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Abstract

Thaw lakes and drained thaw lake basins are ubiquitous on the Arctic Coastal Plain of Alaska. Basins are wet depositional environments, ideally suited for the accumulation and preservation of organic material. Much of this soil organic carbon (SOC) is currently sequestered in the near-surface permafrost but, under a warming scenario, could become mobilized. The relative age of 77 basins on the Barrow Peninsula was estimated using the degree of plant community succession and verified by radiocarbon-dating material collected from the base of the organic layer in 21 basins. Using Landsat-7+ imagery of the region, a neural network classifying algorithm was developed from basin age-dependent spectra and texture. About 22% of the region is covered by 592 lakes (>1 ha), and at least 50% of the land surface is covered by 558 drained lake basins. Analysis of cores collected from basins indicates that (1) organic layer thickness and the degree of organic matter decomposition generally increases with basin age, and (2) SOC in the surface organic layer tends to increase with basin age, but the relation for the upper 100 cm of soil becomes obscured due to cryoturbation, organic matter decomposition, and processes leading to ice enrichment in the upper permafrost.

Introduction

About 20% of the Arctic Coastal Plain of northern Alaska, and large portions of the Arctic Foothills and Seward Peninsula, contain thaw lakes developed in ice-rich permafrost (Livingstone et al., 1958; Black, 1969). Most of these lakes are elliptical (Figure 1), with the major (long) axis oriented a few degrees west of due north and nearly perpendicular to the prevailing summer wind direction (Sellmann et al., 1975). A much larger proportion of the terrestrial surface is scarred by elliptical depressions formerly occupied by thaw lakes; these are known as drained thaw-lake basins. Hussey and Michaelson (1966) estimate that 50–75% of the Coastal Plain is covered either by lakes or former thaw-lake basins. The concentration of lakes and basins decreases inland, with greatest frequency and largest size occurring on the Outer (seaward) Coastal Plain (0–7 m a.s.l.) compared to the Inner Coastal Plain.

This report discusses the initial findings of a coordinated research effort to understand the geomorphological and ecological evolution of thaw lakes on the Arctic Coastal Plain. Emphasis is placed on carbon sequestered in drained thaw-lake basins, which are sites of preferential accumulation of soil organic carbon (SOC). We have focused our efforts thus far on the Barrow Peninsula, a 1600-km^2 region adjoining the Arctic Ocean. Using satellite imagery, field sampling, and laboratory measurements, we obtained a spatial estimate of near-surface carbon reservoirs. If general circulation models are correct in predicting enhanced warming at high latitudes, and if warming increases the depth of summer thaw, this SOC is susceptible to mobilization. Conversion of SOC to atmospheric CO_2 or CH_4 by microbial decomposition would serve as a positive feedback mechanism (Billings et al., 1982).

The Thaw Lake Cycle

Several benchmark publications (Hopkins, 1949; Carson and Hussey, 1962; Britton, 1966; Billings and Peterson, 1980) have con-

tributed to the development of a descriptive model of the thaw-lake cycle for the Arctic Coastal Plain of northern Alaska, a region entirely within the zone of continuous permafrost. The cycle begins with the development of small ponds at the intersection of ice-wedge troughs or in low-center polygons. Ponds eventually coalesce and enlarge to form small lakes. Thermal erosion along the lake margins, combined with melting of permafrost below the standing water, increases lake dimensions over time. Thaw subsidence and thermokarst are favored in sediments supersaturated with ice, and when the ground ice melts, the basin becomes a reservoir for meteoric and snow meltwater runoff (Livingstone et al., 1958; Hussey and Michelson, 1966; O'Sullivan, 1966). The lake can deepen over time through ablation of ground ice and thaw settlement. Once water depth exceeds about 2 m (Brewer, 1958), lakes no longer freeze to their bottoms and a thaw bulb gradually forms beneath the lake basin. Thaw subsidence continues as pore ice, segregation ice, and ice wedges melt. As the thaw front penetrates to depth, the concentration of ice decreases and the subsidence rate is reduced. In shallower basins (<2 m), a thaw bulb does not develop because the winter ice cover freezes to the lake bottom (Lachenbruch et al., 1962).

Lake basins grow laterally by thermoerosion processes and often develop asymmetrical elliptical outlines. In low-relief terrain underlain by ice-rich sediments, wind-oriented waves produce circulation cells that induce currents and concentrated thermoerosion in zones oriented at 50° to the wave approach (Livingstone, 1954; Rex, 1961; Carson and Hussey, 1962; Mackay, 1963). The two-cell circulation pattern results in enhanced erosion approximately normal to the prevailing summer wind direction. Measured bank erosion rates at the degrading lake ends range from several to 25 cm yr^{-1} (Black, 1969; Tedrow, 1969; Brown, 1970; Lewellen, 1972), and bank undercutting yields mats of fibrous organic material that decrease in size with time. Reworked clastics and organic materials are transported by currents from the eroding lake ends and deposited on shallow shelves on the
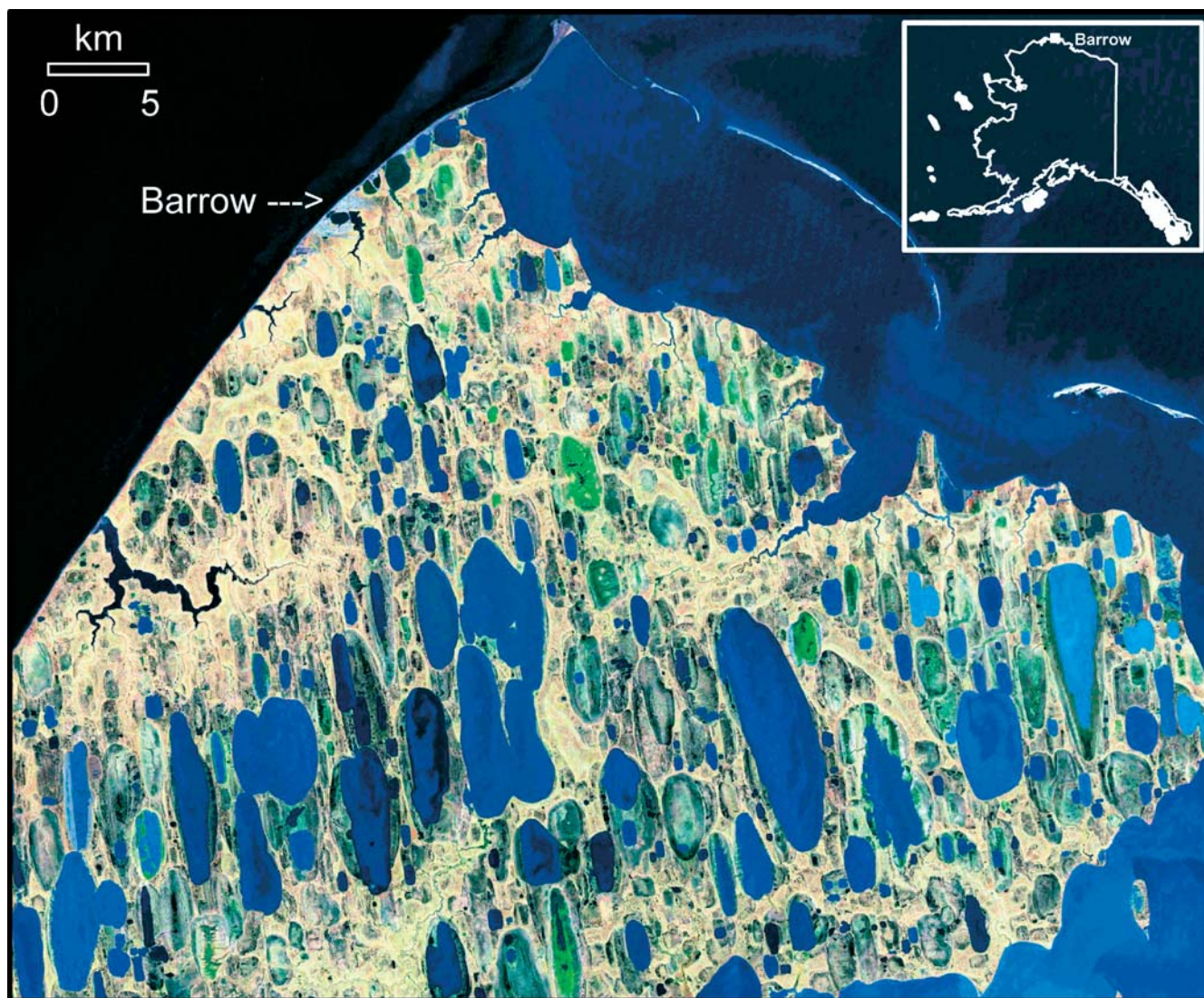
FIGURE 1.   Landsat-7+ image from 30 August 2000 for Barrow region.

upwind and downwind margins of the basin. These sediments lie unconformably above the original subsided surface layer (Murton, 1996). Because they contain reworked organics, these sediments cannot be used for radiometric dating (Anderson, 1982).

As an oriented elliptical lake expands in size, it may encounter and incorporate nearby ponds and lakes; several examples are apparent in Figure 1. Eventually, however, the lake drains. This may be triggered by ice-wedge erosion, headward stream erosion, tapping, bank overflow, or coastal erosion (Hopkins, 1949; Walker, 1978; Mackay, 1988). Drainage via ice-wedge erosion can be catastrophic (Hopkins and Kidd, 1988; Mackay, 1992); Mackay (1988) estimated that 1–2 lakes in the Tuktoyaktuk Peninsula of northwest Canada drain suddenly each year. Complete lake drainage can occur, but most lakes drain partially, leaving a residual lake or several ponds nested within the older basin. Following lake drainage, revegetation and organic matter accumulation begin, and ice-wedge growth is usually re-activated as permafrost aggrades into the unfrozen substrate below the drained lake basin (Mackay and Burn, 2002). Ice-wedge growth eventually yields low-centered polygons with ponds forming preferentially over the resulting troughs, beginning a new phase of the thaw-lake cycle.

Vegetation is established in drained or partially drained basins. Vegetation communities succeed one another as edaphic conditions change, and surface organic material accumulates above the lacustrine sediments. Ground heave, polygon development, and slope processes combine to slowly obliterate the basin, and it eventually appears as wet sedge meadow tundra characterized by *Carex aquatilis* Wahlenb., tall cottongrass (*Eriophorum angustifolium* var. *triste* Honckeny), white cottongrass (*E. Scheuchzeri* Hoppe), and Fisher's tundra grass (*Dupontia fisheri* R. Br.) (Webber, 1978). Because thaw-lake basins often develop in older basins, nested patterns form a palimpsest that dominates the landscape.

Although forming today, many thaw lakes apparently came into existence during the warmer early Holocene around 10,000 BP (Ritchie et al., 1983; Hopkins and Kidd, 1988; Rampton, 1988; Cote and Burn, 2002). (Throughout this paper, BP refers to uncalibrated radiocarbon years, and calibrated dates are given as cal yr BP). In Canada (Mackay, 1992), Alaska (Hopkins, 1949), and Siberia (Czudek and Demek, 1970; Kuzin and Reynin, 1972; Sher, 1992), the formation and drainage of thaw lakes is linked to changes in climate. Their development is apparently associated with regional thickening of the active layer and development of thermokarst in ice-rich permafrost (Burn, 1997). Dating of lacustrine strand lines near the village of Barrow indicates that lakes drained ca. 3500 BP (Carson, 1968; 2001). In northwestern Canada, lake drainage probably occurred preferentially in the colder mid-Holocene, beginning around 4500 BP. This period

Category	Young	Medium	Old	Ancient
Surface organic thickness (cm)	<5	10–15	15–30	15–35
Ice content of uppermost permafrost (%)	30–70	15–50	20–60	30–65
Organic form (decomposition)	fibric	fibric-hemic	hemic-fibric	sapric-hemic
Frost polygons				
Form	none or some erosional high center	flat center	flat center to ponded low center	ponded low centers to coalescing ponds and small lakes
Rim height (m)	none	indistinct	0.25–0.40	0.40–0.75
Vegetation	*Arctophila fulva* on all but the driest surfaces; *Dupontia fisheri*; *Eriophorum scheuchzeri* common; few *Carex aquatilis* but clones <4 m. Drepanocladus and Calligeron mosses in water	Clones of *Eriophorum angustifolium* & *C. aquatilis* (4 to 20-m diameter) Background veg. similar to Young, incl. *A. fulva*, *E. scheuchzeri*, and *D. fisheri*	Continuous *C. aquatalis* in polygon centers w. *Sphagnum*, *Dicranum elongatum* & lichens on rims (more rim species as polygons develop). *A. fulva* restricted to ponds	*C. aquatalis* in low center polygon centers; well developed upland vegetation on rim, incl. *Cassiope tetragona* and *Poa supp*. *A. fulva* absent or only in ponds
Standing dead leaf material	Little except on driest areas	Moderate	Considerable esp. on rims	Much standing dead
Relative plant vigor	High	Medium to high	Moderate to low	Low
Miscellaneous	silts exposed on dry surfaces esp. along lake margins sometimes with *Phipsia algida*			some frost boils
Lakes and ponds	none to irregular ponds with incomplete drainage	generally few ponds except on shelves	generally few ponds except on shelves	numerous ponds in low center polygon centers; steep sides; peat bottoms
Soil great groups	Aquorthel; Aquiturbel	Aquorthel; Aquiturbel	Historthel; Aquiturbel	Sapristel; Hemistel; Historthel
[14]C ages	<50	50–300	300–2000	2000–5500

is associated with peat accumulation in drained lake basins and reactivation of ice-wedge cracking and growth (Mackay, 1992).

The objectives of this study are threefold: (1) to determine the spatial extent of the thaw lakes and drained lake basins on the Barrow Peninsula, (2) to present initial results of radiocarbon dating of basal organic materials to establish basin ages, and (3) to correlate organic layer thickness and SOC content with basin age.

Site Description

LOCATION AND CLIMATE

The Barrow Peninsula is the northernmost region in the United States. The area is underlain by continuous permafrost to a thickness of >400 m, with a maximum seasonal thaw (active layer) depth ranging from 30 to 90 cm in thickness (Nelson et al., 1998; Hinkel and Nelson, 2003). The Barrow Peninsula, extending northward from the Inaru River where it empties into Admiralty Bay at ~71°N, has low relief with elevations ranging from 0 to 22 m a.s.l. About 65% of the surface is covered by polygonal ground (Brown, 1967). Soil cores taken from around Barrow indicate that the volume of ice in pores and ice lenses/veins averages 50 to 75% in the upper 2 m (Sellmann et al., 1975), and ice wedges may contribute an additional 10 to 20%. The soil parent materials are unconsolidated sediments of the Late Pleistocene Gubik Formation.

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 at −26.6°C (NCDC, 2002). Mean annual precipitation is 106 mm, 63% of which falls as rain during the period

July through September. The winter snowpack averages 20 to 40 cm, but snow accumulation is highly variable because of variations in terrain roughness and drifting from strong easterly winds. The climate becomes more continental inland, with warmer summer temperatures (Clebsch and Shanks, 1968; Haugen and Brown, 1980).

VEGETATION

The succession of vegetation in basins following lake drainage follows a predictable sequence. Succession patterns, combined with the degree of ice-wedge polygonization and the surface hydrology, have been used to develop a relative basin-age classification scheme (Bliss and Peterson, 1992; Eisner and Peterson, 1998a). Known henceforth as the "field-based" classification scheme, the 4 classes are young, medium, old, and ancient basins. Characteristics are described in Table 1, illustrated in Figure 2, and summarized below.

In general, the most recently drained basins display early successional assemblages. The grass *Arctophila fulva* is found on all but the driest surfaces, and *Dupontia fischeri* and *Eriophorum scheuchzeri* (sedge) are also common. *Carex aquatilis* (sedge) clones with diameters of <4 m are infrequent, and no *Sphagnum* mosses or lichens are evident. Dry surfaces, especially along the old lake margins, are often bare, silt-enriched sediments where *Phippsia algida*, a vascular plant well adapted to soil movement and needle-ice formation (Bliss and Peterson, 1992), is typical. Plants in young basins have a high relative vigor. There are few lakes and ponds, and there has been insufficient time for development of patterned ground or soil formation. Soil great groups include Aquorthels and Aquiturbels.

Over time, surface organic matter accumulates and basins generally become drier through surface heave and enhanced evapotranspiration. The development of ice wedges in permafrost differentially
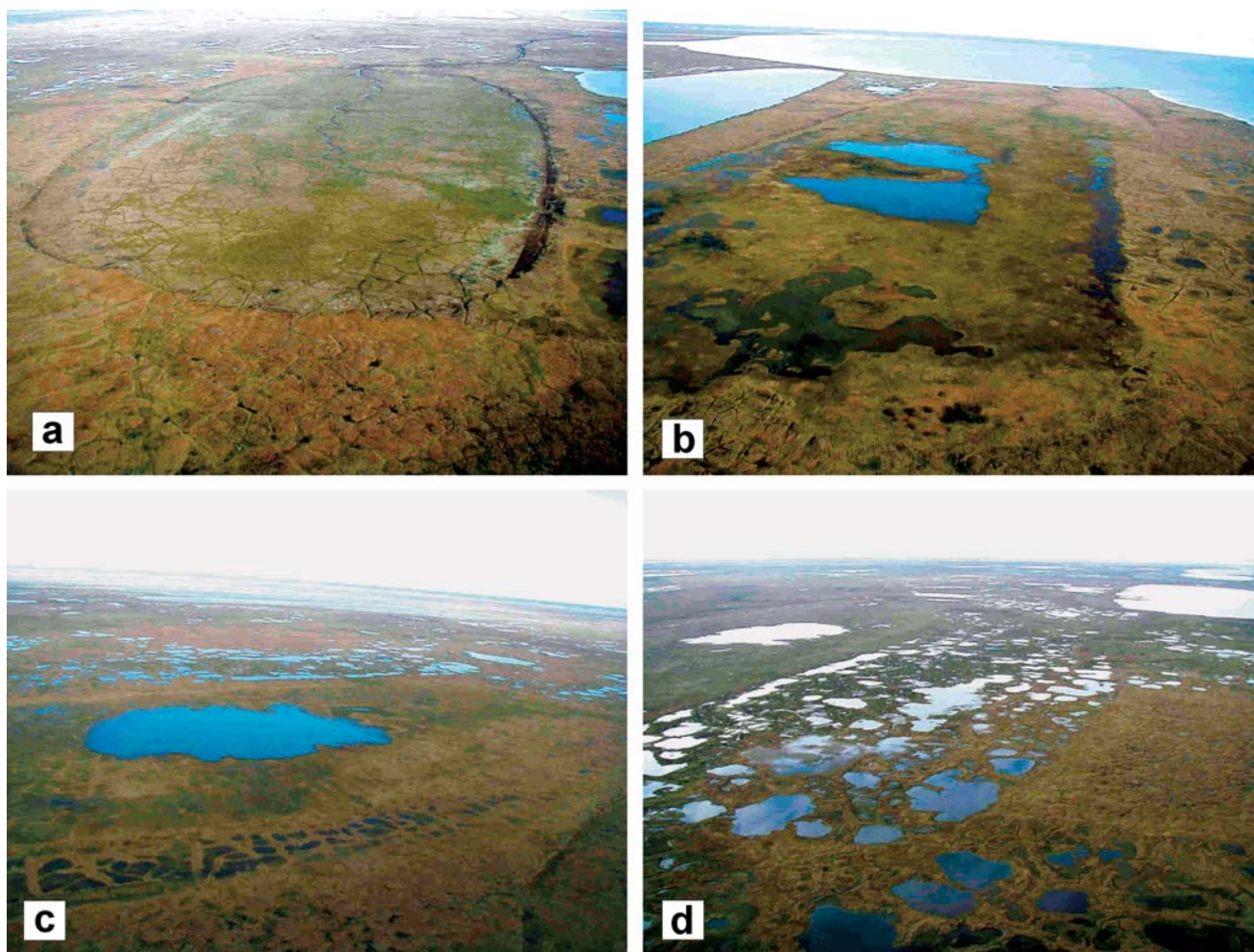
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FIGURE 2. *Oblique photographs of basins classified by relative age: (a) young, (b) medium, (c) old, and (d) ancient basin.*

uplifts the ground surface. Indistinct, flat-centered polygons occur in basins of medium age. The most characteristic vegetation is clones of the sedge *Eriophorum angustifolium* and *Carex aquatilis*, which range between 4 and 20 m in diameter. Otherwise, the vegetation is comparable to young basins with *Arctophila fulva*, *Dupontia fischeri*, and *Eriophorum scheuchzeri* common. Relative plant vigor ranges from medium to high. As with young basins, there are few ponds and lakes in medium-aged basins except on basin shelves, and soils groups are comparable to those in young basins.

Old basins contain flat-center or ponded, low-center polygons with rims ranging between 0.25 and 0.4 m in height. There is continuous *Carex aquatilis* within polygon centers with some *Sphagnum*. *Arctophila fulva* is restricted to ponds, and mosses (*Dicranum elongatum*) and lichens are prevalent on polygon rims. There is considerable standing dead material, especially on polygon rims. Plant vigor is moderate to low. Except on shelves, there generally are few ponds in old basins. Soil great groups include Aquiturbels and Historthels.

Ancient basins contain ponded low-center polygons that may be coalescing into ponds and small lakes. Polygon rims range from 0.4 to 0.75 m in height and support upland vegetation, including *Cassiope tetragona* (heath) and *Poa* spp. (grass). *Carex aquatilis* occupies the low-center polygons. There is considerable standing dead plant material in ancient basins, and plant vigor is low. Frost boils may occur, and organic soils (Sapristels, Hemistels) and Historthels dominate these ancient basins.

The field-based classification scheme is limited to those sites that are easily accessible. Clearly, if this scheme can be converted to the digital realm using an automated classification algorithm, it can be applied to large areas. Underlying this approach is the assumption that the relative field classification scheme is valid, i.e., that the assumed vegetation succession is uniform over time and space. This assumption can be verified by collecting samples from the lacustrine sediment-organic mat interface and subjecting these samples to radiometric dating. In this way, the relative age classification scheme can be bracketed by absolute dates.

Image Analysis

As an initial step, Landsat-7+ imagery from 30 August 2000 was acquired for the Barrow Peninsula. The image was georectified and checked for georegistration using GPS coordinates collected in the field. The 30-m-resolution data were subject to a pan-merging technique to utilize the higher-resolution 15-m panchromatic data; this involves pixel replication to a nominal resolution of 15 m and substitution of the pan band for the near-IR band in the data set. The method allows for high-resolution color display composites but does not corrupt the 30-m data. The resulting image is shown in Figure 1.

The scene was cropped to include only the region north of the mouth of the Inaru River (latitude 70.95°). Lakes south of the Inaru are smaller, and may have developed under different conditions. Sampling

TABLE 2

Descriptive statistics of lakes and thaw-lake basins

	Number	Area (km^2)	% land	Avg size (ha)	Median (ha)	Min (ha)	Max (ha)
Total study area	3191	1572	100				
Lakes (>1.0 ha)	592	345	22	66.30	5.35	1.07	3504.08
Ponds (<1.0 ha)	2599	3.2	<0.1	0.14	0.06	0.017	0.98
Basins	558	792	50	141.90	71.19	1.40	1282.30
Non-basin	—	435	28				

of drained lake basins in 2000 and 2001 was restricted to the peninsula north of 70.95°, defining the southern limit of the study area.

SPATIAL PROPERTIES

A mask was applied to the image to identify all water bodies exceeding 1 pixel in size. The resulting raster data set was vectorized to produce lake polygons, and converted to a shapefile for import into ArcView™. In addition, drained lake basins in the scene were hand-digitized to form a second polygon data file. Although care was exercised to follow basin boundaries, hand-digitizing produces polygons with smoother boundaries. Only larger, obvious basins were digitized from the image. Many older drained basins become obscured with time and are not easy to distinguish. Thus, the data set of 558 drained lake basins is very conservative with respect to the actual number of basins in the study area.

About 1572 km^2 of land is displayed in the study area. Of this, 3191 lakes cover about 348 km^2, or 22% of the land area; this proportion is in agreement with the percentage of lake coverage estimated by Sellmann et al. (1975) using ERTS-1 (LANDSAT) imagery of the area. By contrast, drained basins account for *at least* 50% of the land area, or more than twice the area currently occupied by water bodies. Clearly, nearly the entire landscape is affected by thaw-lake processes. The remaining 28% is classified as "non-basin." In reality, it represents a palimpsest that has been affected by the thaw-lake cycle, but the basins are sufficiently aged and overlapping that individual basins can no longer be discerned as single entities.

In this discussion, water bodies smaller than 1 ha (n = 2599) are eliminated from further consideration; though numerous, they constitute less than 1% of the total areal water coverage and can more properly be considered ponds. The remaining 592 lakes have the statistical characteristics shown in Table 2. Lakes occupying an area of 10 ha or less constitute nearly 400 of the 592 lakes, and frequency decreases rapidly with size (Figure 3). A total of 62 lakes (10%) exceed 100 ha (1 km^2) in areal extent but account for 82% of the total lake coverage. Only 14 lakes have an area exceeding 600 ha, with a maximum size of 3500 ha. By comparison, about 40% (224) of the 558 basins exceed 100 ha (1 km^2) in area and account for 82% of the area covered by basins (Fig. 3). There are 21 basins exceeding 600 ha in size, accounting for 23% of the total basin area. Both lakes and basins were analyzed for mean size (ha), and length and orientation of the longest (major) axis.

CLASSIFICATION OF DRAINED THAW-LAKE BASINS

The Landsat-7+ imagery was also used to develop a multilayered, hierarchical classification scheme for the detection and age-classification of basins. Relative basin age can be distinguished using a number of parameters, including vegetation type, degree of ponding and polygonization, and basin wetness and texture (Table 1). In order to incorporate all of these parameters into an image-based classification scheme, a neural network classifier was used (Frohn et al., 2001). The hyperspectral signatures of 42 field-classified basins were obtained

as training samples. Spectral and textural data transformations were utilized with pattern recognition algorithms to enhance basin features. To test the accuracy of these algorithms and validate the classification scheme, an additional 35 thaw-lake basins, evenly distributed between the 4 age categories and across the entire study area, were field classified. The neural network classifier algorithms produced the correct classification for 71% (25 of 35) of the basins. All of the 10 erroneously classified basins deviated by only one category. The algorithms had the greatest difficulty discriminating medium-aged basins from old and young basins. Thus, we feel confident that the spectral classification scheme provides a reasonable extension of the field-based classification scheme. Results of application of the automated scheme to the study area are presented in the first four columns of Table 3.

Field Sampling, Laboratory Methods, and Radiocarbon Dating

Of the 77 drained thaw lake basins visited for field classification and verification, soil samples were collected from 39 selected basins—approximately evenly distributed by the 4 age categories. Winter coring was conducted primarily in wetter basins near Barrow. In April 2001, 37 cores were collected from 11 basins within a 16-km radius of Barrow. Core sampling was along transects perpendicular to the major basin axis. Site selection was guided by use of ground-penetrating radar (GPR) profiles made along these transects, which helped avoid ice wedges and identify locations with deep organic layers (Hinkel et al., 2001). Cores were taken using a Big Beaver™ earth drill equipped with a 7.5-cm-diameter SIPRE core barrel, mounted on a sledge and pulled by a snow machine. The summer (August 2000 and 2001) soil-sampling program entailed digging soil pits and collecting cores in an additional 28 drained thaw-lake basins accessed by helicopter. In general, 3–5 cores were collected from each of the 39 basins; average core length was 1.3 m.
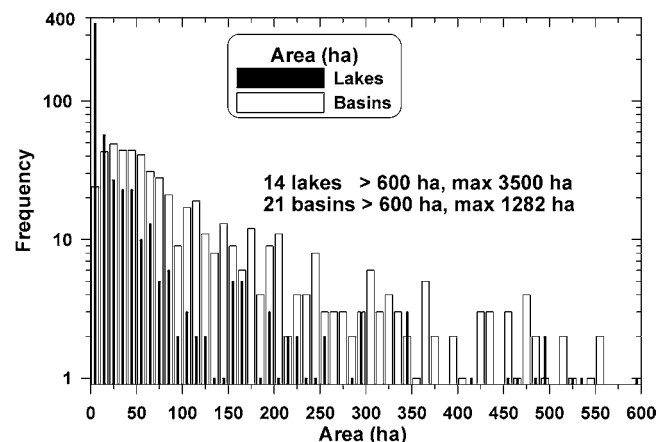

FIGURE 3. *Histogram of area (ha) for lakes and basins in study area.*

TABLE 3

Results from spectral-based classification scheme and basin coring averages

Spectral classification of basins	Number of basins	Area (km²)	% land	Average thickness of organic layer (cm)	Average SOC in organic layer (kg m⁻²)	Average SOC from 0–100 cm (kg m⁻³)	Spatially weighted average SOC, 0–100 cm (kg m⁻³)
Young	88	65	4.1	7	2.5	43.6	1.80
Medium	83	145	9.2	13	5.6	36.2	3.34
Old	305	478	30.4	20	8.0	51.6	15.69
Ancient	82	104	6.6	35	8.1	41.8	2.76
Total basins	558	792	50	—	—	—	—
Total non-basin	—	435	28	5	3.2	57.5	16.04
Total basins + non-basin	—	1227	78	—	—	—	39.64

Drilled cores were described according to soil horizon, and the amount of segregation ice was estimated visually. Cores were then sliced into 10-cm-thick sections for determination of moisture content and bulk density. Samples were also collected from all drilled cores for carbon and pollen analysis. Additional samples for radiocarbon dating were taken at the interface of the lacustrine sediment and the in situ peat, which represents the point in time of lake drainage and revegetation of the basin surface. The interface sample was sieved, and terrestrial plant material, mostly sedge or *Sphagnum*, was selected and pretreated with a weak acid/base wash. These discrete, identifiable biological remnants are not subject to contamination from roots or dissolved humic acids, which can render bulk sediment dating problematic. The material was sent to Lawrence Livermore National Laboratory to undergo radiocarbon Accelerator Mass Spectrometry (AMS) dating. This method is based on the direct detection of ^{14}C in the sample, yielding highly accurate dates with very small sample sizes.

For SOC analysis, core samples were dried at 70°C, and field moisture and bulk density were determined. Subsamples were passed through a 100-mesh (149 μm) sieve. Total carbon was measured for approximately 500 samples on a Dohrmann DC190 carbon analyzer.

FIGURE 4. *Average organic layer thickness (standard deviation ticks) and degree of organic decomposition for 39 basins sampled in study area.*

Samples did not react with 1 M HCl, indicating that there was no inorganic carbon. Thus, the SOC values represent total organic carbon.

Results

SOIL CLASSIFICATION AND CARBON POOLS

In general, the organic layer thickness and degree of organic matter decomposition, as reflected by the dominant kind of organic material, are positively correlated with relative field basin age (Tables 1 and 3, Figure 4). Peat accumulation rates range between 0.9 and 4.7 cm 100 yr⁻¹; these values are less than for temperate environments but are comparable to those reported for other cold regions (Everett, 1983). The rate of organic matter accumulation in arctic Alaska is retarded by the low net primary production due to cold temperatures (Shaver and Chapin, 1991).

The change in organic form reflects increased decomposition with time. All young lake basins analyzed have a fibric surface organic layer, in which the fiber content is 40% or more by volume after rubbing. Similarly, medium-aged and old basins generally contain organic layers that are fibric. In contrast, ancient basins contain either sapric or hemic materials that contain less than 17% or 17–40% fibers after rubbing, respectively.

The change in thickness of the surface organic layer is reflected in the classification of the soils according to *Soil Taxonomy* (Soil Survey Staff, 1999) (Table 1, Figure 4). Young and medium-aged basins contain a <10-cm and 10–15-cm organic layer, respectively, and are classified as Aquorthels and Aquiturbels. Old basins have a surface organic layer ranging between 15 and 30 cm in thickness and contain Historthels, or Aquiturbels in actively cryoturbated areas. Ancient basins contain a surface organic layer ranging from 40 to 50 cm in thickness; Sapristels are common in these basins, along with Hemistels and Historthels. Although organic layer thickness and degree of decomposition do not always increase consistently with time, they are valuable indicators of relative basin age.

Traditionally, soil carbon density is calculated for the upper 100 cm (Post et al., 1982). As shown in Figure 5, there are no significant differences in SOC pools by age category, calculated for the upper 100 cm using equations given by Michaelson et al. (1996). This lack of consistency in SOC storage is likely due to variable amounts of recycled carbon in the lacustrine sediments below the surface organic layer. However, it also highlights a problem often encountered with frozen permafrost samples that has a direct impact on the calculation of SOC. In situ sediments become enriched in ice over time, especially in the near-surface permafrost, and this effect is detectable over decadal scales (Hinkel et al., 1996; Mackay and Burn, 2002). Ice enrichment results from the internal migration of free water and subsequent formation of ice lenses, and also from infiltration of meteoric water down thermal contraction microcracks. Over time, the ice content of
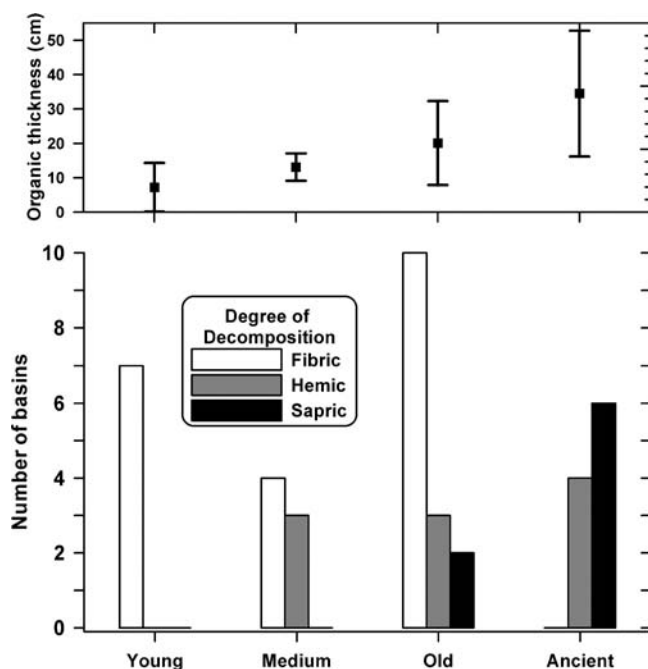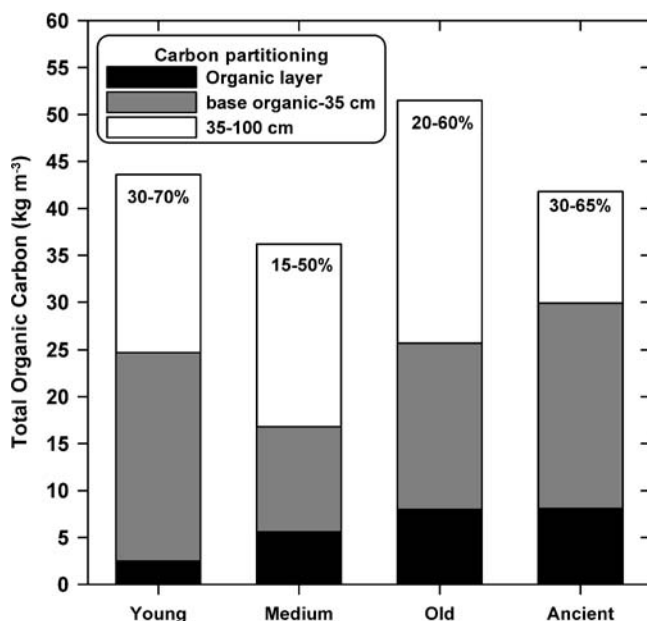
FIGURE 5. Soil organic carbon partitioned by stratum for 39 basins. Percentage values refer to the volumetric ice content in the upper permafrost.

CAMS number	Site	Organic depth (cm)	^{14}C Age	+/−	field class	^{14}C class
77861	VB-1	15	80	90	2	2
77726	VB-2	33	230	50	3	2
77863	VB-3	4	>modern		1	1
77730	VB-4	20	1150	50	3	3
77727	VB-5	22	4770	50	4	4
77866	VB-6	52	3660	60	3	4
77869	VB-7	35	2260	60	4	4
77870	VB-8	18	>modern		1	1
77871	VB-9	16	590	50	3	2
76368	VB-10	14	>modern		3	1
76371	VB-11	26	80	40	3	2
79457	VB-13	53	4520	50	4	4
82903	VB-15	35	5370	100	4	4
79460	VB-18	56	4090	40	4	4
79461	VB-19	49	2970	40	4	4
79463	VB-21	52	4970	160	3	4
79466	VB-24	4	>modern		1	1
79469	VB-25	52	3760	40	4	4
82904	VB-26	48	3230	190	4	4
82905	VB-27	70	1420	40	3	3
82910	VB-33	47	2075	40	4	4

1 = young: 0–50 yr.

2 = medium; 50–300 yr.

3 = old; 300–2000 yr.

4 = ancient; 2000–5500 yr.

the upper permafrost increases; this trend is apparent in Table 1. Cores collected in old and ancient basins often contain thick (>50 cm) lenses of segregation ice (Bockheim et al., 1999), which appear as strong reflection surfaces in GPR profiles (Hinkel et al., 2001). Application of the reporting standard, as SOC with units of kg m^{-3}, can therefore be misleading due to ice enrichment and consequent SOC dilution.

For this reason, SOC was determined separately for each stratigraphic layer, as shown in Figure 5. The surface organic layer represents the accumulation of plant detritus since thaw-lake drainage; SOC generally increases with basin age, and the trend is statistically significant. Owing primarily to cryoturbation, large amounts of SOC exist in the active layer below the organic mat. Approximating the active-layer thickness as 35 cm, we again note an overall increase in SOC with basin age in this stratum. Below the active layer, and extending to a depth of 100 cm, is the near-surface permafrost layer where SOC is sequestered. This layer is often enriched in ice, as indicated by the percentage ice content labels in Figure 5. Thus, there is no statistical trend in SOC with age for the active layer, the upper permafrost, or in the entire upper 100 cm (Statistical Analysis Systems, 1999). This example demonstrates the necessity of developing appropriate sampling protocol for permafrost soils affected by cryoturbation and ice enrichment.

An areally weighted estimate of the amount of near-surface terrestrial carbon was made using the data in Table 3. The basin age-specific estimate of SOC (kg m^{-3}) was weighted by the percentage of land area represented by that basin's spectral age group. Data from Bockheim et al. (1999) for non-basin terrestrial surfaces was also included; this constitutes 28% of the land surface. For the 78% of the land area not covered by lakes, there is a spatially weighted average of approximately 40 kg C m^{-3}. This average is less than the 62 kg C m^{-3} reported by Michaelson et al. (1996) and Bockheim et al. (1997) for the Arctic Coastal Plain. Results are not strictly comparable, however, because most of the samples in the cited studies come from the lower Kuparuk River. That area contains a preponderance of carbonaceous silts of eolian and fluvial origin, which greatly increase the amount of inorganic carbon. Since inorganic carbon is included in the calculation with organic carbon, the total soil carbon is quite high. For example, about 70% of the carbon in the upper meter is found below the active

layer, in carbonaceous mineral soils (Michaelson et al., 1996). By contrast, in addition to dilution of SOC by high amounts of ice, the Barrow region has minimal inorganic carbon but very high organic content in comparison to the Kuparuk. In any event, the SOC in drained thaw-lake basins exceeds that found in most life zones except temperate wetlands (Post et al., 1982).

BASIN CLASSIFICATION AND RADIOCARBON DATING

The results of the radiocarbon dating of 21 drained basins are given in Table 4. All of the organic samples were collected at the base of the organic layer just above the lacustrine sediments. Figure 6 shows the relationship between radiocarbon age and thickness of the surface organic layer for these 21 basins only, and identified by their field age classification. The results show a moderately strong relationship ($r^2 = 0.41$) between ^{14}C basin age and thickness of the surface layer, although the thickest organic layers are always found in the old and the ancient basins.

Certain qualifiers should be taken into account when comparing the radiocarbon dates with the field-based classification scheme. Typically 3–5 cores were collected from each basin, but only 1 per basin was subjected to ^{14}C analysis due to cost constraints. Although we attempted to select this sample from near the center of the basin to ensure that it represented the time of complete lake drainage, there is often considerable interbasin variability in the drainage history. Second, care was taken to ensure that the core sample had not been subject to cryoturbation, but it is sometimes not possible to identify these effects in a core. Finally, the organic sediments are usually extremely decomposed. Discrete plant macrofossils, particularly Sphagnum and sedge species, were therefore used, resulting in a tiny sample size extremely susceptible to contamination. We are in the process of identifying the most reliable plant species to use in our dating, but the choice is limited.
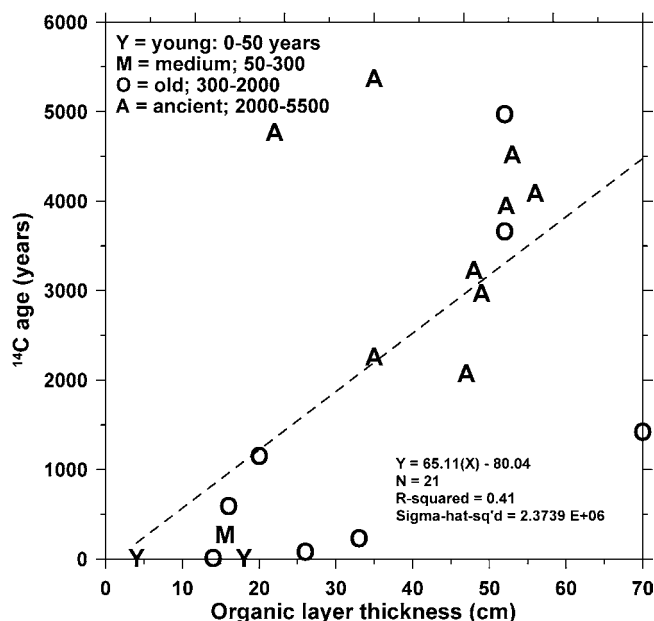
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FIGURE 6. *Correlation between ^{14}C and organic layer thickness, with field-classified basin age.*

Of the 21 dated basins, 4 provide younger-than-modern (post-1950) radiocarbon dates, indicating that the original lake probably drained less than 50 yr ago. These are field classified as young basins, with the exception of VB-10 (medium). Three basins are dated as having drained more than 50 yr ago but less than 230 ± 50 BP. Of these, only 1 is field classified as a medium basin, and the other 2 as old basins. Three thaw-lake basins, drained between 590 ± 50 BP and 1420 ± 40 BP, are field classified as old basins. Finally, 11 of the basins are dated as having drained between 2075 ± 40 and 5370 ± 100 BP. Nine of these are field classified as ancient basins and the other 2 as old basins. Based on the radiocarbon dates, we have tentatively drawn the following basin age boundaries for the Barrow Peninsula: young basins are 0 to 50 cal BP, medium basins 50 cal BP to 300 BP, old from 300 to 2000 BP, and ancient from 2000 to 5500 BP.

Aerial photographs of the Barrow region offer independent confirmation of our classification and dating methods for young basins (0–50 yr). U.S. Navy photographs taken in 1948 show several lakes that are now drained basins. One of these is VB-24, classified as a young basin by field, spectral, and ^{14}C methods. Measurement of the percentage delta ^{14}C in a sample permits tentative calibration of this date using a tropospheric bomb-curve (T. Brown, pers. comm.). The resulting calibration indicates that basin drainage occurred between 1959 and 1962 A.D.

The findings here are reported in radiocarbon years because calibration of late Holocene dates generates a series of probable dates, which, though more accurate, are somewhat cumbersome. Moreover, the resultant calendar-year ranges can also lead to problems with interpretation of the categories. For example, 2 of the basins that were field classified as old but that radiocarbon dates indicate are medium could be open to reinterpretation if their calibrated ranges are examined. VB-2 has a date of 230 ± 50 BP, clearly of medium age. The ranges of the most probable ($>.40$ relative probability) calibrated dates are 1636–1680 A.D. and 1737–1805 A.D., which would place the basin in the old category.

Discussion

Although some of the correlations between field class and radiocarbon dates are problematic, the field classification scheme ap-

pears to be a good indicator of basin age. Owing to cryoturbation and possible contamination, an insufficient number of medium basins are represented; this limitation was compounded by difficulty in accurately identifying medium-aged basins with the neural network classifier. Young and ancient categories are relatively easy to identify since they correspond to end points in the spectra of successional and geomorphic events. The transitional classes of medium and old basins are more difficult to discriminate. This may relate to the high initial rates of vegetation succession and landscape evolution, especially ice-wedge polygonization, early in the basin history (0–500 yr).

As noted earlier, only 3–5 cores were collected from each basin, and the core showing minimal cryoturbation effects was selected for dating analysis. Although this methodology maximizes the number of basins that can be sampled, it implicitly assumes that a single core adequately represents the entire basin. To test this assumption, an intensive basin sampling program was implemented during the second field season in April 2002. Twelve cores were collected from each of 4 basins near Barrow, with 1 basin per age class. These cores are currently being analyzed for organic layer thickness, SOC, pollen content, and drainage date to determine the degree of spatial homogeneity within individual basins (Bockheim et al., in press).

The climate and vegetation history of the Barrow region is poorly understood. Research has been hampered by a shortage of continuous records of lake sediments for paleoecological reconstructions and the low organic content of the drained basins, making chronological control problematic. Much of the earlier radiocarbon dating of thaw lake sediments was carried out on bulk dating of buried organic material excavated from coastal or riverine exposures (Hopkins and Kidd, 1988). Carson (1968, 2001) carried out extensive ^{14}C dating on the strand lines of 4 drained basins near Barrow. His results identified a series of thaw lake cycles that began after widespread lake drainage, interpreted as a response to climatic cooling, occurred 3500 to 4000 yr ago. Our results, based on the 21 dated basins, indicate that the thaw-lake cycle has been operating on the Barrow Peninsula for at least 5500 yr and is currently active. Although sampling was stratified based on the field-classified age, there are no obvious temporal clusters in Figure 6; indeed, drainage events appear uniformly distributed. This implies that basin drainage has not been controlled by climate but occurs at a fairly uniform rate. Further sampling is required to resolve the role of climate forcing of the thaw-lake cycle.

The entire thaw-lake cycle was at one time hypothesized to have a duration of 2000 to 3000 yr (Carson, 1968; Hopkins and Kidd, 1988), but this study indicates that several basins drained prior to 5000 BP and do not appear to be redeveloping as lakes. Conversely, ponds and polygons in some ancient basins are coalescing, and this process may culminate in the reestablishment of a lake. The primary factor encouraging the redevelopment of a lake in the terrestrial basin may be hydrological forcing, not vegetation succession and polygonal development. This forcing could result, for example, from gradient-controlled headward erosion of streams, flooding induced by stream piracy, or changes in the precipitation-evaporation regimes. A better understanding of these forces may be revealed through pollen and microfossil analysis of terrestrial peat (e.g., Eisner and Peterson, 1998b), which could indicate whether gradual or abrupt hydrological changes dominate in the record.

Collection of deep (>2 m) cores from thaw-lake basins does not provide evidence for repeated thaw-lake cycles. Stratigraphic evidence might include a packet of lacustrine sediments overlain by terrestrial peat at depth, with this sequence overlain by a similar packet sequence upsection. Early in this study, we attempted to date buried peat material from deeper in the cores, with the assumption that buried organics represent a previous cycle of drained and revegetation; the results were inconclusive. Of the 37 cores drilled in transects in 11 basins, the sediments underlying the terrestrial peat were permeated with ice

lenses and veins and cryoturbated, making identification of multiple cycles difficult.

In an effort to determine the overall age of the landscape in which thaw lakes have formed and drained, we have visited "non-basin" sites in the study area that appear to have been unaffected by thaw-lake processes. These sites occupy higher elevations, have thick organic layers with very high ice content in the upper permafrost, and have plant communities atypical of the Barrow area. Deep cores were collected in summer that did not penetrate to the base of the organic layer due to the presence of massive segregation ice. Radiometric dating of the lowermost organic sediments yielded a ^{14}C date of 8390 ± 70 BP. Although these sites requires further study, it appears that they are a remnant of an older erosion surface that escaped the effects of the thaw-lake cycle (Eisner et al., in review).

Conclusions

In the Barrow Peninsula north of ~71° latitude, several general conclusions can be made:

1. about 22% of the land area is covered with lakes, and about 50% is scarred by basins formerly occupied by lakes;
2. the spectral classification algorithm produces viable estimates of field-determined basin age. Improvements can be achieved by increasing the number of training samples, especially with medium-aged basins, to improve interclass discrimination;
3. in general, organic layer thickness and the degree of organic matter decomposition increase with basin age;
4. SOC tends to increase with basin age, but the relation becomes obscured due to cryoturbation, organic matter decomposition, and processes leading to ice enrichment in the upper permafrost;
5. spatially weighted estimates of SOC indicate that there are 40 kg C m^{-3} in the 78% of the study area not covered by lakes. This is in contrast to estimates of 62 kg C m^{-3} calculated for soils rich in inorganic carbon (Michaelson et al., 1996; Bockheim et al., 1997). The differences are attributable to the presence of inorganic C in drained thaw-lake sediments of the eastern Arctic Coastal Plain;
6. radiocarbon dating shows a moderate correlation between field-based basin age classification and ^{14}C age. This correlation implies that the factors used to discriminate basin age (Table 1) are valid. Young basins have formed from lakes drained in the past 50 yr, medium basins formed 50–300 BP, old basins 300–2000 BP, and ancient basins 2000–5500 BP;
7. further refinement of the spectral classification is warranted, along with implementation of intensive basin sampling to ascertain basin homogeneity; and
8. the specific age of the basin is a significant but not dominant factor when evaluating the processes involved in carbon and peat accumulation and decomposition rates in these basins. Since time is not the only factor, we must then look to other agents such as surface hydrology, slope, and proximity to coasts, streams, and other lakes.

Acknowledgments

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