GEOMORPHOLOGY OF THE NORTHEAST PLANNING AREA OF THE NATIONAL PETROLEUM RESERVE–ALASKA, 2001

FIRST ANNUAL REPORT

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EXECUTIVE SUMMARY

Permafrost development on the Arctic Coastal Plain in northern Alaska greatly affects the distribution of ground ice, engineering properties of the soil, ecological conditions at the ground surface, thaw-lake formation, and response of the terrain to human activities. Of particular interest for assessing potential impacts from oil development in the National Petroleum Reserve–Alaska (NPRA) are the identification of terrain relationships for predicting the nature and distribution of ground ice across the landscape and the evaluation of disturbance effects on permafrost. Accordingly, this study was designed to determine the nature and abundance of ground ice at multiple spatial scales to develop terrain relationships for predicting ice distribution, to assess the rates of landscape change and develop a conceptual model of how ground ice changes during the evolution of the landscape, and to estimate the amount of thaw settlement likely to occur after disturbance. The study focuses primarily on permafrost dynamics associated with thaw lake development because lacustrine processes, as opposed to eolian, fluvial or marine processes, are currently the dominant geomorphic processes in the study area.

NATURE AND DISTRIBUTION OF GROUND ICE

Classification of materials near the ground surface (<3 m) was done at multiple spatial scales, and included ice structures (microstructures caused by ice segregation), lithofacies (textural-structural assemblages in a stratigraphic profile), and terrain units (depositional and morphological units across the landscape). Eight primary ice structures were identified, including pore, lenticular, vein, layered, reticulate, ataxitic, organic-matrix, and solid ice. Soil materials were classified into 11 lithofacies, the most common of which included: (1) massive organics in the active layer of all terrain types except young eolian sand deposits, (2) massive turbated fines with organics just below the active layer, and (3) massive sands found at 1–3 meters depth at all coring locations. A total of 24 terrestrial terrain units and 7 waterbodies classes were identified, including two colluvial, three eolian, 13 fluvial, five lacustrine, and one alluvial-marine deposits.

When comparing relationships among ice structures, lithofacies, and terrain units using data from 15 cores (2–3 m), there were strong relationships that can be used to partition the variability in distribution of ice across the landscape. Large differences were found in the frequency of occurrence of the various ice structures among lithofacies: pore ice was nearly always associated with massive, inclined, and layered sands; lenticular and ataxitic ice were most frequently associated with massive and layered fines, organic matrix ice was usually found in massive and layered organics and limnic fines. Reticulate ice was broadly distributed among fine and organic lithofacies.

Important differences were observed in the volume of segregated ice among ice structures, lithofacies, and surface terrain units. Among ice structures, mean ice volumes were highest for layered ice (81%) and organic-matrix ice (77%), intermediate for ataxitic ice (72%), reticulate ice (69%), and veined ice (68%), and lowest for lenticular ice (58%) and pore ice (43%). Among lithofacies, mean ice volumes were highest in massive organics (82%), intermediate in massive fines (78%), fines with organics (73%) and layered organics (68%), and lowest in massive (41%) and inclined sands (46%). Among surface terrain units, mean ice volumes were highest in alluvial-marine deposits (71%), intermediate in ice-rich thaw basin margins (64%) and ice-rich thaw basin centers (64%), and lowest in ice-poor thaw basin margins (60%) and eolian sands (54%).

LANDSCAPE CHANGE AND THAW LAKE DEVELOPMENT

Photogrammetric analysis of waterbody changes in three small study areas was used to evaluate shoreline erosion from 1945–1955 to 2001. Overall, 0.74% of the total land area was lost to shoreline erosion during the period (46–56 yr) for which data were available. The average annual erosion rate for the three study areas was 0.04% of the total land area. Average rates of shoreline retreat were very slow (0.02 m/yr), even for large, deep lakes (0.08 m/yr). The maximum observed rate of shoreline retreat was 0.8 m/yr. Deep lakes (mean 26.1% of area) were much more prevalent

than shallow lakes (mean 6.6%), and shallow lakes typically are small (<5 ha).

Our examination of the development of thaw lakes, based on topographic profiles, soil and ground ice surveys, stratigraphic analysis, radiocarbon dating, photogrammetric analysis, and regional comparisons, revealed that lake evolution on the coastal plain was much more complex and less cyclic than suggested by previous investigations. We developed a revised conceptual model for the portion of the Arctic Coastal Plain underlain by extensive sand sheets that includes: (1) initial lakes formation, (2) lateral lake expansion accompanied by mineral and organic sediment accumulation and redistribution, (3) lake drainage, (4) ice aggradation in exposed sediments, (5) secondary development of thaw lakes, and (6) basin stabilization.

THERMOKARST POTENTIAL AND TERRAIN SENSITIVITY

We developed estimates of potential thaw settlement following disturbance in various terrain units, based on (1) volumes of excess segregated ice, (2) volumes of wedge ice, and (3) the equilibrium active layer depth following surface disturbance (active-layer readjustment). Based on an active-layer readjustment to 0.8 m, which is typical of non-flooded highly disturbed surfaces, mean (± SD) thaw settlement is expected to be 0.61 ± 0.32 m for coastal plain, 0.30 ± 0.14 m for ice-rich thaw basin centers, 0.37 ± 0.26 m for ice-rich thaw basin margins, 0.12 ± 0.05 for ice-poor thaw basin margins, and 0.11 ± 0.07 m for eolian sand. Based on data from the literature and field observations in other areas of Arctic Coastal Plain, we estimate that all wedge ice in the top 2 m would be lost, leaving a highly polygonized surface at disturbed sites. The loss of volume due to thawing of ice wedges will be about 40% for old alluvial-marine deposits, 15% for ice-rich thaw basins with well-developed low-centered polygons, and negligible for ice-poor thaw basins.

Based on the differences in the potential for thaw settlement among terrain units, we developed a conceptual model of terrain responses to severe disturbances associated with scraping of the surface or complete removal of the vegetative cover. For example, on coastal plain deposits where the volumes of segregated ice and wedge ice near the surface are high, thermokarst will create highly irregular micro-topography with deep and shallow troughs, and prominent high-centered polygons above the water-filled troughs. In contrast, in ice-poor thaw basins, where segregated ice volumes are much lower and wedge ice is negligible, thermokarst will create uniform, very shallow ponds as the basins settle.

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INTRODUCTION

Permafrost development on the Arctic Coastal Plain in northern Alaska greatly affects the ecological conditions at the ground surface (Billings and Peterson 1980, Webber et al. 1980, Walker 1981), the engineering properties of the soil (Johnson 1981, Kreig and Reger 1982, McFadden and Bennet 1991), and the response of the terrain to human activities (Brown and Grave 1979, Webber and Ives 1978, Lawson 1986). Of particular interest for assessing potential impacts from oil development in the National Petroleum Reserve–Alaska (NPRA) are the identification of terrain relationships for predicting the nature and distribution of ground ice across the landscape, and the evaluation of how permafrost will respond to disturbance. Accordingly, this study was designed to determine the nature and abundance of ground ice at multiple spatial scales to develop terrain relationships for predicting ice distribution, to assess the rates of landscape change, and develop a conceptual model of how ground ice changes affects the evolution of landscape, and to estimate the amount of thaw settlement likely to occur after disturbance as a measure of terrain sensitivity. The study focuses on permafrost modification associated with thaw lake development on the coastal plain portion of the NPRA Development Project Area because lacustrine processes, as opposed to fluvial or marine processes are the dominant geomorphic processes in the study area.

While there is widespread recognition that the soils of the Arctic Coastal Plain have high ice contents, little information is available on the nature, distribution, and dynamics of ground ice. Kreig and Reger (1982) provided a comprehensive analysis of relationships between landforms and soil properties, but their data do not allow the prediction of ice distribution across the coastal plain. At Barrow, the volume of segregated ice in the surface soils often approached 80% (Brown 1968) and overall ice contents associated with segregated ice averaged 62% (Hinkel et al. 1996), but the relation of ice contents to terrain characteristics were not examined. On the Mackenzie Delta, pore and segregated ice constituted 80% and wedge ice 12–16% of total ice volume (Pollard and French 1980), and the structure of the ice has been well documented (Murton and French 1994). Analyses of the nature and distribution of ground ice on the Colville Delta showed that ice content was strongly related to ice structure, soil texture, and terrain unit (Jorgenson et al. 1997, Jorgenson et al. 1998, Jorgenson and Shur 1998). The mean total volume of segregated ice for the most ice-rich terrain unit, abandoned floodplains, was 79%. In a similar study of ground ice patterns and thermokarst potential in the Prudhoe Bay and Kuparuk oilfields, the mean volume of segregated ice in alluvial plain deposits was 76% (Burgess et al. 1998). The present study contributes to this growing body of information by quantifying differences in the structure and abundance of ground ice among terrain units on the coastal plain.

One of the most striking features of the Arctic Coastal Plain is the abundance and recurring oriented pattern of the lakes and drained basins. The abundance of thaw lakes and drained basins has been attributed to a cycle of repeated lake drainage followed by regeneration of ice-rich permafrost (Britton 1957, Tedrow 1969, Billings and Peterson 1980). Everett (1980) conceptualized six stages of thaw lake development that included: (1) original ice-rich elevated conditions, (2) initial thermokarst development along ice wedges, (3) coalescence into small ponds, (4) enlargement into large deep lakes, (5) complete or partial drainage resulting in flat basins, and (6) aggradation of ice, primarily through development of ice wedge polygons. In this study we re-examine and revise this conceptual model by determining rates of thaw lake enlargement, evaluating ice content at different stages of the cycle, and defining the process of ice aggradation in the final stages of the process.

While naturally occurring thermokarst is fundamental to ecological processes on the thaw lake plains of the arctic lowlands (Britton 1957, Billings and Peterson 1980, Walker et al. 1980, Carter et al. 1987), human-induced thermokarst is a concern for land development in the arctic because of the subsequent changes in hydrology, soils, and vegetation (Brown and Grave 1979, Jorgenson 1986, Lawson 1986, Walker et al. 1987). Of specific concern for oil development is the possibility of thermokarst as a result of off-road and seismic trail disturbances (Walker et al. 1987, Emers and Jorgenson 1997), alteration of drainage

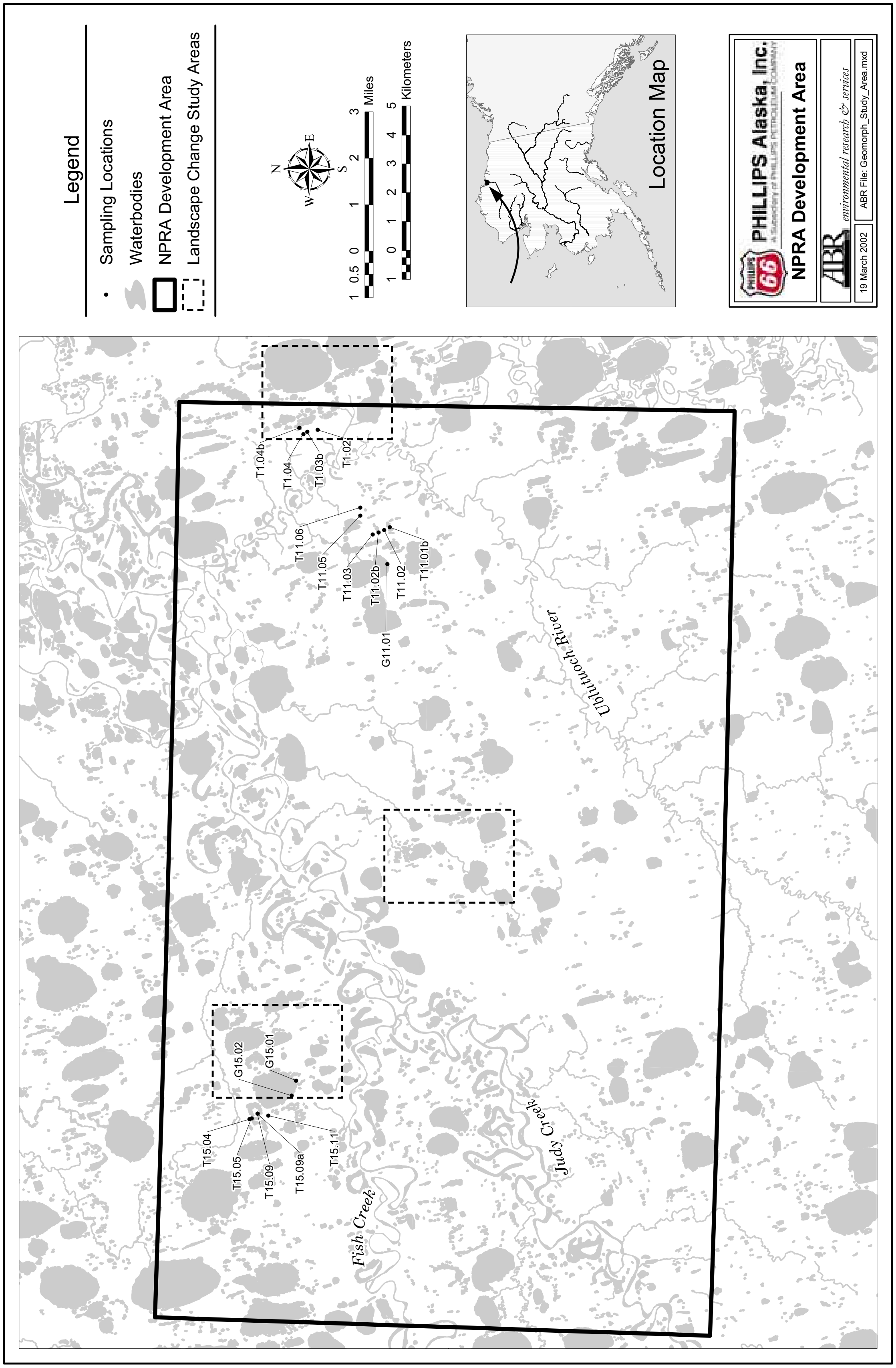


Figure 1. Map of soil sampling locations in the NPRA Study Area, northern Alaska, 2001.

patterns following road development, oil spill cleanups (Jorgenson et al. 1991, Jorgenson et al. 1992), closeout and rehabilitation of reserve pits (Burgess et al. 1999), and gravel removal after site abandonment (Jorgenson and Kidd 1991, Kidd et al. 1997). Thus, to develop land rehabilitation strategies that are site-specific and appropriate to the rapidly changing environmental conditions associated with thermokarst, it is essential to understand the nature and abundance of ground ice across the oilfields, and to relate ice characteristics to terrain characteristics at a level useful for land management. This study contributes to a better understanding of how to manage human activities on permafrost terrain by providing estimates of potential thaw settlement for the dominant terrain units and by outlining a conceptual model of terrain responses to disturbance.

To provide a framework for sampling and analyzing the spatial variability of ground ice across the landscape, and to identify the geographic scales most useful for interpretation and management, we used a hierarchical approach that incorporated regional, landscape, and local ecosystem scales. At the regional level, we used ecodistricts and ecosubdistricts (associations of terrain units and geomorphic processes) developed by Jorgenson et al. (1997) as the basis for allocating our sampling effort. At the landscape scale, we used terrain units (depositional units related to a specific geomorphic process) to stratify our sampling and as the basis for analysis. At the local ecosystem level (areas with relatively uniform soil stratigraphy and vegetation), we classified our samples (soil cores) by microscale characteristics such as ice structures and lithofacies (texture and structure) for use in analysis.

Specific objectives of this study were to:

1. determine the nature and abundance of ground ice at multiple spatial scales, in order to correlate terrain units with associated ice characteristics,
2. quantify the rates of shoreline changes of thaw lakes and develop a conceptual model of the patterns and processes involved in thaw lake development, and to
3. estimate the potential thaw settlement of the various terrain units and develop a conceptual model of thermokarst development after disturbance.

METHODS

NATURE AND DISTRIBUTION OF GROUND ICE

FIELD SURVEYS

Field surveys were conducted during 3–15 August 2001. Data on soil stratigraphy were collected along transects (toposequence) within each of three predominant surficial geology deposits in the NPRA Study Area (Carter and Galloway 1985): alluvial-marine, alluvial terrace, and eolian sand. Along each transect, 3–5 soil cores were collected (15 cores total) in terrain units that covered the entire gradient of landscape development from young thaw basins to the oldest and highest surfaces (Figure 1). The stratigraphy of the near-surface soil (i.e., the active layer) was described from soil pits to assess the depth of thaw, surface organic thickness, and mineral characteristics. For sampling frozen soils below the active layer, a 3-in.-diameter SIPRE corer with a portable power head was used to obtain 1–2.7 m cores. Three profiles also were described from bank exposures after unfrozen material was removed with a shovel to expose undisturbed frozen sediments. Descriptions for each profile included the texture of each horizon, the depth of organic matter, depth of thaw, and visible ice volume and structure. In the field, soil texture was classified according to the Soil Conservation Service system (SSDS 1993).

Soil samples were taken every 20–30 cm along each of the 15 core sections, within soil horizon boundaries (total of 104 samples). Samples were analyzed for volumetric and gravimetric water content, electrical conductivity and pH. Soil volume was determined by measuring core sample length at three points and circumference. EC was measured using an Orion model 290 EC meter and pH was determined using an Orion model 290 pH meter. Additional analyses for particle size, and total carbon were performed on a subset of soil samples by the Palmer Research Station Soils Laboratory (Palmer, AK).

To establish minimum ages for the older stratigraphic terrain units, one sample of basal organic material (sedge peat) was collected from each of six cores. Laboratory analyses were performed by Beta Analytic, Inc. (Coral Gables, FL). Dates in this report are presented as calibrated calendar ages before present (1950) and include the range associated with the 2 sigma error. (Stuiver et al. 1998).

CLASSIFICATION AND MAPPING

The microscale soil textures and structures described from the field profiles were grouped into lithofacies (distinctive suites of sedimentary structures related to a particular depositional environment), and cryostructures (repeating patterns of ice distribution). Lithofacies and cryostructures of the cores were classified in the office using field descriptions and photographs of cores.

Lithofacies were classified according to systems of facies analysis for fluvial deposits developed by Miall (1978, 1985) and Brierley (1991). We added several new classes to these systems to incorporate features specific to the permafrost environment.

Ice structure (form, distribution, and volume of ice) was classified in the field in 2001 following a version of the system developed by Murton and French (1994), modified to better differentiate the structures that we observed (Jorgenson et al. 1997). This classification system was developed during similar work in the Colville River Delta and Kuparuk River Oilfield during 1995-1998.

During classification, primary ice structures were further subdivided by secondary characteristics such as shape and size. Frequently, ice structures occurred in assemblages in which the individual structures were too small to differentiate; these were classified as composite structures. In total, 52 structures and composite structures were identified in the field. This number was too large for practical application, so we aggregated classes through a two-step process. First, we aggregated the non-composite structures into 8 simple primary structure types. Second, the composite types were grouped according to the most complex ice type (in descending order of complexity: solid> ataxitic> reticulate> vein> layered> lenticular> organic-matrix> pore) present in the composite structure. For example, a composite structure that contained ataxitic, reticulate, and layered ice was grouped with the ataxitic structure type. Vertical trending ice structures (vein, reticulate, ataxitic) were considered to be more advanced or complex than horizontal structures (pore, lenticular, layered).

The terrain-unit classification system that we used was adapted from the systems developed by Kreig and Reger (1982) and the Alaska Division of Geological and Geophysical Surveys for their engineering-geology mapping scheme. We modified the system to incorporate the surficial geology units mapped for the area by Carter and Galloway (1985). Because this study focused on permafrost development associated with thaw-lake processes, the analysis was restricted to the following units: (1) alluvial-marine, (2) alluvial terrace, (3) eolian sand, (4) ice-rich thaw basin centers, (5) ice-rich thaw basin margins, and (6) ice-poor thaw basins. Alluvial-marine and alluvial terraces were combined for analysis, due to small sample sizes.

Terrain units were mapped by photo-interpretation of true color photography (1:14,000 scale), and digitizing the units on-screen over a georectified orthophoto mosaic produced by AeroMap, Inc., Anchorage, Alaska. About one-half of the project area was mapped during winter 2001–2002 and we expect the remaining mapping will be completed by December 2002.

LANDSCAPE CHANGE AND THAW LAKE DEVELOPMENT

An analysis of change was performed for three small areas within the larger study area, to assess rates of shoreline erosion resulting from thaw lake processes. The three areas were selected to include three main terrain units (alluvial-marine terrace, alluvial terrace, and eolian sands) (Figure 1). Aerial photographs from 1945 (NARL series, 1:45,000 scale) and from 1955 (USGS, 1:50,000 scale) were controlled to the 2001 orthophoto mosaic using stable features (e.g., polygon intersections, stable lake peninsulas) common to both the old and recent photography. Error assessments were then performed on the co-registered photos to determine the minimum size of change that could be recorded accurately. Five stable tundra features (polygon intersections) were selected (four on the area perimeter and one in the center) for each study area. The average distance between the features on the two dates of photography was used to calculate an average (±□SD) positional error for each study area.

Lake shorelines then were digitized on the 1945 or 1955 controlled photos and compared to shorelines delineated from the 2001 orthophoto mosaic. Lakes were classified as deep (>1.5 m, no bottom features evident,) or shallow (<1.5 m, bottom features evident). This depth corresponds to the critical depth that allows a thaw bulb (talik) to develop beneath the lake. The extent of change was determined by superimposing the two lake shore layers and classifying the resulting features. Any area that was delineated within a waterbody on the 2001 photos but not on earlier photography was defined as eroded. The sum of the eroded areas was used to calculate the percent of the total area lost to erosion. Perimeter-weighted mean erosion rates were calculated for three lake classes: shallow lakes, deep lakes <20 ha, and deep lakes >20 ha. An overall average rate of lateral erosion, was calculated by averaging all erosion rates, weighted by the perimeter of each lake.

THAW SETTLEMENT AND TERRAIN SENSITIVITY

Potential thaw settlement can be calculated from the total volume of excess ice present in the surface sediments and the potential change in the active layer thickness caused by surface disturbance. Excess ice is that portion of the ice that exceeds the pore volume that the soil would have under natural unfrozen conditions, and is equivalent to the maximum volume of soil settlement after thawing. The additional potential surface subsidence due to thawing of wedge ice are computed independently (see below). The term "excess ice" is similar in concept to thaw strain (defined as the decrease in volume a frozen soil sample undergoes when thawed) without external load on soil surface. Excess ice was determined for each of the sampled terrain types (alluvial-marine and alluvial terrace combined, thaw basin, ice-poor margins, thaw basin, ice-rich margins, and thaw basin, ice-rich centers).

When possible, excess ice (% volume) for each permafrost sample was calculated with Crory's (1973) formula:

$$V_x = [(\gamma_{d\,-}\,\gamma_{df\,t})/\,\gamma_{dt}]$$

where: V_x = Volume of excess ice (fractional %)
γ_{df} = Dry unit weight from initial frozen condition (g/cm³)
γ_{dt} = Mean lithofacies dry density values (g/cm³)

The use of this formula requires a dry density value for active layer material measured from each sampled lithofacies. Since many of the lithofacies we encountered were unique to the permafrost zone (due to cryoturbation of organic and mineral materials), the required dry density values could not be derived from measurements of active layer samples. It was found that the relative proportion of organic matter in a sample was more important in determining dry density than particle size composition or texture of active layer samples. Therefore, we developed a regression equation relating dry density to total carbon (%) (n = 17, R^2 = 0.75):

$$\gamma_{dt} = -0.042(C) + 1.4$$

where: γ_{dt} = Dry unit weight for thawed soil (g/ml)
C = % total carbon

Excess ice then was calculated for the samples that were analyzed for % carbon. For permafrost samples, the dry bulk density of the unfrozen soil was calculated from the regression and used to determine % excess ice. For permafrost samples with no determination of % total carbon, % excess ice was determined from the regression of thaw strain to volumetric ice (n = 87, R^2 = 0.70). This approach assumes that during consolidation of soil after thawing the dry density is close to that typically found in the active layer.

In previous investigations in the Colville River Delta (Jorgenson et al. 1997), Kuparuk Oilfields, and Prudhoe Bay (Burgess et al. 1999), visual ice content was well correlated with volumetric measures of excess ice. However, the correlation was poor for the NPRA data, presumably due to the high proportion of organic material in many of the cores. As a result, we were not able estimate thaw settlement for all horizons as we have done in previous studies. Instead we used a statistical approach to estimating total potential thaw settlement, based on determining excess ice volume for 3–5 samples per core. The potential settlement (change in height) for each core was calculated by:

$$\Delta H = [(A_d - A_o)/(1 - V_x)] \times V_x$$

where: A_d = Equilibrium thaw depth of disturbed tundra
A_o = Equilibrium thaw depth of background tundra
V_x = volumetric mean fraction of excess ice

Mean potential thaw settlement (± SD) for each terrain unit was calculated by averaging the Vx values determined for each core.

RESULTS AND DISCUSSION

In presenting and discussing our results, we focus on the patterns of occurrence and processes in permafrost development that are responsible for the evolution of the landscape. First, we classify the types of ground ice and soil materials that occur in near-surface sediments at multiple spatial scales (lithofacies, ice structures, and surface terrain unit), and illustrate how these patterns are interrelated across representative toposequences. The classification of lithofacies and terrain units emphasizes differences in materials that are related to lacustrine (thaw lake), eolian, and marine deposits, and to permafrost aggradation and degradation. We evaluate these classes by comparing differences in soil properties (particle size, salinity, organic content, thaw depths, and ice volumes), with particular attention to ice characteristics. Second, we use the results to develop a conceptual model of the geomorphic evolution of the landscape in the eastern NPRA, which accounts for changes in sediment type, organic matter content, and ice accumulation in the upper 3 meters of the soil. Finally, we evaluate the range of potential thaw settlement resulting from surface disturbance in various terrain units and discuss its implications for land management.

NATURE AND ABUNDANCE OF GROUND ICE

CLASSIFICATION AND MAPPING

Lithofacies

Sediments observed within the NPRA Study Area were classified by texture and structure into 11 lithofacies that reflect depositional environments (e.g., thaw lakes, eolian deposits) and cryogenic processes (e.g., soil turbation)(Table 1). The most common lithofacies in near-surface sediments were (1) massive organics in the active layer on all terrain types except active eolian sand deposits, (2) massive turbated fines with organics just below the active layer, and (3) massive sands at 1–3 meters depth at most coring locations (Figure 2). Other common lithofacies included layered organics; massive fines with organics; and turbated sands with organics. Our interpretations of the processes involved in development of the various lithofacies are presented in Table 1.

Cryostructures

Eight primary ice structure types (based on continuity of ice patterns) were identified in the NPRA: pore, lenticular, vein, layered, reticulate, ataxitic, organic-matrix, and solid ice (Table 2, Figure 3). The same types of ice structure were identified in the Colville Delta (Jorgenson et al. 1997) and the Kuparuk/Prudhoe Bay area (Burgess et al. 1999), indicating that they are widely distributed in arctic Alaska. Therefore, our classification system should be appropriate for describing the range of ice characteristics over a large area of the Arctic Coastal Plain.

Terrain Units

During field surveys, 24 terrestrial terrain units and 7 classes of waterbodies were identified (Table 3). Two of the terrestrial terrain units were related to colluvial processes, three to eolian processes, 13 to fluvial processes, five to lacustrine processes, and one to alluvial-marine processes. Of principal interest for this study were the lacustrine terrain units associated with thaw lake development. These included ice-poor thaw basins, ice-rich thaw basin centers, and ice-rich thaw basin margins (Figure 4). The study also focused on alluvial-marine deposits and alluvial terraces, the original land surface from which thaw basins have developed. For purposes of analysis in other sections of the report, alluvial terraces were combined with alluvial marine deposits because of their similarity and the small sample sizes for each. A preliminary map of terrain units within a portion of the study area is provided in Figure 5.

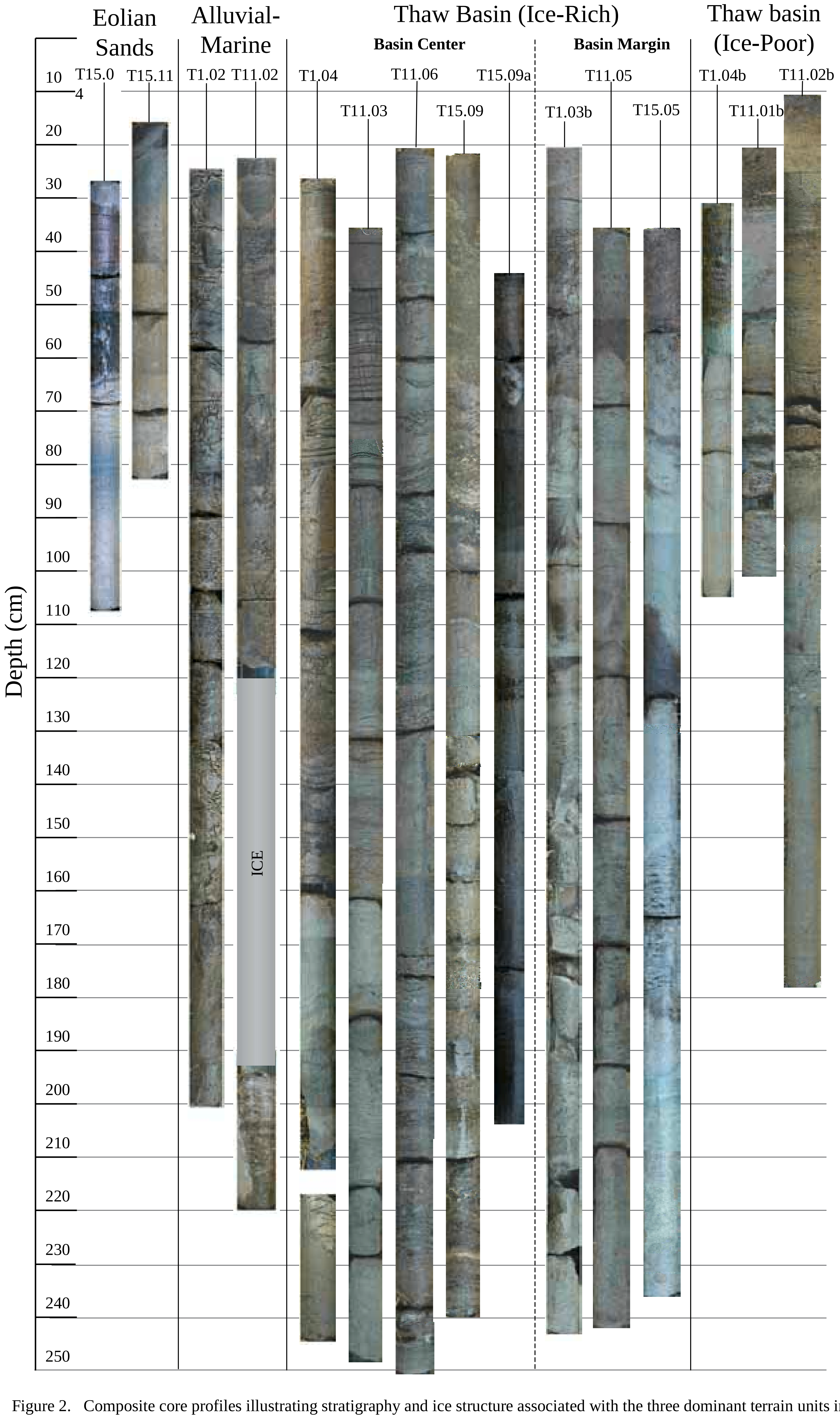


Figure 2. Composite core profiles illustrating stratigraphy and ice structure associated with the three dominant terrain units in the NPRA Development Area, northern Alaska, 2001.

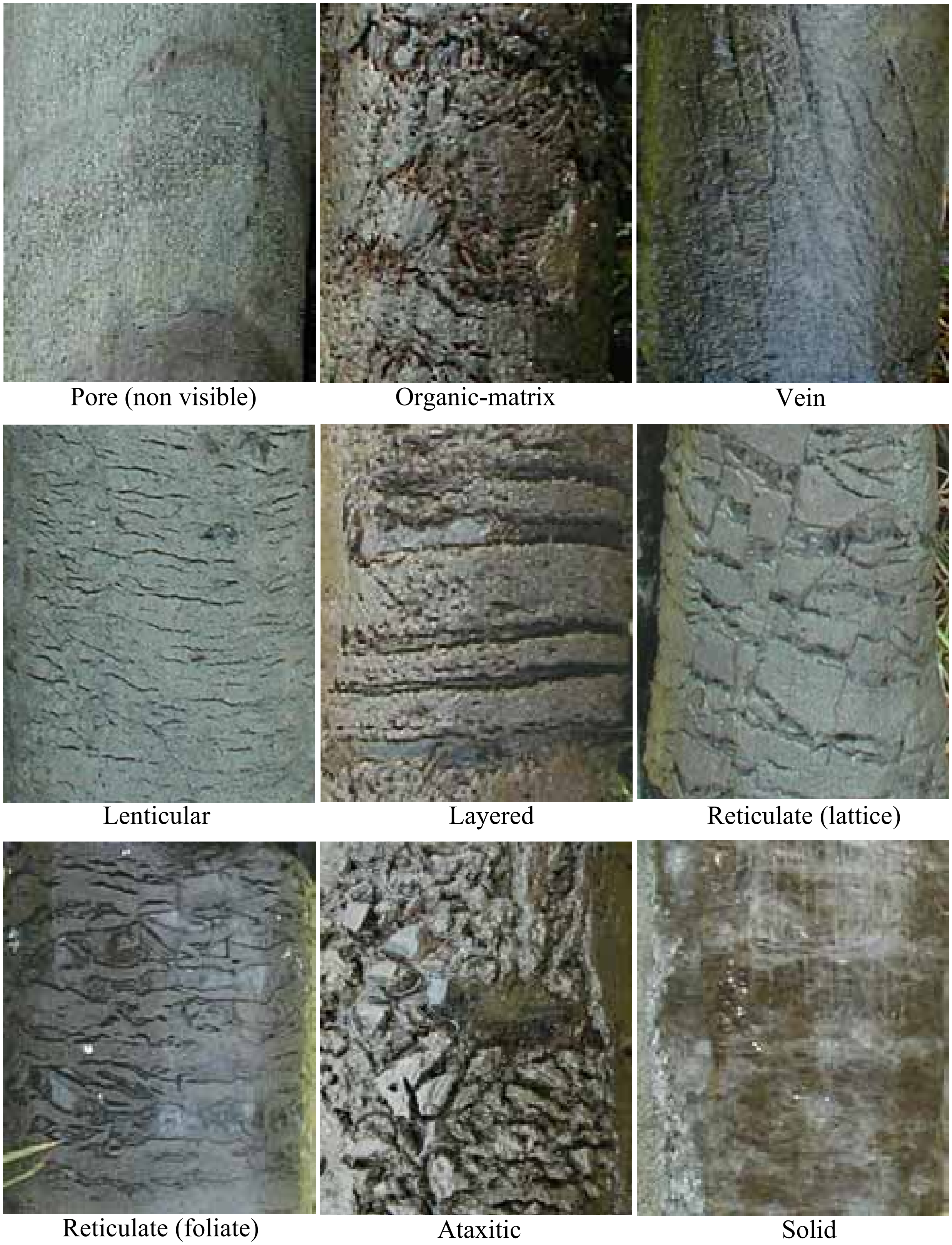


Figure 3. Photographs of simple ice structures found in soils in the NPRA Study Area, northern Alaska, 2001.

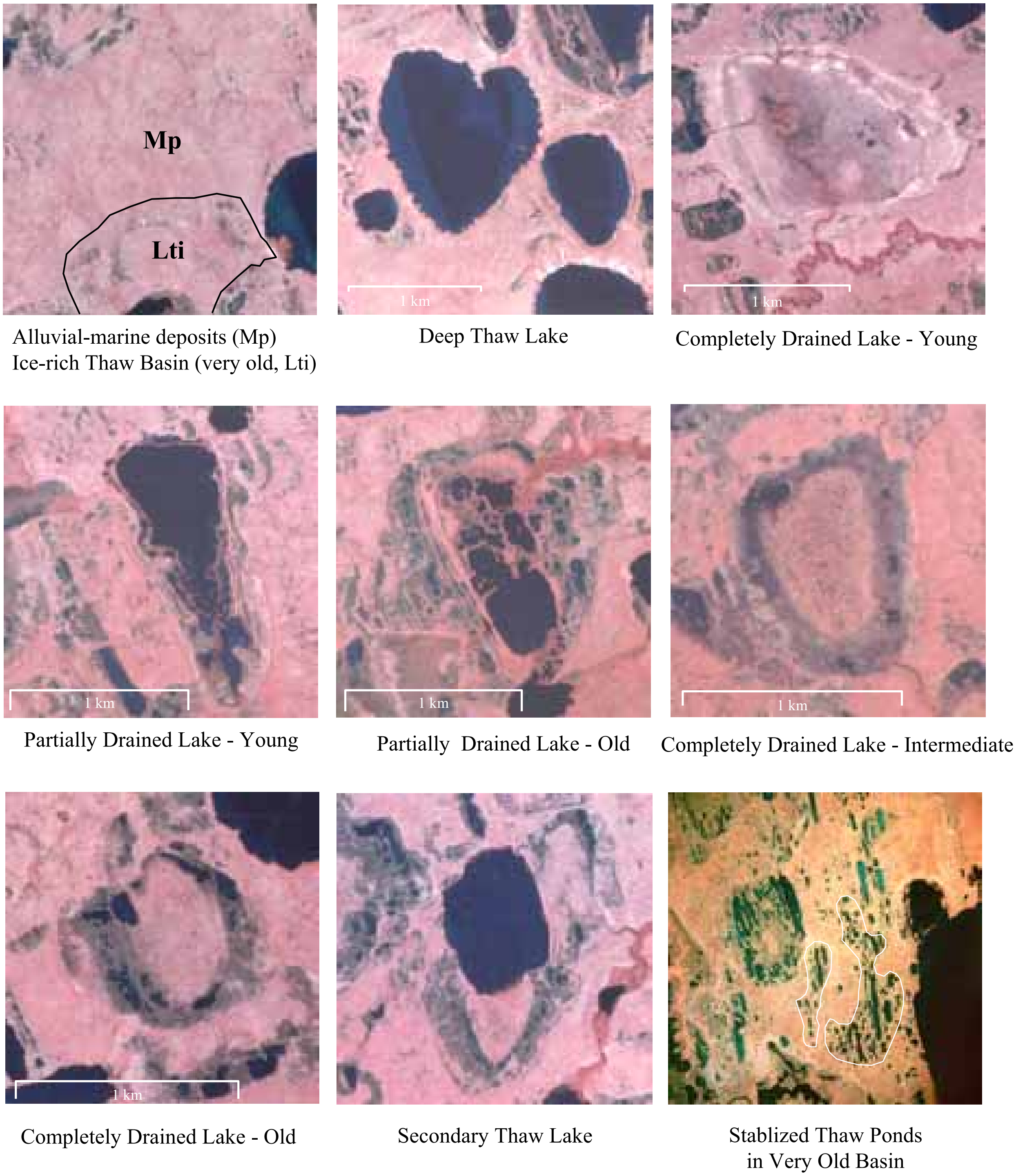


Figure 4. Photographs of dominant terrain units on the coastal plain in the NPRA Study Area, northern Alaska, 2001.

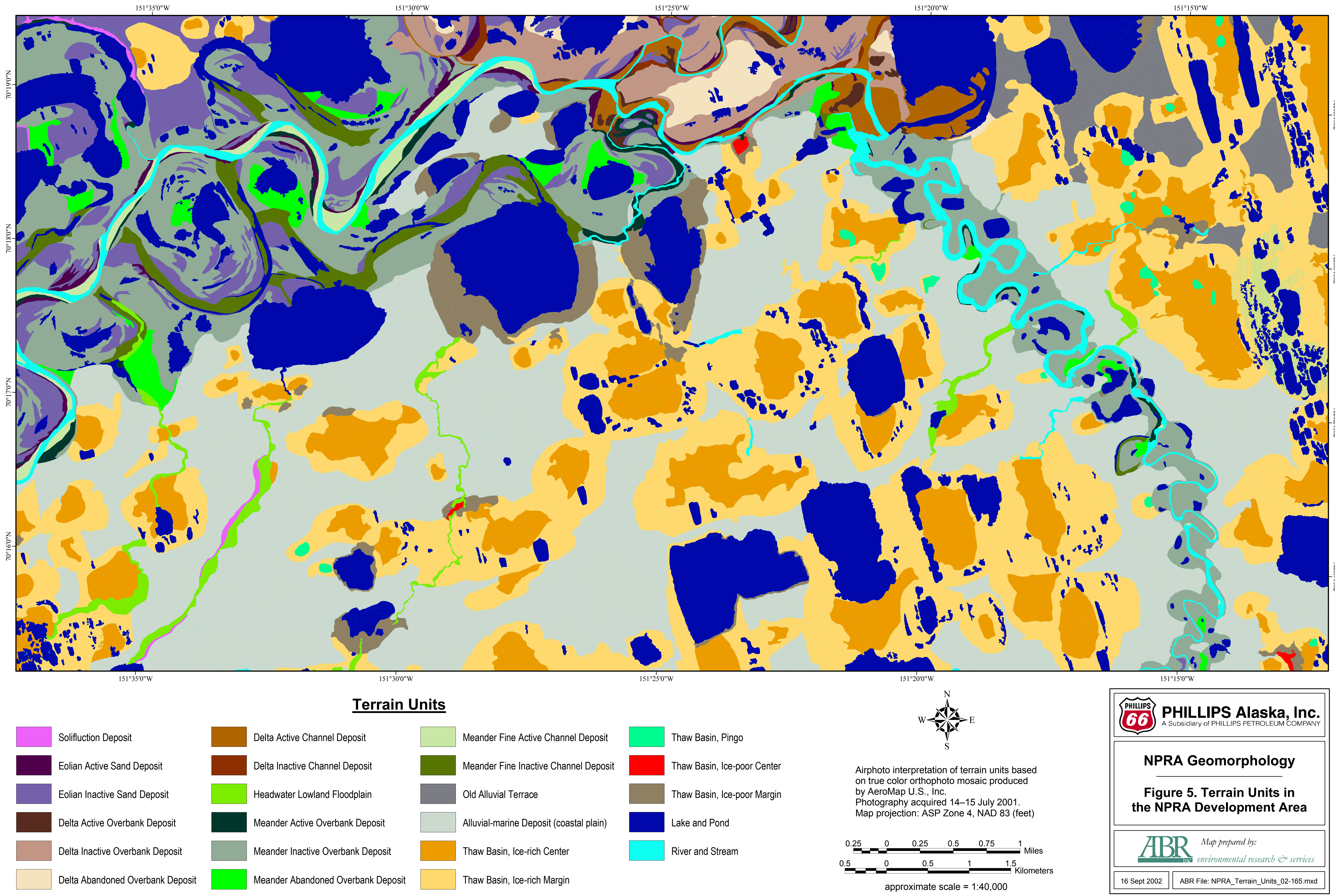


Figure 5. Terrain Units in the NPRA Development Area

Table 1. Classification and description of lithofacies observed in the NPR–A Study Area, Alaska, 1996. Structures less < 4 in. (10 cm) thick are not broken out.

Lithofacies class (and code)	Primary and secondary particle sizes	Sedimentary structures	Mechanism interpretation
Sands-massive (Sm)	Medium-coarse sands	None visible, medium-coarse, light brown sands, may be pebbly	Eolian and marine deposition
Sand, layered (Sl)	Medium-coarse sands	Horizontally stratified layers	Planar bed flow (lower and upper flow regime) or eolian
Sands, inclined (Si)	Medium-coarse sands	Undifferentiated wavy-bedded, ripple, or crossbed stratified layers. Interpretation limited by small size of cores.	Eolian sand dunes
Sands with organics, turbated (Sot)	Medium sands with turbated peat inclusions	Disrupted inclusions or inclined bedding due to cryoturbation	Compression and displacement of material during freezing and thawing
Fines, massive (Fm)	Silts and fine sands	None visible	Eolian, riverine and lacustrine deposits of silts and fine sands
Fines with organics, massive (Fom)	Silts and fine sands with well-decomposed organics	None visible	Soil formation in massive silts
Fines with organic inclusions, massive, turbated (Fomt)	Silts and fine sands with poorly decomposed organic inclusions	Disrupted organic and mineral inclusions due to cryoturbation	Compression and displacement of material during freezing and thawing
Fines, layered (Fl)	Silts and fine sands	Horizontally stratified layers	Lacustrine deposition in shallow thaw lakes
Fines with algae (Fa)	Benthic algal mat and other limnic material	None visible to horizontal lamination	Lacustrine sediments
Organic, massive (Om)	Undecomposed organics, includes trace silt or sand layers	None visible	Autochthonous organic matter, lacustrine or fluvial sedimentation is rare or lacking
Organic, layered (Ol)	Undecomposed organic and fine mineral layers	Horizontal bedding, some mineral redistribution in peat	Lacustrine or eolian deposition in autochthonous organic matter

Table 2. Description of terms used for classifying ground ice in the NPRA Study Area, northern Alaska, 2001.

Cryostructure	Definition
Pore	Ice in minute holes, or pores, within mineral soil matrix that has an almost structureless appearance. May be visible (without hand lens) or non-visible. Visual impression is that ice does not exceed original voids in soil. Forms where pore water freezes *in situ*.
Organic-matrix	Ice formed within organic matrix and has a structureless appearance. May be visible or non-visible. Mostly formed where pore water freezes *in situ*.
Lenticular	Lens-shaped, thin (generally < 0.5 mm), short bodies of ice within a soil matrix. The orientation is generally normal to the freezing front and usually reflects the structure of the sediments.
Vein	Isolated, thin lens, needle-like or sheetlike structures, or particles visible in the face of soil mass. Usually inclined and bisecting sedimentary structures. Differs from layered ice in that they are solitary and do not have a repeated, parallel pattern.
Layered	Laterally continuous bands of ice less than 10-cm thick. Usually parallel, repeating sequences that follow with sedimentary structure or are normal to freezing front. Thicker layers (>10 cm) are described as solid ice. Sparse: ice layers <5% of structure. Medium: ice layers 5–25% of structure Dense: ice layers 25–50% of structure.
Reticulate	Net-like structure of ice veins surrounding fine-grained blocks of soil. Ice occupies up to 50% of surface area. Trapezoidal: ice has distinct horizontal parallel veins with occasional diagonal, vertically oriented veins. Soil blocks have trapezoidal appearance due to fewer vertical veins than lattice-like ice. An incomplete form of latticelike reticulate ice. Latticelike: ice exhibits regular, rectangular or square framework. Foliated: irregular horizontally dominated ice giving soil a platy structural appearance.
Ataxitic	Ice occupies 50–99% of cross-sectional area, giving the soil inclusions a suspended appearance. Sparse: ice occupies 50–75% area, soil inclusions occupy 25–50% of area. Medium Inclusions: ice occupies 75–95% of area, soil inclusions occupy 5–25%. Dense Inclusions: ice occupies 96–99% of area, soil inclusions occupy 1–5%.
Solid	Ice (>10-cm thick) where soil inclusions occupy <1% of the cross-sectional area. Sheet ice: Cloudy or dirty, horizontally bedded ice exhibiting indistinct to distinct stratification. Wedge Ice: V-shaped masses of vertically foliated or stratified ice resulting from infilling of frost fissures. Best identified when large exposures or cross-sections are visible.

Table 3. Classification and description of terrain units in the NPRA Study Area, northern Alaska, 2001. Terrain units modified from Cater and Galloway (1985) and Kreig and Reger (1982). The current report focuses on the six major terrain types.

Unit	Description
Solifluction Deposit	Unconsolidated fine-grained, sandy, or gravelly material, resulting from mass movement of saturated materials. Usually associated with gelifluction processes at the base of slopes and in snowbeds.
Slump Deposit	A type of landslide deposit characterized by downward slipping of unconsolidated fine-grained to gravelly material moving as a unit. Slumps typically are associated with cutbanks along river channels. Areas with slumping often have minor amounts of other mass-wasting processes including debris sliding and falling.
Eolian Active Sand Deposit	Fine to very fine, well-sorted sand containing abundant quartz with minor dark minerals. Sand is stratified with large-scale cross bedding in places. Active dunes are barren or partially vegetated and are undergoing active accretion and deflation. Active dunes usually occur adjacent to exposed sandy channel deposits.
Eolian Inactive Sand Deposit	Fine to very fine, well-sorted sand containing abundant quartz with minor dark minerals. Sand is stratified with large-scale cross bedding in places. Often contains buried soils and peat beds in upper few meters. Inactive dunes are well vegetated, typically have thin to thick organic soil horizons at the surface, and are not subject to active scouring or movement. Inactive dunes occur both on the coastal plain and adjacent to river channels.
Eolian Sand Sheet	Thin to thick deposits of fine to very fine, well-sorted sand on indistinct undulating surfaces. Sand is stratified with large-scale cross bedding in places. Often contains buried soils and peat beds in upper few meters. The inactive surface is well vegetated and have thin to thick surface organic horizons. Sand sheets cover much of the Arctic Coastal Plain. They are mapped in association with distinct dunes. While much of the Arctic Coastal Plain is covered by sand sheets, the sandy surface material is usually included as a component of the Alluvial Plain and Alluvial-Marine Deposits.
Delta Active Channel Deposits	Silty and sandy channel or lateral accretion deposits laid down from the bed load of a river in a deltaic setting under low water velocities. This unit includes point bars, lateral bars, mid-channel bars, unvegetated high-water channels, and broad sandbars exposed during low water. Generally, sediment texture becomes finer in a seaward direction along the distributaries. Organic matter, including driftwood, peat shreds, and other plant remains, usually is interbedded with the sediments. Only those riverbed deposits that are exposed at low water are mapped, but they also occur under rivers and cover deposits. Frequent flooding (every 1–2 yr) prevents the establishment of permanent vegetation
Delta Inactive Channel Deposits	Delta deposits in channels that are only flooded during periods of high flow. Because of river meandering these “high-water” channels are no longer active during low-flow conditions. Generally, there is little indication of ice-wedge development, although a few older channels have begun to develop polygon rims. Very old channels with well-developed low-centered polygons are not included in this unit.
Delta Active Overbank Deposits	Thin (10-50 cm) fine-grained, horizontally stratified cover deposits (primarily silt) that are laid down over sandier channel deposits during flood stages. Relatively frequent (every 3–4 yr.) deposition prevents the development of a surface organic horizon. Supra-permafrost groundwater generally is absent or occurs only at the bottom of the active layer during mid-summer. This unit usually occurs on the upper portions of point and lateral bars and supports low and tall willow vegetation.
Delta Inactive Overbank Deposits	Fine-grained cover or vertical accretion deposits laid down over coarser channel deposits during floods. The surface layers are a sequence (20–60 cm thick) of interbedded organic and silt horizons, indicating occasional flood deposition. Under the organic horizons is a thick layer (0.3-2 m thick) of silty cover deposits overlying channel deposits. Surface forms range from nonpatterned to disjunct and low-density, low-centered polygons. Lenticular and reticulate forms of segregated ice, and massive ice in the form of ice wedges, are common.
Delta, Abandoned-floodplain Cover Deposit	Peat, silt, or fine sand (or mixtures or interbeds of all three), deposited in a deltaic overbank environment by fluvial, eolian, and organic processes. These deposits generally consist of an accumulation of peat 20-60 cm thick overlying cover and riverbed alluvium. Because these are older surfaces, eolian silt and sand may be common as distinct layers or as intermixed sediments. The surface layer, however, usually lacks interbedded silt layers associated with occasional flood deposition. Lenticular and reticulate forms of segregated ice, and massive ice in the form of ice wedges, are common in these deposits. The surface is characterized by high density, low-relief polygons and represents the oldest surface on the floodplain.
Meander Sandy Active Channel Deposits	Sand and mud deposited as lateral accretion deposits in active river channels by fluvial processes. Occassional subrounded to rounded pebbles may be present. Frequent deposition and scouring from flooding usually restricts vegetation to sparse pioneering colonizers. The channel has a meandering configuration characterized by point bars.

Table 3. Continued.

Unit	Description
Meander Sandy Inactive Channel Deposits	Sand and mud deposited as lateral accretion deposits in inactive channels during period of high flow. Because of river meandering these "high-water" channels are no longer active during low-flow conditions. Generally, there is little indication of ice-wedge development, although a few older channels have begun to develop polygon rims. Very old channels with well-developed low-centered polygons are not included in this unit.
Meander Active Overbank Deposit	Thin (0.5–1 ft), fine-grained cover deposits (primarily silt) that are laid down over sandy or gravelly riverbed deposits during flood stages. Deposition occurs sufficiently frequently (probably every 3–4 years) to prevent the development of a surface organic horizon. This unit usually occurs on the upper portions of point and lateral bars and supports riverine willow vegetation.
Meander Inactive Overbank Deposits	Interbedded layers of peat and silty very fine sand material (0.5–2 ft thick), indicating a low frequency of flood deposition. Cover deposits below this layer generally consist of silt but may include pebbly silt and sand and usually are in sharp contact with underlying channel deposits. This unit has substantial segregated and massive ice, as indicated by the occurrence ice-wedge polygons.
Meander Abandoned Overbank Deposits	Sediments are a mixture of peat, silt or fine sand. Surface organic horizon is free of fluvial deposits indicating the terrain is no longer affected by riverine processes. Typically, these areas occupy the highest position on the floodplain, and represent the oldest local terrain. Abandoned floodplain deposits typically have at least 20 cm of surface organics over silt-loam or fine sand alluvium. Low center polygons and small ponds are common.
Headwater Lowland Floodplain	Small streams and tributaries in lowland areas that are too small to be delineated apart from their associated floodplains. These low gradient streams carry little sediment and the floodplain generally is restricted to the immediate vicinity of the stream. The floodp
Alluvial Terrace	Old alluvial deposits, weathered or overlain with eolian and organic material (terrace D of Rawlinson 1993). Soils are cryoturbated loam or sandy loam, buried organics often are present. High-centered polygons are the most common surface form indicating high ice content of surface soils. Thaw basins also are common features.
Alluvial–marine Deposits	Composition is variable but generally consists of a sequence of eolian, alluvial, and marine deposits. Thickness of pebbly eolian sand is highly variable and sometimes absent. Underlying fluvial deposits include gravelly sand, silty sand, and organic silt and occasionally have buried peat beds and logs. Stratified layers of marine gravelly sand, silty sand, silt and minor clay occur in some locations beneath the fluvial deposits and commonly are fossiliferous. This unit is not subject to river flooding. Surface materials can be differentiated as sandy (Mps) or fine-grained (Mpf). This unit includes both the alluvial sand over marine silt and clay (Qam) and alluvial and eolian sand and marine sand and silt (QTas) units of Cater and Galloway (1985).
Loess	Wind-blown silt and very fine sand in homogeneous, nonstratified deposits. On the coastal plain in the study area, loess typically occurs as a layer too thin (<0.5 m) to map as a surficial material.
Thaw Basin Deposit, Ice-poor	Thaw basin deposits are caused by the thawing of ground ice. Soils typically are fine-grained and organic-rich, with stratigraphy re-formed by subsidence. The presence of nonpatterned ground or disjunct polygonal rims indicates that ground ice content is low and that lake drainage has occurred recently. Ponds in these basins typically have irregular shorelines and are highly interconnected. Sandy margins and silty centers are not differentiated because of the lack of micro-topographic features.
Thaw Basin Deposit, Ice-rich Centers	The sediments are similar to those of ice-poor thaw lake deposits but have much more ground ice, as indicated by the development of low-centered or high-centered polygons. The centers of basins usually have organic-rich silty sediments that have high-potential for ice segregation and often are raised by ice aggradation. Surface morphology ranges from low-center polygons at early stages of development to high-centered polygons on distinctly raised domes.
Thaw Basin Deposit, Ice-rich Margins	The sediments are similar to those of ice-poor thaw lake deposits but have much more ground ice, as indicated by the development of low-centered or high-centered polygons. Waterbodies within these basins tend to be rectangular, to have smooth, regular shorelines, and to be poorly interconnected.
Thaw Basin Deposit, Ice-rich Undifferentiated	Sediments similar to ice rich thaw lake deposits but having less ground ice with poorly developed low-centered or high-centered polygons. This type is used when the thaw lake centers and margins are poorly differentiated.
Thaw Basin Deposit, Pingo	Sediments similar to ice-rich thaw basin centers but with much more ground ice indicated by a raised area of well-drained high center polygons.
Tidal River	Permanently flooded channel of lower Judy Creek that is affected by daily tidal fluctuations and has correspondingly variable salinity. The upstream boundary of tidal influence is approximate.

Table 3. Continued.

Unit	Description
Lower Perennial River	Permanently flooded channels of Kalubik Creek. There is no tidal influence, the gradient is low and water velocity is slow, but some water flows throughout the summer. The floodplain is well developed. Rivers of this class generally experience peak flooding during spring breakup and lowest water levels during mid-summer.
Lowland Headwater Stream	Permanently flooded first order tributaries of Judy Creek, Fish Creek, and the Ublutuoch River.
Deep Isolated Lake, Thaw	Deep (≥1.5 m) waterbodies that do not freeze to the bottom during winter. These lakes have no distinct outlets, and are not connected to rivers. The lakes develop from thawing of ice-rich permafrost. Sediments are fine-grained silt and clay.
Deep Tapped Lake w/ Low Water Connection	Deep (≥1.5 m) waterbodies that do not freeze to the bottom during winter. These lakes have distinct outlets (and inlets) connected to rivers. Sediments are fine-grained silt and clay
Shallow Isolated Pond, Riverine	Shallow (<1.5 m) ponds or small lakes with or without emergent vegetation. Water freezes to the bottom during winter, thaws by early to mid-June, and is warmer than water in deep lakes. Sediments are fine-grained silt and clay. These ponds most commonly are found within Ice-rich Thaw Basins and Inactive and Abandoned Overbank Deposits.
Shallow Isolated Pond, Thaw	Shallow (<1.5 m) ponds or small lakes with or without emergent vegetation. Water freezes to the bottom during winter, thaws by early to mid-June, and is warmer than water in deep lakes. Sediments are fine-grained silt and clay. These ponds most commonly are found within Ice-rich Thaw Basins and Inactive and Abandoned Overbank Deposits.

Relationships Among Terrain Components

The terrain components (lithofacies, ice structures, terrain units) generally occur in distinct associations across the landscape. Lithofacies and ice structures were classified by independent characteristics (sediment type versus ice structure) and the relationships between them reveal interrelated processes. However, both lithofacies and terrain units were based on sediment characteristics and lithofacies were used in the definition of terrain units. Thus, components at these two spatial scales are interrelated partly because of the way the components were defined. In the following section we evaluate the interrelationships among lithofacies, ice structures, and terrain units by comparing variation at these three scales across topographic sequences.

Topographic sequences providing information on elevation, terrain, surface forms, and vegetation at selected locations within the NPRA are illustrated in Figures 6–8. Toposequences were oriented to cross both multiple thaw basins and the intervening terrain (Figure 1). Transect 1 (T1), located between Oil Lake and the Ublutuoch River illustrates the topographic progression from ice-poor thaw basin, through ice-rich thaw basin, to the surrounding alluvial terrace (Figure 6). Transect 11 (T11), just west of the Ublutuoch River, crosses ice-poor thaw basins, alluvial-marine deposits and ice-rich thaw basin margins (Figure 7). Transect 15 (T15), located in the northwest quadrant of the NPRA Study Area, crosses eolian sand sheet deposits, an ice-poor thaw basin, and an ice-rich thaw basin (Figure 8).

The profiles provide a useful means of estimating the total amount of ice that has aggraded or degraded during thaw lake terrain development. Differences in relative elevation between the ice-poor thaw basins and the adjacent old surface (alluvial terrace, alluvial-marine deposit, and eolian sand sheet) ranged from 2.5 to 2.7 m, indicating the amount of ice lost during degradation. The difference in relative elevation between ice-poor thaw basins and ice-rich thaw basin centers ranged from 0.7 to 2.7 m, indicating that the amount of ice that can accumulate (causing heaving of the surface) is highly variable.

Analysis of the data obtained from the cores revealed strong relationships among the distributions of ice structures, lithofacies, and terrain units. These relationships can be used effectively to partition the variation in ice structures across the landscape. The frequency of occurrence of most ice structures differed greatly

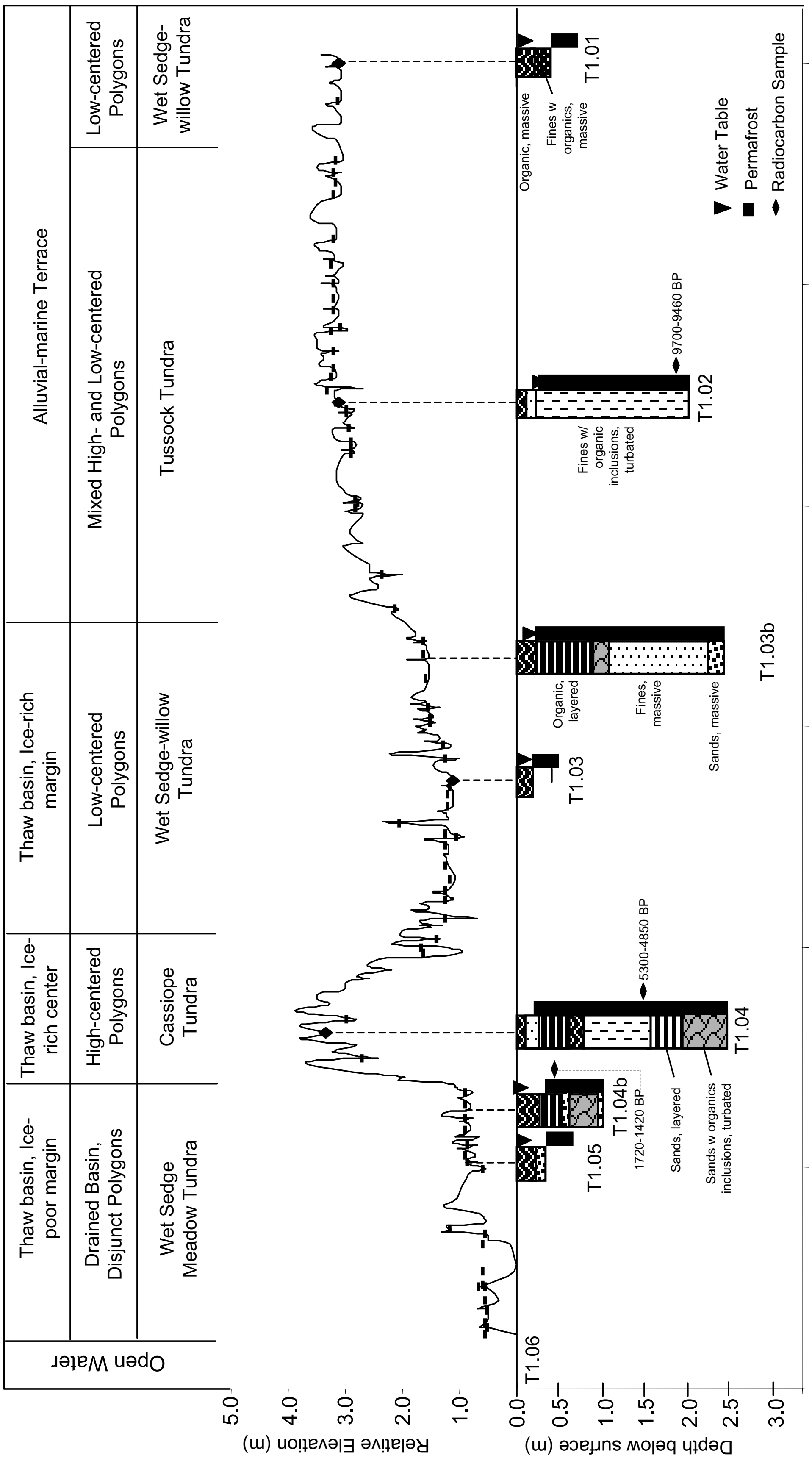


Figure 6. Soil stratigraphy along a terrain sequence (Transect 1) in the eastern portion of the NPRA Study Area, northern Alaska, 2001.

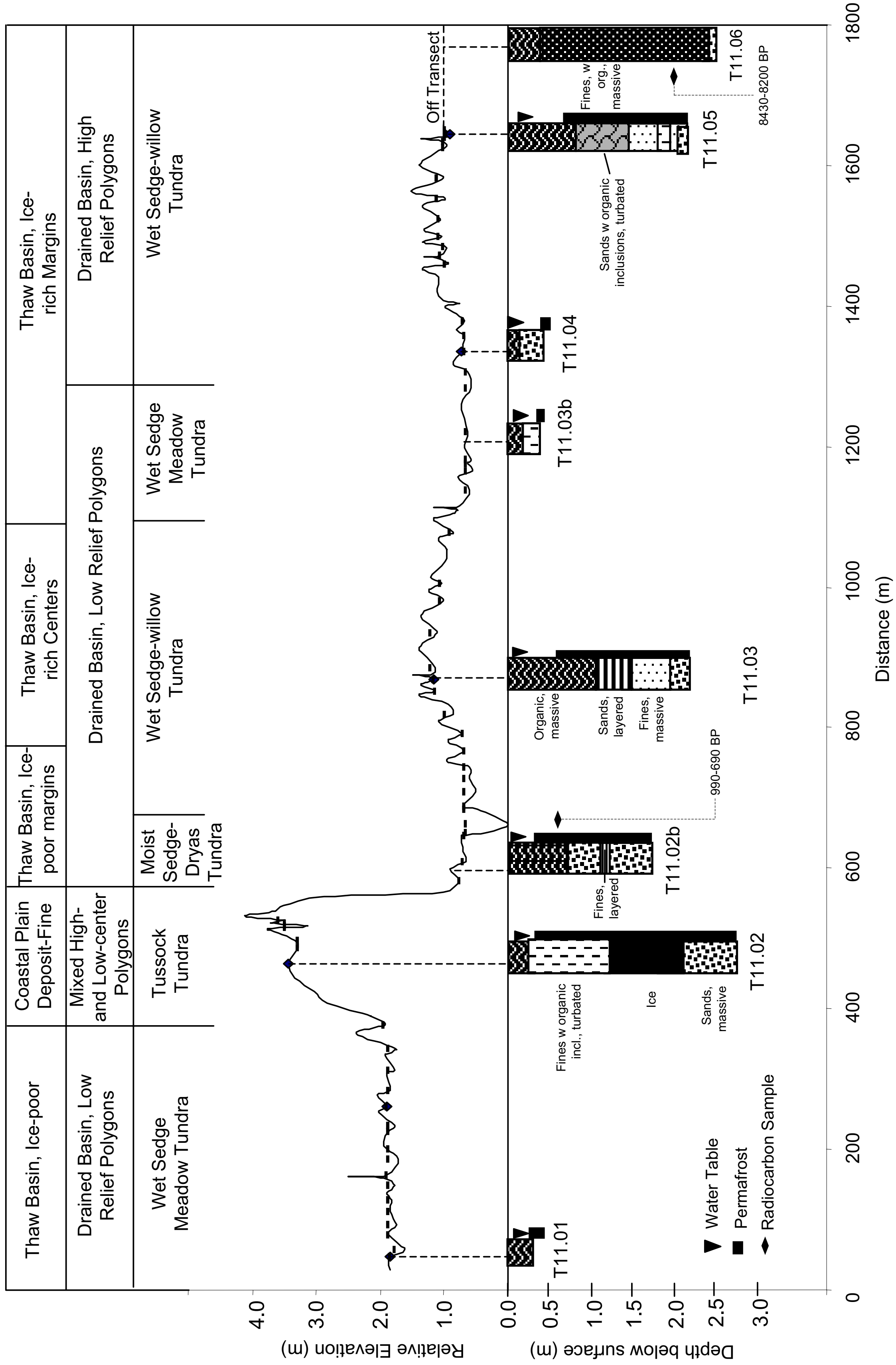


Figure 7. Soil stratigraphy along a terrain sequence (Transect 11) in the central portion of the NPRA Study Area, northern Alaska, 2001.

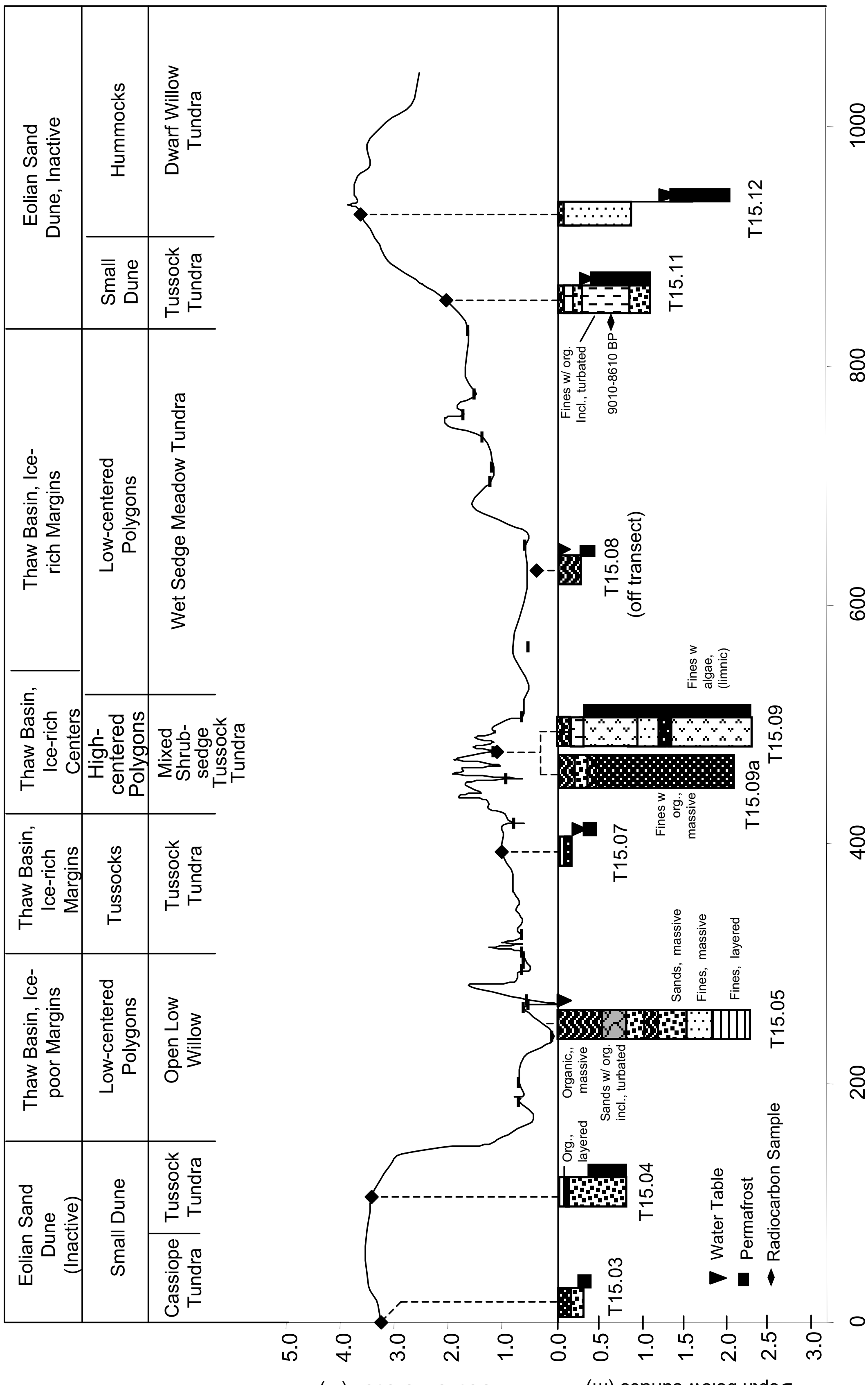


Figure 8. Soil stratigraphy along a terrain sequence (Transect 15) in the western portion of the NPRA Study Area, northern Alaska, 2001.

among lithofacies. Pore ice was almost always associated with massive, inclined, and layered sands; lenticular and ataxitic ice were most frequently associated with massive and layered fines, and organic-matrix ice usually was found in massive and layered organics and limnic fines (Figure 9). In contrast, reticulate ice was broadly distributed among both fine and organic lithofacies.

The associations between ice structures and terrain units were more complex, due to the more complex stratigraphy of soil materials associated

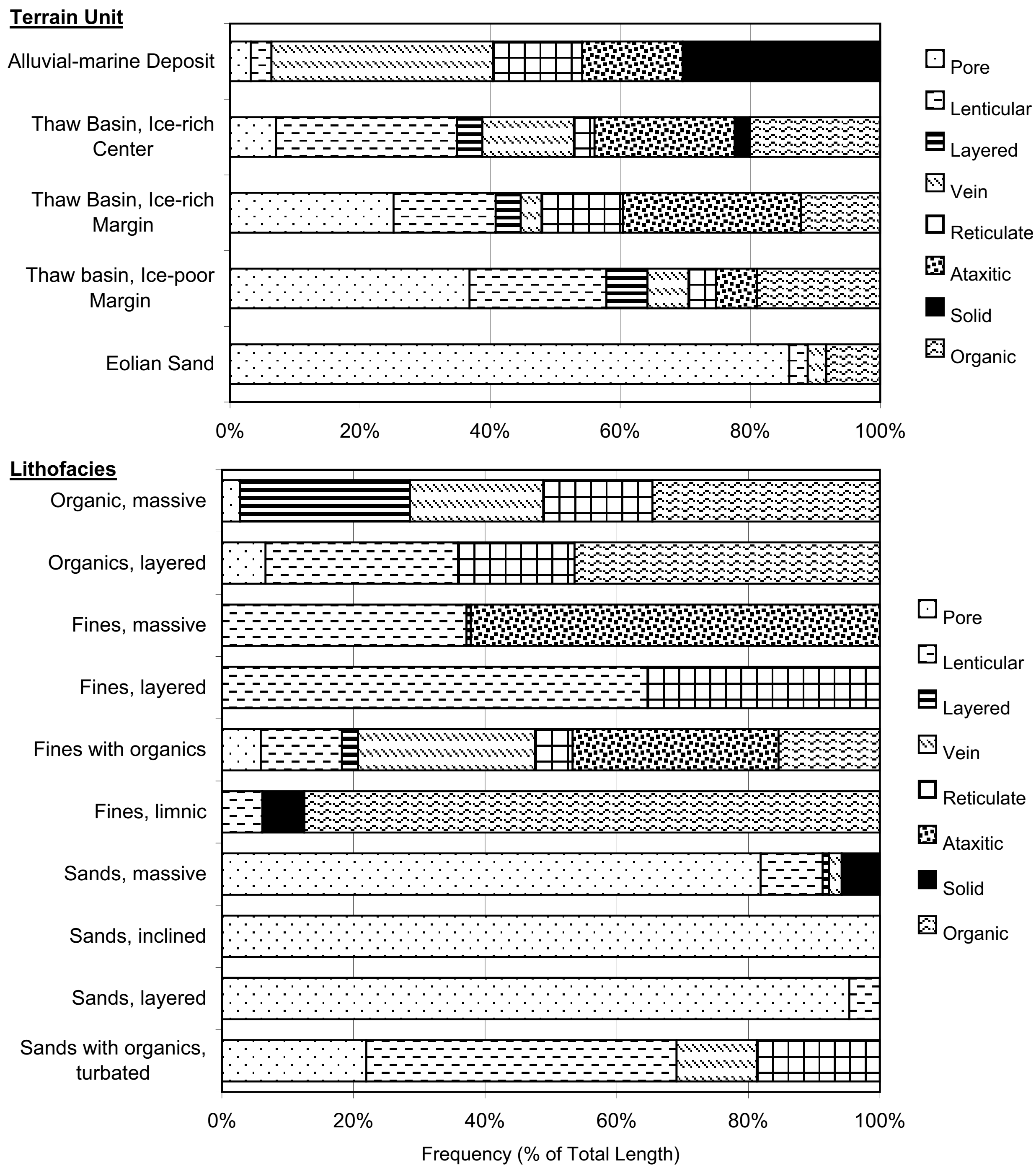


Figure 9. Frequency of occurrence (% of total core length) of ice structures by lithofacies (top) and by surface terrain unit (bottom) for profiles in the NPRA Study Area, northern Alaska, 2001.

with the various terrain units (Figure 9). The simplest association was for eolian sand, which was dominated by pore ice. In contrast, most structures were found in units other than eolian sand but varied somewhat in their frequency. For example, lenticular and ataxitic ice were most frequently found in thaw basins, whereas vein ice was most frequently found in alluvial-marine deposits.

SOIL PROPERTIES

In the following analysis, we compare the physical and chemical properties (particle size, salinity, organic content, thaw depths, and ice volumes) of soils among lithofacies and terrain units. We also discuss landscape processes that contribute to the changes in physical and chemical characteristics of the soil.

Soil analysis was based on lithofacies classifications. The simplest associations between lithofacies and terrain units were seen in alluvial-marine deposits and eolian sand (Figure 10). Eolian sand was dominated by massive, inclined, and layered sand indicative of active dune processes. Alluvial-marine deposits were dominated by turbated fines with organics as a result of extensive ice wedge development and the associated cryoturbation. Lithofacies associations were more complex in thaw basin deposits, reflecting thaw lake and ice aggradation processes.

Particle Size

Due to the small volume of mineral soil present in the samples from some lithofacies, we were able to analyze particle size distribution for only three lithofacies types: massive fines, massive fines with organics, and massive sands (Figure 11). Nearly all samples were composed mainly of sand, most of which was characterized in the field as fine

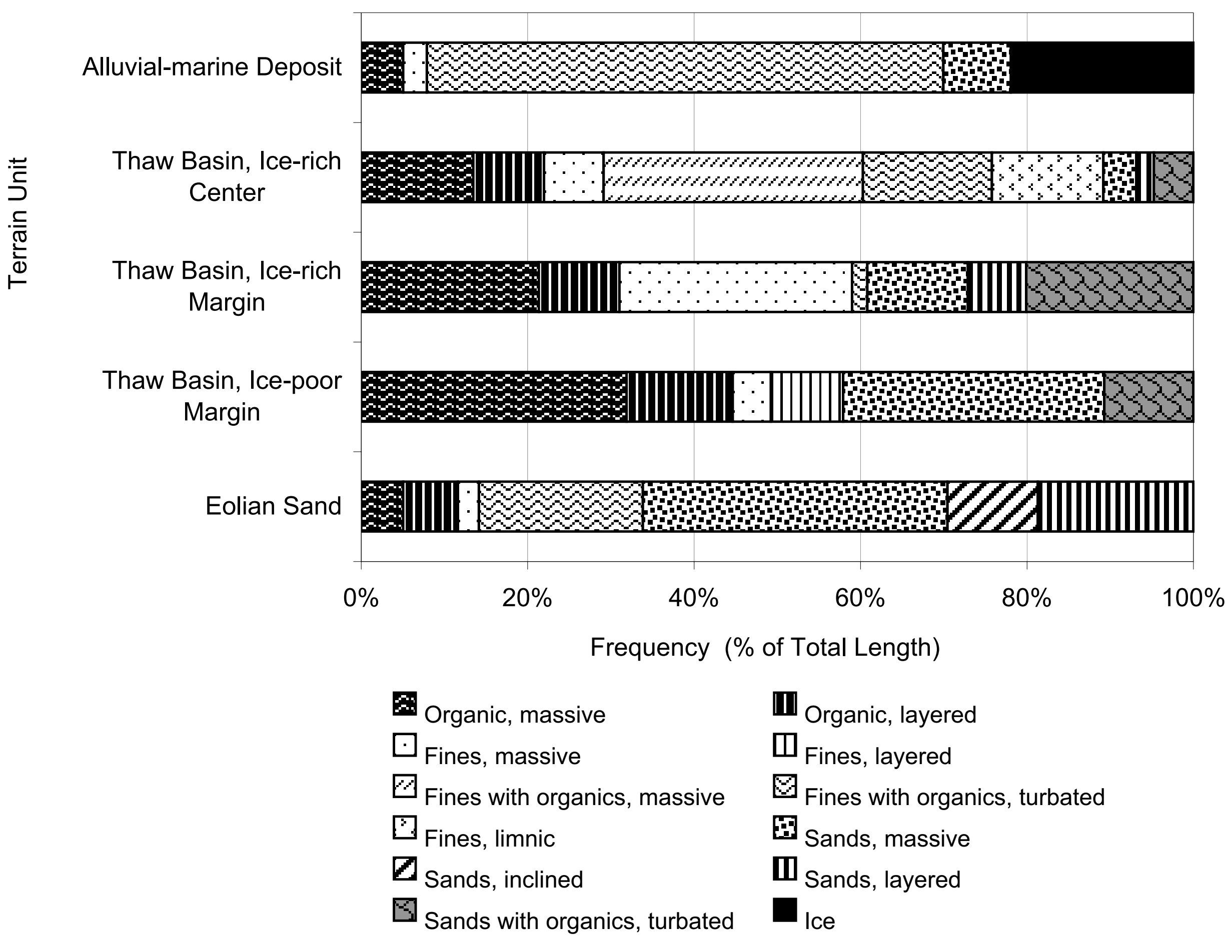


Figure 10. Frequency of occurrence (% of total core length) of lithofacies by surface terrain unit for profiles in the NPRA Study Area, northern Alaska, 2001.

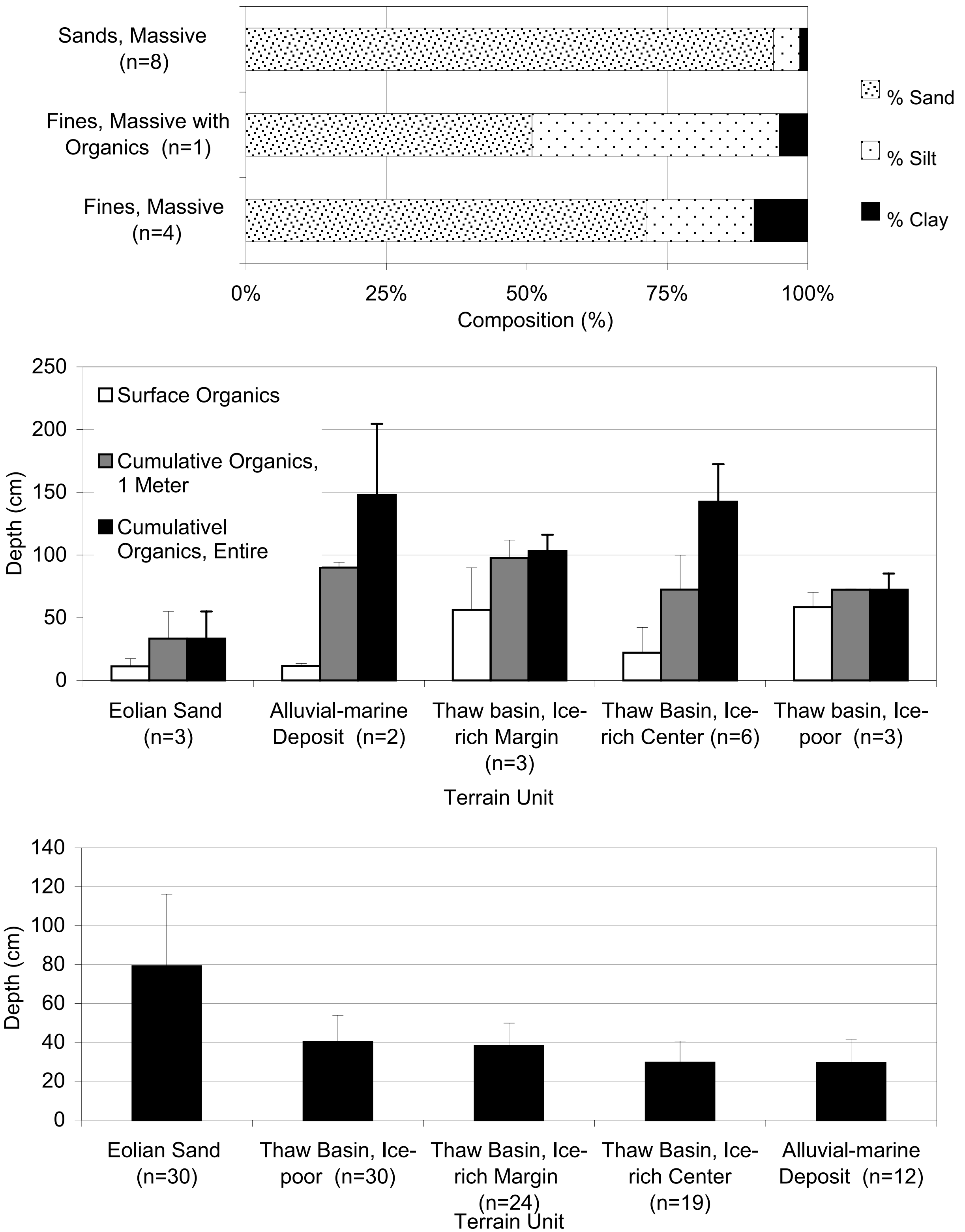


Figure 11. Particle-size characteristics of dominant lithofacies (top), thickness of surface organics and cumulative organics (middle) by surface terrain unit, and thaw depths (bottom) within the soil profiles, NPRA Study Area, northern Alaska, 2001.

or very fine. Medium sand was present in a few samples. The predominance of sand within the study area is likely due to eolian processes associated with the formation of thin loess deposits, sand sheet, distinct sand dunes, and scattered thin loess deposits.

Electrical Conductivity and pH

Salinity, as measured by electrical conductivity (EC), can be used to assess whether the soil material is of marine origin. For all soils near the surface, EC was low (<500 μS/cm) indicating no current marine influence and suggesting that soluble cations may have been leached from the soil by precipitation and drainage. EC generally increased with depth in eolian sands and alluvial-marine deposits, with a maximum value of 1900 μS/cm (2.25 m) in the alluvial-marine deposit. In contrast, profiles for other terrain units generally did not show distinct trends in EC with depth, although there were several elevated values at depth (1100–1500 μS/cm) in the ice-rich thaw basin center (Figure 12).

We attribute the lack of a salinity gradient in the thaw basins to the leaching of soluble cations and redistribution of materials during thaw lake development. The few elevated EC values in the ice-rich thaw basin centers may be related to carbonate deposition in the centers, although the data are insufficient to support a stronger conclusion. We speculate that in deep lakes calcium carbonate is excluded from the ice during winter freezing of the lake surface and precipitated in the sediments. In several cores we observed yellowish granular carbonate materials in thin layers indicative of annual deposition.

Soil pH ranged from 4 at the surface of a ice-rich thaw basin center to 7.8 in the mineral soils of an ice-poor thaw basin and generally increased with depth in all terrain units (Figure 12). We attribute the lower pH values at the surface to result from leaching of cations and production of organic acids in the organic-rich surface layers.

Organic Matter Accumulation

Accumulation of surface organic material was least on eolian sands (11.3 cm) and alluvial-marine deposits (11.5 cm), intermediate in the raised, ice-rich thaw basin centers (22.2 cm) and greatest in ice-poor thaw basins (58.3 cm) and ice-rich thaw basin margins (56.3 cm)(Figure 11).

Subsurface organic horizons within the cores were summed to provide a cumulative measure of thickness, because most cores had discontinuous organic matter disrupted by cryoturbation and thermokarst processes (figure 12). The mean cumulative thickness of organic matter within the top 1-m of the cores was slightly greater in the alluvial-marine deposits (98 cm) and ice-rich thaw basin margins (90 cm) than in the ice-rich thaw basin centers (73 cm) and ice-poor thaw basins (72 cm) (Figure 11). In contrast, cumulative thickness in eolian sand was much less (33 cm) than in other terrain units. The cumulative thickness of all organic layers throughout the core was much greater in the alluvial-marine deposit (148 cm) and ice-rich thaw basin centers (143 cm) than in the ice-rich thaw basin margins (103 cm) and ice-poor thaw basins (72 cm). This comparison by total core length should be interpreted with caution, however, as cores varied in length. For example, some cores in ice-rich thaw basin centers did not reach the maximum depth at which organic layers occurred, so the cumulative thickness of all organic layers was underestimated for this terrain unit.

Thaw Depth

Comparisons of thaw depths are important for evaluating how various landscape surfaces equilibrate thermally with respect to topographical conditions provides a baseline for predicting how the terrain may respond to disturbance. Mean thaw depths were greatest in eolian sands (79.0 cm), intermediate in ice-poor thaw basins (40.1 cm) and least in ice-rich thaw basin centers (29.5 cm) and alluvial-marine deposits (29.4 cm)(Figure 11).

We attribute the greater thaw depths in the eolian sands to soil thermal properties, particularly low moisture contents and the associated low latent heat of fusion. Moderate thaw depths in the ice-poor thaw basin and ice-rich thaw basin margins, despite thick surface organic layers, presumably were due to the presence of standing water at the surface which increases absorption of solar radiation and increases thermal conductivity of material. The relatively low thaw depths in ice-rich thaw basin centers and alluvial-marine deposits presumably were due to their elevated,

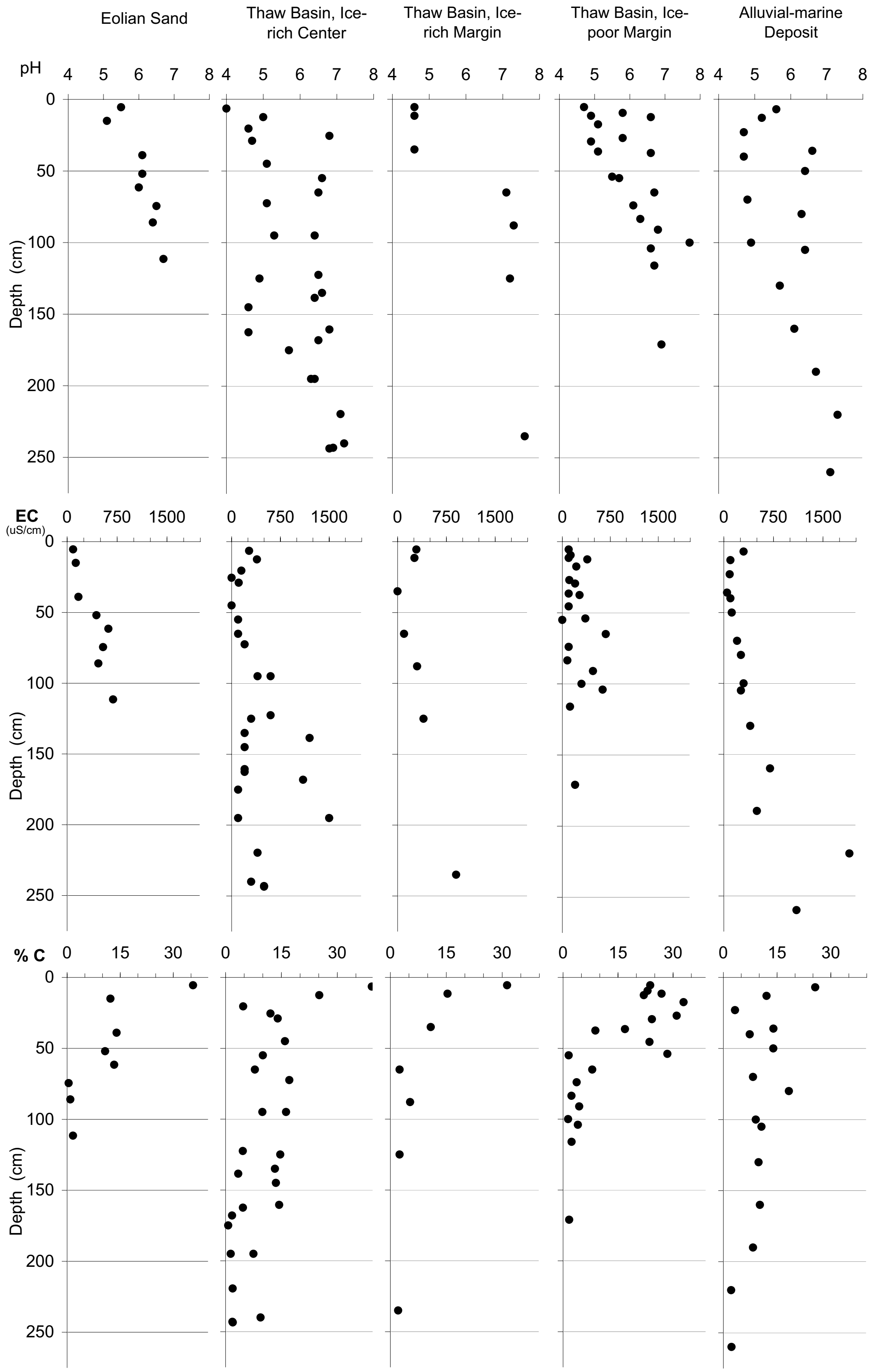


Figure 12. Depth profiles of electrical conductivity, pH, and total carbon by surface terrain unit, NPRA Study Area, northern Alaska, 2001.

relatively well-drained surfaces with abundant vegetation and litter which reduces heat conduction when dry during the summer months.

Ice Volume

To evaluate the distribution of ground ice, we compared the volumes of segregated ice among ice structures, lithofacies (texture/structure classes), and surface terrain units. Among ice structures, mean ice volumes determined from laboratory analysis were highest for layered ice (81%) and organic-matrix ice (77%), intermediate for ataxitic ice (72%), reticulate ice (69%), and veined ice (68%), and lowest for lenticular ice (58%) and pore ice (43%)(Figure 13).

These data show that ice volume is closely related to ice structure and that the ice structure classification alone is valuable for estimating ice volumes. In fact, the classification is better than visual assessments for estimating ice because we found a poor correlation between our visual estimates and laboratory values. Mean ice volumes in this study were similar to results obtained for the coastal plain near Prudhoe Bay and Kuparuk for organic matrix ice (84%), layered ice (81%), reticulate ice (74%), vein ice (70%), lenticular ice (65%), and pore ice (50%), but were somewhat lower for ataxitic ice (83%)(Burgess et al. 1999).

Among lithofacies, the mean volume of segregated ice was highest in massive organics (82%), intermediate in fines with organics (73%) and layered organics (68%), and lowest in massive fines (58%), massive (41%) and inclined sands (46%) (Figure 13). Ice volumes at saturation for pure mineral soils were assumed to be in the range of 40–48%, based on porosity of the silts and fine sands. Thus, volumes above this range represent excess ice (above what the soil would contain at saturation). The large differences in ice content among lithofacies demonstrate that particle size and organic content exert large effects on ice development, although the small sample size resulted in large standard deviations in ice content.

Among terrain units, mean ice volumes were highest in alluvial-marine deposits (71%), intermediate in ice-poor thaw basin margins (60%), ice-rich thaw basin margins (64%) and ice-rich thaw basin centers (64%), and lowest in eolian sand (54%) (Figure 13). Note that these values are for segregated ice (ice wedges were avoided during sampling) in near-surface sediments (1–3 m). Mean ice volumes in this study were very similar to results obtained for the coastal plain near Prudhoe Bay and Kuparuk for alluvial plain (76%), ice-rich thaw basin (74%), and ice-poor thaw basins (58%)(Burgess et al. 1999).

Soil ice volumes generally decrease with depth as soils become dominated by massive sands (Figure 14). In terrain units with thin layers of surface material over the underlying sand sheet (ice-poor thaw basin and eolian sand), ice volumes decrease rapidly with depth. In other terrain types, ice volumes are high throughout the top 1.5–2 meters of the profile.

Accretion Rates

The rates of accumulation of materials were determined using radiocarbon samples obtained from the base of the surface organic deposits. Dates for basal peat were the oldest in alluvial-marine deposits (calibrated calendar years 9700–9460 B.P., n = 1) and eolian sands (9100–8610 B.P., n = 1). Accretion rates (organic and mineral materials) were lowest in the eolian sands, with a rate of 0.07–0.08 mm/yr (Figure 15). Accumulation rates were somewhat higher in the alluvial-marine deposit, ranging from 0.19–0.20 mm/yr. Note that a range in accumulation rate is provided because of the range in calibrated dates. In the ice-rich thaw basin centers, basal peat dates were variable (5300–4850 and 8430–8200 B.P., n = 2). Accretion rates at the two sampling sites were similar (0.28–0.31 and approximately 0.24 mm/yr). Dates from two ice-poor thaw basin margins were younger than other terrain types (990–690 B.P. and 1720–1420 B.P.), and estimated accretion rates were much higher (0.29–0.9 mm/yr, mean 0.54 mm/yr).

LANDSCAPE CHANGE AND THAW LAKE DEVELOPMENT

The oriented thaw lakes of the Arctic Coastal Plain of Alaska have long fascinated scientists because of their importance to ecological processes (Hobbie 1984) and permafrost dynamics (Hopkins 1949, Sellman et al. 1975), their striking pattern (Cabot 1947, Black and Barksdale 1949, Livingstone 1954), apparent cyclic occurrence (Cabot 1949, Britton 1957), and uncertainty about

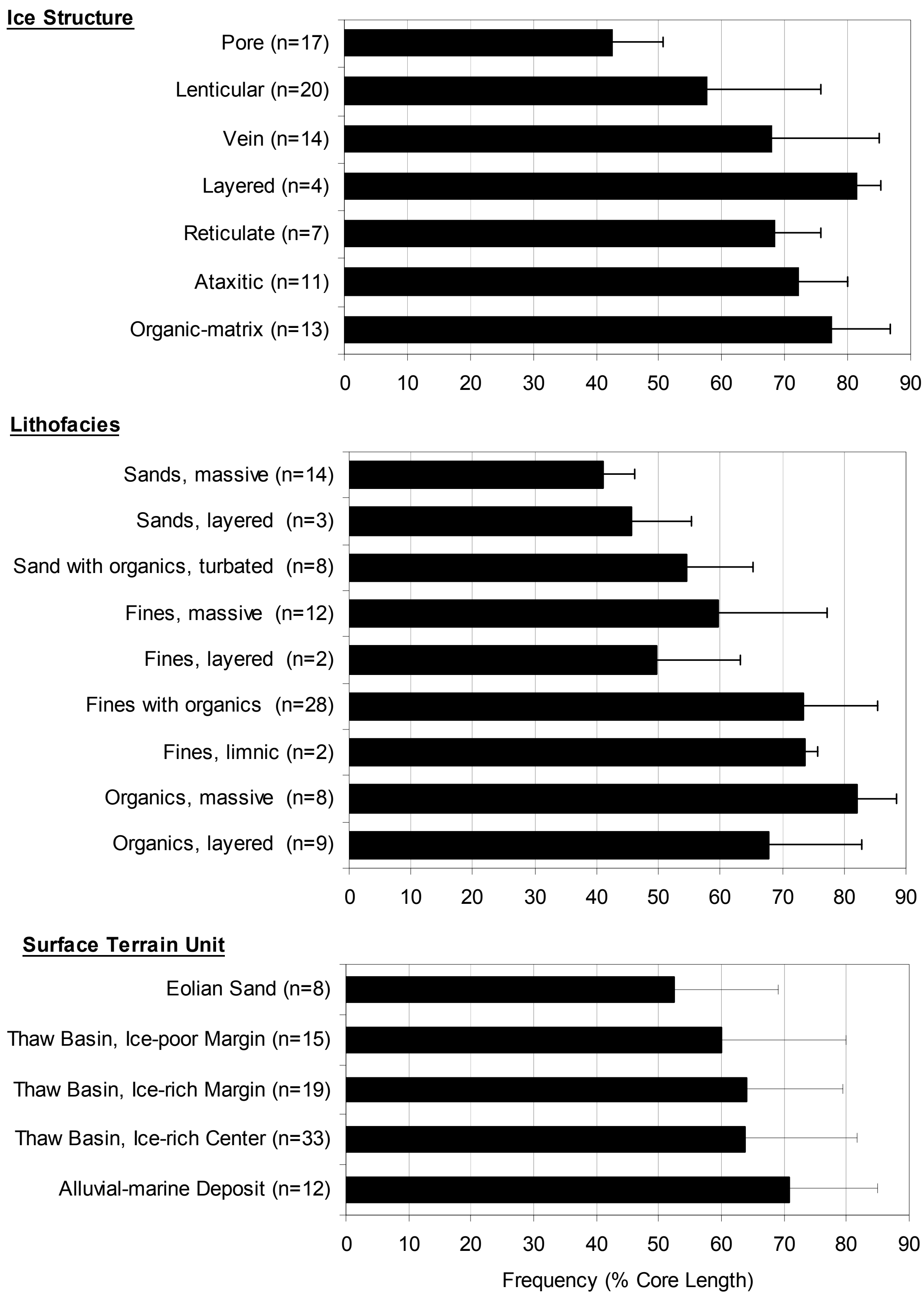


Figure 13. Mean (±SD) volumetric ice contents grouped by primary structure (top), lithofacies (middle) and surface terrain unit (bottom), NPRA Study Area, northern Alaska, 2001.

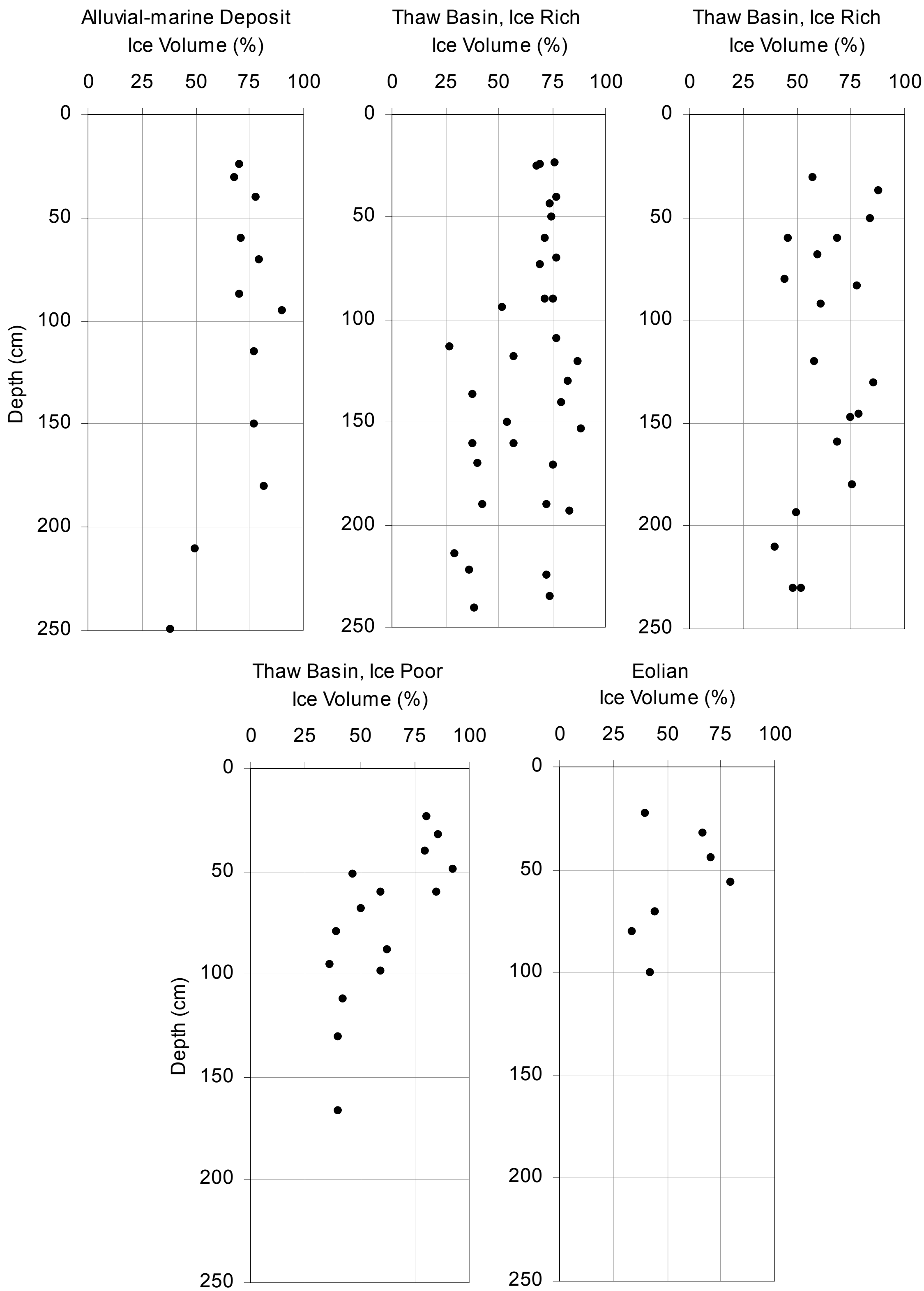


Figure 14. Depth profiles of volumetric ice content by surface terrain unit, NPRA Study Area, northern Alaska, 2001.

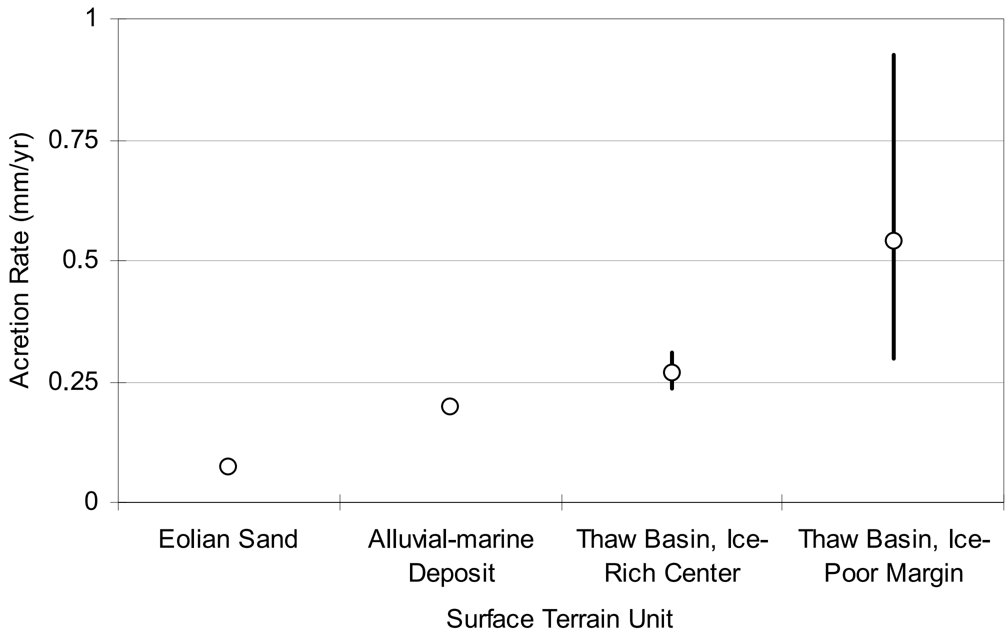


Figure 15. Mean accretion rates of surface materials (organic, mineral, ice) by surface terrain unit, NPRA Study Area, northern Alaska, 2001. Accretion rates were calculated by dividing depth by calibrated radiocarbon age.

their origins (Carson 1968). While a thermokarst origin for the majority of lakes is commonly accepted, the specific mechanisms of ice aggradation and degradation and of lake orientation remain controversial.

Numerous concepts of a "lake cycle" have been proposed, but most lack complete descriptions of the processes by which the surface returns to original conditions. A "lake cycle" was first proposed by Cabot (1947) based on interpretation of lake patterns evident on aerial photographs; this concept emphasized thaw pond formation and drainage and omitted ice aggradation in drained basins. Britton (1957) articulated a more complete "thaw lake cycle" that involved: (1) initial flooding of basins, (2) lake expansion and coalescence through lateral mechanical erosion and thawing accompanied by material sorting, (3) drainage, (4) ice-wedge development in drained basins, and (5) secondary development of thaw ponds from ice-wedge degradation. This concept provided little detail on the ice-aggradation process and does not complete the cycle by recreating the original upland surface. Carson (1968) described a "lacustrine cycle" that included (1) initial development of thaw ponds, (2) expansion and deepening of ponds through lateral erosion and subsurface thawing in a "youthful stage", (3) basin elongation perpendicular to the wind due to longshore currents and thawing at the ends in a north-south direction during the "mature stage", (3) drainage by stream migration, and (4) a secondary cycle of ponds on shelves behind barrier beaches and strands with persistence of the deeper central lake at the "old stage". The role of ice-aggradation and the processes involved in return to the original surface are not included. Everett (1980) defined a more complete cycle beginning with an ice-rich raised surface that includes: (1) climate change or surface disturbance that initiates permafrost degradation, (2) degradation of ice wedges and development of small thaw ponds, (3) expansion of the thaw pond by surface and subsurface thaw, (4) expansion into large lakes by bank erosion and subsurface thawing accompanied by material sorting, (5) partial or complete drainage by stream capture or breaching, (6) and reestablishment of ice-wedges and surface polygon patterns. Finally, Billings and Peterson (1980) described a "thaw-lake cycle" with 10 stages that is similar to Everett's, but emphasizes the role of ice-wedge aggradation and degradation within basins.

While similar, the concepts vary in their explanations of lake formation, in the roles attributed to ice aggradation and degradation, and in their treatments of the return of surfaces to near-original conditions, completing the "cycle". They all assumed that initial conditions had substantial ground ice and were favorable for thermokarst, an assumption that is problematic. Most authors recognized that sediments are sorted and redistributed during lake expansion, yet the importance of this redistribution to ground-ice dynamics has not been generally acknowledged. While attention has focused on ice-wedges, the nature and distribution of other types of ground ice has been overlooked. Finally, previous investigators have hypothesized that the surface returns to near-original conditions, thus creating a "cycle", but have provided little stratigraphic evidence to support the concept. Consequently, the primary reason a consensus has not emerged on the concepts underlying the thaw lake cycle is that these studies have not been supported by the quantitative data on topographic changes, soil stratigraphy, and ice volumes necessary to evaluate the physical processes associated with thaw-lake development.

In this report, we reevaluate the concept of the "thaw-lake cycle" based on detailed terrain analysis, field surveys, and photogrammetry. First, we used a terrain-unit approach to relate surficial materials to landform patterns (see section on Classification and Mapping). Second, we used results of our field surveys to compare differences in elevation and microtopography, sediment characteristics, and ice structure and volume among the terrain units (see section on Nature and Distribution of Ground Ice). Third, we quantified changes in waterbodies over time by comparing aerial photography from 1948–1955 and 2001 to help evaluate rates and patterns of lake development (see section on Landscape Change). We then synthesized this information to develop a modified conceptual model of the thaw lake cycle that incorporates information on both patterns and processes. While orientation is a prominent characteristics of thaw lakes (Sellman et al. 1975, Cabot 1947, Black and Barksdale 1949, Carson and Hussey 1962, and Sellman et al. 1975), it is discussed only briefly here. The specific causes of lake orientation remain controversial, and orientation is not central to understanding the development of thaw lake terrain.

LANDSCAPE CHANGE

Shoreline erosion and waterbody contraction in three small study areas (1450 ha) during the 46–56 year period from 1945–1955 to 2001 were evaluated using photogrammetric analysis (Figure 16). Shoreline erosion rates were greatest in large, deep lakes. These lakes were more prevalent across the landscape and covered a mean of 26.6% of the three study areas (range 14.6–40.4%). Shallow lakes covered a mean of 6.2% (range 5.0–7.8%) in the three map areas (Figure 17) and typically were small (<5 ha). Overall, 0.74% (range 0.45–0.94%) of the land in the three study areas was lost to shoreline erosion. The average annual erosion rate for the three study areas ranged from 0.008–0.017%/yr of total land area (mean 0.013%/yr).

Shoreline erosion was evident in both deep and shallow lakes. Preliminary observations indicated that the amount of erosion depended on type of deposit encountered along the shoreline, and that erosion was most prevalent in ice-rich thaw basin deposits. We attribute this to the fact that most shorelines were comprised of ice-rich thaw basin deposits. However, erosion also was observed in coastal plain deposits (alluvial-marine, alluvial terrace, and eolian sand sheet). The maximum observed erosion rate was 0.8 m/yr, but in most areas erosion rates were much lower (Figure 17). A more detailed analysis of the distribution of erosion among terrain units is planned, pending completion of the integrated-terrain-unit mapping.

A striking feature of the change analysis was the decrease in size of many small, shallow lakes. We attribute this to high water levels in 1945 and to difficulties in interpreting shorelines due to the poor quality of the 1945 photography. Close examination of the photographs revealed that there was much more water in tundra polygons and flooded tundra areas in the 1945 than in 2001. Low quality of the 1945 photography also was a factor: photo-interpretation of shorelines was reliable for deep water bodies where contrast was

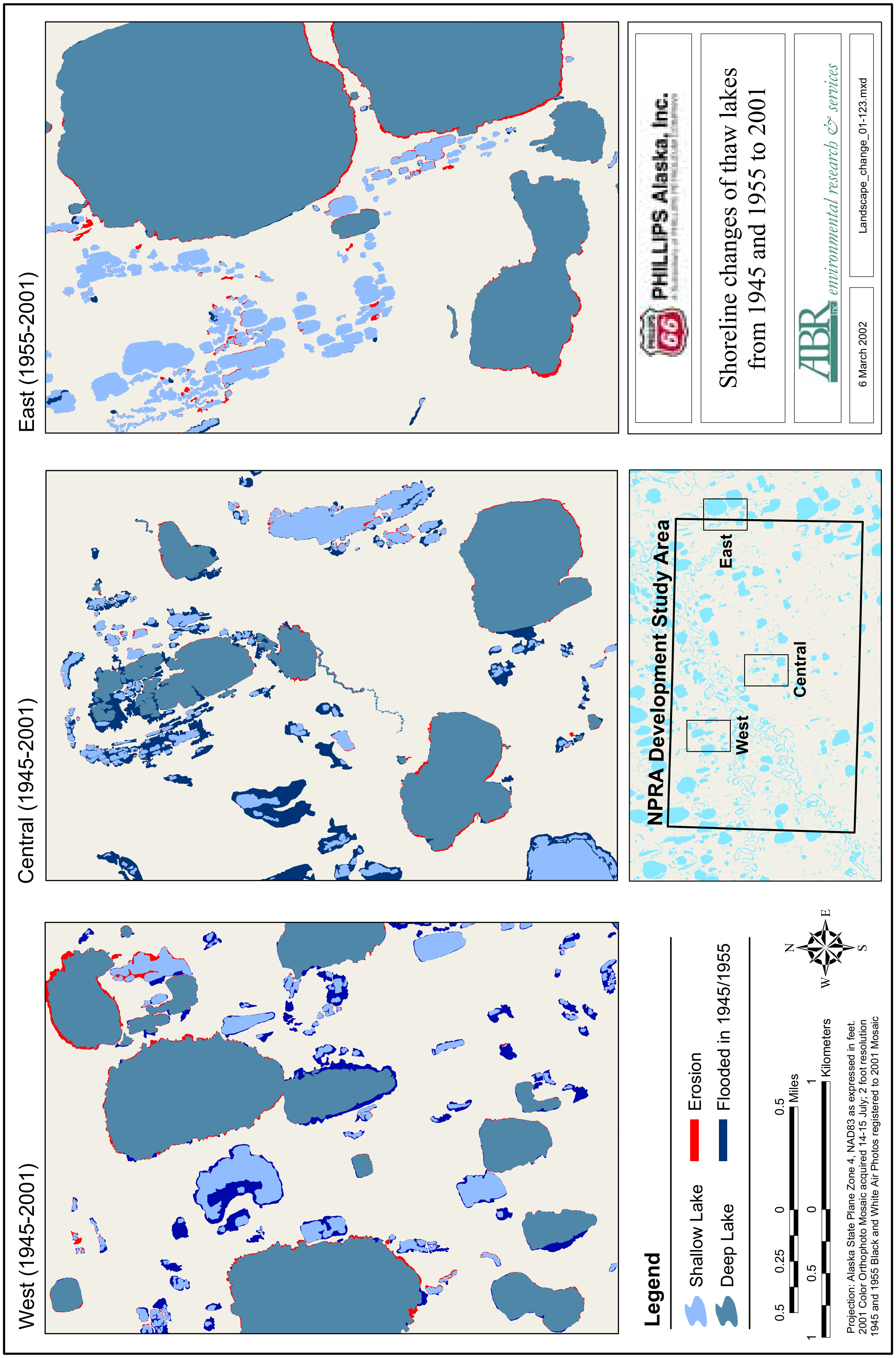


Figure 16. Map of changes in shorelines of thaw lakes from 1948 to 2001, NPRA Study Area, northern Alaska, 2001.

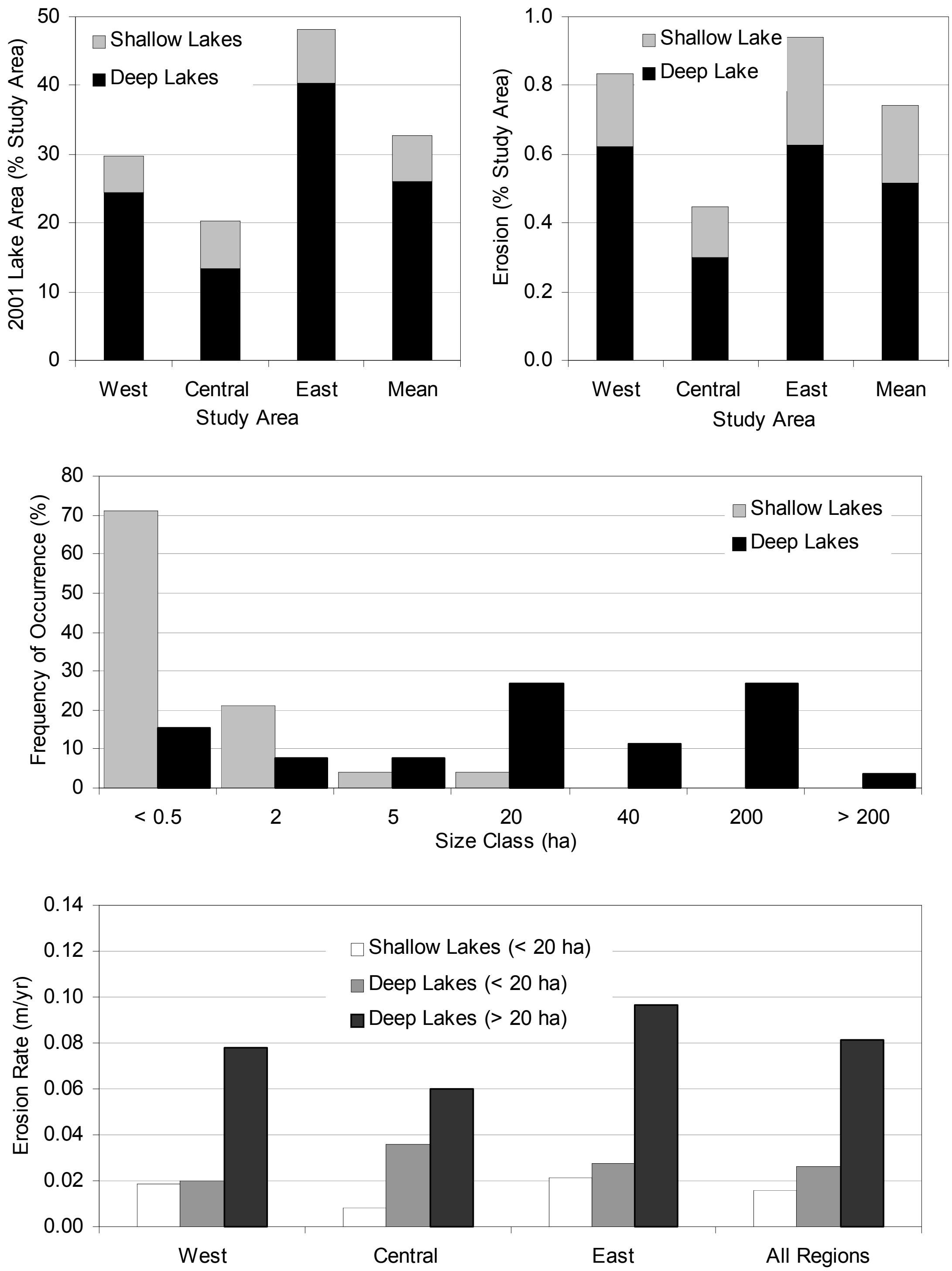


Figure 17. Waterbody characteristics of eroding lakes in the NPRA Study Area, northern Alaska, 2001. Percentage of deep and shallow lakes (top) and their frequency by size class (second from top) are for the year 2001. The percentage of area eroded (second from bottom) and the mean annual rate of shoreline retreat (bottom) are based on changes over 56 years for the western and central areas and 46 years for the eastern area.

high, but was difficult in areas where the shoreline was indistinct or gradational onto flooded tundra. We concluded that the apparent decrease in waterbody size was due primarily to water level changes and limitations of the mapping procedures, rather than contraction of waterbodies due to infilling or heaving of the surface by ice aggradation.

The erosion rates we observed can be used to estimate the ages of the lakes, based on the assumption that the rate of erosion was constant over both time and space. Using the recent rates of erosion, and assuming the lakes started as thaw lakes, the waterbodies would have required 2,466 years (range 1,995–2,875 years) to reach their present sizes. Extrapolated into the future, it would take an additional 5895 years (range 3068–9897 years) to erode all the remaining land. Thus, over a period of 8361 years (range 5961–12427) the landscape could be completely reworked from 100% terrestrial to 100% water. Of course, this process does not occur in a linear fashion and there are numerous factors that cause erosion rates to vary in both time and space. Our data suggest that erosion rates increase in proportion to lake area. As a result, this simplistic approach tends to overestimate the age of the lakes and underestimate the time to complete the erosion; it does provide a rough guide to the age of the lakes, and the rate at which erosion may affect the remaining land.

No recently drained lakes (basins with bare sediments) were present in any of the small study areas, or anywhere in the entire NPRA Study Area. Similarly, ecological mapping of the coastal plain east of the Colville Delta (Jorgenson et al. 1997), revealed that lacustrine barrens (recently drained lakes with barren sediments) also were very uncommon there (<0.1% of area). Drained lakes are more prevalent in some other areas, such as near the coast at Barrow (Sellman et al. 1975) and in the Colville Delta (Jorgenson et al. 1997), presumably because stream gradients are higher or channel migration is more active. The lack of drained lakes in our study areas indicates that the drainage process was more active in the past and that once a lower threshold lake bottom elevation is achieved in the drained basins, repeated drainage is much less likely.

To evaluate the precision of our shoreline change measurements, we calculated the precision of the co-registration of the old and recent photographs. Mean (± SD) positional errors between were 2.6 ± 1.8 m in the west, 1.7 ± 1.7 m in the central and 2.5 ± 1.8 m in the east study area. While these errors are small, a substantial portion of the shoreline changes were within our measurement precision.

In summary, the measurements of shoreline change indicated that erosion is still an active process and continues to alter the landscape. Preliminary examination of the distribution of thaw degradation indicates that reworking of the surface by thaw lakes is mostly confined to the preexisting basins. The lack of drained lakes in our area compared to observation in other areas of the coastal plain also suggests that drainage was more active in the past and set up the conditions for the current thaw lake activity.

CONCEPTUAL MODEL OF THE EVOLUTION OF THAW LAKES

We have revised the earlier conceptual models of thaw-lake development to include six main stages of development (Figure 18). These include: (1) initial flooding of primary lakes, (2) lateral expansion and sediment redistribution, (3) lake drainage, (4) ice aggradation in exposed sediments, (5) secondary development of thaw lakes, and (6) thaw pond stabilization. This conceptual model applies to most of the Arctic Coastal Plain, where sand sheets blanket the surface. For other areas with thick deposits of eolian or fluvial silt, which tend to be extremely ice-rich, such as the Colville Delta and Seward Peninsula, the primary stage (initial flooding of primary lakes) is absent. Alternatively, the primary stage involves initial thawing and coalescing of ice wedges, and the final stage (thaw pond stabilization) does not occur. These developmental stages are explained more fully below, with emphasis on topography, soil stratigraphy, ice content, and relative ages. We expect to refine this preliminary model after accumulation of more fieldwork in 2002.

Initial Flooding of Primary Lakes

Radiocarbon dating of lake sediments indicates that lakes have existed on the coastal plain only since about 12,000 years ago, when the climate warmed from colder glacial conditions at the end of the Pleistocene (Hopkins et al. 1981,

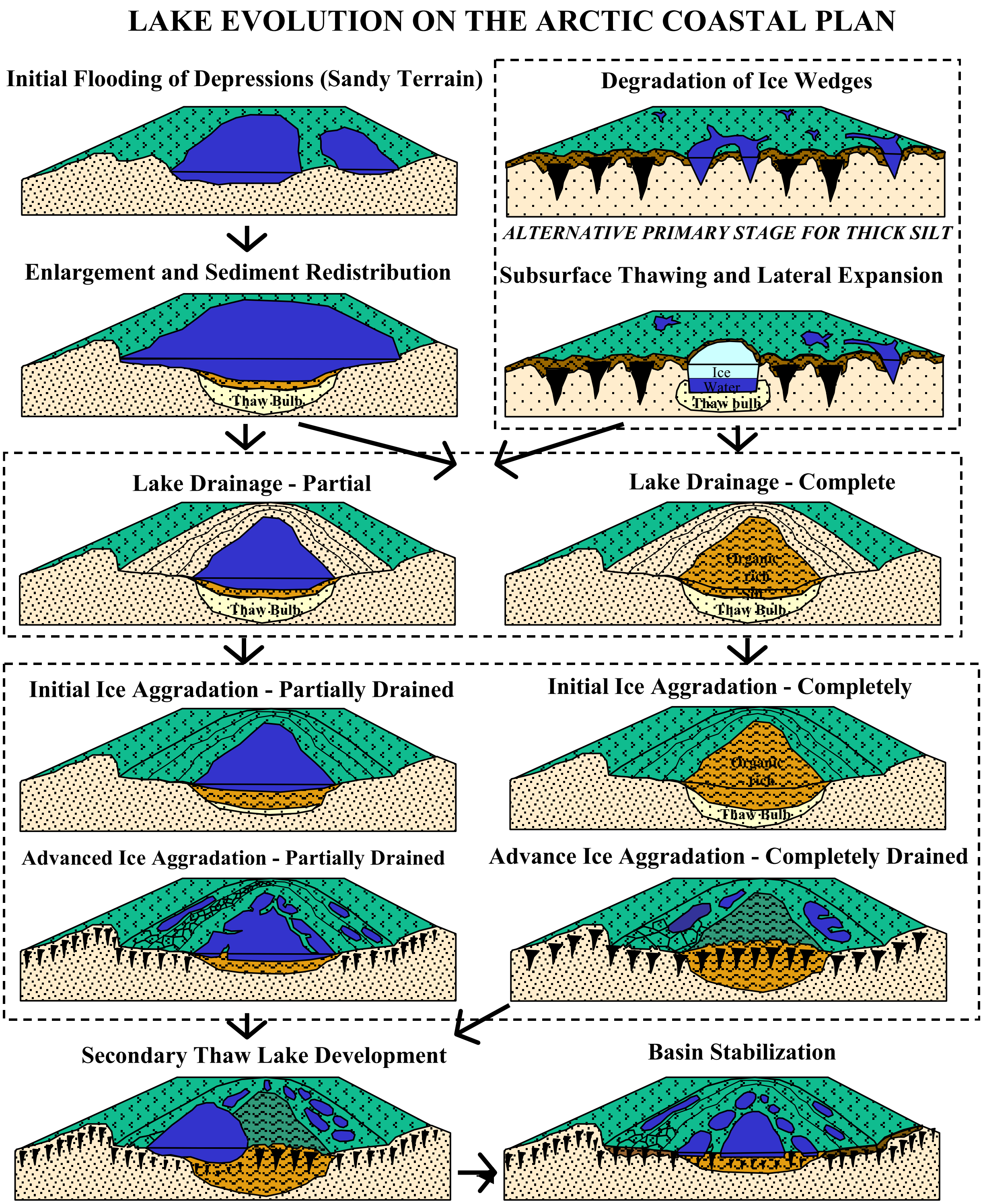


Figure 18. A conceptual model of changes in lithofacies, ice structures, active layer thickness, and ground elevations during thaw lake evolution in the NPRA Study Area, northern Alaska, 2001.

Rawlinson 1983, 1993). Most of the lakes appear to be less than 9,000 years old (Rawlinson 1993). The lack of older lakes is attributed to cold, arid and windy conditions and active eolian deposition near the end of the Pleistocene Era (Cater et al. 1987, Rawlinson 1993).

Eolian sediments are nearly ubiquitous across the Arctic Coastal Plain; in most areas eolian deposition ceased and the ground surface was stabilized about 8,000 years ago (Carter et al. 1984). In our study, calibrated radiocarbon ages for basal organic material from the centers of two old, deep lakes were 4850–5300 BP and 8200–8430 BP (Figures 6 and 7). This material is rich in algal remains and provides an age range for lacustrine material associated with initial lake development. In comparison, basal peat from an adjacent alluvial-marine deposit with an eolian cap yielded radiocarbon ages of 9460–9700 BP (Figure 6), providing an approximate date for stabilization of the oldest surfaces in the area. While the sampling is limited, the older dates indicate that the primary lakes formed soon after stabilization of the land surface during the early Holocene.

During the period when the sand sheets were stabilizing, ground ice content would have been low because: (1) the material was recently deposited, or there would not have been sufficient time to develop much ice, (2) the cold, dry landscape lacked water for ice-wedge development, and (3) the sandy sediments have low potential for development of both segregated and wedge ice. In our field surveys, sandy lithofacies typically had pore ice structures and the amount of excess ice was low. Therefore, the first lakes to develop during the early Holocene could not have been formed by degradation of ground ice. Rather, these lakes formed simply by the accumulation of water in depressions. This mode of development of the first lakes is consistent with the initial basin flooding concept of Britton (1957), Gravis (1978), and Shur (1988), and alluded to by Billings and Peterson (1980), but absent from Everett (1980). Some of the deep lakes that exist today may be remnants of these old primary lakes.

In the classical concept of lake development (alternatively primary stage in thick silt, Figure 18), the formation of waterbodies begins with the degradation of ice wedges in ice-rich upland terrain (Shumsky and Vturin 1963, Tomirdiaro and Riabchun 1978, Everett 1980). Degradation is most intense at the intersections of ice wedges, and deepening of the water in the troughs leads to the formation of small, deep (>1.5 m) ponds. A thaw bulb then develops under the deep water and the thaw lake expands laterally through both mechanical and thermal erosion. While this process is evident in extremely ice-rich, silty sediments, such as those on the Colville Delta (Jorgenson et al. 1997, 1998) and the Seward Peninsula (Hopkins 1949), this stage of thaw lake development is not evident on the sandy coastal plain to the east and west of the Colville. We have not yet been able to identify any small, deep ponds on upland surfaces within the study area that appears characteristic of this stage of development. Even severe disturbances in the region, such as the "peat roads" bladed during early oil exploration, have created only discontinuous deep thermokarst pits and high-centered polygons, not deep thaw ponds.

Lateral Expansion and Sediment Redistribution

After initial flooding, lake levels would have fluctuated in response to changes in precipitation and other components of the water balance, and shorelines would have expanded as a result of wave erosion. In the study area, we observed that the older and higher alluvial-marine deposits (capped with eolian sand) were eroding principally through mechanical erosion, leaving characteristic wave-cut benches. In these locations (i.e., G15.02, T12.10, T13.01, T13.08), the banks were 1.5–2 m high and comprised mostly of fine sands that typically had pore ice with low ice content. At these locations, the thawed portion of the bank had slumped only 0.2–0.3 m due to thaw settlement, indicating that thawing of ground ice was insufficient to account for lake formation. Another indication that erosion was dominated by physical processes (vs. thermal erosion) was the presence of a shallow, sandy wave-cut bench at the foot of the eroding bluffs. Water depths increase very slowly from <0.1 m near the bank to 0.3–0.5 m at a distance of tens of meters. In contrast, bank morphology that typically results from thermal erosion, such as thermal niches which develop from melting of ice rich layers, was observed only at one site where the lake had eroded pingo ice.

Sediments near the wave-cut bank are dominated by fine to medium sands. The soil stratigraphy, shoreline morphology, and sediment composition all indicate that wave-dominated erosion and sediment transport are the dominant processes, not settlement of ice-rich terrain. In contrast, true thaw lakes in ice-rich terrain, such as the Colville Delta, have steep shore profiles and water depths of 3–4 m within 10 m of the banks (Jorgenson et al. 1997).

Redistribution of sediments during early lake formation and expansion, with concomitant accumulation of fines and organics in the deepest portions of the lakes, is key to the dynamics of later ice aggradation and degradation. Numerous investigators have observed the prominence of sandy sediments along the margins of large lakes and the accumulation of fine-grained sediments and organic material in the centers (Britton 1957, Tedrow 1969, Carlson and Hussey 1962, Carson 1968, Hunter and Carter 1985). In our study, sediments near the margins of ice-rich thaw basins were dominated by massive and layered sands (fine and medium sands), whereas the sediments of the ice-rich thaw basin centers were dominated by massive fines with organics, turbated fines with organics, and limnic fines (algal rich)(Figure 10). Also of interest is the fate of organic material from the eroding bluffs. Our sampling, as well as other studies (Britton 1957, Kidd 1990), revealed a continuum of peat bock sizes away from the eroding bluff, apparently due to incorporation of as organic material into the lake sediments. Peat sod mats may be incorporated into the adjacent sandy sediments or may roll and float along the shore, decompose, and become incorporated bottom sediments in the basin center. Drifting ice also can redistribute peat toward deeper water (Shur 1977). Peat-rich sandy sediment was classified as turbated sand with organics and was frequently observed in thaw basin stratigraphic profiles (Figure 10). Most of the organic-rich silt that was observed appeared to contain algal material and small amounts of vascular plant parts, such as shredded leaves and seeds. This sorting of materials during the primary lake stage establishes the conditions for later differential ice segregation and development of ice wedges because particle size is fundamental to the development of ice (Figure 9). Shoreline erosion, lake expansion, and redistribution of sediments were partially recognized in the thaw-lake concept of Britton (1957), but are absent from the concepts of Billings and Peterson (1980) and Everett (1980).

Lake Drainage

The partial or complete drainage of lakes by stream capture, shoreline breaching, or coastal erosion is a dramatic and frequently observed phenomenon in some regions of the coastal plain (Cabot 1947, Hopkins 1949, Britton 1957, Tedrow 1969, Carlson and Hussey 1962, Carson 1968, Everett 1980). While the process of lake drainage appears straightforward, differences in the extent and rate of drainage result in a wide range of subsequent basin water-levels and surface conditions. In addition lakes levels can fluctuate over time due to climatic changes, or be drawn down slowly as the outlet channels progressively erode and lower the base levels of the lakes. The multiple processes causing water-level changes thus create a range of surface ages from initial exposure. This stage of thaw-lake development is incorporated in all conceptual models, although substantial uncertainty remains about the relative importance of tapping versus climatic change in causing lowered water levels in the lake basins.

Radiocarbon dates for two basal peat samples from *in-situ* peat layers on ice-rich thaw basin margins were 690–990 BP and 1420–1720 BP. This sedge peat formed after drainage of the basin and therefore provides approximate dates for the drainage of the basin margins. These dates are consistent with dates of partially drained lakes near Barrow (Carson 1968), where most old basins have sequences of two to four ancient strands. Radiocarbon dating of these strands revealed that most are between 700 and 3500 years old. The radiocarbon dates for both our study area and Barrow, combined with the very low numbers of recently drained lakes (thaw basins with bare or partially vegetated surfaces) both in our NPRA Study Area and on the coastal plain east of the Colville River (Jorgenson et al. 1997), indicates that lake drainage was most active 1000–4000 years BP. It appears that once drainage has occurred in the initial primary lakes and basin bottom elevations have reached a minimum, further drainage is uncommon. Thus, newly forming shallow thaw lakes are unlikely to be drained.

Ice Aggradation

Ice aggradation includes development of wedge ice in polygonal networks, segregated ice within the polygons, and pingo ice in the deepest parts of the lake where thaw bulbs (close taliks) have developed. Differences in the distribution in the nature and volume of these ice structure depends very much on the original active layer dynamics and thaw bulb development in the primary lake before it is drained. Where water is shallower than 1.5–1.8 m, pond sediments will have a thin (~0.6 m) layer that thaws during the summer. This active layer is important because it provides the volume of material where new segregated ice can develop by upward freezing, thereby limiting the potential for ice aggradation. In contrast, under deeper water (>1.5–1.8 m) a thaw bulb will develop. This thaw bulb greatly increases the volume of material in which segregrated and pingo ice can form during downward freezing, and increases the availability of water that can migrate to the freezing front. Differences in sediment texture, from sandy margins to organic-rich silty centers, also plays a large role in ice development.

The development of ice wedges can occur in newly exposed sediments, or under shallow water that freezes to the bottom during winter. Development of the wedges follows a progression of micro-topographic changes that includes: (1) nonpatterned ground, (2) disjunct low rims, (3) low density, low-centered polygons, (4) high density low-centered polygons, and finally (5) mixed high and low-centered polygons. Analyses of floodplain development on the Colville River indicates that disjunct rims take ~300–500 years to develop, low density, low-centered polygons take ~500–1500 years, and complete development of high density low-centered polygons can take 1500 to–3000 years (Jorgenson et al. 1998). At the last stage, wedge ice typically occupies ~20% of the volume of the top 2 m of soil (Jorgenson et al. 1998), although volumes as high as 40% have been estimated for some old and unusual surfaces on sand sheets with thin (<1 m) loess caps near Prudhoe Bay and Kuparuk (Burgess et al. 1999). Development of this wedge ice deforms the adjacent sediments and contributes to heaving of the land surface. This ice then becomes important to surface degradation because it forms very close to the bottom of the active layer and is highly susceptible to partial or complete degradation after surface disturbance. While wedge ice can develop in both the sandy margins and organic-rich centers, they probably can develop larger volume in the centers because of the more easily deformed organic-rich sediments.

Segregated ice (lenticular, layered, reticulate, ataxitic, and organic-matrix ice) forms throughout the basin after drainage, but the nature and volume is highly dependent on initial conditions. In areas where the water was shallow before drainage, initial freezing and readjustment of the active layer occurs within a winter or two after drainage. Afterwards, the active layer readjusts slowly to changing surface conditions resulting from vegetation development and organic accumulation. During this readjustment, thin layers of ataxitic and reticulate ice are prevalent, but the volume of segregated ice is limited by the thickness of the original active layer under the lakes and the presence of sandy sediments near to the surface (Figure 9). In areas where the water was deep and a thaw bulb had developed, the ice aggrades downward into the unfrozen talik (thaw bulb). Because the freezing is downward, free water is able to migrate to the freezing front, and there is a large volume of material to refreeze, the volume of ice can be much larger than in areas of upward freezing. While organic-matrix, ataxitic, reticulate, and ice are still dominant, layered and vein ice often occur in this situation. Ataxitic, reticulate, and layered ice usually were associated with high (70–80%) ice contents (Figure 15).

Pingo ice is an unusual type of ice formed from during freezing of water in coarse sediments within the closed talik that had formed under the deep water zone in a lake. This excess water is injected under pressure into the weakest area where it freezes. During freezing, heaving forces create large mounds (ice-cored hills). The domed surface is characterized by radial tension cracks, which result from the stretching of the surface sediments, and by abundance of large ice wedges that can form in the easily deformed organic-rich sediments. At one exposure that we sampled where the pingo was 4.5 m above the surrounding surface, the surface of the pingo had 3.5 m of organic-rich lacustrine sediments. The ice structures we dominated by organic matrix and layered ice, with prevalent vertical veins presumably formed during heaving and deformation of the sediments. Pingo ice was encountered at a depth of 3.5 m.

Fundamental to the aggradation of ice in the centers of drained lakes is the size of the thaw bulb. The size depends on both water depth and lake area and, therefore, can be highly variable. The size and sediment composition of the thaw bulb then determines the volume and structure of ground ice development after drainage and the subsequent extent of surface heaving of the basin center.

Measurements of surface elevation across three thaw-lake sequences indicate that segregated, wedge, and pingo ice can cause substantial uplifting of the ground surface. Lake centers with thick accumulations of organic-rich, fine grained materials showed as much as 4.0 m of uplift relative to the bottoms of nearby ponds, even exceeding the elevation of the nearby original land surface (Figure 6). We attribute this uplift to ice segregation at both the downward freezing front in the underlying thaw bulb and the upward freezing front in the readjusting active layer, as well as the expansion of wedge ice. Thus, the central portions of most thaw basins can be uplifted and with time appear to resemble the original old surface covered with tussock tundra. In contrast, our limited surveys revealed that the ice-rich margins of thaw basins were only 0.7 to 1.5 m higher than the bottoms of shallow ponds in adjacent ice-poor basins (Figures 6–8).

Overall, our data indicate that ice aggradation can lead to the formation of the types and volumes of ice that form the conditions for subsequent degradation and thaw settlement. The higher volume of ice in the basin centers allows a relatively high potential for settlement, and examination of aerial photography reveals frequent expansion of thaw lakes into these deposits. The ice-rich centers of the basins tend to be convex in shape, however, creating well-drained conditions that are not susceptible to ponding and internal degradation of ice wedges. Instead, the ice-rich centers tend to be degraded by lateral erosion. Along the basin margins, the presence of sandy sediments near the surface prevents substantial accumulation of ice and heaving of the soil. As a result, soils in these sandy margins are not heaved by ice aggradation, and persist as areas of low-lying and usually flooded terrain.

Secondary Development of Thaw Lakes

The secondary development of waterbodies within the ice-rich basins is by far the most complicated stage; multiple opposing hydrologic and pedologic processes can change surface characteristics and lead to disequilibrium in active layer conditions. Predicting the response of the active layer is further complicated by numerous factors that affect the thermal regime, including changes in surface water, accumulation or disruption of surface organic matter, and ice development below the surface. We obtained few field samples from a basin in this stage of development and the environmental factors involved in shaping these features are not well understood. Following fieldwork planned for 2002, we expect to have a better-defined model of this stage of landscape evolution. Based on the available data, the effects of environmental factors on thaw lake development, and their interactions over space and time, are discussed below.

During secondary thaw lake development, water levels in the basins are affected at multiple scales and can vary over time. The development of low-centered polygons and slightly raised shorelines can impede runoff and cause impoundment of water. At a larger scale, water levels in the margins of the basins can rise due to more rapid ice aggradation and soil heaving in the organic-rich silts in the basin centers. Consequently the previously well-drained margins can become partially flooded, as they become the lowest portions of the landscape. We believe this hydrologic readjustment is the principal cause of the prevalence of small, shallow, rounded waterbodies in concentric rings around the margins of thaw basins. This process is reinforced slightly by minor thaw settlement as the thin subsurface layer of segregated ice in the basin margins is degraded by active-layer readjustment. Finally, short and long-term variability in precipitation can cause water levels in the basins to fluctuate.

Opposing this tendency toward thaw degradation due to impoundment of water is the accumulation of subsurface ice and organic material that can raise the surface and reduce water levels. Our limited data indicates surface accretion (through organic matter and ice accumulation) is on the order of 0.5 mm/yr (Figure 16). The accumulation of organic matter alone can reinforce the elevational differences between pond bottoms and the surfaces of the adjacent wet sedge meadows. For many of the shallow ponds that we observed in the basin margins, the thickness of the tundra organic mat (along the pond shore) alone was sufficient account for the difference in the elevation between the pond bottom and the adjacent wet tundra (Figure 11). Also, ice wedges may develop below the ground surface and under shallow ponds, deforming the surface and creating micro-topographic variation that can alter drainage.

The complexity in the development of ground ice, in terms of type, depth, and volume creates a range of initial conditions that can lead to divergent pathways of basin development. While the dynamics of secondary lake development need further study, photogrammetric analysis of changes and the evaluation of the patterns of pond distribution indicate several recurring patterns. First, small ponds can develop from impoundment of water around the margins of old basins and be reinforced by slight thawing of ice beneath the active layer. Second, where the small ponds develop adjacent to the ice-rich centers, they can expand into the centers by thawing the ice-rich materials. These new deep lakes that develop in the original deep portion of the primary lakes (ice-rich centers) can be considered true "thaw" lakes. Third, these larger deeper lakes can continue expansion in the ice-rich materials within the basin, but erosion (primarily mechanical) of higher and older surfaces (alluvial-marine, alluvial terrace, and eolian sand sheet deposits) is minor. Fourth, these newly formed lakes have low likelihood of being drained because they are at a lower elevation and usually well within the confines of an already drained basin. Without drainage, another "cycle" is unlikely to be initiated.

Basin Stabilization

While most basins continue to have active secondary development of thaw ponds resulting in the reestablishment of a deep lake in the ice-rich centers, there is also another old stage of basin development that is not fully understood. In this final stage of basin development, the small, shallow, rounded thaw ponds that have developed within the basins become very persistent features, as indicated by the prevalence of small ponds in indistinct, old basins. They are too small, and the shorelines have too much thick fibrous peat, to be susceptible to wave erosion. In addition, we see no evidence of expansion by thermokarst. Instead, the waterbodies appear to be contracting based on visual comparison of the 1948 and 2001 photography. While we have little data to evaluate this stage of basin development, we speculate that lack of sufficient segregated ice and rapid organic material accumulation across the landscape helps stabilize the surface materials.

During this stage, organic material is added both as algal material in diatomaceous benthic mats within the shallow ponds and as sedge peat in the wet meadows adjacent to the ponds. The surface organic layer within the ice-rich thaw basins typically is >0.5 m deep, and the accumulation of additional subsurface organic material averages nearly 1 m (Figure 11). In numerous soil profiles in the wet tundra adjacent to the ponds in ancient basins, the presence of olive-green limnic material in subsurface horizons indicates that the wet meadows have developed from the thaw ponds. Eventually, the accumulation of organic matter can raise the surface, altering the surface conditions for plant growth and favoring development of moist tundra vegetation. During the accumulation of this organic matter, the depth of the active layer decreases so that some of this organic material becomes incorporated into the permafrost. While this stage is susceptible to erosion from large adjacent lakes, it does not appear to be susceptible to thaw degradation due to intrinsic instability.

More information needs to be collected from basins at this stage of development to evaluation factors leading to stabilization or persistence of the small shallow ponds in the oldest basins. We speculate that incorporation of this organic material makes the surface less sensitive to thaw settlement during active layer readjustment. Thus, more information is needed on the thaw settlement characteristics of fibrous peat. In addition, little is known about the ice structure and volume beneath and adjacent to infilling ponds. Finally, the role of thaw bulb development, or lack thereof, may be important. In shallow basins, where a thaw bulb never developed there may be insufficient ice to allow secondary development of thaw lakes within the basins. This stage has not been recognized by previous investigators.

Summary

Our examination of the development of thaw ponds, based on topographic profiles, stratigraphic analysis, radiocarbon dating, photogrammetric analysis, and regional comparisons, reveals that lake evolution on the coastal plain is much more complex than previous investigators have envisioned. Analysis of the changes in surficial materials, and the rates at which these materials change, also indicate that evolution of the lake basins is less cyclic than previously thought.

In our interpretation, lakes over most of the Arctic Coastal Plain were formed initially by flooding of low-lying terrain during the beginning of the Holocene, when the surface stabilized and climate ameliorated from the cold and dry conditions of the late Pleistocene. Radiocarbon dating of basal peats indicate that the sand sheets stabilized 8–10 thousand years ago and that most lake basins appear to be less than 9 thousand years old. Shoreline erosion, differential transport of fine-grained sediments, and accumulation of organic matter (disseminated peat and algae) in the deep central portion of the lake basins have substantially changed the properties of material near the surface. This reworking of the original sandy deposits and sediment redistribution is required to allow sufficient development of ground ice and the subsequent development of thaw lakes.

Drainage of the lakes, through the development of a drainage network over a lengthy period, created the conditions for permafrost expansion in newly exposed sediments. Radiocarbon dating of basal peats in drained basins and lacustrine strands indicate most basins drained 1–4 thousand years ago. The nature and volume of ground ice that developed in the newly exposed sediments are highly variable across the basins depending on the texture of the redistributed sediments and on variations in water depths and thaw bulb development in the former lakes. Thaw lakes then develop in these ice-rich basins, particularly in the center of old basins where sediments are organic-rich and silty and thaw bulbs had previously developed. Lake development in the flat basins is complicated, however, because surface water conditions are altered by small-scale development of ice-wedge polygons and larger-scale impoundment of water around the margins, due to heaving of the centers of the basins. Photogrammetric analysis of erosion rates indicates that the larger secondary thaw lakes maybe as much as 2 to 3 thousand years old. The common conception of a “thaw-lake cycle” primarily describes this dynamic stage.

We also have identified a final, or at least an alternative pathway, of basin development, where abundant small ponds, become stabilized in very old indistinct basins. Stabilization of the ponds may be due to lack of sufficient ice development in basins that had never developed a thaw bulb, or to accumulation of thick peat deposits.

The concept that thaw-lakes continually migrate across the surface of the coastal plain and that soil materials go through a complete cycle from old upland surfaces to thaw lakes back to conditions similar to the original is not consistent with our analysis of surficial materials and patterns of ground-ice development. Furthermore, the rates at which lakes erode and drain, and at which ground ice develops, are too slow for the entire landscape to have been reworked by multiple cycles during the Holocene. Instead, we conceptualize a landscape which is altered by climatic changes, reworking of surficial materials in topographically constrained lakes, development of integrated drainage networks, drainage, and differential development of ground ice within the basins. This sequential development of unique circumstances formed the conditions for the secondary development of thaw lakes within larger lacustrine basins that currently are widespread across the landscape today.

THAW SETTLEMENT AND TERRAIN SENSITIVITY

THAW SETTLEMENT

We developed estimates of potential thaw settlement based on (1) the amount of excess segregated ice, (2) the amount of wedge ice, and (3) the depth of soil incorporated into the active layer to achieve a new thermal equilibrium (Figure 19) following disturbance. Although ice distribution was highly variable both vertically and

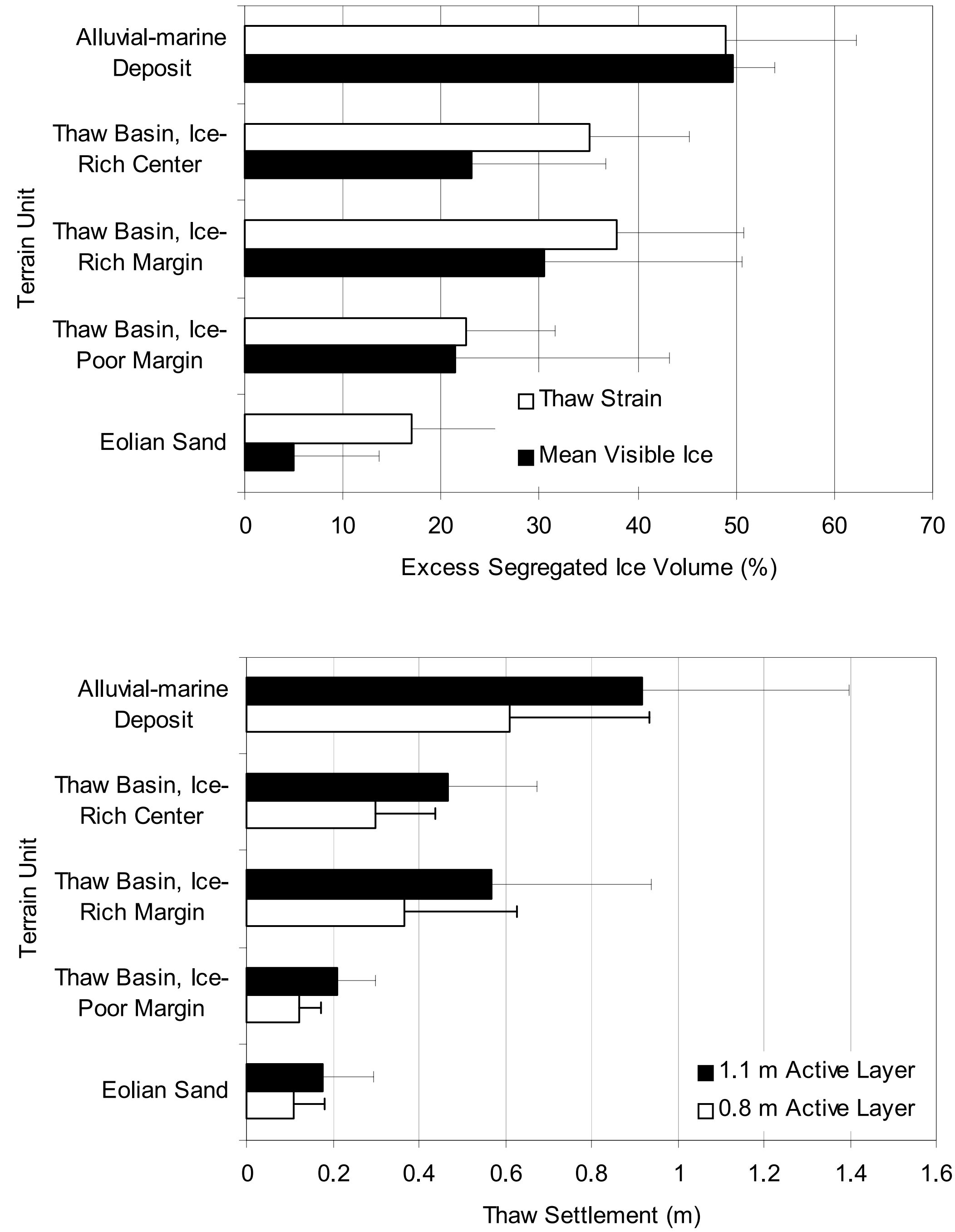


Figure 19. Estimates of mean excess segregated ice volume (±SD), ice wedge volume, and potential thaw settlement due to melting of segregated ice (±SD) in each of five terrain units in the NPRA Study Area, northern Alaska, 2001.

horizontally, there were some clear differences in potential for thaw settlement among terrain units. These differences allowed us to make some generalizations that should be useful for assessing the sensitivity of the terrain to disturbance and for predicting surface responses during future rehabilitation of oilfield facilities. In the following section, we present (1) the assumptions used in the developing estimates of potential thaw settlement, and (2) estimates of thaw settlement associated with segregated ground ice obtained from field samples, followed by (3) conceptual models of the overall potential for thaw settlement in the different terrain units.

We assumed a maximum thaw depth of 1.1 m, after re-establishment of thermal equilibrium in the active layer following a severe surface disturbance (removal of the vegetated surface) . This assumption was based on the maximum thaw depths observed at scraped reserve pits in the Kuparuk Oilfield (Burgess et al. 1999). However substantially lower thaw depths (mean ± 1 SD) were observed at other disturbed sites, including the S.E. Eileen Exploratory Well Site (80 ± 16 cm, 8 years after scraping to the tundra surface (Bishop et al. 1999); the 2U oil spill (46 ± 7 cm, 7 years after scraping to the tundra surface (Cater et al. 1999), and the DS-3O overburden caps (60-70 cm, 7 years after capping (Cater and Jorgenson 1996). In addition, the mean thaw depth in sediments underlying shallow ponds in the Kuparuk oilfield drill sites was 60 ± 31 cm Burgess et al. 1999). Given this range of values, we believe that 1.1 m represents the maximum likely extent of settlement that could occur after severe surface disturbance. For comparison, we also calculated thaw settlement using an expected thaw depth of 0.8 m, which is more typical for disturbed sites without impounded surface water.

Assuming an equilibrium active layer depth of 1.1 m, the maximum thaw settlement from the thermal degradation of segregated ice is expected to be 0.92 ± 0.48 m in alluvial-marine deposits, 0.47 ± 0.20 m in ice-rich thaw basin centers, 0.57 ± 0.37 m in ice-rich thaw basin margins, 0.21 ± 0.09 m in ice-poor thaw basin margins, and 0.17 ± 0.13 m in eolian sand (Figure 19). Based on an active-layer readjustment to 0.8 m, which is more typical of non-flooded highly disturbed surfaces, thaw settlement is expected to be 0.61 ± 0.32 m in coastal plain, 0.30 ± 0.14 m in ice-rich thaw basin centers, 0.37 ± 0.26 m in ice-rich thaw basin margins, 0.12 ± 0.05 in ice-poor thaw basin margins, and 0.11 ± 0.07 m in eolian sand. The relatively low predicted thaw settlement for the ice-rich centers appears at first to be inconsistent with our concept that this is the most ice-rich terrain unit. The thaw basin centers are unusual, however, in that they usually contain accumulations of sand within the top 1.5 m that help stabilize the active layer. In addition, organic soils have smaller thaw strain values without heavy loading. There is a substantial uncertainty associated with some of the estimates, particularly for ice-rich thaw basin centers, because of high organic content and spatial variability in the volume of segregated ice.

The relative abundance of ice wedges is of particular importance to the progression of thermokarst. Although we did not make a systematic assessment of ice wedge volumes across terrain types in the NPRA, we were able to rely on estimates based on an exposure at Mine Site F (Jorgenson, pers. obs.), trenching studies (Everett 1980), and air photo analysis of the Kuparuk Oilfield (Burgess et al. 1999). Because these massive ice bodies occur very close to the bottom of the active layer, increase of the active layer depth after surface modification always causes degradation of the ice-wedges. Small increases in heat flux lead immediately to thaw settlement, which leads to impoundment of water, which in turn leads to increased soil heat flux and additional thermokarst. These processes are likely to cause complete or nearly complete loss of the ice within months to years. Based on a compilation of the available data, we predict that all wedge ice in the top 2 m would be lost, leaving a highly polygonized surface. We predict that the loss of volume due to thawing of ice wedges will be about 20% for old alluvial-marine deposits, 15% for ice-rich thaw basin centers with well-developed low-centered polygons, and negligible for ice-poor thaw basins.

Observations of aerial photographs, exposure on lakeshores, and soil cores suggest that thaw settlement potential in ice-rich thaw basin centers may have been underestimated in this analysis. At a number of core sample sites (4 of 5) we did not encounter the underlying sand sheet due to the depth of lake deposits and ice present in the soil column.

CONCEPTUAL MODEL OF THERMOKARST DEVELOPMENT

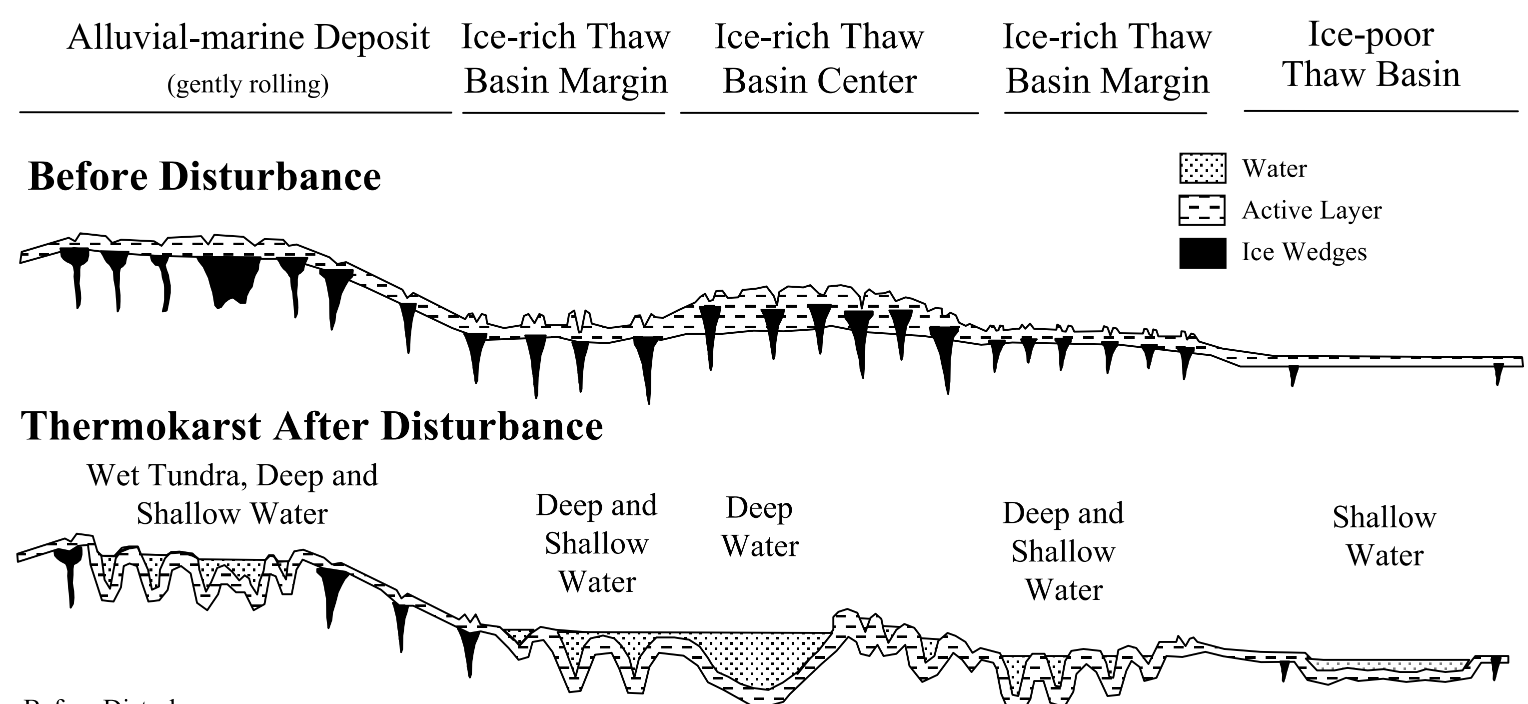
Based on the differences in potential thaw settlement among terrain units we developed a simple conceptual model of terrain response to severe disturbances, such as scraping of the surface or complete removal of the vegetative cover. In this model, the degradation of both segregated and wedge ice results in a mosaic of polygonal troughs and high-centered polygons, depending on terrain type (Figure 20). In alluvial-marine deposits, which have the highest volumes of both segregated ice and wedge ice, the thermokarst micro-topography will be highly irregular, with deep and shallow troughs and prominent high-centered polygons. Because these terrain units occupy slopes and gently rolling uplands between thaw basins, some surface drainage and lowering of the water table is likely. In most cases, this will result in most of the polygon centers being above the water table. In ice-poor thaw basins, where segregated ice volumes are much lower and wedge ice is negligible, the thermokarst topography is likely to be fairly uniform after only a moderate amount of thaw settlement. However, the surface will remain flooded because this terrain unit occurs in the lowest portions of the basins. Thus, even minor thermokarst will result in the development of large, shallow ponds.

The conceptual model is consistent with thermokarst we have observed at numerous sites. At sites on alluvial plains, such as Sinclair Exploratory Well Site (Bishop 1998) and N.W. Eileen State No. 1 (Jorgenson and Cater 1993), thermokarst, settlement and partial drainage have resulted in a highly prominent relief mosaic with deep and shallow water in troughs and patches of wet and moist tundra on the tops of the polygon centers. At a site in an ice-rich thaw basin near the Prudhoe Bay Operations Center, where ~20 cm of gravel was left after gravel removal in 1988, thermokarst has resulted in shallow water over the tops of polygon centers and deep water in the troughs (Kidd and Rossow 1998). At the S.E. Eileen Exploratory Well Site (Bishop 1999), on an inactive floodplain of the Kuparuk River, thermokarst has resulted in mostly moist and wet high-centered polygons with only limited occurrences of shallow water in troughs. This site is somewhat unusual in being located next to the riverbank, so drainage is better than on most flat inactive floodplains. At sites in ice-poor thaw basins, such as the abandoned access road to Drill Site 3K in the Kuparuk Oilfield (Cater and Jorgenson 1993) and the abandoned access road to the Operations Storage Pad in Prudhoe Bay (Kidd and Rossow 1998) where thick gravel was removed from the tundra surface, thermokarst has resulted in level flooded topography with shallow ponds or wet meadows, depending on water depth.

While there were substantial differences in estimated potential thaw settlement among terrain units, the high variability in the estimates indicates that predicting the amount of settlement at a specific location is difficult. Furthermore, the amount of settlement is very sensitive to the amount of thaw that occurs as the active layer reaches a new thermal equilibrium and to the extent of surface disturbance. The value of 0.8 m for the equilibrium thaw depth was based on thaw depths found at some of the more highly disturbed sites, however, the actual increase in thaw depth is likely to vary considerably among sites. Unfortunately, little is known about predicting adjustment of the active layer to disturbance as the numerous factors associated with the energy balance interact during changes in vegetation, soil moisture and hydrology as the surface settles. While these factors make modeling and prediction difficult, we believe our conceptual model based on knowledge of the complexity of ground ice characteristics and simple assumptions of active layer behavior provides reasonable predictions of surface change after severe disturbance. The concepts are consistent with observations at a number of disturbed sites, helping to validate use of the model for land management decisions on the Arctic Coastal Plain.

IMPLICATIONS FOR LAND MANAGEMENT

Differences in thermokarst potential in the various terrain units have important implications for land management, facility planning, and the development of site-specific rehabilitation strategies appropriate to changing site conditions after abandonment. However, evaluation of probable surface responses to specific management



	Alluvial-marine Deposit	Ice-rich Thaw Basin Margin	Ice-rich Thaw Basin Center	Ice-rich Thaw Basin Margin	Ice-poor Thaw Basin
Before Disturbance					
Active Layer (m)	0.3	0.4	0.3	0.4	0.4
Segregated Ice (% vol)	71	64	64	64	60
Wedge ice(% vol)	20	15	20	15	0
After Disturbance					
Active Layer (%)	0.8	0.8	0.8	0.8	0.8
Mound Settlement (m)	0.6±0.3	0.4±0.3	0.3±0.1	0.4±0.3	0.4±0.3
Trough Settement (m)	1-2	1-2	1-3	1-2	0
Drainage	partial	none	none/partial	none	none

Figure 20. Conceptual model of thermokarst development after disturbance of ice-rich permafrost on the dominant terrain units in the NPRA Study Area, northern Alaska, 2001. Typical values are given for active layer depths before and after severe disturbance, volume of excess segregated and wedge ice before disturbance, estimated thaw settlement of mounds or flat surfaces from degradation of segregated ice and settlement in troughs due to degradation of wedge ice, and potential for drainage.

decisions must also consider the nature and extent of water impoundment after thaw settlement, as affected by slope position on the landscape. Some management implications are discussed below.

Appropriate siting of facilities, roads, and cross-drainage structures in relation to thaw stability of the terrain can greatly reduce potential problems with both thermokarst and roadbed performance. While the thaw settlement estimates indicate eolian sand sheets and ice-poor thaw basins have the most thaw stable conditions for development activities, the evaluation is complicated by landscape position. For example, ice-poor thaw basins have low potential for thaw settlement, but even with little settlement these areas tend to become uniformly flooded because of their position in basins. In contrast, thaw settlement estimates are high for upland alluvial-marine deposits, but these are probably better surfaces for road and pad placement because they are better drained. Good drainage reduces the need for cross-drainage structures, with their associated impacts. Culverts on this type of terrain can cause localized thaw degradation, but the relatively good off-site drainage tends to minimize propagation of the disturbance. Perhaps the least favorable terrain for siting facilities is ice-rich thaw basin centers. While estimates indicate that thaw settlement should be small, this terrain is susceptible to severe degradation as indicated by the prevalence of deep thaw ponds in these deposits.

Estimates of potential thaw subsidence (along with other landscape characteristics) also are crucial for directing oil spill response and cleanup operations. Oil spill cleanup often requires weighing the potential risk of thermokarst against the cost and efficiency of the oil recovery techniques. Knowledge of the thaw stability can reduce the risk of thermokarst in ice-rich terrain if appropriate precautions are taken during cleanup operations. In addition, cleanup efficiency can be enhanced in thaw-stable terrain by the use of more aggressive techniques.

Knowledge of the thermokarst potential of a site can help to ensure the long-term success of rehabilitation efforts and enhance the functional relevance of the rehabilitation plan. Thermokarst is a natural process that is integral to development of the landscape on the Arctic Coastal Plain (Britton 1967, Billings and Peterson 1980, Walker et al. 1980), and increases the terrain diversity and can provide useful wildlife habitat (Murphy and Anderson 1993). Thus, thermokarst can be incorporated into the rehabilitation strategies as a natural phenomenon that can increase habitat diversity and productivity. However, minimization of thermokarst propagation beyond the disturbed site should remain an important objective.

When evaluating revegetation options, thermokarst must be recognized as a critical factor because it fundamentally alters soil and hydrologic conditions. If the goal is a moist, well-drained site that is thermally stable, thermokarst can be prevented by adding overburden material (typically about ~0.6 m fines and organics) (Jorgenson 1986). If a patchy mosaic with varying soil and hydrologic conditions is desired, then thermokarst can be allowed to occur naturally. Plant materials adapted to aquatic, wet, or moist conditions can then be selected depending on the rehabilitation strategy and the amount of thermokarst expected.

If gravel removal is selected as a rehabilitation technique, thermokarst will greatly affect the outcome of the effort. Areas on eolian sand sheets and alluvial-marine deposits will likely become diverse mosaics of well-drained high-centered polygons and flooded troughs. Over time, areas in ice-rich thaw basin margins probably would become large shallow ponds, while areas in ice-rich thaw basin centers would likely become large deep ponds. Areas in ice-poor thaw basins would likely become shallow ponds in wet areas, or uniform wet or moist meadows in better-drained areas.

CONCLUSION

Data on ice structures, lithofacies, and terrain units from 15 cores (2–3 m), revealed strong relationships that can be used to partition the variability in ice distribution across the landscape based on the surface terrain classification. Large differences were found in the frequency of occurrence of the various ice structures among lithofacies: pore ice was nearly always associated with massive, inclined, and layered sands; lenticular and ataxitic ice were most frequently associated with massive and layered fines, and organic matrix ice was usually found in massive and layered organics and limnic fines. Reticulate ice was broadly distributed among fine and organic lithofacies. Among surface terrain units, mean ice volumes were highest in alluvial-marine deposits (71%), intermediate in ice-rich thaw basin margins (64%) and ice-rich thaw basin centers (64%), and lowest in ice-poor thaw basin margins (60%) and eolian sand (54%).

Geomorphological processes associated with thaw lake terrain development primarily control the spatial distribution of ground ice in the region. Evaluating terrain stability in the study area required both assessing the volume and distribution of different forms of ground ice, and estimating the rates of landscape change due to the thaw lake expansion. Photogrammetric analysis of waterbody changes in three small study areas was used to evaluate shoreline erosion during a 46–56 year period from 1945–1955 to 2001. For the three areas combined, 0.74% of the total land area was lost to shoreline erosion over 46–56 years. The average annual erosion rate for the three areas, expressed as a percentage of total area, was 0.04%/yr. Average rates of shoreline retreat were very slow (0.02 m/yr), even for large, deep lakes (0.08 m/yr). The maximum rate of shoreline retreat we observed was 0.8 m/yr. Deep lakes (mean 26.1% of area) were much more common than shallow lakes (mean 6.6%), and shallow lakes typically were small (<5 ha).

We developed a conceptual model of thaw lake evolution that enhances our understanding of landscape stability. The model is based on interpretation of aerial photographs and data collected during the 2001 field season, including topographic profiles, stratigraphic analysis, ice structure studies, radiocarbon dating, and regional comparisons. Our analysis revealed that the process of lake development on the coastal plain is more complex and less cyclic than previous investigators believed. Our revised conceptual model for the portion of the Arctic Coastal Plain underlain by extensive sand sheets includes: (1) initial flooding of primary lakes, (2) lateral expansion and sediment accumulation and redistribution, (3) lake drainage, (4) ice aggradation in exposed sediments, (5) secondary development of thaw lakes, and (6) basin stabilization.

Effective land management on permafrost-dominated terrain requires terrain-specific estimates of the potential for thermokarst following disturbance. Knowledge of the likely extent of thaw settlement across the landscape is essential for evaluating facility locations, and effective rehabilitation planning, including surface treatments and selection of appropriate plant materials. This information is also valuable in planning and directing effective and efficient oil spill cleanup and remediation activities. We developed estimates of the amount of thaw settlement likely to occur in various terrain types after severe disturbance. Based on an active-layer readjustment to 0.8 m, which is typical of non-flooded highly disturbed surfaces, mean (± SD) thaw settlement is expected to be 0.61 ± 0.32 m in alluvial-marine deposits, 0.30 ± 0.14 m in ice-rich thaw basins centers, 0.37 ± 0.26 m in ice-rich thaw basin margins, 0.12 ± 0.05 m in ice-poor thaw basin margins, and 0.11 ± 0.07 m in eolian sand. Based on data from the literature and our field observation we estimated that all wedge ice in the top 2 m would be lost, leaving a highly polygonized surface. The expected loss of volume due to thawing of ice wedges is about 40% for old alluvial-marine deposits, 15% for ice-rich thaw basins with well-developed low-centered polygons, and negligible for ice-poor thaw basins.

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