Carbon Stores and Biogeochemical Properties of Soils under Black Spruce Forest, Alaska

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Fifty-two soils under black spruce [*Picea mariana* (Mill.) Britton et al.]-dominated forest communities were examined and assessed for their organic C (OC) stores in relation to soil characteristics. Study sites were located on a variety of parent materials, landscape positions, and drainage conditions. Results indicate that soils at most sites were weakly developed, commonly with organic (O) horizons ranging from 3 to 39 cm (≥100 cm occasionally). Organic C stores tended to increase as drainage changed from somewhat excessive and well to very poorly drained (average to 1 m: 12.6–50.9 kg OC m^{-2}, respectively). The lowest OC store for an individual site was 7.1 kg OC m^{-2} in a well-drained soil on an outwash plain and the highest was 109 kg C m^{-2} in a very poorly drained soil. Surface organic horizons contained 13 to 100% of the total pedon OC stores. In Gelisols, permafrost sequestered an average of 9, 19, and 39% of SOC stores for the somewhat poorly, poorly, and very poorly drained soils, respectively. The presence of permafrost in poorly drained sites increased average OC stores from 27.8 to 50.1 kg OC m^{-2} over those without permafrost. Soil bulk density, cation exchange capacity, and extractable acidity assessed in relation to OC stores of genetic horizons illustrate the significant impact of OC on soil properties. In previous ecological studies in Alaska, OC was determined for only the surface horizons; our data suggest that such shallow sampling may underestimate total OC stores by an average of 26% and up to 68%.

Abbreviations: CEC, cation exchange capacity; OC, organic carbon; OM, organic matter; SOC, soil organic carbon; SOM, soil organic matter.

Boreal forest covers more than 22% of the terrestrial surface (Chapin et al., 2000) and accounts for 30% of the world's total terrestrial soil organic C (SOC) (Oechel and Vourlitis, 1995; Post et al., 1982). It covers 60 to 70% of the land area in Alaska (Van Cleve et al., 1983), of which 44% is dominated by black spruce (Damman and Johnston, 1980; Viereck et al., 1986), the most abundant tree species in Alaska (Zasada and Packee, 1995). The total area of land coverage of the black spruce forest in Alaska is estimated to be 40 million ha. Black spruce is one of the six major tree species found in interior Alaska (Viereck and Little, 2007); this lack of competitors (Chapin et al., 2006) and its shade tolerance (Viereck and Johnston, 1990) allow black spruce to occupy a wide range of sites. Although long associated with poorly drained lowland areas and north-facing aspects (Van Cleve et al., 1991), more recent soil survey reports (Brannen and Swanson, 2001; Clark and Kautz, 1999; Mulligan, 2004, 2007; Pink, 2008) and studies (Rosner, 2004; Hollingsworth et al., 2006) document that black spruce occurs with a wide range of understory species on sites across a broad range of landforms, aspects, and soil wetness and acidity. In Alaska, black spruce grows on soils of the Andisol, Entisol, Gelisol, Histosol, Inceptisol, and Spodosol orders and udic and aquic moisture regimes. It can grow where the rooting zone is limited to just a few decimeters by permafrost, a high water table, acidic mineral soils, or bedrock (Viereck and Johnston, 1990) as well as on well-drained, deep soils. It is commonly the climax species in Alaska. Because of its semi-serotinous cones, it is highly adapted to reproduce by seed after fire (Viereck and Johnston, 1990). In Alaska, black spruce dominates

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wherever soil conditions (e.g., Histisols and Gelisols) limit competitors' ability to grow or fire eliminated or reduced competing tree species and provided an adequate seedbed. Although many black spruce forests in Alaska have low productivity (Viereck, 1983), partly due to black spruce's small stature and its ability to thrive in dense stands of stunted trees on poor sites, well-stocked black spruce forests grown on moderately to well drained upland and bottomland sites are quite productive (Rosner, 2004).

Since the arctic and boreal regions are expected to sustain the greatest impact in terms of increasing ambient temperature and trace gas release due to global climate change (Intergovernmental Panel on Climate Change, 1992; Arctic Climate Impact Assessment Team, 2004), more complete measurements for the estimation of SOC are crucial for predicting future climate effects. Because of the vast global area and the potentially high OC storage capacity of these North American boreal forest communities, a need exists to gain a better understanding of C stores and related biogeochemical properties of black spruce forest soils, which are crucial to both ecological modeling and land management. Furthermore, inventory and turnover rates of this belowground OC determines the present and future roles of boreal forest soils as sources or sinks of C (Rapalee et al., 1998).

In general, there is a lack of studies to investigate deeper SOC stores. Although biased to backslopes, Kane and Vogel (2009) and Hollingsworth et al. (2008) investigated SOC along several toposequences across interior Alaska. Few other studies have included an examination of soils to the depths and across the landforms that inevitably will be affected, in the long term, by fire and drainage changes brought about by thawing permafrost (Ping et al., 2002). Investigations of Canadian boreal forest soils have indicated that only about 30% of SOC stores are in the surface soils (Tarnocai, 2000). Limited studies in Alaska also indicate there are significant SOC stores in subsoils (Ping et al., 2005b, 1998). Nearly 75% of the OC in the boreal forest is estimated to be stored belowground (Apps et al., 1993). Recent studies have found significant relationships between surface SOC stores and soil temperature and plant community assemblages for upland black spruce forests of Alaska (Kane and Vogel, 2009; Hollingsworth et al., 2008). But quantification of deeper stores, particularly in poorly drained sites on lower and more north-facing slopes, has not been widely addressed (Ping et al., 2005b). This is an important data gap to consider, as it is these cooler and more poorly drained ecosystems that are likely to be the most strongly affected by the alteration and recurrence of the permafrost table due to changes in soil temperature and drainage conditions (Schuur et al., 2009).

The boreal forest region of Alaska, with its discontinuous and "warm" permafrost conditions, has and will probably continue to experience rapid climate warming and degradation of permafrost (Osterkamp and Romanovsky, 1999), resulting in land subsidence (thermokarst) and changes in landscape drainage conditions. Such changes in soil temperature and moisture conditions will affect the SOC stocks in both surface and subsurface soils. Studies of SOC occurring deeper than 30 cm in the soil profile (Ping et al., 1997) indicated that increased fire and subsequent thawing of discontinuous permafrost will contribute to the exposure of a significant portion of the sequestered SOC stores (Jorgenson et al., 2001) to oxidation and ultimately a loss from the terrestrial system to the atmosphere (Schuur et al., 2008, 2009). Although SOC stocks have been studied in the arctic regions (north of the boreal forest) of Alaska (Ping et al., 2008, 1998; Michaelson et al., 1996), the spatial extent and amounts of these deeper stores remain unknown across the boreal forest landscape of Alaska.

The objectives of this study were: (i) to better define the range of soils and SOC stores (particularly at depths >20 cm) that occur in association with black spruce communities across the boreal forest of the interior Alaskan landscape; (ii) to develop a basic understanding of how SOC stocks accumulate and their geospatial distribution across interior Alaska; and (iii) to determine how SOC buildup affects soil and soil organic matter (SOM) properties.

MATERIALS AND METHODS

Physiographic Setting

The study sample sites (Table 1) are scattered throughout seven Major Land Resource Areas of Alaska: the Interior Alaska Highlands, Interior Alaska Lowlands, the Interior Brooks Range Mountains, Interior Alaska Mountains, the Yukon–Kuskokwim Highlands, Cook Inlet Lowlands, and Copper River Basin. Narrative descriptions of these areas can be found in NRCS (2006). Study sites covered an east–west span from 141° W in the Tok–Northway area, near the Alaska–Canada border, to 160° W around Aniak in the Yukon–Kuskokwim Highlands, and a north–south span from 67° N in the southern foothills of the Brooks Range to 61° N in the Cook Inlet Lowland (Fig. 1). A detailed description of the physiographic environment of the sampling sites, soil-forming factors, and soil morphological properties were summarized by Ping et al. (2010).

Site Selection and Field Sampling Protocol

Fifty-two forest sites dominated by black spruce were selected to represent a range of physiographic areas and the landforms found in the Alaskan boreal forest. The locations, which are components of different research projects selected for this study, are shown in Fig. 1 and the physical environments of the study sites are summarized in Table 1.

Soil Characterization

At each site, physiographic and environmental characteristics including drainage classes and soil profile descriptions were recorded according to the NRCS soil survey procedures (Schoeneberger et al., 2002). In a representative location at each site, a soil pit was excavated to a depth of >1 m or to lithic or paralithic contact. Soil morphological characteristics were recorded and samples were taken from each genetic horizon for full characterization at either the National Soil Survey Center (Lincoln, NE) or the University of Alaska Palmer Research Center Plant and Soil Analysis Laboratory. Both laboratories followed standard soil survey laboratory procedures (Soil Survey Laboratory

Table 1. Physical environment and total organic C (TOC) of soil study sites associated with black spruce arranged by drainage classes for the sites.

Site identification†	Latitude north	Longitude west	Elevation	LF‡	LP§	PM¶	Slope	Azimuth	Depth to PF#	O layer thickness	Total organic C 1 m	Total organic C O layer	PF
	—— ° ——		m				%	°	——— cm ———		kg m^{-2}	—— % ——	
Somewhat excessively drained													
1-02N0082	63.285	142.513	548	SD	SS	L/S	2	90	NP	10	12.0	34	0
2-05N0092	67.334	150.145	345	AP	MS	Al	0	–	NP	16	15.4	72	0
Well drained													
3-02N1035	63.352	142.983	486	OP	F	Al	1	10	NP	6	7.1	46	0
4-04N0265	63.056	141.827	713	H	BS	T/R	14	199	NP	6	8.0	21	0
5-04N0261	65.094	147.851	490	H	BS	L/R	29	168	NP	14	8.7	56	0
6-UAFMarion	67.325	150.137	351	OT	F	Al/O	0	–	NP	3	9.1	16	0
7-04N0280	64.77	148.265	396	H	SS	L	29	3	NP	14	9.2	42	0
8-S04AK176003	63.061	141.026	740	H	SS	T/R	7	52	NP	11	9.9	34	0
9-04N0264	63.604	141.830	711	H	BS	T/R	14	270	NP	13	10.3	33	0
10-89AK170005	61.417	150.083	67	OP	S	TL	1	180	NP	8	11.2	26	0
11-UCM0742	63.993	148.723	720	M	BS	C	28	197	NP	12	11.9	52	0
12-UCM0731	63.998	148.743	623	AF	MS	Al	8	145	NP	14	12.5	49	0
13-05N0091	68.069	149.580	404	Mo	BS	GO	6	225	NP	10	13.5	53	0
14-02N0081	63.061	141.827	773	H	BS	T/C	6	90	NP	10	16.7	70	0
15-91AK240003	64.019	145.124	383	OP	F	Al	1	20	NP	7	17.3	13	0
16-40A5508	59.983	154.850	107	H	TS	T	15	135	NP	10	29.0	44	0
Moderately well drained													
17–04N0281	64.804	148.331	240	H	BS	L	4	105	NP	16	10.1	43	0
18-UCM0756	63.997	148.737	636	H	T	C	37	154	NP	8	16.1	47	0
19-03AK290006	67.323	150.148	351	OT	F	Al/O	0	–	NP	10	16.8	60	0
20-04N0277	65.175	147.363	741	H	SS	R	6	128	NP	17	18.3	52	0
21-CP09S1–8	67.500	149.833	532	AT	F	Al	1	220	NP	19	36.7	76	0
Somewhat poorly drained													
22–05N0099	63.463	143.468	869	H	RT	R	0	–	NP	21	18.1	67	0
23-UAF0205	65.195	147.497	760	H	BS	L/C	14	345	40	19	20.5	95	<1
24-04N0260	65.103	147.882	445	H	SD	L	9	0	70	22	21.7	33	17
25-01N1175	61.527	159.778	27	H	T	L	2	270	NP	8	26.2	38	0
26-UCM0744	63.994	148.722	722	H	SD	R	10	223	NP	21	31.2	51	0
27-UCM0733	63.966	148.742	638	AT	F	Al	1	194	NP	13	23.1	43	0
Poorly drained													
28–91P0975	61.425	150.137	34	OP	F	L/O	1	180	NP	7	36.0	27	0
29–05N0338	61.347	160.588	37	H	RT	L	2	92	77	22	41.7	58	5
30–93P0617	67.200	150.270	137	Mo	BS	GO	2	136	51	39	65.0	74	18
31-UAF0204	65.175	147.523	394	H	FS	C	55	15	55	30	15.0	64	<1
32–81P0688	62.485	142.485	427	Bn	F	Lu	1	45	54	23	20.7	75	<1
33–40A0650	64.855	147.863	150	AP	F	Al	0	–	NP	13	26.9	56	0
34-S04AK176008	63.156	143.215	592	OP	F	Al/O	0	–	49	34	31.4	67	18
35-UAF0203	65.176	147.523	383	H	FS	C	50	0	52	16	33.4	16	21
36–91P0969	64.851	147.834	134	AP	F	Al	1	180	84	25	36.6	82	2
37–04N0968	64.806	148.418	127	AP	F	Al	0	–	76	6	36.2	26	18
38–00AK068003	63.683	151.517	260	AF	F	Al	1	315	75	21	41.3	62	<1
39–04N0971	64.815	148.043	197	AF	TS	Al	2	190	42	15	42.8	53	22
40–04N0279	64.867	147.875	161	H	TS	L	6	214	70	37	43.3	53	15
41–04N0966	64.443	149.141	130	AP	F	L/Al	0	–	71	27	50.4	75	2
42–04N0964	64.214	148.418	151	AP	F	L/Al	0	–	60	22	51.3	74	8
43–96P0365	64.869	147.857	169	H	BS	L	4	135	58	27	51.8	83	4
44-S04AK176002	64.192	145.868	303	AF	F	Al	0	–	44	16	70.3	26	38
45-01P0012	63.677	151.150	398	H	BS	L	7	270	36	24	78.4	54	39
46-96P0367	64.865	147.853	154	H	TS	O/L	4	0	37	16	98.1	72	26
Very poorly drained													
47-UAF0201	65.152	147.487	217	AF	MS	Al	8	190	55	23	22.8	16	44
48-91AK240002	63.907	145.124	369	OP	F	L/Al	1	20	51	14	29.7	43	33
49-91P0971	64.858	147.882	145	AP	F	Al	1	180	38	24	32.0	69	19
50-92AK068003	63.727	148.841	744	H	FS	L/Tl	13	6	47	27	60.8	70	24
51-91P0970	64.867	147.855	152	H	TS	L	4	270	30	25	109.3	44	53
52-91P0976	61.426	150.139	32	Bn	F	Or	0	–	NP	>100	109.4	100	0

† Site number and identification number from data of Ping et al. (2010).

‡ Landform: AF, alluvial fan; AP, alluvial plain; AT, alluvial terrace; Bn, basin; FS, footslope; H, hill; M, montaine; Mo, moraine; OP, outwash plain; SD, sand dune.

§ Landscape position: BS, backslope; F, flat; MS, midslope; RT, ridgetop; S, slope; SD, saddle; SS, shoulder slope; T, terrace; TS, toeslope.

¶ Parent material: Al, alluvium; Al/O, alluvium over outwash; C, colluvium; GO, glacial outwash; L, loess; L/Al, loess over alluvium; L/C, loess over colluvium; L/O, loess over outwash; L/R, loess over residuum; L/S, loess over sand; L/Tl, loess over till; Lu, lacustrine; Or, organics; O/L, organics over loess; R, residuum; T, tephra; T/C, tephra over colluvium; TL, tephra loess mix; T/R, tephra over residuum.

PF = permafrost; NP = no permafrost present or at >1-m depth.

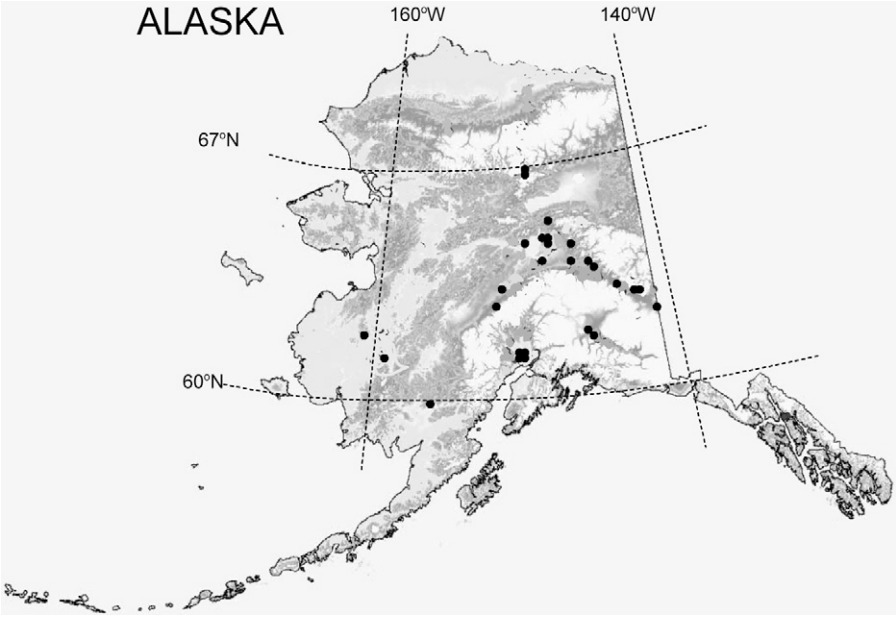

Fig. 1. Locations of sampling sites in Alaska.

Staff, 1996). Detailed soil morphological, physical, and chemical properties for all sites were provided in Ping et al. (2010).

The soil samples were air dried and passed through a 2-mm sieve. Fractions >2 mm were weighed to determine the gravel content. A subsample of the <2-mm fraction was oven dried to 110°C and used to correct the analyses to an oven-dried-weight basis. Subsamples of the air-dried ≤2-mm fraction were ball-milled to 0.425 mm (80 mesh) for total OC and N analyses by the high-temperature ignition method. When only oxidizable OC analyses were available, total OC and N values were corrected using relationships developed between ignition total OC and oxidizable OC by using regional soils data available in the NRCS database. The soil organic matter content was determined as the loss of weight of a sample dried at 110°C then ignited at 450°C. Bulk density (D_b) was determined for unfrozen mineral horizons using either the saran-coated clod method (Soil Survey Laboratory Staff, 1996) or the volumetric soil core method on samples dried at 110°C and organic or frozen horizons. A measured-volume block was cut from the clods with a serrated knife and then oven dried at 110°C (Michaelson et al., 2001). Bulk density was estimated for missing data samples using relationships developed for bulk density as a function of SOC (g kg^{-1}) for groups of similar soils according to Ping et al. (1997). The particle size distribution was determined for the <2-mm fraction of soils using the pipette method (Soil Survey Laboratory Staff, 1996).

Soil pH was measured in distilled water at a 1:1 ratio for mineral soils and in a saturated paste mixture for organic soils. Ground soil subsamples with pH values ≥6.5 were pretreated with 1 mol L^{-1} HCl before total OC and N analysis by dry combustion. The cation exchange capacity (CEC) was determined by the 1 mol L^{-1} NH_4OAc (pH 7) method. Extractable acidity was determined using the $BaCl_2$–triethanolamine (pH 8.3) extraction and acid titration method.

Pedon OC and total N (TN) storage to a depth of 1 m were calculated using the equation of Michaelson et al. (1996): horizon OC storage (kg m^{-2}) = $T\,D_b$ (g OC kg^{-1} × 10^{-2}) [1 − (%CF_v × 10^{-2})], where T is horizon thickness (cm), D_b is bulk density (g cm^{-3}), and CF_v is the volumetric content of coarse soil fragments >2 mm in diameter. Horizon stores were summed to a depth of 100 cm. In cryoturbated soils, where soil horizons are warped or discontinuous, profile diagrams drawn to scale were used to estimate the proportions of each horizon (Kimble et al., 1993; Ping et al., 1997). Soil profile descriptions and classification following Soil Taxonomy (Soil Survey Staff, 2006) for each study sites are available in Ping et al. (2010).

RESULTS AND DISCUSSION

Soil Physical Environment

Black spruce is well known as a species commonly associated with cold and wet environments. Based on our study of boreal forest soils in Alaska, however, forest stands dominated by black spruce occur in a variety of lowland settings, including alluvial fans, terraces, outwash plains, and a broad range of upland slope positions on hills and low mountains. Soil drainage conditions range from somewhat excessively to very poorly drained (Table 1). Sampled sites occupy a wide range of landscape positions from 0 to 55% slope, throughout the range of aspects, with and without permafrost, and elevations between 25 and 870 m above sea level. As suggested above, the wide range of sites occupied by black spruce is due to a combination of factors working separately or in consort: the limited number of competing tree species in interior Alaska, black spruce's ability to occupy a broad range of sites, its strong adaptation to fire (semi-serotinous cones), and greater shade tolerance than associates.

Black spruce occurs on excessively and well drained soils (Table 1, Sites 1–16) formed in coarser textured alluvium (generally <50 cm of loamy material over sand and gravel), sand dunes, moraines, and bedrock uplands. It is particularly abundant in upland sites with a sparse vascular plant understory due to either a poor site (Site 5) or intense shading at a later seral stage (Site 4) and often mixed with white spruce [*Picea glauca* (Moench) Voss] or seral hardwoods. On bedrock uplands, poor drainage is in concavities, north-facing slopes (Table 1, Sites 28, 31, and 36–39), and deep loess covers (Sites 40–43 and 45). Some of the well drained sites are on convexities with a gravelly residuum (Sites 4, 5, 8, and 9) near the surface (Swanson, 1996a). In alluvial lowlands, poorly drained soils with black spruce on permafrost are typically present when there is >50 cm of loamy material insulated with thick organic layers that favor the preservation of permafrost (Sites 36–39, 41, 42, 44, 47, and 49). Permafrost, high water tables, and deep organic layers form a mutually reinforcing system that is vegetated by black spruce due to the tree's unique tolerance of acidity and cold and its ability to root in a very shallow surface layer. For the better drained sites on outwash plains or river terraces, the loamy surface material may be silty loess or alluvium (Sites 2, 3, 6, 10, and 15). In reviewing the site distribution across the boreal forest region of Alaska, it

is apparent that poorly drained lowlands are in most need of future sampling.

Because wildfire frequents the Alaskan boreal forest, especially black spruce dominated ecosystems, a dynamic relationship exists among soil drainage, the soil surface organic layer, thickness, and seasonal freeze–thaw dynamics, and thus the presence or absence of permafrost (Viereck et al., 1983; Swanson., 1996b). These dynamics can result in soil property changes and thus soil classification changes (Ping, 1987). Such relationships were observed in sites in the Tok area of the Upper Tanana Basin in eastern Alaska (Sites 3, 4, and 34).

Soil Organic Carbon

Soil Organic Carbon Stocks and Distribution Patterns

Because black spruce occurs across such a wide range of site conditions, SOC amounts and distribution patterns are highly variable. The general pattern for OC content in a forest soil profile is the result of soil formation. Soil organic matter accumulates and remains on the surface of the mineral soil. Thus the high concentration of OC occurs in the surface organic horizons (usually about 300–500 g OC kg^{-1}), with an abrupt decrease in the uppermost mineral horizon (20–120 g OC kg^{-1}), and a continued decrease to 10 g OC kg^{-1} or less in the lower mineral horizons. The pattern of OC stores on a spatial basis (kg OC m^{-2}) within the organic layer, however, often follows a sometimes subtle but importantly different pattern (Fig. 2). The lower part of the surface organic horizons generally contains a larger amount of OC (Fig. 2a–2d), due in part to the more highly decomposed SOM products that are less vulnerable to oxidation. The SOC stores decrease both downward into the mineral soil and upward into the more fibrous upper surface organic horizons. This inverse pattern of OC storage (in relation to weight-based OC concentration) is largely due to the increased bulk density with depth within the organic horizons (Fig. 2a–2d). In the boreal forest, this could be an important characteristic in that the denser, lower organic horizons can hold more nutrients, C, and water. This affects tree rooting; in some cases, the increased water retention could retard fire and the C losses associated with it (Kasischke and Johnstone, 2005; Kane et al., 2007). The wettest or very poorly drained soils often show a reverse of this storage pattern (Fig. 2e). This is probably due to the shift in rooting pattern associated with saturation to the surface, which contributes to a denser O horizon near the surface. The decrease in both OC concentration and OC stores with depth in the mineral horizons is evident in the better drained soils (Fig. 2a and 2b).

Soils with buried organic-enriched horizons and the more poorly drained soils with permafrost show increases in both OC concentration and OC stores at depth (Fig. 2c and 2e, respectively). These phenomena are also present in arctic soils and are affected by cryoturbation that frost churns the surface organic matter into lower horizons (Ping et al., 1998, 2008; Michaelson et al., 1996). These cryoturbated horizons present at depth are often combination (B/Ajj) horizons with intermediate bulk densities and OC concentrations and usually contain significant SOC stores, as shown in Fig. 2e, where 20% of the profile OC stocks are contained in the cryoturbated horizon at the 47- to 100-cm depth. On the other hand, soils affected by permafrost that lack pronounced cryoturbation build up thick surface organic horizons (Fig. 2d). The differences in SOC distribution in permafrost soils (Gelisols) are related to the presence or absence of frost churning, a characteristic reflected in the soil classification designations of Turbels vs. Orthels, respectively (Fig. 2e vs. 2d) and were discussed in detail by Ping et al. (2010).

Soils formed in fluvial or multiple loess or volcanic ash deposits also have buried organics in multiple buried horizons, and thus they too can show an uneven pattern of SOC throughout the deeper horizons (Ping et al., 1997), and the total pedon OC store can be at least as high as 30 kg OC m^{-2} in well drained soils (Ping et al., 2008). This pattern of increase in SOC at depth is noticeable in the somewhat poorly drained profile (Fig. 2c), which was located on a reworked loess deposit that experienced multiple depositional events. As the Fig. 2 examples illustrate, significant SOC stores are often found at depth in soils with and without permafrost and for well to very poorly drained soils for various reasons. This makes assessment to depth important when quantifying SOC pools in boreal forest soils.

Average stores of SOC and their distribution to a depth of 1 m for the range of drainage classes are summarized in Fig. 3. Although variability within a drainage class is fairly high due to

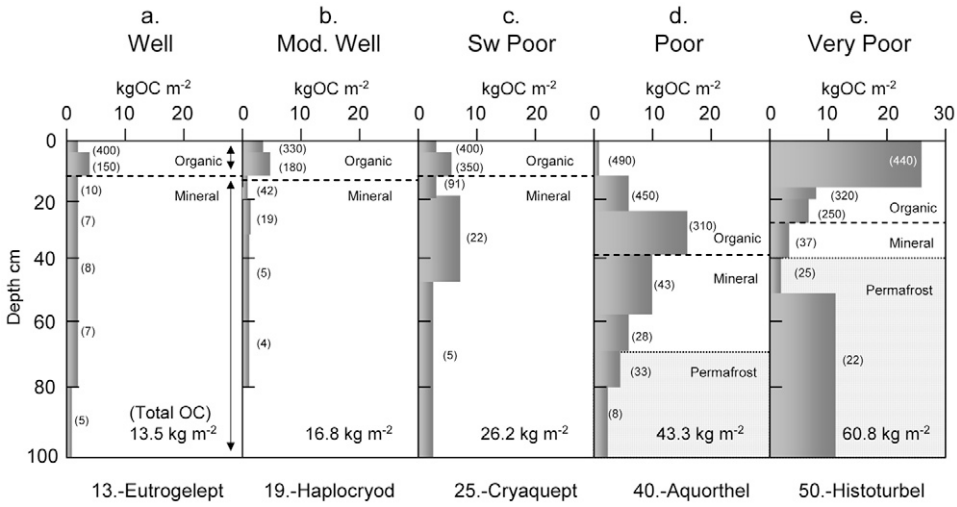

Fig. 2. Soil organic C (OC) storage distribution with depth in representative profiles of the well, moderately well, somewhat poorly, poorly, and very poorly drained drainage classes. Bars represent OC stores for genetic horizons, with total stores for the 1-m profile listed at the bottom for each. In parentheses are weight-based OC contents (g kg^{-1}) for each horizon for comparison with OC stores. Soil site number and USDA soil classifications are provided for each profile.

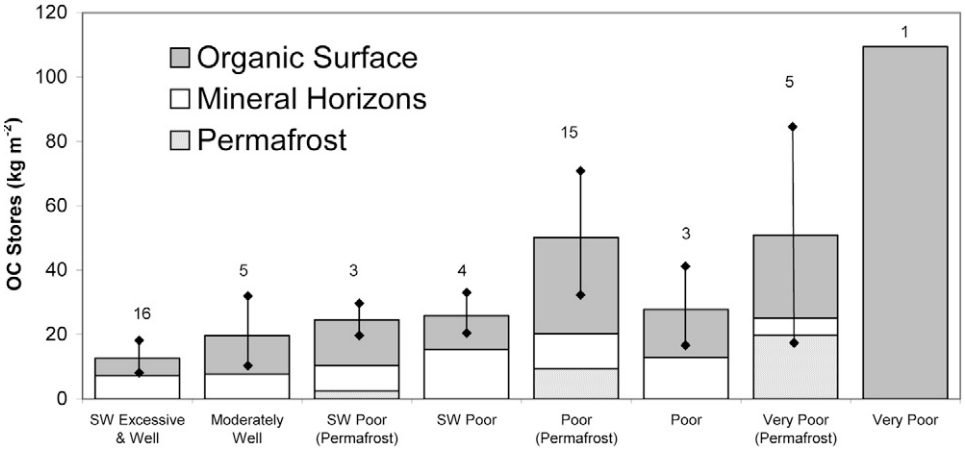

Fig. 3. Average soil organic C (OC) stores and distribution (to 1-m depth) for the somewhat excessively, well, moderately well, somewhat poorly, poorly, and very poorly drained soil drainage classes; bars are one standard deviation above and below the bar for total stores, and *n* values for the averages are given above each bar.

the diverse range of site conditions and locations across the state, a clear trend exists showing an increase in SOC stores with poorer drainage conditions. The presence of permafrost also tends to increase SOC stocks, especially within the poorly drained soils (Table 1, 0–38% of OC stores) and less so within the somewhat poorly drained soils, among which only two sites have permafrost that contains <1 and 17% of their respective profile OC stores. The very poorly drained, permafrost-free site in south-central Alaska (last bar in Fig. 3) has considerably higher SOC stocks than the average for very poorly drained permafrost soils from interior Alaska (Table 1). The range of SOC stores for soils in the very poorly drained condition, however, is high and one very poorly drained permafrost site contained the same high amount of SOC (109 kg OC m^{-2}) as the warmer non-permafrost site (Site 51, Table 1). This very poorly drained permafrost soil was on a gently sloping, north-facing site and had 53% of its 109 kg OC m^{-2} within the permafrost.

In Gelisols, permafrost contains, on average, 9, 19, and 39% of SOC stores for the somewhat poorly, poorly, and very poorly drained soils, respectively. These fractions are likely to be substantial when assessing the potential for change in SOC pools, especially in terms of added vulnerability with the thawing of permafrost. This range in permafrost SOC stores is only somewhat less than the 15 to 47% averages found for arctic Gelisols by Ping et al. (2008). Except for the very poorly drained, non-permafrost site, organic surface layers, on average across the drainage classes, contain from 40 to 60% of the SOC stocks, while conversely, the subsurface mineral layers also contained 40 to 60% of stores. Surface organic horizons in arctic soils contain an average of only 18 to 27% of SOC stocks (Ping et al., 2008), which probably reflects the higher storage of SOC at depth due to the greater prevalence of frost churning in the arctic compared with the boreal forest soils.

Carbon Stores at Different Sampling Depths

The most apparent effect of the buildup of SOC as an organic mat on the soil surface is its insulating and surface-protective qualities (Sharratt, 1997). These qualities have been studied for upland black spruce forest stands on gentle slope sequences across interior Alaska (Kane and Vogel, 2009; Hollingsworth et al., 2008). These studies revealed strong relationships among the thickness of surface organic layers, the temperature, forest understory plant community characteristics, and the amounts of surface SOC storage (the organic layer plus the upper 15 cm of the mineral soil). Moreover, the strong control of site drainage on SOC storage in surface organic horizons in boreal black spruce forests in Canada has been demonstrated by Yi et al. (2009a). Compared with these previous studies, we examined soils and SOC stores under black spruce both to a greater depth and across a larger range of slope and drainage conditions. The importance of measuring OC stores to depth is reinforced by comparing our surface OC measurements with those for the full 1-m depth.

Grouping the SOC stores in only the O horizons for all well drained sites (somewhat excessive, well, and moderately well drained classes) resulted in a large variance for predicting SOC stores to 1 m ($R^2 = 0.43$, $P = 0.001$); adding the SOC stores in the first 15 cm of mineral soil to that of the O horizons reduced the variance ($R^2 = 0.86$, $P = 0.001$). These two measurements of SOC stores represent incremental increases in sampling depth and contain, on average, 60%and 78%, respectively, of the 1-m SOC stores. For poorly drained sites (somewhat poorly, poorly, and very poorly drained classes), the variances were similar ($R^2 = 0.53$, $P = 0.001$ and $R^2 = 0.78$, $P = 0.001$, respectively) for predicting the 1-m OC stores. On average, O horizons and O horizons plus the upper 15 cm of the mineral soil contained 71 and 81% of OC stores measured to 1 m in depth in poorly drained sites.

The proportion of SOC stores to a depth of 1 m contained in the surface varied widely across sites, from 13 to 100% (Table 1), as discussed above and illustrated in the examples in Fig. 2. Our results, again, illustrate the need for soil sampling to 1 m or more for ecological or climate warming impact studies to avoid missing the "deeper" C and other nutrients.

Properties of Soil Organic Matter

The SOM was measured by ignition weight loss for 80 mineral horizons from 16 of the sites across interior Alaska. A good relationship was found for concentrations of OC as a function of SOM: % OC = 0.493(% organic matter), ($R^2 = 0.952$, $P < 0.001$, $n = 80$; data given in Ping et al., 2010; Zaman, 2008). The slope coefficient for this function indicates 493 ± 10 g OC kg^{-1} SOM (95% confidence), which is lower than the average 580 g OC kg^{-1} commonly used for temperate-region soils (Jackson, 1958) and

soils of boreal Canada (Siltanen et al., 1997). This is consistent with the SOM of soils in colder regions that have undergone less extensive decomposition and humification, which increase the OC content of SOM. The lower OC content must be considered when calculating OC stores for soils from Alaska using loss-on-ignition data if it is the only analysis available.

The C/N ratio has been used as a general indicator of the degree of SOM decomposition, and thus average overall SOM quality (Stevenson, 1994). Recently, however, studies have found that SOM is actually a mixture of pools, each with different qualities and C/N ratios due to differences in their physical as well as chemical structure (Sollins et al., 2006; Pineiro et al., 2006), which adds to the complexity of interpreting C/N ratios of bulk soils. With this in mind, it can be stated in only very general terms that surface-accumulated SOM, on average, is largely comprised of less decomposed plant material having higher C/N ratios compared with SOM found in the deeper mineral soil horizons. It is likely that deeper soil horizons contain larger proportions of modified and illuvial SOM and larger amounts of SOM physically protected by association with soil minerals compared with the surface horizons. The average C/N values for soil horizons in this study, presented in Table 2, reflect this general trend of decreasing C/N with depth and from organic to mineral horizons. The cryoturbated horizons (B/Ajj) found in soils with poorly drained permafrost conditions are at intermediate depths in the soil profile (example Fig. 2d) and tend to have higher C/N ratios (average 23), which are similar to the upper mineral A horizons. The C/N ratios decrease moving downward in the soil profile to the mineral B, BC, and C horizons (average 11–17). This pattern is consistent with what could be expected when fresh organic matter (OM) is continuously deposited on the surface to the Oe horizon and where decomposition products generated in the middle to lower depths of O horizons are moved into the A and upper B (mineral) horizons to augment in situ OM deposition by plant roots. In the lower B, BC, and C horizons, direct plant OM inputs decrease with decreased rooting, relative to the input of decomposition products from the upper horizons. Cryoturbation mixes the lower B horizons with the lower O and upper mineral A horizons, resulting in intermediate C/N ratios similar to the A and upper B horizons. The relative susceptibility of SOM in these various horizons to increased decomposition rates under warming conditions cannot be predicted from just C/N ratios. Taken together with a priori knowledge of SOC depth profiles and the apparent ordered relationship between average C/N ratios and soil horizon types, however, C/N ratios could aid in future SOC sensitivity assessment. There are a number of soils studies from the boreal regions that have increased greenhouse gas emissions at depth in the soil profile with warming soil temperatures. Studies include forest soils (Goulden et al., 1998), permafrost soils (Schuur et al., 2009), and peat soils (Dorrepaal et al., 2009). The study by Michaelson and Ping (2003) of soil respiration sensitivity to temperature for arctic and transition arctic to boreal forest soils found that soils grouped by horizon types across a wide area responded similarly, with most of the differences being associated with the SOC of differing horizon types. Thus, additional work associating the distribution of SOC stores with soil horizon types and within soil profiles, and assessing their sensitivity to environmental changes could be most useful in future modeling of the mineralization of SOM and soil thermal properties of boreal region soils (c.f., Fan et al., 2008; Yi et al., 2009b).

Table 2. Average C/N ratios found for soil horizons listed in relative order of depth where found in soil profile sequences from top to bottom.

Soil horizon	*n*	Avg. C/N	SD (range)
O surface	35	40	10 (20–62)
Oe & Oa	42	28	10 (16–20)
A	19	22	6 (10–34)
E	5	23	7 (13–30)
B/Ajj	7	23	3 (17–25)
Bw	30	17	5 (7–27)
Bg	22	14	8 (4–24)
BC	11	11	5 (6–20)
C	34	14	5 (3–23)

For the purposes of scaling the patterns described here to the landscape in general, it is ideal to aggregate soil associations (for similar vegetation community types) into broad units, which could in turn be incorporated into C accounting models. For example, Yi et al. (2009a) recently demonstrated that by grouping organic layer soil characteristics by drainage class, temporal dynamics (~1000 yr) in organic layer thickness could be modeled in Canadian black spruce forests. Moreover, Ju and Chen (2005) have previously aggregated soil associations by drainage class to model spatial patterns of soil C stocks in Canada. In this study, we have grouped soils occurring in different regions of Alaska by drainage class (Table 1), as this is probably the best way to scale patterns at this time. A more detailed understanding of changes in soil taxonomic distinctions occurring in different geomorphic positions is required before finer scale aggregation of soil associations would be meaningful, owing to the relatively small number of full 1-m pedons described across the ~40 million ha of black spruce forest in Alaska.

The CEC, a measure of the ability of the soil exchange complex to retain cations, is generally positively correlated to OC contents in arctic tundra soils (Ping et al., 2005a). As SOM becomes more highly modified and decomposition proceeds, the CEC per unit SOM or OC increases (Stevenson, 1994). Because all OM is decomposed to some degree after being in the soil, it is intuitive to hypothesize that the CEC of organically dominated soils would increase with increasing SOC content. We compared the measured CEC and soil OC concentrations across all horizon samples (n = 245) and found a significant correlation (R^2 = 0.776, P = 0.001). This demonstrates the controlling role of SOM on CEC for these soils, and indicates that soil exchange sites are dominated by organic ligands. This is also consistent with the low clay contents observed for the mineral horizons from the A to lower C. Clay content varied little in these horizons, with a range of only 4to 10% (data from Ping et al., 2010). No significant correlation was found, however, between CEC

and clay. This lack of correlation between CEC and clay content was found, also, for arctic tundra soils (Ping et al., 2002).

With the relatively minor role of clay in the CEC, the amount of CEC in relation to SOC will change as SOM becomes more altered. This CEC/OC relationship should differ among genetically different soil horizon types that contain distinct forms or combinations of forms of SOM (Fig. 4a). The relationship between CEC and the CEC per gram of OC across genetic horizons illustrates that the exchange capacity of SOM (based on OC) increases with the depth of the soil horizon (going from surface O to the lowest C horizons), while the CEC of the soil decreases with decreasing OC. The surface organic horizons have high CEC, due primarily to their highly organic nature (low mineral content), but have lower CEC per gram of OC. Due to advanced decomposition that produces a more concentrated exchange capacity in the SOM, deeper soil horizons have a higher CEC per gram of OC. The higher concentration of CEC per gram of OC of the lower mineral horizons is an indication of the higher state of modification or decomposition in the SOM and, thus, the more exchange sites per gram of OC in the lower mineral horizons. This decomposition is probably the result of in situ biological activity or leaching of more soluble or colloidal decomposition products downward in the profile.

It is interesting to note that both the character of SOM (as indicated by CEC g^{-1} OC) and the soil CEC are intermediate for the near-surface A and E horizons (for soils with better drainage) and the B/Ajj horizons (from cryoturbated Gelisols). This indicates that they are in an intermediate decomposition state and is in agreement with the overall C/N ratio relationships discussed above (Table 2). The amounts and profile distribution of this intermediate-state SOM could be important in assessments of C dynamics and their response to warming or changing soil drainage conditions. There are relatively large amounts of SOM in these horizons that are present in both the warmer, drier soils and the colder, wetter permafrost soils at different landscape positions.

Acidity is produced in soils as SOM decomposes without neutralization by bases that are largely released from parent materials. Soil acidity or saturation of the CEC with H^+ ions increases as the SOM becomes more modified (releasing more H^+) or soil leaching proceeds. The average exchangeable soil H^+ across horizons mirrors the CEC, as does the H^+ retained by the CEC per gram of OC (Fig. 4). Both of these distributions, the exchange complex and the H^+ content in relation to SOC, reinforce the importance of SOC in the chemical and physical properties of these black spruce soils and, to a certain degree, reflect the minimal influence of parent materials, which supports the hypothesis that weathering effects on the mineral components of these cold soils is minimal (DeMent, 1962; Borden et al., 2010).

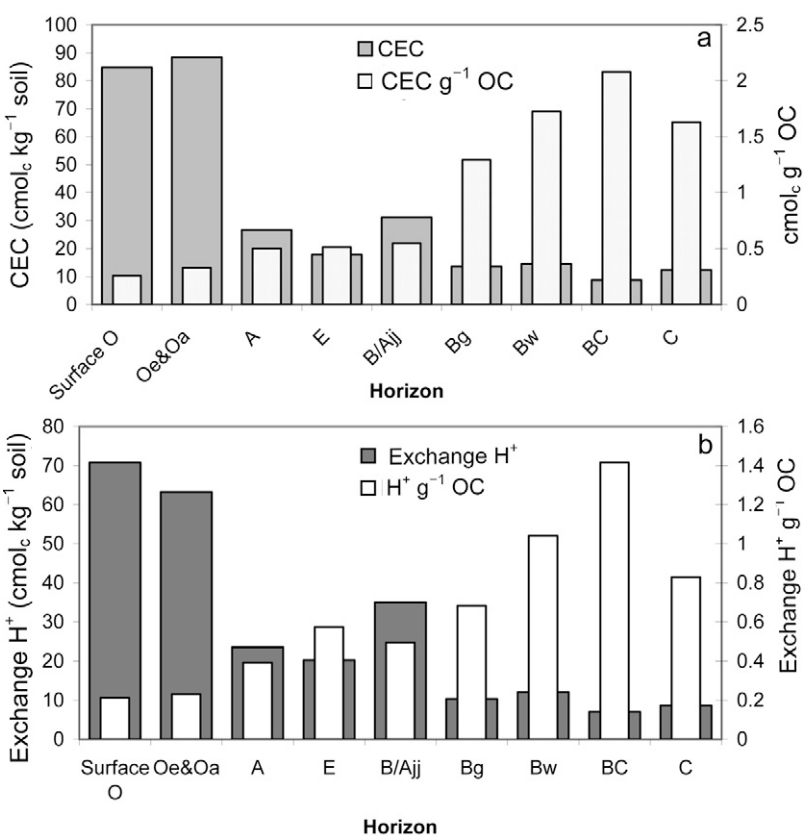

Fig. 4. Relationships between (a) average cation exchange capacity (CEC) and soil organic C (OC), and (b) exchange acidity and OC for genetic horizons shown by increasing depth in the soil profile from left to right.

Effects of Soil Organic Carbon on Soil Physical Properties

Other than the insulating qualities of surface OM buildup, one of the most important effects of SOC on the soil physical properties is on D_b. For mineral soil horizons, soil D_b decreases with the increase in C content across the range of 0 to 120 g OC kg^{-1}. For all soil horizons of the black spruce sites, a significant relationship exists between D_b and OC; D_b (g cm^{-3}) = −0.273 ln(g OC kg^{-1}) + 1.933, (R^2 = 0.863, P = 0.001, n = 304). This is a stronger correlation than reported for arctic Alaska soils (Bockheim et al., 2003). The presence of segregated ice also reduces D_b in soils affected by permafrost and is largely responsible for the greater variation in the relationship for arctic soils compared with the boreal forest soils in this study, where permafrost and segregated ice are limited. Such relationships are useful in calculating SOC stores in samples lacking D_b measurements, which are commonly encountered in many soil databases. With a strong regression relationship for these soils, relatively less variability would be observed between pedons from the black spruce boreal forest soils than that expected for arctic soils. The relationships of D_b as a function of OC content using just surface organic horizons, however, shows a larger variability and a weaker relationship (R^2 = 0.466, P < 0.001, n = 103), and also for only mineral horizons (R^2 = 0.624, P < 0.001, n = 201), so the error will be larger when estimating D_b based on OC for the surface O horizons. This could be significant because often, in up to 87% of Alaska soil pedons, the organic layers are missing D_b data in soil

databases (Ping and Michaelson, 2009). Also, the variation in D_b for organic horizons can vary 10-fold, compared with mineral horizons that vary only two- to threefold (Ping et al., 2010). The vegetation characteristics of the site, e.g., the presence of mosses compared with leaf or needle litter, with various degrees of compaction and degradation, can greatly affect the D_b of surface organic layers. Changes in the physiography (drainage and aspect characteristics) govern changes in the understory or groundcover composition (surface moss and vegetation), which in turn highly influence the D_b of the surface organic horizons.

CONCLUSIONS

This study provides the most complete analysis to date of soils and OC as it occurs under black spruce stands in Alaska. Black spruce forests tolerate a very wide range of soils and soil conditions in the Alaska boreal forest region. These soils formed in a wide variety of parent materials on landscapes from lowlands, outwash plains, and fluvial terraces to hilly uplands and under permafrost and non-permafrost conditions. Soil drainage classes range widely from somewhat excessively to very poorly drained. The soils are generally weakly developed and, thus, key soil properties such as CEC and acidity are closely related to the SOC content.

Surface soil organic horizons vary in thickness, ranging from 3 to 100 cm or more, with thicker organic horizons under more poorly drained conditions. Soil characteristics relating to the state of SOM decomposition, such as the development of CEC and acidity, indicate that the O horizons of surface organic layers contain SOM that is the least decomposed relative to that of the A horizons of the better drained soils and in the cryoturbated B/Ajj horizons of permafrost soils, which are in an intermediate state of decomposition. The most highly altered SOM is in the B and C horizons of the lower soil profile. Total C stores to 1-m depth vary widely, from 7 to 109 kg OC m^{-2}, but generally increase as the soil drainage condition decreases, with the higher OC stores associated with poorly and very poorly drained permafrost sites and very poorly drained non-permafrost sites.

The organic surface horizons contain anywhere from 13 to 100% of the SOC stores across all the soils. The higher SOC values generally found in poorly and very poorly drained permafrost sites are due to additional OC sequestered in the permafrost. Gelisols that are significantly cryoturbated (Turbels as opposed to Orthels) have a relatively higher proportion of SOC stored in the subsoil. In the boreal forest region of Alaska, permafrost exerts a strong control on OC stores due to the wet and cold conditions. Hence, soil sampling to a limited depth across the landscape may miss anywhere from about 40 to 90% of the total OC stores in the boreal forest. Our data indicate that for the most accurate assessment of SOC stores, SOC pools should be directly assessed to a depth of 1 m or more. Estimates can be made from surface stores if differences in profile SOC distribution with depth due to or related to soil formation processes or sequestration effects of drainage and permafrost conditions are taken into account.

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