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Climate and vegetation history from a 14,000-year peatland record, Kenai Peninsula, Alaska

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ABSTRACT

Analysis of pollen, spores, macrofossils, and lithology of an AMS ^{14}C-dated core from a subarctic fen on the Kenai Peninsula, Alaska reveals changes in vegetation and climate beginning 14,200 cal yr BP. *Betula* expansion and contraction of herb tundra vegetation characterize the Younger Dryas on the Kenai, suggesting increased winter snowfall concurrent with cool, sunny summers. Remarkable Polypodiaceae (fern) abundance between 11,500 and 8500 cal yr BP implies a significant change in climate. Enhanced peat preservation and the occurrence of wet meadow species suggest high moisture from 11,500 to 10,700 cal yr BP, in contrast to drier conditions in southeastern Alaska; this pattern may indicate an intensification and repositioning of the Aleutian Low (AL). Drier conditions on the Kenai Peninsula from 10,700 to 8500 cal yr BP may signify a weaker AL, but elevated fern abundance may have been sustained by high seasonality with substantial snowfall and enhanced glacial melt. Decreased insolation-induced seasonality resulted in climatic cooling after 8500 cal yr BP, with increased humidity from 8000 to 5000 cal yr BP. A dry interval punctuated by volcanic activity occurred between 5000 and 3500 cal yr BP, followed by cool, moist climate, coincident with Neoglaciation. *Tsuga mertensiana* expanded after ~1500 cal yr BP in response to the shift to cooler conditions.

Introduction

High latitudes are particularly sensitive to climate change (Moritz et al., 2002). In recent decades climatic change has caused alterations to the hydrological and biogeochemical cycles in the Arctic and subarctic, which have resulted in vegetational changes (Chapin et al., 1995; Klein et al., 2005, Sturm et al., 2005). For example, recent widespread lowering of the water table in south-central Alaska has resulted in increased shrub invasion on wetland sites on the Kenai Peninsula (Klein et al., 2005). These changes are not unique to modern climate, and paleoenvironmental studies aid in investigating ecosystem responses to shifts in temperature and moisture on longer time scales.

Vegetation records from Alaska and western Canada exhibit heterogeneous climatic shifts throughout the late glacial and early Holocene, and spatial variation in synoptic climatic controls that govern temperature and precipitation must be considered when interpreting the paleoclimatic record (Mock et al., 1998). Several sites in Alaska document a cold, dry Younger Dryas signal (Engstrom et al., 1990; Peteet and Mann, 1994; Hu et al., 2002), but this signal is not well defined in other Alaskan records, such as previous Kenai Peninsula records (Ager 2000; 2001). Early Holocene Milankovitch insolation changes poleward of 60° that increased summer and decreased winter insolation by 8% over today's values (Berger, 1978) resulted in widespread warming, which caused range extensions of plant and animal taxa (e.g., Kaufman et al., 2004). While most records are interpreted as reflecting warm and dry conditions, several sites to the west of Cook Inlet, primarily sites surrounding Bristol Bay, are interpreted as moist, with abundant Apiaceae and Polypodiaceae (Hu et al., 1995, 1996; Brubaker et al., 2001).

The Kenai Peninsula in south-central Alaska lies between sites that experienced a warm, dry early Holocene climate [Prince William Sound and southeastern Alaska (Heusser et al., 1985), interior Alaska (Bigelow and Powers, 2001), and Canada (Ritchie et al., 1983)] and those that experienced a moist climate in the early Holocene [southwestern Alaska (Hu et al., 1995; Hu et al., 1996; Brubaker et al., 2001; Axford and Kaufman, 2004)]. On the Kenai Peninsula, Kachemak Bay sites (Ager, 2000) are interpreted as warm and moist in the early Holocene based on pollen assemblages composed of Apiaceae and Polypodiaceae, while the climate at a single Kenai lowland site (Anderson et al., 2006) was recently regarded as warm and dry based on increased charcoal.

Aspects of the early Holocene vegetation remain poorly understood because modern analogs for the pollen and spore record are lacking, particularly the high fern spore interval observed in several early Holocene records across southwestern and south-central Alaska. The Kenai Peninsula lies in the central southern coastal region and is

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thus positioned to provide information about spatial temperature and precipitation patterns, and about whether these configurations remain constant over time. A combination of pollen, spores, and plant macrofossils in conjunction with loss-on-ignition (LOI), AMS ^{14}C dates, and lithology are used to address late-glacial and Holocene hydrologic and temperature shifts from Swanson Fen, a Kenai Peninsula lowland site.

The following questions are addressed in this study: (1) What does this multi-proxy record add to pre-existing vegetational history on the Kenai Peninsula? (2) How can the lithologic and macrofossil shifts aid our interpretation of hydrology and ecosystem shifts? (3) What are the implications for understanding regional atmospheric circulation patterns throughout the late glacial and Holocene, including changes to the strength and position of the Aleutian Low?

Setting

Swanson Fen is a small (300 m^2), nutrient-poor fen situated on the northern Kenai Peninsula lowlands in south-central Alaska (60°47.35′ N, 150°49.91′W) (Fig. 1). The Kenai Peninsula is connected to the mainland to the north by Portage Passage and bordered to the west by Cook Inlet. The Kenai Mountains, which reach 2015 m and are mantled by the Harding, Sargent, and Spencer-Blackstone icefields, extend across the eastern Kenai Peninsula.

The Kenai Mountains are composed of Mesozoic bedrock, while the lowlands to the west exhibit a flat to rolling surface formed by several Wisconsin-age glaciations, and comprised of organic-rich late-glacial and Holocene peat, interbedded with layers of volcanic ash (Reger et al., 2007). The Turnagain Arm lobe of the Naptowne

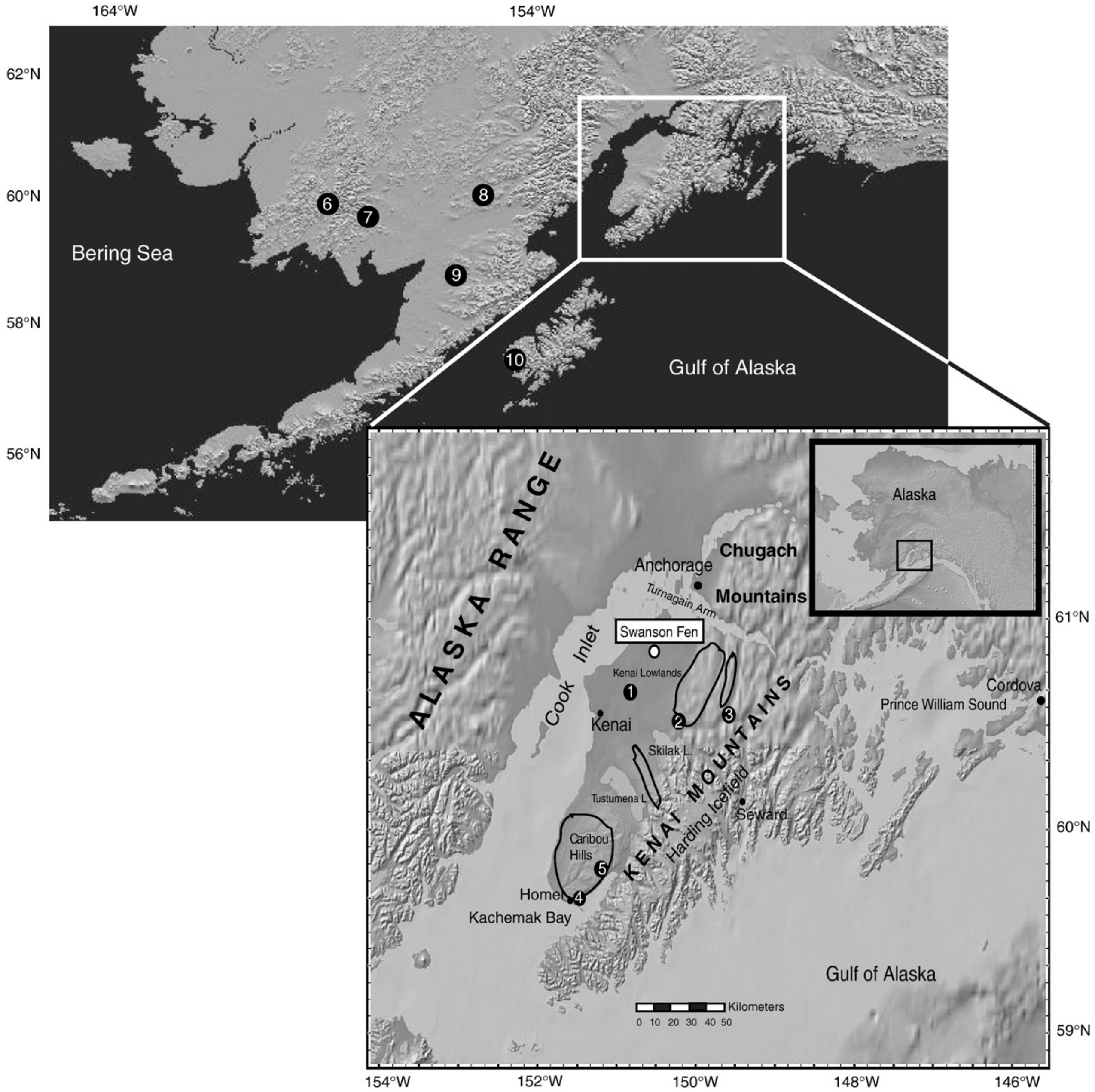

Figure 1. Kenai Peninsula map and core locations. Swanson Fen (this study) is marked with a white dot, and numbered circles indicate locations of other palynological studies on the Kenai Peninsula—1: Paradox Lake (R.S. Anderson et al., 2006); 2: Hidden Lake (Ager, 1983); 3: Tern Lake (Ager, 2001); 4: Homer Spit Peat Section (Ager, 2000); 5: Circle Lake (Ager, 2000). Regional map of selected studies on Holocene vegetation change—6: Little Swift Lake (Axford and Kaufman, 2004); 7: Grandfather Lake (Brubaker et al., 2001); 8: Snipe Lake (Brubaker et al., 2001); 9: Idavain Lake (Brubaker et al., 2001); 10: Phalarope Pond (Peteet and Mann, 1994). Areas outlined in black are approximate locations of glacial refugia on the Kenai Peninsula, modified from Reger et al. (2007).

Table 1
Weather station data from selected south-central Alaska locations.

Weather station location	M Jan T (°C)	M Jul T (°C)	MAT (°C)	MAP (cm)
Kenai, AK	− 11	12.5	1.9	48
Homer, AK	− 6	11	1.8	65
Seward, AK	− 3.2	13.6	4.6	182
Cordova, AK	− 4.9	12.5	3.5	228
Kodiak, AK	− 1	12.5	4.9	193

Mean annual temperature (MAT), mean July temperature, mean January temperature, and mean annual precipitation (MAP) for five southern coastal Alaskan sites: Kenai, Homer, Seward, Kodiak, and Cordova, AK. Measurements are based on a 30-year average.

glaciation (~32,000–16,000 cal yr BP) did not extend much farther west than the Kenai Mountains, as indicated by the absence of till or moraines of Naptowne age (Reger et al. 2007). Ice moved across Cook Inlet from the west to cover much of the northern Kenai lowlands during the Naptowne glaciation (Reger et al., 2007), and geologic evidence suggests small ice-free refugia existed in the northwestern Kenai Mountains, in the upland between Skilak and Tustumena Lakes, and in the Caribou Hills (Fig. 1, Reger et al., 2007).

Modern vegetation

The Kenai Peninsula is divided into three ecoregions (Ager, 2000; Viereck et al., 2007): *Picea sitchensis-Tsuga* forests along the high-precipitation southern and southeastern coasts; *Picea glauca* and *Picea mariana-Betula* forests west of the Kenai Mountains; and both shrub and alpine herb tundra communities in the Kenai Mountains. Lowland forests include *P. mariana*, bottomland *Picea X lutzii* (*P. glauca* x *P. sitchensis*), and *P. mariana-Populus* adjacent to larger river drainages. *P. glauca*, *Populus tremuloides*, *Betula neoalaskana* and *Betula kenaica* (Viereck et al., 2007) commonly occur on well-drained upland soils (Ager, 2001). Wetter areas and riparian zones contain *Alnus* and *Salix* thickets. Some wet tundra communities also exist along the coastline of Cook Inlet. Swanson Fen is surrounded by a *P. mariana* forest and an isolated *Tsuga mertensiana* stand, a species typical of deep winter snows in regions with cool, moist climates, such as the eastern Kenai Mountains.

Modern climate patterns

The Kenai Mountains create a rain shadow, with high precipitation on the eastern side of the mountains and much drier conditions to the west. The Kenai lowland experiences a semi-continental climate, with lower precipitation (48 cm), colder winter temperatures (− 15.6 to − 6.2°C mean January temperatures), and warmer or comparable summer temperatures (8.5–16.5°C mean July temperatures) than surrounding coastal locations, such as either Kodiak to the south or Cordova to the east (Table 1; http://www.Cdc.noaa.gov/USclimate).

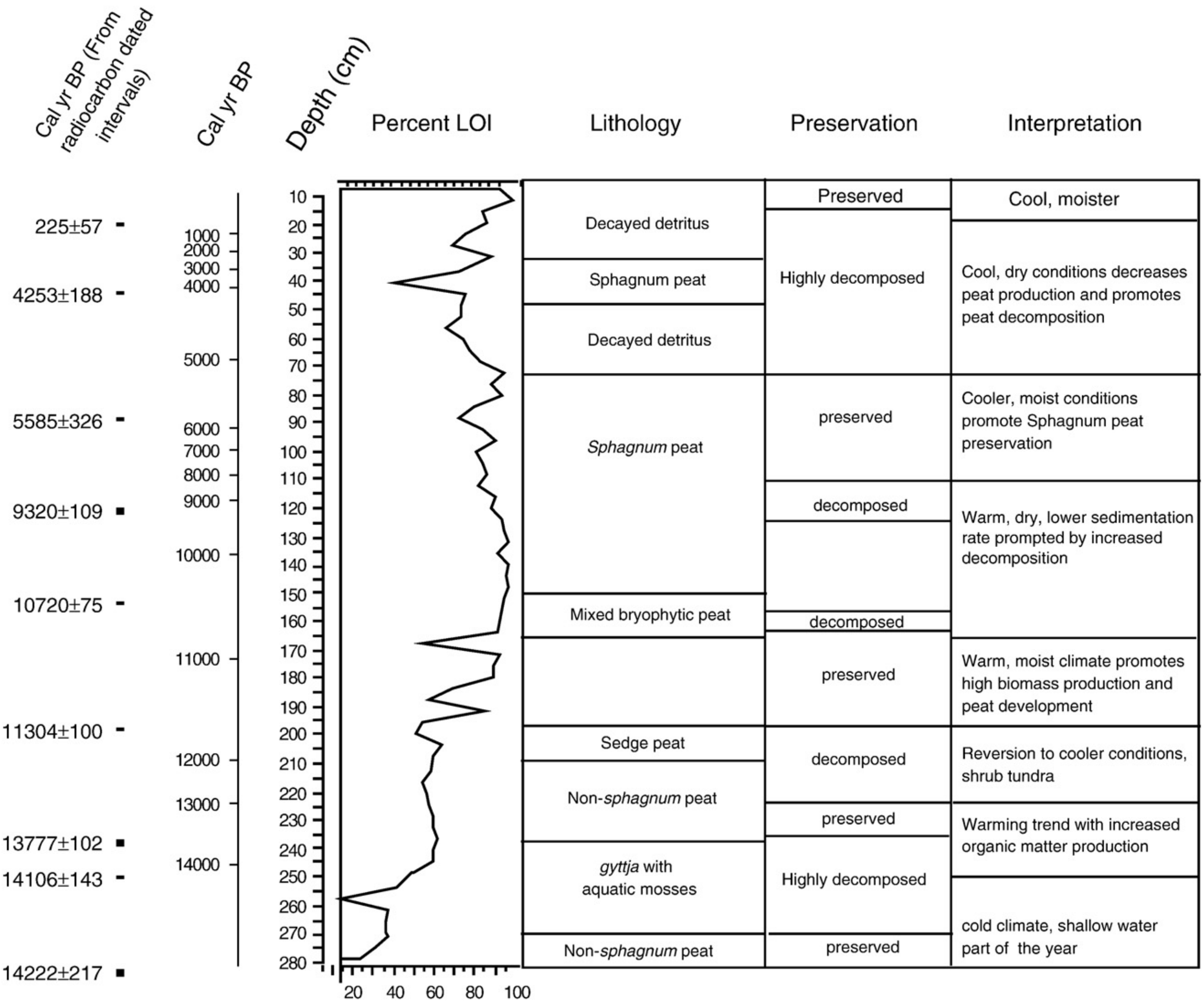

Figure 2. Lithostratigraphy of the Swanson Fen, Alaska core following the Troels-Smith (1955) classification. Lithology describes the composition of the core, and the preservation was determined based on degree of recognizable material.

Table 2
Radiocarbon and calendar age ranges, calculated using Calib 5.1 (Stuiver and Reimer, 1993).

Lab no.	Depth (cm)	$\delta^{13}C$ (estimated)	Material dated	Uncorrected ^{14}C year BP	Calibrated 2-sigma age range (cal yr BP)
CAMS-114297	20–22	−25	Trigonous *Carex* seed	140 ± 40	225 ± 57 (45%)
CAMS-131915	44–46	−25	*Betula* bark	3840 ± 80	4253 ± 188 (95%)
CAMS-123586	88–90	−25	*Vaccinium oxycoccus* seed, trigonous *Carex* seed	4810 ± 160	5585 ± 326 (94%)
CAMS-131916	108–110	−25	*Carex* seed, *Betula* seed, insect part	3310 ± 90	3545 ± 185 (97%)
CAMS-131917	120–122	−25	Wood	8305 ± 40	9320 ± 119 (93%)
CAMS-123587	152–154	−25	*Menyanthes* seeds	9480 ± 35	10720 ± 75 (78%)
CAMS-114299	196–198	−25	*Menyanthes*, *Hippuris* seeds	9890 ± 45	11304 ± 100 (100%)
CAMS-114300	236–238	−25	*Carex* seeds	11910 ± 40	13777 ± 102 (100%)
CAMS-114301	248–250	−25	*Hippuris* seeds	12245 ± 45	14106 ± 143 (100%)
CAMS-123588	281–283	−25	*Hippuris* seeds	12290 ± 40	14222 ± 217 (100%)

The Kenai Peninsula weather patterns are influenced by the Aleutian Low (AL), which is a semi-permanent low-pressure center situated over the North Pacific (Overland et al., 1999). The AL moves from its position in the Bering Sea eastward to the Gulf of Alaska in the winter and early spring months when it strengthens, and westward toward the Aleutian Islands in July when it weakens (Trenberth and Hurrell, 1994). The AL exhibits decadal variability in its strength and position (Overland et al., 1999), and a stronger AL is linked to warmer sea surface temperatures (SSTs) and greater precipitation on the Kenai Peninsula (Trenberth and Hurrell, 1994). Gulf of Alaska land temperatures and SSTs are strongly correlated to the Pacific Decadal Oscillation (PDO), an indicator of the mode of North Pacific SSTs, and to the North Pacific Index (NPI), an indicator of the intensity of the AL, which influence the decadal variability in temperature and precipitation in this region (Overland et al., 1999).

Methodology

Field methods

A 2.5-m sediment core from Swanson Fen was recovered in the summer of 2005 using a Hiller corer. Because the first core did not reach the basal clay, a second core 2.9 m long was taken directly adjacent to the first core in 2006. The sediment of each core was subsampled in 2-cm intervals.

Pollen, spore, and macrofossil analyses

Samples for pollen analysis were processed every 4 cm using standard procedures (Faegri and Iverson, 1989). A minimum of 300 terrestrial pollen grains was counted at 400× magnification. Higher counts to a 500-grain pollen sum on selected samples revealed pollen percentage shifts of <1.5% from a 300-grain count for the dominant pollen taxa. Pollen sums were calculated to include all terrestrial pollen types, and spore percentage calculations were based on the sum of pollen and spores. These percentages were plotted in Tilia Graph (Grimm, 1992).

Macrofossils were analyzed from 20 cm^3 of sediment according to the methods of Watts and Winter (1966) and Janssens (1990), identifying seeds, fruits, needles, stems, bryophytes, and charcoal. Macrofossils were picked and counted using a binocular microscope and were plotted based on influx or presence, in the case of low influx. For each sample, the matrix was analyzed and characterized using a modified Troels-Smith (1955) classification system. Primary divisions included peat moss, herbaceous peat, decayed detritus (<1 mm), and *gyttja* (lake material). These types were further subdivided based on the degree of humification and presence of other defining characteristics, such as the presence of pteridophytes, *Daphnia* ephippia (aquatic crustaceans), tephra, and fungal sclerotia (Figs. 2, 4). Selected macrofossils were dated using AMS radiocarbon dating techniques at Lawrence Livermore Labs, California (Table 2). Radiocarbon ages were calibrated to calendar ages using CALIB (version 5.0) (Stuiver and Reimer, 1993) and an age model was constructed in Tilia Graph (Grimm, 1992) based on linear interpolation between the midpoints of the 2-sigma calibrated age range.

Results

Chronology and lithology

Ten radiocarbon ages from plant macrofossils (Table 2) were obtained for Swanson Fen, excluding a date of 3310 ± 90 ^{14}C yr BP at 110–108 cm. The small size of this sample may have resulted in high background contamination, or the variety of the material dated may have influenced the age, both resulting in a younger age (Oswald et al., 2005). Emergent macrofossils (*Hippuris* and *Menyanthes trifoliata* seeds) were dated in four samples because of a lack of other suitable upland species. *Hippuris* does not generally use HCO_3^- and therefore is unlikely to contain "old" carbon, and while *Menyanthes* can use respired CO_2 from sediments, the consistent chronology suggests "old" carbon did not contaminate these samples (Birks, 2001). High sedimentation in the late glacial corresponds to a low LOI (Fig. 3), and no pollen or spores were found at the core base (290–280 cm).

The lithology (Fig. 2) changes from *gyttja* to peat at 240 cm core depth and remains bryophytic for the remainder of the core. The transition from Amblystegiaceae-dominated to *Sphagnum*-dominated bryophyte species begins at 150 cm and most minerotrophic brown moss species disappear by 70 cm (~5000 cal yr BP, Fig. 4), indicating a decline in pH (Janssens, 1990). The greatest peat preservation occurs from 148 to 72 cm, and it is most degraded in the upper 70 cm of the core.

Loss-on-ignition

LOI in Swanson Fen (Fig. 2) is low at the base (12%) and increases (>80%) in the early Holocene (11,500–9500 cal yr BP),

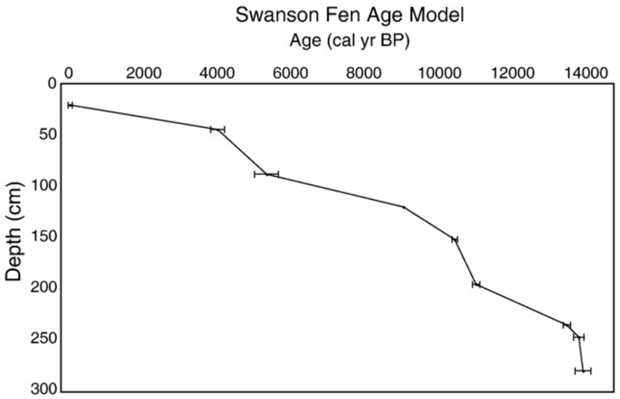

Figure 3. Swanson Fen age model. Sedimentation rate based on AMS radiocarbon-dated macrofossils.

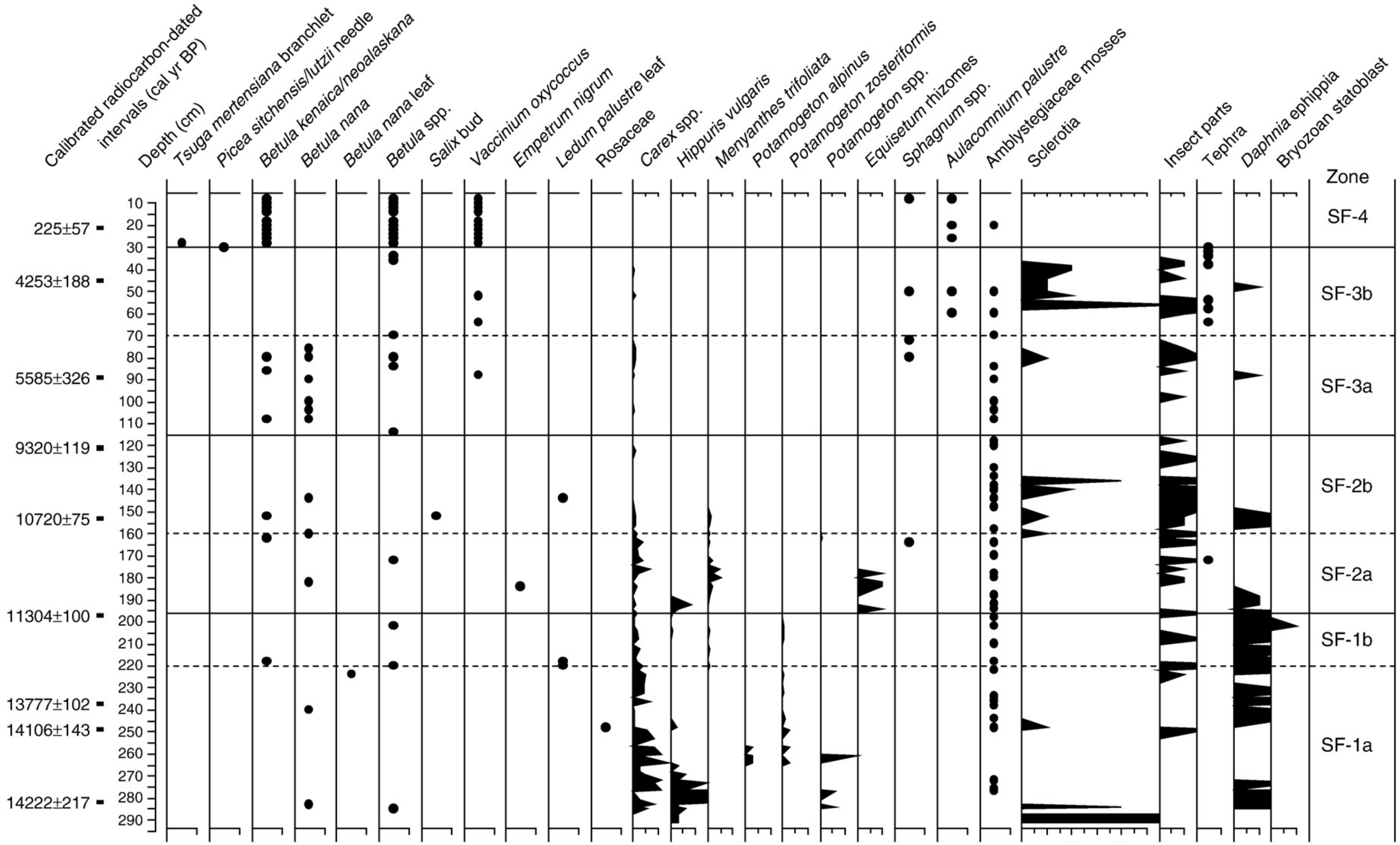

Figure 4. Macrofossil influx ($10^3\ \mathrm{cm}^{-2}\ \mathrm{yr}^{-1}$) from Swanson Fen, Alaska, including identifiable plant remains, algae and fungal remains, bryophytes, insect remains, and tephra. Macrofossils are seeds unless otherwise indicated. Taxa with low influx were plotted as presence (black dots).

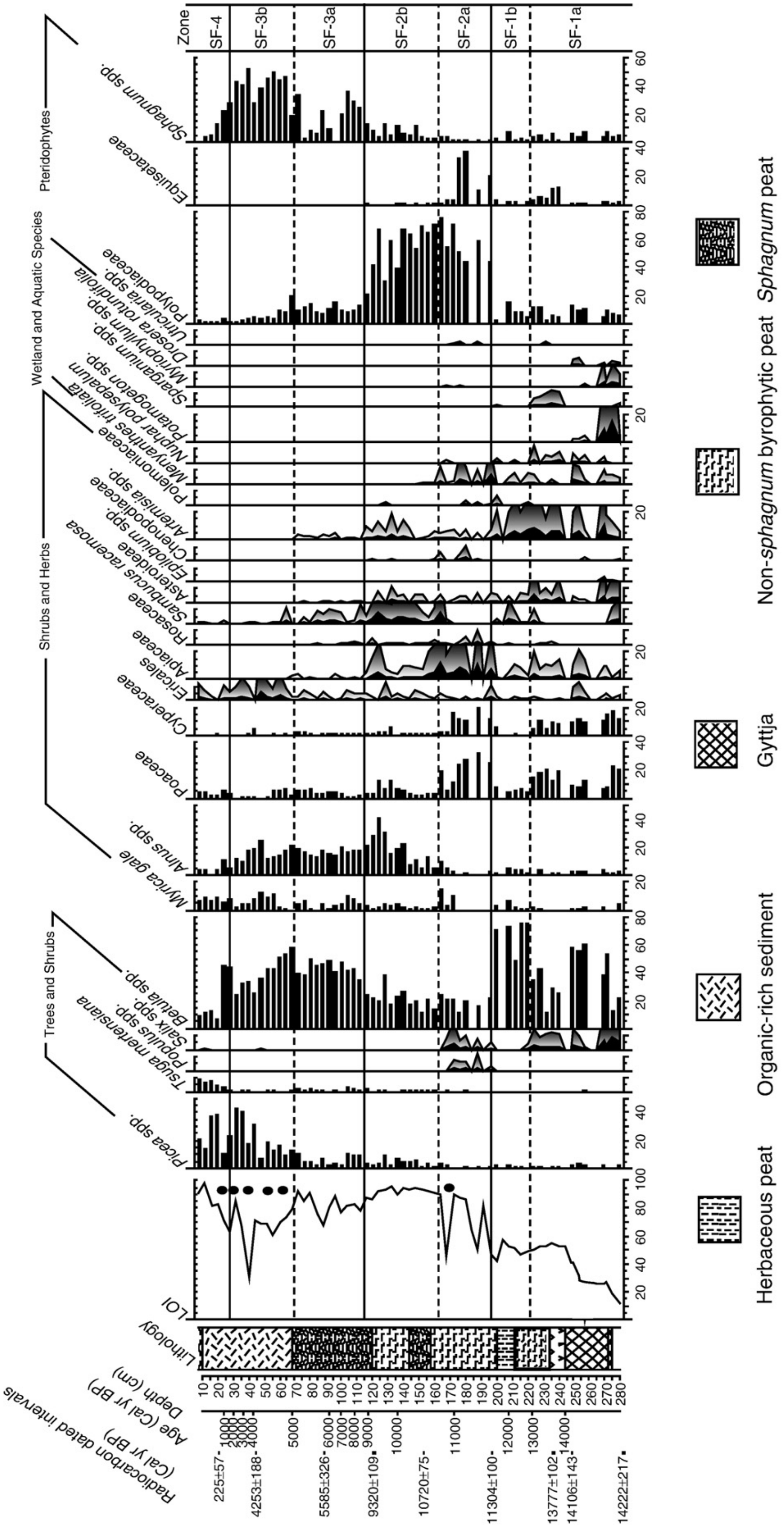

Figure 5. Pollen, spore, and LOI percentages from Swanson Fen, Alaska. Minor pollen taxa magnified 10×, and dots adjacent to the LOI indicate depths at which tephra was found.

mid-Holocene (5500–5000 cal yr BP), and late Holocene (<500 cal yr BP) (Fig. 5). Visible tephras at 170–168 cm, 58–56 cm, 54–52 cm, 40–38 cm, and 28–26 cm explain the low LOI at these depths. Anomalously low LOI at 196 cm does not coincide with a noticeable tephra, although fine tephra could have washed through the sieve during macrofossil processing. LOI significantly increases at the late glacial–Holocene boundary (198 cm, 11,500 cal yr BP) from ~50% to >80%. The LOI remains generally high except for a decline at 115 cm to ~70% and a subsequent decrease to ~50% from 60–40 cm.

Swanson Fen pollen, spore and macrofossil analysis

Because the pollen, spore, and macrofossil analyses were performed on two parallel adjacent cores, duplicate samples from overlapped segments below 200 cm were both counted to ensure that the cores followed the same chronology (Figs. 4, 5).

Zone SF-1a (280–220 cm, 14,200 to ~12,800 cal yr BP)

Herbaceous (Cyperaceae, Poaceae, *Artemisia*, Asteroideae) and shrub tundra (*Betula* [30–60%] and *Salix* [5–10%]) pollen dominate from 280 to 220 cm, along with aquatic and wetland species (*Nuphar*, *Myriophyllum*, *Potamogeton*, *M. trifoliata*, and *Sparganium*). *Picea* and *Alnus* pollen percentages are low (<5%). Polypodiaceae percentages are ~10% throughout this zone, and *Equisetum* increases to <15% toward the upper portion of the zone. *Hippuris vulgaris* and *Potamogeton* spp. seeds in the basal macrofossil record indicate the presence of fresh water. High influx of *H. vulgaris* and *Carex* spp. seeds between 280 and 285 cm coincides with the depth at which enough pollen was preserved in the sediment for analysis. An increase in the influx of *Potamogeton alpinus* and *Potamogeton zosteriformis* seeds was recorded from 265 to 247 cm, directly above the *Hippuris* peak. *Carex* spp. seeds, present throughout, have the highest influx at 260 cm. The high *Carex* influx coincides with the interval with high *P. alpinus* and *P. zosteriformis* influx, although low influx of *P. zosteriformis* seeds occurs from 226 to 224 cm. A *Betula nana* leaf at 226–224 cm confirms its on-site presence during the late glacial. *Daphnia* spp. ephippia occur in this zone, except between 270–244 cm and 230–220 cm. Influx of fungal sclerotia is high at the base of the core (290 cm), but only two other samples in this zone contain sclerotia. Sclerotia, most likely *Cenococcum*, are reproductive bodies formed under unfavorable conditions and suggest soil disturbance by frost activity, solifluction, and water transport from terrestrial sources (Birks, 2000).

Zone SF-1b (220–196 cm, ~12,800 to ~11,500 cal yr BP)

Betula pollen dominates (~80%) this zone, while Poaceae and Cyperaceae pollen (~10%) decline, and aquatic taxa disappear. *Equisetum* decreases to ~10%. *Carex* seeds steadily decline until the beginning of the Holocene. *M. trifoliata* appears and *Daphnia* spp. ephippia are present throughout zone SF-1b, indicating the presence of standing water.

Zone SF-2a (196–160 cm, 11,500 to ~10,700 cal yr BP)

Polypodiaceae spores increase drastically (from 10% to 70–80%), and the lower and mid-portion of this zone contains abundant *Equisetum* spores (20–40%). *Betula* pollen percentages decline precipitously from ~80% in the previous zone to ~20%, and *Myrica gale* increases to >20%. Wet meadow plants (Apiaceae, Cyperaceae, and *M. trifoliata*) increase during this interval. The macrofossil record shows increased influx of *Hippuris* seeds just above the Pleistocene–Holocene boundary (196–192 cm). After this maximum, *Hippuris* disappears from the record. Influx of *M. trifoliata* seeds is highest immediately following the disappearance of *Hippuris* (185–180 cm). Influx of *Carex* spp. seeds increases contemporaneously with the *M. trifoliata* peak.

Zone SF-2b (160–115 cm, ~10,700 to ~9000 cal yr BP)

As *Alnus* increases to become the dominant pollen type by ~9500 cal yr BP, *Betula* remains around 20%, and the abundances of Poaceae, Cyperaceae, and Apiaceae decline. *Picea* pollen fluctuates between 5 and 10%, and *T. mertensiana* pollen fluctuates between 0 and 2%, while *Sambucus* and Ericales pollen increase and become consistently present. *Artemisia* pollen and *Sphagnum* only slightly increase here, while Polypodiaceae spores begin to decline. *M. trifoliata* seeds disappear at 150 cm, concurrent with the decline of Cyperaceae. Macrofossils of plants typically grown on well-drained soil were found in this interval, including *Betula nana* seeds at 160–158 cm and 146–144 cm, *Betula kenaica/neoalaskana* seeds at 150–148 cm, a *Salix* bud at 150–148 cm, and a *Ledum palustre* leaf at 146–144 cm.

Zone SF-3a (115–70 cm, ~9000 to ~5000 cal yr BP)

Betula pollen increases (50–70%), while *Alnus* (15%) and *Picea* (10%) remain stable, and *T. mertensiana* and Ericales pollen occur in low amounts (<5% and <2%, respectively). *Betula nana* and *Betula kenaica/neoalaskana* seeds become more prevalent, with *Betula kenaica/neoalaskana* and *Betula* spp. (undifferentiated) influx increasing after ~5500 cal yr BP. Influx of *Carex* spp. seeds is low but increases concomitantly with *Betula*, and a *Vaccinium oxycoccus* leaf was found at 88–86 cm.

Zone SF-3b (70–30 cm, ~5000 to ~1500 cal yr BP)

Picea pollen increases to 40%, concurrent with the decline of *Betula* from 60 to 20%. *Alnus* fluctuates while *Sphagnum* spores reach their highest values (25–40%). This zone contains few macrofossils, but tephras occur throughout (66–64 cm, 60–58 cm, 56–54 cm, and 30–28 cm). *V. oxycoccus* leaves were found at 64–62 cm and 54–52 cm.

Zone SF-4 (30 cm-surface, ~1500 cal yr BP to present)

Picea spp. pollen increases to 40%, *Betula* pollen decreases to <15%, and *Alnus* declines to <5%. *T. mertensiana* increases to 10% at the surface. At the boundary between zones SF-3 and SF-4, a *P. sitchensis* needle and *T. mertensiana* branchlet were found. These remains are the sole macrofossil evidence for coniferous trees in the core. A greater presence of *Betula kenaica/neoalaskana* seeds, *Betula nana* seeds, *Betula* spp. (undifferentiated) seeds, and *V. oxycoccus* leaves occur in the upper 20 cm of the core.

Discussion

Late-glacial vegetation and climate change in Swanson Fen

The Swanson Fen record begins as a shallow pond 14,200 cal yr BP but shifts to peat ~14,000 cal yr BP, which is earlier than most circumarctic peatland sites (MacDonald et al., 2006). Pacific coastal sites, such as Homer Spit Peat Section, southern Kenai lowlands (Ager, 2000), and western Kodiak Island (Peteet and Mann, 1994) indicate similar early peatland development. The mild maritime climate of the sites fostered both early deglaciation and peat initiation.

Prior to 14,000 cal yr BP, conditions were cold and dry, as evidenced by abundant *Artemisia*, fungal sclerotia, and low LOI. A 20% increase in LOI and good peat preservation ~14,000–13,700 cal yr BP suggest increased productivity resulting from warmer temperatures, and the higher LOI is concurrent with a change from *gyttja* to bryophytic peat (Figs. 2, 5). Bryophyte remains and *Sphagnum* spores are found in the base of this record, indicating that peat-forming mosses were regionally present immediately following deglaciation. The glacial record suggests that a portion of the Kenai Peninsula near the Caribou Hills (Fig. 1) remained ice-free during the last glacial maximum (LGM) (Reger et al., 2007), harboring aquatic species (Ager, 2000) and perhaps also a number of bryophytes, which allowed for rapid colonization of deglaciated sites following ice retreat.

Low *Picea* pollen percentages occur in Swanson Fen during the late glacial, and regional records suggest that *Picea* did not migrate to the Kenai until ~8500 cal yr BP (Ager, 2000). However, *Picea* pollen is found in other late-glacial sites in southwestern Alaska (Brubaker et al., 2001) and eastern Beringian (Brubaker et al., 2005) records. Brubaker et al. (2005) propose that *Picea* populations persisted in eastern Beringia through the LGM, and that the apparent migration rates are too rapid for the species to have expanded from south of the Cordilleran Ice Sheet following ice retreat. Furthermore, recent DNA research (Anderson et al., 2006a) contends that high haplotype diversity of *P. glauca* indicates that it survived the last glaciation in an Alaskan refugium. Macrofossils from mature *Picea* trees found in the Yukon Territory date to the onset of the LGM, the youngest local Pleistocene *Picea* evidence (Zazula et al., 2006). The authors point out that while palynologists often adhere to a 10% *Picea* pollen rule to indicate presence, numerous cases exist where *Picea* pollen is <5%, but where a presence of *Picea* macrofossils indicates nearby populations (Zazula et al., 2006). Thus, it is possible that *Picea* pollen was windblown from interior sites to the Kenai during the late glacial, or that a refugium could have existed on the Kenai.

Younger Dryas evidence in Swanson Fen

A subtle negative excursion in LOI, an increase in *Betula*, and a decrease in Poaceae, Cyperaceae, and *Salix* coincide with the Younger Dryas (YD) chronozone (Zone SF-1b in Figs. 4, 5), and similar LOI and vegetation shifts are observed in other Kenai Peninsula sites (Ager, 2000; Anderson et al., 2006b). However, the change to shrubbier vegetation contrasts with the pattern observed in most of Alaska during the YD. The pollen record from Phalarope Pond, located on Kodiak, exhibited a decline in fern abundance concurrent with an increase in *Empetrum*, which, along with well-dated coastal section LOI profiles, Peteet and Mann (1994) interpreted as indicating a YD shift to a cold, dry climate. Other Alaskan profiles have shown a reversal to colder, drier conditions, with herb expansion and diminishing trees and shrubs (Engstrom et al., 1990; Hu et al., 1995, 2002; Brubaker et al., 2001; Bigelow and Edwards, 2001), but a YD signal is absent from numerous records in Alaska, suggesting significant spatial climatic heterogeneity (Kokorowski et al., 2008).

While the YD was first associated with the North Atlantic and its surroundings, North Pacific oceanic records have also indicated cooler sea surface temperatures (SSTs) (Kallel et al., 1988), and global climate model (GCM) results have simulated the effects of North Pacific SST changes during the YD (Mikolajewicz et al., 1997; Peteet et al., 1997). In the North Pacific YD experiment, cooler North Pacific SSTs resulted in reduced cloudiness and increased snowfall along the southern Alaskan coast (Peteet et al., 1997). Other GCM simulations (Mikolajewicz et al., 1997) show a pattern consistent with an eastward shift in the AL associated with YD oceanic cooling, implying decreased cloudiness on the Kenai Peninsula. Greater sea ice extent in the Bering Sea during the YD could have influenced atmospheric circulation patterns that contributed to the differences in the southwestern and south-central coastal sites. Alternatively (or in addition), an expanded and stronger subtropical high (Kokorowski et al., 2008) during the summer months may have resulted in warm, sunny summers.

The growth and expansion of *Betula nana* and other shrubs appears to be positively correlated with warm temperatures in arctic Alaska (Sturm et al., 2001; 2005) and with drying in wetter, subarctic peatlands, such as those found on the Kenai (Klein et al., 2005). However, because the YD is widely recognized as a cooling event, other factors must have caused the *Betula* increase. Modern studies have shown the importance of nutrients and light over temperature and precipitation in plot manipulations (Chapin et al., 1995). Greater growth of *Betula nana*, *Salix* spp., and *Alnus crispa* shrubs in the Arctic today (Sturm et al., 2001) creates a positive feedback with winter snowfall, which insulates the ground, increases microbial activity and plant-available nitrogen, and promotes shrub development (Sturm et al., 2005). Enhanced southerly atmospheric flow during the YD (Peteet et al., 1997) may have increased winter snowfall on the Kenai, which would have insulated the ground enough to result in shrub survival. Dry, sunny summers may have also fostered shrub proliferation, as shading today reduces *Betula nana* growth (Chapin et al., 1995). Alternatively, *Betula* expansion and contraction of graminoids may reflect the formation of palsas under the cold YD climate. In the Arctic today, *Betula nana* commonly grows on the margins of elevated water tracks that receive seasonal drainage (such as run-off from spring snowmelt) and on well-drained palsas, whereas graminoids are more typical of wet sedge tundra (Walker et al., 1989).

Polypodiaceae-dominated early Holocene: paleoclimatic implications

A sharp increase in Polypodiaceae (fern) spore percentages (~80%, Fig. 5) occurs 11,500–8500 cal yr BP. This feature appears in cores throughout southwestern and south-central Alaska (Hu et al., 1995; Ager, 2000, 2001; Brubaker et al., 2001; Anderson et al., 2006b), but the meaning of such abundant Polypodiceae spores is unclear, as no modern analog exists on the Kenai today. The conspicuous Polypodiaceae interval raises questions as to whether the early Holocene was a period of high disturbance or whether the ferns reflect a climatic signal (or some combination). Polypodiaceae are often early successional plants found on disturbed or barren ground such as in thermokarst sites or newly deglaciated terrain. Thermokarst development results in species changes similar to those observed in the early Holocene in Swanson Fen, beginning with taxa such as *Carex* spp., *M. trifoliata*, and *Equisetum* spp. forming in the paludified sites, followed by *M. gale* and *Salix* spp. (Jorgenson et al., 2001). However, ecological succession occurs over a period of decades or centuries, and thus is an unlikely explanation for the millennial duration of the fern peak.

One possibility is that the presence of wet meadow taxa and disappearance of aquatic plants represents infilling of the pond at Swanson Fen. Warming and drying in the early Holocene could have increased evaporation, converting much of the kettle lake landscape into kettle fens by lowering the water table and allowing terrestrial peatland plants, such as *M. trifoliata*, wet meadow taxa, Polypodiaceae, and *Equisetum* spp. to colonize drying lakebeds. This process is unlikely to have lasted for over a thousand years, however.

This interval on the Kenai can be divided into two distinct periods based on the pollen assemblages and will be discussed separately. The first (11,500–10,700 cal yr BP) is characterized by a decrease in *Betula*, an increase in wet meadow taxa, *Salix* spp. and *Populus* spp. The second (10,700–8500 cal yr BP) is characterized by a decrease in wet meadow taxa, an increase in *Sambucus racemosa* and *Artemisia*, and an expansion of *Alnus* spp.

In the first half of the early Holocene (11,500–10,700 cal yr BP), an increase in wet meadow and wetland taxa, along with high accumulation of well-preserved peat, is indicative of higher effective moisture. Regionally, records from Kodiak (Peteet and Mann, 1994) and southwestern Alaska (Hu et al., 1995; Brubaker et al., 2001) show similar increases in wet meadow taxa, *Populus*, and *Salix*, and the spatial pattern of these assemblages on the Kenai and the Alaska Peninsula suggests a regional climatic shift. Expansion of *Populus* has previously been attributed to warming (Ritchie et al., 1983) and/or soil disturbance (Hu et al., 1996). The presence of *Populus* and *Salix* in the early Holocene may result from the colonization of outwash as alpine glaciers continued to recede upslope, since *Populus* is found along stream banks and recently deposited alluvium (Viereck and Little, 2007). Polypodiaceae may have also thrived on these newly exposed surfaces (Ager, 2000). Tree line and avalanche scar sites with high fern cover adjacent to the Malaspina Glacier show percentages of Polypodiaceae near 75%, but *Alnus* (75% pollen) also dominates the sites (Peteet, 1986), which is not the case for this interval on the Kenai. Hu et al. (1995) suggest that Polypodiaceae and Apiaceae communities thrive in areas with thick winter snow cover followed by rapid snow melt in the spring warming, a plausible scenario for the highly seasonal early Holocene (Berger, 1978). The dry summers that ensued could explain the presence of Chenopodiaceae on the Kenai Peninsula, as they occur in drier settings.

The early Holocene pattern of vegetation change on the Kenai Peninsula 11,500–10,700 cal yr BP may suggest a strengthening of the Aleutian Low over the Bering Sea region, which is consistent with atmospheric observations that enhanced land–sea temperature contrasts create a deeper AL (Trenberth and Hurrell, 1994). Warm SSTs result in advection of warm, moist air northward, with increased moisture west of the Kenai Mountains (Trenberth and Hurrell, 1994). This agrees with other records from Alaska, particularly southwestern Alaska lake records (Hu et al., 1996; Axford and Kaufman, 2004), which are interpreted as warm and moist. The opposite moisture regime is cited from eastern Beringia (Ritchie et al., 1983), south-central, and southeastern Alaska (Heusser et al., 1985), where pollen records indicate warm and dry conditions. No paleo-lake level records exist for the Kenai, but Birch Lake from interior Alaska shows a brief interval of increasing water levels from the beginning of the Holocene to ~9500 ^{14}C BP (~11,000 cal yr BP) (Abbott et al., 2000), which roughly corresponds to our interpretation of an initially wet early Holocene, followed by a drier climate.

Rising sea level from global ice retreat contributed to the demise of the Bering land bridge around the Pleistocene–Holocene boundary and would have resulted in large-scale changes in ocean–atmosphere dynamics when the Arctic Ocean connected with the North Pacific (Peteet et al., 1997). The combination of warm temperatures caused by the maximum of summer solar insolation (Berger, 1978), the melting Laurentide Ice Sheet, and decline of sea ice in the Bering Sea during this time are thought to have enhanced moisture availability to the region, resulting from the weakening of the Laurentide glacial anticyclone (Heusser et al., 1985). One scenario for increased moisture in southwestern, interior, and south-central Alaska is an eastward shift of the East Asian trough, and a stronger than normal Pacific subtropical high. This pattern increases the frequency of moisture and storms from the southwest, permitting this moisture to reach the interior of Alaska because of the lack of orographic obstacles (Mock et al., 1998).

Lower LOI, diminished peat preservation, and decreased Apiaceae, Cyperaceae, *M. trifoliata* and fungal sclerotia between 10,700 and 8500 cal yr BP signify a change to a drier climate. However, abundant ferns and *Alnus* expansion across the Kenai Peninsula (Ager, 2000, 2001; Anderson et al., 2006b) and high peat accumulation in Swanson Fen indicate high moisture availability. This disparity may be explained by the continuation of a strong seasonal gradient. The shift in amplitude of solar insolation at this time likely altered winter sea ice extent, which would have influenced the position and strength of the AL, bringing winter precipitation to the Kenai Peninsula. Meanwhile, weak southerly flow in the summer may have brought sufficient moisture to the region to maintain high peat accumulation and abundant ferns, while the warm climate and high evapotranspiration would have resulted in periodic dry conditions, as evidenced by fires (Anderson et al., 2006b), peat humification, and *Cenoccocum* sclerotia. Alternatively, the high ferns may indicate increased disturbance due to fire. Increased fire frequency is recorded in Paradox Lake on the Kenai lowlands beginning 10,500 cal yr BP (Anderson et al., 2006b), which may suggest that ferns were colonizing these disturbed forest sites. The Swanson Fen record does not contain charcoal, which indicates that this catchment may only record very localized fire signals and that the fires recorded in Paradox Lake did not reach as far north as Swanson Fen. The fern peak occurs in sites across the Kenai Peninsula while the fire signal does not, suggesting that fires are not the cause of the widespread fern expansion.

Mid- to late Holocene vegetation changes

A major decline in Polypodiaceae abundance occurs ~8500 cal yr BP, contemporaneous with an increase in *Sphagnum*, a decline in *Alnus*, and an expansion of *Betula*, including both *Betula nana* and *Betula neoalaskana* or *B. kenaica* (Figs. 4, 5). Well-preserved peat (Fig. 2) after ~8200 cal yr BP suggests either cooler or moister conditions (or a combination), which would decrease decomposition. *T. mertensiana* pollen increases 8500 cal yr BP, but its presence cannot be substantiated because of a lack of macrofossil evidence. Peteet (1986) found that despite low *T. mertensiana* pollen rain near Yakutat, trees of that species were growing near the site. Its potential presence on the Kenai lowlands raises the possibility that a small population may have survived in Kenai lowland refugium and may not have been previously recorded because of its low pollen rain. This topic requires further investigation. However, because pollen percentages remain low, it is also likely that *T. mertensiana* pollen could have blown across the Kenai Mountains to this lowland site.

Numerous lines of evidence, including decreasing LOI, highly decomposed peat, and lack of macrofossils other than fungal sclerotia, suggest drier conditions 5500–4000 cal yr BP. Although the LOI declines during this interval, a sharp negative excursion at ~40 cm coincides with increased tephra horizons, as revealed by LOI and macrofossil analysis (Fig. 5). Abundant *Sphagnum* and the disappearance of most brown moss species except *Aulacomnium palustre* occur concurrently with Ericales expansion, which implies acidification and drying of the fen, as certain *Sphagnum* species and *Aulacomnium palustre* form hummocks (Janssens, 1990) and *Sphagnum* acidifies its environment through cation exchange (Clymo, 1966). While *Picea* colonized the Kenai Peninsula much earlier, it did not expand until ~5000 cal yr BP on the Kenai lowlands, as evidenced by sustained pollen percentages >10%.

In recent millennia, the expansion of *T. mertensiana*, concurrent with the declines in *Betula*, *Alnus*, and sclerotia, suggest a cooler, moister climate. Increased precipitation is inferred from Prince William Sound and southeastern Alaska pollen records (Heusser et al., 1985; Peteet, 1986). The Kenai Peninsula and Prince William Sound experienced glacial advances beginning 3600 cal yr BP and culminating in the Little Ice Age, which are attributed to increased winter precipitation and cooler summer temperatures (Wiles and Calkin, 1994). This may explain the most recent expansion of *T. mertensiana*, as cool, moist conditions would have allowed this species to expand its range from the Kenai Mountains to the lowlands.

Conclusions and implications

1. A maritime climate on the Kenai lowlands promoted early peat initiation following deglaciation around 14,200 cal yr BP, and cool late-glacial temperatures resulted in shrub tundra dominance until the Holocene warming.
2. The YD on the Kenai is characterized by *Betula* expansion and a decline in herb tundra, which contrasts with many other Alaska sites. The YD response highlights the regional differences and complexity in Alaskan climate during this interval. Increased snowfall may have raised ground temperatures and promoted *Betula* proliferation through enhanced nutrient availability. Alternatively, a cold, dry YD may reflect local permafrost dynamics with the formation of palsas, which were elevated above the peat surface, decreasing wet sedge habitat and favoring *Betula nana* (Walker et al., 1989).
3. Remarkable Polypodiaceae abundance from 11,500 to 8500 cal yr BP implies a significant change to the climate. Enhanced peat preservation and the occurrence of wet meadow species suggest high moisture from 11,500 to 10,700 cal yr BP, in contrast to drier conditions in southeastern Alaska; this pattern implies a possible repositioning of the Aleutian Low and a strengthening and repo-

sitioning of the Pacific subtropical high. Drier Kenai climate 10,700–8500 cal yr BP signifies a weakening of southerly flow in the summer, but elevated fern abundance was sustained due to high seasonality with substantial snowfall and enhanced glacial melt.

4. *Sphagnum* and *Betula* expansion, along with the presence of *T. mertensiana*, suggests the prevalence of a cooler, moister climate after 8200 cal yr BP, owing largely to a decrease in the insolation-induced seasonality (Berger, 1978). Higher effective moisture is recorded throughout the southern coast of Alaska (Heusser et al., 1985; Hu et al., 1995), suggesting a change to more uniform spatial climate patterns. An interval of increased volcanic activity after 5000 cal yr BP coincides with drier conditions, as indicated by poor peat preservation, lower LOI, increased fungal sclerotia, and a lack of plant macrofossils.
5. Establishment of *T. mertensiana* on the Kenai Lowlands in the late Holocene (1500 cal yr BP) corresponds with the widespread Neoglaciation, characterized by glacial advance associated with a cool, moist climate on the Kenai (Ager, 2000; Wiles and Calkin, 1994).

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