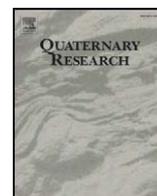




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Quantitative assessment of precipitation seasonality and summer surface wetness using ombrotrophic sediments from an Arctic Norwegian peatland

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ABSTRACT

Seasonality of precipitation is an important yet elusive climate parameter in paleoclimatological reconstructions. This parameter can be inferred qualitatively from pollen and other paleoecological methods, but is difficult to assess quantitatively. Here, we have assessed seasonality of precipitation and summer surface wetness using compound specific hydrogen and carbon isotope ratios of vascular plant leaf waxes and *Sphagnum* biomarkers extracted from the sediments of an ombrotrophic peatland, Bøstad Bog, Nordland, Norway. Our reconstructed precipitation seasonality and surface wetness are consistent with regional vegetation reconstructions. During the early Holocene, 11.5–7.5 ka, Fennoscandia experienced a cool, moist climate. The middle Holocene, 7.5–5.5 ka, was warm and dry, transitioning towards cooler and wetter conditions from the mid-Holocene to the present. Changes in seasonality of precipitation during the Holocene show significant coherence with changes in sea surface temperature in the Norwegian Sea, with higher SST corresponding to greater percentage of winter precipitation. Both high SST in the Norwegian Sea and increased moisture delivery to northern Europe during winter are correlated with a strong gradient between the subpolar low and subtropical high over the North Atlantic (positive North Atlantic Oscillation).

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Introduction

Seasonality of precipitation has a profound impact on agriculture, natural vegetation, and regional moisture balance. It is an integral part of the climate system that can be assessed using climate models, but paleoclimate datasets to test model output of precipitation seasonality are rare. Changes in the amount of summer versus winter precipitation also have tremendous impacts on vegetation and lake levels (Shuman and Donnelly, 2006). Therefore, knowledge of precipitation seasonality is also crucial for interpreting past records of plant ecology and lake level. Many methods are available to assess total precipitation or overall moisture balance in the past. For example, records of lake level, diatom assemblages and mineralogy from lake sediments (Fritz, 1996; Shuman and Donnelly, 2006), and testate amoebae assemblages (Booth, 2002) and *n*-alkane distributions from peat bogs (Nichols et al., 2006) can provide total water balance information. However, existing proxies from these paleoclimate archives do not provide a sufficient level of quantification for precipitation seasonality on the continents.

Because it is isolated from groundwater, the water level in an ombrotrophic peatland is extremely sensitive to changes in moisture balance (i.e., precipitation–evaporation) (e.g.: Barber et al., 2000). The

response to this change in water level is manifested at the surface by a change, for example, in the vegetation assemblage (Charman, 2002). Both testate amoebae and vegetation assemblages in ombrotrophic peatlands have been used successfully to assess the past hydrological variations (Blackford, 2000; Charman 2002; Mauquoy and Barber, 2002). Recently, Nichols et al. (2006) demonstrated that the relative distribution of vascular and *Sphagnum* biomarkers in peatland sediments can be used to assess paleohydrology. The biomarker approach is characterized by its exceptionally high efficiency and sample throughput; up to 200 samples can be processed in a week. Such detail is often more difficult for conventional methods based on time-consuming microscopic observation. Despite major recent progresses, however, none of these approaches can provide specific information on precipitation seasonality using samples from ombrotrophic peatlands.

We report a new approach based on compound specific stable isotope ratios of lipid biomarkers to assess the seasonality of precipitation from a peat bog (Bøstad Bog) in Arctic Fennoscandia. Although isotope ratios of biomarker compounds have been used extensively in lakes (e.g., Seppä and Hammarlund, 2000; Shuman et al., 2004; Hou et al., 2006, 2007) and, preliminarily, in peatlands (Xie et al., 2000) to infer temperature changes, so far, quantitative paleohydrological reconstructions using compound specific isotopic ratios of biomarkers in peatlands have not been reported. We show, from our analyses of Bøstad Bog sediments, that stable isotope ratios of *Sphagnum* lipids can be used to quantify bog surface moisture and

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vascular plant waxes to reconstruct precipitation seasonality, taking advantage of the large seasonal differences in precipitation H isotopic ratios in high latitude regions. We will also synthesize our results with regional climate records to better understand the underlying driving mechanism of climatic change for the past 11.5 ka in the study region.

Study area

Bøstad Bog ($68^{\circ}14.8'N$, $13^{\circ}44.5'E$), an ombrotrophic peatland, is located on the island of Vestvågøy, which is part of the Lofoten Islands, off the coast of mainland Norway (Fig. 1). Peat has been accumulating at this site since the early Holocene, ~ 11 ka (Vorren, 1979). In the modern system, the hydroclimate in the region of Bøstad Bog is closely controlled by conditions in the North Atlantic. Variations in moisture delivery to the Bøstad Bog region are controlled, in part, by the North Atlantic Oscillation (NAO). The NAO describes the gradient in atmospheric pressure between the subtropical high and subpolar low over the North Atlantic Ocean during winter. A strong gradient (high index) is correlated with increased moisture delivery to northern Europe, while a weak gradient is associated with a moisture deficit (Hurrell and Dickson, 2004). A positive-index condition is also related to increased transport of warm water to the Norwegian Sea by the Gulf Stream (Flatau et al., 2003).

Figure 2 shows the monthly mean precipitation amount for Bøstad Bog as well as the monthly mean hydrogen isotope values for precipitation as calculated by the Online Isotopes in Precipitation Calculator (Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005). Isotope values of precipitation in this region are most strongly influenced by the temperature effect (Clark and Fritz, 1997). The modern monthly temperature and precipitation data used in this study is from the Global Historical Climatology Network (GHCN-monthly, version 2).

Methods

Field methods

A 122 cm peat monolith was recovered from Bøstad Bog in the Lofoten Islands, Norway. The monolith was described in the field and

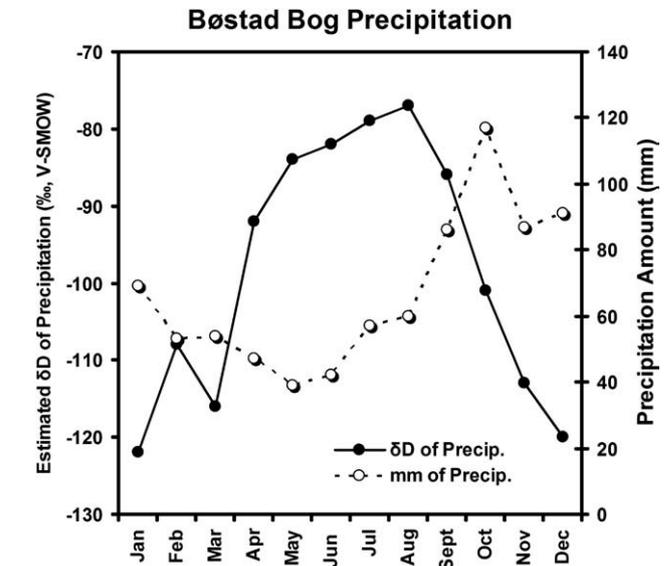


Figure 2. Monthly mean values for precipitation amount and δD of precipitation at Bøstad Bog. δD values are calculated from the Online Isotopes in Precipitation Calculator (Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005).

then sampled in 1 cm slices. Slices for geochemical analysis were taken every 8 cm for the first meter and every 4 cm thereafter.

Leaves of vascular plants and whole *Sphagnum* plants were collected from various ombrotrophic bog locations throughout the upper Midwestern U.S. and New York State. *Sphagnum* species collected included *S. angustifolia*, *S. capillifolium*, *S. fallax*, *S. girgensohnii*, *S. norvegicum*, *S. russowii*, and *S. viridum*. Sedge species included *Carex limosa*, *C. livida*, *C. oligosperma*, and one unidentified species of *Carex*. Ericads included *Kalmia angustifolia*, *Kalmia polifolia*, and *Rhododendron groenlandicum*. The distribution of *n*-alkanes in these plant samples was reported in Nichols et al. (2006). Ombrotrophic bogs are characterized by low species richness (Glaser et al., 1997), and the vascular species studied here account for the vast majority of the non-*Sphagnum* biomass input to peats (Davis and

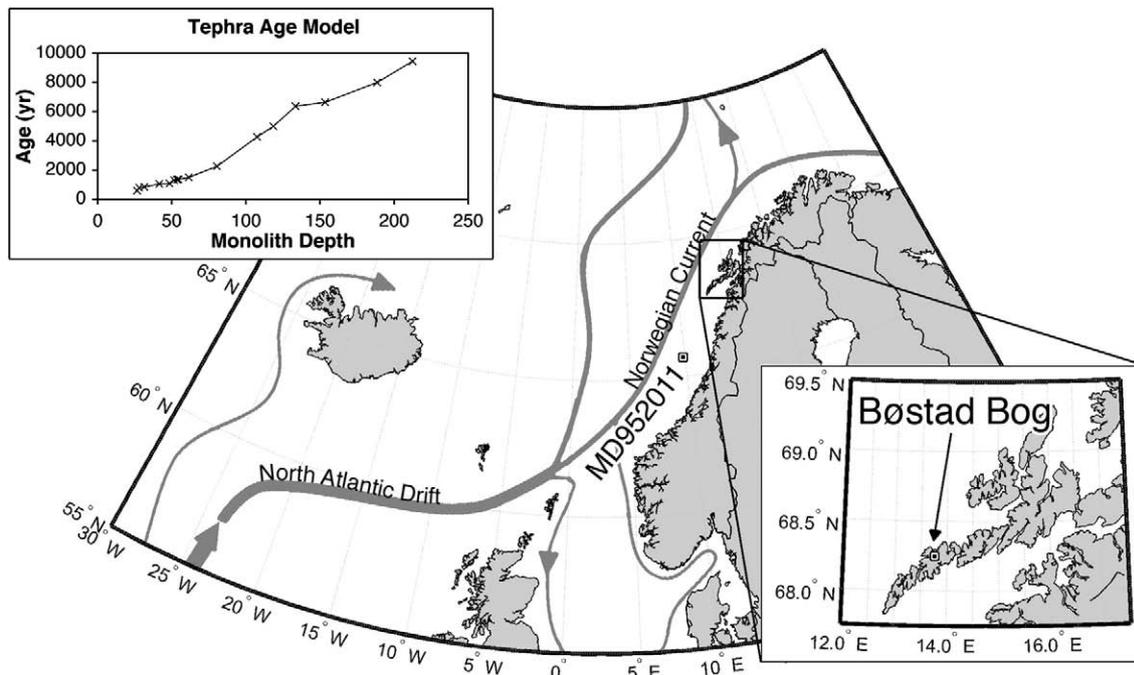


Figure 1. Site map indicating the locations of Bøstad Bog and core MD952011. Inset: The tephra age model for this core, established by Pilcher et al. (2005).

Anderson, 2001). Peat bog vegetation is also quite similar between North America and Europe (Charman, 2002).

Laboratory methods

Extractable lipids were dissolved from freeze-dried peat samples using the Dionex Accelerated Solvent Extractor (ASE) with a 2:1 mixture of dichloromethane and methanol. Plant samples were extracted by ultrasonic agitation in a 9:1 solution of dichloromethane and methanol. The total lipid extracts of both plants and peats were separated into acid and neutral fractions by solid phase extraction (Aminopropyl Bond Elute®). The carboxylic acid fraction was methylated using anhydrous 2% hydrochloric acid in methanol. Hydroxy-acids were removed using silica gel column chromatography (DCM as solvent) in order to further purify the resulting fatty acid methyl esters and avoid chromatographic coelution (Hou et al., 2006). The neutral fraction was separated into aliphatic compounds; ketone, ester, and aromatic compounds; alcohols; and methanol-soluble compounds on a silica gel column. The hydrocarbon fraction was oxidized with Ruthenium tetroxide to remove unsaturated compounds (Huang et al., 2007).

Quantification and identification of compounds were carried out using GC and GC–MS, respectively. An HP 6890 GC interfaced with a GC Combustion III module to a Finnigan Delta^{Plus}XL stable isotope mass spectrometer through a high-temperature pyrolysis reactor was used for hydrogen and carbon isotopic analysis (Huang et al., 2002, 2004). The accuracy was routinely checked by an injection of laboratory isotopic standards between every six sample injections. $\delta^{13}\text{C}$ values obtained from individual fatty acids (as methyl esters) were corrected by mathematically removing the isotopic contributions from added groups before reporting. The $\delta^{13}\text{C}$ value of the added methyl group was determined by acidifying and then methylating (along with the samples) the disodium salt of succinic acid with a predetermined $\delta^{13}\text{C}$ value (using TC/EA-IRMS) (Huang et al., 2002).

The particularly robust chronology for this core was previously established using 16 well-dated tephra layers (Pilcher et al., 2005). The humification index was determined by transmittance spectroscopy of basic extracts of peats (e.g., Caseldine et al., 2000).

Results and discussion

Development of quantitative proxies for precipitation seasonality and surface wetness

n-Alkanes and *n*-acids as biomarkers for *Sphagnum* and vascular plants

Sphagnum and vascular plants show distinct differences in the distributions of *n*-alkanes (Baas et al., 2000; Nott et al., 2000; Pancost et al., 2002), with the former containing greater amounts of C_{23} *n*-alkanes and the latter more C_{29} *n*-alkanes (Fig. 3A). The mass balance corrected $\text{C}_{23}:\text{C}_{29}$ *n*-alkane *Sphagnum* to Vascular plant Ratio (henceforth referred to as SVR) can be used to track changes in contribution of the two different plant types to the peat (Nichols et al., 2006). During dry periods, *Sphagnum*, with no vascular system, cannot take advantage of the available water at the lowered water table, and its growth is impeded. Consequently, vascular plants contribute a greater proportion of leaf waxes to the peat deposited during drier times. When the water table is high, the oxygenated zone near the surface of the peatland becomes smaller, making the peatland less hospitable to vascular plants whose root systems require adequate oxygen. During these wetter intervals, the *Sphagnum* flourishes, contributing in greater proportion to the peat. Therefore, changes in the amount of C_{23} and C_{29} *n*-alkane, the most abundant alkanes in *Sphagnum* and vascular plants respectively, reflect changes in moisture balance. Higher values indicate more contribution by *Sphagnum* and thus, wetter conditions during the time of deposition.

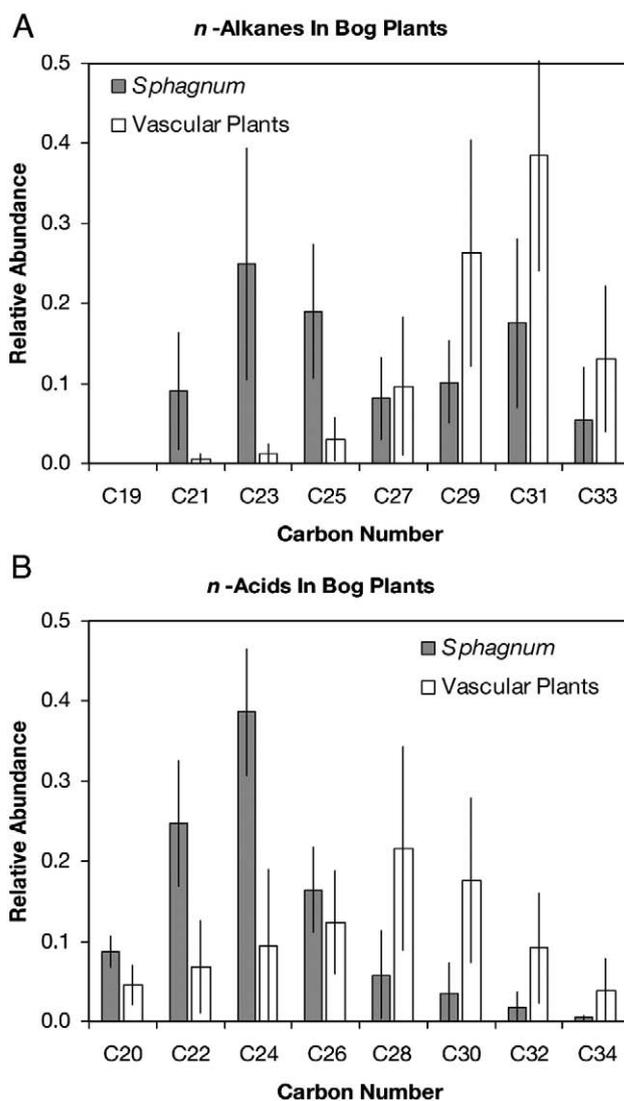


Figure 3. Average distribution of *n*-alkanes (A) and *n*-acids (B) in peat-forming plants. The data are from all samples of *Sphagnum* and vascular plants we collected and analyzed (Some *n*-alkane data were reported in Nichols et al., 2006). A total of 7 *Sphagnum* species and 7 vascular plant species are used to obtain the average distributions (species names are listed in the “Methods” section).

Because *n*-fatty acids are the chemical precursors to *n*-alkanes (e.g., C_{23} *n*-alkane is derived from decarboxylation of C_{24} *n*-acid), we investigated the distributions of fatty acids from species (listed in “Methods”) of *Sphagnum* and vascular plants found in peatlands. Figure 3B shows the relative abundance of *n*-acids from a variety of species of *Sphagnum* and vascular plants collected from ombrotrophic peatlands throughout the Midwestern U.S. Among the *Sphagnum* species, the distribution of fatty acids has a mode at C_{24} , the precursor of C_{23} *n*-alkane, as expected. This indicates that C_{24} *n*-acid is also an effective biomarker for *Sphagnum* as suggested by Pancost et al. (2002).

Hydrogen isotope ratios of C_{29} *n*-alkanes and precipitation

Generally, the δD values of vascular plant leaf waxes (e.g., C_{29} *n*-alkane) have been shown to track the δD of the source water used by vascular plants during their growth period (Sachse et al., 2006; Shuman and Donnelly, 2006; Hou et al., 2008). Several factors, such as leaf-water evaporation and species distribution, could also affect δD values of vascular plant leaf waxes. However, ombrotrophic peatlands are unique in a number of ways that minimize the effect of leaf-water

evaporation and species effect. Leaf-water evaporation is mainly controlled by the leaf-to-air vapor pressure difference (Ripullone et al., 2008). This difference is dependent on relative humidity and stomatal conductance. Growth experiments show that the impact of relative humidity on plant leaf waxes is in fact quite small (Hou et al., 2008). Further, the effect of changing relative humidity on the δD of leaf waxes in peatland plants is particularly minimized, as ombrotrophic peatlands typically only grow in regions with 75 to 85% relative humidity (Halsey et al., 2000). Because the major environmental effect on stomatal conductance is moisture stress, it is also unlikely that this parameter would vary substantially in an ombrotrophic peatland, where drought stress on vascular plants is all but nonexistent. Though the regional vegetation changes substantially throughout the time period covered by our sediments, the site itself remained a peatland for the length of the interpreted record. Peatlands are known to have a remarkably low diversity, with most of the community change represented by shifts between vascular plants and *Sphagnum* (Davis and Anderson, 2001). Though the pollen in the sediments of this peatland comes from regional vegetation, the *n*-alkanes in the sediment are mainly from in-situ production by peatland plants. For these reasons, we interpret the δD of vascular plant leaf waxes in our ombrotrophic peatland to essentially reflect the δD of stored peatland water.

It is also important to note that the stored peatland water represents mean annual precipitation. Vascular plants' root systems afford them to access the water stored in the peatland that is protected from evaporation by the *Sphagnum*-covered surface (Kim and Verma, 1996). This water, which has been stored in the peatland over the seasons, is representative of the annual average precipitation (Clark and Fritz, 1997; Charman, 2002). Therefore, a distinction must be made between water sources of vascular plants growing in peatlands and those growing on normal soils in the northeastern United States. Non-peatland vascular plants in the northeastern United States use a combination of groundwater (often representing annual mean precipitation) and summer water, depending on topographic and climatic conditions as well as plant physiology (White et al., 1985; Hou et al., 2008). The fact that vascular plants growing on ombrotrophic peatlands use annual mean precipitation as source water greatly simplifies our interpretation of leaf wax hydrogen isotopic records. The ~20% offset between summer and annual precipitation is reflected in the downcore δD values of C_{29} *n*-alkane of Bøstad samples. During the early Holocene (11–9.2 ka), the mean δD of C_{29} *n*-alkane is -165‰ , while during the drier period from 9ka to 6ka, the mean δD of C_{29} alkane is -185‰ . Presently, the June, July, August (JJA) average δD of rainfall over Bøstad Bog is -78.3‰ , while the annual average is -98.3‰ (Fig. 2). Thus, we interpret the large variation of δD of annual average precipitation to be driven by changing mass balance between warm and cool season precipitation. This interpretation will be further discussed in the "Isotope modeling" section.

Carbon isotope ratios of C_{24} *n*-acid as a proxy for surface methanotrophy

We interpret the changes in $\delta^{13}C$ values of C_{24} *n*-acid (produced mainly by *Sphagnum*) as influenced primarily by assimilation of methane-derived CO_2 . During dry times, the $\delta^{13}C$ values of C_{24} *n*-acid increase, and during wetter intervals, values decrease. This interpretation is based on the following justifications. Changes in the $\delta^{13}C$ values of C_{24} *n*-acid in the Bøstad core are concurrent with changes in the SVR, our established proxy for paleohydrology (Figs. 4A and B), with lower $\delta^{13}C$ values corresponding to wetter times. We interpret this relationship to derive from the utilization of methane-derived CO_2 by *Sphagnum* under wet conditions. Methane is produced in peatlands in the anoxic zone below the water table, and is oxidized in the aerated zone above the water table (Roulet et al., 2007). A high water table in the peatland means a small zone of oxidation. Therefore, during wet conditions, more methane produced at depth

reaches the peatland surface, whereas during dry times, it can be oxidized before it reaches the surface. Recently, a methanotrophic bacterium was found to live symbiotically with *Sphagnum* in its hyaline (water-holding) cells (Raghoebarsing et al., 2005). As more methane is released from the peatland, the methanotrophic symbiont provides more methane-derived (^{13}C -depleted) CO_2 to the *Sphagnum*. Therefore, higher methane fluxes under wet conditions lead to more assimilation of CH_4 -derived CO_2 , and thus lower carbon isotopic ratios of *Sphagnum* lipids. Our interpretation of carbon isotopic changes of *Sphagnum* biomarkers coincide in direction with those applied to vascular plants, but for very different reasons. Vascular plants become less selective of $^{12}CO_2$ (hence higher $\delta^{13}C$ values) as they keep their stomata more closed to prevent water loss during dry times (Clark and Fritz, 1997). *Sphagnum* does not have stomata, so this interpretation is not valid; therefore, we deduce that it is primarily the degree of methane utilization that determines the *Sphagnum* $\delta^{13}C$ values.

Our interpretation of the carbon isotope ratios of *Sphagnum* biomarkers appears to be the opposite of the published effect of water film thickness (Williams and Flannagan, 1996). Indeed, a number of papers have suggested that bulk *Sphagnum* C isotopic ratios increase when *Sphagnum* is in a wetter environment (e.g., Loisel et al., 2009), when we would have predicted a lower $\delta^{13}C$ values due to increased recycling of methane under wetter conditions. However, a thorough literature survey has failed to yield paleomoisture reconstructions actually using the water film interpretation; all studies have been based on modern *Sphagnum* plants from wet and dry spots in individual peatlands. Studies describing the water film interpretation are based on the comparison of samples from hummock and hollow microenvironments. While binning carbon isotope data into these microtopographical designations and comparing them supports this water film hypothesis, comparison of carbon isotope measurements with absolute water-table depths, or comparisons of measurements made within one microtopographical designation do not yield the same relationship (Supplemental Fig. 1). Downcore studies rely on the assumption that the microtopographic designation of a particular spot (the core location) does not change through time, therefore, a carbon isotope interpretation, which relies solely on microtopographical designation, and not on absolute water-table depth cannot be applied. This assumption that positions of hummocks and hollows are constant over time is supported by paleoecological studies (Charman, 2002; McMullen et al., 2004). The one published downcore record of compound specific carbon isotopic data of *Sphagnum* lipids (Xie et al., 2004) presents results that are consistent with our interpretations. We have plotted the carbon isotopic ratios of *Sphagnum* lipids from Xie et al. (2004) with regional precipitation during the same time period (Supplementary Fig. 2). At this location, times of reduced precipitation are contemporary with higher $\delta^{13}C$ values of *Sphagnum* lipids, and vice versa. Overall, we conclude that rate of methane recycling is the dominant factor controlling the $\delta^{13}C$ values of individual *Sphagnum* lipids found in downcore peat samples.

Humification

Humification, a long-used proxy for peatland moisture balance, is based on the fact that when dried, the surface of a peat bog decays. The humification index measures the amount of decay which the peat has undergone based on the amount of humic substances present. More humic substances indicate drier conditions at a particular interval (Blackford and Chambers, 1995). Most of this decay occurs during the summer. In a broad sense, both $\delta^{13}C$ of C_{24} *n*-acid and humification index in the Bøstad core show similar variations (Fig. 4). However, the transitions between dry and wet intervals appear gentler in the humification record. The drying trends, especially those of the early Holocene into the middle Holocene, appear to begin earlier than in the $\delta^{13}C$ record. This is likely due to overprinting of the humification signal, as the decay during dry intervals occurs throughout the unsaturated zone of the peat, which, during dry times, can be up to

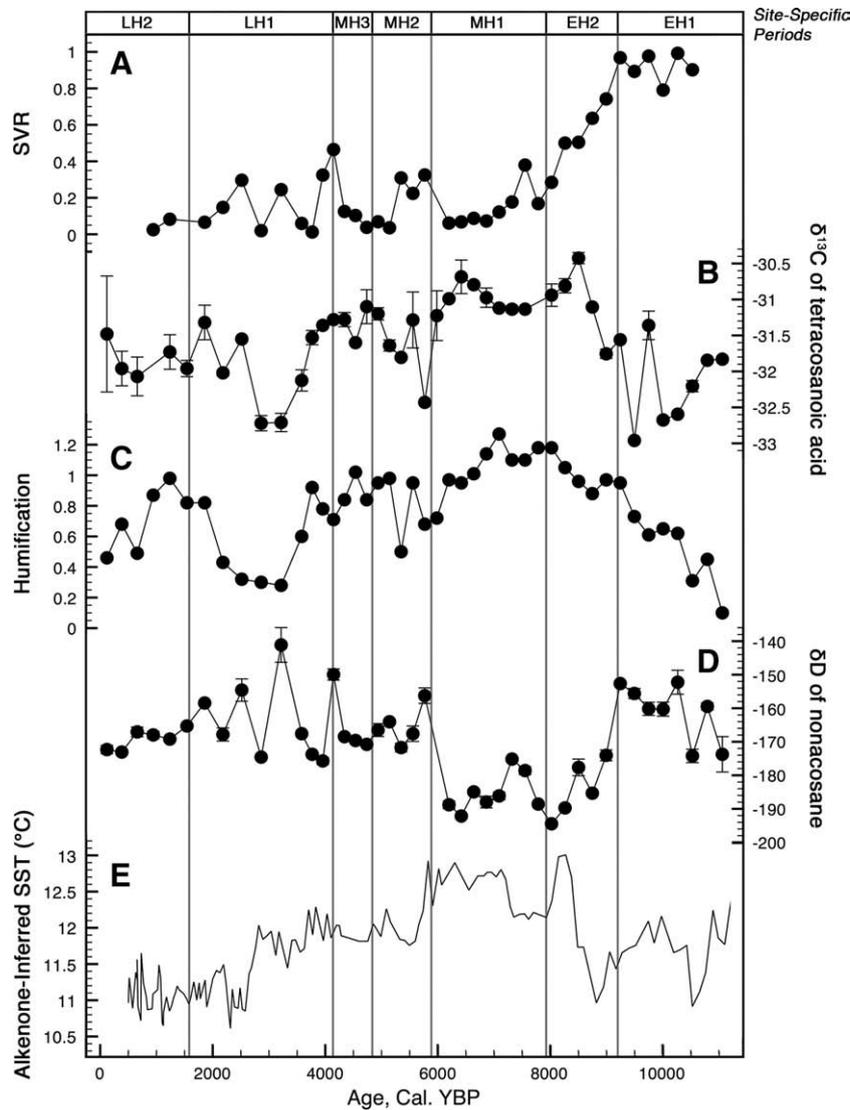


Figure 4. Downcore plots of various proxy data for the past 12 ka. (A) *Sphagnum* to Vascular *n*-alkane ratio (SVR) indicating wetness/dryness of Bøstad Bog. (B) The $\delta^{13}\text{C}$ values of C_{24} *n*-acids extracted from Bøstad Bog sediments. (C) The humification index for Bøstad Bog, as determined by the transmittance method. (D) The δD values of C_{29} *n*-alkanes extracted from Bøstad Bog sediments. (E) Sea surface temperatures at core location MD952011 based on alkenone paleothermometry (Calvo et al. 2002).

50 cm (Booth, 2002). The $\delta^{13}\text{C}$ signal, however, is not affected by diagenetic overprinting (Huang et al., 1997), and hence is a more precise and finely resolved recorder of past moisture variations.

Isotope modeling

We have developed two models to convert our compound-specific stable isotope methods into climate parameters. One model translates carbon isotope values of *Sphagnum* biomarkers into growing season bog surface wetness, and another translates hydrogen isotope ratios of vascular plant biomarkers into annual average δD of precipitation and subsequently to seasonality of precipitation.

To determine the wetness of the surface of the bog at a given interval, we use a mass balance model to find the amount of methane-derived CO_2 the *Sphagnum* is using. To calculate the wetness fraction, first we calculate the $\delta^{13}\text{C}$ value of the CO_2 assimilated by the *Sphagnum* by converting the $\delta^{13}\text{C}$ value of C_{24} *n*-acid using a biosynthetic fractionation factor, $\epsilon = -25\%$. Because no biosynthetic fractionation factor has been determined for *Sphagnum* specifically, we use an average fractionation factor for C3 photosynthetic plants (Collister et al., 1994). Raghoebarsing et al. (2005) determined in vitro that the maximum input of methane-derived CO_2 used by *Sphagnum* is approximately 15% of the total. We generalize for this model that

this maximum input occurs when the *Sphagnum* is wettest. We believe this to be a reasonable assumption because at low-water conditions, much of the methane produced at depth is oxidized before reaching the surface, whereas when the water table is high, much more methane oxidation takes place at or near the surface, where the *Sphagnum* can take advantage of the resultant CO_2 (Blodau and Moore, 2003). Our mixing model assumes that methane-derived CO_2 has a $\delta^{13}\text{C}$ value of -40% and atmospheric CO_2 is constantly -6.5% (Clark and Fritz, 1997). With a 1.0 surface wetness fraction (completely wet), we assume the *Sphagnum* is assimilating 15% methane-derived CO_2 and 85% atmospheric CO_2 . At 0.0 wetness fraction (completely dry), the *Sphagnum* is assimilating 100% atmospheric CO_2 . The results from our surface wetness fraction model are shown in Figure 5. Our “surface wetness” is therefore a normalized relative scaling factor based on maximum and minimum (15 and 0%) assimilation of methane-derived CO_2 by *Sphagnum* during photosynthesis.

To calculate the fraction of warm season (May–October, hereafter referred to as “summer”) precipitation from the δD of vascular plant biomarkers, we first convert the δD of C_{29} *n*-alkane into the δD of the water assimilated by the vascular plants using an apparent fractionation factor, $\epsilon = -73.5\%$. Recent studies suggest the fractionation of plant leaf waxes can range from -70 to -130% (Hou et al., 2007,

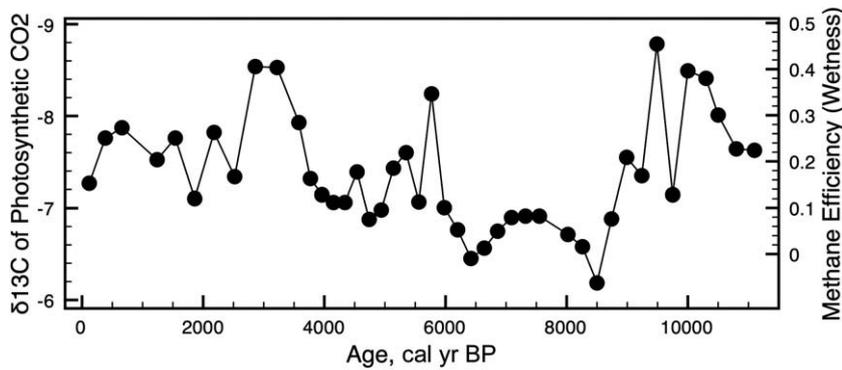


Figure 5. Results of our surface wetness model.

2008). There is currently no reported value for vascular plants living in bogs. We choose $\epsilon = -73.5\%$ to force the model such that the most recent fraction summer precipitation values we find in our core lie at the center of the range of modern values. If we have incorrectly estimated this fractionation factor, it will not change the relative trends in reconstructed downcore seasonality of precipitation, and hence will not affect our interpretation of hydrological changes over time for Bøstad Bog. The δD of precipitation at high latitudes is determined primarily by the temperature effect (Clark and Fritz, 1997). We remove the effect of changing annual temperature from our δD record (Fig. 4) using the changes in temperature from the SST record of Calvo et al. (2002), and the modern temperature– δD of precipitation relationship at the Bøstad location ($r^2 = 0.8$): $\delta D = 2.9 * T - 113.8$ (calculated from OIPC, Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005).

The remaining δD variation, after removing the temperature effect, can then be interpreted as a mass balance effect of the changing amount of precipitation falling during the warm and cold seasons. We use modern meteorological data to build a relationship between the fraction summer precipitation and the annual average δD of precipitation. We use monthly temperature and precipitation data from the nearest weather station at Bodo, Norway, and the local relationship between average monthly temperature and average monthly δD of precipitation (calculated from OIPC, Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005) to create a precipitation amount-weighted average δD of precipitation for each year AD 1951–2006. We regressed the fraction precipitation falling from May to October (fraction summer precipitation, FSP) for each year against the δD of mean annual precipitation (MAP) for that year, yielding a linear relationship ($r^2 = 0.5$):

$$\text{FSP} = 0.02 * \delta D_{\text{MAP}} + 2.4.$$

We use this linear relationship to translate our δD of precipitation derived from vascular plant alkanes into fraction summer precipita-

tion. The modeled fraction summer precipitation from the Bøstad location is shown in Figure 6. Large changes in δD of precipitation can also be interpreted as changes in moisture source. However, since our site is so close to the coast, it is unlikely an influence of moisture from sources other than the North Atlantic can overwhelm the signal from local moisture.

Paleoclimatological interpretations

The Holocene climate history of Arctic Fennoscandia has been extensively investigated using a large array of continental and oceanic proxies. Snowball et al. (2004) recently summarized these results. During the early Holocene, 11,500–7500 yr, Fennoscandia experienced a cool, moist climate. The climatic conditions are warm and dry during the Holocene thermal maximum, 7500–5500 yr. From 5500 yr to present, regional climate trends toward cooler temperatures and overall wetter conditions (Snowball et al., 2004). Published pollen data from Bøstad Bog (Johansen and Vorren, 1986) are consistent with these general climate trends: early Holocene vegetation was dominated by *Betula* forest, which requires a cool, moist climate; *Pinus*, a more dry-tolerant taxon, becomes dominant in the mid-Holocene; in the later Holocene, taxa representing cooler and wetter climate reappear (Johansen and Vorren, 1986). Building on these well-established general climate scenarios, our new data provide more refined reconstructions of hydrological variations with respect to precipitation seasonality and surface wetness.

Early Holocene 11,000–8200 yr

The early Holocene period can be divided into two parts. During the first part of the early Holocene, 11,000–9200 yr (Fig. 4, EH1), the SVR shows the highest level of *Sphagnum* productivity for the entire record. There is very little contribution of alkanes to the sediment by vascular plants. This low contribution by vascular plants is accompanied by high fraction summer precipitation values. Between 9200 and 8000 yr (EH2), there is a sharp increase in contribution of *n*-

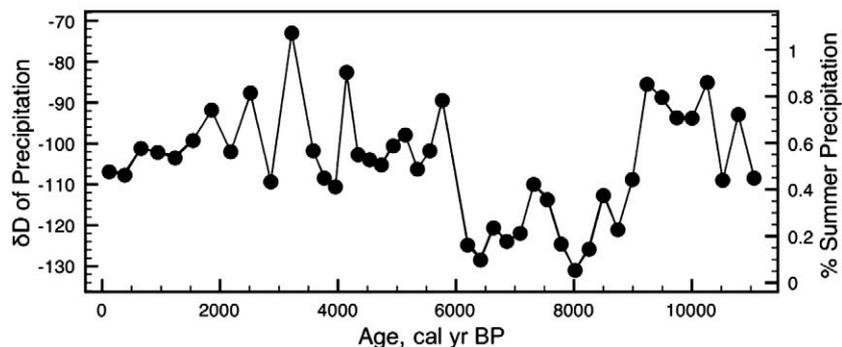


Figure 6. Results of our precipitation seasonality model.

alkanes by vascular plants, as well as high bog surface wetness. During this early Holocene period, the climate of this region transitions from a cool, wet summer with a dry winter, to a more evaporative summer with most of the moisture recharge occurring during the winter.

The climate of the early Holocene period described here is consistent with the vegetation assemblage identified at Bøstad by Vorren (1986). The regional vegetation during this time transitions from tundra to birch forest. This vegetation change is in agreement with the moisture regimes suggested by the biomarker and isotope data. Tundra is supported during cold, dry winters and cool moist summers, as in EH1. The transition to wetter winter and warmer,

more evaporative summers in EH2 is consistent with the climate associated with birch forest.

Mid-Holocene 8000–4300 yr

The mid-Holocene, between 8000 and 4300 yr (MH1–3) is characterized by low input of *n*-alkanes by *Sphagnum* (high input by vascular plants). This indicates an overall dry condition at the surface of the bog. This dry condition, suggested by the low productivity of *Sphagnum* as compared with vascular plants, is supported by a low surface-wetness fraction, indicating an evaporative summer growing season.

The mid-Holocene dry period is interrupted by an approximately 1000-yr interval of wetter conditions, between 6000 and 5000 yr

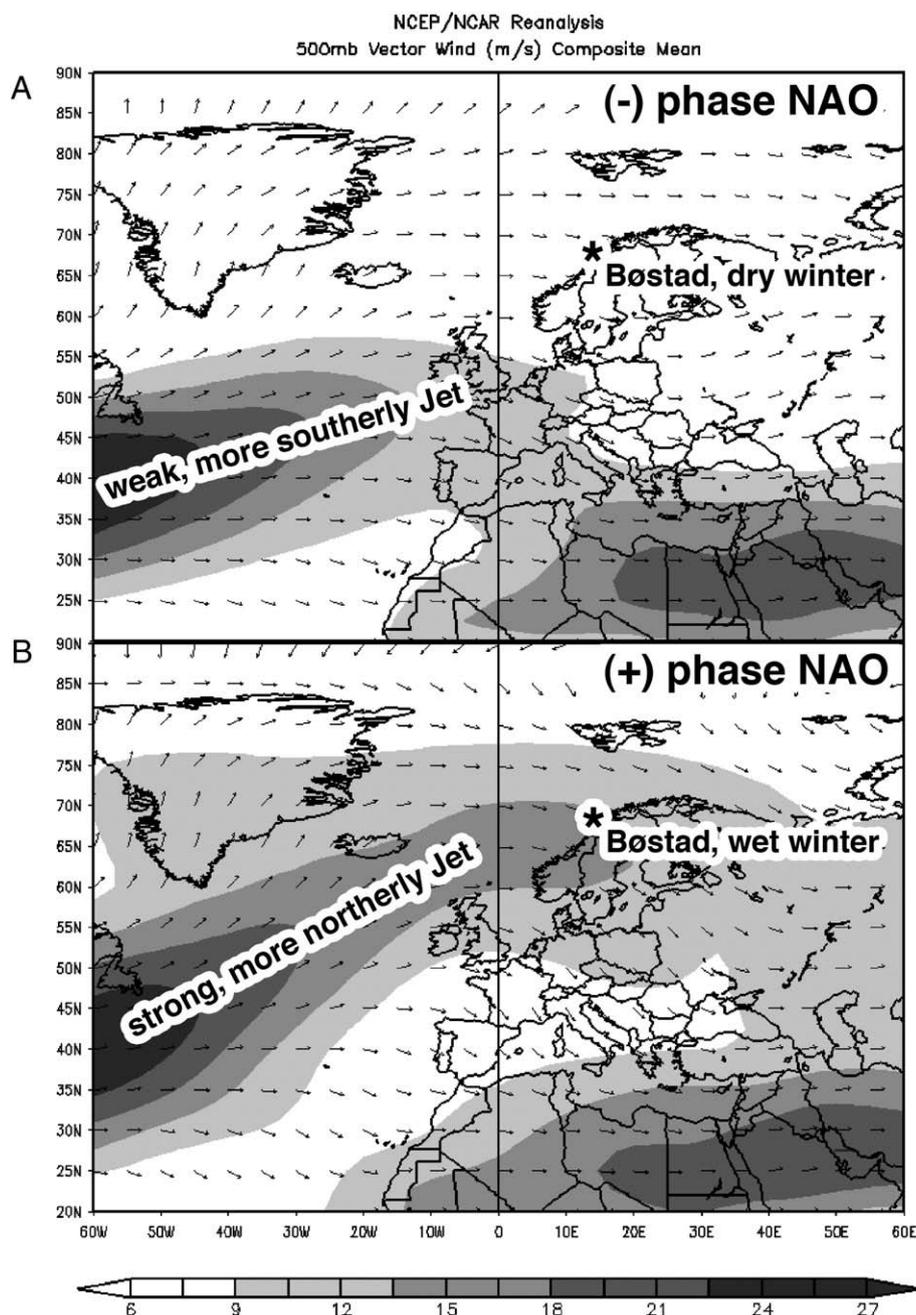


Figure 7. Average 500mb wind vectors for the winters (DJF) with the least amount of precipitation, AD 1966, 1967, 1969, 1977, 1985, 1999, (A) and the most precipitation, AD 1963, 1964, 1992, 1993, 2000, (B) of the past 50 years as measured at Bodo, Norway. The fastest wind speeds define the position of the Jet Stream, which is responsible for steering the storms that deliver moisture to the continent. Plot A, the dry winters, with a weaker, more southerly Jet are illustrative of negative NAO condition, while plot B, with the wettest winters, and a strong, northerly Jet, are illustrative of the conditions which prevail during a positive NAO condition (Hurrell and Dickson, 2004). Images are modified from those provided by the NOAA-ESRL Physical Sciences Division, Boulder Colorado from their Web site at <http://www.cdc.noaa.gov/>, using data from the NCEP/NCAR 40-year reanalysis project (Kalnay et al., 1996).

(MH2). This wet period is characterized by an increased SVR and summer wetness fraction, indicating increased overall wetness at peatland surface and decreased summer evaporation. The start of the wet period is also marked by a shift in the δD of vascular biomarkers towards higher values, indicating a shift of precipitation seasonality in favor of summertime.

Late Holocene, 4300 yr–present

The late Holocene is characterized by higher variability in the paleohydrologic record than previously, but with overall wetter conditions than in the mid-Holocene. Though the overall hydrologic balance indicated by the SVR is wetter, the contribution of winter precipitation is greatly reduced, as indicated by the high FSP values (Fig. 5). The contribution of summer precipitation peaks during this time at ~3200 yr. This age also marks the peak in summer wetness as indicated by the $\delta^{13}C$ of *Sphagnum* biomarkers. Though the overall state of the peatland is not as wet as it was in the early Holocene, the seasonality of precipitation is similar. This precipitation regime is supported by the return of the *Betula* forest, peaking at ~3000 yr.

Norwegian Sea surface temperature and winter precipitation

The Norwegian Sea is the moisture source for the precipitation falling on Bøstad Bog. As this location is an island in the Norwegian Sea, it is unlikely that the moisture source has changed significantly through time. The δD of vascular plant biomarkers, however, changes dramatically throughout the Holocene. These changes in the δD of vascular plant biomarkers reflect changes in the δD of water stored in the peatland. These shifts are most likely a consequence of an isotope mass balance effect, as the peatland recharge fluctuates between times when the peatland is recharged more by winter precipitation, to times when it is more by summer precipitation. The amount of precipitation falling during the winter at a particular location is determined by how many storms track over that location. The winter storms, which are responsible for the delivery of moisture from the ocean to the land surface, are steered by the Jet Stream. The position and strength of the Jet is determined by the pressure gradient between the subtropical high-pressure system, and the subpolar low, referred to as the North Atlantic Oscillation (Hurrell and Dickson, 2004). In turn, the average position of the jet over a winter season determines where the greatest amount of winter precipitation will fall that season. Generally, during a positive phase (stronger gradient) of the North Atlantic Oscillation (NAO), the jet stream is forced northeastward (Flatau et al., 2003), consequently bringing more precipitation to Bøstad. Figures 7A and B show the average position of the jet for the five wettest, and the five driest winters at Bodo, Norway (the closest weather station to the Lofoten Islands with long records of monthly climate) of the past 50 years. During the five wettest winters (Fig. 7B), the Jet stream tracks directly over the Lofoten Islands and Bodo, but during the driest winters (Fig. 7A), the Jet tracks much farther south and brings the storms, and the precipitation, with it.

Changes in SST of the Norwegian Sea occur contemporaneously with changes in the record of the δD of vascular plant biomarkers from Bøstad (Fig. 4D), even though the SST record is reflecting changes in summer SST, and the δD record of vascular plant biomarkers is recording changes in winter precipitation. We propose that these two responses are driven by a similar atmospheric mechanism. Positive NAO, which is a wintertime phenomenon, is correlated with stronger transport of water by the Gulf Stream to the Norwegian Sea and, hence, warmer Norwegian Sea SST (Calvo et al., 2002; Flatau et al., 2003). The changes in the Gulf Stream occur during the spring following the positive NAO condition (Frankignoul et al., 2001). This interaction between the NAO and Gulf Stream transport describes a mechanism that would produce high summer SST in the Norwegian Sea coincident with increased winter precipitation at Bøstad Bog. The

influence of this mechanism on continental climate is consistent with other interpretations of climate data from this region of the continent (Seppä and Hammarlund, 2000), as well as of Norwegian Sea SST (Birks and Koç, 2002; Calvo et al., 2002).

Conclusions

Compound specific stable isotope ratios of *Sphagnum* and vascular plant biomarker compounds from the sediments of ombrotrophic peatlands can be used to reconstruct changes in the seasonality of precipitation and bog surface wetness. Changes in the δD of vascular plant biomarkers record changes in the δD of stored water in the peatland, which represents annual average precipitation. Changes in δD values of this reservoir are affected by mass balance changes of precipitation falling during warm and cold seasons. Using our model, we can translate the downcore changes in δD of plant leaf waxes into changes in the fraction summer precipitation. The $\delta^{13}C$ of *Sphagnum* biomarkers records the relative contribution of methane-derived CO_2 from symbiotic methanotrophs. The activity of these methanotrophs is related to the moisture content of the *Sphagnum* at the surface of the peatland during the growing season (summer). Wetter conditions (higher water table) leads to enhanced methanotrophic activity, a more ^{13}C depleted source CO_2 for *Sphagnum* and lower $\delta^{13}C$ values for *Sphagnum* biomarkers (C_{23} *n*-alkane and C_{24} *n*-acid). The amount of precipitation Bøstad Bog receives during winter is strongly related to the surface temperature of the Norwegian Sea, as this is the moisture source for the Lofoten Islands. Warmer sea surface temperatures are also related to a prolonged positive condition of the North Atlantic Oscillation, which also directs more moisture delivery to the continent, especially in winter (Hurrell and Dickson, 2004).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.yqres.2009.07.007.

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