Carbon Pools and Accumulation Rates in an Age-Series of Soils in Drained Thaw-Lake Basins, Arctic Alaska

J. G. Bockheim,* K. M. Hinkel, W. R. Eisner, and X. Y. Dai

ABSTRACT

We examined soils in 20 drained thaw-lake basins (DTLBs) near Barrow representing four age classes: young [0–50 years before present (BP)], medium (50–300 BP), old (300–2000 BP), and ancient (2000–5500 BP). In addition, we examined an erosional remnant that is about 8000 yr in age with deep organic deposits. There are significant age-related differences in thickness of the organic layer, which for DTLBs represents surface organic accumulation since lake drainage. The lower portion of the organic layer progresses in decomposition stage from fibric in young basins to sapric in ancient basins and the erosional remnant. Whereas extractable organic matter (humic- and fulvic-acid fractions) in the organic layer decreases, the nonextractable portion (humin) increases with basin age. There are significant age-related differences ($p = 0.027$) in soil organic carbon (SOC) of the organic layer; however, there were no significant differences in C pools for other depths or within the upper 100 cm. Profile SOC pools in DTLBs average 48 kg m^{-3}, which is less than elsewhere on the Arctic Coastal Plain (62 kg m^{-3}). From a depth of 100 to approximately 160 cm, SOC pools average 3.2 kg C m^{-2} dm^{-1}, confirming that there are large amounts of SOC below the traditional reporting depth of 100 cm. The spatial variability of SOC pools increases with relative basin age class and is likely because of increased variability in hydrology related to enrichment of ground ice in the upper permafrost. The long-term net accumulation (during the last 5500 yr) of SOC is 13 g m^{-2} yr^{-1}.

About 20% of the Arctic Coastal Plain of northern Alaska contains thaw lakes developed over ice-rich permafrost (Livingstone et al., 1958; Black, 1969). Near the village of Barrow, about 22% of the area is comprised of thaw lakes (Fig. 1). The long axis of these lakes has a preferred orientation of 352° (Hinkel et al., 2003), which is nearly perpendicular to the easterly winds that prevail in summer (Sellmann et al., 1975). Thaw lakes play a critical role in the livelihood of the native people as they contain fish and are the loci for migrating waterfowl and ungulates such as caribou. Lakes eventually drain to form what are called drained thaw-lake basins, that subsequently have accumulated organic matter. About 50% the Barrow area contains DTLBs (Hinkel et al., 2003).

The development and drainage of thaw lakes appear to be linked with changes in climate (Hopkins, 1949; Hopkins and Kidd, 1988; Sher, 1992), particularly with a regional deepening of the surface zone of seasonal thaw, known as the *active layer* (Burn, 1997). Lake drainage appears to have occurred preferentially in the colder part of the Holocene, beginning around 5000 BP (Mackay, 1992). In northwestern Canada, this period appears to be associated with peat accumulation in drained lake basins and reactivation of ice-wedge cracking and growth (Mackay, 1992). In arctic soils, C accumulation rates were higher in the warm early-to-middle Holocene than the cooler late Holocene, but it is not known whether this was because of changes in temperature, precipitation, plant succession, or local depositional factors (Marion and Oechel, 1993).

A conceptual model of the thaw-lake cycle has been developed for the Arctic Coastal Plain of northern Alaska (Hopkins, 1949; Britton, 1966; Billings and Peterson, 1980; Hinkel et al., 2003). In this model, the cycle begins with ponding over ice-wedge troughs and low-center polygons. Eventually, the ponds coalesce to form a small lake. Thermal erosion along the lake margins, combined with thawing of permafrost below the standing water, increases the lake dimensions across time. If the lake is sufficiently deep, ice wedge growth will cease and the wedges may begin to ablate as the thaw bulb below the lake deepens. In time, the lake drains by ice-wedge erosion, stream piracy, tapping, bank overflow, or coastal erosion (Mackay, 1988). Vegetation and organic matter accumulation begin after the lake drains, and ice-wedge growth is usually reactivated as permafrost aggrades into the unfrozen substrate below the drained lake basin. Ice-wedge growth yields low-centered polygons with ponds forming preferentially over the resulting troughs. In this way, thaw-lake development may begin anew. Billings and Peterson (1980, p. 426) explained and described graphically the various stages of the thaw-lake cycle.

Although soils of the Arctic Coastal Plain have been sampled to determine the C content (Michaelson et al., 1996; Bockheim et al., 1997, 1999, 2002, 2003), no studies have been undertaken within thaw-lake basins that systematically examine the spatial and temporal variations in SOC. Accordingly, the objective of this study was to examine changes in organic C in a developmental sequence of DTLBs on the Arctic Coastal Plain.

STUDY AREA

The study was undertaken on the Barrow Peninsula, Alaska, which is underlain by continuous permafrost to a thickness of around 400 m (Brewer, 1958). The active layer ranges from 30 to 90 cm in thickness (Nelson et al., 1998). The Barrow Peninsula has low relief with elevations ranging from 0 to 20 m above sea level. Soil

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Abbreviations: BP, years before present; DTLB, drained thaw-lake basin; SOC, soil organic carbon; PER, Peterson Erosional Remnant.

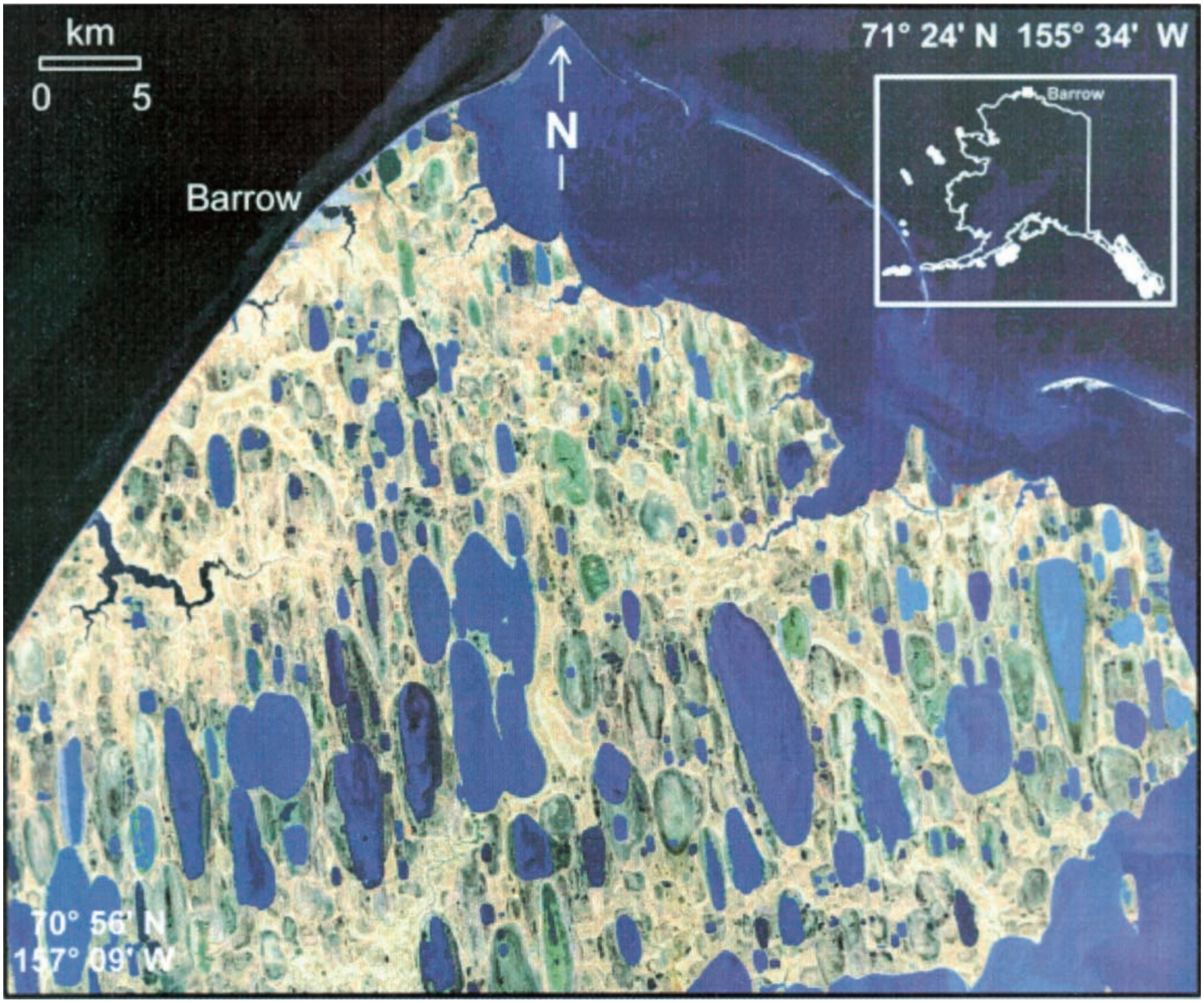

Fig. 1. Landsat7 ETM+ image from 30 Aug. 2000 for the Barrow Peninsula, Alaska, showing oriented thaw lakes (blue) and shadows of drained thaw lake basins (green).

parent materials are marine sediments of Pleistocene age that have been reworked by thaw-lake processes (Sellmann and Brown, 1973). Barrow has a cold maritime climate. Winters are long, dry, and cold, and summers are short, moist, and cool. The mean annual air temperature is −12.0°C; July is the warmest month at 4.7°C, and February is the coldest month at −26.6°C (National Climate Data Center, 2002). Mean annual precipitation is 106 mm, 63% of which falls as rain during July through September. The winter snowpack averages 20 to 40 cm, but snow accumulation is highly variable because of terrain roughness and drifting from strong easterly winds.

Eight major vegetation associations have been recognized in the Barrow region, including *Luzula* heath, *Salix* heath, *Carex-Poa* meadow, *Carex-Oncophorum* meadow, *Dupontia* meadow, *Carex-Eriophorum* meadow, *Arctophila* pond margin, and *Cochlearia* meadow (Webber, 1978). C. Tweedie mapped five major landcover types in the Barrow region using high-resolution satellite imagery (IKONOS), including dry heath, dry meadow, moist meadow, wet meadow, and emergent aquatic (http://www.cevl.msu.edu/ael/posters/ikonos-landcover.html, verified 28 Oct. 2003).

Soils of the Barrow region have been mapped at a scale of 1:20 000 (Drew, 1957; Bockheim et al., 1999, 2002) and include 12 soil subgroups, of which four are classified as Turbels, four as Orthels, and four as Histels. There is considerable spatial variability in the soils due to the presence of low-, flat-, and high-centered ice-wedge polygons.

MATERIALS AND METHODS

Field

We examined 76 DTLBs in the Barrow region that ranged between 70°56′ and 71°24′ N latitude and 155°34′ and 157°09′ W longitude. The basins were classified into one of four age classes, including young (0–50 BP), medium (50–300 BP), old (300–2000 BP), or ancient (2000–5500 BP) (Hinkel et al., 2003). The development and verification of the classification scheme, including ^{14}C dating, are described by Hinkel et al. (2003).

A total of 124 cores were collected from 53 of the basins in August 2000, April 2001, August 2001, and April 2002 with a 7.5-cm-i.d. SIPRE auger barrel powered by either a standard posthole digger (August sampling) with helicopter access, or a Big Beaver (Little Beaver, Inc., Livingstone, TX)[1] earth drill apparatus that was mounted on a sledge and pulled by a snow machine. Cores were taken near the center of the basin while avoiding large ponds; they ranged from 48 to 170 cm long (average = 116 cm). Coring was terminated when massive ice, generally in the form of ice wedges, was encountered.

To evaluate spatial patterns of variability, one basin from each age category was selected for intensive study. Three east-to-west transects were surveyed across the northern, middle, and southern reaches of each basin, and three cores were collected from each transect. In addition, a core was taken from the northern and southern ends of the basin. A 12th core was obtained adjacent to the coring location where the organic layer was thickest.

In addition, we identified an area south of Barrow informally named the Peterson Erosional Remnant (PER). It is a prominent topographic feature (>20 m of relief) that is characterized by well-developed ice-wedge polygons, high ice content of the upper permafrost, and plant communities that differ from those in DTLBs. The PER appears to have escaped the effects of thaw-lake processes, and is thus an erosional remnant. A ^{14}C date on organic material from the 69-cm depth yielded an age of 8070 BP (Eisner, 2003, unpublished data). Seven cores were collected from the PER in August 2001 and April 2002.

Laboratory

In a cold room at Barrow, the cores were photographed and cut with a chop-saw into sections representing soil horizons within 12 h of collection. Core sections were dried at 70°C. Field moisture content was determined, and bulk density was estimated on an oven-dry basis. Emphasis was placed on the organic layer, which represents the amount of organic matter that has accumulated since lake drainage. The thickness of this layer was measured, and the maximum degree of decomposition (e.g., fibric, hemic, or sapric) was estimated from rubbed and unrubbed fiber contents (Lynn et al., 1974). The maximum degree of decomposition normally occurred at the base of the organic layer. Soils were classified using the *Keys to Soil Taxonomy* (Soil Survey Staff, 1998).

Soil organic C determinations were made on more than 500 samples from 88 cores representing 20 basins roughly evenly distributed by age class. Five cores were taken from the PER, one of which was sectioned into 1-cm intervals for SOC determination. Samples were passed through a 100-mesh (149-μm) sieve. Total C was determined on a Dohrmann DC190 C analyzer. None of the soil samples reacted with 1 *M* HCl, indicating that the values represent organic C exclusively.

Thirty-five samples from the surface organic layer were selected from 31 cores and 11 basins representing the four age classes for characterizing extractable SOC. The samples were extracted with 0.1 *M* NaOH at a solution-to-soil ratio of 15:1. The extract was separated by centrifugation and filtration through a 0.45-μm polysulfanone membrane filter; the extraction was repeated three times (Ping et al., 1995). The filtrate was acidified with 6 *M* HCl to pH 1 and left overnight in an ice-bath to allow for precipitation of humic acids. Humic acids were then separated by centrifugation and filtration. The extraction yielded three fractions: humic-acid C, fulvic-acid C, and a nonextractable fraction representing humin.

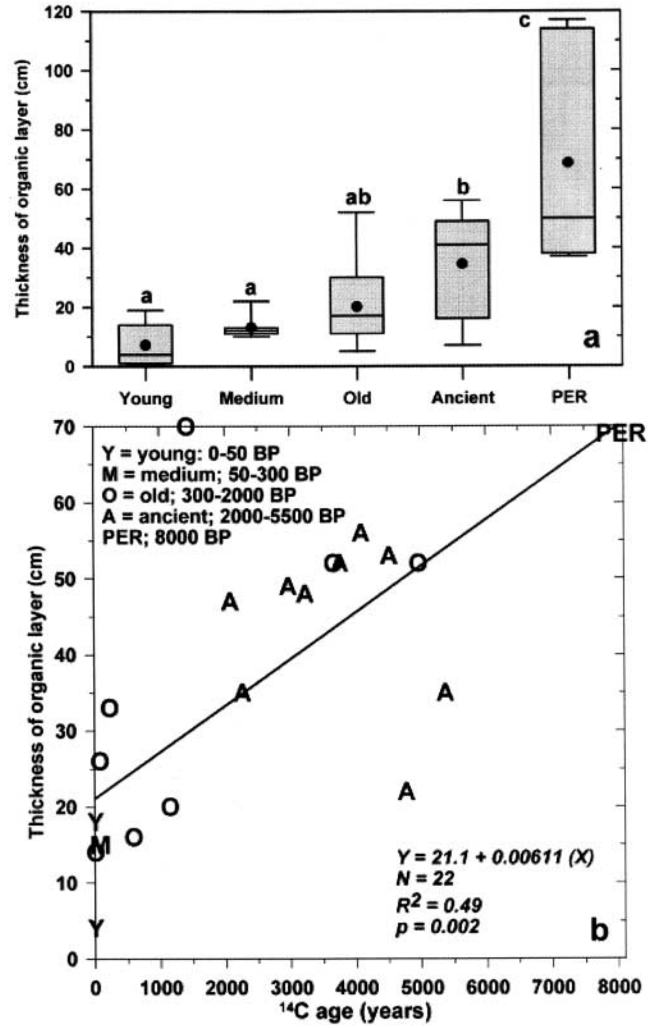

Fig. 2. Relation between thickness of the organic layer and basin age class (upper panel) and between thickness of the organic layer and ^{14}C age (lower panel) (after Hinkel et al., 2003). In the upper panel, the median value with quartiles is shown by horizontal bars, the range of values by *whiskers*, and the mean with a black dot; differences in small case letters are statistically significant at p = 0.05. PER, Peterson Erosional Remnant.

Twenty-two samples of organic matter, primarily in the form of decomposed plant fiber, were collected from just above the organic–mineral interface and analyzed at the Lawrence Livermore National Laboratory for radiocarbon dating via the accelerator mass spectrometer technique (Hinkel et al., 2003; Eisner, 2003, unpublished data). These dates are assumed to represent the date of lake drainage and subsequent initiation of organic matter accumulation.

Computations and Statistics

Soil C density was determined for four layers: the organic layer (representing the buildup of organic matter since thaw-lake drainage), from the base of the organic layer to a depth of 35 cm (35 cm approximates the mean seasonal thaw depth

[1] Mention of trade names does not imply endorsement by the Universities of Wisconsin or Cincinnati.

Table 1. Morphological characteristics, C pools, and classification of soils of drained lake basins, arctic Alaska.

Age class	Site name	Degree of decomp.†	Thickness of organic layer	C content of organic layer	C content base of organic to 35 cm	C content of 35- to 100-cm layer	C content of 0- to 100-cm layer	Soil subgroup classification
			cm	kg m^{-2}			kg m^{-3}	
Young	VB-3	i	1	0.1	28.4	20.9	49.4	Typic Aquorthels
	VB-8	i	14	2.6	19.5	30.5	52.6	Typic Aquiturbels
	VB-16	i	19					Typic Historthels
	VB-24	i	4					Typic Aquorthels
	VB-35	i	1					Typic Aquorthels
	Footprint	i	9	6.2	14.6	12.4	33.2	Typic Aquorthels
	Dry	i	3	0.9	26.5	11.9	39.3	Typic Aquorthels
	Mean		7	2.5	22.3	18.9	43.6	
Medium	VB-1	i	12	3.1	13.7	14.3	31.2	Typic Aquiturbels
	VB-22	e	13					Typic Aquorthels
	VB-23	e	11					Typic Aquorthels
	VB-30	I	12					Typic Aquorthels
	VB-34	i	12					Typic Aquorthels
	Central Marsh	e	10	5.3	12.8	32.7	50.8	Glacic Historthels
	Golf Course	i	22	8.3	7.2	11.2	26.7	Typic Fibristels
	Mean		13	5.6	11.2	19.4	36.2	
Old	VB-2	i	18	12.1	12.3	41.4	65.8	Typic Aquiturbels
	VB-4	i	17	8.2	12.6	17.2	37.9	Typic Aquiturbels
	VB-6	e	12	5.0	20.5	27.0	52.5	Typic Aquiturbels
	VB-9	i	14	8.4	13.8	21.8	44.0	Typic Aquiturbels
	VB-10	i	11	6.2	16.1	24.2	46.4	Typic Aquiturbels
	VB-11	i	11	7.2	20.4	33.4	61.1	Typic Aquiturbels
	VB-14	e	10					Typic Historthels
	VB-17	i	31					Typic Historthels
	VB-20	a	29					Typic Historthels
	VB-21	a	52					Typic Sapristels
	VB-27	e	30					Hemic Glacistels
	VB-28	i	5					Typic Historthels
	VB-31	i	30					Typic Fibristels
	VB-32	i	20					Typic Historthels
	Truncated	i	12	9.1	28.2	15.9	53.2	Glacic Fibristels
	Mean		20	8.0	17.7	25.8	51.6	
Ancient	VB-5	e	15	10.7	15.5	11.4	37.6	Glacic Histoturbels
	VB-7	e	7	5.4	28.1	12.4	45.9	Glacic Aquiturbels
	VB-13	a	20					Glacic Sapristels
	VB-15	a	35					Typic Historthels
	VB-18	a	56					Typic Hemistels
	VB-19	a	49					Typic Sapristels
	VB-25	a	52					Typic Hemistels
	VB-26	a	48					Typic Sapristels
	VB-29	e	16					Typic Aquorthels
	VB-33	e	47					Sapric Glacistels
	Mean		35	8.1	21.8	11.9	41.8	
PPP	01-01	a	38					Sapric Glacistels
	01-02	a	50					Sapric Glacistels
	02-01	a	114					Glacic Sapristels
	02-02	a	41					Sapric Glacistels
	02-03	a	37					Sapric Glacistels
	02-04	a	83	139+			139.0	Typic Sapristels
	02-05	a	117					Glacic Sapristels
	Mean		69					

† i = fibric (least decomposed); e = hemic; a = sapric (most decomposed).

for these soils), the 35- to 100-cm layer (near-surface permafrost), and for the upper 100 cm (traditional depth for reporting SOC by life zone; Post et al., 1982). Carbon density was estimated using equations given by Michaelson et al. (1996) and Bockheim et al. (2003). The mean long-term net accumulation was estimated by dividing SOC of the organic layer by radiocarbon age (Oksanen et al., 2001). These estimates were made for basins > 100 BP because of uncertainty of very young radiocarbon ages.

Age-related differences in thickness of the organic layer and SOC pools were evaluated using analysis of variance (ANOVA) followed by post hoc separation of means with Fisher's LSD test (Minitab, 2000). The relation between thickness of the organic layer and basin age class was assessed by regression analysis.

RESULTS AND DISCUSSION

Thickness of Organic Layer

There were highly significant ($p < 0.001$) differences between thickness of the organic layer and basin age class (Fig. 2a). A linear function best explained the relationship between surface organic layer thickness and ^{14}C age (Fig. 2b). The equation in Fig. 2b suggests that organic layers in DTLBs form at a rate of approximately 10 cm per 1000 yr, which is comparable with values reported in other arctic regions during the same time interval (Oksanen et al., 2001; National Wetlands Working Group, 1988). Two of the basins classified in the *old* category were radiocarbon dated as *young* or *medium* and one basin was dated as *ancient* (Fig. 2b). In general,

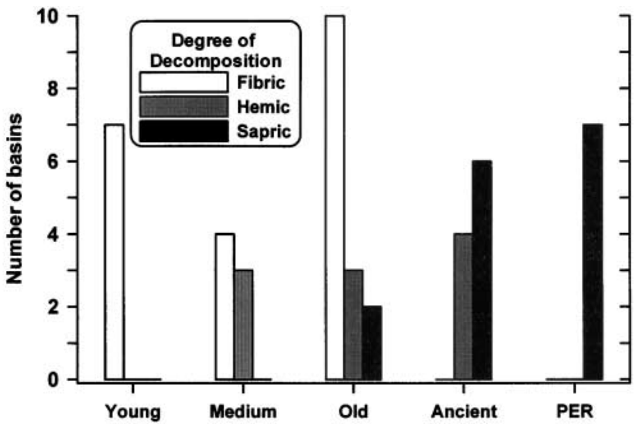

Fig. 3. Relation between degree of decomposition of the organic layer and basin age class. PER, Peterson Erosional Remnant.

however, radiocarbon dating validated our assignment of the basins to specific age classes (Hinkel et al., 2003).

Morphological and Chemical Changes in Organic Layer

Not only did the thickness of the organic layer increase with basin age, but also the morphology of the organic horizons and forms of extractable C changed across time. All of the lake basins classified as young contained a fibric (Oi) horizon. Medium and old basins had either a fibric or a hemic (Oe) horizon, and ancient basins generally had a sapric (Oa) horizon. The PER, an erosional remnant, contained deep (average = 68 cm; Table 1) sapric materials (Fig. 3).

The change in morphology of the organic layer with time was accompanied by changes in the forms of extractable C. The proportion of extractable C (humic- and fulvic-acid C) decreases, whereas the proportion of nonextractable C (humin) increases with basin age (Fig. 4). These trends likely reflect the gradual *ripening* of organic matter to more recalcitrant forms (van Heuveln et al., 1960; Pons and van der Molen, 1973). The ranking of organic matter forms by abundance is: humin > humic-acid C > fulvic-acid C, which is comparable with fractions of tundra soils in arctic Alaska as determined from carbon-13 nuclear magnetic resonance spectroscopy (Dai et al., 2001). The composition of the more-resistant humin fraction is not known, but may include paraffinic C derived from algal or microbial sources (Ping et al., 1995; Dai et al., 2001).

Carbon Pools and Accumulation Rates

There are significant ($p = 0.027$) age-related differences in SOC content of the organic layer (Table 1; Fig. 5). However, there were no consistent age-related trends in SOC storage at any of the other three depths. These lack of trends in SOC storage are likely because of variable amounts of recycled C in the lacustrine sediments below the organic layer.

On average, DTLBs contained 48 kg C m^{-3} to a depth of 100 cm (Table 1), which is less than the 62 kg C m^{-3} reported for the Alaskan Arctic Coastal Plain by Michaelson et al. (1996) and Bockheim et al. (1999).

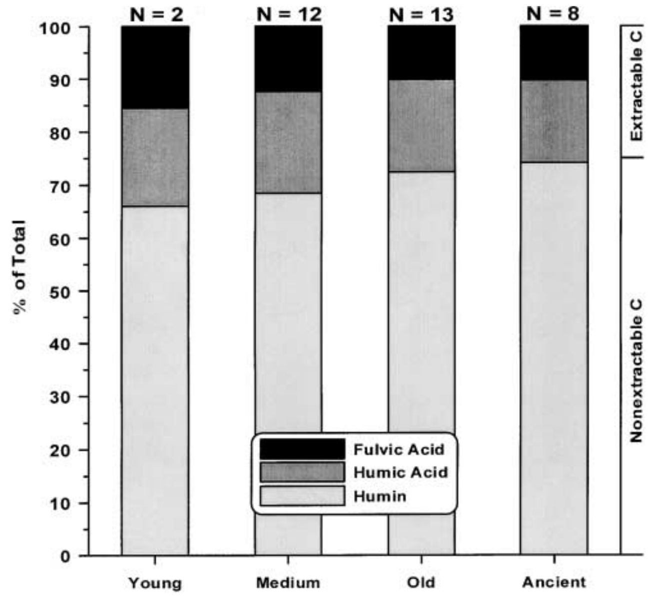

Fig. 4. Relation between forms of extractable organic matter in the organic layer and basin age class.

However, the cited studies were conducted in the Kuparuk River valley, where loess-derived inorganic C contributes substantially to the total C content of soils.

Traditionally, SOC pools are reported for the upper 100 cm (Post et al., 1982). While this may be appropriate for most life zones, our data suggest that there is considerable SOC below the standard depth of 100 cm (Table 1; Fig. 6). This SOC likely originates from lacustrine sediments and is sequestered by upward aggradation of permafrost across time.

The SOC density from 100 to approximately 160 cm ranges between 1.2 and 8.0 kg C m^{-2} dm^{-1} and averages

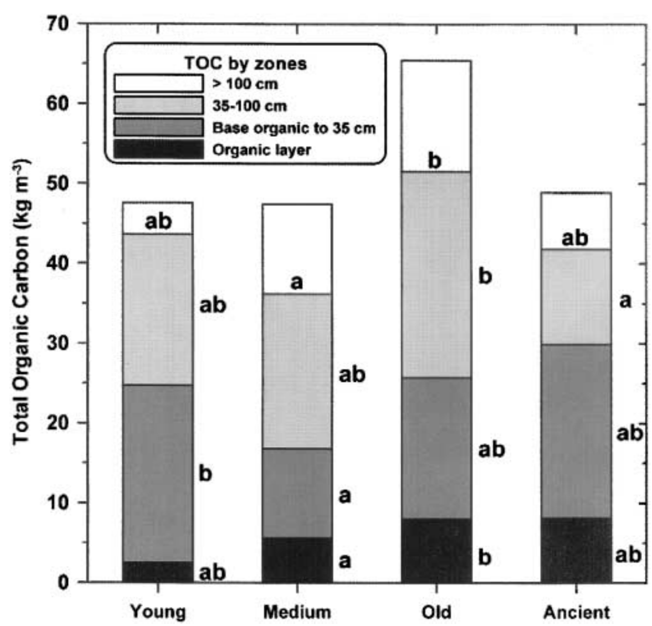

Fig. 5. Carbon density of the organic layer, the base of the organic layer to a depth of 35 cm, 35 to 100 cm, 0 to 100 cm, and >100 cm in relation to basin age class. Differences in small-case letters distributed across the basin age classes for a particular layer in the profile are statistically different at $p < 0.05$. TOC, total organic C.

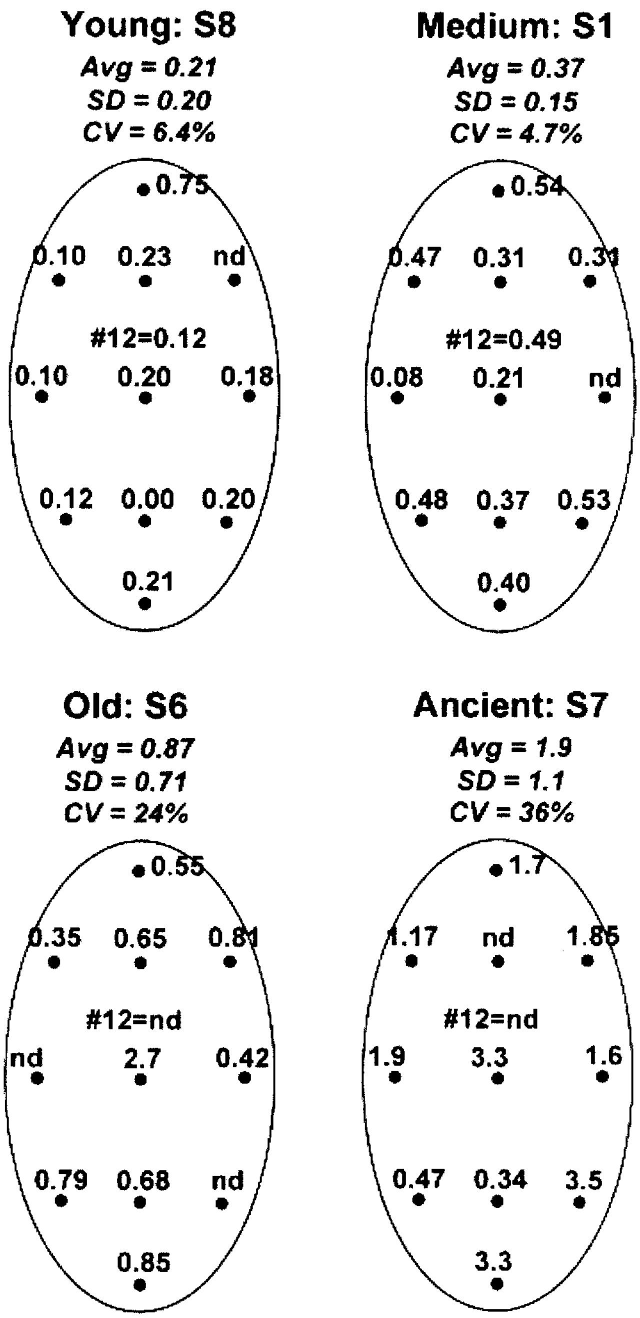

Fig. 6. Spatial variability in SOC of 12 samples collected from the organic layer of young, medium, old, and ancient drained thaw-lake basins. CV, confidence value; nd, not determined.

3.2 kg C m^{-2} dm^{-1}. These results agree with those of Tarnocai (2000), who reported that in some Mackenzie Valley soils in Canada, the 0- to 100-cm depth contains only 55% of the total C mass present in the 0- to 300-cm depth. We conclude that estimates of SOC pools for the arctic tundra should be reported to depths in excess of 1 m.

Our results indicate considerable intra- as well as interbasin variability in SOC pools in the organic layer of DTLBs (Fig. 6), although the observed pattern of increasing SOC with basin age is again apparent. The coefficient of variation of the SOC pool in the surface organic horizon increases with basin age. These patterns of spatial and temporal variation likely reflect the dynamics of partial or differential lake draining (Billings and Peterson, 1980) and post-drainage recovery. Local changes in surface hydrology may be attributed to the accumulation of ground ice, differential development of microtopography, and vegetational succession.

Table 2. Long-term net C accumulation rates in drained thaw-lake basins.

Basin age class	^{14}C age range	Long-term net C accumulation†
	yr	g m^{-2} yr^{-1}
Young	0–50	10.0, 260, 620 (267)
Medium	50–300	38.8, 52.6, 90 (60)
Old	300–2000	7.1, 14.2 (10.6)
Ancient	2000–5500	2.2, 1.3, 2.3 (1.9)
Mean of all basins >100 yr		13.2

† Estimated by dividing C content of the organic layer by 14C age [data from Hinkel et al. (2003) and Eisner (2003, unpublished data)]. Mean values shown in parentheses.

Immediately following lake drainage, the accumulation of SOC in the organic layer is high, averaging 267 g m^{-2} yr^{-1} in young basins (ca. 50–300 BP) and 60 g m^{-2} yr^{-1} in medium-aged basins (ca. 50–300 BP) (Table 2). The mean accumulation of SOC in the organic layer was 11 g m^{-2} yr^{-1} in old (ca. 300–2000 BP) and 1.9 g m^{-2} yr^{-1} in ancient (ca. 2000–5500 BP) basins. This decline in accumulation of SOC is common in arctic peatlands (Oksanen et al., 2001) and commonly is attributed to changes in decomposition and quality of organic matter from plant succession. However, we suggest that the increase in the amount of segregation ice in the near-surface permafrost of DTLBs across time (e.g., Mackay, 1992) effectively dilutes the organic C. For example, massive ice was ubiquitous in cores taken from the PER. Increasing the content of pore and segregation ice lenses following the deposition of organic matter increased the mass and causes upward soil displacement, thus reducing the concentration of SOC.

The mean long-term net accumulation of SOC for basins > 100 BP is 13 g m^{-2} yr^{-1}, which is comparable with values reported for peat sequences in the European and Russian arctic (Botch et al., 1995; Makila et al., 2001; Oksanen et al., 2001).

Soil Classification

Soils in young and medium DTLBs are dominantly Aquorthels (Table 1). In contrast, soils in old basins are Aquiturbels and Historthels; soils in ancient basins commonly are Hemistels and Sapristels containing a glacic layer, and the PER contains either Sapric Glacistels or Glacic Sapristels. These developmental changes in soil classification reflect (i) the thickening of the surface organic layer with time, (ii) a higher degree of decomposition of organic matter, especially in ancient basins and the PER, and (iii) a gradual increase in the amount of ground ice in the upper permafrost.

CONCLUSIONS

The existence of a Holocene-aged sequence of soils in DTLBs in arctic Alaska offered a unique opportunity to examine the development and evolution of organic soils underlain by permafrost (Histels). With the exception of soils on a high erosional remnant, all of the soils examined range between 10 and 5370 BP and could be arrayed into four age classes from remote sensing and ^{14}C dating (Hinkel et al., 2003). The following properties of the organic layer increased with basin age: thickness of the organic layer, morphology of the organic horizons (i.e., degree of decomposition), proportion of recalcitrant extractable C (humin), and spatial variability in SOC in the organic layer. Profile quantities of SOC averaged 48 kg m^{-3}, which is less than values reported for the Arctic Coastal Plain (62 kg m^{-3}). The average long-term net accumulation of C in basins $>$ 100 BP was 13 g m^{-2} yr^{-1}, which is comparable with values reported in the European and Russian arctic.

The results of this study have important implications regarding the possible impacts of climate warming in arctic regions. The reported increase in summer temperature and seasonal thaw depth (Serreze et al., 2000) could enhance decomposition of SOC in the seasonal thaw layer and release CO_2 from decomposition in the near-surface permafrost, thereby exacerbating warming. In addition, the observed increase in seacliff erosion (Serreze et al., 2000) could initiate drainage of thaw lakes. Thus far, studies indicate that thaw lakes form during periods of warming; however, thawing of ice wedges could either enlarge existing thaw lakes or contribute to lake drainage. Increased precipitation could create changes in hydrology with uncertain effects on the landscape, such as an increase in stream piracy, leading to more drainage or higher lake levels, leading to bank overflow. Given these uncertainties, it is essential that thaw lakes and DTLBs be monitored. Rapid climate changes will have a profound effect on the landscape, vegetation, soils, and lakes of the Arctic Coastal Plain, as well as on the animals and human cultures that depend on them.

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