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Comparison of stable carbon and oxygen isotopes in *Picea glauca* tree rings and *Sphagnum fuscum* moss remains from subarctic Canada

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ABSTRACT

Stable isotope ratios from tree rings and peatland mosses have become important proxies of past climate variations. We here compare recent stable carbon and oxygen isotope ratios in cellulose of tree rings from white spruce (*Picea glauca*), growing near the arctic tree line; and cellulose of *Sphagnum fuscum* stems, growing in a hummock of a subarctic peatland, in west-central Canada. Results show that carbon isotopes in *S. fuscum* correlate significantly with July temperatures over the past ~20 yr. The oxygen isotopes correlate with both summer temperature and precipitation. Analyses of the tree-ring isotopes revealed summer temperatures to be the main controlling factor for carbon isotope variations, whereas tree-ring oxygen isotope ratios are controlled by a combination of spring temperatures and precipitation totals. We also explore the potential of combining high-frequency (annual) climate signals derived from long tree-ring series with low-frequency (decadal to centennial) climate signals derived from the moss remains in peat deposits. This cross-archive comparison revealed no association between the oxygen isotopes, which likely results from the varying sensitivity of the archives to different seasons. For the carbon isotopes, common variance could be achieved through adjustments of the *Sphagnum* age model within dating error.

Introduction

Compared to the dense network of instrumental stations and large number of climate reconstructions from various archives in western Eurasia, information about past climate and environmental changes is sparse in sub-arctic North America. This poor coverage not only limits conclusions about climatic changes over the observational period (i.e., the past few decades), but also constrains the evaluation and calibration of proxy archives that could be used to assess Holocene climate variations in arctic and subarctic North America.

Twentieth century high-latitude climatic and environmental changes are characterized by increases in air temperature (mainly during winter and spring), reductions of spring snow cover, a northern expansion of the arctic tree line, and permafrost temperature increases that stimulate carbon fluxes in arctic environments and increase CO_2 and CH_4 releases into the atmosphere (Serreze et al., 2000). There are, however, quite substantial regional deviations from this general pattern, such as eastern Canada where surface temperatures and permafrost have cooled since the 1960s (Brown and Braaten, 1998; Chapman and Walsh, 1993) until quite recently. There is thus a need for more paleoclimate information from this huge area to evaluate long-term climate variability and to assess the full range of possible future changes.

Stable isotopes in tree rings are sensitive to climatic changes and can be used as quantitative proxies for various parameters such as growth season temperature, relative humidity, soil moisture and irradiance (McCarroll and Loader, 2004; Treydte et al., 2007; Buhay et al., 2008; Holzkämper et al., 2008; Kirdyanov et al., 2008; Esper et al., 2010). Other studies have shown that the isotopic signature of *Sphagnum* peat can also be used to reconstruct summer temperature, precipitation amounts, and water table depth, depending on the environmental situation and species studied (Skrzypek et al., 2007a; Loisel et al., 2009; Moschen et al., 2011), in some cases over several millennia (Kaislahti Tillman et al., 2010a) and likely without long-term biases as long as a single moss species and the same moss components are analyzed (Kaislahti Tillman et al., 2010b). However, peat proxies suffer from relatively low temporal resolution (centennial to decadal, exceptionally annual). Tree rings, in contrast, are annually resolved (Fritts, 1976) but suffer from limitations in retaining low-frequency (centennial-scale) climate variations (Esper et al., 2002) and cover shorter periods (usually several centuries to millennia (Frank et al., 2010). The integration of stable isotope analyses from both peat bog and tree-ring archives offers the possibility to assess and combine high- and low-frequency climate signals, which would likely reduce uncertainties typical for reconstructions based on individual archives (Jones et al., 2009).

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To test this possibility, we investigated tree-ring and peat isotopes from two close-by locations in the west-central Canadian sub-Arctic. The area is suited for this analysis as trees and peat hummocks grow in adjacent habitats, and human disturbance is minimal. Moreover, the extreme climatic conditions (extremely low winter temperatures and short growing season) likely lead to high sensitivities of vegetation parameters even to relatively small changes. Here, we compare annually resolved carbon ($\delta^{13}C$) and oxygen ($\delta^{18}O$) isotope data from *Picea glauca* with respective isotopes in *Sphagnum fuscum* mosses dated by ^{210}Pb and correlate the time series with observational climate data from nearby climate stations. We presume that $\delta^{13}C$ in tree-ring and moss cellulose correlates with growing-season temperatures (Skrzypek et al., 2007b; Holzkämper et al., 2008; Tardif et al., 2008; Kaislahti Tillman et al., 2010b), whereas $\delta^{18}O$ signals in tree rings and moss may be more difficult to combine, due to the more complex mechanisms occurring during the transfer of oxygen isotopes and related fractionation processes from precipitation to the plant body (Ménot-Combes et al., 2002; McCarroll and Loader, 2004).

Study sites and methods

Location

The two sampling sites are located in the west-central Canadian sub-Arctic, ~400 km west of Hudson Bay. The area is underlain by crystalline rock of the Canadian Shield, which is covered by Quaternary glaciofluvial sediments, forming a rather flat landscape (topographic relief <100 m) with a variety of periglacial features (Dyke and Prest, 1987) (Fig. 1). Discs (EL1–3) from three white spruce (*Picea glauca* (Moench) Voss) trees were collected in summer 2005 near the northern shore of Ennadai Lake (61°03′N, 101°41′W, 350 m.a.s.l.) in the northern forest tundra zone (MacDonald et al., 1998), a region that is underlain by continuous permafrost (Zoltai, 1995). Black spruce (*Picea mariana*), which is also a common species in the study area, was disregarded as it preferably grows in wetter lowland areas and might thus show a dampened or different reaction to climatic changes compared to white spruce. All three *Picea glauca* trees grew on a large, ~5–10 m high esker, where the hydrological and microclimatic conditions result in the absence of near-surface permafrost supporting development of taller trees. Trees were growing within a 1-km radius; EL1 was ~11 m tall, growing on a SE-facing steep and well-drained slope near the top of the esker. EL2 was ~12 m tall and growing on the lower part of the esker complex, on a gentle SW-facing slope. EL3 reached ~12 m, growing near the top of the esker on a NW-facing slope. Ring counting and cross dating revealed that trees EL1–3 cover the 1931–2005, 1930–2005, and 1856–2005 periods, respectively.

In this study, we compare data derived from trees at Ennadai with *S. fuscum* ((Schimp.) Klinggräff) hummock samples (ML3), which were also collected in summer 2005, near Misaw Lake (59°52′N, 102°34′W), located ~140 km southwest of Ennadai Lake, in the southern forest tundra zone. The ML3 hummock was found in a permafrost underlain peatland within the zone of discontinuous permafrost (Kaislahti Tillman et al., 2010b). It consists mainly of well-preserved *S. fuscum* mosses, but some *Polytrichum* sp. mosses and rootlet layers are also present. More details of the stratigraphy are described in Kaislahti Tillman et al. (2010b). Vegetation in the peatland includes *Picea glauca* (up to ~6 m high) and numerous *Sphagnum* hummocks with *Ledum* spp., *Betula glandulosa* and *Vaccinium vitis-idaea* as ground vegetation. The active layer depth was 51 cm in July 2005.

Climate

Climate data were obtained from the Canadian Weather Service, where they were quality-controlled. The data are integrated in the global collections of GISS and GHCN and used to assess regional to large-scale climate change considered in a number of publications (including the IPCC). At the closest meteorological station to the collected trees, Ennadai Lake (61°22′N, 100°54′W, 325 m.a.s.l.), 50 km northeast of the EL site, mean annual temperature was −9.2°C, and mean January and July temperatures were −31.3°C and +12.9°C over the 1950–1978 period, respectively (Fig. 2). Meteorological observations were interrupted in 1979 and resumed in 1998. Mean annual

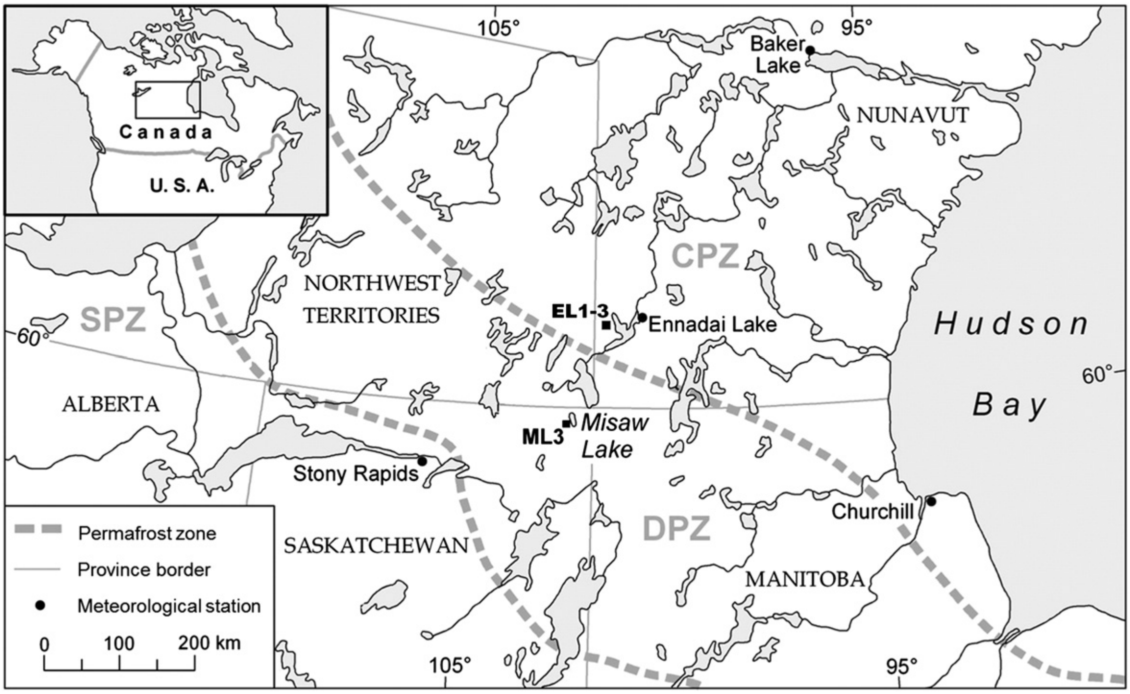

Figure 1. Study sites Misaw Lake (ML) and Ennadai Lake (EL) in western–central Canada. Dotted lines indicate boundaries of the present permafrost zones: SPZ = sporadic, DPZ = discontinuous and CPZ = continuous permafrost zone (Zoltai, 1995). Trees EL1–3 are located at Ennadai Lake, ~140 km northeast of ML. Map redrawn after Sannel and Kuhry (2008).

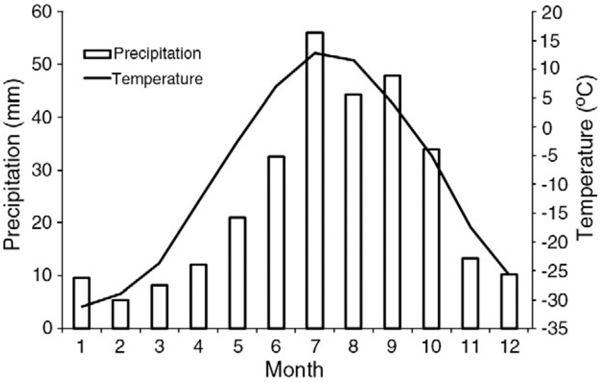

Figure 2. Mean monthly temperature and precipitation at climate station Ennadai Lake over the 1950–1978 period (data source: www.knmi.nl).

precipitation was 291 mm, with ~60% occurring during the June–September growing season. At Stony Rapids climate station (59°15′N, 105°50′W, 245 m.a.s.l.), ca. 200 km southwest from the collected peat hummock at Misaw Lake, mean annual temperature was −3.3°C and mean annual precipitation was 430 mm for the 1987–2002 period. Climate observations at Stony Rapids were interrupted from 1978–1986. Correlations between the two climate stations for the years with overlapping records are $R^2 = 0.9$ for annual temperature and $R^2 = 0.29$ for annual precipitation. Similar patterns of correlations between climate stations in the region were reported by Kaislahti Tillman et al. (2010b).

Sample preparation and statistical analysis

For chronological control (cross-dating), ring-width measurements were carried out with an accuracy of 0.01 mm on three radii for each tree. Annual growth was vigorous for all three trees and no signs of frost or other damages were visible. The tree-ring series were visually compared against each other and against a larger set of samples collected during the field trip, using the graphical and statistical programs CATRAS (Aniol, 1983) and COFECHA (Holmes et al., 1986). We extracted α-cellulose from single *Picea glauca* tree rings (cut from two radii per tree using a surgery knife) and from *S. fuscum* stems (following procedures outlined in Loader et al., 1997). A detailed description of the extraction method for the moss material based on sodium chlorite is provided in Kaislahti Tillman et al. (2010b). To ensure a complete chemical separation of tree-ring holo- and α-cellulose, we also applied sodium hydroxide (NaOH 17% m/v) for 45 min at room temperature to the annual wood samples. The $\delta^{13}C$ values were corrected for the increasing atmospheric CO_2 concentration and the related depletion (following procedures outlined in McCarroll and Loader, 2004). In addition, $\delta^{13}C$ data were corrected for enhanced water-use efficiency with 0.007‰ per ppm CO_2 increase relative to pre-industrial values (Kürschner, 1996). The so-called pin-correction for the plant-physiological response to increased atmospheric CO_2 during the industrial era (McCarroll et al., 2009) was not applied here. This is because (1) the effect of such detrending is small given the relatively short time period covered by our data, and (2) there is a risk of removing trends that are related to low-frequency climatic change, as the magnitude of the pin-correction is statistically linked to the calibration target (i.e., climate) (Treydte et al., 2009).

The ten innermost (oldest) tree rings of EL1 and EL2 were not considered for isotope measurements to avoid juvenile effects, which arise from rapid changes in CO_2 use and hydraulic conductivity of young trees and shading effects of surrounding vegetation (McCarroll and Loader, 2004). The extent and duration of the juvenile effect differ widely, depending on micro-site conditions and species. Even though it is preferable to omit 20–30 yr, we decided to use a shorter period to increase the overlap with observational climate data for calibration and because of the low ground vegetation and small shading effects at tree line, which likely lead to smaller and shorter juvenile effects compared to other environments (Holzkämper et al., 2008; Porter et al., 2009). From EL3, only the tree rings common with EL1 and EL2 were considered, restricting the analysis period to 1936 to 2004.

Correlation analyses of the tree-ring isotope time series with observed climate data were conducted using the software DENDROCLIM 2002 (Biondi and Waikul, 2004). Signal strength of the isotope data was assessed using a 25-yr moving window of the inter-series correlation (*Rbar*) and the expressed population signal (*EPS*) (Wigley et al., 1984). For tree-ring calibration, instrumental data from the Ennadai Lake station were used, where climate was observed continuously between 1949 and 1979 (Environment Canada, 2010). The isotopic signals derived from *S. fuscum* stems at site ML3 were compared to meteorological observations at Stony Rapids over the 1986–2004 period. We considered using the Ennadai Lake station data for both archives as well as using precipitation data averaged from several sites (Kaislahti Tillman et al., 2010b), but decided against those options because (1) the station at Ennadai Lake was closed in 1980–1997 and could not be used for the calibration of the most recent *Sphagnum* samples, (2) temperature patterns are quite homogeneous over central Canada but there is a large spatial variability in precipitation amounts (R^2 as low as 0.12 between station records from Ennadai Lake, Stony Rapids, Baker Lake, and Churchill), which hampered the use of Stony Rapids' precipitation data for tree ring calibration, and (3) the large distance to neighboring climate stations implies that important climatic and ecological boundaries, such as the arctic front and the arctic tree line, are crossed in some cases (MacDonald et al., 1998). Shifts of these boundaries over time might cause problems when longer reconstructions are generated using calibration data sets from stations that are far away from the climate archive.

Results

The $\delta^{13}C$ values of the three trees range between −21.1‰ and −23.9‰ (after correction) (Table 1) and show a generally good match with each other, with an inter-series correlation (*Rbar*) of $R^2 = 0.45$. However, EL3 features substantially (0.5–1‰) higher $\delta^{13}C$ values compared to EL1 and EL2 from about 1970 and onwards. The running inter-series correlation was calculated over 25-yr time windows lagged by one year. Results show a wide range of R^2 from 0.43 to 0.78 with a positive trend after 1960 (Fig. 3). A slight trend towards more negative $\delta^{13}C$ values is recorded in EL1 and EL2 over the 1936–1995 period, a feature that is absent in EL3. After 1997, $\delta^{13}C$ values decrease by ~1‰ in all three trees. Inter-annual variability is in the order of 1.5‰ throughout the past 70 yr.

Table 1
Average values, ranges and standard deviations for $\delta^{13}C$, $\delta^{18}O$ and ring width (TRW, period 1935–2004) for *Picea glauca* trees EL1–3.

	Average	Range	Standard deviation
$\delta^{13}C$ (‰)			
EL1	−22.79	−22.93 to −21.97	0.36
EL2	−22.66	−23.44 to −21.68	0.44
EL3	−22.34	−23.35 to −21.14	0.51
$\delta^{18}O$ (‰)			
EL1	18.79	17.01–21.33	0.96
EL2	19.02	15.73–21.75	1.17
EL3	19.44	17.14–22.42	0.96
TRW (all values in 1/100 mm)			
EL1	152	95–220	31
EL2	208	66–407	72
EL3	146	91–214	30

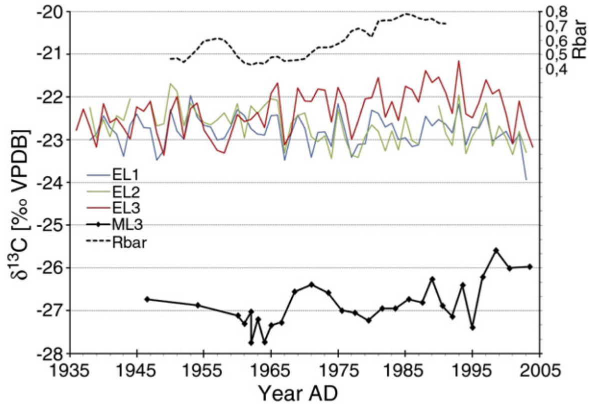

Figure 3. Tree-ring $\delta^{13}C$ values derived from α-cellulose of three *Picea glauca* trees growing near Ennadai Lake (EL1–3) compared with those of *Sphagnum fuscum* from a peat hummock located near Misaw Lake (ML3). Note the reduced and temporally changing resolution in the moss data. The values were corrected for atmospheric ^{13}C decrease after McCarroll and Loader (2004) and for plant physiological effects due to atmospheric CO_2 -concentration increase after Kürschner (1996). Running inter-series correlation (*Rbar*) between the three trees is calculated over a 25-yr moving window.

$\delta^{18}O$ values of the sampled trees range from 15.7‰ to 22.4‰, with high-frequency variance between 1 and 2‰ (Table 1, Fig. 4). The time series are characterized by common decadal-scale variations superimposed on a positive long-term trend. This visual impression is corroborated by an increase of the inter-series correlation after application of a 10-yr smoothing filter from $R^2 = 0.52$ to $R^2 = 0.59$, whereas the residuals show a lower correlation ($R^2 = 0.39$). The running inter-series correlation (calculated in the same way as for $\delta^{13}C$) varies over time, with a R^2 range from 0.84 in 1964 down to 0.29 in 1985 (Fig. 4).

To assess whether the isotopic signals are representative for the study site, *EPS* values were calculated for $\delta^{18}O$ and $\delta^{13}C$. For $\delta^{18}O$, the *EPS* values are 0.77 for the 1938–2003 period and 0.93 for the 1951–1978 period (i.e., the calibration period with climate data), respectively. For $\delta^{13}C$ the values depend strongly on whether raw data or corrected data are used: *EPS* for the 1951–1978 period is 0.83 and 0.74 for the corrected and raw data, respectively. Commonly, an *EPS* value of 0.85 is used as a threshold to provide a site representative climate signal (Leavitt and Long, 1984; McCarroll and Loader, 2004). The choice of this number is just a general guideline (Wigley et al., 1984), especially since no attempt is made to utilize the found relationships between stable isotopes and climate for reconstruction purposes in the presented study. According to our data, climate reconstruction with *EPS* values >0.85 would require five trees for $\delta^{18}O$ and seven trees for $\delta^{13}C$ for the 1938–2003 period (assuming that all trees show a similar correlation to each other).

Our comparison of peat and tree-ring archives revealed a ~3–6‰ offset for $\delta^{13}C$, and an offset of ~1–5‰ for $\delta^{18}O$ (Figs. 3 and 4). There is no significant statistical correlation between the two archives, with $R^2 = 0.02$ ($p = 0.47$) for $\delta^{13}C$ and $R^2 = 0.09$ ($p = 0.12$) for $\delta^{18}O$. We correlated tree-ring and moss data by averaging tree-ring data from those years during which one layer of moss has formed. If we consider the dating uncertainty of 3–5 yr of the ^{210}Pb-age–depth model for the time period 1985–2004 (Kaislahti Tillman et al., 2010a), then a 0.5–1 yr shift of the ML3-chronology (i.e., well within dating uncertainty, Fig. 5) reveals a possible correspondence between the tree-ring and *Sphagnum* isotopes. The adjustment is based on the synchronization of the most dominant maxima and minima of the tree-ring and moss records. The coherence is particularly improved in the $\delta^{13}C$ data, where there is an increase in the correlation with summer temperatures (July, August, July–August) (Table 3).

S. fuscum $\delta^{13}C$ values are mainly controlled by July temperatures, but $\delta^{18}O$ values are related to both July temperature and precipitation at Misaw Lake (Kaislahti Tillman et al., 2010b). The relationships between tree-ring $\delta^{13}C$ and climate varied between sampled trees: all three trees have significant ($p<0.05$) positive correlations with August temperatures, EL1 and EL3 are correlated with June precipitation, and EL2 is negatively correlated with July precipitation (Table 2). Tree-ring $\delta^{18}O$ values are controlled by spring (April to June) temperatures and May precipitation.

Discussion

$\delta^{13}C$

The $\delta^{13}C$ values of tree-ring and moss cellulose appear to correspond through time, particularly with the minor adjustments to the chronology of the moss record (Table 3). However, the offset of ~4‰ between the tree-ring and moss records deserves further explanation.

Both trees and mosses use the C_3 pathway during carbon fixation, favoring the lighter ^{12}C isotope and discriminating against ^{13}C.

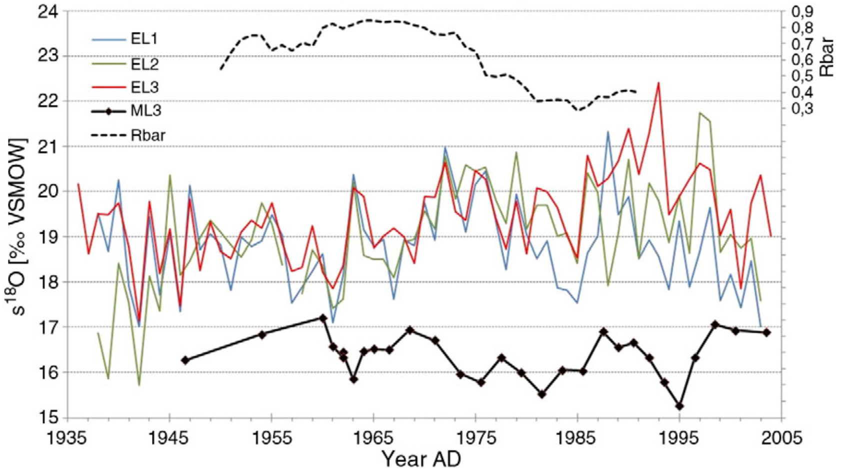

Figure 4. Tree-ring $\delta^{18}O$ values derived from α-cellulose of three *Picea glauca* trees growing near Ennadai Lake (EL1–3) compared with those of *Sphagnum fuscum* from a peat hummock located near Misaw Lake (ML3). Running inter-series correlation (*Rbar*) between the three trees is calculated over a 25-yr moving window.

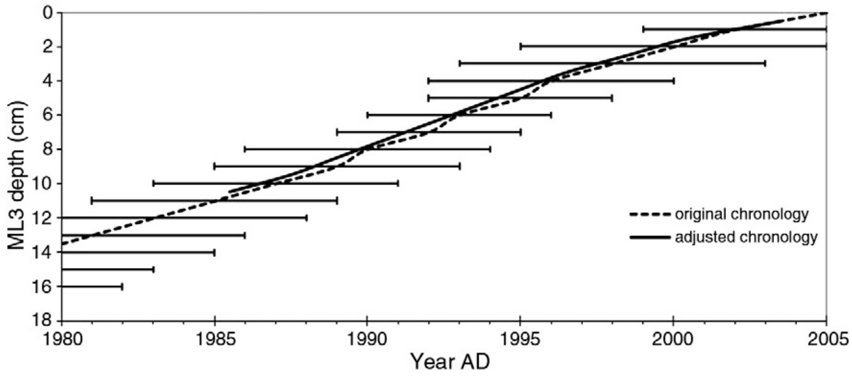

Figure 5. Age-depth model for the upper part of a peat hummock located near Misaw Lake (ML3) according to the original ^{210}Pb chronology (with error bars) and adjusted chronology (after Kaislahti Tillman et al., 2010a).

Following Farquhar et al. (1982), the δ^{13}C values of plant cellulose are defined as:

$$\delta^{13}C_{plants} = \delta^{13}C_{air} - a(1 - c_i/c_a) - b(c_i/c_a) \qquad (1)$$

where $\delta^{13}C_{air}$ is the δ^{13}C value in atmospheric CO_2 (at present ca. −8‰ VPDB), *a* is the fractionation coefficient describing the diffusion of air through stomata (ca. −4.4‰), *b* denotes net discrimination caused by carboxylation (ca. −27‰), and c_i/c_a denotes the ratio between intracellular and ambient CO_2 concentrations. The higher the CO_2 concentration in the internal leaf air (c_i relative to c_a), the stronger the stomatal conductance and the more negative the resulting δ^{13}C value in the plant cellulose. Discrimination against heavier isotopes also occurs during direct CO_2 uptake through water-filled hyaline cells surrounding the photosynthetic cells of *Sphagnum* leaves (Ménot-Combes and Burns, 2001; Moschen et al., 2011). Additional fractionation occurs when cellulose and other plant components are produced from leaf sugars, yielding lower δ^{13}C values. Because *Sphagnum* mosses do not have stomata, a theoretical equation for discrimination in moss plants would be:

$$\delta^{13}C_{plants} = \delta^{13}C_{air} - b(c_i/c_a) - d(f) \qquad (2)$$

where "*d*" denotes the species specific fractionation coefficient and depends furthermore on the cell water content *f*. Based on this simplified equation without a stomatal effect for mosses, the theoretical fractionation between atmospheric CO_2 and plant cellulose in tree and moss leaves would be ~ −35‰, if the ratio c_i/c_a is close to 1.

In terrestrial plants, the variation of c_i/c_a has been estimated to be 0.95–0.36‰ (Farquhar et al., 1982), while δ^{13}C values range from −20 to −30‰ for wood (McCarroll and Loader, 2004) and from −21 to −38‰ for non-vascular plants (Farquhar et al., 1989). The coefficient *d*(*f*) may be one of the causes of the lower δ^{13}C values in *Sphagnum* compared with tree cellulose. In addition, some studies suggest that differences in δ^{13}C of *Sphagnum* species are mainly governed by the surface moisture gradient, with the most negative values occurring in species growing on hummocks (as in our study) (e.g., Loisel et al., 2009, 2010). In arctic regions, where the growing season is short, high temperatures and sunlight intensity increase the rate of assimilation and thus deplete interior CO_2 concentration, decreasing c_i/c_a and fractionation (Loader et al., 2007). This leads to higher δ^{13}C values and positive correlations with temperature. The ML3 samples were collected from a hummock in the continental interior, where summer temperature is an important factor for the photosynthesis rates, and where variations in precipitation amounts are low (Schoning et al., 2005; Kaislahti Tillman et al., 2010a).

Table 2
Pearson correlations between tree-ring stable isotope values (δ^{13}C and δ^{18}O) and monthly temperatures and precipitation recorded at Ennadai Lake. Only correlations significant at the 95% level ($p \leq 0.05$) are shown.

	Temperature					Precipitation				
	Apr	May	June	July	Aug	Apr	May	June	July	Aug
$\delta^{13}C$										
EL1					0.51			0.38		
EL2					0.42				−0.42	
EL3				0.31	0.53			0.43		
Mean					0.61			0.49	−0.37	
$\delta^{18}O$										
EL1	0.43	0.5	0.49				0.51	0.35		
EL2		0.54	0.44				0.45			
EL3	0.35	0.47	0.41				0.51			
Mean	0.41	0.54	0.48				0.52			

An increased rate of photosynthesis due to increased temperatures (and sunlight) may also play a role in the positive δ^{13}C trend in ML3 over the past 70 yr. Summer temperatures rose by ~1.5°C over the 1986–2003 period (Stony Rapids climate station). Interestingly, the tree-ring δ^{13}C values do not reflect this positive trend, but show a decrease by ~1‰ over this period (Fig. 3). This obviously contradictory finding could be related to recent permafrost thawing (Laberge and Payette, 1995), leading to changing soil moisture conditions in the elevated esker complex and, in turn, drought stress for the trees during the growing season.

Although the radial growth rate in boreal forests reaches its maximum around the time of maximum day length (Rossi et al., 2006), a significant fraction of mass of the annual tree ring at high latitudes is formed in late summer as latewood (Briffa et al., 1998). According to our correlation analysis, the tree-ring δ^{13}C is most sensitive to temperature variations in August. There is, thus, a one-month delay compared to the period which shows highest correlations to the *S. fuscum* δ^{13}C composition. However, July and August temperatures are cross-correlated (particularly at lower frequencies) ($R^2 = 0.29$ in the Stony Rapids climate data), hampering an unambiguous correlation of tree-ring and *Sphagnum* isotopes with climate of a specific month (Table 3).

After the minor adjustment of the ML3 age–depth model, the coefficients of correlation between δ^{13}C and δ^{18}O with summer temperature increase compared to the original age model (Table 3). With such

Table 3
Coefficients of correlation (R^2) and ANOVA *t*-test values for a slope (p) for the linear relationship of the isotope data from the Misaw Lake peat record and observed temperatures from the Stony Rapids weather station for the period AD 1986–2004. *values from the previous age model by Kaislahti Tillman et al., 2010b and **values after age model adjustment by 0–1 yr to fit isotope data with observed data. The adjustment is within the age error (3–5 yr) of the ^{210}Pb chronology, see also Figs. 3 and 4.

	July			August			July–August		
	R^{2*}	R^{2**}	P^{**}	R^{2*}	R^{2**}	P^{**}	R^{2*}	R^{2**}	P^{**}
δ^{13}C	0.60	0.75	<0.01	0.45	0.46	0.02	0.60	0.66	<0.01
δ^{18}O	0.38	0.51	0.01	0.25	0.29	0.09	0.36	0.43	0.03

temporal adjustment of ML3, there is a strong match between the tree-ring and moss isotope curves (Figs. 3 and 4). However, the differing temporal resolution of the two records remains an obstacle for comparing the two archives. The tree-ring time series are annually resolved and precisely dated, but the resolution of the peat varies from annual to decadal through time, and dating uncertainties make it difficult to assign a sample to a specific year.

The tree ring δ^{13}C data are in line with results from a study on white spruce from nearby Churchill, where δ^{13}C values range between −22‰ and −24‰, and where a positive correlation was found between δ^{13}C and July–August temperature conditions (Tardif et al., 2008) and with a study on larch from northeastern Yakutia, where the positive correlation between δ^{13}C and June–August temperatures was ascribed to drought stress (Sidorova et al., 2008). However, viewing the records from Churchill and Ennadai together reveals some discrepancies in both long-term and inter-annual variability (Fig. 6). These discrepancies may be caused by differences in the amount and timing of precipitation, but they may also be caused by differences in the micro-site conditions. The latter hypothesis is supported by the lack of correlation between δ^{13}C and water-deficit variables (drought status, precipitation, relative humidity) at the Churchill site (Tardif et al., 2008), whereas part of the variability in δ^{13}C at Ennadai Lake can be explained by summer precipitation (Table 2). The Ennadai trees are likely sensitive to moisture due to their exposed location on the top or upper flanks of well-drained esker complexes. A study on white spruce from the Mackenzie Delta region in northern Canada also concluded that temperature-induced drought stress is an important factor for stomatal conductance and, with that, ^{13}C/^{12}C fractionation (Porter et al., 2009). Thus, the stable-isotope ratios of trees growing in xeric locations, like those in the Ennadai study site, may reflect variations in both summer temperature and moisture conditions.

$\delta^{18}O$

Three factors control the δ^{18}O composition of plant cellulose: (1) the isotopic composition of the source water (i.e., precipitation and/or soil water), (2) evaporative enrichment of the leaf water, and (3) the biochemical fractionation between source water and cellulose (Zanazzi and Mora, 2005). Assuming that the biochemical enrichment during cellulose formation remains at a constant level of +27‰ for both tree rings and peat as earlier investigations suggested (Epstein et al., 1977; Zanazzi and Mora, 2005), the observed offset of ~1–5‰ between tree ring and peat δ^{18}O must be due to differences in source water δ^{18}O and/or evaporative enrichment. Evaporative enrichment might be more significant for trees considering the larger leaf surface, drier air, and higher wind speed several meters above ground.

The weak association between δ^{18}O variations in tree rings and moss is likely related to different sensitivities to climate conditions at different times of the year. δ^{18}O values in precipitation are mainly controlled by condensation temperature but also by precipitation amount and cloud track history of the air masses (Rozanski et al., 1993). Whereas tree-ring δ^{18}O values were mainly influenced by both temperature and moisture conditions in spring and early summer (i.e., during times when the ground is still covered by snow; Table 1) (McCarroll and Loader, 2004), *Sphagnum*-δ^{18}O values at Misaw Lake were most sensitive to moisture and temperature conditions in July (i.e., snow-free conditions) (Kaislahti Tillman et al., 2010b), a period when much of the annual growth occurs (Lindholm, 1990). The observed discrepancy may thus be related to soil moisture reservoir effects on the trees, dampening the meteorological signal compared to the *Sphagnum* mosses. Even small differences in micro-site conditions, such as slope, aspect, soil moisture, and nutrient supply can influence tree growth at the arctic tree line (Wilmking et al., in press) and it is likely that such changes also influence the δ^{18}O signature of tree rings. Correlations between δ^{18}O in tree cellulose and early-spring to mid-summer temperature conditions in the Mackenzie Delta region were ascribed to the temperature-dependent isotopic composition of snowmelt and summer rain feeding the soil and groundwater reservoir (Porter et al., 2009), which is in line with results from our and other studies (Holzkämper et al., 2008; Sidorova et al., 2008). Since *S. fuscum* mosses grow under ombrotrophic conditions, they take up precipitation without any significant delay and without being influenced by ground water or pond water. Thus, there is a temporal sensitivity offset between the moss and tree-ring record.

In addition, an assessment of dominant wind directions in the study region suggests that air-mass trajectories differ widely between the Misaw Lake and Ennadai Lake sites (Kaislahti Tillman et al., 2010a), which may imply a differing isotopic signal in precipitation. According to a model that predicts δ^{18}O in precipitation based only on latitude and altitude, δ^{18}O would be about −15.0‰ at Misaw Lake and −15.8‰ at Ennadai Lake for the June–July period (Bowen and Wilkinson, 2002). It has been proposed that moisture source regions (and with that, δ^{18}O in precipitation) changed significantly during the Holocene (Edwards et al., 1996). Such changes and shifts of

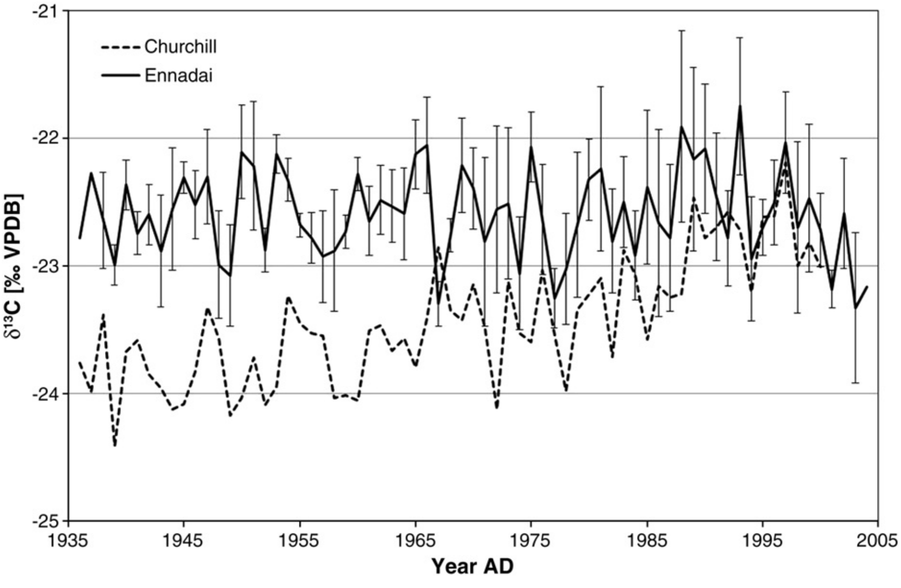

Figure 6. Comparison of tree-ring δ^{13}C values (corrected) near Ennadai Lake (mean values from this study, with standard deviations) and near Churchill (Tardif et al., 2008).

the arctic front are feasible even on shorter time scales, so that the average offset of $\delta^{18}O$ between the two study sites cannot be regarded as stable over time.

Conclusions

The comparison of climate data from stations across central Canada suggests that surface air temperature from a single site can be used to draw conclusions about past temperature variability over a large region, whereas precipitation patterns feature more spatial heterogeneity. Thus, the robust calibration of precipitation proxy data cannot be carried out using remote station data. Furthermore, since the stable isotope time series are influenced by both temperature and precipitation, it remains a challenge to disentangle temperature and precipitation signals in paleoclimate records. This points to the need for the analysis of multiple climate proxies from individual peat records, including testate amoeba, plant macrofossils, stable isotopes, and proxies for decomposition (Chambers and Charman, 2004; Loisel and Garneau, 2010), and the synthesis of different archives from the region (e.g., tree rings, peats, and lake sediments).

This study highlights the possibility of combining stable carbon and oxygen isotope data from tree rings and peat, allowing the integration of both high- and low-frequency climatic changes in a single, long record. The isotopic record of peat tracks long-term environmental changes, while tree rings serve as a proxy for high-frequency variability. The prospect is that the combination of both archives may compensate for the main shortcomings of the single archives (i.e., low temporal resolution of peat records and biased low frequency signal in tree rings). Our results show that $\delta^{18}O$ and $\delta^{13}C$ data from peat and tree rings have slightly different response periods, and somewhat varying sensitivities to climatic and environmental conditions. The use of both tree rings and peat will therefore lead to a better understanding of climatic variability and seasonality than if just a single archive were analyzed. This approach may also improve the continuity of peat records, as gaps caused by fire or rootlet layers in peat may be filled by tree-ring proxies. In the future, additional and longer isotope time series need to be developed and analyzed, for example from subfossil wood in lakes and along rivers, so that reconstructions can be extended beyond the short period of observational data. Furthermore, it seems advisable to sample peat cores at the highest possible resolution (realistically 1 cm), and to improve the dating of the upper parts of the records by applying ^{210}Pb, ^{137}Cs, and post-bomb ^{14}C analyses, thus facilitating well-constrained calibration with observed climate data.

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