The Holocene 13,6 (2003) pp. 801–808

Carbon sequestration in western Canadian peat highly sensitive to Holocene wet-dry climate cycles at millennial timescales

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Received 21 August 2002; revised manuscript accepted 13 April 2003

A HOLOCENE FAST-TRACK REPORT

Abstract: A high-resolution fen peat record and 79 basal peat dates from paludified peatlands in continental western Canada provide evidence for cyclic change in moisture conditions and in peat carbon accumulation. The ash-free bulk density, a proxy for degree of peat decomposition and thus moisture conditions, shows periodicities at both millennial (from 1500 to 2190 yr, with a mean of 1785 yr) and century scales (386 yr and 667 yr). Wet periods of 200–600 yr in duration, especially at ~6900, 5500 and 4000 cal. BP, correlate with rapid peat accumulation, new peatland initiation and declines in the rate of increase of atmospheric CO_2 concentrations. The wet periods in western North America are coeval with warm periods in the North Atlantic, a phasing relationship that has been documented in other published palaeorecords for the glacial period and late Holocene, probably in response to variations in solar activity. These results indicate a strong connection between climate and the global carbon cycle at the millennial scale, mediated in part by peatland dynamics. This is the first demonstration that peatland carbon sequestration rates are highly sensitive even to minor climatic fluctuations, which are too small to produce detectable changes in major species in the peatland. That global atmospheric CO_2 concentrations have in the past responded to these changes in peatland dynamics implies a strong potential for peatlands to be a major player in affecting future global change.

Key words: Fen peat, climate cycle, millennial scale, carbon sequestration, sensitivity, teleconnection, *Scorpidium scorpioides*, Holocene, Canada.

Introduction

Climate interacts with various components of the global carbon cycle at different timescales. This interaction has been illustrated by the effect of air temperature on the amplitude and timing of the seasonal cycle of high-latitude atmospheric CO_2 during the historical time (Keeling *et al*., 1996) and by the close associations of climate and atmospheric CO_2 concentrations during the glacial–interglacial cycles (Petit *et al.*, 1999) and in the last glacial period (Stauffer *et al*., 1998). Recent palaeoclimate records indicate that Holocene climate has also varied with regularity at millennial scales (O'Brien *et al.*, 1995; Mayewski *et al*., 1997; Bond *et al.*, 1997; 2001). The high-resolution atmospheric CO_2 record from Taylor Dome in Antarctica indicates significant variations in the

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10.1191/0959683603hl667ft

global carbon cycle during the Holocene (Indermühle *et al*., 1999). Northern peatlands are one of the largest terrestrial carbon reservoirs and have accumulated up to ~400 Gt ($\times 10^{15}$ g) of carbon during the Holocene (Gorham, 1991; Clymo *et al*., 1998; Roulet, 2000). Understanding their dynamics and climate sensitivity is key to understanding feedbacks to global change. Here we use the first well-dated high-resolution fen peat record (Yu *et al*., 2003) and 79 basal peat dates (Halsey *et al.,* 1998) from western Canada to investigate connections of Holocene climatic variability, vertical peat accumulation, regional peatland initiation (Campbell *et al.,* 2000) and the global carbon cycle.

The Upper Pinto Fen (UPF; unofficial name) is located near the northern edge of the northern Great Plains on the eastern slope of the Rocky Mountain Foothills (Figure 1A; 53°35’ N, 118°01’ W; elevation ~1310 m). It is an extremely rich fen dominated by brown mosses. The region has a semi-humid continental climate with an annual precipitation of ~550 mm. Its headwater position and small surface catchment area (~4 km^2; Figure 1C) is likely to make the site sensitive to changes in regional climate, especially available moisture.

Peat core stratigraphy and climate proxy

The chronology of a 397 cm peat core from the UPF was controlled by 19 (plus one post-bomb) accelerator mass spectrometry (AMS) ^{14}C dates from 1 cm slices taken mostly at 20 cm intervals (Table 1; Yu *et al*., 2003). The age model is based on linear interpolation between pairs of ages calibrated with the INTCAL98 data set (Stuiver *et al*., 1998; Figure 2A). The temporal sampling resolution ranges from ~3 to 60 years for each contiguous 1 cm interval. Ash-free bulk density shows a pattern of cyclic change between ~0.08 and 0.16 g/cm^3 (Figure 2B). Macrofossil results show that the site was paludified from a spruce-dominated forest. From 6500 to 1300 cal. BP, the site was dominated by the moss *Scorpidium scorpioides*, succeeded from a diverse moss-*Larix* phase and followed by a *Tomenthypnum nitens*-*Larix* fen (Yu *et al*., 2003).

The state of preservation of moss macrofossils such as *Scorpidium* (Figure 2D) and the abundance of unrecognizable debris (Figure 2C) indicate the degree of peat decomposition. Our results show that high bulk density corresponds with highly decomposed peat (up to 90% unrecognizable debris), whereas low density corresponds with well-preserved bryophytes (as little as only 10% debris). As a characteristic species of rich fens, *Scorpidium scorpioides* is most common at or just below the water surface, rarely occurring more than 10 cm above the water table (Gignac *et al.,* 1991; Vitt *et al.,* 1993). The productivity of *S. scorpioides* is sensitive to moisture conditions: growth rates are high with high water levels but strongly reduced during prolonged summer drought (Kooijman and Whilde, 1993). Moist conditions cause high moss production and decreased decomposition (anaerobic condition), resulting in low bulk density. Thus the peat bulk density in our record is interpreted as reflecting local moisture conditions on the peatland surface, with high density corresponding

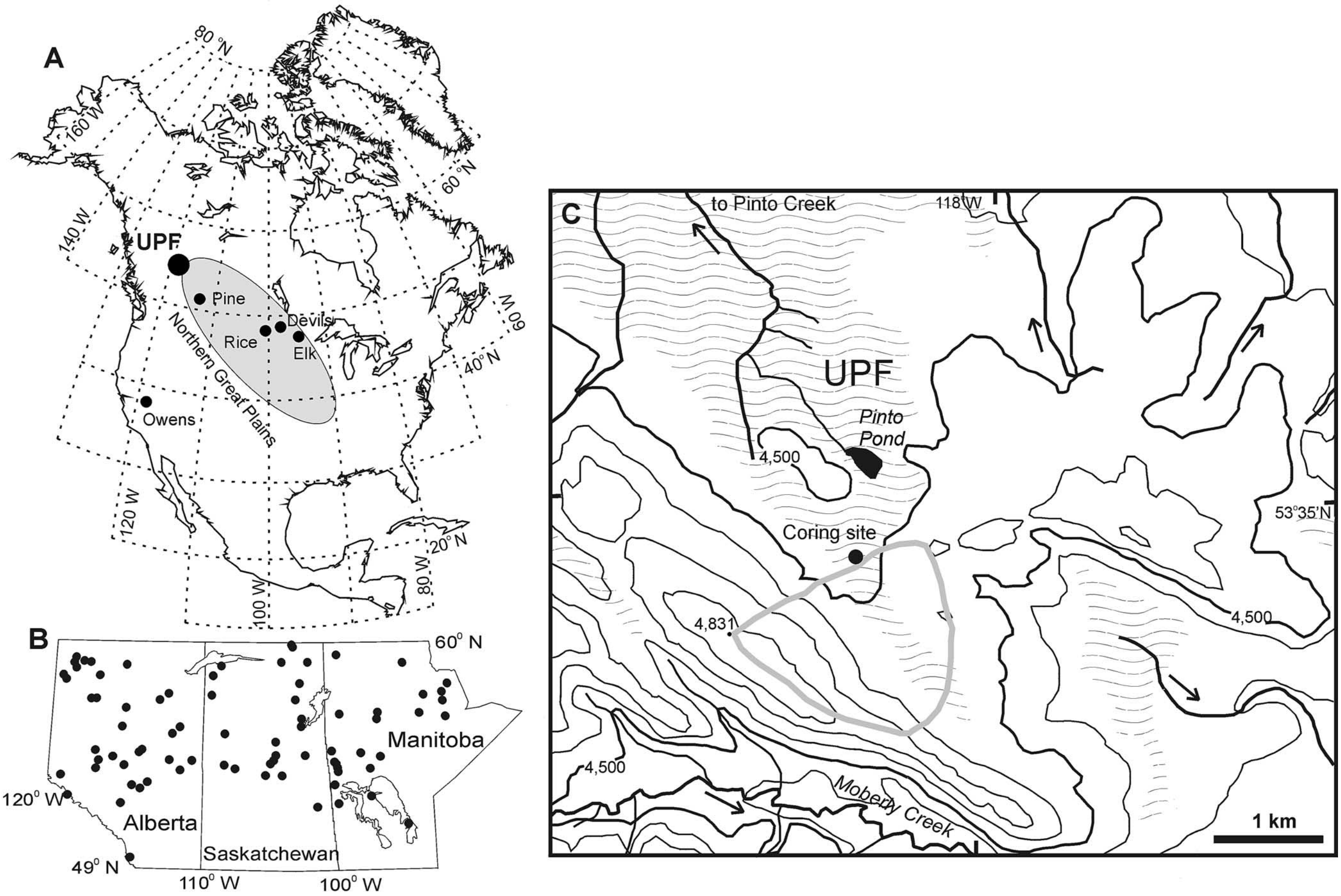

Figure 1 Location maps. (A) Location of the Upper Pinto Fen (UPF) and other selected sites in central North America: Pine Lake, AB (Campbell *et al*., 1998); Rice Lake, ND (Yu and Ito, 1999; Yu *et al.,* 2002); Devils Lake, ND (Fritz *et al*., 1991; Haskell *et al*., 1996); Elk Lake, MN (Dean, 1997); and Owens Lake, CA (Benson *et al.,* 1997). The shaded area shows approximate extension of the northern Great Plains. (B) Locations of 79 paludified peatlands in western Canada (based on maps in Halsey *et al.,* 1998, and Campbell *et al*., 2000). (C) Detailed topographic map of the UPF peatland, showing its headwater location with a small surface catchment area (grey line). Hatched areas are peatlands. Contours are at 100 ft (= ~30 m). The coring site is close to the 4500 ft contour line at the edge of the water-shedding topographic plateau.

Table 1 AMS radiocarbon dates from the Upper Pinto Fen, Alberta, Canada (from Yu *et al*., 2003)

Depth (cm)	Lab. number	^{14}C date ± 1SD (yr BP)	Age range at 2 σ (95%)
24–25	AA37423	Post-bomb 1.3169 fraction of modern ^{14}C	
44–45	AA37428	305 ± 40	462–297
64–65	AA37108	975 ± 40	955–789
84–85	AA37429	1315 ± 40	1295–1174
104–105	AA37104	1630 ± 40	1611–1413
109–110	Beta-144856	2060 ± 40	2130–1920
118–119	Beta-144857	2320 ± 40	2360–2315
124–125	AA37419	2620 ± 45	2848–2707
144–145	AA37105	3125 ± 45	3450–3242
164–165	AA37420	3775 ± 60	4308–3974
189–190	Beta-135661	3850 ± 40	4365–4152
214–215	AA37421	4270 ± 50	4968–4803
234–235	AA37422	4600 ± 50	5467–5249
254–255	AA37106	4690 ± 50	5486–5314
274–275	AA37424	5035 ± 50	5899–5661
294–295	AA37425	5325 ± 50	6201–5985
314–315	AA37426	5750 ± 50	6662–6438
334–335	AA37107	5990 ± 50	6945–6720
364–365	AA37427	6075 ± 55	7029–6783
384–385	Beta-135662	6750 ± 40	7670–7568

to drier conditions (Figure 3). The single species *S. scorpioides* dominates most of the record, suggesting that the pattern in observed bulk density and inferred local moisture conditions was not associated with hummock-hollow species cycles, as seen in some peatlands (Aaby, 1975), nor with change in botanical composition of peat (Barber *et al*., 1994). Instead, the variations were probably caused by changes in local hydrology responding to regional climatic variations.

A spectral analysis of the ash-free bulk density was performed; the extrapolated 8066-yr-long time series was first detrended by linear subtraction of fitted mean values of the whole series. The detrended bulk-density time series was directly analysed without interpolation, as the interpolation procedures usually lead to an underestimation of high-frequency components (Schulz and Stattegger, 1997), using the Lomb-Scargle algorithm for unevenly spaced data in AutoSignal v.1.0 (by the SPSS) and SPECTRUM (Schulz and Stattegger, 1997). Both methods produced similar results. Significant peaks occur in a broad band between 1500 and 2190 years (average 1785 yr) above 99% significance level and at 386 and 667 years above 90% level, with the 1500-yr peak having the highest spectral power (Figure 4). The 1500–2190-year cycles are visible in the graph (Figure 5B), as indicated by low bulk-density values, and thus wet periods (from UPF events 5 to 1 and post-LIA). The middle points of these wet periods are at 6900, 5500, 4000, 2500, 1000 and 150 cal. BP, with a variable duration ranging from 200 to 600 yr.

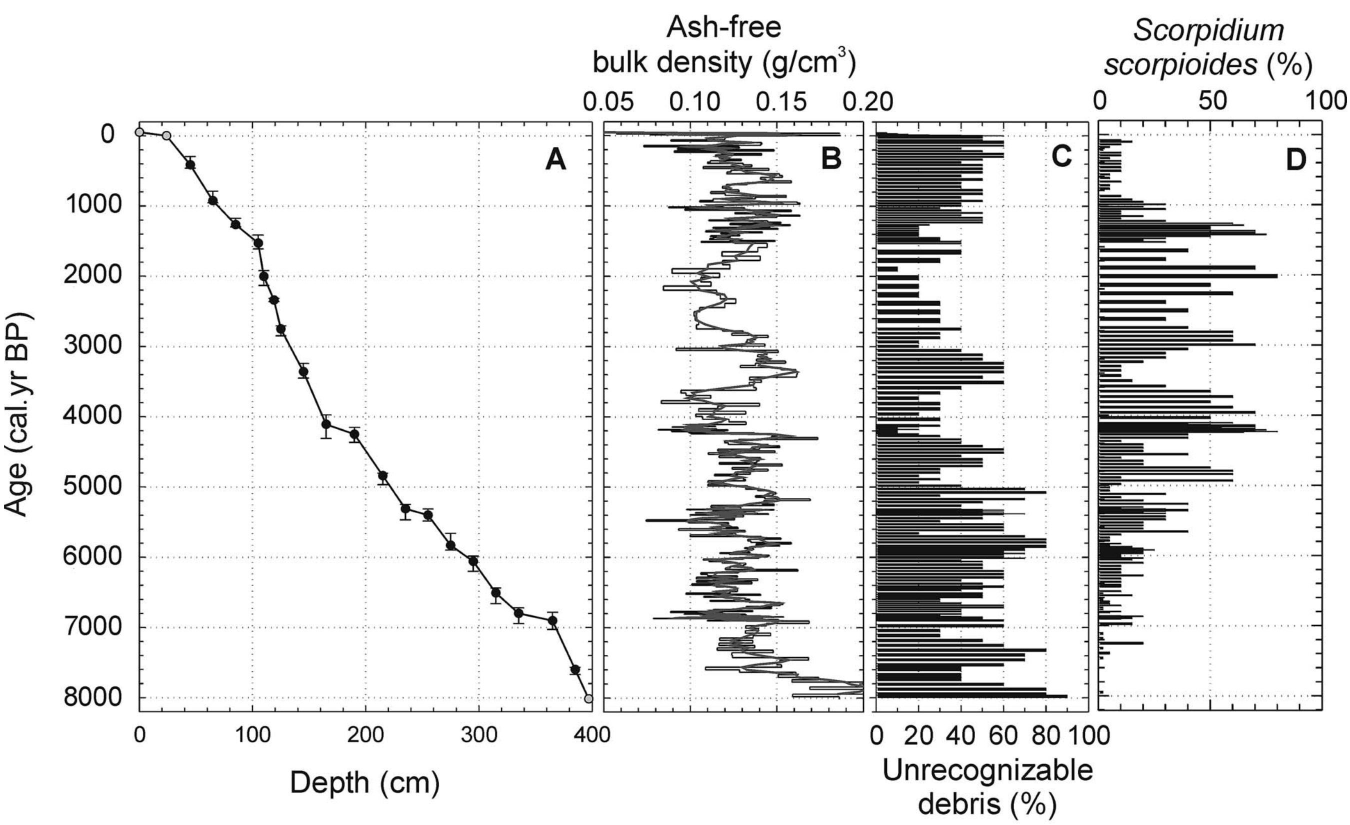

Figure 2 Palaeoecological records from the UPF core. (A) Age-depth model. The error range (2 σ) is represented by vertical bars. The top two grey circles represent the peat surface (–49 cal. BP) and a post-bomb date (assuming 0 cal. BP), respectively. (B) Ash-free bulk density (bars are raw data measured at 1 cm intervals, and heavy curve is simple three-point running mean). The loss-on-ignition at 550°C was used to estimate organic content. The ash-free bulk density was calculated from the measurements of sample volume, dry weight and organic contents. (C) Relative abundance of unrecognizable/decomposed debris. (D) Relative abundance of brown moss *Scorpidium scorpioides*. A semi-quantitative method was used to estimate relative abundance of different macroscopic components on a volume basis, using ~1 cm^3 samples of peat. We first estimated the decomposed/unidentifiable debris and then subdivided the remaining sample to various identifiable components.

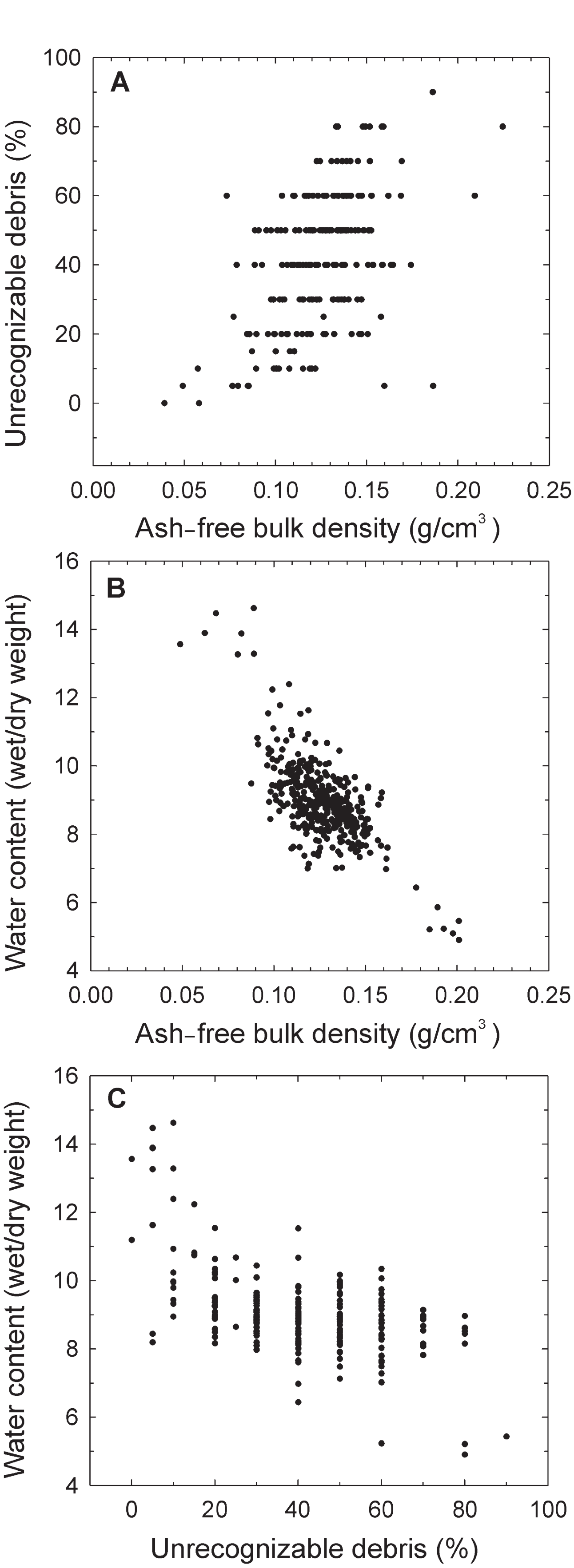

Figure 3 Proxies of peatland moistures at the UPF core. (A) Scatter plot of ash-free bulk density and unrecognizable debris. (B) Scatter plot of ash-free bulk density and water content (wet sediment weight/dry sediment weight). (C) Scatter plot of unrecognizable debris and water content.

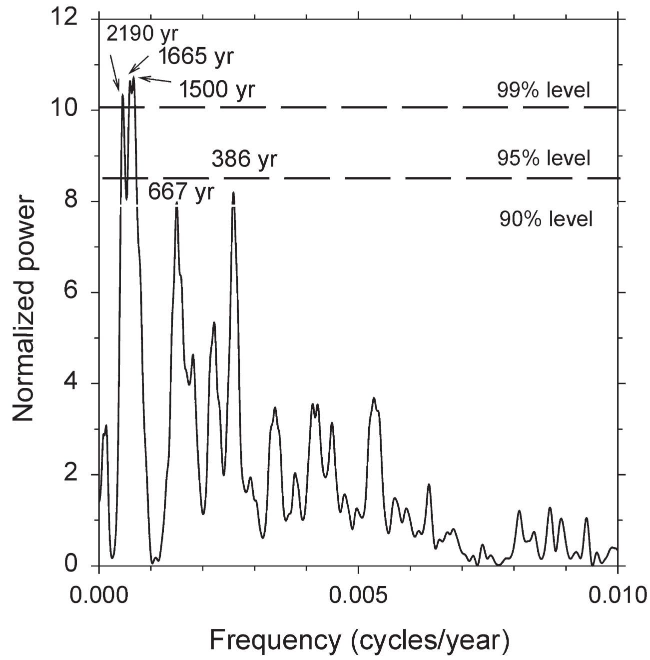

Figure 4 Power spectrum of ash-free bulk density time series from the UPF core. The spectral analysis was performed using the Lomb-Scargle algorithm for unevenly spaced data in AutoSignal program (version 1.0) by the SPSS. The Lomb-Scargle periodogram was generated by setting to fast algorithm and using a cs2 Hamming data window, with a spectrum size of 8192.

Peat accumulation and peatland initiation

We calculated the long-term mass-accumulation rates of peat on the basis of paired dating points at the UPF. The calculated rates show three prominent peaks at ~6900, 5400 and 4200 cal. BP (Figures 5C and 6), corresponding to the three earlier wet periods (Figure 5B). This suggests that peak accumulation rates were caused by high plant production or low decomposition under moist condition. We recalculated the peat-accumulation rates considering the probability distribution of calibrated radiocarbon ages. For all dates within 2 σ (95% probability) of calibrated ages, the peak accumulation rates are always at least double the values of the adjacent periods. Therefore the high accumulation rates observed during these wet periods are real features (>95% probability).

The basal dates from paludified peatlands in continental western Canada (Halsey *et al.*, 1998; see Figure 1B for distributions of these peatlands) suggest that Holocene peatland initiation followed the same regularity at a millennial timescale (Campbell *et al.*, 2000; Figure 5D). Most peatlands were evidently initiated during wet periods, especially at ~7000, 5200 and 3800 cal. BP, with ~400–500-yr timelag after the beginning of UPF wet events 4 and 3 at 5600 and 4200 cal. BP (Figure 5B). This lag may relate to the tendency of incorporating younger peat into the compacted basal samples for dating. The peatland-initiation pattern does not appear to be related to any spatial sampling bias, with dates from any given region spanning a wide range of values (Figure 1B; Halsey *et al.*, 1998).

The apparent absence of accumulation peaks (Figure 5C) and of basal dates younger than ~3000 cal. BP (Figure 5D), especially for the strong UPF event 2 around 2500 cal. BP, is perplexing. In the case of basal dates, this could be a simple matter of sampling bias, with few shallow peat sites having been dated (Campbell *et al.,* 2000). In the case of the UPF data, sampling bias is less likely, as two additional radiocarbon dates were obtained between 2750 and 1530 cal. BP (Figure 2A; Table 1)

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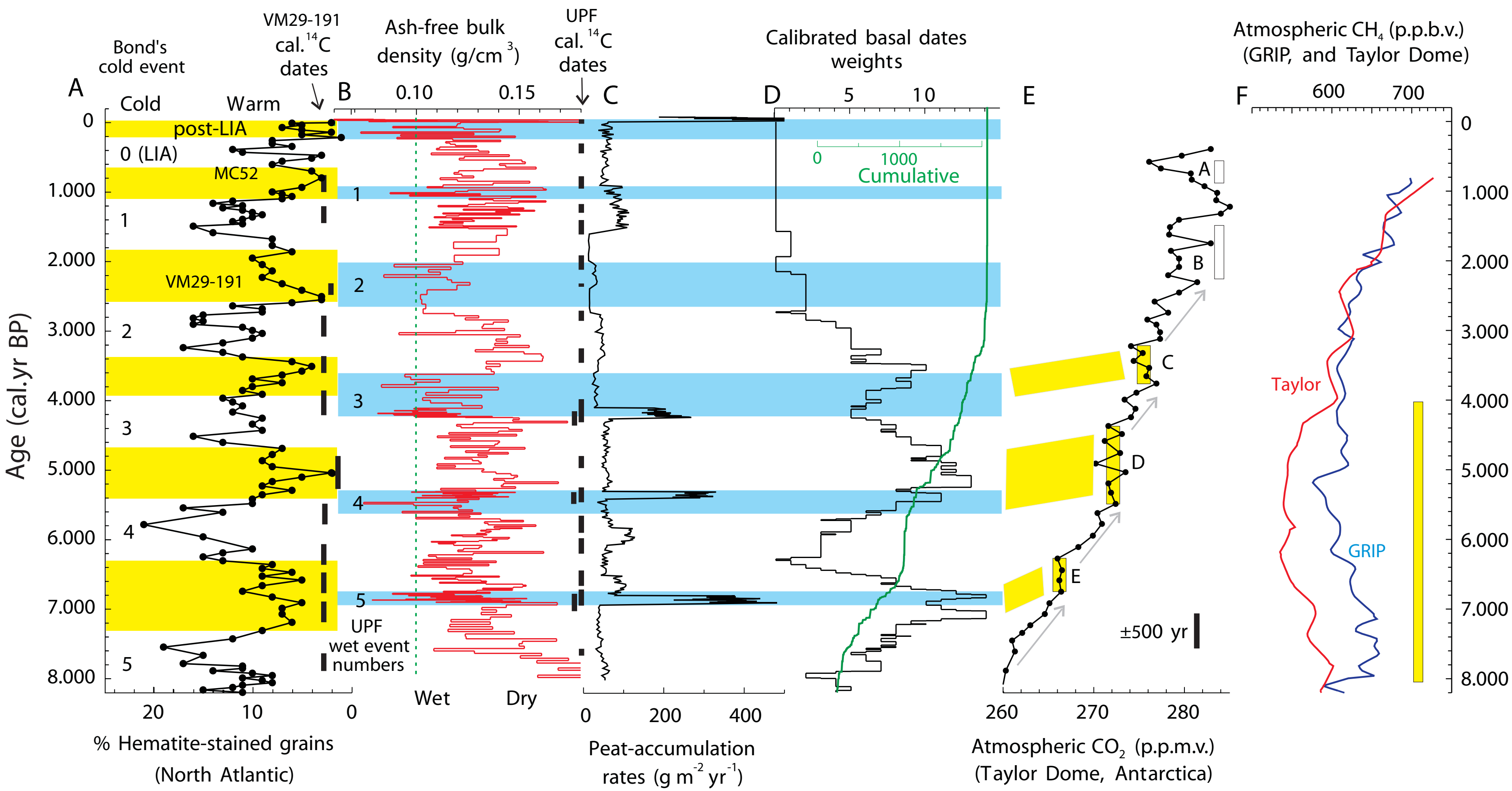

Figure 5 Climate, peatland and carbon-cycle correlation. (A) Percentage of hematite-stained grains of combined core MC52/VM29–191 from the North Atlantic (Bond *et al.*, 2001). Numbers show the cold events for the North Atlantic region, together with the 'Little Ice Age' (LIA). Rectangles on the right are locations and 2 σ ranges (180–480 years, with a mean of 300 years) of 12 calibrated ^{14}C dates (Stuiver *et al.*, 1998) from VM29–191 (Bond *et al.*, 2001). There might be an additional age error of $\pm$200 years owing to variable reservoir correction. Thus the correlation with the UPF record (yellow bands) is suggestive within the dating uncertainty. (B) Ash-free bulk density from the UPF core in central Alberta. The UPF wet events (shaded blue bands) were defined as the lowest ash-free density values (<0.1 g/cm^3 with >1 sample). Rectangles on the right are locations and 2 σ ranges (45–335 years, with a mean of 190 years) of calibrated ^{14}C dates (Stuiver *et al.*, 1998). (C) Peat-accumulation rates from the UPF core. Accumulation rates were calculated by multiplying ash-free bulk density (g/cm^3) and vertical peat-growth rates (cm/yr), assuming constant vertical growth rates between pairs of dates as in the age model (Figure 2A). (D) Weight of calibrated basal peat dates from 79 paludified peatlands in continental western Canada listed in Halsey *et al.* (1998) as a measure of probability of peatland initiation (Campbell *et al.*, 2000). The probability histogram was constructed by summing the number of calibrated date ranges including all individual years (Campbell *et al.*, 2000). Cumulative curve (green) can be used as a proxy for the increase of new peatland areas. See Halsey *et al.* (1998) for detailed information on location, peat depth and reference of each basal date. (E) Atmospheric CO_2 concentrations from Taylor Dome in Antarctica (Indermühle *et al.*, 1999). Rectangles indicate periods of decreases in the CO_2 rising rate, especially during the CO_2 plateaus E, D and C (yellow) shortly after the peat accumulation and peatland initiation peaks, and during the phase of rapid increase in new peatland area (C, D). The suggestive correlation with the UPF and western interior Canadian peatlands is shown as yellow bands. The uncertainty of the CO_2 chronology is $\pm$500 years (black bar on the low right) from 8000 to 1000 cal. yr BP, with analytical uncertainty of 1.5–3.0 p.p.m.v. in CO_2 concentrations (Indermühle *et al.*, 1999). (F) Atmospheric CH_4 concentrations from GRIP core in Greenland (Chappellaz *et al.,* 1997) and from Taylor Dome core in Antarctica (Brook *et al.*, 2000). The interpolar gradients indicate relative strengths of methane sources from tropical and northern wetlands. During 8000–4000 cal. yr BP, higher CH_4 concentration in Greenland than in Antarctica suggests more CH_4 from expanding northern peatlands.

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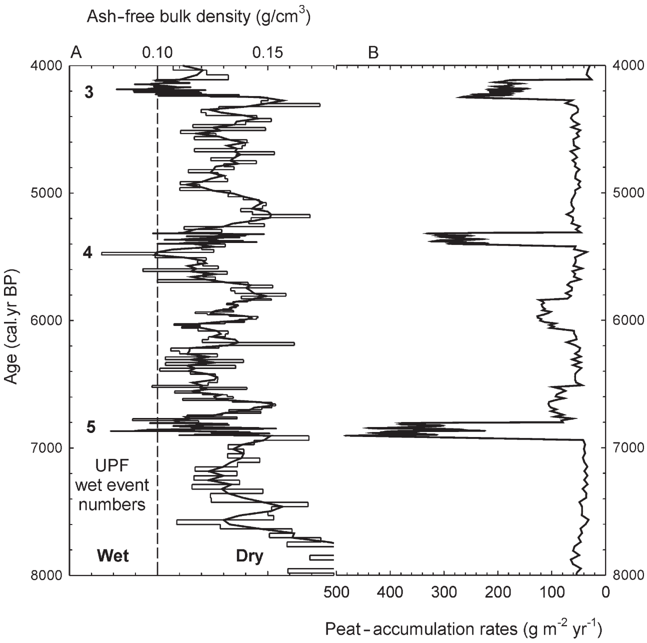

Figure 6 Close-up of ash-free bulk density and peat-accumulation rates for the period of 8000–4000 cal. yr BP at the UPF core. High peat accumulation is the result of high vertical peat increment rate, rather than high bulk density, so in a sense the two curves were derived independently.

specifically to determine if a sampling bias was present. It is more likely that, in both cases, a threshold in sensitivity had been crossed. The regional moistening trend since 6000 cal. BP (Schweger and Hickman, 1989) may have progressively reduced peatland sensitivity to modest climatic fluctuations. Also most sites suitable for peatland development by paludification may have been occupied by peatlands by 3000 cal. BP, as peatlands reached their southern limit by then (Halsey *et al*., 1998).

Holocene millennial-scale climate cycles and mechanisms

This millennial-scale climatic variability as indicated by the peat bulk density data at the UPF has also been documented at other sites in the northern Great Plains (see Figure 1A for site locations). These palaeoclimate records appear to show similar millennial-scale wet-dry cycles (Fritz *et al*., 1991; Haskell *et al.*, 1996; Dean, 1997; Campbell *et al*., 1998), though with different proxies and variable quality of dating control. At Pine Lake in central Alberta, a sediment grain-size record, reflecting regional moisture conditions, showed ~1450-yr wet-dry cycles in late-Holocene sediments (Campbell *et al.,* 1998). The Mg/Ca ratios from ostracode shells, as a proxy of lake salinity and thus regional effective moisture, at Devils Lake, North Dakota (Haskell *et al.*, 1996), showed a close match with the UPF moisture record, especially after 4000 cal. yr BP, both at millennial and multicentennial scales. Geochemical records from Elk Lake, northwestern Minnesota, showed three peaks of Al accumulation rates at 7740, 5900 and 4240 cal. BP (Dean, 1997; Keeling and Whorf, 2000), which recorded dust layers under dry conditions, occurring anywhere from 700 to 40 years before our three earlier Holocene wet periods (UPF wet events 5, 4 and 3; Figures 5B and 6), which started at 7000, 5600 and 4200 cal. BP. The ~400-yr drought cycles have been shown in several palaeoclimate records in the northern Great Plains (Dean, 1997; Yu and Ito, 1999).

Our results correlate in time pacing and phase with Holocene climate records from the North Atlantic, where the petrographic tracers showed millennial-scale variations during the Holocene, with high spectral power in a broad band centred at ~1800 years (Bond *et al*., 1997) or simply 1–2 kyr climate cycle (Bond *et al*., 1999). Among other proxies of drift ice (detrital carbonate, Icelandic volcanic glass), for example, high abundance of hematite-stained grains reflects increased ice-rafting associated with ocean-surface cooling events (Figure 5A; Bond *et al*., 1997; 2001). The wet periods in the UPF peat core appear to correlate with these petrographic records, all occurring during the warm (UPF-5 and -2) or warming phases (UPF-4, -3 and -1) immediately after the Bond's cold events 5, 4, 3, 2, 1 and the 'Little Ice Age' (LIA = event 0; Figure 5, A and B). If we accept the chronologies at their face values in both records, it implies that for some peatland wet periods either there is a timelag from the warm periods in the North Atlantic (e.g., UPF-5) or peatland events occur at times of cold-to-warm transitions in the North Atlantic (e.g., UPF-4, -3 and -1). However, these marine cores have a dating uncertainty of ±200 years in oceanic reservoir correction on top of calibration uncertainty of 300 years (ranging from 180 to 480 years; Figure 5A), whereas the UPF record has a calibration uncertainty of 190 years, with a range of 45–335 years (Figure 5B). Given these dating errors, the peak warm periods in the North Atlantic may match the peak wet periods in Alberta. The most recent UPF wet period since ~200 cal. BP immediately followed the LIA, the most recent cold phase of the Holocene 1–2 kyr cycle in the North Atlantic as recorded in a high-resolution ocean core (Bond *et al*., 1999; 2001). A similar teleconnection (cold periods in Greenland corresponding with dry periods in western North America) has been documented in the hydrologic balance of the Owens basin, California, at millennial scales during the last glacial termination (Benson *et al*., 1997) and of Rice Lake, North Dakota, at centennial scales for the last 2100 years (Yu and Ito, 1999; Yu *et al*., 2002).

Although the nature of the forcing mechanism for the millennial cycles is still under debate, various ideas have been proposed. For example, Bond *et al*. (1999) suggested that these events were associated with expansion of the subpolar cyclonic gyre in the North Atlantic, possibly caused by shifts in the North Atlantic's thermohaline circulation (Bianchi and McCave, 1999). Broecker *et al*. (1999) argued that they might be a Holocene manifestation of a bipolar seesaw phenomenon in the North Atlantic. Keeling and Whorf (2000) proposed that a tidal forcing mechanism could explain the broad 1800-yr climate cycle by increasing vertical mixing in the oceans to cause cooling at the sea surface. Alley *et al.* (2001) argued that a stochastic resonance hypothesis could explain these millennial-scale changes, in which switches between warm and cold climate modes were produced by a combination of a weak periodicity signal and stochastic processes (noise). Recently, Bond *et al*. (2001) have found a close correlation between solar proxy (production rates of cosmogenic nuclides ^{14}C and ^{10}Be) and changes in proxies of drift ice from the North Atlantic, and proposed that atmospheric changes induced by variations in solar irradiance were amplified and transmitted through their impact on sea ice and North Atlantic thermohaline circulation.

In any case, simulations indicate that cooling of the North Atlantic, such as occurred during the Younger Dryas, could result in cooling of the North Pacific (Mikolajewics *et al*., 1997), which in turn could decrease the temperature and moisture content of air passing over the middle latitudes, leading to a drier climate in interior North America (Benson *et al.,* 1997; Peteet *et al*., 1997). Benson *et al.* (1997) suggested that linkage of the climatic regimes of the North Atlantic and western North America weakened during the Holocene and perhaps disappeared. However, our observations suggest that this teleconnection mechanism operated during the Holocene. Alternatively, both western North America and North Atlantic might have responded independently to common solar forcing during the Holocene. GCM simulation (Rind and Overpeck, 1993), together with palaeorecords for solar–

drought linkage in the northern Great Plains during the late Holocene (Yu and Ito, 1999), suggested that the continental interior is more sensitive to small changes in the solar irradiance.

Impact of peatland dynamics on the global carbon cycle

Peatland dynamics in continental western Canada have significantly affected the global carbon cycle during the Holocene as suggested by stepwise increase of atmospheric CO_2 concentrations (Indermühle *et al.*, 1999; Smith *et al.*, 1999). The CO_2 record from Taylor Dome, Antarctica, shows a 25 p.p.m.v. rising trend from 8000 to 1000 cal. BP (Figure 5E), attributed by Indermühle *et al.* (1999) to cumulative biospheric release of 195 Gt of carbon, as constrained from CO_2 carbon-isotope data. During the same time period, however, northern peatlands have taken up ~400 Gt of carbon (Gorham, 1991; Roulet, 2000), of which ~50 Gt are from continental western Canada (Vitt *et al.*, 2000), the same region as the peatland initiation data (Halsey *et al.,* 1998; Figure 5D). Three accumulation and initiation peaks at *c.* 7000, 5400 and 4000 cal. BP were followed by declines in the rising rates of CO_2 concentrations (CO_2 plateau E, D and C; Figure 5E), suggesting detectable impacts of peatlands on the global carbon cycle by reducing the CO_2 rising rate, whatever are the causes of that rise (e.g., Broecker *et al.*, 1999; Beerling, 2000). Discussion of possible timelags in atmospheric CO_2 response is hampered by the large dating uncertainty of $\pm$500 years in CO_2 record owing to synchronization by CH_4 with Greenland ice cores (Indermühle *et al.,* 1999). Interestingly, atmospheric CO_2 concentrations reached the pre-industrial level of 280 p.p.m.v. between ~3000 and 1000 cal. BP with no obvious CO_2 plateau (B and A; Figure 5E), while there was no peat-accumulation or initiation peak. The interpolar gradients in atmospheric methane (CH_4) concentration also suggest diminishing role of northern peatlands as dominant sources of CH_4 after 4000 cal. yr BP (Figure 5F).

If the high-resolution carbon-isotope data of CO_2 collected in the future from polar ice cores or from fossil stomatal records (e.g., Rundgren and Beerling, 1999) confirm the findings of ~200 Gt biospheric C release by Indermühle *et al.* (1999), other biomes in the world must have released >600 Gt C in order to account for the carbon uptake of ~400 Gt by northern peatlands from 8000 to 1000 cal. BP. Recent estimates of changes in terrestrial carbon reservoirs since the last glacial maximum from palaeoecological records and modelling exercises (Adams and Faure, 1998; Beerling, 2000) do not support the hypothesis that terrestrial biosphere released that large amount of carbon during the mid- and late Holocene. If so, oceans must have played a major role in controlling atmospheric CO_2 concentration. Clearly, further studies are needed to understand Holocene carbon cycle dynamics, and a full explanation of the relative importance of oceans and terrestrial biota would require a coupled ocean–biosphere–atmosphere biogeochemical model.

Conclusion and implications

These results indicate that during the earlier part of the Holocene small changes in moisture conditions resulted in a large shift in the carbon source-sink relations of a continental fen in western Canada. This response occurred in the absence of any apparent change in dominant species on the fen, indicating that peat-accumulation rates are more sensitive to minor climatic fluctuations than is species composition and that future climatic change could significantly reduce peat carbon sequestration rates without any telltale changes in the species composition of peatland vegetation. These changes in peatland carbon dynamics have influenced global atmospheric CO_2 concentration, implying that projected future climatic warming and especially drying of peatlands in some regions may significantly affect peat carbon sequestration, producing a potential impact on global warming.

Acknowledgements

We thank N.L. Cleavitt, R.C. Errington, L.A. Halsey and M.R. Turetsky for field and laboratory assistance; and J.M. Adams, P.H. Glaser, E.T. Hogg, F.S. Hu, J.T. Overpeck, D.T. Price, P.J.H. Richard, N.T. Roulet, T. Webb III and H.E. Wright Jr for comments and suggestions. This work was initiated while the senior author was at the University of Alberta and Canadian Forest Service (Edmonton, Alberta), supported by the Climate Change Action Fund of Canada.

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