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Abstract

The dominant landscape process on the Arctic Coastal Plain of northern Alaska is the formation and drainage of thaw lakes. Lakes and drained thaw lake basins account for approximately 75% of the modern surface expression of the Barrow Peninsula. The thaw lake cycle usually obliterates lacustrine or peat sediments from previous cycles which could otherwise be used for paleoecological reconstruction of long-term landscape and vegetation changes. Several possible erosional remnants of a former topographic surface that predates the formation of the thaw lakes have been tentatively identified. These remnants are characterized by a higher elevation, a thick organic layer with very high ground ice content in the upper permafrost, and a plant community somewhat atypical of the region. Ten soil cores were collected from one site, and one core was intensively sampled for soil organic carbon content, pollen analysis, and ^{14}C dating. The lowest level of the organic sediments represents the earliest phase of plant growth and dates to ca. 9000 cal BP. Palynological evidence indicates the presence of mesic shrub tundra (including sedge, birch, willow, and heath vegetation); and microfossil indicators point to wetter eutrophic conditions during this period. Carbon accumulation was rapid due to high net primary productivity in a relatively nutrient-rich environment. These results are interpreted as the local response to ameliorating climate during the early Holocene. The middle Holocene portion of the record contains an unconformity, indicating that between 8200 and 4200 cal BP sediments were eroded from the site, presumably in response to wind activity during a drier period centered around 4500 cal BP. The modern vegetation community of the erosional remnant was established after 4200 cal BP, and peat growth resumed. During the late Holocene, carbon accumulation rates were greatly reduced in response to the combined effects of declining productivity associated with climatic cooling, and increased nutrient stress as paludification and permafrost aggradation sequestered mineral nutrients.

Keywords: Alaska, arctic environment, Holocene, permafrost, pollen analysis, soil organic carbon, tundra

1. Introduction

The Arctic Coastal Plain of Alaska (Wahrhaftig, 1965) is a landscape scarred by dynamic cryogenic processes and, above all, by the development and evolution of thaw lakes. A suite of geomorphic processes govern the thaw-lake cycle, including the formation, growth, orientation, and eventual drainage of the ubiquitous lakes in this region. On the Barrow Peninsula north of the 71st parallel, approximately 22% of the land area is covered with thaw lakes and ponds developed atop permafrost; an additional 50% is composed of thaw-lake basins formerly occupied by lakes (Sellman et al., 1975; Hinkel et al., 2003). Thaw lake features, therefore, account for nearly three-quarters of the modern surface expression of the region. Lake formation and drainage processes have likely been operating at least since the end of the Late Glacial (Ritchie et al., 1983; Hopkins and Kidd, 1988; Rampton, 1988), but the dynamic nature of the thaw-lake cycle has made it difficult to study landscapes that existed prior to the present thaw-lake cycle. The processes associated with the thaw lake cycle usually obliterate lacustrine or peat sediments from previous cycles, which could otherwise be used for paleoecological reconstruction. As a result, long-term landscape and vegetation changes are imperfectly understood.

There remains about 28% of the landscape that appears to have been unaffected by thaw-lake processes. Much of the region undoubtedly represents old drained lake basins that cannot be readily identified because basin boundaries (old lake shorelines) have been obliterated through time. Using Landsat 7+ imagery of the Barrow region (Figure 1), we used the spectral signature and textural variation of the vegetation to identify and locate regions unaffected by the thaw-lake cycle (Frohn et al., 2001; Hinkel et al., 2003; Bockheim et al., 2004). The morphology of some features indicates that they are former barrier islands or spits that emerged following recent marine regression (Brigham, 1985; Brigham-Grette and Hopkins, 1995; Figure 2). However, several inland sites have substantially higher elevations, a deep organic layer with very high ice content in the upper permafrost, and a plant community not typical of the area (Hussey and Michelson, 1966; Bockheim et al., 2004).

2. Background

The Coastal Plain has been subjected to several marine transgressions during the late Cenozoic (Brigham-Grette and Carter, 1992; Brigham-Grette and Hopkins, 1995; Brigham-Grette, 2001). Hopkins (1973; 1982) and others (e.g., Péwé, 1975; Carter, 1993) have identified ancient shorelines, dunes, and wave-cut scarps that can be discontinuously traced across northern Alaska. The Pelukian transgression (isotope substage 5e; 123 ka), evidenced as former beach ridges now at an altitude of 10-13 m asl, is associated with the Sangamon interglacial highstand. The coastal region of the landscape was again inundated to a depth of about 7-8 m during the Simpsonian interstadial transgression (isotope substage 5a/4, ~58-75 ka), and this event is associated with deposition of the glaciomarine fossiliferous Flaxman member of the Gubik Formation (Brigham-Grette, 2001; Hayden and Brigham-Grette, 2001). More recently, beach ridges located ~2 m asl appear to have been deposited in the late Wisconsinan-early Holocene (Péwé, 1975). INSERT The Barrow Peninsula has remained unglaciated throughout this period and therefore not been directly affected by glacial depression or rebound.

An increase in sea level by 7 m would inundate much of the Barrow Peninsula, leaving topographic prominences as isolated uplands or islands (Figure 2). These sites would potentially have surface organic deposits that might provide a long-term, uninterrupted record of regional climate change. Our most detailed records from the area are collected within drained thaw-lake basins (Eisner et al., 2004), but the oldest basins are only 5500 years old (Hinkel et al., 2003).

The Arctic experienced warmer temperatures than present during the early Holocene and northern Alaska experienced warming before other parts of the Arctic beginning around 11,500 cal BP. (Note: unless otherwise indicated, dates in this paper are given as calendar years before present (cal BP), with the age being derived as the midpoint of the 2-sigma calibrated age range from Stuiver and Remier (1993) and with present being defined as 1950 AD). This period (ca. 11,500 – 9000 cal BP), the Holocene thermal maximum (HTM), has generated particular scrutiny because it may be an analogue to present-day climate warming (Kaufman et al., 2004). The evidence for the HTM is based on a variety of sources and data types. Pollen and macrofossil (wood, leaves) evidence from northern Alaska show that *Populus balsamifera* (balsam poplar) was present beyond its modern range during this interval (Hopkins et al., 1981; Brubaker et al., 1983; Edwards et al., 1985; Anderson, 1988; Mann et al. 2002). Beetles found beyond their modern limits on the Arctic Coastal Plain indicate an increase in summer temperatures of +2 - +3°C around 10,800 cal BP (Nelson and Carter, 1987; Elias, 2000). A recent synthesis investigation of the West Siberian Arctic shows “explosive, widespread peatland establishment between 11.5 and 9 thousand years ago” (Smith et al., 2004).

Organic materials began to accumulate ca. 13,000 cal BP at Imnavait Creek in the Arctic Foothills physiographic province (Eisner, 1991), and ca. 10,800 cal BP at Atqasuk on the Meade River bluffs about 120 km SSW of Barrow (Eisner and Peterson, 1998b). Carbon accumulation rates (CARs) were high at the Meade River site throughout the early Holocene and slowed after 8300 cal BP, near the end of the HTM. Dunes in the Meade River region have a well-developed near-surface buried soil dating to ca. 9200 cal BP. This presumably developed in response to moister and warmer conditions, enabling re-vegetation and stabilization of the sand surface (Carter, 1993). Subsequently, the soil was buried by renewed aeolian activity after ca. 5600 cal BP.

Although the period following the HTM has not been examined with the same intensity, there is evidence of cooler and possibly drier conditions on the Arctic Coastal Plain after 8000 cal BP. CARs for arctic Alaskan soils show a striking temporal variation at Prudhoe Bay, ranging from 7.9 g C m^{-2} yr^{-1} in the mid-Holocene (8000 to 5000 cal BP) to 1.4g C m^{-2} yr^{-1} during the late Holocene (ca. 5000 cal BP to present) (Marion and Oechel, 1993). In the Meade River area, peat accumulation was high during the mid-Holocene, and slowed after reactivation of sand dunes (Carter, 1981; Galloway and Carter, 1993; Eisner and Peterson, 1998b). During the late Holocene, the Coastal Plain experienced overall cooling, with intermittent periods of dune reactivation and generally drier conditions after 4500 cal BP (Eisner, 1999).

Only a few sites on the Arctic Coastal Plain of Alaska provide convincing palynological evidence of the HTM (Nelson and Carter, 1987; Eisner and Peterson, 1998b), and these are both derived from peat deposits. Tundra vegetation lacks sensitive indicators of climatic variation, which are more prevalent at treeline. Further, the dynamic nature of the landscape, dominated as it is by freeze-thaw cycles, development of thermokarst, and thaw-lake processes, renders most sediment accumulation areas (lakes and peatlands) susceptible to cryoturbation, disruption, and obliteration. Finally, pollen analysis of tundra vegetation tends to be an insensitive proxy for examination of events such as the shift from wet sedge tundra to a shrub tussock-tundra. These potentially crucial landscape changes may not be strongly reflected in pollen assemblages.

Holocene landscape and vegetation changes on the Arctic Coastal Plain are therefore imperfectly understood, despite years of intensive study. Where undisturbed sediments are available, preferably in the form of peat layers preserved in permafrost, and when pollen analysis can be coupled with microfossil analysis and changes in soil carbon storage, reliable evidence of landscape change becomes obtainable (Eisner and Peterson, 1998b).

The objective of this study was to combine geomorphic, palynological, and organic carbon analyses to study paleoenvironmental changes at a site on the Arctic Coastal Plain that was not influenced by thaw-lake processes.

3. Site Description

The Arctic Coastal Plain is developed in unconsolidated Quaternary marine sediments of varying thickness. Known as the Gubik Formation, these sediments lie unconformably on nearly flat Upper Cretaceous sedimentary rock (Black, 1964; O'Sullivan, 1961). Marine sands cover most of the Barrow Peninsula, although marine silt occurs at the surface as a band along the eastern coast (Figure 2). These marine deposits are often capped by silt.

We focused our activities on one erosional remnant (informally named: Peterson Erosional Remnant; PER; 71°04' N, 156°24' W; Figures 1 & 2), lying 36 km south of Point Barrow near the center of the Barrow Peninsula. The feature has an elongated triangular shape, with the long axis oriented WNW and extending for about 4.5 km. The maximum width of 1.3 km is near the southern end. The broad, flat upland is at an altitude of about 20 m asl, and stands 15-16 m above the surrounding thaw lake-effected landscape; the slopes of the PER are very steep for this area of the Coastal Plain. The surface exhibits high-frequency microrelief of about 0.5 m owing to the presence of small (2-5 m diameter) high-centered ice-wedge polygons. By comparison, nearby drained lake basins are characterized by low- and flat-centered polygons with diameters of 20-30 m.

The mean annual temperature at Barrow, a coastal site 30 km to the north, is –12.0° C (NCDC, 2002); mean monthly temperatures range from 4.7° C in July to –26.6° C in February. Most of the annual precipitation (106 mm) falls as rain in summer; winter snow accumulation ranges from 20-40 cm. The climate becomes more continental inland, with warmer summer temperatures and reduced precipitation (Clebsch and Shanks, 1968; Haugen and Brown, 1980). Given a geothermal gradient of 3°C/100 m, permafrost thickness is estimated to be around 400 m thick in the region. The active layer, or seasonal thaw zone, was measured at the PER in mid-August 2002 and averaged 27 cm.

Vegetation on the PER is typical moist acidic tussock tundra dominated by a sedge-shrub assemblage consisting of *Salix glauca* (willow), *Cassiope tetragona*, *Eriophorum vaginatum*, and *Ledum palustre*, with non-vascular plants such as *Dicranum elongatum*, *Sphagnum* and lichen species. Ice-wedge troughs contain *Carex aquatilis* (sedge). The presence of *Ledum palustre* on the PER is unique for the Barrow peninsula. In a study of the Barrow area vegetation, *Ledum* was not part of the modern vegetation (Webber, 1978). By contrast, vegetation in the lowlands surrounding the PER is dominated by *Carex aquatilis* and the tundra is a highly patterned mosaic of wet and dry vegetation with mesic components restricted to narrow boundaries between the two.

A total of ten soil cores were examined on the PER within a 200 m radius. Soils ranged from 37 to 150 cm thick, and were composed of highly decomposed organic matter (sapric material) over segregated ice and/or gleyed mineral soil. Using the *Keys to Soil Taxonomy* (1998), the soils were classified as either Sapric Glacistels or Glacic Sapristels.

4. Field And Laboratory Methods

Two exploratory cores were taken in August 2001 using a gas-powered post-hole digger. These cores (PER 01-1&2 in Table 1) indicated the presence of thick peat and massive segregation ice in the upper permafrost. In contrast, the fine-grained soils near Barrow contain abundant ice as both pore ice and segregation ice lenses. Sellman et al. (1975) reported that the volumetric ice content in the upper 2 m of permafrost in the Barrow region averages 50-75%, and ice wedges may account for an additional 10-20%. Ice lens growth, or formation of ice layers developed from the lateral injection of water, causes the sediments to expand vertically. This "frost heaving" displaces the ground surface upward, and may have contributed to the topographic expression of the PER. The internal growth of ice layers also disrupts the sedimentary sequence by post-depositional inflation of the stratigraphy. This complicates age-depth relations and makes calculation of deposition rates problematic.

In an effort to acquire a non-inflated, non-cryoturbated, continuous core of the organic layer and underlying mineral sediments, eight additional cores were collected from the PER in April 2002 and 2003 (samples PER 02 and PER 03, respectively). To obtain sufficiently deep cores, heavy drilling equipment was required to prevent barrel freezing during the drilling operation. We used a Big Beaver™ earth drill equipped with a 7.5-cm-diameter SIPRE core barrel. The drill and motor are permanently mounted on a specially designed sledge and towed by snowmachine. Cores were collected from the crest of the PER, and snowcover obscured the microrelief. In general, however, they were taken from within polygon centers to avoid ice wedges.

All of the cores were described according to soil horizon, and the amount of segregation ice was estimated visually. The cores were photographed and processed in a cold room at

Barrow. Core PER 02-3 contained mostly massive segregation ice and was not subject to further analysis. Three cores (PER 02-1, PER 02-2, and PER 02-4) were analyzed for determination of moisture content and bulk density. These were separated into soil horizons, and cut with a chop-saw into 10-cm sections for processing. The sections were oven-dried at 70˚C. Moisture content was determined, and bulk density was calculated on an oven-dry basis. The degree of decomposition of organic horizons (e.g., fibric, hemic, or sapric) was estimated from rubbed and unrubbed fiber contents (Lynn et al., 1974), and soils were classified using the *Keys to Soil Taxonomy* (Soil Survey Staff, 2003).

One promising core (PER 02-5) was selected for intensive analysis, described, photographed, and sliced into 1-2 cm sections. Each resulting disk was split for (1) pollen analysis and radiocarbon dating, and (2) determination of total organic carbon (TOC). Total organic carbon (TOC) was measured on 78 samples from cores PER 02-5 and adjacent core PER 02-4, which are separated on the landscape by about 1 m. Samples were passed through a 100-mesh (149 μm) sieve. Total carbon was measured on a Dohrmann DC190 C analyzer. None of the soil samples reacted with 1 M HCl, indicating that the values represent organic carbon exclusively.

Percent TOC of Core PER 02-5 was determined at 1-2 cm intervals for the entire core length. Carbon storage (kg m^{-2}) was calculated as the product of sampling depth, bulk density, and TOC using equations in Michaelson et al. (1996). Interval-specific TOC accumulation rates (g m^{-2} y^{-1}) were determined by dividing the soil carbon pool for a particular pollen zone by the time interval represented by that zone, as established from radiocarbon dating. The mean long-term net accumulation of TOC for the PER was estimated by dividing the soil carbon pool of the entire section by the basal age.

Peat samples were sieved and plant material, mostly sedge or moss leaf epidermis, was selected, pretreated, and dated by radiocarbon accelerator mass spectrometry (AMS) at Lawrence Livermore National Laboratory. Pollen samples were prepared using standard procedures (Berglund and Ralska-Jasiewiczowa, 1986; Moore et al., 1991). Identification of algal, fungal, and zoological remains was carried out using an extensive collection of photographs and reference material (http://www.ngdc.noaa.gov/paleo/parcs/atlas/beringia/fungus.html). Descriptions and environmental significance of these fossils are from van Geel (1978; 1986), van Geel et al. (1981; 1989), and Eisner and Peterson (1998a; 1998b).

5. Results

5.1. Ice content of cores

If the degree of ice-wedge polygon development is more advanced on the PER owing to its greater age, near-surface permafrost should be enriched with ground ice (Hinkel et al. 2003; Bockheim et al. 2004). This would result from the lateral or vertical migration and concentration of water, which eventually freezes to form lenses or layers of massive ground ice. Meteoric water can also penetrate into the permafrost from above. Thermal microcracks that form in winter serve as conduits for snow meltwater infiltration in spring. This process is evidenced by the presence of ice veins or sills in the frozen active layer and upper permafrost.

As shown in Table 1, massive ground ice was encountered in nine of the ten cores. Although ice lenses dominated, vertical foliation indicated the presence of ice wedges in several cases. The thickness of the massive ice depended on the length of the core collected, which was dictated by the time of year that the core was collected. A relatively short (1.0 m) core barrel was used in summer 2001 to prevent barrel freezing, while a longer (1.35 m) barrel was used in winter when deeper drilling with the heavy-duty drill was possible. Typically, if thick ice was encountered, the coring operation was abandoned in favor of a new site.

The data in Table 1 indicate the widespread occurrence of massive ground ice, which was encountered at a frequency far exceeding that typical of sampling sites on the Coastal Plain. In thaw-lake basins on the Barrow Peninsula, for example, massive ice was typically encountered only as ice wedges or as thin (<5 cm) ice lenses; and the amount of massive ice in the near-surface permafrost tends to increase with basin age (Hinkel et al., 2003; Bockheim et al., 2004). These findings suggest that the PER has been subject to ice-enrichment processes for a much longer time period than the surrounding region, and consequently is a much older surface. Because the base of the massive ice layer was not encountered in most cases, it is likely that frost heave has contributed to the topographic prominence of the PER. The magnitude of heave cannot be evaluated without collecting deeper cores or use of ground-penetrating radar (Hinkel et al., 2001).

5.2. Textural analysis of the mineral horizon

Including Core PER 02-4, which lacked massive ice, the average soil thickness of the ten cores in Table 1 was 68 cm (standard deviation of 38 cm). Textural analysis of the mineral sediments at the base of Core PER 02-4, sampled from the Cg horizon immediately beneath the organic layer (depth = 83 cm) is shown in Figure 3. The particle size analysis indicates that the parent material of the basal mineral sediment is a sandy loam. The high degree of sorting, lack of coarse particles, and the dominance of the fine sand (34%) and silt (33%) fractions, suggest an aeolian origin.

5.3. Stratigraphy, carbon content, and dating of Core PER 02-5

The near-surface stratigraphy of the PER is distinctive because (1) there is more segregated ice compared to the surrounding landscape dominated by drained thaw-lake basins, and (2) it contains deep, highly decomposed (sapric) organic material. The degree of humification at the PER was comparable to that of the most ancient drained lake basins, whose dates range from 2000 to 5500 cal BP (Hinkel et al., 2003; Bockheim et al., 2004).

The total original length of Core PER 02-5 was 170 cm. However, a 34-cm thick layer of massive segregation ice occurs from 64 to 98 cm. A second ice lens, containing fine suspended organic material, occurred at 118-143 cm depth. This 25-cm thick layer shattered during the coring operation and was impossible to reconstruct. The "inflation" of the sediments due to the migration of water into the section therefore resulted in the vertical extension of the sediments by at least the sum of the thickness of the two ice lenses, or 59 cm. Further analyses require that depth-stratigraphic relations be reconstructed by eliminating the inflation effect. This was done by subtracting the ice thickness from the profile depths and collapsing the core length from 170 cm to 111 cm. Deflated core depths will be used for all subsequent figures and analyses.

The upper 7 cm of the Core PER 02-5 is composed of hemic materials (Oe horizon). Sapric material occurs from 7 to 64 cm depth (Oa1 horizon), and several pebbles (~1 cm diameter) were found between 13 cm and 39 cm. Below 39 cm is ice-rich sapric material (Oa2 horizon) extending from 64 cm to 84 cm. Sapric material continues from 84 to 102 cm, but the ice content is lower. The basal section from 102 to 111 cm is mineral sediment (Cg), and was not sampled for pollen or total organic content.

Percent TOC of Core PER 02-5 is shown in Figure 4a. Interval-specific carbon storage (kg m^{-2}) and TOC accumulation rates (g m^{-2} y^{-1}) are shown in Figures 4b and 4c, respectively. The mean long-term net accumulation of TOC for the PER yielded a value of 15.0 g m^{-2} y^{-1}.

AMS dating of samples collected from the organic sediments indicates that the record spans nearly the entire Holocene (Table 2). Our first set of radiocarbon dates were obtained primarily from wood fragments and showed serious inconsistencies. The oldest date obtained in this first set, 9935±40 ^{14}C yr BP (CAMS #86720), was from the sample at basal depth, the next oldest date, 9180±80 ^{14}C yr BP (CAMS #90994), was obtained from the uppermost sample, near the surface. Anomalous dates from wood fragment samples are not uncommon in arctic peatlands. Oswald *et al.* (2002) observed discrepancies in Arctic sediments, where wood samples were found to yield older dates than other plant materials. Where conditions favor the re-deposition of "old" wood, its influence has resulted in bulk samples yielding ^{14}C dates that are thousands of years older than *in situ* material (Nelson et al., 1988).

Our inability to obtain consistent dates from wood fragment samples motivated greater stringency in sample selection criteria. For establishing the chronostratigraphy of the PER 02-5 core, we relied exclusively on ^{14}C dates obtained from *in situ* peat material that was part of the peat-forming vegetation (Table 2). In view of the highly decomposed (sapric) character of the peat matrix, this limited us to very small samples sizes (<<1 mg) and resulted in some dates having relatively large uncertainties.

The calendar age ranges in the lower portion (30-100 cm depth) of Core PER 02-5 show only a slight trend towards younger ages with decreasing depth. In fact, graphical representation of the cal BP age ranges against depth (Figure 4c) shows that the age ranges for depths of 30-100 cm all fall within a period of a few centuries, suggesting very rapid peat growth which is not unprecedented for the early Holocene Arctic (*e.g.* Ovenden, 1990; Vardy et al., 1997; Eisner and Peterson, 1998b; Lozhkin et al., 2001). Since both the pollen stratigraphy and sedimentological analyses do not indicate mixing or cryoturbation of sediments, disturbance processes cannot be invoked to suggest homogenization of the sediments in this depth zone.

As noted above, the ^{14}C dates show only a slight decreasing trend in age with decreasing depth below 30 cm. In addition, the small sample sizes (and attendant relatively large uncertainties of some dates) lead to relatively large calibrated age ranges for some samples and considerable overlap of 2σ age ranges throughout the 30-100 cm section. Because of these factors, detailed analysis of the age-depth trends within the 30-100 cm depth range is not possible. As a result, the best available estimate of the age span represented by the 30-100 cm sediments is derived from the basal dates and the date obtained from the 30 cm depth sample.

The three basal dates are not statistically distinguishable and have a weighted mean of 7880±60 ^{14}C yr BP, resulting in a basal calibrated 2σ age range of 8540-8980 cal BP (essentially identical to the age range of CAMS #97577). On the basis of this age range, we assign peat initiation at the PER site as occurring between 8540 and 8980 cal BP (midpoint 8760 cal BP). Our confidence in the basal age range for this core is strengthened by the basal date from Core PER 02-1, taken several meters away, which yielded a calibrated age range of 8710-9130 cal BP (midpoint 8920 cal BP). The uppermost level for this period of peat accumulation is 30 cm, with the layers immediately above 30 cm depth providing an indication of a hiatus (see below). The date obtained for the 30 cm sample provides a calibrated age range of 8190-8660 cal BP (midpoint 8420 cal BP) for the uppermost level of peat accumulation. Thus, the peat accumulation represented by the 30-100 cm section of the PER 02-5 core occurred between 8540-8980 cal BP (midpoint 8760 cal BP) and 8190-8660 cal BP (midpoint 8420 cal BP).

The pattern of the more precise dates within the 30-100 cm section can be used to suggest refinements of the basal and 30 cm depth age ranges. In particular, the 8980-9260 cal BP age range obtained for the 67 cm sample suggests that materials deposited deeper in the core would be older than 8980 cal BP. On this basis, it is likely that the actual basal age of the peat accumulation is close to the older endpoint of the 8540-8980 cal BP age range obtained from the basal samples; i.e., is likely to be close to 8980 BP. Likewise, the 8170-8390 cal BP age range obtained for the 57 cm sample suggests that materials deposited higher in the core would be younger than 8390 cal BP. Thus, the actual age of the peat at 30 cm depth is closer to the younger endpoint of the 8190-8660 cal BP age range obtained from the basal samples; i.e., is likely to be close to 8190 BP. Based on these arguments, the 2σ cal BP age ranges that are most appropriate to use in subsequent calculations of accumulation rates and in designating the ages of the boundaries of the sediments between 30-100 cm are 8980 cal BP and 8190 cal BP, respectively.

We were unable to obtain viable material between 30 cm and 15 cm for dating; only materials which were clearly not from the soil matrix were available (e.g., wood fragments) and the dates from these materials were rejected. The large span of time represented by the 30-15 cm section (over 4000 years), coupled with our sedimentological analyses of the sediments, leads us to conclude that this period contains an unconformity; i.e., that sediment removal took place some time between 8190 and 3890-4540 cal BP (midpoint 4240 cal BP).

The age ranges of the uppermost section of the core, from 15 cm to the surface, span from 3890-4540 cal BP (midpoint 4240 cal BP) to 0-300 cal BP (midpoint 150 cal BP).

5.4. Pollen and microfossil diagrams

Vegetation and landscape reconstructions are based on pollen and spore assemblages, as well as the environmental significance of fungi, algae, and zoological remains. The pollen percentage diagram (Figure 5) for the core profile is based on the sum of all terrestrial pollen types. The pollen was generally well preserved and reasonably abundant. A local zonation, based on pollen percentages, was constructed using stratigraphically constrained cluster analysis (Grimm, 1987). A pollen concentration diagram for selected taxa is also presented (Figure 6), but a pollen accumulation rate diagram was not produced owing to the limitation imposed by the overlapping calendar year ranges.

Utility of microfossils in this study

A number of fungi, algae, and other microfossil remains were identified in conjunction with the palynological analysis of the PER 02-5 core. Those identified microfossils used in our interpretations are listed in Table 3. One freshwater microfossil was particularly useful in environmental reconstruction, and should be noted. Large numbers of oocytes (eggs) of a small freshwater flatworm (order Neorhabdocoela, class Turbellaria) were identified in the pollen preparation and have been counted as percentages of the total pollen (Hass, 1996). The two species represented in the PER sediments are *Micodalyellia armigera* and *Gyratrix hermaphroditus*. Both species inhabit aquatic environments, prefer low lake levels, and respond favorably to eutrophication.

Zone PR-1 (100–30 cm), 8980 to 8190 cal BP.

Zone PR-1 shows relatively consistent pollen percentages. It is characterized by high percentages of Cyperaceae (44-65%), Ericaceae (heaths; 2-28%), and Gramineae (grasses; 6-21%). *Betula* (birch; 6-22%), and *Salix* (willow; 2-6%) are periodically present as are herbs such as *Polygonum bistorta*, *Artemisia*, and *Rubus chamaemorus* (cloudberry). The brief appearance of *Populus* (poplar; 5.7%) at 67 cm is notable. *Alnus* (alder; 2-9%) and *Picea* (spruce; 2%) pollen are both present at levels generally under 5% and are the result of long-distance transport.

Non-vascular plants are represented by *Lycopodium* (club mosses), *Sphagnum*, and *Equisetum* (horsetail; 2-8%). The CAR for the entire Zone PR-1 is 205 g m^{-2} y^{-1}(Figure 4c).

We split Zone PR-1 into subzones 1a and 1b based on the divergence of several indicators: pollen concentration, microfossils, and TOC. Zone PR-1a is characterized by Type 28 (spermatophore of Copepoda), the testate amoebae *Centropyxis*, and Type 353B (Turbellaria), which are present at very high levels (6-38%). Type 315 (*Zygnema* type; algae) is also present. The TOC throughout this subzone was very high (Figure 4a), ranging from a low of 11% at peat inception to 43% near the Zone PR-1a/PR-1b boundary. For Zone PR-1a, carbon storage is 127 kg m^{-2} (Figure 4b).

There is a increase in pollen concentration, from less than 5,000 in Zone PR-1a to over 40,000 grains cm^{-3} in Zone PR-1b, which also exhibits very high concentrations of *Betula* and Ericaceae pollen. The zoological remains for Types 28, *Centropyxis*, and 353B are present in Zone PR-1b, but in much lower amounts. Some algal spores are notable in this zone: Type 215 (*Debarya* sp.), and Type 229 (a probable algal spore). At the transition from Zone PR-1a to Zone PR-1b, Type 55A (fungal ascospore *Sordaria*) makes an appearance. While TOC levels are generally increasing during Zone PR-1a, they are decreasing throughout Zone PR-1b (Figure 4a). Carbon storage in Zone PR-1b is 35 kg C m^{-2} (Figure 4b).

Zone PR-2 (30-15 cm), 8190 cal BP to 4240 cal BP

Results from our sediment description (appearance of pebbles), low carbon storage (17 kg C m^{-2}; Figure 4b) and relatively low CAR (4.3 g m^{-2} y^{-1}; Figure 4c), and the palynological analyses all indicate this zone contains a disruption. It appears that sediments were eroded from the PER surface at some time during this period, presumably by wind action, to create an unconformity. This event took place between 8190 and 4240 cal BP, after which peat accumulation resumed.

The pollen and microfossil percentage diagram (Figure 5) shows a steep drop in *Betula* to 1.6%, its lowest level in the entire core. All of the major shrub percentages drop in favor of Cyperaceae and Gramineae within this zone; however, an inspection of the concentration diagram (Figure 6) shows that Cyperaceae also experienced a net decline in pollen. Grasses and herbs (*Artemisia, P. bistorta*, several species of Rosaceae, and Saxifragaceae undiff.) display increases in percentages and concentration of pollen grains.

Zone PR-3: (15-0 cm), 4240 cal BP to 150 cal BP

This zone, encompassing the Late Holocene, displays a very different assemblage of pollen and microfossil from the earlier zones. Cyperaceae dominates but the percentage range increases (53-65%). Gramineae percentages (6-16%) also increase. *Betula* is low (6-12%) and Ericaceae percentages are very low (3-7%). *Alnus* pollen attains its highest percentage level (7-9%). Most microfossils disappear or are present only in very low amounts. Exceptions are Type 229 which is present in high percentages, and small increases in the uppermost level of Type Wh2-8 and Type 353B. Little change from Zone PR-2 is evident in carbon storage (18 kg C m^{-2}) or CAR (4.4 g m^{-2} y^{-1}).

6. Discussion

Organic deposits from the pre-Holocene were not found at the PER. The basal date, 8980 cal BP, extends back to the early Holocene, and is significantly older than dates collected from thaw-lake basins on the Barrow Peninsula (Hinkel et al., 2003). Since the organic layer is underlain by mineral soil, it seems likely that coring penetrated to the base of the unit. However, it is possible that buried soils exist at greater depth, and earlier sediments may have been removed by erosion.

The Holocene vegetation and landscape changes at local and regional scales are summarized in Table 4. The mineral sediments at the base of the PER cores is a sandy loam probably of aeolian origin. These sediments could have been deposited at the end of the late Wisconsinan, although dry conditions conducive to landform aggredation may have continued into the early Holocene. During the period from 8980 to 8190 cal BP, the development of a sedge-grass heath tundra helped to stabilize the exposed mineral soils. This was a landscape diverse enough to sustain mesic vegetation as well as *Equisetum,* which is a typical component of mire communities and is often found growing in shallow water (~50 cm depth) of ponds and lake margins (Bliss and Matveyeva, 1992). Further evidence of persistent wet conditions are the high levels of zoological remains, all of which are typical of aquatic or semi-aquatic environments and, in the case of *Turbellaria*, also of high nutrient conditions.

Populus pollen appears briefly as a sharp spike in Zone PR-1a. Given the short duration of this event and the inadequate chronological control within this zone, it is impossible to fix a precise date for the *Populus* peak at the PER site or to assign climatological significance. Poplar macrofossils collected from peat deposits overlooking the Ikpikpuk River (150 km south of the study site) are dated to about 10,800 cal BP (Hopkins et al., 1981; Nelson and Carter, 1987). Poplar may have been present on the Arctic Coastal Plain during the HTM, but its distribution was likely restricted to river and stream valleys (Bockheim et al., 2003). The pollen percentages found in this zone (>5%) could be considered borderline evidence of a local poplar stand (Edwards and Dunwiddie, 1985).

Microfossil analysis indicates that, in the early Holocene, the local landscape was wet and conditions were eutrophic. Permanent standing water may result from higher precipitation or could be caused by soil water saturation due to poor drainage above permafrost. These conditions may well have encouraged the high TOC, C storage and CAR of 205 g m^{-2} y^{-1}. The mean long-term TOC value of 15.0 g m^{-2} y^{-1} is comparable to those reported for peat sequences from drained thaw-lake basins (Bockheim et al., 2004) as well as the European and Russian Arctic (Botch et al., 1995; Oksanen et al., 2001). Likewise, higher-than-modern peat accumulation and CARs is characteristic of Arctic peatland studies for the early to middle Holocene (Zoltai and Tarnocai 1975; Ovenden 1990; Sher 1992; Vardy et al. 1997; Lozhkin et al. 2001).

Zone PR-1b is notable for exhibiting the highest pollen concentrations of *Betula* and Ericaceae, which may represent the presence of shrub tundra. An important question is whether *Betula nana* (dwarf birch) was part of the local vegetation. *Betula nana* is absent today at the PER site, although its modern range does include the Atqasuk region, 80 km south of the study site (Komarkova and Webber, 1980). The pollen rain of the modern birch-free tundra vegetation, sampled from the mud-water interface of lake sediments, registers <15% *Betula* pollen on the Barrow Peninsula. Where dwarf birch is present, pollen increases to between 15 and 25% (Oswald et al., 2003). Thus, pollen percentages of over 20%, as well as the high concentrations of birch pollen, indicate that dwarf birch was likely present on the landscape between 8980 to 8190 cal BP. We hypothesize that this time span saw a more northerly expansion of birch on the Arctic Coastal Plain than previously reported.

The fungal ascospore, *Sordaria*, which is mainly limited to the earlier half of Zone PR-1b, has been found in Holocene sediments from the Netherlands. There, it is consistently found in association with not only heath and dwarf birch, but also animal dung (van Geel et al., 1981). Walker et al. (2001) identified caribou's primary foraging plants in moist acidic tundra as *Betula nana, Salix pulchra, Carex bigelowii,* and *Cassiope tetragona,* all of which are likely sources of the pollen identified in Zone PR-1b. This evidence, coincident with the rise in herbs and disappearance of open water indicators, points to lower soil moisture toward the end of Zone PR-1b. With increased drainage and the formation of ice-rich sediments, nutrient availability became limited.

Sometime between 8190 and 4240 cal BP (Zone PR-2), the vegetation of the PER changed from shrub tundra to herb tundra. Plants with a high tolerance of either dry or disturbed conditions become more prevalent, and the disappearance of birch during this period is significant. Productivity, as represented by TOC and CAR, also drop precipitously. The appearance of pebbles also indicate a severely disturbed surface; it is likely they were left as a lag deposit after finer sediments were removed by wind action. Because it is uncertain when the sediments were lost, it is impossible to say whether the dry or disturbed herb tundra represented by this pollen assemblage was established before or after the erosion event. We can imagine the conditions required to cause sediment loss would be a reduction in soil moisture caused by regional climate aridity around 4500 cal BP (Carter, 1993; Galloway and Carter, 1993). This condition may have been enhanced by local landscape processes such as aggradation of the PER surface by frost heave and attendant slope steepening.

The final stage of vegetation development on the PER, Zone PR-3, is a brief record of the development of late Holocene vegetation and landscape. Vegetation, dominated by grasses, sedges, and herbs, and slow accumulation of organic matter resumed by 4240 cal BP. This may represent a community increasingly under stress from declining nutrient levels and colder climatic conditions. The re-introduction of wetland indicators may indicate that the coring site was becoming influenced by ice-wedge growth and the formation of small high-center polygons, which dot the present surface of the PER.

We propose the following working hypothesis for vegetation development and carbon sequestration during the Holocene for the Alaskan Arctic Coastal Plain. During the early Holocene (9000 to 8200 cal BP) the active loess and sand dunes were largely stabilized through climatic amelioration. In upland and sheltered regions, carbon accumulation was rapid due to high net primary productivity in a relatively nutrient-rich environment. The middle part of the Holocene, represented as an unconformity in the PER record, shows that at some point after 8200 cal BP but before 4200 cal BP, profound landscape changes occurred. Studies done in the Coastal Plain (Carter, 1993; Eisner and Peterson, 1998b) provide strong evidence that sediment erosion occurred only slightly prior to 4500 cal BP. Galloway and Carter (1993) dated a series of buried soils showing that aeolian activity decreased in the western Arctic Coastal Plain between 4300 and 3900 cal BP, reactivating during several subsequent intervals until the present.

During the late Holocene (4200 cal BP to 150 cal BP), carbon accumulation resumed but was reduced in response to the combined effects of declining productivity associated with climatic cooling, and increased nutrient stress as paludification and permafrost aggradation sequestered mineral nutrients. Further corroborative evidence from northern Alaska and northwestern Canada indicate increased ice-wedge development, permafrost aggradation, and thaw lake drainage, presumably occurring preferentially as a response to cooling conditions (Ritchie, 1984; Mackay, 1992; Eisner et al. 2003).

It is integral to our analysis of the PER sediments and landform that we understand the indigenous processes effecting the vegetation and soils before we invoke climate change as the primary agent for change. The aggradation of the local surface may be partially due to frost heave. Drainage may have been enhanced as thaw lakes developed on either side of the PER, eroding the "plateau" laterally and steepening the sides. Aeolian accumulation continued intermittently in the Holocene, and could also have contributed to the increasing relief of the PER above the surrounding landscape. As the PER became recognizably elevated above its surroundings, it may have become a grazing and congregation area for caribou, resulting in the increase in *Sordaria*.

7. Conclusions

The PER record provides evidence for warmer conditions between ca. 9000 and 8200 cal BP which we interpret as representing the Holocene thermal maximum on the Arctic Coastal Plain. The PER sediments lack the earliest part of the HTM, but nonetheless several lines of evidence correlate well with previous studies in other parts of northern Alaska. These include the development of a highly productive (as indicated by high carbon storage) mesic shrub tundra, the probable presence of dwarf birch beyond its modern limit, and high carbon storage and accumulation rates throughout the period.

Microfossils and sedimentological analyses have also enabled us to reconstruct the Holocene vegetation history and geomorphic development of this unique feature of the Arctic Coastal Plain. The period of sediment erosion between ca. 4200 and 8200 years ago is striking in terms of its possible regional significance but also because it allows us some insight into how these erosional remnants respond to changes in local and regional hydrological and climatic changes. These findings reflect local changes from a single site, but correlation of the regional pollen signal with other proxy data from the Arctic Coastal Plain including dated loess deposits, macrofossils, beetle analyses (Hopkins et al., 1981; Nelson and Carter, 1987; Carter, 1993), and the palynological study at the Meade River, allows us to reconstruct a tentative regional history of the region.

Our findings indicate that carbon accumulation rates in the arctic tundra may be closely linked to wet and eutrophic conditions as well as temperature. We recommend that as research into the response of terrestrial soil carbon stocks under changing climatic conditions continues, investigations should focus on changes in soil moisture as well as climatic warming.

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TABLES

Table 1: Property of cores collected from PER, 2001-2003

Profile Number	Soil thickness (cm)	Depth of massive ice (cm)	Thickness of massive ice (cm)	Comments
PER 01-1	38	38 to >71	>33	short core barrel
PER 01-2	50	50 to >67	>17	short core barrel
PER 02-1	124	124 to >134	>10	basal peat dated (119 cm)
PER 02-2	41	41 to >160	>119	mostly massive ice
PER 02-3	37	37 to >120	>83	mostly massive ice
PER 02-4	150	--	0	cryoturbated, textural analysis of Cg
PER 02-5	64	64 to 98 (34) 118-143 (25)	59	soil resumes 98-118 cm & 143-170 cm; TOC, pollen, dating
PER 03-1	62	62 to >74	>12	soil resumes 74-101 cm
PER 03-2	52	52 to >90	>38	
PER 03-3	58	58 to >105	>47	

Table 2. Original sampling depths, deflated depths, radiocarbon dates, and calibrated dates[a] from PER 02-5 and from PER 02-1

CAMS[b] lab no.	Depth (cm)	Deflated depth (cm)	^{14}C yr B.P.	±	cal BP 2_ midpoint	cal BP 2_ range	Material Sampled
90992	3	3	184	40	150	0-300	sedge
86718	6	6	2845	45	2965	2850-3080	sedge
92174	15	15	3860	130	4240	3890-4540	sedge
92175	30	30	7670	110	8420	8190-8660	sedge
92176	44	44	7620	110	8395	8180-8610	sphagnum
87123	57	57	7470	60	8280	8170-8390	sphagnum
86719	101	67	8100	50	9120	8980-9260	sedge
92177	109	75	8480	410	9425	8450-10400	sphagnum
97577	158	100	7870	70	8765	8540-8990	sphagnum
97578	158	100	7910	140	8720	8410-9030	sedge
92179	159	100	7910	210	8820	8340-9300	sedge
90998	PER 02-1	119	8070	70	8920	8710-9130	sphagnum
90994 [c]	15		9180	80			wood
90995 [c]	30		8820	70			seed
90996 [c]	44		8140	70			wood
90997 [c]	109		8390	70			wood
86720 [c]	159		9935	40			wood

[a]CALIB rev. 4.4 (Stuiver & Reimer, 1993)

[b]CAMS, Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory

[c]Denotes radiocarbon dates not incorporated into paleoecological reconstruction because sample composition has indicated the material is not part of the sample matrix (see text).

Table 3. Selected diagnostic microfossils from the PER cores

Type	Name	Environmental Significance/Association
28, spermatophore	Copepoda (Crustacea)	(temporary) open water
55A, fungal ascospore	*cf. Sordaria*	eu- to mesotrophic peat, animal dung
215, algal spore	*Meesia triquetra*	low nutrients
229, algal spore	Species undetermined	sandy, shallow pools
315, algal spore	*Zygnema* type	mesotrophic, shallow pools
353B, flatworm oocytes	Turbellaria (Neorhabdocoela)	shallow open water, eutrophication
testate amoeba	*Centropyxis*	aquatic or very wet conditions

Table 4: PER zonation and landscape/climate reconstruction

Zone	**Depth (cm)**	**date (midpoint cal BP)**	**C storage (kg C m^{-2})**	**Mean C accum. rate (g m^{-2} y^{-1})**	**Vegetation & local landscape reconstruction**	**Regional landscape reconstruction**
PR-3	0-15	150-4240	18	4.4	Grass-sedge-herb tundra; upland with high-center polygons	cooling from previous periods; permafrost aggradation
PR-2	15-30	4240-8190	17	4.3	**Unconformity:** grass-herb tundra; upland w. eroded surface	climate drier; regional eolian activity
PR-1b	30-57	8190-8980	35	205	Birch-heath shrub tundra; incipient upland, decreased soil moisture	Warmer than present, thaw lake formation; HTM
PR-1a	57-100		127		Sedge-grass tundra (heath, birch, willow); low-lying wetland	

FIGURE CAPTIONS

Figure 1. Landsat 7+ image from 30 August 2000 of Barrow Peninsula

Figure 2: Shaded DEM of Barrow Peninsula showing primary geomorphic features. Manley et al., 2004.

Figure 3. Textural analysis of mineral soil from Cg horizon of PER 02-4.

Figure 4: (a) Total organic carbon (TOC, %); (b) Carbon storage (kg m^{-2}); and (c) age of sediments (cal BP) for deflated Core PER 02-5. Numbers in boxes in 4c represent interval-specific carbon accumulation rate (g m^{-2} y^{-1}).

Figure 5: Pollen and microfossil percentage diagram for Core PER 02-5. Depth is based on deflated measurements after ice thickness is subtracted (see text). Percents are calculated based on all identified and unidentified pollen, with spores, fungi, algae, and rhizopods expressed as a percentage of the sum. Shaded pattern indicates 7X exaggeration. Arrow denotes probable unconformity gap..

Figure 6: Summary pollen concentration diagram of selected pollen spectra for deflated Core PER 02-5. Scale is expressed as grains per cm X 100.

Figure 1.

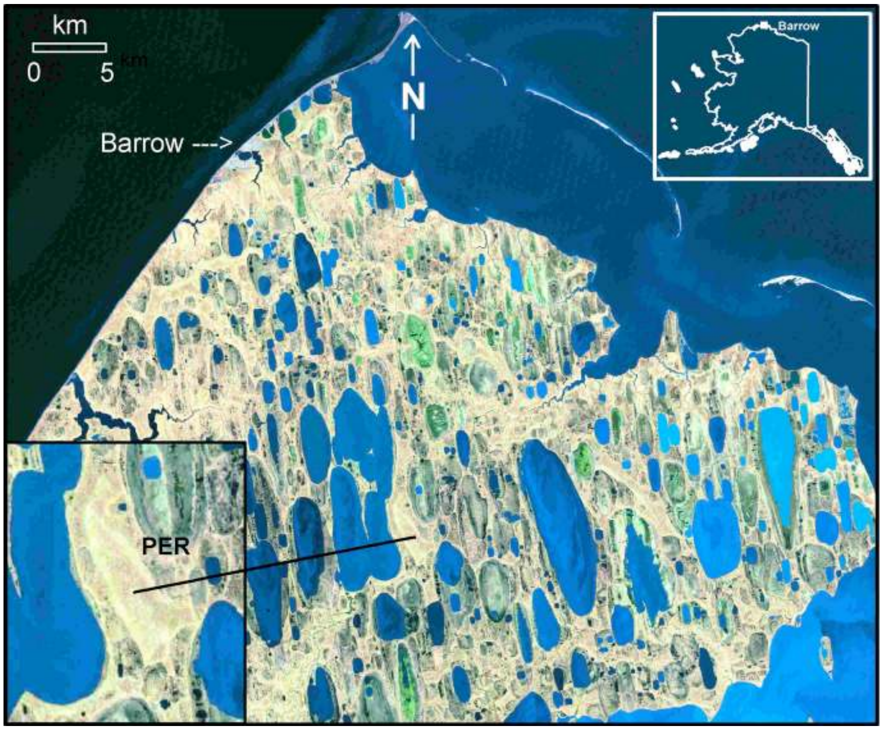


Figure 2.

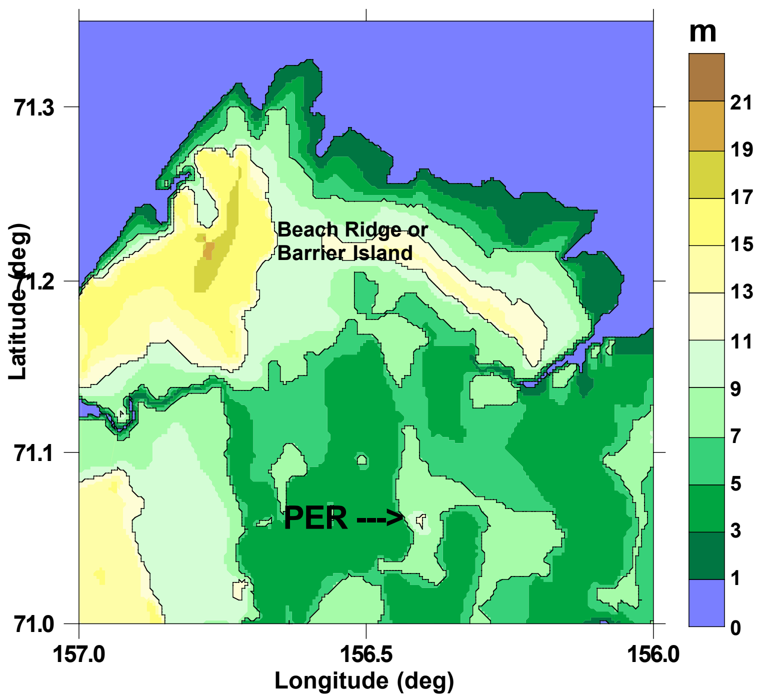

m
21
19
17
15
13
11
9
7
5
3
1
0
71.3
71.2
71.1
71.0
Latitude (deg)
157.0
156.5
156.0
Longitude (deg)
Beach Ridge or
Barrier Island
PER --->

Figure 3.

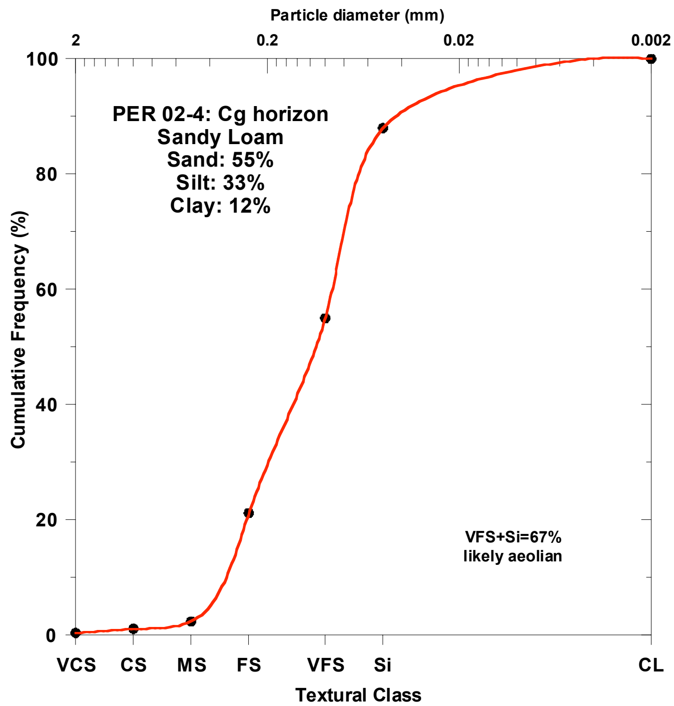
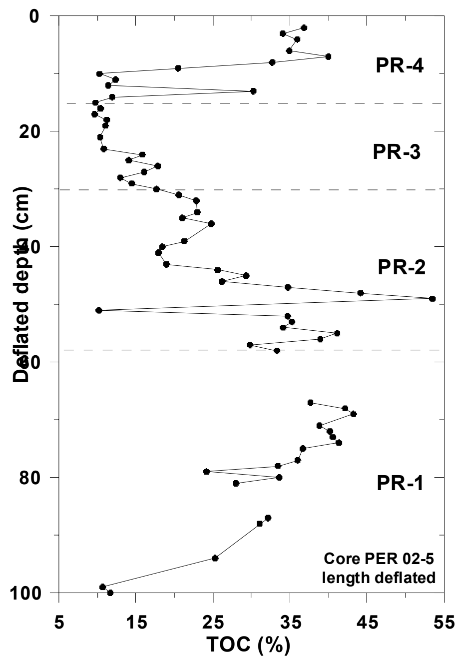
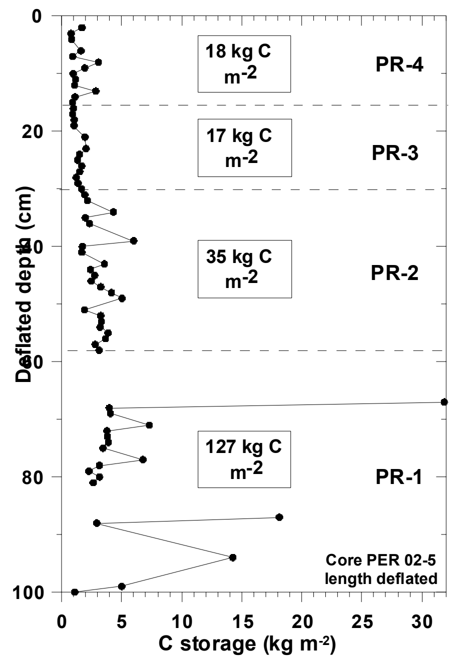

Particle diameter (mm)
2
0.2
0.02
0.002
100
80
60
40
20
0
PER 02-4: Cg horizon
Sandy Loam
Sand: 55%
Silt: 33%
Clay: 12%
Cumulative Frequency (%)
VFS+Si=67%
likely aeolian
VCS
CS
MS
FS
VFS
Si
CL
Textural Class

Figure 4a.



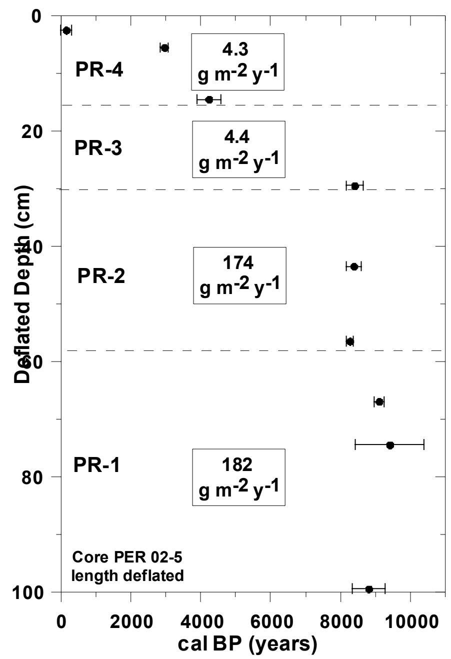
PR-4
PR-3
PR-2
PR-1
Core PER 02-5
length deflated
Deflated depth (cm)
TOC (%)
0
20
40
60
80
100
5
15
25
35
45
55

Figure 4b.



18 kg C m-2
PR-4
17 kg C m-2
PR-3
35 kg C m-2
PR-2
127 kg C m-2
PR-1
Core PER 02-5
length deflated
Deflated depth (cm)
C storage (kg m-2)
0
20
40
60
80
100
0
5
10
15
20
25
30

Figure 4c.



PR-4
4.3
g m-2 y-1
PR-3
4.4
g m-2 y-1
PR-2
174
g m-2 y-1
PR-1
182
g m-2 y-1
Core PER 02-5
length deflated
Deflated Depth (cm)
0
20
40
60
80
100
0
2000
4000
6000
8000
10000
cal BP (years)

Figure 5.

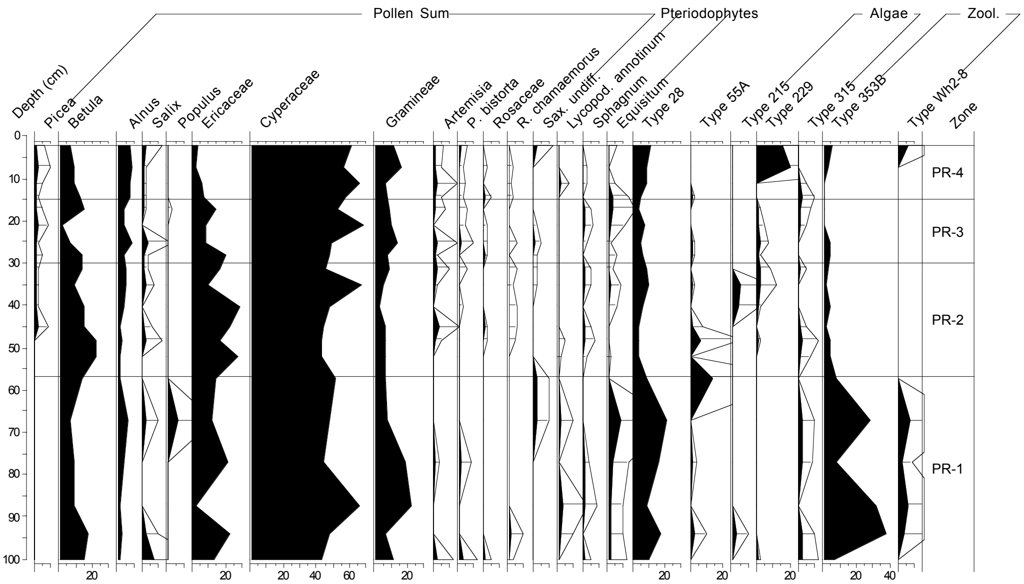
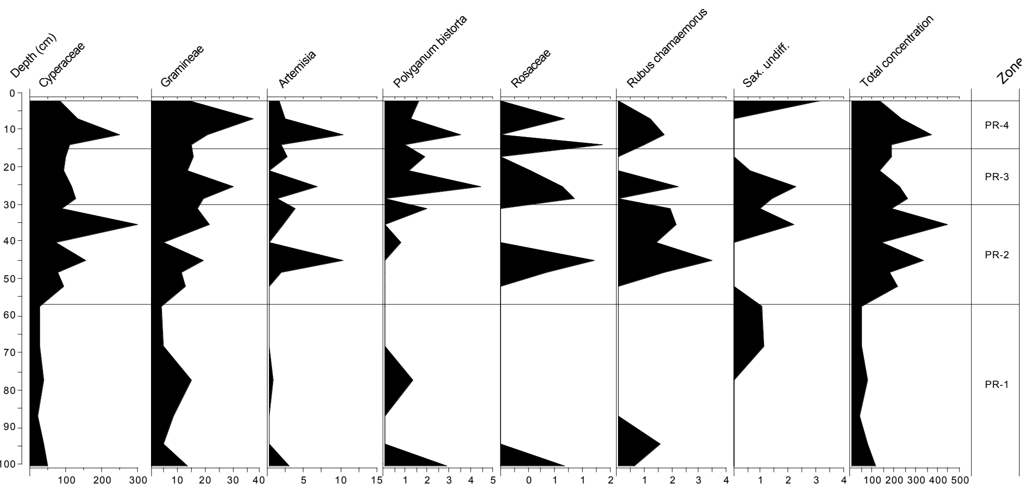

Pollen Sum
Pteriodophytes
Algae
Zool.
Depth (cm)
Picea
Betula
Alnus
Salix
Populus
Ericaceae
Cyperaceae
Gramineae
Artemisia
P. bistorta
Rosaceae
R. chamaemorus
Sax. undiff.
Lycopod. annotinum
Sphagnum
Equisitum
Type 28
Type 55A
Type 215
Type 229
Type 315
Type 353B
Type Wh2-8
Zone
PR-4
PR-3
PR-2
PR-1
0
10
20
30
40
50
60
70
80
90
100
20
40
60

Figure 6.


Depth (cm)
Cyperaceae
Gramineae
Artemisia
Polyganum bistorta
Rosaceae
Rubus chamaemorus
Sax. undiff.
Total concentration
Zone
PR-4
PR-3
PR-2
PR-1