SOIL Discuss., 1, 709–756, 2014
www.soil-discuss.net/1/709/2014/
doi:10.5194/soild-1-709-2014

SOIL Discussions Open Access

Permafrost soils and carbon cycling

C. L. Ping[1], J. D. Jastrow[2], M. T. Jorgenson[3], G. J. Michaelson[1], and Y. L. Shur[4]

[1]Agricultural and Forestry Station, Palmer Research Center, University of Alaska Fairbanks, 1509 Geogeson Drive, Palmer, AK 99645, USA
[2]Biosciences Division, Argonne National Laboratory, Argonne, IL 60439, USA
[3]Alaska Ecoscience, Fairbanks, AK 99775, USA
[4]Department of Civil and Environmental Engineering, University of Alaska Fairbanks, Fairbanks, AK 99775, USA

Received: 4 October 2014 – Accepted: 15 October 2014 – Published: 30 October 2014

Correspondence to: C. L. Ping (cping@alaska.edu)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

Knowledge of soils in the permafrost region has advanced immensely in recent decades, despite the remoteness and inaccessibility of most of the region and the sampling limitations posed by the severe environment. These efforts significantly increased estimates of the amount of organic carbon (OC) stored in permafrost-region soils and improved understanding of how pedogenic processes unique to permafrost environments built enormous OC stocks during the Quaternary. This knowledge has also called attention to the importance of permafrost-affected soils to the global C cycle and the potential vulnerability of the region's soil OC stocks to changing climatic conditions. In this review, we briefly introduce the permafrost characteristics, ice structures, and cryopedogenic processes that shape the development of permafrost-affected soils and discuss their effects on soil structures and on organic matter distributions within the soil profile. We then examine the quantity of OC stored in permafrost-region soils, as well as the characteristics, intrinsic decomposability, and potential vulnerability of this OC to permafrost thaw under a warming climate.

1 Introduction

Permafrost is defined as ground (soil or rock and included ice) that remains at or below 0 °C for at least two consecutive years (Washburn, 1973). In this review the permafrost material that receives the most attention is ice-cemented permafrost because of the phenomena associated with ice formation, freeze-thaw cycles, and frost heave, which affect soil forming processes and resulting soil properties. Permafrost underlies nearly 24 % of the exposed land surface on Earth, mostly in the lowlands of the circumpolar north regions, the boreal regions, high alpine and plateau regions in the Northern Hemisphere, and limited areas in the high alpine regions of the Southern Hemisphere and Antarctica (Bockheim, 1995; Brown et al., 1998). The total ice-free land areas of the northern circumpolar region encompass 17.8×10^6 km^2 and are estimated to store

SOILD
1, 709–756, 2014

1034 Pg organic carbon (OC) in the surface 0–3 m. The permafrost-affected soils account for only 58 % of this land area yet they store 70 % of the soil OC (Hugelius et al., 2014). Recently, permafrost and permafrost-affected soils have received greater attention because of their large OC stocks and the potential, with warming, for releasing significant amounts of this C as the greenhouse gases carbon dioxide (CO_2) and methane (CH_4) (McGuire et al., 2009; Schuur et al., 2011). The release of C is exacerbated by rising soil temperatures and accelerated rates of permafrost thawing, creating a strong positive feedback loop, especially for CH_4 with its greater warming potential compared to CO_2 (Oechel et al., 1993; Koven et al., 2011). But, recent reports (Walter et al., 2006, 2007; Zimov et al., 2006) suggest that while a large amount of CH_4 is produced in thermokarst features created by permafrost thawing, this could be mitigated over the long term by increased primary production (Walter Anthony et al., 2014). However, what controls the rate of gas release is not necessarily total OC stocks but rather, environmental factors such as changes in soil hydrology, temperature, and vegetation communities (Olefeldt et al., 2013). In addition, differences in landscape position and the ice content of permafrost create widely varying conditions that affect soil drying or wetting after permafrost thaw, and ultimately affect the accumulation or loss of soil OC (Jorgenson et al., 2013).

The terms “permafrost soils” and “permafrost-affected soils” are commonly used synonymously. Permafrost-affected soils are referred to as Cryosols in both the World Reference Base (WRB) (IUSS Working Group WRB, 2006) and the Canadian soil classification system (Soil Classification Working Group, 1998), Gelisols in the US system (Soil Survey Staff, 1999), Cryozem in the Russian system (Shishov et al., 2004), and permagelic suborders in the Chinese system (Gong et al., 1999). Despite differences in nomenclature and the properties used for classification, these systems share the basic requirement of the presence of permafrost at a certain depth, generally within 2 m of the surface. However, these systems differ substantially in their classification hierarchy. In the Russian and WRB systems the genetic lineage or soil material type, such as organic vs. mineral, supersedes soil thermal regimes. For example, organic

SOILD
1, 709–756, 2014

soils (Histosols) key out first and those with permafrost key out in the subclass as Cryic Histosols. But, in the Canadian and US systems, Cryosols and Gelisols, respectively, key out first because permafrost is recognized as the controlling factor in land use interpretation and ecosystem functions.

Gelisols of the US classification system have three suborders: Histels, Turbels, and Orthels. Histels are organic soils, Turbels are mineral soils affected by cryoturbation, and Orthels are mineral soils lacking cryoturbation (Soil Survey Staff, 1999). Gelisols are ubiquitous throughout the continuous permafrost zone of the northern circumpolar Arctic region. In this zone, Gelisols also form in lowlands with restricted drainage (Rieger et al., 1979; Ping et al., 2004, 2005a). Within the discontinuous and sporadic permafrost zones of the boreal regions, slope and aspect of uplands and mountains play a controlling role in distributions of permafrost and Gelisols (Péwé, 1975; van Cleve et al., 1983; Ping et al., 2005b). In the Southern Hemisphere, permafrost and Gelisols occur on the Antarctic continent, in the sub-Antarctic islands, and in mountain areas (Beyer et al., 1999, 2000; Blume et al., 1997; Bockheim, 1995). However, the OC content of permafrost-affected soils in the southern circumpolar region is extremely low when compared to the northern permafrost region (Bockheim, 1990). Hence, the focus of this review is on permafrost soils of the northern permafrost region.

2 Permafrost characteristics and transformations

2.1 Permafrost formation and distribution

There are two general types of permafrost. The most common one is a paleoclimate remnant of the Quaternary age. This old permafrost is widespread throughout the lowlands and hills of the circumpolar region (Gubin, 1993; Péwé, 1975; Kanevskiy et al., 2011; Winterfeld et al., 2011). Most notable for this kind are the Yedoma formations in northeast Russia, Arctic and boreal Alaska (Reyes et al., 2010; Schirrmeister et al., 2011; Kanevskiy et al., 2011, 2014), and the Yukon regions in Canada (Froese et al., 2008). Yedoma is also called "Ice-Complex" because of the huge ice wedges formed in thick deposits of syngenetic origin (i.e., concurrent upward growth of deposits and the permafrost surface). Old permafrost that consists of stratified sediments is also found in deep loess deposits, such as those in central Alaska and the Yukon Territory of Canada (Péwé, 1975). The second, more modern kind of permafrost was formed during the Holocene as a result of deglaciation followed by vegetation succession under cold climatic conditions and is mostly located in the sub-Arctic and boreal regions (Zoltai and Tarnocai, 1975). This more recently formed type has been termed "ecosystem-driven permafrost" (Shur and Jorgenson, 2007). In general, most permafrost in the Arctic is of late-Pleistocene age (Brown, 1965; Gubin, 1993; Kanevskiy et al., 2011; Ping et al., 1997, 2008a). In the Dry Valley region of Antarctica, the permafrost is even older as it formed during the pre-late Quaternary (Bockheim, 1990). However the upper permafrost of the old formations is subject to contemporary climate fluctuations.

2.2 Cryogenesis and cryostructure

Permafrost affects soil formation through physical and biological processes. Permafrost exerts a controlling role on soil physical and morphological properties through cryogenesis; i.e., frost cracking and formation of ice crystals and lens cryostructures, resulting in frost heaving and cryoturbation (Ping et al., 2008b). Cryostructures are defined as patterns formed by ice inclusions in the frozen soil (Kanevskiy et al., 2011). These inclusions form as ice crystals and lenses and as layers of segregated ice in the mineral soil matrix due to freezing conditions within the soil (French and Shur, 2010; Shur and Jorgenson, 1998; Shur et al., 2005). There are seven major types of ice cryostructures in the active layer and the upper permafrost: structureless (pore), lenticular (microlenticular: < 0.5 mm thick and lenticular: > 0.5 mm thick), layered, reticulate, irregular reticulate (braided), crustal, and suspended (ataxitic) (Mackay, 1974, French and Shur, 2010). Some examples of these cryostructures are shown in Fig. 1. Upon desaturation after thawing, these cryostructures generally leave soil structures that include platy, blocky, wedge-shaped, granular and massive (Ping, 2013).

SOILD
1, 709–756, 2014

According to French and Shur (2010) a permafrost soil profile commonly has three layers referred to as the active, the transient and the intermediate layers. The active layer is defined as the zone of the soil profile above the permafrost that is subject to annual freeze-thaw cycles (Burn, 1998). The active layer is the modern soil, mostly formed since the Holocene and a product of the contemporary climate that undergoes seasonal thawing. The active layer typically consists of a surface organic horizon (O) – with or without a humus-rich surface mineral horizon (A) – and a cambic horizon (Bw, Bg), or the entire active layer can consist of just mineral or organic matter (Ping et al., 1998, 2013). The transient layer (uppermost part of the permafrost) results from fluctuations of the permafrost table that have occurred on a decadal scale and has distinct layered, lenticular and reticulate cryostructures (Shur et al., 2005). The intermediate layer is caused by aggradation of permafrost due to soil climate changes resulting from the buildup of surface organic horizons on a decadal to century scale. The intermediate layer is characterized by ice-rich cryostructures that are referred to as "suspended" (or "ataxitic") because blocks of soil are suspended in an ice matrix (Fig. 1). Together the transient and intermediate layers comprise the uppermost section of permafrost (Shur, 1988) and below it is the "true" permafrost that has remained frozen and not been subjected to freeze-thaw cycles on century to millennial scale (French and Shur, 2010).

2.3 Periglacial processes and patterned ground

Toward the end of the growing season solar radiation decreases and the active layer starts to freeze. In the continuous permafrost zone with cold permafrost, the active layer begins to refreeze both from the top down and bottom up, while in the discontinuous zone with warmer permafrost, the active layer freezes only from the top down. The part of the active layer that has not yet frozen is isothermal and called the "zero-curtain envelope" (Rieger, 1983; Davis, 2001). It generally takes 3 to 4 months for the zero-curtain to close, i.e., freeze-up (Davis, 2001). During the freeze-up process, lenticular, reticulate and/or layered cryostructures form in both the upper and lower parts of the

SOILD
1, 709–756, 2014

active layer, drawing soil water away from the middle part of the active layer and leaving it relatively dry and desiccated. This leaves a dessicated layer with either coarse platy or massive soil structure (Shur and Ping, 1994). The exception is in poorly drained soils where there is adequate water to sustain layered cryostructure development throughout the entire active layer.

Frost heave, caused by ice segregation and ice lens formation, often results in the deformation of the ground surface. With horizontal expansion limited, there is often enough stress to produce crooked or tilted lenticular and reticulate structures (French and Shur, 2010). Differential frost heave eventually deforms originally flat horizons into warped or wavy horizons. When soil of the active layer freezes on two freezing fronts in conditions where water is not limiting, a saturated zone becomes sandwiched between the fronts and builds up cryostatic pressure that eventually causes the soil to become gelified in this zone, and this soil can deform through weak spots or cracks (Ping et al., 2003). When ice-rich layers below the active layer thaw, the release and movement of water leads to a type of micro-scale diapirism where water and saturated materials are forced through brittle layers or cracks while heavier mineral and organic rich soils fill the ensuing voids (Swanson et al., 1999).

Patterned ground derived from freeze/thaw processes greatly affects soil formation and the complexity of soils at the micro-scale. Common patterned ground types include ice-wedge polygons, sorted and nonsorted circles, stone nets, stripes, and gelifluction lobes (Washburn, 1973). Ice-wedge polygons form when contraction and expansion of the ice/soil permafrost matrix creates a large-scale net of cracks (5–50 m diameter polygon units) in winter. The cracks fill with water during the warm season and refreeze the following winter. This process repeats itself annually in new cracks in roughly the same areas to build wedges that control surface water distribution and deform developing soils (Kanevskiy et al., 2013; Ping et al., 1998).

Circle pattern formation generally is initiated by frost cracking of the active layer ground followed by repeated freeze-thaw cycles, and thermo-hydrologic conditions around the cracks are modified by vegetation and ice aggradation (Shur et al., 2008;

SOILD
1, 709–756, 2014

Title Page
Abstract Introduction
Conclusions References
Tables Figures
Back Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

Walker et al., 2008). With these modified thermo-hydrologic conditions, the surface across circles heaves at differential rates with greater heaving in the centers relative to the margins. For soils without rock fragments, materials within and outside the circles have the same general texture, and are classified as "nonsorted" circles or more commonly "frost boils". For soils with coarse or rock fragments, these fragments tend to be heaved up and eventually pushed to the outer edge of the circle by repeated cycles of frost action and thus form segregated domains that are referred to as "sorted" circles (Washburn, 1973). In the initial stage, circle formation leaves the soil surface devoid of vegetation. Progressively vegetation communities establish and organic horizons start to build. However with the continuing annual differential heaving of the circle surface, often the surface organic mat becomes ruptured, forming discontinuous surface organic horizons. On slopes, gravitational forces can cause both nonsorted and sorted circles to deform downslope as they go through their annual freeze-thaw cycles, and these develop into stripes or stone stripes when rock fragments are present (Geisler and Ping, 2013).

Oriented rocks are common in sorted circles. Orientation of rocks is caused by thermally-induced frost heave, similar to a common phenomenon observed in soils wherever seasonal frost occurs. The mechanism driving rock orientation is that the longest rock axis tends to orient itself parallel to the frost table, while the short axis lies perpendicular in order to minimize resistance (Washburn, 1973; Davis, 2001). Silt-capped rocks commonly occur in fragmental soils on exposed landscapes in the Arctic and alpine regions (Munn and Spackman, 1990). The freeze-thaw cycles contribute to the grinding of rock fragments into silt-sized particles which are then transported by percolating water and accumulate on the upward facing surfaces of stones.

Patterned ground formation, can cause large- and small-scale variation in soil types across the Arctic landscape. For example, across ice-wedge polygons, soils in the polygon troughs often have moderately thick peat underlain by massive ice, soils on the rims are better-drained with well-decomposed organics forming A horizons (Fig. 2), while soils in the polygonal centers have thick peats over mineral sediment with different

SOILD
1, 709–756, 2014

degrees of cryoturbation (Ping et al., 1998, 2011). For circle patterns, the small-scale net of cracks (0.5–5 m) in the active layer affects moisture availability, plant community distributions, and differential active layer dynamics (Walker et al., 2008) and creates different soil profile characteristics across the circles (Michaelson et al., 2008).

2.4 Thermokarst

Where the water of ground ice in fine-grained sediments exceeds the pore space of the soil, thawing of the permafrost can cause the surface to settle or liquefy, and the amount of settlement is directly related to the amount and type of ice (Shur and Osterkamp, 2007). The irregular topography resulting from the melting of excess ground ice and subsequent ground collapse is called thermokarst. The patterns and amount of settlement or loss of surficial material are related to complex interactions of slope position, soil texture, hydrology, and vegetation over time (Shur and Jorgenson, 2007). The highly variable terrain and permafrost factors have led to a wide variety of thermokarst landforms that include degrading ice wedge troughs, thermokarst pits, thermokarst lakes, thermokarst bogs, thaw slumps, and thermal erosion gullies (Jorgenson et al., 2013). Thermokarst is widespread throughout Arctic and boreal regions and has large implications for soil hydrology and C balance (Sannel and Kuhry, 2011; Schuur et al., 2008; Jorgenson et al., 2013; Walter Anthony et al., 2014).

Permafrost dynamics have implications for soil formation and soil properties, including soil OC storage. Patterning of the ground surface at larger scales can lead to fine-scale variations in moisture distributions and temperature profiles, and thus the development and distribution of soils with differing properties.

3 Cryopedogenesis

Cryopedogenesis refers to soil formation processes as affected by freezing temperatures and freeze-thaw processes or cryogenic processes. Some of the most important

SOILD
1, 709–756, 2014
Permafrost soils and carbon cycling
C. L. Ping et al.

direct effects result from the physical barrier the permafrost zone imposes on (1) water movement and biogeochemical processes at low temperatures and (2) the expansion and contraction of the active layer during the formation of seasonal ice. Permafrost also exerts seasonal cooling effects on the bottom of the active layer, which affects ice formation and the freeze up of the active layer.

3.1 Biochemical process at low temperatures

The lower temperatures and higher moisture contents in the active layer can favor development of anaerobic conditions, that slow decomposition and enhance accumulation of soil organic matter (SOM). Cooler soil temperatures also slow chemical alteration of soil minerals and biological activity. The zero-curtain zone of the active layer provides favorable conditions for heterotrophic soil respiration during the shoulder winter season. Thus, appreciable gas fluxes have been measured on the snow surface during winter (Zimov et al., 1993, 1996; Fahnestock et al., 1999). It is likely that frost cracks provide passages for gases to reach the atmosphere as there is often a strong pulse of gases during early spring thaw. Under saturated conditions SOM is oxidized, whereas mineral elements are reduced even at sub-zero temperatures. Most commonly manganese (Mn) and iron (Fe) are first reduced followed by sulfur and then C (as CH_4) (Patrick and Jugsujinda, 1992). Elements like Mn and Fe in minerals are reduced by microbial processes and can release into the aquatic system (Lipson et al., 2010). This reduced Fe becomes oxidized upon in contact with surface water and forms ferrihydrite (Stiles et al., 2011). The hydromorphism also results in the gleyed color of the mineral matrix in the lower active layer due to Fe reduction (Ping et al., 1993, 2008b).

3.2 Soil structure in the active layer

Cryogenesis commonly results in some unique soil structures that are limited to the Arctic, sub-Arctic, Antarctic and alpine regions. Ice formation, as the active layer

SOILD
1, 709–756, 2014

refreezes in the fall and winter, varies in response to the different conditions that occur in different parts of the active layer, and this variation results in development of different soil cryogenic structures. On the exposed ground surface of silty mineral soils with no or only thin surface organic horizons, crumb and granular structures develop due to needle ice formation (Ping et al., 2008b). Ice crystals form on frozen soil particles or structural units near the surface when water is drawn from below to build or form the needle-like ice crystals that lift and move soil particles or structural units (Brink et al., 1967). Under snow cover, sublimation ice can build near the surface from water vapor in the soil surface atmosphere. In this process small amounts of SOM at the surface can be mixed with the surface few centimeters of mineral soil and can promote formation of soil biotic crusts (Michaelson et al., 2008, 2012). In the High Arctic, where vegetative and snow cover are scant, a persistent net of soil surface cracks form due to freeze-desiccation contraction creating small frost polygons that in turn produce microenvironments that support different vegetation communities (Walker et al., 2008). In areas where snow banks occur on slopes, the micro-cracking and vegetation pattern can interact with erosion and deposition patterns to form small < 1 m diameter turf hummocks (Tarnocai et al., 2006; Broll and Tarnocai, 2002). But these small polygons and hummocks caused by frost-desiccation cracks result in little cryoturbation, largely due to rapid freeze up and insufficient moisture to develop segregated ice during the freeze up. These surface cryogenic processes are most important on the surface of exposed mineral soils, such as those of active frost boils in the Low Arctic and more generally across the landscapes of the High Arctic.

3.3 Soil structure in the upper permafrost

As mentioned above the transition layer is the upper zone of permafrost that has formed more recently and can be thawed during extremes of the climate cycle but remains frozen most years. For this reason this layer may be more ice rich and be similar in ice and soil structures to the lower active layer when it is frozen. Angular blocky, lenticular, or platy soil structures are common. The upper permafrost below the intermediate

SOILD
1, 709–756, 2014

layer however, is usually devoid of visible or larger ice bodies or lenses. Rather, it is ice-cemented or has porous visible or invisible ice crystals (Murton and French, 1994). The upper permafrost is most often massive or structureless. The most common exceptions occur in permafrost landscapes that have developed for long periods of time or under many thaw-lake cycles or periodic depositional cycles, where the permafrost has moved steadily upward encompassing many previous transient and active layers (Hinkel et al., 2003). Gelisols in each of the suborders can contain very large proportions of ice in the upper-permafrost horizons (transient and intermediate layers of Shur, 1988). Investigations over large exposed areas of Alaska's Arctic coastline found the overall average ground ice content of Gelisols to be 77 % by volume in the upper few meters of soil, ranging from 89 % in Yedoma formations to 43 % in eolian deposits (Kanevskiy et al., 2013).

4 Modes of soil organic carbon accumulation particular to permafrost soils

Similar to other soils, the deposition of surface litter, rhizodeposits, and the turnover of belowground biomass are sources of OC inputs to SOM for permafrost soils. These inputs from the existing plant community are deposited in the active layer. However, in permafrost soils other processes, such as cryoturbation, deformation by massive ice growth, and intermittent burial and syngenetic permafrost growth, result in the storage of large quantities of this surface and near-surface produced SOM at depth.

4.1 Cryoturbation

Ice in the soil profile, especially segregated ice in the upper permafrost and seasonal ice formation in the active layer, are cryogenic processes that result in redistribution or mixing of soil horizons, generically termed cryoturbation. Cryoturbation is a suite of processes involving freeze-contraction cracking, ice-lens formation resulting in frost heave in the active layer, deformation by massive ice, and diaparism (Washburn, 1973;

SOILD
1, 709–756, 2014
Permafrost soils and carbon cycling
C. L. Ping et al.

Rieger, 1983; Swanson et al., 1999). Freeze contraction allows deep cracks or voids to develop in the soil during the winter, then during spring thaw water and saturated mineral and organic materials can flow into the voids. Frost heave in extreme cases can cause the surface to heave as much as 20 cm in circles compared to 3 cm in vegetated soils surrounding the circles (Romanovsky et al., 2008).

The most striking soil cryoturbation features are associated with patterned ground (i.e., nonsorted and sorted circles). These features include broken, involuted or warped horizons within the soil profile (Tarnocai and Smith, 1992; Ping et al., 2008b, 2013) that are caused by differential thermal convection processes in the soils. Commonly, circle centers are devoid of vegetation or are only sparsely vegetated, and the areas outside and at the margins of circles have tundra vegetation with thick organic horizons (Fig. 3). Due to the lack of surface thermal insulation, the mineral-dominated circles tend to thaw faster and deeper and also freeze faster, resulting in deeper active layers. In contrast, areas outside the circle, which are insulated by vegetation cover and organic layers, have shallow active layers and tend to have higher soil moisture content during thaw cycles (Shilts, 1978). Because the mineral-dominated circles freeze faster, they draw available water from circle margin areas to form ice lenses, which displace greater volumes in circle centers. As a consequence, greater heaving and deformation occurs within the circles than in the adjacent insulated inter-circle soil. This differential frost heave often causes the organic mat and, in some cases, the vegetative cover to be frost-churned down from the depressed inter-circle zone to the lower active layer under the circle and eventually can become part of the upper permafrost (Ping et al., 2008b). Generally, nonsorted circles form in fine-textured (loamy) soils, particularly silt loam soils. With increasing clay contents, the centers of such circles tend to be more elevated. Circles with elevated centers are termed "earth hummocks" with hummock height directly related to clay content (Ping et al., 2008b).

The importance of cryoturbation has long been studied and stressed by soil scientists and geocryologists. Classic examples of these studies were carried out in the 1970s in the Canadian Arctic by Tedrow (1974), Pettapiece (1974), Tarnocai and Zoltai (1978),

SOILD
1, 709–756, 2014

and summarized by Tarnocai and Smith (1992). These researchers found that the depth of the permafrost table is inversely related to the surface microtopography, i.e., the active layer is deepest in the center of the earth hummock and shallowest in the depressions around the hummocks. They also noted that buried and discontinuous organic horizons occurred at depth and attributed this "burial" phenomenon to cryoturbation. Concurrently in northeastern Russia, Makeev and Kerzhntser (1974) also found similar accumulation of cryoturbated organic matter at the permafrost table, which often occurs at depths of 100 to 200 cm. Rieger et al. (1979) also recognized the occurrence of "black streaks of frost-churned organic matter" at depth and discontinuous surface organic horizons in association with nonsorted circles in the tundra of Arctic Alaska. This work led to creation of the subgroup of "Ruptic-Histic" in the US taxonomic system (Soil Survey Staff, 1975). Later, more-detailed pedological studies in Arctic Alaska asserted that cryoturbation is the primary factor controlling the amount of C sequestered in the soils of tundra dominated by nonsorted circles (Bockheim et al., 1998; Michaelson et al., 1996; Ping et al., 1998, 2008a). Thus, the presence/absence of cryoturbated features was adopted as differentia for the Turbel suborder of Gelisols in US taxonomy (Bockheim and Tarnocai, 1998; Ahrens et al., 2004) and for the prefix of Cryosols in the WRB (IUSS Working Group WRB, 2006). However, these phenomena are not limited to the continuous permafrost zone. Cryoturbation and C translocation/sequestration associated with sorted circles, stripes, and solifluction are also commonly found in landscapes affected by periglacial processes, such as alpine environments in the boreal regions of Alaska (Geisler and Ping, 2013), Mongolia (Maximovich, 2004), and southern Siberia (Gracheva, 2004). Deeply subducted organic masses are even found in rocky permafrost soils lacking surface evidence of patterned ground (Jorgenson et al., 2013). Even when permafrost is absent under current climate conditions, cryoturbated profiles are often present as a remnant of past permafrost environments (Munn, 1987; Rieger, 1983).

Redistribution of SOM within the soil profile due to cryoturbation is most common in the Arctic, where patterned ground processes are active – such as sorted and

SOILD
1, 709–756, 2014

nonsorted circles and ice-wedge polygons (Washburn, 1973; Michaelson et al., 2008; Walker et al., 2008; Kanevisky et al., 2011; Ping et al., 2014). In early studies (Post et al., 1982; Brown, 1969; Everett and Brown, 1982), soil C stocks were measured only at shallower depths – mostly limited to the rooting zones at depths < 50 cm due to logistics. However, as concerns about global warming increased, this raised fears that the OC stored in permafrost-region soils might become a source rather than a sink for atmospheric C (Oechel et al., 1993).

Consequently, a series of studies were conducted in North America to explore the depth distribution of stored biogenic C in Gelisols (Michaelson et al., 1996, 2008; Ping et al., 1998, 2008a; Bockheim and Hinkel, 2007; Bockheim et al., 1999). Generally, on gentle to moderate slopes of glaciated uplands, SOM was cryoturbated to depths of mostly 80 to 120 cm. But, cryoturbated SOM was found to reach depths of 3 m or more on exposed ridge tops where vegetation is sparse and protection from snow cover in the winter is lacking (Michaelson et al., 1996), on floodplains (Shur and Jorgenson, 1998), and in drained-lake basins (Jorgenson et al., 2013). Most of these mineral soils with cryoturbated SOM and broken surface organic horizons are classified as Turbels (Turbic Cryosols). But, in the poorly drained valleys or basins among the hilly uplands, thick organic horizons (> 40 cm) can build up due fen or bog formation, and these soils are classified as Histels (Cryic Histosols).

Ping et al. (2008a) measured the C stores along a north-south transect through five bioclimatic subzones from the High Arctic to the boreal regions in North America as part of a larger study investigating the interrelationships between patterned ground formation and vegetation zonation (Walker et al., 2008). As the patterned ground transitioned from simple frost cracking in the most northerly area (subzone A) to well-formed nonsorted circles at lower latitudes, the landcover types changed from polar-desert to tundra and to shrub tundra in the south (subzone E). Ping et al. (2008a) found that soil C stores were directly related to biomass production but the amount of cryoturbated C did not follow the same trend. Rather, the proportion of cryoturbated C reached a peak near the middle of the transect (in subzones C to D) and then dropped off (Michaelson

SOILD
1, 709–756, 2014

et al., 2008; Ping et al., 2008a) due to increased insulation of the surface organic horizon in subzones D to E (Kade and Walker, 2008).

4.2 Deformation by massive ice formation

Ice-wedge polygons, with diameters ranging from a few meters to > 20 m across, dominate the landscapes of the Arctic coastal plains, which are widespread across the lowlands of the northern circumpolar regions, especially in Alaska, Canada and Russia (French, 2007). During ice-wedge development, soils on both sides of the ice wedge are pushed apart and heaved to form rims on both sides, creating a trough between the rims over the ice wedge that delineates the polygon (Fig. 2). The polygon center is less affected and remains flat or slightly lower than the rims. At this developmental stage, the polygons are flat- or low-centered, where the trough (and often the center) is wet during the growing season. Thus, three different soil types develop across this micro-toposequence: (1) organic or organic-rich soils over the ice wedge (generally < 50 cm across) in the polygon trough (Glacistels), (2) cryoturbated soils along the polygon rims (Histoturbels or Aquiturbels), and (3), soils usually lacking cryoturbation in polygon centers (Aquorthels, Historthels). However, with time, the ice wedges can degrade, forming deep troughs (Jorgenson et al., 2006). Polygon interiors then become high centered, surface cracking increases, and greater cryoturbation leads to formation of soils (Histoturbels) with greater accumulations of organic matter (Ping et al., 2008a, 2011; Zubrzycki et al., 2013).

4.3 Intermittent burial and syngenetic permafrost

Intermittent burial by eolian, alluvial, colluvial, and lacustrine sediments can lead to significant soil OC stocks in these depositional environments (Shur and Jorgenson, 1998; Schuur et al., 2008; Tarnocai et al., 2009; Grosse et al., 2011; Schirrmeister et al., 2011; Zubrzycki et al., 2013). These processes are often accompanied by syngenetic

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

SOILD

1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

permafrost growth, where the upward growth of the permafrost surface follows the accumulation of sediments and peat layers at the surface (Shur, 1988).

Deltas and thaw lakes with ice-rich permafrost are abundant along the Arctic coastal plains, such as at Barrow in northwest Alaska (Hinkel et al., 2005), the Colville Delta in northern Alaska (Shur and Jorgenson, 1998; Ping et al., 2011) and the Lena River Delta in northern Russia (Boike et al., 2013; Zubrzycki et al., 2013). On the Colville Delta, buried organic soils are found as deep as 3 m (Shur and Jorgenson, 1998).

Flat landscapes in the Arctic and boreal regions also are dotted with thaw lakes and drained thaw-lake basins. Thaw lakes form as the underlying permafrost thaws and ice wedges degrade (Washburn, 1973), with a cycle estimated at about 3000 years (Brown and Kreig, 1983; Hinkel et al., 2003). In some cases, soils formed in the polygons of these basins have experienced multiple thaw sequences, as evidenced by the presence of multi-layered organic-mineral horizons with cryoturbated organic matter to depths of 2–3 m (Ping et al., 2014) (Fig. 4). As a consequence of the processes occurring during the thaw-lake cycle, large soil OC stocks can accumulate, as high as 90 kg C m^{-2} from 0–3 m deep (Ping et al., 2011). Much of the organic matter in the central portions of drained-lake basins derive from limnic sediments comprised of algal material and detrital peat eroded from the collapsing lake shores (Jorgenson et al., 2013).

Extremely ice-rich silt deposits of Pleistocene age, known as Yedoma or Ice Complex, are particularly noteworthy because of the significant stocks of “fossil” OC. These deposits (average depth ~20–25 m) developed during the late Pleistocene in unglaciated areas of the Beringia, including Northeast Russia, Arctic and Interior Alaska, and the Yukon Territory, Canada (Zimov et al., 2006; Froese et al., 2008; Schirrmeister et al., 2011; Kanevskiy et al., 2011; Vonk et al., 2012). Yedoma deposits are accumulations of mainly windblown dust, or occasionally alluvial or colluvial sediments, that settled in the interglacial periods and froze once air temperatures dropped below zero. Yedoma ice content can be as high as 80–90 %, including both segregated ice within soil and massive ice wedges that can reach depths of 50 m or more (Kanevskiy et al., 2013). Mineral horizons in Yedoma typically contain over 70 % silt, although sandy deposits also occur. Fast sedimentation rates and simultaneous freezing often buried and preserved organic matter that is typically more decomposable than that of non-permafrost mineral soils. The concentrations of OC in these deposits average 1–5 % (Strauss et al., 2013), but concentrations can reach 15 % in peaty horizons within some deposits (Schirrmeister et al., 2002). Strauss et al. (2013) estimated an OC pool of 83 (+61/ − 57) Gt C for the Yedoma regions of Siberia and Alaska, with an average OC density of 19 (+13/ − 11) kg C m^{-3} that was reduced to 10 (+7/−6) kg C m^{-3} when ice-wedge volumes were included in the estimate. The large ice content of Yedoma makes this soil particularly vulnerable to climate warming, and deep thermokarst lakes are common on the Yedoma landscape. Some thermokarst lakes eventually turn into drained basins, or "alases", a Yakutian term for a depression with deep peat (Smith et al., 1995).

5 Quantity of organic carbon in permafrost soils

Major efforts to estimate global soil OC storage during the twentieth century yielded values ranging from 400 to 9120 Pg C stored in soils worldwide (Amundson, 2001). Amundson (2001) calls attention to the wide variation in the amounts of soils data, analyzed depths, and summation methods used in these early analyses of the global soil OC pool. Some estimates applied overall soil averages to the entire land surface (Rubey, 1951), others averaged storage according to ecosystem types (Schlesinger, 1977; Post et al., 1982; Jobbágy and Jackson, 2000), while others were based on soil maps and soil taxonomy (Atjay et al., 1979; Bohn, 1982; Eswaran et al., 1993). However, the influence of permafrost on soil OC storage was not directly recognized until relatively recently in studies such as Michaelson et al. (1996, 2013), Bliss and Maursetter (2010), Johnson et al. (2011), Ping et al. (2008a), and Mishra and Riley (2012) for Alaska and North America, Stoblovoi (2002) for Eurasia, and Tarnocai et al. (2009) for

SOILD
1, 709–756, 2014

the entire northern circumpolar region. Indeed, the Gelisol order of Soil Taxonomy was only established in 1999 (Soil Survey Staff, 1999).

As more soil pedon data became available, especially those evaluated for soil OC storage, estimates of the permafrost-region soil OC pool increased substantially – primarily due to a fuller accounting of deeper cryoturbated OC and permafrost OC stocks. As an example, Jobbágy and Jackson (2000) estimated only 294 Pg C were stored at 0–3 m depth in the boreal forest and tundra biomes combined; whereas, Tarnocai et al. (2009) estimated 1024 Pg C stored at 0–3 m for the northern circumpolar permafrost region. At present, the largest spatially distributed soils data set from the northern circumpolar region is the Northern Circumpolar Soil Carbon Database (NCSCD), which was initially developed for the Tarnocai et al. (2009) estimate. The NCSCD (Hugelius et al., 2014) was recently updated with new data for soil depths of 1–3 m as well as deeper Yedoma and deltaic-alluvium deposits. They used the updated NCSCD to produce the current best overall estimate of 1300–1370 Pg C stored in regional soils (0–3 m depth together with C in the deeper Yedoma and deltaic deposits). However, this estimate is not well constrained as it has a substantial uncertainty range (819–1690 Pg C) and significant data gaps (Hugelius et al., 2014). Among the reasons for this large uncertainty range are the remoteness and vastness of the region (17.8×10^6 km^2 or 16 % of global soil area), which is represented in the NCSCD by very sparse C data available from only a total of 341 Gelisol pedons and 177 pedons of non-permafrost soils (Hugelius et al., 2014).

In addition to the low sampling density due to difficulties associated with remote access and the vast extent of the circumpolar permafrost area, there are technical and sampling challenges unique to Gelisols. These challenges can contribute to increased variability and decreased confidence in C storage calculations and overall estimates for Gelisols. Two of the more significant technical challenges are the lack of measured soil bulk densities (Db) and the differences in methods of soil C determination. Both of these challenges are common in most soil data bases, necessitating the estimation of these key soil properties essential to soil C storage calculation (Tarnocai et al., 2009; Johnson et al., 2011; Hugelius et al., 2012; Mishra and Riley, 2012). Michaelson et al. (2013) found that the commonly used chemical oxidation method for determination of soil C concentration in older datasets (used for some pedons by both Tarnocai et al., 2009, and Hugelius et al., 2013) can over-estimate the OC in Gelisols by an average of 12 % ($R^2 = 0.98$) compared to more modern ignition methods. Until new pedotransfer functions were developed and used by Michaelson et al. (2013) to update Alaska USDA-NRCS soil pedon data, the C contents of 57 % of the pedons representing Alaska were determined only by the chemical oxidation method, and thus were over-estimated. In addition, soil C concentrations are commonly used to estimate and fill in missing soil Db data. But for Gelisols the correlation between soil C concentration and Db is poor in the permafrost layer ($R^2 = 0.46$) because of the large and variable ice contents (Michaelson et al., 2013), especially in the often C-rich transition layer of the upper permafrost. Ping et al. (1998, 2013) outline state-of-the-art protocols for permafrost soils that address the issues of sampling across pedon cryogenic patterns of variability, calculation of C storage in cryoturbated profiles, and the sampling and measuring of Db in frozen Gelisol horizons.

Another major issue affecting estimates of the permafrost-region soil C pool is fine-scale vertical and horizontal heterogeneity. Estimates of the vertical distribution of OC stocks in permafrost region soils are extremely important for predicting the amount of C that will become vulnerable to mineralization upon top-down thaw or losses due to increased incidence of fire (Koven et al., 2011; Harden et al., 2012). Yet, the content and distribution of C within the soil profile vary widely among Histels, Turbels, and Orthels (Harden et al., 2012), making this variation a major source of uncertainty in current estimates. In addition, as described above, patterned-ground features can exhibit various combinations of these suborder soils, resulting in large local and landscape heterogeneity. A number of studies have looked at the C content of patterned soil ground features, for example frost boils (Dyke and Zoltai, 1980; Walker et al., 2004; Kaiser et al., 2005; Michaelson et al., 2012), circles (Kimble et al., 1993; Hallet and Prestrud, 1986), stripes (Walmsley and Lavkulich, 1975, Horwath et al., 2008), ice-wedge polygons

SOILD
1, 709–756, 2014

(Ping et al., 2014; Zubrzycki et al., 2013), and earth hummocks (Kimble et al., 1993; Landi et al., 2004). However, explicit efforts to account for and incorporate local-scale spatial heterogeneity into sampling designs and up-scaling approaches are only beginning to be considered in the estimation of soil OC stocks at landscape or regional scales (Horwath et al., 2008; Zubrzycki et al., 2013).

6 Characterization of the quality and decomposability of organic matter in permafrost soils

Soil organic matter is a complex and heterogeneous mixture derived from plant litter inputs and microbial residues that exist in various stages of decomposition and associations with soil minerals (Baldock and Skjemstad, 2000; Sutton and Sposito, 2005; Kelleher and Simpson, 2006; Jastrow et al., 2007; Lehmann et al., 2008). Traditionally, the "quality" of SOM (and by inference its intrinsic potential to be further decomposed, transformed and mineralized) has been evaluated by assessing its molecular composition and by applying some type of physical, chemical, or biological fractionation approach to partition the bulk SOM pool into "labile" vs. "recalcitrant" or "stable" forms (Kleber and Johnson, 2010; von Lützow et al., 2007; Simpson and Simpson, 2012). Although the chemical composition of SOM can affect the rate of decomposition, it is now recognized that even readily degradable "labile" C forms can be stabilized in soil by a variety of mechanisms or conditions that physically or chemically limit microbial access to the C or impact microbial activity, and that even the most "recalcitrant" C forms can be mineralized given the "right" conditions (von Lützow et al., 2006; Marschner et al., 2008; Kleber, 2010).

Thus, while precise definitions and standardized measurements of "labile", "recalcitrant", or "stable" SOM pools are not possible, these terms are nonetheless useful in a relative sense for comparing the degradation state or potential decomposability of SOM – particularly for permafrost-affected soils. This is because, for permafrost soils, the dominant factor limiting decomposer activity is the cold and often wet anoxic

SOILD
1, 709–756, 2014

environments (Davidson and Janssens, 2006). In addition, low soil pH can limit decomposition rates in some soils (Grosse et al., 2011). Hence in permafrost soils, organic matter is often preserved in a relatively undecomposed state both in surface horizons and in buried forms (e.g., peat deposits, cryoturbated organic matter). In addition, much of this organic matter is uncomplexed or only poorly associated with soil minerals (Diochon et al., 2013; Höfle et al., 2013). Furthermore, subzero microbial activity in unfrozen water films can lead to accumulation of easily decomposable, soluble byproducts in frozen soils, and soluble organics produced in surface horizons (e.g., root exudates) might accumulate due to drainage limitations imposed by the permafrost (Michaelson et al., 1998; Mikan et al., 2002; Michaelson and Ping, 2003). Alternatively, in some soil environments, cryochemical precipitation of dissolved OC might result in the deposition of rather stable SOM near the permafrost table (Gundelwein et al., 2007).

Because so much SOM is currently stabilized simply by environmental conditions, most efforts to evaluate the quality of OC stored in permafrost region soils have focused on characterizing the chemical composition of the organic matter (e.g., Dai et al., 2002a; Turetsky et al., 2007; Pedersen et al., 2011). Chemical characterization studies have run the gamut of methodologies in efforts to assess the current state of degradation – from wet chemical fractionations (acid/base extractions, molecular biomarkers, and water or solvent extractions; e.g., Uhlířová et al., 2007; Paulter et al., 2010; Hugelius et al., 2012) to various types of spectrochemical analysis (nuclear magnetic resonance spectroscopy, pyrolysis-gas chromatography/mass spectrometry, mid-infrared spectroscopy; e.g., Dai et al., 2002a, b; Anderson and White, 2006; Waldrop et al., 2010; Pedersen et al., 2011; Pengerud et al., 2013). Even though organomineral associations might play a role in the relative persistence of SOM upon thawing and warming (Davidson and Janssens, 2006), relatively few studies have employed physical fractionations to characterize the amount of soil OC stabilized by mineral associations or by aggregation (e.g., Dutta et al., 2006; Xu et al., 2009b; Höfle et al., 2013).

SOILD
1, 709–756, 2014

These varied techniques provide different perspectives into the composition of organic matter stored in permafrost soils and its potential resistance to decomposition. In general, wet chemical and physical fractionations enable estimates of pool sizes for SOM of different qualities, whereas spectrochemical analyses provide more specific information on SOM chemistry. From the perspective of the permafrost region as a whole, applications of these methods to a range of soils and horizons provide similar overall conclusions for permafrost soils. In general, wet chemical and spectrochemical methods indicate (1) the bulk of the organic matter in permafrost-affected soils is relatively less decomposed or "humified" than that of more temperate soils; (2) plant-derived materials are often more prevalent than microbial residues; and (3) environmental conditions have promoted the preservation of relatively labile SOM constituents (e.g., Ping et al., 1997; Dai et al., 2002a, b; Pedersen et al., 2011; Hugelius et al., 2012; Paulter et al., 2010). Physical fractionation approaches have provided comparable information. For example, Diochon et al. (2013) found that lightly decomposed particulate organic matter (POM) was a greater component (usually $> 30\,\%$) of the total soil OC in active-layer and permafrost horizons of Canadian Turbic Cryosols than is typical for temperate soils. For similar horizons in two Alaskan Turbels, Xu et al. (2009a, b) reported even greater proportions of total soil OC ($> 70\,\%$) in the POM fractions, and analysis of this material by pyrolysis-gas chromatography/mass spectrometry indicated it was only lightly decomposed. Moreover, Höfle et al. (2013) concluded that organomineral associations and aggregation are of lesser importance for SOM stabilization in permafrost soils than in temperate and tropical soils. Indeed, this is not surprising because most clay minerals in Arctic tundra soils are inherent from the parent material rather than pedogenic (Borden et al., 2010). The lack of reactivity of these clay minerals was further demonstrated by the lack of correlation between clay content and cation exchange capacity (Ping et al., 2005c).

More direct assessments of the potential decomposability of SOM have been made by measuring soil OC mineralization in laboratory incubation studies (e.g., Michaelson and Ping, 2003; Lee et al., 2012; Elberling et al., 2013; Knoblauch et al., 2013). To

SOILD
1, 709–756, 2014

date, most incubation studies have been relatively short term (weeks to a few months) and thus are assessments of the most bioavailable components of SOM. Some studies have combined chemical or physical characterizations with incubations in an effort to relate SOM composition to decomposition rates or to identify indices of decomposability (e.g., White et al., 2002, 2004; Weintraub and Schimel, 2003; Waldrop et al., 2010; Diochon et al., 2013; Paré and Bedard-Haughn et al., 2013; Pengerud et al., 2013; Treat et al., 2014). In general, SOM of higher quality (less-decomposed and rich in polysaccharides and proteins) is positively related to mineralization rates.

Long-term (> 1 year) incubation studies provide better assessments of the integrated effects of multiple SOM chemistries and stabilization mechanisms on mineralization rates and also can inform the estimates of turnover rate functions used in process models. But, to date, the number of studies of this length for permafrost soils are quite limited (Schädel et al., 2014). Initial or short-term mineralization is often related to the amount of dissolved or water-extractable OC, but in longer term incubations, mineralization from thawed permafrost soils has been related to total soil OC concentration and to differences in SOM quality at the time of permafrost incorporation (Dutta et al., 2006; Lee et al., 2012; Knoblauch et al., 2013). In a synthesis of eight long-term aerobic incubation studies (> 1 year) including 121 samples from upland sites in 23 high-latitude ecosystems, Schädel et al. (2014) estimated pool sizes and turnover times from a three-pool decomposition model. They projected 20–90 % loss of initial soil OC within 50 incubation years at 5 °C, with greater losses occurring for soils with higher C : N ratios. However, the importance of oxygen availability on mineralization rates was demonstrated by Elberling et al. (2013). At the end of their 12.5-year incubation study, only 9 % of the initial OC in a saturated upper permafrost soil from a wet grassland site was mineralized compared to 75 % OC lost when the same soil was drained before incubation.

In addition to intrinsic soil differences (i.e., SOM quality and mineral composition) and incubation time, the outcome and conclusions drawn from incubation studies are highly dependent upon experimental conditions including sample handling/disturbance,

SOILD
1, 709–756, 2014

temperature, moisture conditions, oxygen availability, and experimental additions of labile C substrates and nutrients. For example, Dai et al. (2002b) found the most bioreactive compound was polysaccharides at 4 °C whereas at 25 °C more resistant fractions such as lignin were consumed. Such shifts in substrate utilization at higher temperatures likely reflect changes in the active microbial community that may not be realistic under most field conditions. In another case, Wild et al. (2014) showed that the mineralization rates of surface organic soil, mineral subsoil, and subsoil cyroturbated organic material (all from the active layer) were each differentially limited by the availability of labile OC substrates or nitrogen. Importantly, however, many of the conditions impacting mineralization rates in incubation studies are indicative of factors responsible for variations in overall OC storage and the intrinsic relative degradation state of SOM at subregion, landscape, and local scales (e.g., Hugelius et al., 2012; Paré and Bedard-Haughn et al., 2013; Pengerud et al., 2013).

7 Soil organic carbon stocks of the permafrost region in the context of climatic change

Climatic change at high latitudes is causing region-wide warming, hydrologic changes, and other related disturbances (e.g., fires) that are triggering widespread degradation and thawing of permafrost with potential global impacts (Jorgenson et al., 2010; Romanovsky et al., 2010; Rowland et al., 2010). One of the most likely and important consequences of sustained warming in circumpolar regions is the thawing of permafrost soils and subsequent release of CO_2 and CH_4 to the atmosphere due to enhanced microbial mineralization of previously frozen soil OC stocks (Oechel et al., 1993; Zimov et al., 1993; Goulden and Crill, 1997; Melillo et al., 2002; Eliasson et al., 2005; Zhuang et al., 2007; Schuur et al., 2009, 2011).

Model simulations of C losses from thawing permafrost are highly uncertain and vary widely – with predictions of cumulative net transfers to the atmosphere ranging from 7 to 250 Pg C by 2100, 121 to 302 Pg C by 2200, and 180 to 380 Pg C by 2300 for intermediate to high fossil fuel emission scenarios (Zhuang et al., 2006; Koven et al., 2011; Schaefer et al., 2011; Burke et al., 2012; MacDougall et al., 2012; Schneider von Deimling et al., 2012). Such large releases of greenhouse gases are expected to have a positive feedback to Earth's atmosphere, leading to further warming (Schuur et al., 2008, 2011; Schaefer et al., 2011; Burke et al., 2012; MacDougall et al., 2012; Schneider von Deimling et al., 2012). But there are large uncertainties in model predictions of C-climate feedbacks caused by warming of northern circumpolar regions (Koven et al., 2011; Burke et al., 2012). Further, a discrepancy exists between the baseline soil OC stock estimates generated by Earth system models and observation-based estimates of soil OC stocks for the permafrost region (Mishra et al., 2013). Resolving these issues will require improved empirical estimates as well as better model representations of the unique processes controlling the formation and stabilization of soil OC stocks in permafrost regions (Mishra et al., 2013).

In addition to the uncertainties surrounding current estimates of OC stocks, even less is known about the potential vulnerability of OC stored in permafrost soils to climatic change, and how this varies across land cover classes or soil types within different ecoregions (Schuur et al., 2008; Kuhry et al., 2010). Upon thawing, the initial intrinsic degradation state of previously frozen soil OC pools depends on the type of permafrost formation, the origin and chemistry of the SOM, and the extent of mineralization and transformation that occurred before these materials were incorporated into permafrost (Kuhry et al., 2009; Hugelius et al., 2012; Knoblauch et al., 2013). Often SOM stored in permafrost has undergone some level of decay before its incorporation in perennially frozen horizons (Pedersen et al., 2011; Hugelius et al., 2012). This may be particularly true for SOM stabilized in epigenetic permafrost deposits (permafrost formed after deposition of soil material) (Schuur et al., 2008). But, in cryoturbated soils or syngenetic permafrost deposits (permafrost formed more or less concurrently with deposition of soil material, such as Yedoma or some peat deposits), relatively undecomposed organic materials were sometimes rapidly buried and frozen (Zimov et al., 2006; Sannel and Kuhry, 2009). Even in epigenetic permafrost, the extent of SOM degradation can

SOILD
1, 709–756, 2014

be limited if anoxia or other factors constrained decomposition before permafrost formation (Hugelius et al., 2012).

Thus, there is a need to develop biogeochemical indicators that reflect differences in the genesis and past history of SOM before incorporation into permafrost and that represent its intrinsic relative degradation state (Kuhry et al., 2010; Hugelius et al., 2012). With such indicators, maps of the spatial and vertical distributions of soil OC stocks could be coupled with an indication of the potential decomposability of these C stocks. Taken together, this information would substantially improve observationally based predictions of the impacts of soil warming and permafrost thawing and contribute significantly to the calibration and validation of regional and Earth system models (Schuur et al., 2008; Burke et al., 2012; Hugelius et al., 2012; Mishra et al., 2013).

Ultimately, the vulnerability of OC stocks in permafrost soils to changing climatic conditions will depend on the interactions of SOM composition with numerous other controlling factors that also are likely to respond to climatic changes (such as temperature, aeration, water and nutrient availability, new C inputs from changing plant communities, and changing associations with soil minerals). These interactions will affect the integrated activities of the microbial community, as well as the physical access of decomposers and their enzymes to thawed C pools (Davidson and Janssens, 2006; Schmidt et al., 2011). All of these factors and interactions can only be assessed through a combination of laboratory studies, integrated observational and manipulative field studies, geospatial upscaling and mapping of OC stocks and indicators of their decomposability, as well as ecosystem, regional, and Earth system modeling studies.

Acknowledgements. This material is based upon work supported by the US Department of Agriculture NIFA Hatch funds and by the Alaska State Soil Classification project (CLP and GJM), and by the US Department of Energy, Office of Science, Office of Biological and Environmental Research under Contract No. DE-AC02-06CH11357 (JDJ).

References

Ahrens, R. J., Bockheim, J. G., and Ping, C. L.: The Gelisol order in soil taxonomy, in: Cryosols – Permafrost-Affected Soils, edited by: Kimble, J. M., Springer-Verlag, Berlin, 627–635, 2004.

Amundson, R.: The carbon budget in soils, Annu. Rev. Earth Planet. Sc., 29, 535–562, 2001.

Andersen, S. K. and White, D. M.: Determining soil organic matter quality under anaerobic conditions in arctic and subarctic soils, Cold Reg. Sci. Technol., 44, 149–158, 2006.

Atjay, G. L., Ketner, P., and Duvigneaud, P.: Terrestrial primary production and phytomass, in: The Global Carbon Cycle, edited by: Bolin, B., Degens, E. T., Kempe, S., and Ketner, P., SCOPE, 13, Wiley, Chichester, UK, 129–182, 1979.

Baldock, J. A. and Skjemstad, J. O.: Role of the soil matrix and minerals in protecting natural organic materials against biological attack, Org. Geochem., 31, 697–710, 2000.

Beyer, L., Bockheim, J. G., Campbell, I. B., and Claridge, G. G. C.: Genesis, properties and sensitivity of Antarctic Gelisols, Antarct. Sci., 11, 387–398, 1999.

Beyer, L., Pingpank, K., Wriedt, G., and Bölter, M.: Soil formation in coastal continental Antarctica (Wilkes Land), Geoderma, 95, 283–304, 2000.

Bliss, N. B. and Maursetter, J.: Soil organic carbon stocks in Alaska estimated with spatial and pedon data, Soil Sci. Soc. Am. J., 74, 565–579, 2010.

Blume, H. P., Beyer, L., Bölter, M., Erlenkeuser, H., Kalk, E., Kneesch, S., and Schneider, D.: Pedogenic zonation in soils of the Southern Circum-Polar Region, Adv. GeoEcol., 30, 69–90, 1997.

Bockheim, J. G.: Soil development rates in the Transantarctic Mountains, Geoderma, 47, 59–77, 1990.

Bockheim, J. G.: Permafrost distribution in the southern circumpolar region and its relation to the environment: a review and recommendations for further research, Permafrost Periglac., 6, 27–45, 1995.

Bockheim, J. G. and Hinkel, K. M.: The importance of "deep" organic carbon in permafrost-affected soils of arctic Alaska, Soil Sci. Soc. Am. J., 71, 1889–1892, 2007.

Bockheim, J. G. and Tarnocai, C.: Recognition of cryoturbation for classifying permafrost-affected soils, Geoderma, 81, 281–293, 1998.

Bockheim, J. G., Walker, D. A., Everett, L. R., Nelson, F. E., and Shiklomanov, N. I.: Soils and cryoturbation in moist nonacidic and acidic tundra in the Kuparuk River Basin, arctic Alaska, U.S.A., Arctic Alpine Res., 30, 166–174, 1998.

Bockheim, J. G., Everett, L. R., Hinkel, K. M., Nelson, F. E., and Brown, J.: Soil organic carbon storage and distribution in arctic tundra, Barrow, Alaska, Soil Sci. Soc. Am. J., 63, 934–940, 1999.

Bohn, H. L.: Estimate of organic carbon in world soils: II, Soil Sci. Soc. Am. J., 46, 1118–1119, 1982.

Boike, J., Kattenstroth, B., Abramova, K., Bornemann, N., Chetverova, A., Fedorova, I., Fröb, K., Grigoriev, M., Grüber, M., Kutzbach, L., Langer, M., Minke, M., Muster, S., Piel, K., Pfeiffer, E.-M., Stoof, G., Westermann, S., Wischnewski, K., Wille, C., and Hubberten, H.-W.: Baseline characteristics of climate, permafrost and land cover from a new permafrost observatory in the Lena River Delta, Siberia (1998–2011), Biogeosciences, 10, 2105–2128, doi:10.5194/bg-10-2105-2013, 2013.

Borden, P. W., Ping, C. L., McCarthy, P. J., and Naidu, S.: Clay mineralogy in arctic tundra Gelisols, northern Alaska, Soil Sci. Soc. Am. J., 74, 580–592, 2010.

Brink, V. C., Pearce, D. G., Mackay, J. R., and Freyman, S.: Needle ice and seedling establishment in southwestern British Columbia, Can. J. Plant Sci., 47, 135–139, 1967.

Broll, G. and Tarnocai, C.: Turf hummocks on Ellesmere Island, Canada, in: Transactions of the 17th World Congress of Soil Science, Bangkok, Thailand, 14–21 August 2002, CD-ROM paper 1049, 2002.

Brown, J.: Radiocarbon dating, Barrow, Alaska, Arctic, 18, 36–48, 1965.

Brown, J.: Soil properties developed on the complex tundra relief of northern Alaska, Biuletyn Perygacjalny, 18, 153–167, 1969.

Brown, J. and Kreig, R. A. (Eds.): Guidebook to Permafrost and Related Features Along the Elliott and Dalton Highways, Fox to Prudhoe Bay, Alaska, Alaska Division of Geological and Geophysical Surveys, Department of Natural Resources, Fairbanks, Alaska, 1983.

Brown, J., Ferrians Jr., O. J., Heginbottom, J. A., and Melnikov, E. S.: Circum-Arctic Map of Permafrost and Ground Ice Conditions, National Snow and Ice Data Center, Boulder, Colorado, Digital media, 1998, revised February 2001.

Burke, E. J., Hartley, I. P., and Jones, C. D.: Uncertainties in the global temperature change caused by carbon release from permafrost thawing, The Cryosphere, 6, 1063–1076, doi:10.5194/tc-6-1063-2012, 2012.

Burn, C. R.: The active layer: two contrasting definitions, Permafrost Periglac., 9, 411–416, 1998.

Dai, X. Y., Ping, C. L., and Michaelson, G. J.: Characterization of soil organic matter in Arctic tundra soils by different analytical approaches, Organic Geochem., 33, 407–419, 2002a.
Dai, X. Y., White, D., and Ping, C. L.: Evaluation of soil organic matter composition and bioavailability by pyrolysis-gas chromatography/mass spectrometry, J. Anal. Appl. Pyrol., 62, 249–258, 2002b.
Davidson, E. A. and Janssens, I. A.: Temperature sensitivity of soil carbon decomposition and feedbacks to climate change, Nature 440, 165–173, 2006.
Davis, N.: Permafrost – A Guide to Frozen Ground in Transition, University of Alaska Press, Fairbanks, Alaska, 351 pp., 2001.
Diochon, A., Gregorich, E. G., and Tarnocai, C.: Evaluating the quantity and biodegradability of soil organic matter in some Canadian Turbic Cryosols, Geoderma, 202–203, 82–87, 2013.
Dutta, K., Schuur, E. A. G., Neff, J. C., and Zimov, S. A.: Potential carbon release from permafrost soils of Northeastern Siberia, Glob. Change Biol., 12, 2336–2351, 2006.
Dyke, A. S. and Zoltai, S. C.: Radiocarbon-dated mudboils, central Canadian Arctic. Current Research, Part B, Geological Survey of Canada, Paper 80-1B, 271–275, 1980.
Elberling, B., Michelsen, A., Schädel, C., Schuur, E. A. G., Christiansen, H. H., Berg, L., Tamstorf, M. P., and Sigsgaard, C.: Long-term CO_2 production following permafrost thaw, Nature Climate Change 3, 890–894, 2013.
Eliasson, P. E., McMurtrie, R. E., Pepper, D. A., Stromgren, M., Linder, S., and Agren, G. I.: The response of heterotrophic CO_2 flux to soil warming, Glob. Change Biol., 11, 167–181, 2005.
Eswaran, H., Van Den Berg, E., and Reich, P.: Organic carbon in soils of the world, Soil Sci. Soc. Am. J., 57, 192–194, 1993.
Everett, K. R. and Brown, J.: Some recent trends in the physical and chemical characterization and mapping of tundra soils, arctic slope of Alaska, Soil Sci., 133, 264–280, 1982.
Fahnestock, J. T., Jones, M. H., and Welker, J. M.: Wintertime CO_2 efflux from arctic soils: Implications for annual carbon budgets, Global Biogeochem. Cy., 13, 775–779, 1999.
French, H. M.: The Periglacial Environment, 3rd Edn., John Wiley & Sons, 2007.
French, H. and Shur, Y.: The principles of cryostratigraphy, Earth Sci. Rev., 101, 190–206, doi:10.1016/j.earscirev.2010.04.002, 2010.
Froese, D. G., Westgate, J. A., Reyes, A. V., Enkin, R. J., and Preece, S. J.: Ancient permafrost and a future, warmer Arctic, Science 321, p. 1648, doi:10.1126/science.1157525, 2008.

SOILD
1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Geisler, E. and Ping, C. L.: Effects of periglacial processes on soil formation in subarctic Alaska, ASA, CSSA, SSSA International Annual Meetings, Tampa, FL, 3–6 November 2013, Abstract no. 126-3, 2013.

Gong, Z. T., Chen, Z. C., Lou, G. B., Zhang, G. L., and Zhao, W. J.: The reference of Chinese soil taxonomy, Soil, 31, 57–63, 1999.

Goulden, M. L. and Crill, P. M.: Automated measurements of CO_2 exchange at the moss surface of a black spruce forest, Tree Physiol., 17, 537–542, 1997.

Gracheva, R. G.: Cryosols of the mountains of Southern Siberia and far Eastern Russia, in: Cryosols – Permafrost-Affected Soils, edited by: Kimble, J. M., Springer-Verlag, Berlin, 231–252, 2004.

Grosse, G., Harden, J., Turetsky, M., McGuire, A. D., Camill, P., Tarnocai, C., Frolking, S., Schuur, E. A. G., Jorgenson, T., Marchenko, S., Romanovsky, V., Wickland, K. P., French, N., Waldrop, M. , Bourgeau-Chavez, L., and Striegl, R. G.: Vulnerability of high-latitude soil organic carbon in North America to disturbance, J. Geophy. Res., 116, G00K06, doi:10.1029/2010JG001507, 2011.

Gubin, S. V.: Late Pleistocene soil formation on coastal lowlands of northern Yakutia, Pochvovedeniye, 10, 62–70, 1993 (in Russian).

Gundelwein, A., Müller-Lupp, T., Sommerkorn, M., Haupt, E. T. K., Pfeiffer, E.-M., and Wiechmann, H.: Carbon in tundra soils in the Lake Labaz region of arctic Siberia, Eur. J. Soil Sci., 58, 1164–1174, 2007.

Hallet, B. and Prestrud, S.: Dynamics of periglacial sorted circles in Western Spitsbergen, Quaternary Res., 26, 81–99, 1986.

Harden, J. W., Koven, Ping, C. L., Hugelius, G., McGuire, A. D., Michaelson, G. J., Kanevskiy, M., Kuhry, P., O'Donnell, J., Tarnocai, C., and Jorgenson, T.: Field information links permafrost carbon to physical vulnerabilities of thawing, Geophy. Res. Lett., 39, L15704, doi:10.1029/2012GL051958, 2012.

Hinkel, K. M., Eisner, W. E., Bockheim, J. G., Nelson, F. E., Peterson, K. M., and Dai, X. Y.: Spatial extent, age, and carbon stocks in drained thaw lake basins on the Barrow Peninsula, Arct. Antarct. Alp. Res., 35, 291–300, 2003.

Hinkel, K. M., Frohn, R. C., Nelson, F. E., Eisner, W. R., and Beck, R. A.: Morphometric and spatial analysis of thaw lakes and drained thaw lake basins in the western Arctic Coastal Plain, Alaska, Permafrost Periglac., 16, 327–341, 2005.

SOILD

1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Höfle, S., Rethemeyer, J., Mueller, C. W., and John, S.: Organic matter composition and stabilization in a polygonal tundra soil of the Lena Delta, Biogeosciences, 10, 3145–3158, doi:10.5194/bg-10-3145-2013, 2013.

Horwath, J. L., Sletten, R. S., Hagedon, B., and Hallet, B.: Spatial and temporal distribution of soil organic carbon in nonsorted striped patterned ground of the High Arctic, J. Geophy. Res. 113, G03S07, doi:10.1029/2007JG000511, 2008.

Hugelius, G., Routh, J., Kuhry, P., and Crill, P.: Mapping the degree of decomposition and thaw remobilization potential of soil organic matter in discontinuous permafrost terrain, J. Geophy. Res., 117, G02030, doi:10.1029/2011JG001873, 2012.

Hugelius, G., Tarnocai, C., Broll, G., Canadell, J. G., Kuhry, P., and Swanson, D. K.: The Northern Circumpolar Soil Carbon Database: spatially distributed datasets of soil coverage and soil carbon storage in the northern permafrost regions, Earth Syst. Sci. Data, 5, 3–13, doi:10.5194/essd-5-3-2013, 2013.

Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J. W., Schuur, E. A. G., Ping, C. L., Schirrmeister, L., Grosse, G., Michaelson, G. J., Koven, C. D., O'Donnell, J. A., Elberling, B., Mishra, U., Camill, P., Yu, Z., Palmtag, J., and Kuhry, P.: Improved estimates show large circumpolar stocks of permafrost carbon while quantifying substantial uncertainty ranges and identifying remaining data gaps, Biogeosciences Discuss., 11, 4771–4822, doi:10.5194/bgd-11-4771-2014, 2014.

IUSS Working Group WRB: World Reference Base for Soil Resources 2006, World Soil Resources Reports No. 103, FAO, Rome, 2006.

Jastrow, J. D., Amonette, J. E., and Bailey, V. L.: Mechanisms controlling soil carbon turnover and their potential application for enhancing carbon sequestration, Climatic Change, 80, 5–13, 2007.

Jobbágy, E. G. and Jackson, R. B.: The vertical distribution of soil organic carbon and its relation to climate and vegetation, Ecol. Appl., 10, 423–436, 2000.

Johnson, K. D., Harden, J., McGuire, A. D., Bliss, N. B., Bockheim, J. G., Clark, M., Nettleton-Hollingsworth, T., Jorgenson, M. T., Kane, E. S., Mack, M., O'Donnell, J., Ping, C. L., Schuur, E. A. G., Turetsky, M. R., and Valentine, D. W.: Soil carbon distribution in Alaska in relation to soil-forming factors, Geoderma, 167–168, 71–84, 2011.

Jorgenson, M. T., Shur, Y. L., and Pullman, E. R.: Abrupt increase in permafrost degradation in Arctic Alaska, Geophys. Res. Lett., 33, L02503, doi:10.1029/2005GL024960, 2006.

Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S.: Resilience and vulnerability of permafrost to climate change, Can. J. Forest Res., 40, 1219–1236, 2010.

Jorgenson, M. T., Harden, J., Kanevskiy, M., O'Donnell, J., Wickland, K., Ewing, S., Manies, K., Zhuang, Q., Shur, Y., Striegl, R., and Koch, J.: Reorganization of vegetation, hydrology and soil carbon after permafrost degradation across heterogeneous boreal landscapes, Environ. Res. Lett., 8, 035017, doi:10.1088/1748-9326/8/3/035017, 2013.

Kade, A. and Walker, D. A.: Experimental alteration of vegetation on nonsorted circles: Effects on cryogenic activity and implications for climate change in the Arctic, Arct. Antarct. Alp. Res., 40, 96–103, doi:10.1657/1523-0430(06-029)[KADE]2.0.CO;2, 2008.

Kaiser, C., Meyer, H., Biasi, C., Rusalimova, O., Barsukov, P., and Richter, A.: Storage and mineralization of carbon and nitrogen in soils of a frost-boil tundra ecosystem in Siberia, J. Appl. Soil Ecol., 29, 173–183, 2005.

Kanevskiy, M., Shur, Y., Fortier, D., Jorgenson, M. T., and Stephani, E.: Cryostratigraphy of late Pleistocene syngenetic permafrost (yedoma) in northern Alaska, Itkillik River exposure, Quaternary Res., 75, 584–596, doi:10.1016/j.yqres.2010.12.003, 2011.

Kanevskiy, M., Shur, Y., Jorgenson, M. T., Ping, C. L., Michaelson, G. J., Fortier, D., Stephani, E., Dillon, T., and Tumskoy, V.: Ground ice of the upper permafrost of the Beaufort Sea coast of Alaska, Cold Reg. Sci. Technol., 85, 56–70, doi:10.1016/j.coldregions.2012.08.002, 2013.

Kanevskiy, M., Jorgenson, T., Shur, Y., O'Donnell, J. A., Harden, J. W., Zhuang, Q., and Fortier, D.: Cryostratigraphy and permafrost evolution in the lacustrine lowlands of West-Central Alaska, Permafrost Periglac., 25, 14–34, 2014.

Kelleher, B. P. and Simpson, A. J.: Humic substances in soils: Are they really chemically distinct?, Environ. Sci. Technol., 40, 4605–4611, 2006.

Kimble, J. M., Tarnocai, C., Ping, C. L., Ahrens, R., Smith, C. A. S., Moore, J. P., and Lynn, W.: Determination of the amount of carbon in highly cryoturbated soils, in: Proceedings Joint Russian-American Seminar on Cryopedology and Global Change, Pushchino, Russia, 15–16 November 1992, edited by: Gilichinsky, D., Russian Academy of Sciences, Moscow, 277–291, 1993.

Kleber, M.: What is recalcitrant soil organic matter?, Environ. Chem., 7, 320–332, 2010.

Kleber, M. and Johnson, M. G.: Advances in understanding the molecular structure of soil organic matter: Implications for interactions in the environment, Adv. Agron., 106, 77–142, 2010.

SOILD

1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Knoblauch, C., Beer, C., Sosnin, A., Wagner, D., and Pfeiffer, E.-M.: Predicting long-term carbon mineralization and trace gas production from thawing permafrost of Northeast Siberia, Glob. Change Biol., 19, 1160–1172, 2013.

Koven, C. D., Ringeval, B., Friedlingstein, P., Ciais, P., Cadule, P., Khvorostyanov, D., Krinner, G., and Tarnocai, C.: Permafrost carbon-climate feedbacks accelerate global warming, P. Natl. Acad. Sci. USA, 108, 14769–14774, 2011.

Kuhry, P., Ping, C. L., Schuur, E. A. G., Tarnocai, C., and Zimov, S.: Report from the International Permafrost Association: Carbon pools in permafrost regions, Permafrost Periglac., 20, 229–234, 2009.

Kuhry, P., Dorrepaal, E., Hugelius, G., Schuur, E. A. G., and Tarnocai, C.: Potential remobilization of belowground permafrost carbon under future global warming, Permafrost Periglac., 21, 208–214, 2010.

Landi, A., Mermut, A. R., and Anderson, D. W.: Carbon distribution in a hummocky landscape from Saskatchewan, Canada, Soil Sci. Soc. Am. J., 68, 175–184, 2004.

Lee, H., Schuur, E. A. G., Inglett, K. S., Lavoie, M., and Chanton, J. P.: The rate of permafrost carbon release under aerobic and anaerobic conditions and its potential effects on climate, Glob. Change Biol., 18, 515–527, 2012.

Lehmann, J., Solomon, D., Kinyangi, J. Dathe, L., Wirick, S., and Jacobsen, C.: Spatial complexity of soil organic matter forms at nanometer scales, Nat. Geosci., 1, 238–242, 2008.

Lipson, D. A., Jha, M., Raab, T. K., and Oechel, W. C.: Reduction of iron (III) and humic substances plays a major role in anaerobic respiration in an Arctic peat soil, J. Geophy. Res.-Biogeo., 115, G00I06, doi:10.1029/2009JG001147, 2010.

MacDougall, A. H., Avis, C. A., and Weaver, A. J.: Significant existing commitment to warming from the permafrost carbon feedback, Nat. Geosci., 5, 719–721, 2012.

Mackay, J. R.: Reticulate ice veins in permafrost, Northern Canada, Can. Geotech. J., 11, 230–237, 1974.

Makeev, O. W. and Kerzhentsev, A. S.: Cryogenic processes in the soils of Northern Asia, Geoderma, 12, 101–109, 1974.

Marschner, B., Brodowski, X., Dreves, A., Gleixner, G., Gude, A., Grootes, P. M., Hamer, U., Heim, A., Jandl, G., Ji, R., Kaiser, K., Kalbitz, K., Kramer, C., Leinweber, P., Rethemeyer, J., Schaffer, A., Schmidt, M. W. I., Schwark, L., and Wiesenberg, G. L. B.: How relevant is recalcitrance for the stabilization of organic matter in soils?, J. Plant Nutr. Soil Sc., 171, 91–132, 2008.

SOILD

1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Maximovich, S. V.: Geography and ecology of cryogenic soils of Mongolia, in: Cryosols – Permafrost-Affected Soils, edited by: Kimble, J. M., Springer-Verlag, Berlin, 253–270, 2004.
McGuire, A. D., Anderson, L. G., Christensen, T. R., Dallimore, S., Guo, L., Hayes, D. J., Heimann, M., Lorenson, T. D., Macdonald, R. W., and Roulet, N.: Sensitivity of the carbon cycle in the Arctic to climate change, Ecol. Monogr., 79, 523–555, 2009.
Melillo, J. M., Steudler, P. A., Aber, J. D., Newkirk, K., Lux, H., Bowles, F. P., Catricala, C., Magill, A., Ahrens, T., and Morrisseau, S.: Soil warming and carbon-cycle feedbacks to the climate system, Science, 298, 2173–2176, 2002.
Michaelson, G. J. and Ping, C. L.: Soil organic carbon and CO_2 respiration at subzero temperature in soils of Arctic Alaska, J. Geophys. Res., 108, 8164, doi:10.1029/2001JD000920, 2003.
Michaelson, G. J., Ping, C. L., and Kimble, J. M.: Carbon storage and distribution in tundra soils of Arctic Alaska, USA, Arctic Alpine Res. 28, 414–424, doi:10.2307/1551852, 1996.
Michaelson, G. J., Ping, C. L., Kling, G. W., and Hobbie, J. E.: The character and bioavailability of dissolved organic matter at thaw and in spring runoff waters of the arctic tundra north slope, J. Geophy. Res., 103, 18939–28946, 1998.
Michaelson, G. J., Ping, C. L., Epstein, H. E., Kimble, J. M., and Walker, D. A.: Soils and frost boil ecosystems across the North American Arctic Transect, J. Geophys. Res., 113, G03S11, doi:10.1029/2007JG000672, 2008.
Michaelson, G. J., Ping, C. L., and Walker, D. A.: Soils associated with biotic activity on frost boils in Arctic Alaska, Soil Sci. Soc. Am. J., 76, 2265–2277, doi:10.2136/sssaj2012.0064, 2012.
Michaelson, G. J., Ping, C. L., and Clark, M. H.: Soil pedon carbon and nitrogen data for Alaska: An analysis and update, Open J. Soil Sci., 3, 132–142, doi:10.4236/ojss.2013.32015, 2013.
Mikan, C. J., Schimel, J. P., and Doyle, A. P.: Temperature controls of microbial respiration in arctic tundra soils above and below freezing, Soil Biol. Biochem., 34, 1785–1795, 2002.
Mishra, U. and Riley, W. J.: Alaskan soil carbon stocks: spatial variability and dependence on environmental factors, Biogeosciences, 9, 3637–3645, doi:10.5194/bg-9-3637-2012, 2012.
Mishra, U., Jastrow, J. D., Matamala, R., Hugelius, G., Koven, C. D., Harden, J. W., Ping, C. L., Michaelson, G. J., Fan, Z., Miller, R. M., McGuire, A. D., Tarnocai, C., Kuhry, P., Riley, W. J., Schaefer, K., Schuur, E. A. G., Jorgenson, M. T., and Hinzman, L. D.: Empirical estimates to reduce modeling uncertainties of soil organic carbon in permafrost regions: a review of

recent progress and remaining challenges, Environ. Res. Lett., 8, 035020, doi:10.1088/1748-9326/8/3/035020, 2013.
Munn, L. C.: Soil genesis associated with periglacial ice wedge casts, southcentral Wyoming, Soil Sci. Soc. Am. J., 51, 1000–1004, 1987.
Munn, L. C. and Spackman, L. K.: Origin of silt-enriched alpine surface mantles in Indian Basin, Wyoming, Soil Sci. Soc. Am. J., 54, 1670–1677, 1990.
Murton, J. B. and French, H. M.: Cryostructures in permafrost, Tuktoyaktuk coastlands, western arctic Canada, Can. J. Earth Sci., 31, 737–747, 1994.
Oechel, W. C., Hastings, S. J., Vourlitis, G., Jenkins, M., Riechers, G., and Grulke, N.: Recent change of arctic tundra ecosystems from a net carbon dioxide sink to a source, Nature, 361, 520–523, 1993.
Olefeldt, D., Turetsky, M. R., Crill, P. M., and McGuire, A. D.: Environmental and physical controls on northern terrestrial methane emissions across permafrost zones, Glob. Change Biol., 19, 589–603, doi:10.1111/gcb.12071, 2013.
Paré, M. C. and Bedard-Haughn, A.: Soil organic matter quality influences mineralization and GHG emissions in cryosols: a field-based study of sub-to high Arctic, Glob. Change Biol., 19, 1126–1140, 2013.
Patrick, W. H. and Jugsujinda, A.: Sequential reduction and oxidation of inorganic nitrogen, manganese, and iron in flooded soil, Soil Sci. Soc. Am. J., 56, 1071–1073, 1992.
Pautler, B. G., Simpson, A. J., Mcnally, D. J., Lamoureux, S. F., and Simpson, M. J.: Arctic permafrost active layer detachments stimulate microbial activity and degradation of soil organic matter, Environ. Sci. Technol., 44, 4076–4082, 2010.
Pedersen, J. A., Simpson, M. A., Bockheim, J. G., and Kumar, K.: Characterization of soil organic carbon in drained thaw-lake basin of Arctic Alaska using NMR and FTIR photoacoustic spectroscopy, Org. Geochem., 42, 947–954, 2011.
Pengerud, A., Cécillon, L., Johnsen, L. K., Rasse, D. P., and Strand, L. T.: Permafrost distribution drives soil organic matter stability in a subarctic palsa peatland, Ecosystems 16, 934–947, 2013.
Pettapiece, W. W.: A hummocky permafrost soil from the subarctic of northwestern Canada and some influences of fire, Can. J. Soil Sci., 54, 343–355, 1974.
Péwé, T.: Quaternary Geology of Alaska, US Geological Survey, US Govt. Printing Office, Washington, DC, Geological Survey Professional Paper 835, 1975.

SOILD
1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Ping, C. L.: Gelisols: Part I. Cryogenesis and state factors of formation, Soil Horizons, 54, doi:10.2136/sh2013-54-3-gc, 2013.
Ping, C. L., Lynn, W., and Smith, C. A. S.: Redoximorphic features in permafrost soils, in: Proceedings Joint Russian-American Seminar on Cryopedology and Global Change, Pushchino, Russia, 15–16 November 1992, edited by: Gilichinsky, D., Russian Academy of Sciences, Moscow, 233–244, 1993.
Ping, C. L., Michaelson, G. J., and Cherkinsky, A.: Characterization of soil organic matter by stable isotopes and radiocarbon ages of selected soils in Arctic Alaska, in: The Role of Humic Substances in the Ecosystem and in Environmental Protection, edited by: Drozd, J., Gonet, S. S., Senesi, N., and Weber, J., Polish Society of Humic Substances, Wroclaw, Poland, 475–480, 1997.
Ping, C. L., Bockheim, J. G., Kimble, J. M., Michaelson, G. J., and Walker, D. A.: Characteristics of cryogenic soils along a latitudinal transect in arctic Alaska, J. Geophys. Res., 103, 28917–28928, doi:10.1029/98JD02024, 1998.
Ping, C. L., Michaelson, G. J. Overduin, P. P., and Stiles, C. A.: Morphogenesis of frostboils in the Galbraith Lake area, Arctic Alaska, in: Proceedings of the Eighth International Conference on Permafrost, edited by: Phillips, M., Springman, S. M., and Arenson, L. U., Zürich, Switzerland, 21–25 July 2003, Balkema Publishers, Lisse, The Netherlands, 897–900, 2003.
Ping, C. L., Clark, M. H., Swanson, D. K.: Cryosols in Alaska, in: Cryosols – Permafrost-Affected Soils, edited by: Kimble, J., Springer-Verlag, Berlin, 71–94, 2004.
Ping, C. L., Boone, R., Clark, M. H., Packee, E. C., and Swanson, D. K.: State factor control of soil formation in Interior Alaska, in: Alaska's Changing Boreal Forest, edited by: Chapin III, F. S., Oswood, M. W., van Cleve, K., Viereck, L. A., and Verbyla, D. L., Oxford University Press, New York, 21–38, 2005a.
Ping, C. L., Michaelson, G. J., Kimble, J. M., and Walker, D. A.: Soil acidity and exchange properties of cryogenic soils in Arctic Alaska, Soil Sci. Plant Nutr., 51, 649–653, 2005b.
Ping, C. L., Michaelson, G. J., Packee, E. C., Stiles, C. A., Swanson, D. K., and Yoshikawa, K.: Soil catena sequences and fire ecology in the boreal forest of Alaska, Soil Sci. Soc. Am. J., 69, 1761–1772, 2005c.
Ping, C. L., Michaelson, G. J., Jorgenson, M. T., Kimble, J. M., Epstein, H., Romanovsky, V. E., and Walker, D. A.: High stocks of soil organic carbon in North American Arctic region, Nat. Geosci., 1, 615–619, doi:10.1038/ngeo284, 2008a.

Ping, C. L., Michaelson, G. J., Kimble, J. M., Romanovsky, V. E., Shur, Y. L., Swanson, D. K., and Walker, D. A.: Cryogenesis and soil formation along a bioclimate gradient in Arctic North America, J. Geophys. Res., 113, G03S12, doi:10.1029/2008JG000744, 2008b.

Ping, C. L., Michaelson, G. J., Guo, L., Jorgenson, M. T., Kanevskiy, M., Shur, Y., Dou, F., and Liang, J.: Soil carbon and material fluxes across the eroding Alaska Beaufort Sea coastline, J. Geophy. Res.-Biogeo., 116, G02004, doi:10.1029/2010JG001588, 2011.

Ping, C. L., Clark, M. H., Kimble, J. M., Michaelson, G. J., Shur, Y., and Stiles, C. A.: Sampling protocols for permafrost-affected soils, Soil Horizons, 54, 13–19, 2013.

Ping, C. L., Jastrow, J. D., Jorgenson, M. T., Matamala, R., Michaelson, G. J., and Shur, Y.: Soils and cryostratigraphy of ice wedge polygons in Arctic Alaska, Book of Abstracts of EUCOP4 – 4th European Conference on Permafrost, Evora, Portugal, 18–21 June 2014, 228, 2014.

Post, W. M., Emanuel, W. R., Zinke, P. J., and Stangenberger, A. G.: Soil carbon pools and world life zones, Nature, 298, 156–159, 1982.

Reyes, A. V., Froese, D. G., and Jensen, B. J. L.: Permafrost response to last interglacial warming: field evidence from non-glaciated Yukon and Alaska, Quaternary Sci. Rev., 29, 3256–3274, 2010.

Rieger, S.: The genesis and classifi cation of cold soils, Academic Press, San Diego, CA, 1983.

Rieger, S., Schoephorster, D. B., and Furbush, C. E.: Exploratory soil survey of Alaska, US Department of Agriculture, Soil Conservation Service, Washington, DC, 1979.

Romanovsky, V. E., Marchenko, S. S., Daanen, R., Sergeev, D. O., and Walker, D. A.: Soil climate and frost heave along the Permafrost/Ecological North American Arctic Transect, in: Proceedings of the Ninth International Conference on Permafrost, Vol. 2, Fairbanks, Alaska, 28 June–3 July 2008, edited by: Kane, D. L. and Hinkel, K. M., Institute of Northern Engineering, University of Alaska Fairbanks, 1519–1524, 2008.

Romanovsky, V. E., Smith, S. L., and Christiansen, H. H.: Permafrost thermal state in the polar Northern Hemisphere during the international polar year 2007–2009: A synthesis, Permafrost Periglac., 21, 106–116, 2010.

Rowland, J. C., Jones, C. E., Altmann, G., Bryan, R., Crosby, B. T., Geernaert, G. L., Hinzman, L. D., Kane, D. L., Lawrence, D. M., Mancino, A., Marsh, P., Mcnamara, J. P., Romanovsky, V. E., Toniolo, H., Travis, B. J., Trochim, E., and Wilson, C. J.: Arctic landscapes in transition: Responses to thawing permafrost, Eos, 91, 229–230, 2010.

Rubey, W. W.: Geologic history of sea water an attempt to state the problem, Geol. Soc. Am. Bull., 62, 1111–1148, 1951.

Sannel, A. B. K. and Kuhry, P.: Holocene peat growth and decay dynamics in sub-arctic peat plateaus, west-central Canada, Boreas, 38, 13–24, 2009.
Sannel, A. B. K. and Kuhry, P.: Warming-induced destabilization of peat plateau/thermokarst lake complexes, J. Geophys. Res., 116, G03035, doi:10.1029/2010JG001635, 2011.
Schädel, C., Schuur, E. A. G., Bracho, R., Elberling, B., Knoblauch, C., Lee, H., Luo, Y., Shaver, G. R., and Turetsky, M. R.: Circumpolar assessment of permafrost C quality and its vulnerability over time using long-term incubation data, Glob. Change Biol., 20, 641–652, 2014.
Schaefer, K., Zhang, T., Bruhwiler, L., and Barrett, A. P.: Amount and timing of permafrost carbon release in response to climate warming, Tellus B, 63, 165–180, 2011.
Schirrmeister, L., Siegert, C., Kunitzky, V. V., Grootes, P. M., and Erlenkeuser, H.: Late Quaternary ice-rich permafrost sequences as a paleoenvironmental archive for the Laptev Sea Region in northern Siberia, Int. J. Earth Sci., 91, 154–167, 2002.
Schirrmeister, L., Grosse, G., Wetterich, S., Overduin, P. P., Strauss, J., Schuur, E. A. G., and Hubberten, H.-W.: Fossil organic matter characteristics in permafrost deposits of the northeast Siberian Arctic, J. Geophys. Res., 116, G00M02, doi:10.1029/2011JG001647, 2011.
Schlesinger, W. H.: Carbon balance in terrestrial detritus, Annu. Rev. Ecol. Syst., 8, 51–81, 1977.
Schmidt, M. W., Torn, M. S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I. A., Kleber, M., Kogel-Knabner, I., Lehmann, J., Manning, D. A. C., Nannipieri, P., Rasse, D. P., Weiner, S., and Trumbore, S. E.: Persistence of soil organic matter as an ecosystem property, Nature, 478, 49–55, 2011.
Schneider von Deimling, T., Meinshausen, M., Levermann, A., Huber, V., Frieler, K., Lawrence, D. M., and Brovkin, V.: Estimating the near-surface permafrost-carbon feedback on global warming, Biogeosciences, 9, 649–665, doi:10.5194/bg-9-649-2012, 2012.
Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V., Hagemann, S., Kuhry, P., Lafleur, P. M., Lee, H., Mazhitova, G., Nelson, F. E., Rinke, A., Romanovsky, V. E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J. G., and Zimov, S A.: Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle, BioScience, 58, 701–714, 2008.
Schuur, E. A. G., Vogel, J. G., Crummer, K. G., Lee, H., Sickman, J. O., and Osterkamp, T. E.: The effect of permafrost thaw on old carbon release and net carbon exchange from tundra, Nature, 459, 556–559, doi:10.1038/nature08031, 2009.

SOILD
1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

Schuur, E. A. G., Abbott, B. W., Bowden, W. B., Brovkin, V., Camill, P., Canadell, J. P., Chapin III, F. S., Christensen, T. R., Chanton, J. P., Ciais, P., Crill, P. M., Crosby, B. T., Czimczik, C. I., Grosse, G., Hayes, D. J., Hugelius, G., Jastrow, J. D., Kleinen, T., Koven, C. D., Krinner, G., Kuhry, P., Lawrence, D. M., Natali, S. M., Ping, C. L., Rinke, A., Riley, W. J., Romanovsky, V. E., Sannel, A. B. K., Schädel, C., Schaefer, K., Subin, Z. M., Tarnocai, C., Turetsky, M., Walter Anthony, K. M., Wilson, C. J., and Zimov, S. A.: High risk of permafrost thaw, Nature, 480, 32–33, doi:10.1038/480032a, 2011.

Shilts, W. W.: Nature and genesis of mudboils, central Keewatin, Canada, Can. J. Earth Sci., 15, 1053–1068, 1978.

Shishov, L. L., Tonkonogov, V. D., Lebedeva, I. I., and Gerasimova, M. I.: Classification and diagnostics of soils of Russia, Oikumena, Smolensk, Russia, 2004.

Shur, Y. L.: The upper horizon of permafrost soils, in: Proceedings of the Fifth International Conference on Permafrost, Vol. 1, edited by: Senneset, K., Tapir Publishers, Trondheim, Norway, 867–871, 1988.

Shur, Y. L. and Jorgenson, M. T.: Cryostructure development on the floodplain of the Colville River Delta, northern Alaska, in: Proceedings of Seventh International Conference on Permafrost, Yellowknife, Canada, edited by: Lewkowicz, A. G. and Allard, M., Universite Laval, Centre d'études nordiques, Collection Nordicana, 57, 993–999, 1998.

Shur, Y. L. and Jorgenson, M. T.: Patterns of permafrost formation and degradation in relation to climate and ecosystems, Permafrost Periglac., 18, 7–19, 2007.

Shur, Y. and Osterkmap, T. E.: Thermokarst. Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, AK, Rep. INE 06.11, 50 pp., 2007.

Shur, Y. L. and Ping, C. L.: Permafrost dynamics and soil formation, in: Proceedings of the Meetings on the Classification, Correlation, and Management of Permafrost-affected Soils, edited by: Kimble, J. M. and Ahrens, R. J., USDA Soil Conserv. Serv.-Nat. Soil Surv. Cent., Lincoln, NE, 112–117, 1994.

Shur, Y., Hinkel, K. M., and Nelson, F. E.: The transient layer: Implications for geocryology and climate-change science, Permafrost Periglac., 16, 5–17, doi:10.1002/ppp.518, 2005.

Shur, Y., Jorgenson, T., Kanesvskiy, M., and Ping, C. L.: Formation of frost boils and earth hummocks, in: Ninth International Conference on Permafrost, Fairbanks, Alaska, 28 June–3 July 2008, Extended Abstracts, edited by: Kane, D. L. and Hinkel, K. M., Institute of Northern Engineering, University of Alaska Fairbanks, 287–288, 2008.

SOILD
1, 709–756, 2014

Simpson, M. J. and Simpson, A. J.: The chemical ecology of soil organic matter molecular constituents, J. Chem. Ecol., 38, 768–784, 2012.

Smith, C. A. S, Swanson, D. K., Moore, J. P., Ahrens, R. J., Bockheim, J. G., Kimble, J. M., Mazhitova, G. G., Ping, C. L., and Tarnocai, C.: A description and classification of soils and landscapes of the Lower Kolyma River, Northern Russia, Polar Geography and Geology, 19, 107–126, doi:10.1080/10889379509377563, 1995.

Soil Classification Working Group: The Canadian System of Soil Classification. Agriculture and Agri-Food Canada Publication 1646, NRC Research Press, National Research Council of Canada, Ottawa, Canada, 1998.

Soil Survey Staff: Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys, 1st Edn., Agric. Handb. 436. USDA-SCS, US Gov. Print. Office, Washington, DC, 1975.

Soil Survey Staff: Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys, 2nd Edn., Agric. Handb. 436. USDA-NRCS, US Gov. Print. Office, Washington, DC, 1999.

Stiles, C. A., Dunkinson, E. T., and Ping, C. L.: Initial field installation of manganese reduction indicators in soils (MRIS) in the Brooks Range of Alaska, Soil Survey Horizons, 51, 102–107, 2011.

Stolbovoi, V.: Carbon in Russian soils, Climatic Change, 55, 131–156, 2002.

Strauss, J., Schirrmeister, L., Grosse, G., Wetterich, S., Ulrich, M., Herzschuh, U., and Hubberten, H.-W.: The deep permafrost carbon pool of the Yedoma Region in Siberia and Alaska, Geophys. Res. Lett., 40, 6165–6170, doi:10.1002/2013gl058088, 2013.

Sutton, R. and Sposito, G.: Molecular structure in soil humic substances: The new view, Environ. Sci. Technol., 39, 9009–9015, 2005.

Swanson, D. K., Ping, C. L., and Michaelson, G. J.: Diapirism in soils due to thaw of ice-rich materials near the permafrost table, Permafrost Periglac., 10, 349–367, doi:10.1002/(SICI)1099-1530(199910/12)10:4<349::AID-PPP318>3.0.CO;2-N, 1999.

Tarnocai, C. and Smith, C. A. S.: The formation and properties of soils in the permafrost regions of Canada, in: Cryosols: The Effects of Cryogenesis on the Processes and Peculiarities of Soil Formation, Proceedings of the 1st International Conference on Cryopedology, Pushchino, Russia, 10–16 November 1992, edited by: Gilichinsky, D. A., Russ. Acad. of Sci., Pushchino, Russia, 21–42, 1992.

Tarnocai, C. and Zoltai, S. C.: Earth hummocks of the Canadian Arctic and Subarctic, Arctic Alpine Res., 11, 581–594, 1978.

Tarnocai, C., Walker, D. A., and Broll, G.: Turf hummocks in Arctic Canada: Characteristics and development, Eos Trans. AGU, 87, Fall Meet. Suppl., Abstract C51A-0381, 2006.

Tarnocai, C., Canadell, J. G., Schuur, E. A. G., Kuhry, P., Mazhitova, G., and Zimov, S. A.: Soil organic carbon pools in the northern circumpolar permafrost region, Global Biogeochem. Cy., 23, GB2023, doi:10.1029/2008GB003327, 2009.

Tedrow, J. C. F.: Soil of the Polar Regions, Rutgers Univ. Press, New Brunswick, N. J., 1974.

Treat, C. C., Wollheim, W. M., Varner, R. K., Grandy, A. S., Talbot, J., and Frolking, S.: Temperature and peat type control CO_2 and CH_4 production in Alaskan permafrost peats, Glob. Change Biol., 20, 2674–2686, 2014.

Turetsky, M. R., Wieder, R. K., Vitt, D. H., Evans, R. J., and Scott, K. D.: The disappearance of relict permafrost in boreal north America: Effects on peatland carbon storage and fluxes, Glob. Change Biol., 13, 1922–1934, 2007.

Uhlířová, E., Šantrůčková, H., and Davidov, S. P.: Quality and potential biodegradability of soil organic matter preserved in permafrost of Siberian tussock tundra, Soil Biol. Biochem., 39, 1978–1989, 2007.

van Cleve, K., Dyrness, C. T., Viereck, L., Fox, J., Chapin III, F. S., and Oechel, W.: Taiga ecosystem in interior Alaska, Bioscience, 33, 39–44, 1983.

von Lützow, M., Kogel-Knabner, I., Ekschmitt, K., Matzner, E., Guggenberger, G., Marschner, B., and Flessa, H.: Stabilization of organic matter in temperate soils: Mechanisms and their relevance under different soil conditions – A review, Eur. J. Soil Sci., 57, 426–445, 2006.

von Lützow, M., Kögel-Knabner, I., Ekschmitt, K., Flessa, H., Guggenberger, G., Matzner, E., and Marschner, B.: SOM fractionation methods: relevance to functional pools and to stabilization mechanisms, Soil Biol. Biochem., 39, 2183–2207, 2007.

Vonk, J. E., Sánchez-García, L., van Dongen, B. E., Alling, V., Kosmach, D., Charkin, A., Semiletov, I. P., Dudarev, O. V., Shakhova, N., Roos, P., Eglinton, T. I., Andersson, A., and Gustafsson, Ö.: Activation of old carbon by erosion of coastal and subsea permafrost in Arctic Siberia, Nature, 489, 137–140, 2012.

Waldrop, M. P., Wickland, K. P., White III, R., Berhe, A. A., Harden, J. W., and Romanovsky, V. E.: Molecular investigations into a globally important carbon pool: Permafrost-protected carbon in Alaskan soils, Glob. Change Biol., 16, 2543–2554, 2010.

Walker, D. A., Epstein, H. E., Gould, W. A., Kelley, A. M., Kade, A. N., Knudson, J. A., Krantz, W. B., Michaelson, G. J., Peterson, R. A., Ping, C. L., Raynolds, M. K., Romanovsky, V. E., and Shur, Y.: Frost-boil ecosystems: complex interactions between landforms, soils, vegetation, and climate, Permafrost Periglac., 15, 171–188, 2004.

Walker, D. A., Epstein, H. E., Romanovsky, V. E., Ping, C. L., Michaelson, G. J., Daanen, R. P., Shur, Y., Peterson, R. A., Krantz, W. B., Raynolds, M. K., Gould, W. A., Gonzalez, G., Nickolsky, D. J., Vonlanthen, C. M., Kade, A. N., Kuss, P., Kelley, A. M., Munger, C. A., Tarnocai, C., Matveyeva, N. V., and Daniëls, F. J. A.: Arctic patterned-ground ecosystems: a synthesis of field studies and models along a North American Arctic Transect, J. Geophys. Res. 113, G03S01, doi:10.1029/2007JG000504, 2008.

Walmsley, M. E. and Lavkulich, L. M.: Chemical, physical, and land-use investigations of organic terrain, Can. J. Soil Sci., 55, 331–342, 1975.

Walter, K. M., Zimov, S. A., Chanton, J. P., Verbyla, D., and Chapin III, F. S.: Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming, Nature, 443, 71–75, 2006.

Walter, K. M., Edwards, M. E., Grosse, G., Zimov, S. A., Chapin III, F. S.: Thermokarst lakes as a source of atmospheric CH_4 during the last deglaciation, Science, 318, 633–636, 2007.

Walter Anthony, K. M., Zimov, S. A., Grosse, G., Jones, M. C., Anthony, P. M., Chapin III, F. S., Finlay, J. C., Mack, M. C., Davydov, S., Frenzel, P., and Frolking, S.: A shift of thermokarst lakes from carbon sources to sinks during the Holocene epoch, Nature, 511, 452–456, doi:10.1038/nature13560, 2014.

Washburn, A. L.: Periglacial processes and environment, St. Martin's Press, New York, 1973.

Weintraub, M. N. and Schimel, J. P.: Interactions between carbon and nitrogen mineralization and soil organic matter chemistry in Arctic tundra soils, Ecosystems, 6, 129–143, 2003.

White, D. M., Garland, D. S., Dai, X.-Y., and Ping, C. L.: Fingerprinting soil organic matter in the Arctic to help predict CO_2 flux, J. Cold Reg. Sci. Technol., 35, 185–194, 2002.

White, D. M., Garland, D. S., Ping, C. L., and Michaelson, G. J.: Characterizing soil organic matter quality in arctic soil by cover type and depth, J. Cold Reg. Sci. Technol., 38, 63–73, 2004.

Wild, B., Schnecker, J., Eloy Alves, R. J., Barsukov, P., Bárta, J., Čapek, P., Gentsch, N., Gittel, A., Guggenberger, G., Lashchinskiy, N., Mikutta, R., Rusalimova, O., Šantrůčková, H., Shibistova, O., Urich, T., Watzka, M., Zrazhevskaya, G., and Richter, A.: Input of easily available organic C and N stimulates microbial decomposition of soil organic matter in arctic permafrost soil, Soil Biol. Biochem., 75, 143–151, 2014.

Winterfeld, M., Schirrmeister, L., Grigoriev, M. N., Kunitsky, V. V., Andreev, A., Murray, A., and Overduin, P. P.: Coastal permafrost landscape development since the Late Pleistocene in the western Laptev Sea, Siberia, Boreas, 40, 697–713, doi:10.1111/j.1502-3885.2011.00203.x, 2011.

Xu, C. H., Dou, F., Guo, L., and Ping, C. L.: Potential DOC production from size fractionated arctic tundra soils, Cold Reg. Sci. Technol., 55, 114–150, 2009a.

Xu, C., Guo, L., Ping, C. L., and White, D. M.: Chemical and isotopic characterization of size-fractionated organic matter from cryoturbated tundra soils, northern Alaska, J. Geophy. Res.-Biogeo., 114, G03002, doi:10.1029/2008JG000846, 2009b.

Zhuang, Q., Melillo, J. M., Sarofim, M. C., Kicklighter, D. W., McGuire, A. D., Felzer, B. S., Sokolov, A., Prinn, R. G., Steudler, P. A., and Hu, S.: CO_2 and CH_4 exchanges between land ecosystems and the atmosphere in northern high latitudes over the 21st century, Geophy. Res. Lett., 33, L17403, doi:10.1029/2006GL026972, 2006.

Zhuang, Q., Melillo, J. M., McGuire, A. D., Kicklighter, D. W., Prinn, R. G., Steudler, P. A., Felzer, B. S., and Hu, S.: Net emissions of CH_4 and CO_2 in Alaska: Implications for the region's greenhouse gas budget, Ecolog. Appl., 17, 203–212, 2007.

Zimov, S. A., Zimova, G. M., Daviodov, S. P., Daviodova, A. I., Voropaev, Y. V., Voropaeva, Z. V., Prosiannikov, S. F., Prosiannikova, O. V., Semiletova, I. V., and Semiletov, I. P.: Winter biotic activity and production of CO_2 in Siberian soils: a factor in the greenhouse effect, J. Geophy. Res.-Atmos., 98, 5017–5023, doi:10.1029/92JD02473, 1993.

Zimov, S. A., Davidov, S. P., Voropaev, Y. V., Prosiannikov, S. F., Semiletov, I. P., Chapin, M. C., and Chapin, F. S.: Siberian CO_2 efflux in winter as a CO_2 source and cause of seasonality in atmospheric CO_2, Climatic Change, 33, 111–120, 1996.

Zimov, S. A., Schuur, E. A. G., and Chapin, F. S.: Permafrost and the global carbon budget, Science, 312, 1612–1613, doi:10.1126/science.1128908, 2006.

Zoltai, S. C. and Tarnocai, C.: Perennially frozen peatlands in the Western Arctic and Subarctic of Canada, Can. J. Earth Sci., 12, 28–43, 1975.

Zubrzycki, S., Kutzbach, L., Grosse, G., Desyatkin, A., and Pfeiffer, E.-M.: Organic carbon and total nitrogen stocks in soils of the Lena River Delta, Biogeosciences, 10, 3507–3524, doi:10.5194/bg-10-3507-2013, 2013.

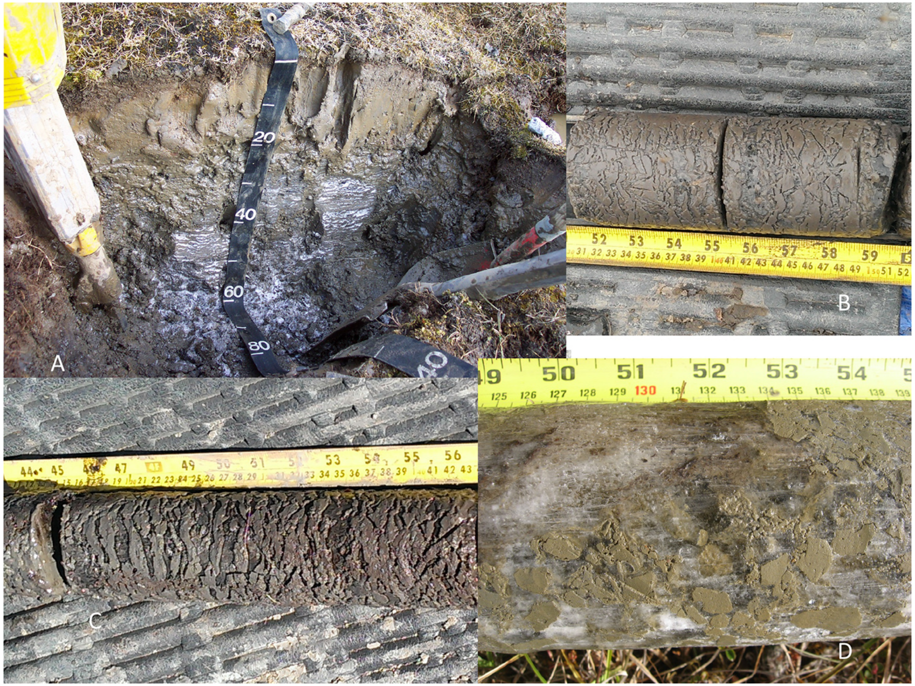

Figure 1. Examples of common cryostructures: **(a)** lenticular structures exposed along a freshly exposed profile from an earth hummock, Mould Bay, arctic Canada, **(b)** lenticular structures transition to reticulate structure at 130–150 cm showing in a core extract from the center of a low centered ice wedge polygon, **(c)** lenticular structure, and **(d)** suspended structure (ataxitic).

SOILD
1, 709–756, 2014

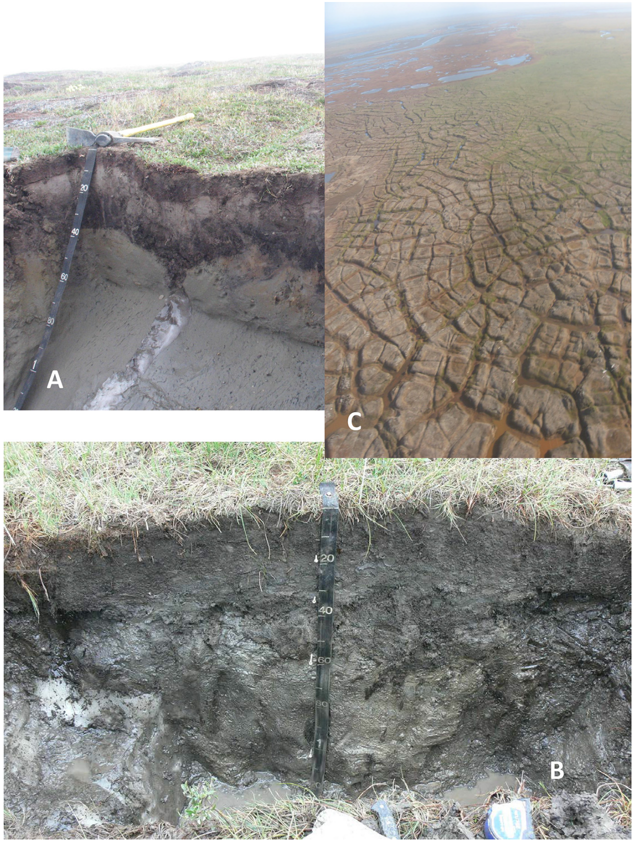

Figure 2. Soils formed in ice-wedge polygons, **(a)** active layer deformed by newly formed ice wedge (7 cm wide), **(b)** cryoturbated polygon rim of a flat-centered ice-wedge polygon, (ice wedge 85 cm wide), and **(c)** an aerial view of the ice-wedge polygon-dominated landscape of the Arctic Coastal Plain, Alaska.

SOILD
1, 709–756, 2014

Figure 3. The cryoturbated soil profiles of the thawed active layers of an earth hummock (top) and a nonsorted circle (bottom). Note the distribution of organic matter in the low micro-topographic positions and the insulating effect of the organic horizon on thaw depths.

SOILD

1, 709–756, 2014

Permafrost soils and carbon cycling

C. L. Ping et al.

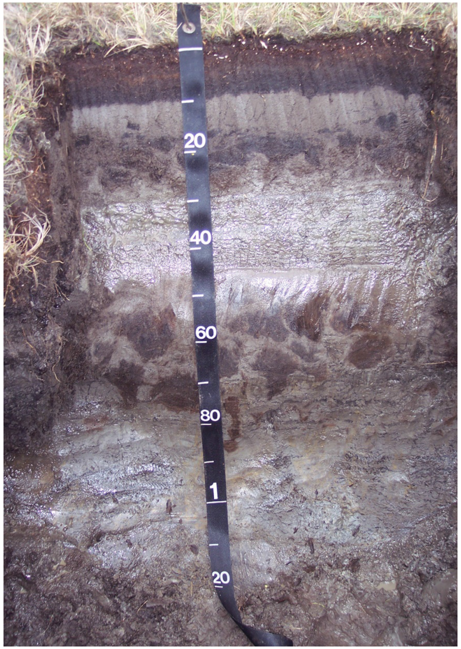

Figure 4. Gelisol profile from the Arctic Coastal Plain of Northwest Alaska exhibiting a complex stratigraphy developed through a sequence of different environments coupled with cryoturbation caused by freeze/thaw/heave forces. The strata suggest: (1) old coastal plain sediments below 80 cm thawed and collapsed into a thaw lake as indicated by the mixed up sediments with scattered fragmental peat chunks; (2) after drainage an organic mat developed at 50 to 80 cm; (3) this mat was flooded or collapsed within the lake basin and was covered by limnic silts at 30–50 cm; (4) after a second drainage another organic mat covered the limnic silts at 15 to 30 cm; (5) windblown eolian silt covered this mat at 10–16 cm; and (6) a final organic mat developed at the surface. The fluctuating and rising permafrost table disrupted the underlying organic mats.

SOILD
1, 709–756, 2014