs largely occur in the form of gas exchange or transport of dissolved gases, dissolved organic and inorganic C, and particulate organic and inorganic C. In particular, processes of C sequestration from and C release to the atmosphere can have important implications for global climate.

[3] Permafrost-affected soils and peatlands in the terrestrial northern high-latitude region have sequestered soil organic C (SOC) from organic litter and vegetation over thousands to tens of thousands of years during the Late Quaternary [*Harden et al.*, 1992; *Smith et al.*, 2004; *Gorham et al.*, 2007; *Schirrmeister et al.*, 2010]. Recent panarctic assessments [*Tarnocai et al.*, 2009; *McGuire et al.*, 2009] have demonstrated the global significance of northern high-latitude soils as long-term C sinks. These assessments indicate that about 1400–1850 Pg of SOC are stored in both perennially frozen and unfrozen soils and frozen near-surface sediments in permafrost regions, approximately doubling previous estimates and now accounting for about 50% of the global below ground organic carbon. The transfer of 1 Pg C from a long-term pool to the atmosphere translates to an about 0.471 ppm increase in atmospheric CO_2 [*Battle et al.*, 2000]. This suggests that a 1% loss of C (14 to 18.5 Pg) from the SOC pool to the atmosphere could results in a significant (6.6 to 8.7 ppm) increase in atmospheric C.

[4] The northern high-latitude region has generally been considered to have been a sink for atmospheric CO_2 in recent decades due to tree growth and peat accumulation in the boreal region and to have been a source of CH_4 due to emissions from widely distributed wetlands and lakes [*McGuire et al.*, 2009; *Smith et al.*, 2004; *Walter et al.*, 2006]. However, there is concern that the CO_2 sink may be weakening in the last decade because of the loss of soil organic matter from enhanced decomposition and combustion [*McGuire et al.*, 2010; *Turetsky et al.*, 2011]. Fundamental changes in the C cycle of northern high-latitude soils in response to a rapidly changing climate could have strong feedbacks to global climate, thereby exacerbating (positive feedback) or mitigating (negative feedback) climate change.

[5] The northern high-latitude SOC pool is a dynamic stock affected by variations in inputs (organic litter quality and quantity), the mode and timing of C stabilization (permafrost aggradation, cryoturbation, peat accumulation, sedimentation) and C destabilization (microbial decomposition, combustion), and C export (via dissolved and particulate phases, inorganic and organic state, gas fluxes). Individual processes of these C pool dynamics can act on different timescales, complicating an overall assessment. As a result, the net increases or decreases of the SOC pool are inexorably tied to both the processes and form of C stabilization. Rates of burial, cryoturbation, inundation, and permafrost aggradation become important factors determining soil organic matter quality (form and relative lability or recalcitrance) and together with disturbance history can dictate whether organic C in a soil is most vulnerable to combustion, leaching, or microbial decomposition if subjected to thawing and subaerial exposure. Moreover, interactions among soil processes and the physical and biological systems occur at different times with different responses and durations. For example, one possible local scenario might be: slow increase in active layer depth → slow landscape drying → slow increase in fire probability → fast fire occurrence → fast thermokarst → slow C accumulation and fast CH_4 emissions under anaerobic conditions. Meanwhile other landscapes less vulnerable to fire owing to poor soil drainage may respond only to changes in temperature (more plant production and/or decomposition). Thus, heterogeneity of the landscape is further complicated by the timing and duration of various physical and biogeochemical feedbacks.

[6] Disturbance, or the perturbation of a normal state or regime, of the northern high-latitude soil C cycle is an important natural mechanism rapidly facilitating such changes. Important research issues concern the degree to which disturbances affect SOC storage or loss in northern high-latitude soils, and how past disturbance regimes determine current soil organic matter quality. We define disturbance of the northern high-latitude SOC pool as an event or process forced onto this system that results in a significant redistribution of C among major reservoirs. Natural disturbances alter key ecosystem factors in ways that affect the dynamics of the northern high-latitude SOC pool (Figure 1). Disturbances can be categorized by their time-horizon of influence, the complexity of involved processes, the size of area affected, and the total impact on a SOC pool or the C cycle. Furthermore, press and pulse disturbances can be distinguished [*Collins et al.*, 2007]. *Presses* are impacts driven by decadal to century-scale changes of factors controlling a system. *Pulses* are one-time or episodic short-term events that are often extreme in nature and impact a system very rapidly. Press disturbances with impact on northern high-latitude SOC include, for example, paludification, aridification, top-down permafrost thaw, and changes in pedological processes and vegetation structure (Figure 1). Pulse disturbances include wildfires, thermokarst formation, thermokarst lake drainage, flooding, and rapid erosion of soils. By crossing environmental thresholds, press disturbances could trigger pulse disturbances. For example, widespread top-down permafrost thawing could trigger local rapid thermokarst or increased soil erosion, or changes in vegetation structure could result in increased risk of wildfires. There are also feedbacks between different pulse disturbance processes (Figure 1). For example, wildfires can trigger local thermokarst in some permafrost types or floods trigger soil erosion. Though natural disturbances are an important component in shaping the northern high-latitude SOC pool, changes in their frequency and intensity may have important impacts on this C pool. It is therefore essential to understand how global change will alter disturbance frequencies and intensities in the northern high-latitude region.

[7] *Chapin et al.* [2010] defines vulnerability as the "degree to which a system is likely to change due to exposure and sensitivity to a specific perturbation," such as a disturbance or sustained warming. A SOC pool has a high vulnerability to disturbance if the physical, chemical and/or biological change in soil characteristics results in a SOC loss over the long-term. Specific vulnerabilities for northern high-latitude SOC pools involve different types and locations of SOC: Permafrost thaw affects the thermal state of the SOC in shallow and deep horizons; fire affects near-surface organic matter in soils by combustion; drought and inundation impact shallow and deep SOC pools because of links between organic C sequestration and decomposition to aerobic, anaerobic, and thermal soil conditions; physical and chemical preservation of organic matter can affect its vul-

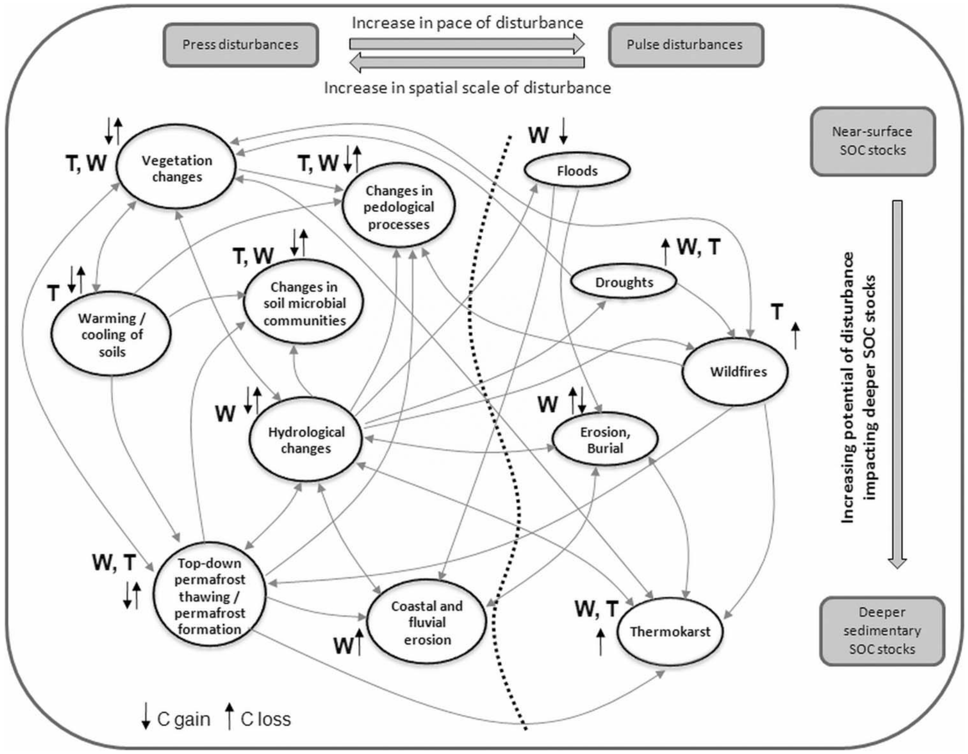

Figure 1. Key disturbance processes impacting northern high-latitude SOC stocks and feedbacks between them. Press disturbances are slower (decades to centuries) but affect larger areas, whereas pulse disturbances can be fairly rapid (seasons to decades) but are often local with sometimes widespread occurrence. Disturbances have different potential to impact SOC stocks at greater depths. Feedbacks between disturbances include for example the impacts of wildfires on vegetation changes, and subsequent soil warming and top-down permafrost thawing. Disturbances affect SOC stocks mainly via changes related to water (W), or temperature (T), or both. For example, wildfires affect SOC in the upper soil organic layers by combustion (temperature); or thermokarst affects SOC stocks by warming soils (temperature) and melting ground ice (water). Small black arrows indicate whether SOC is lost or sequestered due to certain disturbances; some disturbance types can result in SOC losses or gains depending on environmental preconditions.

nerability to disturbance. In contrast, resilience of a SOC pool is the capacity of this system to resist disturbances, and maintain its fundamental structure and functioning [*Chapin et al.*, 2010]. Resilience also describes the capability of the system to recover from disturbances to the previous state.

[8] In this paper, we assess the vulnerability of the high-latitude SOC pool in North America to disturbances in the region. Our approach is to (1) review the current organic C storage in high-latitude soils of North America; (2) describe key ecosystem, climate, and soil processes that characterize SOC pools in this region; (3) discuss major press and pulse disturbances and their impacts on northern high-latitude SOC; (4) evaluate model projections of disturbance impacts on northern high-latitude SOC; and (5) discuss research and data gaps that need to be addressed to better predict the near-future trajectory of SOC in the North America high-latitude region.

[9] For the scope of this paper we here define the northern high latitudes as the region approximately covering the tundra and boreal forest ecoregions (Figure 2a). In North America, these zones encompass regions from about 45–83° N and 53–170° W. A large part of the northern high latitudes is characterized by permafrost (Figure 2a), classified into continuous permafrost (90–100% cover), discontinuous permafrost (50–90% cover), sporadic permafrost (10–50% cover), and isolated permafrost (0–10% cover) [*Brown et al.*, 1997]. There is a general zonal trend in permafrost distribution with more complete cover in the high Arctic and decreasing coverage in lower latitudes. Parts of the northern high latitudes, i.e., in the boreal zone, have only isolated permafrost or do not currently have permafrost. For this synthesis, which is a contribution to the North American Carbon Program (NACP), we primarily base our assessment on the numerous process and modeling studies that have been conducted in North America while recognizing that a large body of related research has been conducted in northern high latitudes of Eurasia.

2. Review of SOC Storage in High-Latitude Regions of North America

[10] The distribution of SOC must be well characterized for any assessment of C vulnerability. A detailed characterization of different C stores within the northern high-latitude soil organic C pool allows a more process-oriented perspective on both the formation and vulnerabilities of C in

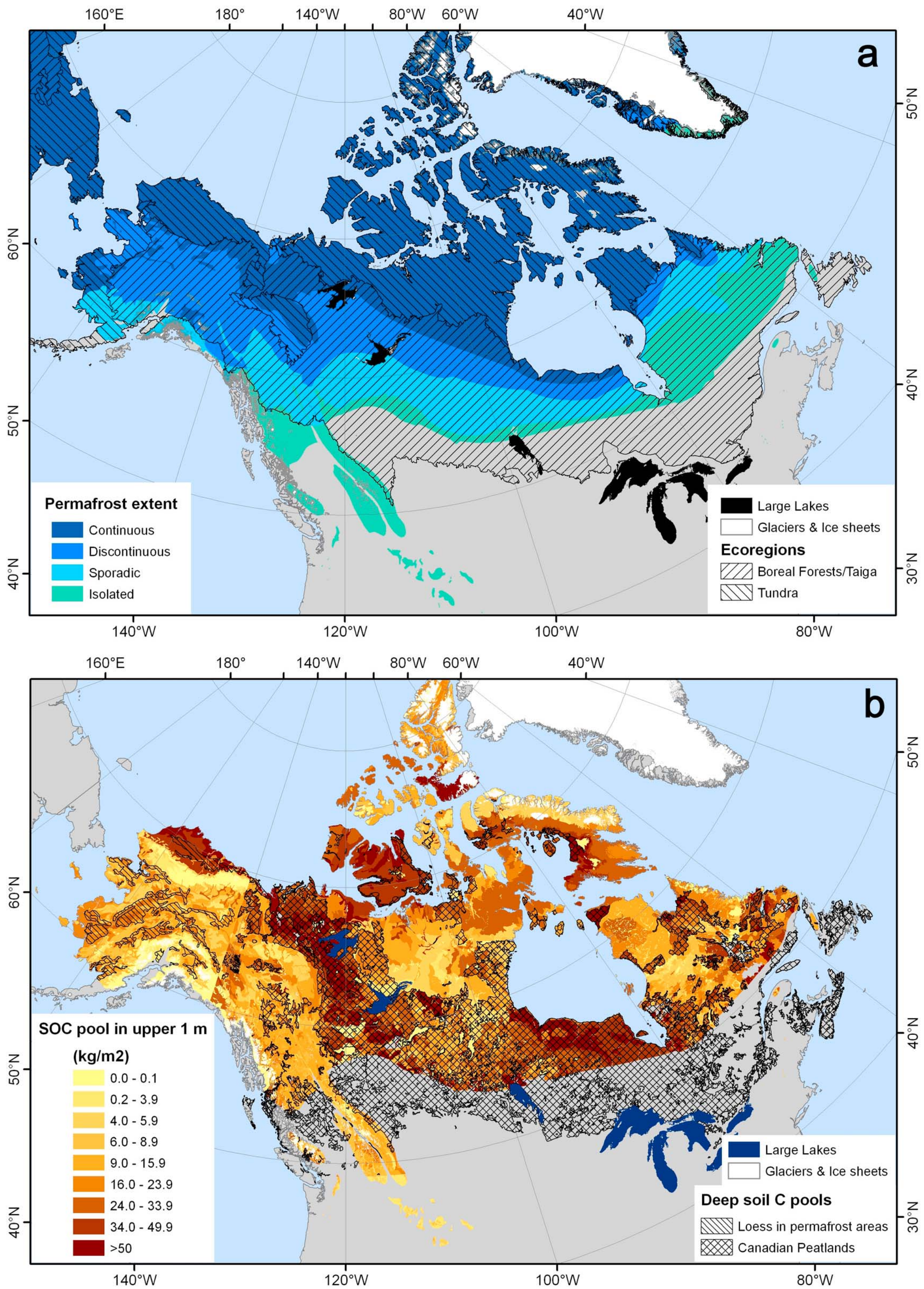

Figure 2. Characteristics of the high latitudes in North America. (a) Permafrost extent [*Brown et al.*, 1997] and tundra and boreal forest ecoregions [*The Nature Conservancy*, 2008]. (b) SOC pools of the North America high-latitude region, showing SOC content in upper 1 m in permafrost regions [*Tarnocai et al.*, 2007a, 2009]; peatlands in Canada [*Tarnocai et al.*, 2004]; loess-paleosoil deposits in Alaska and Canada [*Wolfe et al.*, 2009].

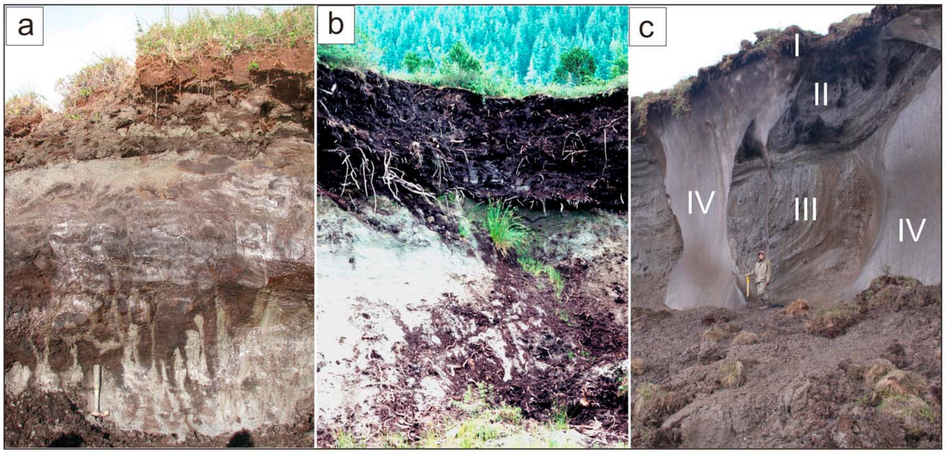

Figure 3. Near-surface SOC pools and deeper organic C pools in peatlands and frozen sediment-paleosoil horizons in North American high latitudes. (a) Strongly cryoturbated soil exposed on a thermokarst lake bluff in a tundra lowland on the Seward Peninsula, Alaska (hammer for scale) (photo: G.G.). (b) About 3 m thick peat exposed at a bluff, Diana Lake in Prince Rupert area, British Columbia, Canada (photo: D. Vitt). (c) Deep carbon-rich permafrost sediments exposed in a thaw slump on the Colville River, Alaska North Slope (person for scale). I, modern peaty soil; II, perennially frozen, ice-rich silt with peat horizons and organic-rich soils; III, perennially frozen, ice-rich sand with peat balls and sedge tussocks; IV, syngenetic ice wedge (photo: G.G.).

relation to climate and both press and pulse disturbances. SOC stocks in the northern high latitudes can be divided into stocks found in near-surface soils (largely consisting of the upper soils down to 1 m depth) and stocks found in deeper soils and stores such as thick peat (up to several m depth) or thick sediment sequences of mineral and organic-rich horizons (up to several tens of m depth). A major distinction in the character of SOC stores in different regions results from the presence or absence of permafrost. Also, while the deeper sedimentary stores are either perennially frozen or perennially unfrozen, portions of the near-surface soils can be fully or partially influenced by seasonal freeze-thaw cycles, resulting in a seasonally thawed layer (active layer) in permafrost regions and a seasonally frozen layer in nonpermafrost regions. Portions of the near-surface soil can also be influenced by seasonally changing moisture status and thus aerobicity. These active portions of the near-surface soils are dynamic and can change in thickness between years. To calculate SOC content and mass, depth categories of 0–30 cm, 0–100 cm, and 0–300 cm are commonly used for the upper C pools.

2.1. Context of Panarctic Northern High-Latitude SOC Stocks

[11] While some initial progress has been made in understanding the distribution of C in northern high-latitude terrestrial ecosystems, substantial uncertainties remain, often related to depth distribution. First, estimates are commonly based on the analysis of organic C in soil profiles, and different analyses consider different depth ranges in assessing SOC stocks. Second, large stores of SOC are found in northern high-latitude peatlands, yet the spatial and depth distributions of peat have not been well quantified. Finally, there are large uncertainties in estimating the substantial SOC that has accumulated in deeper permafrost sedimentary deposits mainly during the last glacial period.

[12] *McGuire et al.* [2009] estimate SOC storage of northern high-latitude terrestrial ecosystems to be between 1400 and 1850 Pg. Another detailed recent estimate of 1672 Pg organic C in soils of the northern circumpolar permafrost region [*Tarnocai et al.*, 2009] falls within this range. It has been estimated that nonpeatland regions in northern high latitudes contain 750 Pg organic C in the upper 3 m of soil [*Schuur et al.*, 2008]. SOC stocks in northern frozen and unfrozen peatlands were recently estimated at 547 Pg [*Yu et al.*, 2010], while peatland C in permafrost regions was estimated at 277 Pg [*Tarnocai et al.*, 2009]. These peatland C estimates account for total peat depth (up to several meters) but not the SOC of underlying mineral soils. According to *Tarnocai et al.* [2009], all permafrost-zone soils estimated to 3 m depth (including peatlands, but with variable depth) contain 1024 Pg organic C, with 818 Pg of that contained in permafrost-affected soils (Cryosols/Gelisols). Another 407 Pg is contained in thick silt-loess-paleosol accumulations below 3 m in Siberia [*Zimov et al.*, 2006], and 241 Pg is estimated for alluvial sediment accumulations below 3 m in river deltas of the seven major arctic rivers [*Tarnocai et al.*, 2009]. Recent data by *Schirrmeister et al.* [2011] suggests that the estimates for the organic C pool in thick silt-loess-paleosols in Siberia may have to be revised downward based on updated carbon and bulk density data. Estimates for deeper (>1 m) organic C pools are poorly constrained at this time, but their inclusion doubles or triples the estimate of permafrost zone C over previous estimates [*Tarnocai et al.*, 2009].

2.2. Near-Surface SOC Stocks in North American High Latitudes

[13] Soils cover approximately 6,211,340 km^2 of the North American permafrost region with approximately 58% of the area being occupied by permafrost-affected soils and the remainder by nonpermafrost soils [*Soil Carbon Database*

Table 1. Average Organic Carbon Content of Soils in the Various Ecological Regions of North America [*Tarnocai*, 1998, 2000]

	Average Carbon Content (kg/m^2)			
	Mineral Soils[a]		Organic Soils (Peatlands)[b]	
Ecological Regions	Perennially Frozen	Unfrozen	Perennially Frozen	Unfrozen
Arctic	49	12	86	43
Subarctic	61	17	129	144
Boreal	50	16	81	134

[a]For the upper 1 m.
[b]For the total depth of the peat deposit.

Working Group, 1993]. Approximately 17% of the total soil area in the North American permafrost region is associated with organic soils (peatlands), the remainder with mineral soils. Figure 2b shows the distribution of various important high-latitude near-surface SOC pools in North America. Large portions (approximately 86% in Canada) of the permafrost-affected soils are Turbic Cryosols/Turbels. In these soils cryoturbation can move surface organic matter into the deeper soil layers through slow freeze-thaw processes over 100s to 1000s of years [*Ping et al.*, 2008a] or abruptly through thawing of ice-rich permafrost after fire [*Swanson*, 1996a] (Figure 3a). In deeper horizons, low soil temperatures, high soil moisture, and in some cases low pH, slow down or restrict decomposition (see section 3 on role of environmental factors for soil formation). As a result, these soils have acted as C sinks over the past millennia.

[14] Organic C storage in the upper meter of high-latitude soils varies by at least tenfold across North American landscapes [*Ping et al.*, 2008b] based on variations in local to regional soil types, soil drainage status, permafrost, climate and disturbance histories. Vegetation and microhabitat also play a key role in spatial variation of SOC stocks, because plant production is the source of organic C input to soil and because plants moderate moisture and heat exchange with the atmosphere. Organic soil layers typically have accumulated over several cycles of climate and disturbance [*Harden et al.*, 2000]. The mean organic C content of permafrost mineral soils (49–61 kg/m^2) is about 4 times as high as in unfrozen mineral soils within the three main high-latitude ecoregions (arctic, subarctic, and boreal) in North America [*Tarnocai*, 2000] (Table 1). Perennially frozen organic soils can have mean organic C contents from 81 to 129 kg/m^2, however unfrozen organic soils in the same eco-region have organic C contents ranging from 43 to 144 kg/m^2 (Table 1).

[15] *Ping et al.* [2008b] estimate that 98.2 Pg C is stored in the upper one meter of soils in the North American treeless Arctic. In a more complete survey for the entire permafrost region of the North America high latitudes, *Tarnocai et al.* [2007a, 2009] estimates the organic C mass of the upper meter of mineral soils to be 104 Pg, with about two thirds of that stored in permafrost soils and one third in unfrozen soils (Table 2). In addition, organic soils add an organic C mass in the upper meter of 61 Pg C that is split equally between perennially frozen and unfrozen landscapes [*Tarnocai et al.*, 2007a, 2009].

2.3. Deeper Soil and Sedimentary SOC Stocks in North American High Latitudes

[16] Large uncertainties in SOC stocks are associated with sparse data on distribution and thickness of deeper sedimentary deposits and their organic C contents in the North America high-latitude region. Major organic C pools are found in thick Holocene peatlands and Pleistocene silt-soil alternations preserved in permafrost (Figure 2b).

[17] The vertical distribution of SOC in permafrost soils varies as a function of soil type (Figure 4). Peatlands tend to have uniformly high SOC concentration with little depth variation, typically extending to the bottom of the peat deposit (Figures 3b and 4). While SOC concentration decreases somewhat with depth, this is typically offset by increases in soil bulk density such that organic C density (kg/m^3) in peatlands tends to be similar for each meter increment deeper in the peat profile. Peatlands in North America cover ~1,372,000 km^2, with permafrost peatlands storing approximately 54 Pg C and nonpermafrost peatlands storing about 125 Pg C [*Bridgham et al.*, 2006].

[18] Mineral soils, which may also contain a thin organic horizon at the surface, tend to decrease in SOC concentration with depth, but often have much higher bulk densities than peat soils (Figures 3c and 4). SOC densities in mineral soils are usually highest in the top meter of soil compared to the second and third meters. Nevertheless, large amounts of SOC can be stored in permafrost soils below 1 m due to cryoturbation [*Bockheim and Hinkel*, 2007; *Ping et al.*, 2008b]. In some unglaciated regions of Alaska and NW Canada, deposits of alluvial, aeolian, and slope sedimentation created paleosol-rich silt layers that are tens of meters thick [*Fraser and Burn*, 1997; *Kanevskiy et al.*, 2011]. These deposits are rich in ground ice and tend to have high SOC density throughout the profile (Figures 3c and 4). In some sites these sequences represent several glacial-interglacial cycles [*Froese et al.*, 2008], indicating long-term SOC sinks in regions where permafrost was less vulnerable to thaw. A first order estimate for deeper C pools in deposits equivalent to

Table 2. Soil Organic Carbon Mass in the North American Permafrost Region [*Tarnocai et al.*, 2007a, 2009]

	Soil Carbon Mass[a] (Pg)						
	Mineral Soils			Organic Soils (Peatlands)			
Depth (m)	Perennially Frozen	Unfrozen	Total	Perennially Frozen	Unfrozen	Total	All Soils, Total
0–0.3	33	16	49	9	9	18	67
0–1.0	75	29	104	31	30	61	165
0–3.0	177	48	225	77	85	162	387

[a]These estimates include Greenland.

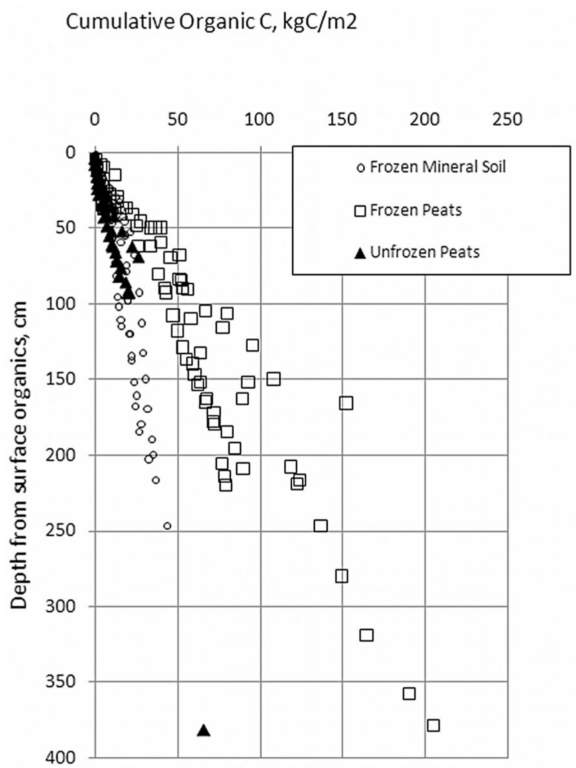

Figure 4. Cumulative carbon densities of three arctic/subarctic soil types as a function of depth. C contents of permafrost and peatland soils can remain high at depths of several meters. Percent organic carbon was multiplied by bulk density (g/cm^3) and by 10 (to get kgC/m^2) for each soil horizon and summed from the surface downward. Data are shown for frozen and unfrozen peat from Manitoba (n = 5 profiles from J. Harden (unpublished data, 2010)), Inuvik (n = 2), Norman Wells (n = 2) Canada [*Tarnocai*, 2009] and for frozen mineral soil from Hess Creek, interior Alaska (n = 4 [*O'Donnell et al.*, 2010b]).

Siberian Yedoma [*Kanevskiy et al.*, 2011] can be based on the distribution of loess-like and retransported silty deposits within the current permafrost regions of northwestern North America (258,000 km^2) (based on *Wolfe et al.* [2009]). This is about one fourth of the estimated area of 1 million km^2 for Yedoma in the Arctic according to *Zimov et al.* [2006], thus, assuming the same average base data for organic carbon content, ice content, and thickness, the North American portion of Yedoma deposits below 3 m depth may contain about 100 Pg C. Recent data by *Kanevskiy et al.* [2011] and so far unpublished data by others indicate that average organic C contents in such deposits strongly vary between different regions in North America, suggesting that the 100 Pg C estimate is still associated with high uncertainties.

3. Factors Controlling Formation of Northern High-Latitude SOC Pools

[19] SOC inputs are strongly dependent on vegetation distribution and structure, which in turn are governed by climatic and hydrological processes (Figure 1). Similarly, storage, decomposition, and C outputs such as dissolved organic carbon export are governed by physical, biological, and chemical soil conditions, hydrology, sedimentation processes, and various types of press and pulse disturbances (Figure 1). Hence, the genesis and character of the current North American high-latitude SOC pool is strongly related to the changing distribution of permafrost and ice sheets during the last glacial/interglacial cycle and general soil-forming conditions in cold environments as determined by climate, hydrology, ecosystem distribution, and landscape dynamics during the Holocene. Understanding these factors and their history in the study region allows better characterization of the distribution of SOC pools and their vulnerability to disturbance.

3.1. Permafrost and Peatland History

[20] During the Late Pleistocene, permafrost in North America was more widespread than today [*Péwé*, 1973]. While the glaciated regions themselves were unsuitable for soil development, many regions outside the glaciated areas developed permafrost soils that accumulated organic C in perennially frozen ground. Soil accretion and syngenetic permafrost growth occurred in areas subject to active aeolian, alluvial, or colluvial sedimentation. Thick paleosol-sediment sequences with syngenetic ice wedges and high ground ice and organic C content accumulated in parts of Alaska [*Kanevskiy et al.*, 2011] and NW Canada [*Fraser and Burn*, 1997]. Soon after the Last Glacial Maximum (LGM, ~26.5–19 ky ago), the Laurentide Ice Sheet retreated [*Dyke et al.*, 2003] and permafrost started to thaw rapidly at its southernmost limits. Deglaciation of the Laurentide Ice Sheet resulted in fresh exposure of large land areas, effectively increasing the sequestration of SOC in northern high-latitude soils due to new mineral and organic soil formation in these formerly ice-covered regions [*Harden et al.*, 1992; *Gorham et al.*, 2007]. By the time of the Holocene Thermal Maximum (depending on region between 5 and 11 ky BP [*Kaufman et al.*, 2004]), permafrost completely disappeared from large regions of the continent, including some northern high-latitude regions [*Péwé*, 1973]. Several relatively cold periods during the mid and late Holocene, of which the most recent and perhaps coldest was the Little Ice Age (LIA, 1650–1850 AD), resulted in formation of new permafrost in some parts of the landscape within the present day isolated to discontinuous permafrost zones. During the LIA, shallow permafrost formed in places that were predominantly unfrozen for most of the Holocene [*Zoltai*, 1995]. Recent warming has initiated thaw of this LIA permafrost, as identified through radiocarbon dating of plant materials associated with recent thermokarst features [*Jorgenson et al.*, 2001; *Turetsky et al.*, 2007]. While we highlight some broad trends above, permafrost history is greatly complicated by geomorphic processes associated with glacial, eolian, and colluvial activity [*Reyes et al.*, 2010], with large climatic changes associated with glacial and interglacial periods [*Froese et al.*, 2008], and with powerful feedbacks involving vegetation and standing water [*Jorgenson et al.*, 2010]. Such processes have allowed permafrost to persist in some sites within the discontinuous permafrost zone for hundreds of thousands of years [*Froese et al.*, 2008].

[21] The genesis of northern high-latitude peatlands relates primarily to spatial and temporal development of

deglaciation, permafrost, and lake and wetland distributions. Peatland initiation in North America started slowly with the end of the LGM and peaked around 7–8 ky BP, when about 90% of the landscape was deglaciated [*Gorham et al.*, 2007] and climatic and hydrological conditions were favorable for peat growth in many regions of North America [*Jones and Yu*, 2010; *Camill et al.*, 2009]. The colonization of drained lakebeds by vascular plants and mosses played a particularly important role in areas where glacial ice-dammed lakes deposited clayey impermeable sediments, then subsequently drained, and provided a wet environment for sustained peat growth [*Kratz and DeWitt*, 1986]. Permafrost peatlands in boreal and subarctic regions also reflect formation by paludification, or the landward encroachment of peat, which has occurred since the mid Holocene [*Sannel and Kuhry*, 2008; *Zoltai and Vitt*, 1990; *Korhola et al.*, 2010]. On the Canadian Arctic islands and in Alaska, many peatlands initiated before 8.6 ky BP [*Tarnocai*, 1978; *Jones and Yu*, 2010], but in general, C uptake by North American peatlands was highest during the late Holocene [*Korhola et al.*, 2010; *Harden et al.*, 1992]. In the southern portions of North America's high latitudes, permafrost aggradation occurred after peatland initiation, as most peatlands at 6 ka were nonpermafrost fens [*Zoltai*, 1995]. Further north, however, Holocene peat accumulation has occurred under permafrost conditions [*Zoltai*, 1995].

3.2. Northern High-Latitude SOC Formation and Role of Permafrost, Hydrology, Vegetation, and Topography

[22] Northern high-latitude soils range from unfrozen to frozen states and can be mineral (<20% organic C by weight) or organic (>20% organic C by weight) soils. For unfrozen soils and noncryoturbated frozen soils in the permafrost region organic matter is deposited on the soil surface or derived from roots penetrating the soil. Some soluble organic matter moves downward with unsaturated flow, but because these soils are not affected by cryoturbation, movement of particulate organic matter from the surface into the deeper soil layers is limited. Most belowground organic C of noncryoturbated soils may originate from roots and residence times are relatively short [*Tarnocai et al.*, 2007b]. Nevertheless, unfrozen mineral soils of northern high-latitude regions can still store vast amounts of SOC in the uppermost soil.

[23] Cryoturbated soils, by contrast, can have organic horizons intermixed with mineral soil at depth. *Cryoturbation*, or the mixing of soil layers in response to repeated freeze-thaw cycles, redistributes organic C away from the surface to greater depth and incorporates organic C into permafrost over periods of 100s to 1000s of years [*Bockheim*, 2007; *Ping et al.*, 2008a]. Organic C-rich horizons are typically at or below the thin active layer (few tens of cm) that thaws annually during the short summer growing season. Syngenetic permafrost growth with ongoing sedimentation incorporates SOC into permafrost over periods of 100s to 10,000s of years. A number of recent studies note the importance of cryoturbation or sedimentation for formation of deeper SOC stocks in northern high latitudes [*Bockheim and Hinkel*, 2007; *Ping et al.*, 2008a; *Tarnocai et al.*, 2009].

[24] In deep soils of peatlands, SOC storage is typically very high because productivity has exceeded decay over long periods of time. The water-saturated conditions, low soil temperatures often including permafrost, and acidic conditions of northern peatlands provide an environment in which decomposition is limited. The litter is converted to peat and preserved in the organic soil. This gradual accumulation has been ongoing often for millennia, resulting in peat deposits that are an average of 2–3 m thick and, in some cases, up to 10 m thick. Organic soils (peat) have acted as very effective C sinks for many thousands of years. The gradual buildup processes in both unfrozen and permafrost-affected organic soils are similar, resulting in similar soil organic carbon contents [*Tarnocai et al.*, 2007b].

[25] In many regions the character and spatial distribution of soil types, permafrost, and soil carbon are controlled by landscape physiography and soil texture [*Brown and Péwé*, 1973; *Harden et al.*, 2006]. In interior Alaska's discontinuous permafrost, coarse-textured materials, convex and steep slopes, and south facing aspects are associated with warm, dry mineral soils, while the driest, well-drained soils that are located on convex, upper-slope positions often contain no permafrost [*Swanson*, 1996b]. In contrast, sites with fine-textured materials, concave and gentle slopes, and north facing aspects exhibit cold and wet soils. SOC distribution shows strong correlation with these landscape characteristics, with decreased mineralization and substantial carbon accumulation in colder, wetter soils [*Ping et al.*, 2005; *Harden et al.*, 2006].

[26] Large northern high-latitude SOC pools are generally a result of slow decomposition and low outputs rather than high inputs. The main limitations to SOC decomposition in northern high-latitude soils are found in abiotic factors and the quality (relative decomposability) of plant C inputs. The two most important abiotic constraints include low temperature (including below freezing) that limits decomposition rates, and high water saturation that leads to oxygen limitation to microbes, lower aerobic decomposition, and lower combustion during wildfires. Input from nonvascular plants is relatively high in many ecosystems in the northern high latitudes, providing poorly decomposable plant litter [*Hobbie*, 1996; *Turetsky et al.*, 2010]. The large accumulation of SOC in peatlands reflects a combination of anaerobiosis and slow decomposition, structural recalcitrance of vegetation, i.e., *Sphagnum*, and retardation of combustion during wildfires caused by inundated soils and mosses with high water retention [*Shetler et al.*, 2008; *Turetsky et al.*, 2010; *Blodau*, 2002]. Permafrost presence in peatlands also strongly affects vegetation structure and soil drainage characteristics, and thus SOC accumulation rates [*Robinson and Moore*, 2000].

[27] Complex feedbacks of these environmental factors to SOC input, storage, and output result from their interaction with each other and their general variation along climatic, ecologic, and successional gradients over different spatial and temporal scales [*Walker et al.*, 2008; *Jorgenson et al.*, 2010]. For example, in the discontinuous permafrost zone with ground temperatures near the freezing point the presence of vegetation and organic soil layers strongly influence the ground thermal regime due to insulation effects of thick organic layers. SOC lability in permafrost likely varies according to the rates and timing of C burial/exposure. Fresh C substrates that are enclosed by syngenetic, aggrading permafrost are likely the most labile, whereas substrates

subjected to repeated fire or freeze/thaw cycles are likely somewhat more recalcitrant. Data from Alaska shows that permafrost soils have similarly or even more labile organic matter compared to samples from the active layer of the same site [*Waldrop et al.*, 2010]. Nevertheless, even soils associated with discontinuous permafrost in the boreal forest have relatively labile forms of SOC in deeper organic layers [*Fan et al.*, 2008]. For peatland soils, the form of SOC is more variable according to water table, freeze/thaw and combustion histories, and associated plant assemblages, and can be highly humified with a lower decomposition potential.

[28] A number of studies in interior boreal Alaska suggest that the establishment of deep, cold organic layers in wetter, poorly drained sites may result in a self-perpetuating system of resistance to disturbance [*Ping et al.*, 2005; *Harden et al.*, 2006]. Nevertheless, both hydrologic and ground thermal regimes are especially sensitive to disturbances.

4. Disturbance Processes Affecting SOC in High Latitudes of North America

[29] In this section we characterize key disturbances according to our earlier definition of press and pulse disturbances, and we analyze their impact on input, storage and output of SOC. We note that two main drivers for both disturbances and C dynamics are temperatures in and above the ground, and water in the form of soil moisture, ground ice, or as transport agent (Figure 1). All disturbances characterized in the following sections are strongly related to either temperature or moisture, or both.

4.1. Key Press Disturbances

[30] Press disturbances impacting SOC in a warming climate include vegetation change, soil warming, and hydrological change (Figure 1). Gradual vegetation changes such as expansion of shrubs in arctic tundra impact organic C inputs and thermal insulation of soils. Soil warming, even without permafrost thawing, is expected to have impacts on biological processes (enzyme kinetics) and plant and microbial communities, but warming also enhances release of nitrogen (N) from mineralization and can contribute to increased plant production. For example, results from long-term fertilization experiments at Toolik Lake, Alaska, suggest that increased nutrient mineralization arising from warming-induced decomposition of SOM could lead to a significant loss of soil carbon from deeper layers, while offsetting some of this loss with an increase in aboveground biomass and fine roots [*Mack et al.*, 2004; *Nowinski et al.*, 2008].

[31] The coupling of warming with hydrological change also introduces adjustment to many soil and plant processes. Hydrological changes related to evapotranspiration, precipitation, storage, and runoff strongly interact with the soil thermal regime to affect plant and microbial communities, peat growth, cryoturbation, soil aerobicity, and export of dissolved organic C. The state of water, and its freeze-thaw transitions, has profound effects on abundance and activity of microbial communities (bacteria, fungi, and Archaea), and the rates and processes by which they decompose SOC [*Yergeau et al.*, 2010; *Waldrop et al.*, 2010]. Differences in microbial activity for perennially frozen (permafrost) versus seasonally unfrozen (active layer) soil horizons generally cannot be attributed to lower SOC concentrations or SOC quality in permafrost, but could be due to the difficulty of surviving in a frozen environment for long periods of time [*Waldrop et al.*, 2010; *Yergeau et al.*, 2010]. Warming of soils and extension of the seasonal thawing period can lift this limitation in microbial activity and thus enhance SOC decomposition.

[32] Coastal and fluvial erosion of soils could be classified as press disturbances that likely will accelerate as sea ice cover, sea surface temperature, and storminess increase and hydrological runoff and storage patterns fundamentally change (Figure 1). Coastal erosion rates are especially high in the northwest American Arctic, where permafrost soils exposed along coastal bluffs are rapidly eroded with rates sometimes exceeding 20 m/yr [*Jones et al.*, 2009], resulting in substantial soil-to-ocean C fluxes such as the total flux of 1.8×10^5 Mg C/yr from coastal erosion along the Alaskan Beaufort Sea coast alone [*Jorgenson and Brown*, 2005]. An increase in storminess could even trigger more frequent individual strong erosion events, or the breaching and submergence of low-lying thermokarst basins along the coast, both of which could be considered a pulse disturbance to the SOC stored there. High-latitude fluvial dynamics are strongly connected to changes in precipitation and cryosphere characteristics [*McNamara and Kane*, 2009], and can affect erosion of soils in river floodplains and fluvial export of SOC [*Guo et al.*, 2007].

[33] Two of the most important press disturbances for SOC in northern high latitude are top-down permafrost thawing and landscape drying and wetting due to hydrological disturbances.

4.1.1. Permafrost Warming and Widespread Top-Down Thawing

[34] Long-term thawing of permafrost starts when the seasonal active layer does not refreeze completely even during the most severe winters. At this tipping point, permafrost will thaw from top down, thereby fundamentally changing the environmental regime of soils and their organic C stocks. The switch from perennially frozen to perennially unfrozen soil can facilitate major changes in SOC inputs (vegetation composition and production; soil rooting), storage, and losses (decomposition, export of dissolved carbon, combustion). In addition, processes such as cryoturbation and oxygenation will affect translocation and transformation of SOC as well as the potential sources of greenhouse gases. Ground temperatures [*Romanovsky et al.*, 2010] and active layer depths [*Nelson et al.*, 2004] are considered the best direct indicators for detecting permafrost thawing. A steady increase in the active layer thickness allows remobilization of root biomass and SOC and a more active decomposition. Continued thawing of permafrost with high ice content can trigger and increase the frequency of related pulse disturbances such as thermokarst, thaw slumps, or active layer detachments and thus involve a more complex and spatially heterogeneous response to warming (see section 4.2 on thermokarst) (Figure 1). Permafrost temperatures monitored in boreholes in North America show a warming trend similar to other arctic regions during the last decades [*Romanovsky et al.*, 2010; *Smith et al.*, 2010]. The strongest temperature increases were observed in cold, high arctic continuous permafrost regions. In contrast, warming

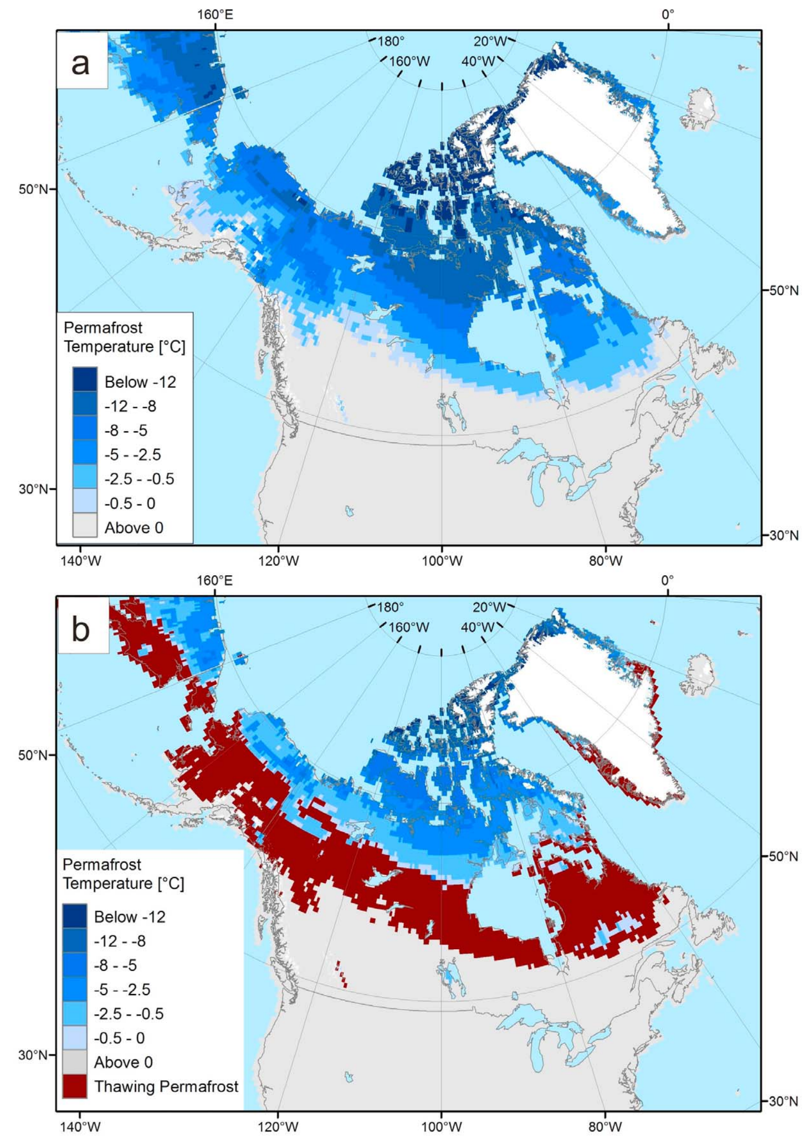

Figure 5. Permafrost model result for North America. (a) Average distribution of permafrost and ground temperatures for the period 1981–2000; (b) for the period 2081–2100 (model by S. Marchenko, in the work of *Romanovsky et al.* [2007]). See also Table 3 for Cryosol areas and near-surface SOC stocks affected.

was slower in warm, discontinuous permafrost already close to 0°C. Permafrost represented by fine-grained ice-rich sediment showed generally smaller warming rates because of latent heat effects of the thawing of constituent ground ice in a range of negative (°C) temperatures [*Romanovsky et al.*, 2010]. Local conditions and snow influence the response of permafrost temperature to air temperature changes [*Jorgenson et al.*, 2010]. For example, increase in snow depth at northern arctic sites resulted in stronger increases in permafrost temperatures, whereas lower snow cover in interior and southern Alaska during recent years resulted in a slowing of permafrost temperature increase. In the absence of pulse-response disturbances, modeling results of the temporal and spatial transient response of permafrost to projected climate changes suggest significant changes in near-surface permafrost within this century, with vast territories of present-day discontinuous permafrost actively thawing [*Romanovsky et al.*, 2007; *Lawrence et al.*, 2008; *Zhang et al.*, 2008; *Marchenko et al.*, 2008] (Figure 5). Field studies corroborate these modeled trends [*Halsey et al.*, 1995; *Payette et al.*, 2004; *Thibault and Payette*, 2009]. According to our permafrost modeling scenario [*Romanovsky et al.*, 2007], Holocene permafrost in Northern America may be actively thawing at all locations and Late Pleistocene permafrost could start to thaw at some locations by the end of the 21st century (Figure 5). Based on these model projections the area of Cryosols in zones of actively thawing surface permafrost in North America will be about 385,000 km^2 by 2050 and 1,132,000 km^2 by 2100 (Table 3). Cryosol organic C stocks in the upper 1 m in these areas affected by permafrost thawing are estimated to be 10 Pg and 29 Pg C, or 10% and 29%, respectively, of the total North American Cryosol SOC pool in the upper 1 m (100 Pg)

Table 3. Soil Organic Carbon Stocks in Upper 1 m of Cryosols Impacted by Permafrost Thawing in North America by 2050 and 2100[a]

	Current Extent (~2000)[b]	Zone of Thawing Surface Permafrost by 2050[c]	Zone of Thawing Surface Permafrost by 2100[c]
Cryosol area[b,d]	680,000 km^2 (AK); 2,645,000 km^2 (CAN); 3,324,000 km^2 (total)	224,000 km^2 (AK); 161,000 km^2 (CAN); 385,000 km^2 (total)	387,000 km^2 (AK); 745,000 km^2 (CAN); 1,132,000 km^2 (total)
Total SOC mass for the upper 1 m in these Cryosols[b,e]	17 Pg C (AK); 83 Pg C (CAN); 100 Pg C (total)	4 Pg C (AK); 6 Pg C (CAN); 10 Pg C (total)	6 Pg C (AK); 23 Pg C (CAN); 29 Pg C (total)

[a]AK, Alaska; CAN, Canada.

[b]Based on Cryosol area and SOC mass provided in Circum-Arctic Soil Organic Carbon Map [*Tarnocai et al.*, 2007a].

[c]Zone of thawing permafrost based on projection with the GIPL permafrost model at 0.5° by 0.5° latitude/longitude resolution (model by S. Marchenko, in the work of *Romanovsky et al.* [2007]). Surface permafrost thawing is defined here as reaching a seasonal thawing depth in excess of 2 m depth. Areas were calculated by clipping GIS-polgyons of the target layers with a polygon derived from all 0.5 by 0.5 degree grid cells characterized by thawing permafrost at a time slice.

[d]Summary of all permafrost-affected soils in the North American permafrost zones; Cryosol area for each polygon was calculated by multiplying the area of a landscape polygon with the percentage cover of Cryosols occurring in this polygon area.

[e]Summary of SOC mass for all Cryosols; SOC mass for each polygon was calculated by multiplying area of Cryosols with SOC content for the upper 1 m for each polygon.

(Table 3). Depending on a number of topographic, hydrologic, and biotic conditions, soils and their carbon pools in these areas are likely to be affected in various ways by this thawing. Using a more advanced, higher resolution model of permafrost thaw and taking in to account deeper peatland C pools, *Wisser et al.* [2011] estimate that about 670 km^3 of peat soils containing 33 Pg C, and representing about 20% of all peat in Alaska and Canada, will be seasonally thawed by 2100.

[35] The impact of top-down thaw on the carbon balance likely involves a hydrologic response (see section on thermokarst below), but active layer thickening alone enables deeper rooting, enhanced C and N mineralization, and seasonal changes to plant available moisture. For example, over a fire cycle (see section 4.2 on disturbances by wildfire) with a transient deepening of the active layer, near-surface mineral soils are significantly warmer both in summer and in winter and their carbon turns over at much higher rates than preburn soils [*Kane et al.*, 2005]. C sources of decomposition are significantly enhanced by deepening of the active layer [*Schuur et al.*, 2009; *O'Donnell et al.*, 2010a]. In lowlands and frozen peats, widespread top-down thawing of permafrost and development of thermokarst often leads to paludification and an increase in peat accumulation rates, and thus enhanced C sequestration [*Hinkel et al.*, 2003; *Turetsky et al.*, 2007; *Camill et al.*, 2001], but also stimulates methane emissions [*Liblik et al.*, 1997; *Wickland et al.*, 2006]. Despite changes in peat accumulation rates, however, *Johansson et al.* [2006] concluded that increased methane emissions following permafrost thaw increased radiative forcing by 47% on a 100 year time horizon. Whether permafrost thaw in peatlands leads to a positive or negative feedback to atmospheric greenhouse gas concentrations will depend on spatial and temporal variation in water table position as well as soil temperature, which both serve as important controls on methane release in peatlands.

[36] Permafrost degradation can alter drainage conditions in both directions, affect hydrologic pathways, deepen water tables and increase groundwater contribution to streamflow, and result in increased mineralization or sorption of dissolved organic carbon and shifts in the relative exports of dissolved organic and inorganic carbon [*Walvoord and Striegl*, 2007; *Frey and McClelland*, 2009; *Keller et al.*, 2010]. Dissolved organic carbon dynamics are also affected by permafrost thaw through changes in C sources and sinks. Permafrost horizons can release as much dissolved organic carbon upon thaw as overlying active layer horizons [*Waldrop et al.*, 2010], depending on SOC content. However, dissolved organic carbon released from newly thawed soils may be readily metabolized or sorbed to mineral soils [*Kawahigashi et al.*, 2006], with either no net change or a relative decrease in dissolved organic carbon transport to surface waters [*Striegl et al.*, 2005].

4.1.2. Hydrological Disturbances of High-Latitude SOC in North America

4.1.2.1. Drying Landscapes

[37] A number of remote sensing analyses of surface hydrology suggest that landscape dryness has increased over the past several decades in the North American boreal and some tundra regions [*Yoshikawa and Hinzman*, 2003; *Riordan et al.*, 2006; *Goetz et al.*, 2005; *Verbyla*, 2008]. While one study suggests links between shrinking ponds and permafrost thawing in discontinuous permafrost [*Yoshikawa and Hinzman*, 2003], others have related increased landscape drying to a decreasing water balance due to changes in evaporation and precipitation in a warming and drying climate (Figure 6a). Ground-based studies corroborate landscape drying resulting from an increase in evaporation over precipitation [*Hinzman et al.*, 2005; *Smol and Douglas*, 2007].

[38] Increased warming and drying impacts vegetation, and thus affects soil C inputs. While a longer and warmer growing season potentially benefits vegetation growth, increased summer water deficits reduce growth rates [*Barber et al.*, 2000]. In some cases tree growth can respond positively or negatively depending on landscape position [*Wilmking et al.*, 2004]. Multiyear eddy covariance studies in boreal forests highlight the strong interactions among soil moisture, temperature, and vegetation type in governing net ecosystem production and ecosystem respiration processes [*Welp et al.*, 2007; *McMillan et al.*, 2008]. Water stress not only suppresses photosynthesis but also can suppress ecosystem respiration by limiting plant growth and/or suppressing microbial respiration [*Gaumont-Guay et al.*, 2006]. *Oquist et al.* [2009] and *Wickland and Neff* [2008] documented the strong effects of moisture availability on the temperature sensitivity of soil CO_2 production in northern soils. An increase in landscape dryness to

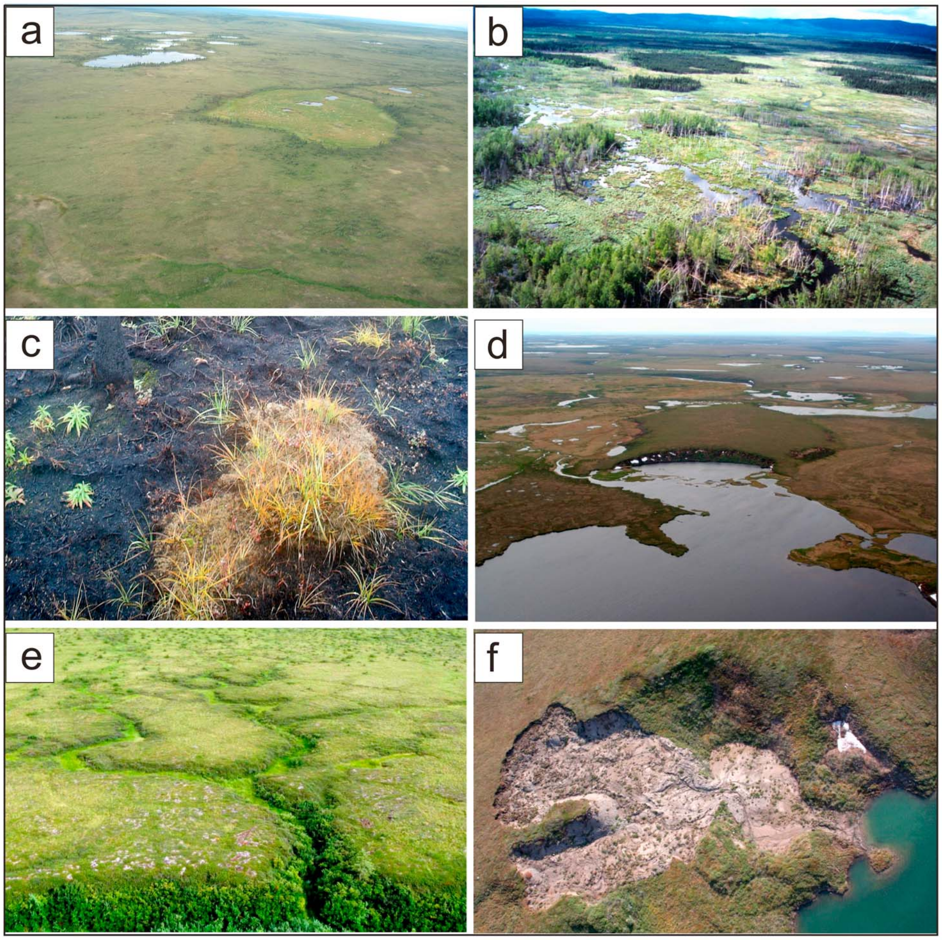

Figure 6. Common disturbances affecting SOC in northern high latitude. (a) Drying lakes; (b) thermokarst resulting in fens and bogs; (c) fires combust surface organic layers and have a long-term impact on the ground thermal regime; (d) thermokarst lakes; (e) thermoerosional gullies; (f) retrogressive thaw slump associated with glacial thermokarst (photographs: Figure 6a, G.G.; Figure 6b, M.T.J.; Figure 6c, S.M.; Figures 6d–6f, M.T.J.).

some extent can promote greater release of CO_2 through enhanced decomposition and will result in lower CH_4 emissions due to lower rates of methanogenesis and enhanced methane oxidation [*Billings et al.*, 2000].

[39] In northern peatlands and wetlands, water table position is an important factor influencing vegetation structure and productivity. A declining water table position can either decrease or increase Gross Primary Productivity (GPP) depending on whether it triggers moisture stress on photosynthetic processes or stimulates the productivity of woody species or belowground productivity [cf. *Griffis et al.*, 2000; *Weltzin et al.*, 2000]. However, vegetation and associated carbon responses to drying often depend on microform, as hummock and hollow species often can respond to and recover from drought differently [*Strack and Waddington*, 2007]. An experimental drought study of a fen ecosystem in Alaska found reduced C inputs to soil, primarily due to declines in vegetation productivity, whereas an increased water table caused net C storage [*Chivers et al.*, 2009]. The "Enzyme Latch Hypothesis" predicts that lower water table position in peatlands will trigger the loss of soil C by releasing oxygen constraints on phenol oxidase activity [*Freeman et al.*, 2001]. However, results from experimental and natural drying studies deliver mixed results so far, some measuring net C loss, and some showing no relation between water table and GPP [*Chivers et al.*, 2009; *Lafleur et al.*, 2005].

[40] Peatlands in some situations may be resistant to drought due to the high water holding capacity of some moss species, particularly some *Sphagnum*. Lower water table positions during drought can trigger subsidence of the surface peat layer, which reduces acrotelm thickness and inhibits further water loss and soil respiration. While short-term responses in plant physiology and photosynthetic water relations will be important in driving short-term C responses to drying, ecosystem succession and changing plant community structure with sustained drying will have strong influences on C inputs and soil transformations on decadal or longer timescales. Last, widespread drying of landscapes will lead to decreased export of dissolved organic carbon from terrestrial systems to surface waters, as dissolved organic

carbon export is limited mostly by water availability rather than by C source [*Striegl et al.*, 2007].

4.1.2.2. Inundating Landscapes

[41] While drying is likely to occur throughout many parts of the North American boreal region due to increased summer water deficits, landscapes can experience inundation in areas with strong topographic relief such as interior Alaska due to increased upwelling of meltwater from adjacent uplands [*Jorgenson and Osterkamp*, 2005] or to local thawing of permafrost in lowlands (Figure 6b). An increase in landscape wetness will tend to promote increased CH_4 emissions through enhanced methanogenesis [*Myers-Smith et al.*, 2007; *Turetsky et al.*, 2008a] while slowing CO_2 release because of anaerobic conditions that hinder decomposition. Semi-aquatic mosses such as those found in peatland hollows tend to have higher net primary production rates than hummock *Sphagnum* or feather moss species [*Turetsky et al.*, 2010]. However, these fast growing moss species also tend to produce more labile biomass that decomposes more quickly than hummock mosses [*Turetsky et al.*, 2008b]. Thus, increases in aquatic or semi-aquatic groundcover may be transient for ecosystem C storage [*Camill et al.*, 2009].

4.2. Key Pulse Disturbances

[42] Pulse, or event-based, disturbances, are sensitive to changes in overall climate but are fundamentally complex in their spatial and temporal distribution, with the total net effect resulting from numerous individual events of local nature. Wildfires and rapid permafrost thawing from thermokarst are two pulse disturbances that significantly affect C budgets of North American high latitudes (Figure 1).

4.2.1. Wildfire Impacts: Combustion of SOC and Permafrost Thawing

[43] Significant numbers and areas of wildland fires occur in North America's boreal forest regions [*Kasischke et al.*, 2010; *Flannigan et al.*, 2009], many of which are underlain by discontinuous to sporadic permafrost. Wildfire plays an important role in the interannual and decadal scale variability of C source/sink relationships in the boreal region [*Balshi et al.*, 2007], and in some regions has been associated with changes in peat accumulation rates and long-term SOC storage [*Robinson and Moore*, 2000]. Patterns of burning and successional recovery are in part controlled by the presence and depth of peat layers and permafrost [*Mack et al.*, 2008; *Harden et al.*, 2003] and the topographic patterns controlling soil temperature and moisture [*Kane et al.*, 2007; *Turetsky et al.*, 2011]. Previous studies have shown that typical fire cycles in North America's upland boreal forests are about 100 years in length [*Turetsky et al.*, 2011]. Estimates for fire return intervals obtained from charcoal reconstructions in peat as well as remote sensing range from 120 to 1100 years in continental peatlands [*Kuhry*, 1994; *Zoltai et al.*, 1998]. Recent studies, however, indicate a shift in fire regimes that include a doubling of mean annual burn area, increased frequency of large fire years, and an increase in burning late in the fire season over the past ~60 years [*Kasischke et al.*, 2010; *Kasischke and Turetsky*, 2006]. Such changes in annual burn area in Canada have been related to anthropogenic climate change [*Gillett et al.*, 2004], and fire models predict increases in burn area and severity in some North American boreal regions with implications for C emissions and SOC stocks [*Amiro et al.*, 2009].

[44] In general, fires release soil and plant organic C to the atmosphere immediately through combustion, and over the longer-term reduce autotrophic respiration and alter heterotrophic respiration by influencing soil climate, substrate quality, and microbial populations [*O'Neill et al.*, 2006; *Mack et al.*, 2008]. The depth of combustion of organic soils is dependent upon many factors, including fire behavior, fuel moisture conditions, and organic soil characteristics including bulk density and ash content [*Miyanishi and Johnson*, 2002; *Benscoter et al.*, 2011]. Vegetation can also have important effects on organic soil combustion rates, for example by influencing fuel moisture and organic soil properties. Precipitation interception by tree crowns can influence organic soil combustion [*Miyanishi and Johnson*, 2002]. *Sphagnum* moss species, particularly those that dominate hummock microforms, tend to inhibit combustion due to high soil moisture retention, leading to the protection of thick peat layers even after severe fire activity [*Shetler et al.*, 2008]. *Turetsky et al.* [2011] summarized empirical estimates of depth of organic soil combustion in both uplands and peatlands in Canada and Alaska. As has been found at landscape scales [*Kane et al.*, 2007; *Boby et al.*, 2010], mean prefire organic soil depth was a strong predictor of mean depth of burn across these studies. Prefire organic soil depths were higher in studies published for Alaskan upland forests compared to Canadian forests (AK: 19.7 cm; CAN: 15.9 cm), which also corresponded to higher mean depth of burn and SOC losses from Alaskan than from Canadian forests (AK: 11.7 cm and 2.88 kg C/m^2; CAN: 8.3 cm and 1.65 kg C/m^2). In general, few studies have quantified SOC losses from peatlands [see *Turetsky et al.*, 2011].

[45] Combustion of vegetation, dead woody fuels, and the soil organic layer directly impacts C budgets. Characteristics of burning and the magnitude of C emissions depend on vegetation fuel type, structure, and fuel moisture as well as burning conditions (e.g., fire weather), all of which lead to variations in fire extent and severity [*French et al.*, 2011]. Terrain topography strongly influences burn area and burn severity [*French et al.*, 2008]. In general, substrate and microbial communities are altered by the burn and the depth of burn, resulting in either enhanced or reduced decomposition immediately after burning. In the pervasive black spruce forests of boreal and subarctic North America, fires can penetrate into any of a number of organic layers that are typically comprised of moss detritus, fibrous root litter, or decomposed organics that are denser and less porous. Last, the successional patterns of stand age reflect the process of burning (which reduces the organic layers) and regrowth (which allows them to reaccumulate). Thus, fires alter both soil and vegetation in a succession pattern that plays out for decades and is reflected spatially in stand age distributions.

[46] The radiative surface energy balance of an area also is impacted by fire. Initially, summertime surface reflectivity (albedo) of the terrain is reduced by the conversion of the surface material to dark char [*Chambers et al.*, 2005] (Figure 6c). Postfire vegetation recovery subsequently produces a general increase in albedo, which may persist for several decades due to the higher reflectance of early successional vegetation types compared to prefire forest types [*Randerson et al.*, 2006]. Wintertime albedo at these sites is much higher for many years due to snow cover in a site with low-stature vegetation, especially in springtime when solar inputs are relatively high and

Table 4. Distribution of Cryosols and Their SOC Stocks in North American Regions Characterized by Medium or High Ground Ice Content[a]

	Medium Ground Ice Content[b]	High Ground Ice Content[b]
Area with Cryosols[c]	406,000 km^2 (AK); 308,000 km^2 (CAN); 714,000 km^2 (total)	73,000 km^2 (AK); 814,000 km^2 (CAN); 887,000 km^2 (total)
Total SOC mass for the upper 1 m in these Cryosols[d]	9 Pg C (AK); 13 Pg C (CAN); 21 Pg C (total)	4 Pg C (AK); 30 Pg C (CAN); 33 Pg C (total)

[a]These areas and C stocks are potentially highly vulnerable to thermokarst and thermo-erosion. AK, Alaska; CAN, Canada.
[b]Ground ice distribution is based on the Circum-Arctic Map of Ground Ice and Permafrost Conditions [*Brown et al.*, 1997]. In this map, medium ground ice content is defined 10–20% ice by soil volume, and high ground ice content is >20% (for lowlands) or >10% (for mountains and plateaus).
[c]Cryosol distribution is based on the Circum-Arctic Soil Organic Carbon Map [*Tarnocai et al.*, 2007a].
[d]SOC mass is based on soil organic carbon contents in the upper 1 m as mapped in the Circum-Arctic Soil Organic Carbon Map [*Tarnocai et al.*, 2007a].

snowpack is still deep. As vegetation recovers, the snow covered burn site slowly darkens as the vegetation height reaches above the annual snowpack. Studies in the boreal black spruce vegetation types, where annual changes in energy exchange have been measured, have shown that fire lowers net radiation annually [*Liu et al.*, 2005]. In addition to influences on albedo, which directly effects net radiation at the surface, the modification of the vegetation structure from fire modifies evapotranspiration and surface roughness, which along with net radiation impact the sensible and latent heat fluxes [*Liu et al.*, 2005]. Fire reduces the vegetation cover and surface organic layer, thereby changing surface insulation properties and evapotranspiration regimes, and altering soil density, heat capacity, and thermal conductivity [e.g., *Viereck*, 1982; *Yoshikawa et al.*, 2003; *Kasischke and Johnstone*, 2005; *O'Donnell et al.*, 2009]. These parameters are all relevant for the subsurface thermal regime and for soil C input and decomposition over the burn cycle.

[47] The nonlinear impacts of fires on active layer and permafrost dynamics range from no effect, to short-term disturbance, to severe long-lasting permafrost degradation. The stability of permafrost after such disturbance depends upon burn severity, but also on permafrost properties. Wildfires can cause active-layer deepening and a shallow talik formation in the years following the disturbance. In ice-rich permafrost rapid mass wasting and thermo-erosion can occur immediately after the disturbance, mobilizing old organic carbon stored in deeper permafrost layers [*Liljedahl et al.*, 2007; *Lacelle et al.*, 2010].

[48] Permafrost recovery is also sensitive to revegetation and regeneration of insulating layers, both of which in turn are a function of preexisting vegetation, slope characteristics, fire characteristics, damage to organic layers, and timing of the fire [*Racine et al.*, 2004]. A fire-caused increase in active layer depth also can lead to an increase in plant rooting depths, enhancing C input to soils [*Yoshikawa et al.*, 2003; *Jorgenson et al.*, 2010]. Prefire and postfire variations in soil moisture influence both the carbon and thermal budgets that occur in relation to fire and succession. Thermal conductivity of soils is related to soil moisture and its physical state [*Yoshikawa et al.*, 2003; *O'Donnell et al.*, 2009]. In peat plateaus, which are generally dry above the permafrost table, increase in wildfires may result in decreases in peat formation [*Sannel and Kuhry*, 2008; *Robinson and Moore*, 2000]. However, if there is a loss of ground ice after burning, surface subsidence can result in impoundment and enhanced C accumulation as discussed above. Alternatively, enhanced drainage conditions and drier systems could result from taliks which ultimately may penetrate thin permafrost or from further subsidence which increases topographic gradients.

[49] Last, decreases in dissolved organic carbon export have been measured from watersheds after wildfire due to removal of shallow soil organic C, and possibly due to greater water holding capacity with increased active layer thickness from fire-induced thaw [*France et al.*, 2000].

4.2.2. Rapid Permafrost Degradation From Thermokarst

[50] Degradation of permafrost with high ground ice content has profound impacts on topography, vegetation structure, and hydrology and thus can strongly determine dynamics of carbon sequestration and release [*Jorgenson and Osterkamp*, 2005; *Shur and Jorgenson*, 2007; *Turetsky et al.*, 2007; *Walter et al.*, 2007a; *Schuur et al.*, 2009]. Though high ground ice content is a protective feature if slow top-down thawing takes place, it can render ground highly vulnerable to disturbances. Excess ground ice, which can include segregated ice, ice wedges, and other massive ice bodies, is often irregularly distributed in the ground and its melting can strongly reshape the local landscape by forming thermokarst terrain. Thermokarst landforms are extremely variable in their form and nature and occur as a result of interactions among climate, topography, soil texture, hydrology, amounts and types of ground ice, heat transfer mechanisms, and local disturbances. Particularly organic Cryosols (in peat plateaus) and yedoma-like deposits can have a very high ground ice content and their significant C stocks below 1 m depth can be affected by thermokarst formation, lateral erosion, and deep talik formation, which are typical for these areas. In North America, high ground ice contents (for lowlands, >20% by soil volume) occur in Cryosol-covered areas totaling about 887,000 km^2 and containing 33 Pg C in the upper 1 m of soil (33% of the North American Cryosol SOC stock down to 1 m depth) (Table 4). In addition, medium ground ice contents (10–20% by soil volume) occur in Cryosol-covered areas totaling another 714,000 km^2 and containing 21 Pg C (21%) in the upper 1 m of soil. Thermokarst in these regions will have complex and varied impacts on soil, vegetation, and erosion, but changes in C storage are likely to be large as thermokarst is capable to tap into deeper C pools of peatlands and Yedoma-like sediments (see section 2). The extent and depth of future thermokarst processes remain very

poorly constrained, but will likely affect a significant proportion of the areas with high ground-ice content.

[51] Thaw processes are active on annual to decadal timescales, occur at mean annual air temperatures as high as +2°C and as low as −20°C, and vary greatly in completeness of degradation. Thermokarst can result from pooling of surface water into thermokarst ponds; an increase in seasonal snow depths resulting in thermal insulation of permafrost from cold winter temperatures; a removal of insulating organic layers and vegetation by wildfires or erosion allowing deeper summer thaw; or man-made impacts connected to infrastructure or construction and transportation activities. The resilience and vulnerability of permafrost to climate warming and disturbance is complex due to strong positive feedbacks from water and negative feedbacks from vegetation [*Jorgenson et al.*, 2010].

[52] Impoundment of water in subsiding thermokarst terrain floods collapsing soils and alters the surface energy balance (see also section 4.1 on hydrological disturbances). In some regions, thermokarst results directly in paludification and peat accumulation. In other regions, coalescing ponds can grow into large thermokarst lakes (1 to ~25 m deep, few 100 m^2 to several 100 km^2 in surface area), providing strong positive feedbacks to permafrost thawing around and beneath the lake (Figure 6d). Thermokarst lakes massively reshaped Late Pleistocene accumulation plains of nonglaciated regions in North America (Alaska North Slope, the Seward Peninsula, the Alaska Interior, and northwestern Canada) during the Holocene, thereby rearranging hydrological patterns and degrading vast stretches of ice-rich carbon-storing permafrost.

[53] Thermokarst environments provide unique sites for microbial communities and C decomposition. Thaw bulbs developing beneath thermokarst lakes within a few tens to hundreds of years [*West and Plug*, 2008] are anaerobic environments in which microorganisms produce methane year-round by decomposing labile, formerly permafrost-stored organic carbon [*Walter et al.*, 2007a]. Another rapid mechanism of permafrost-stored organic matter entering thermokarst lake systems is through lateral lake expansion and thermal erosion of lake banks consisting of exposed soils and sediments. Decomposition and release of methane from freshly thawed organic matter in thaw bulbs and lake bottoms starts more or less immediately. Thermokarst lakes are capable of mobilizing very deep and very old Late Pleistocene permafrost-stored carbon [*Walter et al.*, 2006] and were identified as an important source of atmospheric CH_4 in past warming periods [*Walter et al.*, 2007b]. Photosynthetic microorganisms can form microbial mats over thaw ponds and thereby lower rates of CO_2 emissions compared to ponds without mats [*Laurion et al.*, 2010]. *Schuur et al.* [2009] report that fluxes of old permafrost-stored C in thermokarst areas on tundra uplands were offset initially by enhanced plant growth in wetter thermokarst zones. However, once surface carbon from thaw-induced vegetation change equilibrated, the loss of old SOC shifted overall ecosystem carbon storage from a net C sink to a C source on decadal timescales.

[54] Lake ice conditions under a warming climate are an important factor for permafrost thawing under shallow thermokarst lakes and related C mobilization in thaw bulbs. As maximum lake ice thickness decreases in a warming climate, many shallow lakes will not freeze to the bottom anymore and could start developing a thaw bulb and emitting methane year-round. Future increase in permafrost thawing might cause rapid lake area growth due to increased water temperatures and new lake formation in some regions, or increased lake drainage due to active-layer deepening and thermal erosion and subsequent lake seepage in other regions. Current remote sensing observations show both growing and shrinking lake area trends in different North America regions. However, most lake level changes observed so far were attributed to shifting evaporation and precipitation patterns [*Riordan et al.*, 2006; *Plug et al.*, 2008] rather than ongoing permafrost degradation [*Yoshikawa and Hinzman*, 2003].

[55] Thermokarst lake drainage [*Hinkel et al.*, 2007] through tapping and breaching by river and coastal erosion or through slow degradation of surface permafrost and subsequent formation of drainage pathways can profoundly change the carbon dynamics of the impacted area. Terrestrial peat forming in drained lake basins sequesters C [*Hinkel et al.*, 2003] while new wetlands seasonally release methane. In addition, refreezing of the thaw bulb after lake drainage within decades [*Ling and Zhang*, 2004] is preventing organic C in these sediments from further decomposition. A large portion of arctic and subarctic lowlands is covered with drained-lake basins [*Hinkel et al.*, 2005] formed over thousands of years, indicating the significance of this process for C dynamics in permafrost regions.

[56] Thermal erosion gullies (Figure 6e), active layer detachment slides, and retrogressive thaw slumps (Figure 6f) can be rapidly formed within a summer season or a few years and often deeply erode ice-rich permafrost soils in areas with sufficient relief gradient. Volumes of soil, sediment, and organic C mobilized with such processes are comparably small and effects on the total northern high-latitude C pool are of local nature so far. However, increasing abundance of these features has been reported from various regions, including the Yukon, Northwest Territories, Canadian Archipelago, and the foot hills of the Brooks Range in Alaska, indicating that their impact becomes more important in a warming climate [*Liljedahl et al.*, 2007; *Lantz and Kokelj*, 2008; *Bowden et al.*, 2008; *Lacelle et al.*, 2010]. Erosion and thermokarst is often connected with a redeposition of sediments and partial reburial of C in downslope areas, such as toe slopes, lakes, streams or the sea.

5. Soil Organic Carbon Fate

[57] A central question is the extent to which disturbance regimes and their soil and plant successional trajectories will change in response to warming and hydrology and how this will impact northern high-latitude SOC pools. While landscapes and soils have sequestered carbon for past millennia, particularly in peatlands and permafrost soils (Figure 4), future disturbance regimes in a warming climate will impact the northern high-latitude SOC pool through a complex series of physical and biological responses. An important factor here will be the linkage between press and pulse disturbances (Figure 1), i.e., how climate warming may trigger pulse disturbances from already intensified press disturbances, for example as increased dryness or active layer depth in boreal forest regions leads to more fires and enhanced biomass burning. The fate of northern high-latitude SOC in a post-disturbance setting depends particularly on (1) whether the carbon remains in unaffected layers or is subjected to changes in moisture, temperature, or leachates, (2) whether the carbon

is frozen or subject to freeze-thaw cycles, (3) whether the ice or water content is high or low, because ice and water dictate both thermal, geomorphic, and biological processes, and (4) soil organic matter quality, which depends both on botanic origin and long-term decomposition trajectories. As successional stage is important to how plants and soils respond to disturbance [*Chapin et al.*, 2006] widescale vulnerabilities will likely involve changes in either successional timing or associations. For example, over recent decades many boreal forests generally release C in early stages of succession and accumulate C in later stages to sustain a small net sink over repeated fire cycles [*Goulden et al.*, 2010]. Modeling that is capable of assessing disturbance processes, resilience mechanisms, and timing and association of successions will be key to accurate prediction of the fate of northern high-latitude SOC.

6. Predicting Northern High-Latitude SOC Dynamics in Response to Disturbance

[58] There are many issues involved with incorporating the dynamics of organic soils into earth system models [*Frolking et al.*, 2009]. In the last decade there has been much progress in integrating soil freeze-thaw dynamics, hydrology, and biogeochemistry in large-scale ecosystem models that simulate the carbon dynamics of northern high-latitude organic rich soils [e.g., *Zhuang et al.*, 2002; *Balshi et al.*, 2007; *Beer et al.*, 2007; *Koven et al.*, 2009; *Wania et al.*, 2009a, 2009b]. Only one large-scale model has included a representation of cryoturbation as a disturbance process, representing it as a diffusional redistribution of organic carbon down the soil profile [*Koven et al.*, 2009]. They found that cryoturbation coupled with the strong effect of organic layers on soil thermal properties led to a significant increase in SOC sequestration over millennia, particularly in northeastern Eurasia, and to a decrease in active-layer thickness throughout the north. However, the model did not simulate the effect of organic rich soils on soil hydrology, or the impacts of fire. *Wania et al.* [2009a, 2009b] explicitly modeled peatlands and peatland hydrology along with soil freezing and thawing and permafrost dynamics, but did not consider disturbances like fire and cryoturbation. Although many of the recent modeling efforts considered fire in their large-scale applications [e.g., *Zhuang et al.*, 2002; *Balshi et al.*, 2007; *Beer et al.*, 2007], the models generally treat organic soils as static in response to disturbances like fire and do not consider how the loss and reaccumulation of soil organic matter following disturbance affect hydrology, water table depth, and unsaturated zone water content in organic rich soils. Thus no models are yet able to fully consider how changes in hydrology and soil thermal dynamics associated with disturbance influence soil carbon dynamics at high latitudes.

[59] There are numerous challenges in modeling the fate of SOC of northern high-latitude ecosystems in response to disturbances like fire. Since fire in northern high-latitude ecosystems combusts some of the surface organic layer, models must simulate burn depth and therefore need to represent the distribution of SOC with depth. The response of SOC after disturbance depends, in part, on the quality of remaining SOC, and therefore the quality of SOC with depth also needs to be represented. It is also important to represent how soil thermal and hydrological dynamics vary with depth as these dynamics depend, in part, on the structure of the SOC horizons above the mineral soil. Modeling the accumulation of SOC after disturbance requires representation of inputs with depth, and models need to represent the trajectory of vegetation succession after disturbance as different trajectories may affect both the quantity and pattern of inputs.

[60] *Harden et al.* [2000] took the first step in developing a spread-sheet model to evaluate how fire influences the long-term (6000 years) dynamics of SOC in a black spruce forest in Manitoba. This model primarily used a long-term mass balance approach to represent the dynamics of SOC horizons, including inputs to a particular horizon, the quantity of SOC in that horizon, and the turnover rate of SOC in that horizon. Thus, the model was able to represent some aspects of the distribution of SOC quantity and quality with depth as well as some aspects of inputs with depth. That model implementation showed the importance of fire disturbance in affecting long-term dynamics of organic C in fire-affected black spruce forests, and that this accumulation was sensitive to inputs, fire return interval, and the turnover of organic C in the soil. *Carrasco et al.* [2006] built on the approach of *Harden et al.* [2000] by having decomposition more explicitly driven by variability in soil temperature with depth. The study found that the C accumulation in soils depends on the presence of permafrost and the SOC need not be recalcitrant to accumulate. *Fan et al.* [2008] further extended the approach of *Carrasco et al.* [2006] to evaluate the role of soil moisture variability across sites by using inversion techniques. The results of *Fan et al.* [2008] indicate that decomposition (especially in wetter sites) was not accurately represented with standard moisture and temperature controls and that other important protection mechanisms (e.g., limitation of O_2, redox conditions, and permafrost) rather than inherent recalcitrance are responsible for the low decomposition rates of deep organic C in black spruce ecosystems of Manitoba, suggesting that SOC can be quite labile and potentially decomposable upon changes in these conditions due to disturbances.

[61] While the studies of *Carrasco et al.* [2006] and *Fan et al.* [2008] made much progress in understanding the role of permafrost in protecting SOC at depth, they did not consider how soil temperature and moisture vary during succession and how this may affect SOC dynamics. *Zhuang et al.* [2002] used a coupled soil thermal-biogeochemistry model to explore the potential effects of several factors associated with fire disturbance in black spruce forests of Alaska. Sensitivity analyses with this model indicate that, along with differences in fire and climate history, a number of other factors influence the response of carbon dynamics to fire disturbance, including nitrogen fixation, the growth of moss, changes in the depth of the organic layer, soil drainage, and fire severity. *Yi et al.* [2009a] further refined and tested the soil thermal model of *Zhuang et al.* [2002] with the goal of better representing soil thermal and hydrological dynamics with depth. Several differences between model simulations and field measurements identified key challenges for evaluating/improving model performance, including (1) proper representation of discrepancies between air temperature and ground surface temperature; (2) minimization of precipitation biases in the driving data sets; (3) improvement of the measurement accuracy of soil moisture in surface organic horizons; and (4) proper specification of organic horizon depth/properties and soil thermal conductivity.

[62] Permafrost model analyses [*Marchenko et al.*, 2008] show that vegetation succession provides strong negative feedbacks that make permafrost resilient to even large increases in air temperatures [*Jorgenson et al.*, 2010]. The proper specification of organic horizon properties during succession requires models to dynamically represent the accumulation of soil organic matter during succession, as soil organic matter acts as an insulator to heat exchange between the deeper soil horizons and the atmosphere. To address this issue, *Yi et al.* [2009b] developed relationships for how organic soil properties vary within soil vertical profiles as well as between two different age classes (young versus mature) and between moisture conditions (dry versus wet) for black spruce forests. To demonstrate the use of these relationships, *Yi et al.* [2009b] incorporated this understanding into a terrestrial ecosystem model to simulate the dynamics of organic horizons and associated variations in the soil environment for black spruce stands on dry and wet soils. *Yi et al.* [2010] conducted sensitivity analyses with the model to investigate issues related to spatial heterogeneity of carbon dynamics including soil drainage and fire frequency. Simulations for wet black spruce ecosystems showed a threefold increase in decomposition immediately after fire due to increased active layer depth, and subsequent decrease in decomposition as soils get colder during succession due to the reaccumulation of the surface organic horizon. In contrast, simulations for dry black spruce ecosystems showed slight decreases in decomposition immediately after fire (due to lower levels of soil carbon) and slight increases in decomposition as soil carbon accumulates. These simulations indicate that the soil environment is a more important control in wet black spruce ecosystems while the quantity/quality of organic matter as affected by fire frequency is a more important control in dry black spruce ecosystems.

[63] To date, several modeling efforts have looked at the regional effects of historical fire disturbance on the storage of soil carbon in the boreal forest of North America [*Kurz and Apps*, 1999; *Balshi et al.*, 2007; *Bond-Lamberty et al.*, 2007; *Hayes et al.*, 2011]. There have also been efforts to model the response of soil organic matter to future projections of fire disturbance [*Balshi et al.*, 2009a, 2009b]. While these efforts have treated the response of soil organic matter to disturbance with respect to the balance of inputs of carbon to the soil versus releases of carbon from the soil, they have not represented the SOC system in terms of how soil hydrological, thermal, and biogeochemical dynamics covary with depth during succession. Other challenges with respect to regional assessments of soil carbon response to fire include the modeling of soil organic horizon combustion across the landscape [*Barrett et al.*, 2010]. It will be important to develop models capable of predicting combustion across the landscape in a changing climate, particularly with respect to how combustion relates to a deepening of soil thaw and possible drying of currently wet organic soils in peatlands. To date, little progress has been made with respect to modeling the effects of fire or other disturbances on SOC dynamics in peatlands. Besides fire, other disturbances like insect disturbance [*Kurz et al.*, 2008] and thermokarst (see section 4.2) have important effects on carbon storage, but have not received much attention with respect to modeling the effects of disturbance either directly on soil carbon pools (except for *Koven et al.* [2009] on cryoturbation), or indirectly on how soil hydrological, thermal, and biogeochemical dynamics covary during succession following these types of disturbance.

7. Research and Data Gaps

[64] Landscapes are variable with respect to the vulnerability of SOC to press or pulse disturbances, and an assessment of overall net C balance requires insights and models to capture such heterogeneity. For all disturbances, subgrid-scale processes need to be meaningfully integrated with macro-scale models. However, large uncertainties are still connected to some of the data needed for models. For example, knowledge on spatial and depth distributions of SOC stocks in large and often very remote regions of the North American high latitudes is incomplete and needs to be enhanced. Similarly, the distribution and properties of permafrost (e.g., temperature and ground ice content) is mostly known from sparse point measurements only. Often, feedbacks of disturbance processes to SOC stocks tested and confirmed for one study site turn out to be more complex as more data is collected over wider environmental gradients. Transferring existing local information to large regions using appropriate models or techniques such as remote sensing is therefore one of the key research goals.

[65] In order to capture spatial heterogeneity of landscape-scale processes, more detailed spatial scales must be used and integrated mathematically into the coarser scales required for regional and global models. Spatial monitoring of aboveground organic C stocks and vegetation-soil-water interactions using remote sensing including hyperspectral, synthetic aperture radar (SAR) and Light Detection and Ranging (LIDAR) sensors, coupled with dedicated field campaigns, would greatly improve understanding SOC input dynamics related to terrestrial land surface dynamics under disturbance regimes. Since thermal and hydrological regimes are critical to vulnerability and resilience of SOC, remote sensing applications aiming at quantifying these regimes in northern high latitudes would enhance our capability to characterize, quantify, and potentially predict disturbances and their impacts on SOC at large scales. Upcoming soil moisture and surface water satellite missions (e.g., SMOS, SWOT, SMAP), spaceborne synthetic aperture radar missions to detect aboveground biomass (in C band, L band, P band), and existing thermal satellite missions (e.g., Terra MODIS, AVHRR) should be fully exploited to assess changes in land surface properties in northern high latitudes and their impacts on the large SOC pool. Gravity remote sensing of surface water dynamics over large watersheds for example could help detecting macro-scale changes in permafrost thaw, active layer thickness, soil water storage, and runoff. Remote sensing methods, in combination with field data on processes, could provide an excellent database for northern high-latitude disturbance models as well as terrestrial C-cycle models.

[66] Spatial heterogeneity is confounded with temporal heterogeneity of processes, some of which are carbon based and others of which exert controls on carbon processing. For example, the magnitude of CO_2 release from enhanced decomposition of SOC in thawing soils may be governed first by the rate of thaw and the length of the thaw season; second by substrate availability, e.g., labile carbon; and third by nutrient availability. The longevity of enhanced decomposition is further complicated by the timing of water availability for microbial processing and by nutrient-plant-microbial

interactions. Such complexities are vastly understudied and require complex models that account for spatiotemporal interactions at various spatial and temporal resolutions.

[67] Field data on SOC processes need to be enhanced as well. For example, the occurrence of fire and combustion as a function of physiography and soil type is potentially complex. The buildup of deep, moist organic profiles cools subsurface mineral soils, facilitating the persistence of permafrost and initiating a feedback among cold conditions, poor drainage, thick organic layers and reduced susceptibility to fire [*Turetsky et al.*, 2005; *Harden et al.*, 2006]. Climate warming can potentially uncouple physiography from susceptibility to fire if hydrological conditions change and poorly drained areas become more prone to burning [*Turetsky et al.*, 2004].

[68] Also, the role of cryoturbation in C source/sink processes are not well understood, yet cryoturbation dynamics on centennial to millennial timescales are likely affected by both warming and hydrological change. Whether warming could increase cryoturbation causing high-latitude soils to become stronger sinks of carbon is an active area of investigation. Cryoturbation appears to be more common in North America during past warm events in the Holocene [*Bockheim*, 2007], suggesting that this process might sequester more carbon in northern high-latitude soils in a warmer climate. *Ping et al.* [2008a] also found that cryoturbation increased along a north to south warming gradient in Alaska. *Bockheim* [2007] suggests that movement of organic matter to deeper and colder parts of the soil column by cryoturbation results in a decrease of decomposability and long-term SOC storage. Future studies should target dating and modeling of cryoturbation in context of regional warming, fire, and hydrological shifts.

[69] One of the challenges of understanding the susceptibility of deep SOC to thaw, cryoturbation, decomposition, and leaching is the robust evaluation of existing stocks in relation to active layer and deeper processes [e.g., *Ping et al.*, 2008b] rather than in relation to depth alone. While increased thaw and changes in vegetation can increase organic C accumulation in surface soils (see section 4), the effects of warming or thaw are propagated to deeper soils and increased respiration from these layers might shift these ecosystems to carbon sources [*Schuur et al.*, 2009]. This can be studied through the measurement of ^{14}C in soil respiration, which can provide a sensitive fingerprint for detecting changes in the respiration of C deep in the soil as permafrost thaws. Laboratory soil incubations allow for experimental studies to test a variety of hypotheses regarding organic C lability and microbial activity under a simulated disturbance regime. Incubations reduce system complexity and increase the interpretability of results to quantify potential rates of decomposition under varying conditions (temperature, moisture, and oxygen availability), to understand the importance of substrate quality (vegetation, parent material, composition trajectories), and to study microbial communities in context soil carbon processing. Taken together, these studies provide insight into the vulnerability of shallow and deep soil carbon pools under various disturbance regimes including warming temperatures and permafrost thaw, draining or flooding, and wildfire. What laboratory incubations do not resolve is the importance of in situ rates of gaseous and liquid diffusion, enzyme and substrate mobility, both in context of thermal gradients. For such in situ studies, depth profiles of biogeochemistry should be combined with core/borehole and C flux measurements.

[70] Ground temperature data are a crucial element of understanding the current extent of permafrost, its future trajectory, and the vulnerability of SOC pools stored in permafrost and permafrost-affected soils on local, regional, and circum-arctic scales. Similarly, ground ice content and ice distribution are key parameters for determining the vulnerability of permafrost, but still hard to estimate for large regions. Therefore, data collection on physical permafrost properties (temperature, ice content, thickness) in various environments and climatic settings need to be enhanced, including continuation and enhancing of existing permafrost and active layer temperature monitoring programs and better coordination of these programs with ecological and C cycle research projects. Similarly, the mapping of thermokarst remains challenging because of the heterogeneity of processes involved. Geophysical and remote sensing technologies, in combination with field studies and numerical modeling, need to be further developed to allow mapping of permafrost physical parameters and detection of changes for large regions [e.g., *Bartsch et al.*, 2010; *Romanovsky et al.*, 2010]. We need to enhance our understanding of permafrost-vegetation-hydrology feedbacks under current and future climate scenarios. Vegetation and hydrology are known to have extensive interactions with permafrost, particularly in the zone of discontinuous permafrost. Thresholds for permafrost stability are also depending on these interactions.

[71] Last, we need to determine the role of terrestrial permafrost and also submarine permafrost of terrestrial origin in the stability of old C pools which are not currently part of the active C cycle. Future reduced permafrost extent and thickness, and increased permeability of permafrost will potentially result in the release of old C from long-term stores in and beneath permafrost into the active C cycle. This also requires a better quantification of C pools of different origin in shallow and deep terrestrial permafrost soils and subsea shelf permafrost, as well as C pools beneath permafrost. Variability, vulnerability to various disturbance types, and bioavailability of these C pools need to be understood.

[72] Appropriate numerical models for permafrost, C cycle, hydrology, and vegetation, and their integration in advanced Earth System models are required. Permafrost models are highly dependent on quality input derived from climate models. In addition, all current permafrost models are incapable so far of incorporating local subgrid disturbances such as thermokarst, and thus only include widespread but gradual top-down thawing. Improving regional and global numerical permafrost models by integrating thermokarst and other disturbances is therefore necessary to better quantify and predict the impact of disturbances on northern high-latitude SOC pools.

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