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Late-glacial and Holocene δ^{15}N and δ^{13}C variation from a Kenai Peninsula, Alaska peatland

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ABSTRACT

We use stable carbon and nitrogen isotopes from a Kenai Peninsula, Alaska fen spanning the late-glacial and Holocene to enhance our understanding of peatland hydrology and nutrient cycling. We find that isotopic values for carbon and nitrogen are lighter in the late-glacial and early Holocene (−28 to −34‰ and 0 to −4‰, respectively) than the remainder of the record. The light δ^{13}C value corresponds with submerged and emergent aquatic vegetation during the late-glacial, suggesting high respired carbon input. Similarly, light δ^{15}N values imply that nitrogen is derived from sediment and groundwater sources. Early Holocene δ^{13} values ranging from −30‰ to −34‰ infer high recycled carbon uptake from methanogenesis, related to the warm climate and the wet swale environment. A shift to heavier δ^{15}N and δ^{13}C isotopic values ~9600 cal yr BP coincides with a change in the lithology to *Sphagnum*-dominated peat, indicative of peat accumulation above the water table. A shift to lighter values from 5500 to 4200 cal yr BP coincides with a period of glacial recession related to warmer conditions, whereas a subsequent period of heavier values is correlative with glacial advance culminating at 3000 cal yr BP, suggesting that the hydrology of this peatland is closely linked with the glacial meltwater. δ^{15}N remains near 0‰, relatively close to the atmospheric value of δ^{15}N, throughout the Holocene, suggesting that the fen has been supported by N-fixation. We propose that groundwater on the Kenai lowlands was fed by subglacial meltwater, and the warm early Holocene summers resulted in high glacial ablation and groundwater recharge, sustaining a high Swanson Fen water table. Cooler temperatures resulted in a decrease of glacial meltwater input at the close of the early Holocene.
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1. Introduction

Pollen, spores, and plant macrofossils are useful tools in peat core analysis to determine past environmental and climatic changes. However, no-analog ecological communities during the early Holocene in Alaska (Anderson et al., 2004; Edwards et al., 2005) have made it difficult to determine the nature of Alaskan early Holocene climate. While it is well established that the early Holocene was warmer than present (Kaufman et al., 2004), the precipitation and/or moisture availability on the landscape is more difficult to decipher. Low lake levels from the Kenai Peninsula (Anderson et al., 2006) and central Alaska (Bigelow, 1997; Hu et al., 1998; Abbott et al., 2000) suggest that the climate was relatively dry, but peat cores from the Kenai Peninsula (Jones et al., 2009; Jones and Yu, 2010) reveal rapid accumulation of well-preserved peat as well as rapid peatland expansion across parts of Alaska, suggesting that sufficient moisture was available for both. Detailed records of mid- to late Holocene changes rely mostly on glacial evidence to interpret changes in temperature and moisture (Barclay et al., 2009), although it can be difficult to decipher whether ice margin changes result from increased or decreased moisture or from seasonal changes in temperature (or some combination).

The use of stable carbon isotope ratios as a tool for independent climatic information from wetland environments has only recently been explored (Sukumar et al., 1993; White et al., 1994; Pancost et al., 2003; Sharma et al., 2005; Skrzypek and Jedrysek, 2005; Loader et al., 2007; Skrzypek et al., 2007a,b; Loisel et al., 2009, 2010; Tillman et al., 2010), but has the potential to improve our understanding of carbon and nitrogen cycles in peatlands, as well as to provide a method for interpreting long-term temperature, moisture, and nutrient changes. The goal of this study is to better characterize the nature of Holocene climate and peatland environmental conditions. We utilize stable carbon and nitrogen isotopes from a peat core on the Kenai Peninsula, Alaska as an independent proxy for paleoenvironmental change.

Fens and bogs are often dominated by mosses, which account for a significant portion of productivity, influence successional and community dynamics, and play an important role in nutrient cycling (Clymo and Hayward, 1982; Rice, 2000). In vascular plants, the carbon isotopic signature is controlled by stomatal conductance, which is

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related to temperature and moisture availability (Farquhar et al., 1989), but because mosses are nonvascular, they lack stomata and are thus unable to regulate the uptake of atmospheric CO_2 (White et al., 1994; Rice and Giles, 1996). Carbon isotopic discrimination in mosses therefore is controlled by the presence and thickness of a water film on the leaf, which creates a barrier to carbon dioxide diffusion (Rice and Giles, 1996; Williams and Flannagan, 1996). In addition to diffusion, enzymatic reactions such as the carbon fixation by Rubisco contribute significantly to the fractionation (Rice and Giles, 1996). Carbon isotopic values have also been shown to correspond with the degree of recycled carbon uptake from methanogenesis (Raghoebarsing et al., 2005; Nichols et al., 2009), as found in particularly wet environments (Nichols et al., 2009). Carbon isotopic relationships between hummocks and hollows have been observed (Price et al., 1997; Rice, 2000; Loisel et al., 2009), related to physiological and morphological differences in hummock and hollow species (Rice, 2000). Lawn and hollow species are not as efficient as hummock species in drawing up water from below through capillary action and therefore display higher sensitivity to water table fluctuations (Rydin, 1985).

Data on nitrogen isotopes in peat are more limited, but nitrogen stable isotopic ratios can be effectively utilized together with C/N, carbon isotopic ratios, and paleoecological tools (Wooller et al., 2003), and have been used to trace the flow and utilization of organic matter in food webs and as an indicator of past nutrient status in lakes (Brenner et al., 1999), estuaries (Orem et al., 1999), and wetlands (Novak et al., 1999; Wooller et al., 2003). More recently, nitrogen isotopic ratios in peat-forming *Sphagnum* mosses have been compared to water table position, and their respective ratios reflect differences in nitrogen acquisition strategies (Asada et al., 2005).

1.1. Setting

The Kenai Peninsula (Fig. 1), located in south-central Alaska, is connected to the mainland to the north by Portage Passage and

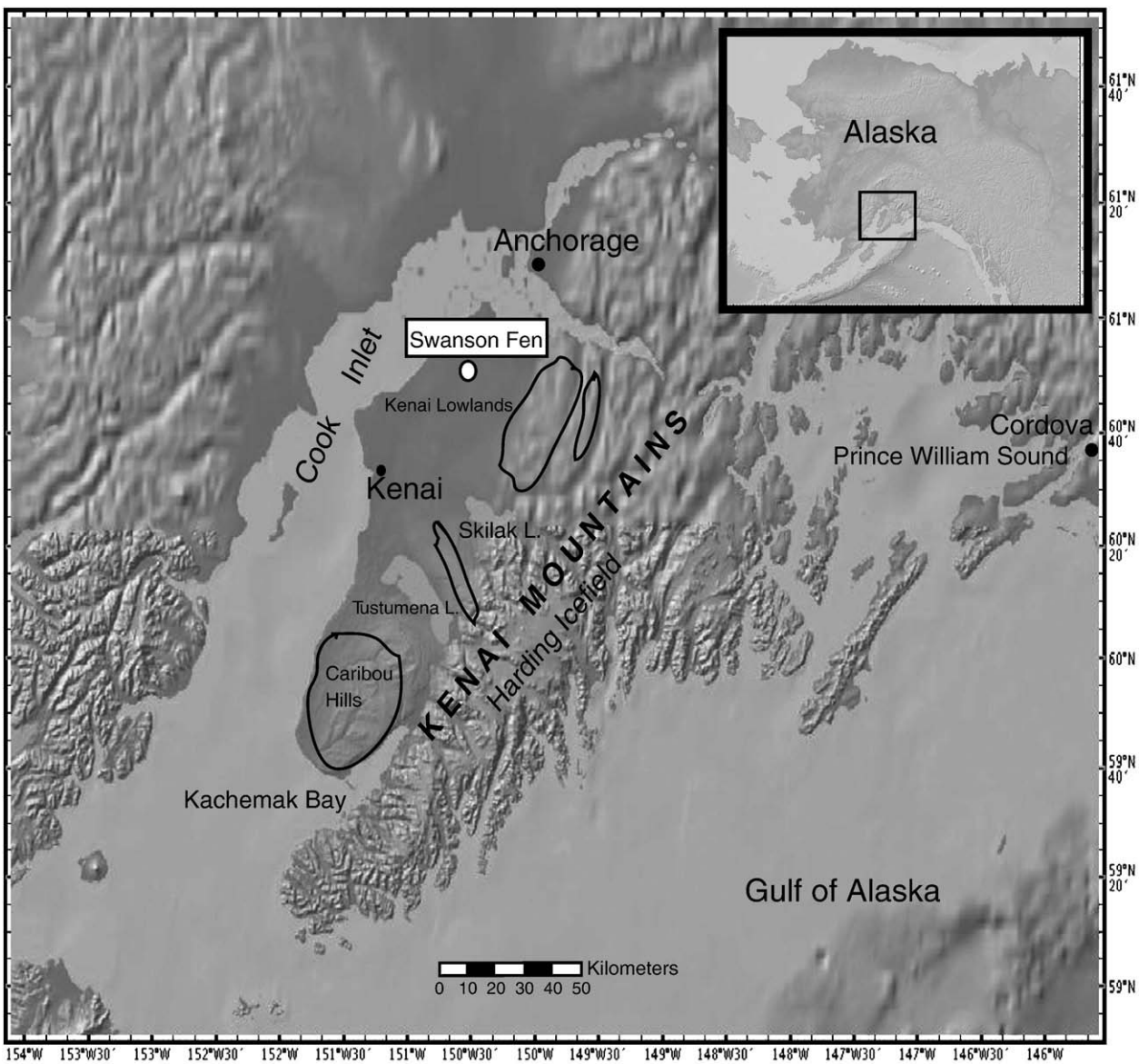

Fig. 1. Kenai Peninsula Map. Swanson Fen, marked with a white dot, is located on the northern Kenai lowlands. The black lines outline glacial refugial areas, according to the glacial geology (Reger et al., 2007).

bordered westward by Cook Inlet. To the south of the Peninsula is the expansive Gulf of Alaska, and Prince William Sound lies to the east. The Kenai Mountains, which reach 2015 m and are mantled by the Harding, Sargent, and Spencer-Blackstone Icefields, extend across the eastern side of the Kenai Peninsula, creating a rain shadow with high precipitation on the eastern side of the mountains and much drier conditions on the western lowlands. The northern Kenai lowlands were deglaciated during the Killey Stade of the Naptowne glaciation, between 18,500 and 17,500 cal yr BP (Reger et al., 2007) and Swanson Fen is a small (300 m^2) nutrient poor fen situated within these lowlands (60°47.35′N 150°49.91′W) (Fig. 1). A mixed *Tsuga mertensiana* and *Picea x lutzii* forest surrounds the fen today. Dominant modern fen bryophyte species are *Sphagnum capillifolium* and *Aulacomnium palustre*, while *Polytrichum strictum* occupies some of the higher hummocks. The water table is ~20 cm below the peat surface in the center and the measured pH value in August of 2007 was 4.81. The low water table accounts for the shrubbiness of the fen today, with *Chamaedaphne calyculata* and *Betula nana* representing the shrub species and *Equisetum arvense* and *Calamogrostis* occupying the wetter site margin.

Kenai lowland climate is semi-continental, with mean January temperatures for the nearby town of Kenai ranging from −15.6 °C to −6.2 °C, mean July temperatures ranging from 8.5 °C to 16.5 °C, and annual total water equivalent precipitation at Kenai equaling 48 cm (30-year average) (Alaska Climate Research Center, http://www.Cdc.noaa.gov/USclimate). Snowfall occurs from October to May, with maximum snow depths occurring in December and total annual snow depths reaching 154 cm (NOAA-CIRES Climate Diagnostics Center, http://climate.gi.alaska.edu/climate/). Growing season length on the Kenai lowlands averages 212–230 days per year. The Kenai Peninsula weather patterns today 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).

2. Methods

The peat core in Swanson Fen (290 cm) was extracted using a Hiller corer in the summer of 2005. The core was sub-sampled and bagged in the field at 2-cm intervals, and stored in refrigeration in the Lamont-Doherty Earth Observatory (LDEO) core repository. In the lab, loss-on-ignition (LOI) was performed by first drying each sample, followed by burning the sample at 550 °C to yield the organic matter content of each sample (Dean, 1974). Pollen was processed using standard procedures (Faegri and Iverson, 1989) and is discussed fully in Jones et al. (2009). The lithology was described using the Troels-Smith's method (1955).

2.1. Stable isotope analysis

The samples were dried at 50 °C for 48 h and homogenized. Approximately 2 μg of each sample were measured out into tin cups and weighed using a Cahn electrobalance. Samples were analyzed using a Europa 20/20 continuous flow (CF) isotope ratio mass spectrometer (IRMS) coupled with an ANCA NT combustion system (PDZ-Europa, Cheshire, UK) at LDEO. Secondary standards calibrated against NIST primary standards were included in every analytical run. Results are based on the mean of three replicates of each sample and expressed as δ-values relative to the VPDB standard for $\delta^{13}C$ and atmospheric nitrogen (AIR) for $\delta^{15}N$. The δ-values are defined as

$$\delta^{15}N(\delta^{13}C) = \left[\left(R_{\text{sample}} / R_{\text{standard}}\right) - 1\right] \times 1000, \quad (1)$$

where R_{sample} and R_{standard} are the $_{15}N/_{14}N$ ($_{13}C/_{12}C$) ratios of the samples and standards, respectively. C/N ratios were calculated by dividing the moles/mg of carbon by the moles/mg of nitrogen, as determined by the mass spectrometer. The average analytical error for the $\delta^{15}N$ was 0.33‰ and 0.29‰ for $\delta^{13}C$. Samples were not treated with HCl, as the underlying geology is composed primarily of greywacke-argillite granite and greenstone rock types, and extensive geologic mapping has found the lowlands to be devoid of detectable amounts of calcium carbonate (Karlstrom, 1964). Holocene tephra deposits can be traced to the five main, active volcanoes in the Cook Inlet region, which do not contain carbonates (Riehle, 1985).

2.2. Macrofossils and chronology

Macrofossils were separated from the matrix using a 500 μm sieve and selected samples were dated using AMS radiocarbon dating techniques at Lawrence Livermore Labs, California (Table 1). Radiocarbon dates were calibrated to calendar ages using CALIB (version 5.1) (Stuiver and Reimer, 1993; Reimer et al., 2004).

Bryophytes picked from the macrofossil analysis were mounted onto slides. A dissecting microscope was used to isolate the moss specimens at 60× and a Zeiss microscope was used for moss identification at 400×. Species were identified using Crum and Anderson (1981) and a reference collection of mosses from the Kenai Peninsula, housed at the Kenai National Wildlife Refuge headquarters. The chronology used in this study is based on the AMS ^{14}C chronology from plant macrofossils.

3. Results

Detailed lithological, pollen, spore, and macrofossil data for the Swanson Fen core are given in Jones et al. (2009). As a brief summary

Table 1
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}Cyear 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 (3)	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 (12)	12290 ± 40	14222 ± 217 (100%)

the record begins ~14,200 cal yr BP, and semi-aquatic macrofossils indicate that the site was a wet swale environment during the late-glacial, with a progression from a wet sedge and brown moss-dominated fen with little *Sphagnum* moss during the early Holocene (11,300 to ~9600 cal yr BP) to a brown moss-*Sphagnum*-mixed bryophyte fen from 9600 cal yr BP to 9000 cal yr BP, to a mostly *Sphagnum*-dominated fen (~9000 cal yr BP to present; Fig. 4). The highest peat accumulation rates occur during the early Holocene (11,300 to 9320 cal yr BP).

3.1. $\delta^{13}C$ changes in Swanson Fen

The isotopic information from Swanson Fen is derived from bulk peat samples, and although many studies rely on cellulose extracted from individual species to perform isotopic analysis, the results of recent studies suggest that $\delta^{13}C$ values of bulk organic matter of peat are highly correlated with the values of extracted cellulose (Skrzypek et al., 2007a), with the bulk peat of living samples having a lighter $\delta^{13}C$ signature, consistently offset by <1.5‰ (Menot and Burns, 2001). In a comparison of $\delta^{13}C$ values of *Sphagnum* cellulose from stems and branches, a similar offset in absolute isotopic values exists (Loader et al., 2007; Moschen et al., 2009), but the offset occurs in a consistent direction. Thus while differences in carbon isotopic composition were found in stems and branches of living *Sphagnum* plants, the same degree of common forcing among the bulk *Sphagnum* is preserved downcore, suggesting that bulk *Sphagnum* peat can still show significant environmental changes (Loader et al., 2007). Therefore, we assume the bulk samples represent an average over a growing period represented by the peat, whereas the isotopic values of the stems and branches represent differences in growing conditions over random growing periods within the sample period. The slight offset in overall isotopic composition may occur because of slight differences in the decomposition rates (Loader et al., 2007). These findings suggest that the relative isotopic shifts in the Swanson Fen record can be interpreted based on environmental and climatic changes. Based on the observations in the aforementioned studies, we only will discuss relative changes in isotopic composition instead of absolute values.

Carbon isotopic ratios (Fig. 3a) during the late glacial period show high variability before 14,000 cal yr BP, but stabilize to ca. −28‰ between 14,000 and 11,500 cal yr BP. Carbon isotopic values become extremely negative during the early Holocene, from 11,500 to 10,500 cal yr BP, initially decreasing to −34‰ followed by an average closer to −30‰ until 10,000 cal yr BP. A significant shift to less negative values (−24‰ to −22‰) occurs between 10,000 and 9800 cal yr BP. Isotopic values then remain constant until 5500 cal yr BP, when they decrease to ca −26‰. The C/N ratio, reported by dividing moles/mg of C by moles/mg of N as reported by the mass spectrometer, in Swanson Fen (Fig. 3) is lowest (near 1) in the late-glacial (14,200–11,500 cal yr BP), which suggests that the source of the sediment is aquatic (Meyers, 1994, and agrees with the lithology and macrofossils in this portion of the core that includes aquatic macrofossils. Given the abundance of brown mosses and aquatic and semi-aquatic plant macrofossils in this section of the core profile (Fig. 4), it is likely that this was a fen with seasonally-flooded mud-bottomed pools, which may have high algal productivity resulting in low C/N (Meyers, 1994. The C/N ratio increases briefly to over 200 at the beginning of the Holocene before returning to values between 30 and 50 for the remainder of the Holocene. This significant excursion in the C/N ratio at ~11,500 cal yr BP corresponds to a large negative excursion in $\delta^{13}C$ (−34‰).

3.2. $\delta^{15}N$ changes in Swanson Fen

Nitrogen isotopic ratios can be more difficult to interpret than C because of the complexity of the nitrogen sources, transfer, and plant uptake (Sharma et al., 2005). Nitrogen enters peatlands both in organic and inorganic forms, and plants use different assimilation strategies in these nitrogen-limited ecosystems (Jauhiainen et al., 1998; Asada et al., 2005). Inorganic forms commonly occur in solutions as nitrate (NO_3^-) and ammonium (NH_4^+) to a lesser extent, and a significant amount of nitrate enters the system from precipitation, groundwater, and surface water (Kendall, 1998; Sharma et al., 2005). Without any anthropogenic input for the core analyzed in this study, the nitrate levels resemble background atmospheric deposition. A difference in uptake of NO_3^- and NH_4^+ in hummocks versus hollows may exist under similar climatic conditions (Jauhiainen et al., 1998), as nitrogen assimilation in peat is higher in hummock species than in lawns or hollows (Jauhiainen et al., 1998). Because of the complexity of fractionation within the nitrogen cycle, we will only discuss the significance of relative isotopic shifts.

The nitrogen isotopic (Fig. 2) values show variability during the late-glacial from 14,200 to 13,000 cal yr BP, ranging from 0 to −6‰. From 13,000 to ~11,000 cal yr BP, $\delta^{15}N$ values average close to −2‰. From 11,000 cal yr BP to ~10,000 cal yr BP, values display a large range from −1‰ to −4‰. A significant shift in $\delta^{13}\delta^{15}N$ occurs from −4‰ to almost 1‰ at 10,000 to 9800 cal yr BP. After 9800 cal yr BP $\delta^{13}\delta^{15}N$ values stabilize near 0‰, which is the atmospheric isotopic value. This shift occurs simultaneously with the $\delta^{13}C$ shift.

3.3. Vegetation changes from selected pollen and spore profiles

The major isotopic changes in Swanson Fen correspond to changes in pollen and spore percentages (Jones et al., 2009). The early Holocene (11,500 to 9500 cal yr BP) is marked by a 20–40% decrease in *Betula* pollen and a 40–60% increase in Polypodiaceae fern spores. The close of the early Holocene shows a significant decline in fern spore and increase in *Alnus* pollen, contemporaneous with the large shift in $\delta^{15}N$ and $\delta^{13}C$ to isotopically heavier values. It also reveals an increase in *Sphagnum* spores, which could represent a regional shift of brown moss-dominated fens to *Sphagnum*-dominated fens and an increase in summer precipitation (Sundberg, 2002). While lower spore production is sometimes a result of low temperatures, the early Holocene temperatures in Alaska were warmer than present (Kaufman et al., 2004), so a colder climate cannot explain the low *Sphagnum* spore values during the early Holocene.

Smaller-scale isotopic shifts occur in the mid- to late-Holocene, and while we exercise caution in placing too much value on these smaller shifts, some distinct relationships emerge. A ~2‰ shift to lighter values from ~6500 cal yr BP to 4800 cal yr BP corresponds to a period of higher peat accumulation, a distinct spike in ferns (Polypodiaceae), similar to the early Holocene. A subsequent increase of ~2‰ culminates at ~3000 cal yr BP, which corresponds to a lower accumulation rate, an increase of *Sphagnum* spores and a decrease in ferns, *Betula* pollen and an increase in *Picea* pollen. These changes may reflect a drier climate, as Polypodiaceae are more often found in areas of high moisture. The increase in *Picea* could reflect an expansion of *Picea mariana* on drier peatlands.

4. Discussion

The geochemical proxies used in this analysis are controlled by a number of environmental and diagenetic factors. Diagenetic processes in anaerobic sediments were first examined to determine their significance in the Swanson Fen core and whether these factors appreciably obscure the original environmental imprint. Second, the environmental factors are assessed in the context of late Pleistocene and Holocene changes on the Kenai Peninsula, Alaska using existing lithological and paleoecological information (Jones et al., 2009).

4.1. C/N and diagenesis

Diagenesis may play a significant role in the isotopic composition of peat, and numerous studies have examined how degradation

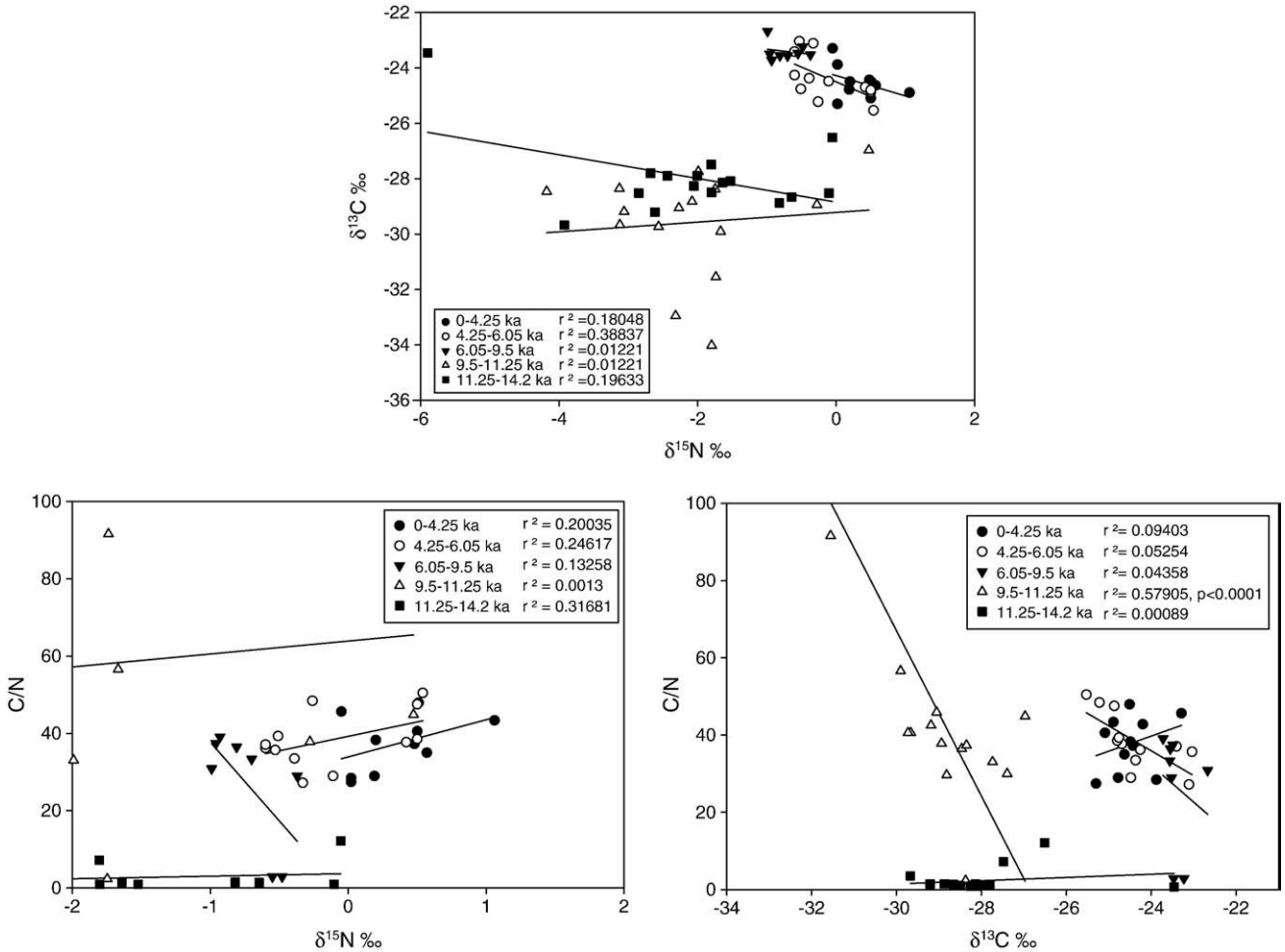

Fig. 2. Relationships among δ^{13}C, δ^{15}N, C/N for key climatic intervals in the Swanson Fen record. High correlation would indicate that the isotopic signature is largely that of decomposition.

influences these signatures (Herczeg, 1988; Kuhry and Vitt, 1996; Meyers, 1997; Sharma et al., 2005). Kuhry and Vitt (1996) suggest that C/N can be used as a diagenetic tool, since an increase in C/N in the acrotelm, the peat layer still receiving aeration due to fluctuating water table levels, indicates a preferential loss of nitrogen during aerobic decomposition. Anaerobic decomposition in the catotelm, the portion of the peat that receives little aeration and has invariable water table fluctuations, would result in a gradual decrease in C/N for bulk peat because of methanogenesis (Kuhry and Vitt, 1996) and because nitrogen becomes immobilized in the peat (Urban and Eisenreich, 1988). This trend of decreasing C/N with depth is not observed in Swanson Fen (Fig. 3), suggesting that diagenesis may not have played a major role. The Holocene values, excluding the values at the Pleistocene–Holocene transition, are typical for bryophytic peat (Kuhry and Vitt, 1996).

A high correlation between C/N and δ^{13}C may further indicate if diagenesis significantly altered the original record (Herczeg, 1988; Meyers, 1997; Sharma et al., 2005). Microbial degradation acts mostly on the nitrogen-bearing compounds in the peat and the preferential removal of $_{12}$C to ^{13}C, resulting in lower δ^{13}C in the peat (Herczeg, 1988; Meyers, 1997). Furthermore, a gradual increase in δ^{13}C occurs as a result of aerobic decomposition, while the opposite occurs in anaerobic conditions (Libes and Deuser, 1988; Lehmann et al., 2002; Sharma et al., 2005). The r^2 values for the relationship between C/N and δ^{13}C are small for four of the five stratigraphic units in this core (Fig. 2), suggesting that decomposition is largely not driving the stable isotopic values (0–4250 cal yr BP, $r^2=0.09403$; 4250–6050 cal yr BP, $r^2=0.05254$; 6050–9670 cal yr BP, $r^2=0.04358$; 11,250–14,125 cal yr BP). For the period 9670–11,250 cal yr BP the relationship has an $r^2=0.7443$ ($p<0.0001$), which may suggest that decomposition is driving the isotopic signature (Engel et al., 2010), but the low p-value suggested that this return is not significant. When removing the data points that contain the anomalously high C/N and depleted δ^{13}C values at the beginning of the Holocene, $r^2=0.12073$. Late-glacial and early Holocene r^2 values for the C/N versus δ^{13}C are not highly correlated, suggesting diagenesis did not significantly alter the record.

Decomposition of the plant material in the peat by microbial activity can alter the original δ^{15}N isotopic signal. In situ bacterial growth has been shown to increase δ^{15}N under oxic conditions, which corresponds to an increase in the C/N ratio, as a preferential loss of nitrogen over carbon occurs (Lehmann et al., 2002). Denitrification selectively removes ^{14}N, therefore resulting in a heavier residual N pool (Lehmann et al., 2002; Bernard-Haughn et al., 2003). In the Swanson Fen record, the δ^{15}N values are lighter rather than heavier, suggesting that the signal is not dominated by microbial degradation. A strong relationship between C/N and δ^{15}N indicates decomposition is high, but a relatively low correlation in Swanson Fen suggests that decomposition did not significantly influence the N isotopic values (0–4250 cal yr BP, $r^2=0.20035$; 4250–6050 cal yr BP, $r^2=0.24617$; 6050–9670 cal yr BP, $r^2=0.13258$; 9670–11,250 cal yr BP, $r^2=0.0013$;11,250–14,125 cal yr BP, $r^2=0.31681$) (Fig. 2). Plotting the δ^{15}N versus δ^{13}C shows data falling into two distinct clusters (Fig. 2). The samples in the lower portion of the core, the late-glacial

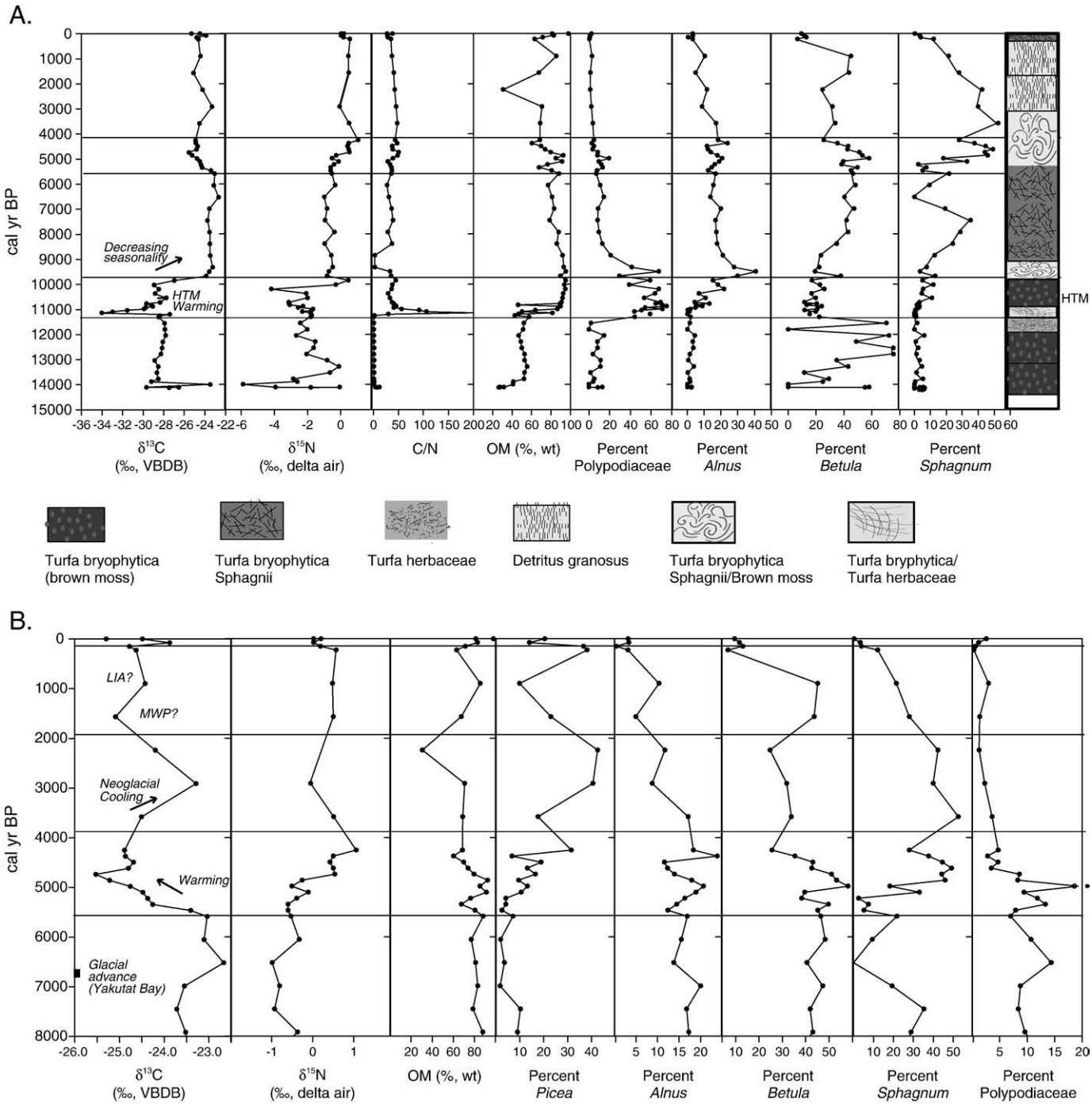

Fig. 3. A. Swanson Fen lithological and isotopic changes over the late-glacial and Holocene. Carbon and nitrogen isotopic values (expressed as ‰, VPDB), C/N ratio, percent organic matter (OM) by weight, and pollen and spore percentages from selected species in Swanson Fen. B. Swanson Fen δ^{13}C and δ^{15}N, C/N, selected pollen and spore percentages for the last 8000 years.

and early Holocene, have generally lower δ^{13}C and δ^{15}N than the rest of the core, showing an isotopic separation between the wetter, semi-aquatic and rich fen-like portions of the core and the mostly *Sphagnum*-dominated upper portion of the core.

4.2. Late-glacial δ^{13}C and δ^{15}N variation in Swanson Fen

The late-glacial peat is composed primarily of humified brown mosses and the late-glacial macrofossil record indicates that Swanson Fen was a shallow pond (Jones et al., 2009). The low and fluctuating δ^{13}C values below 244 cm (14,000 cal yr BP) suggests aquatic uptake, where carbon sources can be highly variable (Hecky and Hesslein, 1995). Abundant aquatic and semi-aquatic macrofossils are present in this portion of the core (Jones et al., 2009), which further supports the above interpretation. C/N ratios depend on source material, and the aquatic source material, such as aquatic algae, in the late-glacial may account for the low C/N, particularly when productivity of macro-autotrophs, and thus carbon content, was low. The low δ^{13}C and low C/N ratios in the late-glacial fall within the range of lacustrine algae (Meyers, 1994), which suggests that the fen in this period may have primarily been a mud-bottomed pool that also supported plants such as *Hippuris vulgaris* and *Carex* spp. (Jones et al., 2009). The relatively stable (−28‰ to −29‰) carbon isotopic values from 14,000 to 12,000 cal yr BP may suggest a constant CO_2 source, and the values fall

into the range of atmospheric uptake by C_3 plants. This interval corresponds with the interval of high sedge achenes in the macrofossil record (Jones et al., 2009), which may indicate that the isotopic signal was controlled by sedges (vascular plants) rather than byrophytes, as in the remainder of the record. The relatively low LOI is further evidence for the aquatic nature of the site.

Late-glacial δ^{15}N values in Swanson Fen are negative (−2‰ to −4‰, on average), which may suggest high nitrogen availability. The nitrogen isotopic signal can be more difficult to decipher because nitrogen can enter fens from numerous sources and in different organic and inorganic forms. When transported in surface water, it can take a dissolved or particulate form, existing in either organic or inorganic molecules (Robinson, 2001). The inorganic forms are dissolved as either NO_3^- or NH_4^+, and NO_3^- in swales can come from allochthonous sources, including groundwater, surface water, and precipitation (Sharma et al., 2005). Nitrogen also becomes available to plants through N_2-fixation, mineralization, and translocation (Malmer, 1988), and nitrogen is lost from the system through root efflux, denitrification, and N transfer through ecosystem pools by ingestion, excretion, and groundwater export (Robinson, 2001). When considering the δ^{15}N in hummock versus hollow *Sphagnum* species, hummock species tend to be more depleted than hollow species (Asada et al., 2005), since δ^{15}N in rainwater is more depleted than in groundwater (Kendall, 1998). However, the abundance of aquatics and brown mosses in this portion of the Swanson Fen record suggest that groundwater may have played a larger role than rainwater. One possible explanation for this discrepancy is that in the late-glacial interval, a relatively large amount of the sediment (40–80%, see LOI, Fig. 3) was fine clay derived from glacial outwash. These clay minerals have both large surface area and negative surface charges, which allow them to adsorb ammonia, resulting in depressed C/N ratios (Meyers, 1997). High nitrogen availability is in agreement with the elevated nitrogen concentrations in the late-glacial sediments, as enhanced nitrate availability allows for greater biotic discrimination in favor of $_{14}$N, resulting in lighter δ^{15}N values (Inglett et al., 2007). The low C/N ratios during the late-glacial likely also suggest that the peat is primarily derived from aquatic sources (Meyers, 1994; Herczeg et al., 2001). The abundance of brown mosses in the record (Fig. 4) suggests that the site was likely a seasonally flooded, mud-bottomed pool, where algae commonly grow.

Another explanation for the δ^{15}N signal in the late-glacial is that it indicates a terrestrial rather than aquatic source. More specifically, ericoid and ectomycorrhizal plants have been shown to be $_{15}$N-depleted (Emmerton et al., 2001; Nordbakken et al., 2003; Asada et al., 2005; Hobbie and Hobbie, 2006) because of the isotopic fractionation associated with the assimilation of organic nitrogen by mycorrhizal fungi and the subsequent transfer to plant roots (Emmerton et al., 2001; Hobbie and Hobbie, 2006). Ericoids are terrestrial, and several studies have also found mycorrhyizal associations in *Carex* species from groundwater-fed communities, such as fens and wet meadows (i.e. Sondergaard and Laegaard, 1977; Turner et al., 2000). The low-nutrient environment associated with a newly deglaciated landscape may have favored plants with mycorrhizae in order to promote plant establishment. Indeed, ericoid and mycorrhizal plants are common in subarctic and arctic tundra ecosystems (Michelson et al., 1998; Emmerton et al., 2001), where productivity is generally low and nutrients are limited, with up to 80% nitrogen coming from microbial symbionts (van der Heijden et al., 2008). Thus, the light late-glacial isotopic values may be the result of the colonization and expansion of plants with mycorrhizal symbioses, such as *Carex* spp., which are abundant in the late-glacial sediments in Swanson Fen (Jones et al., 2009).

4.3. Isotopic variation at the Pleistocene–Holocene transition

A sharp δ^{13}C decrease at the beginning of the Holocene (~11,500 cal yr BP) ca. −34‰ corresponds with an increase in the C/N ratio over 200. The high C/N ratio, high LOI, and low δ^{13}C at this interval indicate high biomass accumulation and nutrient utilization, related to warm early Holocene temperatures (Kaufman et al., 2004). The sudden influx in organic matter may have enhanced decomposition with higher evaporation as a result of warm early Holocene temperatures, ultimately resulting in more negative δ^{13}C values.

Calibration experiments from peat studies suggest that temperature changes play a major role in carbon isotopic fractionation (Skrzypek and Jedrysek, 2005), with a 1 °C temperature increase resulting a 0.6‰ decrease in δ^{13}C in *Sphagnum* peat and a 1.5‰ decrease in δ^{13}C in living mosses (Skrzypek et al., 2007b). While the early Holocene peat is dominated by brown mosses, a similar temperature fractionation could provide a plausible explanation for the change in fractionation at the end of the HTM. The early Holocene isotopic values coincide with the maximum of summer insolation north of 60° (11,000–9000 cal yr BP; Berger and Loutre, 1991), which increased seasonality and resulted in 2–4 °C warmer summer temperatures in Alaska (Kaufman et al., 2004). However, temperatures increased by more than 1 °C in the early Holocene, so the thermodynamic fractionation is probably offset by other factors. Fractionation in δ^{13}C of peat has also been shown to occur with changes in atmospheric CO_2 partial pressure (0.9‰ to 1.1‰ for *Sphagnum* per 3 ppmv pCO_2 Menot and Burns, 2001), but Holocene CO_2 values have not varied more than 30 ppmv and the associated shift expected in organic matter δ^{13}C is <0.3‰ (Indermühle et al., 1999), so these changes cannot explain the observed variation.

A likely explanation for the decrease in early Holocene δ^{13}C values is that methanogenesis was high. The δ^{13}C values from methane ebullition are extremely light (−58‰ to −80‰) in arctic lakes today (Walter et al., 2006), and increased atmospheric methane during the Pleistocene–Holocene transition was partially caused by thermokarst formation stemming from rapid warming (Walter et al., 2007). While it is unclear whether permafrost existed in this location, methanogenesis, a process driven by increased temperatures, was likely high. This scenario is reasonable, as early Holocene temperatures were higher than today (Kaufman et al., 2004). Evidence for increased methane emission in Swanson Fen is a bryophytic assemblage, including *W. exannulata* and *D. capillifolius* mosses, characteristic of high water table and correlated with high CH_4 production environments (Fig. 4; Bubier, 1995).

In order for the lighter δ^{13}C values to be recorded in the plants, a pathway must exist for this light carbon to be assimilated during photosynthesis. Wet swale environments can cause bryophytes to use recycled carbon, which leads to lighter carbon isotopic values (Price et al., 1997; Raghoebarsing et al., 2005). While numerous studies have interpreted lighter δ^{13}C in mosses as indicating drier conditions, as explained by a thickness of water film on the *Sphagnum* leaf (Williams and Flannagan, 1996; Rice and Giles, 1996; Loisel et al., 2009), this study is consistent with the interpretation of Nichols et al. (2009), which showed lighter values in wetter environments where recycling of methane is high. A recent surface moisture and testate amoebae-δ^{13}C calibration study in *Sphagnum* peatlands from Alaska found no strong correlation between surface moisture and δ^{13}C, and suggested recycled carbon use from methanogenesis as a potential reason, as the lowest δ^{13}C value was recorded in a partially thawed permafrost site (Markel et al., 2010). The particularly low values of δ^{13}C during the early Holocene in this study would suggest that at least a fraction of the carbon is from recycled carbon uptake from methane.

Early Holocene δ^{15}N values do not differ significantly from late-glacial values, despite a sizeable negative excursion of the δ^{13}C signal and a large increase in the C/N ratio. The >200 C/N ratio (corresponding with a decline in overall nitrogen content, Fig. 3) at the beginning of the Holocene indicates high nitrogen consumption relative to carbon. Bodelier and Laanbroek (2004) show that methanotrophic bacteria have high N requirements, and high methanogenesis and N consumption would result in lighter δ^{15}N, as the methanotrophs would consume the lighter isotope.

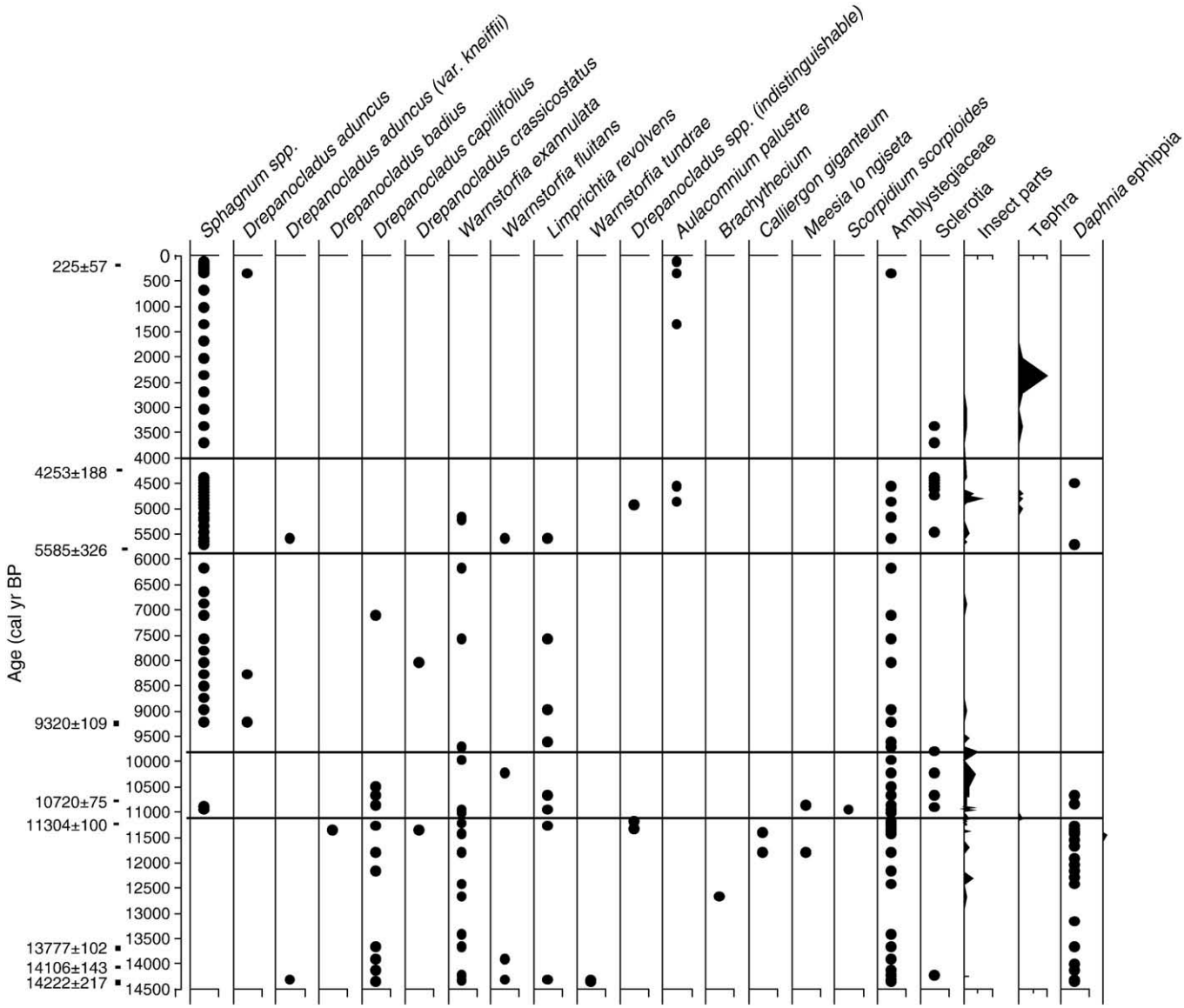

Fig. 4. Bryophyte macrofossils for Swanson Fen recorded as presence versus absence.

The early Holocene vegetation on the Kenai Peninsula is dominated by a large increase in fern spores, a pattern observed throughout the region, such as from rapidly retreating mountain glaciers (Hu et al., 1995; Ager, 2000; Jones et al., 2009). While it is unclear if the high ferns represent moist growing season conditions (Hu et al., 1995; Jones et al., 2009) or high disturbance (Ager, 2000), well-preserved peat and high peat accumulation during the early Holocene (Jones, 2009; Jones and Yu, 2010), suggests that moisture was not limiting in these peat-forming processes. One explanation for the high peat accumulation in this groundwater-fed fen, as well as for the observed vegetation patterns, is that Kenai Mountain glacial ablation was high during the early Holocene. Evidence for glacial retreat in the Kenai Mountain beyond its modern limit (Barclay et al., 2009) during the early Holocene implies high rates of early Holocene glacial ablation. If the Kenai lowlands were fed either directly or indirectly through groundwater input by water ultimately stemming from melting glaciers, the lowland fens would have been inundated. From a recent hydrologic model of the peatland hydrogeologic setting on the Kenai Peninsula (Reeve and Gracz, 2008), it is plausible that water originating from the Kenai Mountains would replenish the lowland fens. The combination of warm and wet conditions likely contributed to high methanogenesis, and carbon released through this process and the uptake of recycled carbon (Price et al., 1997; Raghoebarsing et al., 2005) would have resulted in highly negative δ^{13}C values. The low δ^{15}N values are in agreement with this hypothesis, as the negative and variable δ^{15}N suggests that the primary source of N were combined sources from the soil, rather than atmospheric sources.

4.4. Interpretation of isotopic shifts at end of early Holocene

A ~5‰ shift to less negative δ^{13}C values concurrent with a ~4‰ shift to heavier δ^{15}N takes place between 10,000 cal yr BP and 9600 cal yr BP. These shifts correspond with a large decrease in fern spore abundance, an increase in *Alnus* pollen, and a slight increase in *Betula* pollen (Fig. 3). A lithological change from brown-moss to *Sphagnum*-dominated bryophytes is concurrent with increasing *Sphagnum* spores (Fig. 3). The similarity in timing of these changes may indicate that a significant change in regional climate occurred at this time. An oxygen isotopic record from Hundred Mile Lake in the Matanuska Valley shows a similarly large oxygen isotopic shift of 2 to 4‰ to lighter values, albeit slightly earlier (10,200 cal yr BP) (Yu et al., 2008), but the difference in timing may be within the dating uncertainty, suggesting a regional shift in climatic conditions occurred

at roughly the same time or slightly earlier to the north. The authors of this oxygen isotope record suggest that a change in regional atmospheric circulation patterns could have caused the shift, either resulting in an increase in winter precipitation compared to summer precipitation or an increase in moisture from the west (Yu et al., 2008). These results suggest a large-scale change in the moisture regime. A record from interior Alaska suggests that the climate at this time was warm and dry, a pattern that can occur from a weakened subtropical high (Edwards et al., 2001). Edwards et al. (2001) suggest an increase in moisture occurred ~9500 cal yr BP, which is in agreement with increasing lake levels from interior Alaska (Bigelow, 1997; Abbott et al., 2000), although these sites may have experienced a different climate pattern than the Kenai Peninsula. The lithology for Swanson Fen (Figs. 3 and 4) shows a change from brown moss-dominated to a peat matrix mostly composed of *Sphagnum* at this time. Jones et al. (2009) interpreted this change, along with the associated change in pollen assemblages, to drier conditions. However, assuming part of the carbon isotopic shift to heavier values is associated with an increase in peat moisture (Rice and Giles, 1996; Loisel et al., 2009, 2010), the region may have experienced a change in seasonal precipitation input.

The shift to heavier isotopic values at the end of the early Holocene may largely be explained by a lithological change, from brown moss dominated to *Sphagnum* dominated peat. The abundance of brown mosses during the early Holocene suggests that the fen was mostly fed by groundwater, sustaining a higher water table. A slowdown or cessation of glacial ablation at the end of the early Holocene may explain the isotopic shift, as well as the shift in vegetation. The subsequent shift in both δ^{15}N and δ^{13}C to more positive values suggests that climate cooled, decreasing methanogenesis. It also suggests that the moisture regime changed, so that mosses became more isolated from groundwater sources and precipitation became the dominant source of moisture. *Sphagnum* has the ability to store large amounts of water (Titus and Wagner, 1984) even in drier climates and is generally more associated with hummocks than brown mosses, which rely on nutrients from the groundwater to survive (Janssens, 1989). A decrease in brown moss abundance and increase in *Sphagnum* dominance suggests that as the surface of the fen became isolated from the groundwater, less recycled carbon uptake took place.

The ~4‰ shift in δ^{15}N results in a nitrogen isotopic signature that resembles the atmospheric signal, suggesting that N_2 from the atmosphere became the primary nitrogen source rather than combined soil pools (Robinson, 2001). A recent study showed that a long-term depletion of the N source eventually leads to reduction and cessation of methane consumption (Bodelier and Laanbroek, 2004), as may have been the case ~9600 cal yr BP at temperatures began to decline. The lower water table, as possibly suggested by the shift from a brown moss to a *Sphagnum*-dominated fen environment relies on the assumption that brown moss (*Drepanocladus* spp.) typically occur at the water table surface or submerged a few centimeters, whereas *Sphagnum* mosses are more commonly found at or several centimeters above the water table. Because nutrient conditions are closely linked to groundwater influx to the peatland, *Sphagnum*, which is commonly found in poor fens and bogs, also (Figs. 3 and 4) indicates an evolution to nutrient-poor conditions. Although significant differences exist among *Sphagnum* species, they are more typically associated with low nutrient environments. They contribute to the fen acidity through cation exchange capacity (Clymo, 1963) and to nutrient-poor conditions through the formation of decay-resistant complexes that result in N immobilization (Aerts et al., 2006), both of which inhibit vascular plant growth. Kendall (1998) suggests that a small pool of nitrogen is quickly converted from reactant to product with little or no fractionation, as is observed in this record. In ombrogenous bogs, *Sphagnum* mosses take up airborne nitrogen through N_2-fixation by cyanobacterial symbionts (Basilier, 1980), which are the primary nutrient input in these rain-fed systems (Aerts et al., 1992) and they have the ability to use organic nitrogen (Kielland, 1997), resulting in δ^{15}N of 0‰.

4.5. Mid to Late Holocene shifts

Isotopic shifts from 8000 cal yr BP to present are likely more related to changes in environmental and climatic conditions rather than large-scale lithological changes. Slightly more enriched δ^{13}C values occur between 7000 and 6000 cal yr BP, which corresponds with the earliest known Holocene glacial expansion at 6850 cal yr BP in Yakutat Bay, although this surging glacier may not reflect climatic changes (Calkin et al., 2001). However, decreasing Holocene temperatures, associated with decreasing insolation seasonality (Kaufman et al., 2004), could have slowed the rate of glacial ablation around this time, resulting in less moisture input to the peatland. Alternatively, the shift could reflect a change in atmospheric circulation patterns, related to a shift or weakening of the Aleutian Low or a movement in the subtropical high (Trenberth and Hurrell, 1994; Overland et al., 1999).

A shift to more depleted values occurs from 5500 to 3500 cal yr BP when mountain glaciers were under a period of retreat. Higher peat accumulation during this interval suggests warmer conditions and greater dominance of brown mosses suggests greater groundwater input (Figs. 3 and 4). The direction of the shift is the same as the warm early Holocene interval when the climate was warmer and groundwater input was also likely higher. Although the exact mechanism of this shift is not well understood, the consistency in the direction of the shift suggests that the same mechanism was in control of the shift. These changes also correspond to increases in fern spores (Polypodiaceae), which have been interpreted as disturbance related to glacial retreat and/or peatland expansion (Jones et al., 2009). The increase in *Sphagnum* spores at that time suggests peatland expansion occurred under a relatively warm and potentially moist climate, and a change to more brown mosses in the lithology (Figs. 3 and 4) further support an increase in water table depth. The interpretation that the thickness of the water film on the *Sphagnum* leaf controls δ^{13}C signature would suggest that this climatic interval is drier, but it seems to be the opposite of our interpretation. Using several independent lines of evidence, Nichols et al. (2009) suggest that wetter conditions lead to lower δ^{13}C values, citing greater methane recycling as the reason for the lower values. This idea also seems to be consistent with our early Holocene interpretation, as much lighter isotopic values during the early Holocene probably cannot be explained by a change in lithology alone. The δ^{15}N becomes more enriched at this time, which could reflect a change in the source of nitrogen. It also is more consistent with the interpretation that the groundwater input becomes more important, as other studies report that rainwater has a more depleted δ^{15}N signature than groundwater (Kendall, 1998; Asada et al., 2005). Neoglaciation on the Kenai began between 3600 and 3000 cal yr BP (Wiles et al., 1999), which corresponds with ~2‰ shift to more enriched δ^{13}C values. The low peat accumulation rate after the neoglaciation prevents us from making any robust interpretation about the Medieval Warm Period or Little Ice Age. The decomposed lithology and low peat accumulation may even suggest that a hiatus exists in this portion of the record, perhaps related to cooler, drier growing season conditions as glaciers advanced during the LIA (Barclay et al., 2009).

5. Paleoclimatic implications and conclusions

The isotopic ratios of C and N shift significantly at the beginning of the Holocene, indicative of the regional change in climate which affected vegetation and lithology. Warm early Holocene temperatures resulted in high glacial ablation, ultimately maintaining high water table levels in Kenai lowland fens. The combination of warm, wet

conditions in these fens would have led to high rates of methanogenesis, which is supported by the very light δ^{13}C values. Likewise, the negative and variable nitrogen isotopic signal suggests plant uptake of combined nitrogen soil pools from the nutrient-rich environment rather than atmospheric nitrogen.

The extremely low δ^{13}C values in the early Holocene in Swanson Fen correspond with a time when atmospheric methane concentrations, as recorded by the Greenland ice cores, were high (Brook et al., 2000). Numerous studies (Chappellaz et al., 1997; Petrenko et al., 2009) have suggested that boreal wetlands are a likely early Holocene source of atmospheric methane due to rapid peatland development (Smith et al., 2004; MacDonald et al., 2006; Jones and Yu, 2010). Furthermore, δ^{13}C measurements in methane within the Greenland ice core during the early Holocene reveal lighter isotopic values during warmer intervals such as the early Holocene, compared to colder intervals such as the Younger Dryas (Fischer et al., 2008), suggesting higher methanogenesis during warmer intervals, but also that the source of methane is largely from boreal wetlands. This study may provide evidence for high methanogenesis during the warm early Holocene, but more studies with refined sampling techniques are needed to assess the validity of this hypothesis and to determine how widespread these peatland processes were at that time.

As temperatures cooled at the close of the early Holocene, glacial ablation slowed, and the fens became dominated by atmospheric moisture, a shift to heavier δ^{13}C values occurred and nitrogen isotopic values became centered around 0‰, the atmospheric N isotopic value, indicating a change in predominant nitrogen source from the soil to the atmosphere. Evidence for further climatic cooling, as documented by glacial advance around 6850 cal yr BP, results in an enrichment of δ^{13}C. A subsequent warmer, wetter climatic interval around 4800 cal yr BP, results in a δ^{13}C shift in the same direction as the early Holocene shift and also contains a smaller fern spike as in the early Holocene, which may suggest that warmer, wetter conditions result in lighter δ^{13}C values. These results suggest that these Kenai lowland peatlands respond to changes in glacier advance and retreat. Thus, under increased warming observed today, these peatlands may increase accumulation as long as there is a glacial meltwater supply of water to these peatlands, as glacial ablation is highest during the warmest part of the summer when peatlands are under high evapotranspirative stress. These observations, particularly when placed into the context of the global carbon cycle, merit further attention. Thus this study indicates how climate mediated a significant response in both carbon and nitrogen isotopic ratios at one site, and further investigation will enhance the full potential of these proxies.

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