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Periglacial fires and trees in a continental setting of Central Canada, Upper Pleistocene

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ABSTRACT

Fire is a key factor controlling global vegetation patterns and carbon cycling. It mostly occurs under warm periods during which fuel builds up with sufficient moisture, whereas such conditions stimulate fire ignition and spread. Biomass burning increased globally with warming periods since the last glacial era. Data confirming periglacial fires during glacial periods are very sparse because such climates are likely too cold to favour fires. Here, tree occurrence and fires during the Upper Pleistocene glacial periods in Central Canada are inferred from botanical identification and calibrated radiocarbon dates of charcoal fragments. Charcoal fragments were archived in sandy dunes of central Saskatchewan and were dated >50 000–26 600 cal BP. Fragments were mostly gymnosperms. Parallels between radiocarbon dates and GISP2-$\delta^{18}O$ records deciphered relationships between fire and climate. Fires occurred either hundreds to thousands of years after Dansgaard–Oeschger (DO) interstadial warming events (i.e., the time needed to build enough fuel for fire ignition and spread) or at the onset of the DO event. The chronological uncertainties result from the dated material not precisely matching the fires and from the low residual ^{14}C associated with old sample material. Dominance of high-pressure systems and low effective moisture during post-DO coolings likely triggered flammable periglacial ecosystems, while lower moisture and the relative abundance of fuel overshadowed lower temperatures for fire spread. Laurentide ice sheet (LIS) limits during DO events are difficult to assess in Central Canada due to sparse radiocarbon dates. Our radiocarbon data set constrains the extent of LIS. Central Saskatchewan was not covered by LIS throughout the Upper Pleistocene and was not a continental desert. Instead, our results suggest long-lasting periods where fluctuations of the northern tree limits and fires after interstadials occurred persistently.

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INTRODUCTION

Fire is a key component that occurs mostly during warm periods and closely controls carbon cycling (Bowman *et al.*, 2009) and global vegetation (Bond *et al.*, 2005). If moisture is sufficient, warm climates stimulate fuel build-up and in turn, fire ignition and spread. Since the last glacial era, biomass burning has therefore increased globally with warming periods (Daniau *et al.*, 2012). Periglacial climates are generally thought to be too cold to favour fires. There are a few reports of fire activity in modern shrubby tundra (Payette *et al.*, 1989; Higuera *et al.*, 2008; Hu *et al.*, 2010; Mack *et al.*, 2011) and in periglacial environments during glacial times (Wang *et al.*, 2005). Overall, however, few studies (Daniau *et al.*, 2010) have examined the response of fire to climate variability across glacial periods. Fire modelling during the LGM indicated a slightly lower

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biomass burning activity compared to interglacials, resulting in reduced greenhouse gas (GHG) fluxes towards the atmosphere (Thonicke *et al.*, 2005). The model is calibrated with modern vegetation processes, but does not consider the vegetation distribution at biome edges, especially in the tundra–taiga transition. Trees might have survived within tundras or cold steppes during glacial times (Kullman, 2002; Brubaker *et al.*, 2005). Combined with soil organic matter, these scattered trees could have offered fuel for ignition and fire spread. Periglacial fires are currently uncommon, characterized by long fire return intervals and cycles (Payette *et al.*, 1989), but could become a large source of GHG (Mack *et al.*, 2011; Schuur *et al.*, 2013). Future climate could also support more fires in periglacial environments (Hu *et al.*, 2010) because of woody plant expansion into the tundra (Sturm *et al.*, 2001; Naito & Cairns, 2011). Increased GHG fluxes to the atmosphere from periglacial fires could lead to a positive feedback in fire activity, especially at high latitudes where warming is amplified.

During the Last Glacial Maximum (LGM), 26 500–19 000 cal BP (Clark *et al.*, 2009), the Laurentide and Cordillera ice sheets covered most of northern North America and eventually coalesced in the southern Canadian Rockies and, despite some uncertainties, in northern areas as well (Dyke *et al.*, 2002). Dating of erratics that formed along the coalescence of the two ice sheets (i.e. Foothills Erratics Train) supports that the deposition occurred during the Late Wisconsinian, close to the LGM (Jackson *et al.*, 1997). Late Pleistocene climatic records indicate that large temperature changes at about 100 000, 41 000 and 23 000 years BP were caused by orbital variation in insolation (Imbrie *et al.*, 1984). High-resolution paleoclimatic records from the circum-North Atlantic region also indicate many abrupt climatic oscillations during marine isotope stage 3 (MIS 3, ca. 59 000 to 30 000 years BP). The rapid and significant warming events in Greenland and the North Atlantic region [known as the Dansgaard–Oeschger (DO) warming events (Dansgaard *et al.*, 1993; Huber *et al.*, 2006)] were initially associated with a thermohaline circulation (THC) moving from a weak (stadials) to a strong (interstadials) state (Broecker, 2006). Within some stadials, very large volumes of ice from the Laurentide ice sheet (LIS) surged into the Atlantic Ocean, referred as the Heinrich (H) events (Heinrich, 1988; Hemming, 2004). Despite that the causes of the millennial-scale climatic oscillations during MIS 3 are not yet fully elucidated (Broecker, 2006; Clement & Peterson, 2008), they had a definite impact on the extent and timing of ice advance or recession. Dyke *et al.* (2002) suggested that interstadials prior to the LGM left the southern parts of the Canadian prairies ice-free. However, there is still considerable uncertainty in regard to the extent and timing of ice advance and recession during MIS 3 in North America due to a geographically sparse database of ^{14}C dates (Burns, 1996; Dyke *et al.*, 2001) and the lack of studies trying to relate the ^{14}C dates to climatic records from ice cores.

The main objective of the study was to document fire disturbance of ecosystems from central Saskatchewan, Canada, using ^{14}C dating and microscope identifications of paleosoil charcoal in dunes (up to a depth of 2 m) at five different locations. Based on these data, we tested the following two hypotheses: (i) the extent of recession of the southwestern LIS during glacial times in Central Canada was significant enough for the development of climatic conditions that were conducive to the establishment of needleleaf trees as well as the build-up of sufficient fuels to support fires on the landscape; and (ii) fires were well timed with DO warming events and ice recession episodes, although perhaps slightly asynchronous because of a delay necessary for tree establishment and thus fuel build-up.

MATERIALS AND METHODS

The study area (54°40′53″N, 105°29′55″W, coordinates for soil profile 1) is located in the mid-boreal upland of the boreal plain ecozone, approximately 20 km northeast of the northern edge of Montreal Lake (Fig. 1). The area investigated was relatively small, covering about 5 km^2. The topography of the area is relatively flat with occasional hummocky terrains (460–490 m asl). The climate is characterized by mean monthly temperatures ranging from −19.7 °C in January to 16.2 °C in July and mean annual precipitations of 467 mm (National Climate Data and Information Archive, http://www.climate.weatheroffice.gc.ca/). Approximately 30% of precipitations fall as snow, which covers the ground from November to April. The local dominant trees are jack pine (*Pinus banksiana*), which is largely associated with Eutric Brunisols/Cambisols (or Fluvaquentic Eutropepts) developed from very sandy and excessively drained fluvioglacial parent material with low carbon levels except for layers with charcoal. Trembling aspen (*Populus tremuloides*), white spruce (*Picea glauca*) and black spruce (*Picea mariana*) are also found in the area but generally are associated with finer (silt and clay bands) and/or moderately well-drained glaciolacustrine parent material. The understory vegetation is typically bearberry (*Arctostaphyllos uva-ursi*) and reindeer lichens (*Cladina* sp.). There is no earthworm activity in these soils due to the relatively low pH and the remoteness from agriculture. Tree ring analyses and the topsoil charcoal abundance indicate that the sampled stands regenerated after fire about 45 years ago.

Five paleosoils containing buried charcoal were investigated in the Montreal Lake area (Fig. 2). The profiles were archived in sand dunes and were undisturbed by erosion, pedoturbation by freeze-thaw cycles or burrowing animals, suggesting that charcoal fragments were time-stratified.

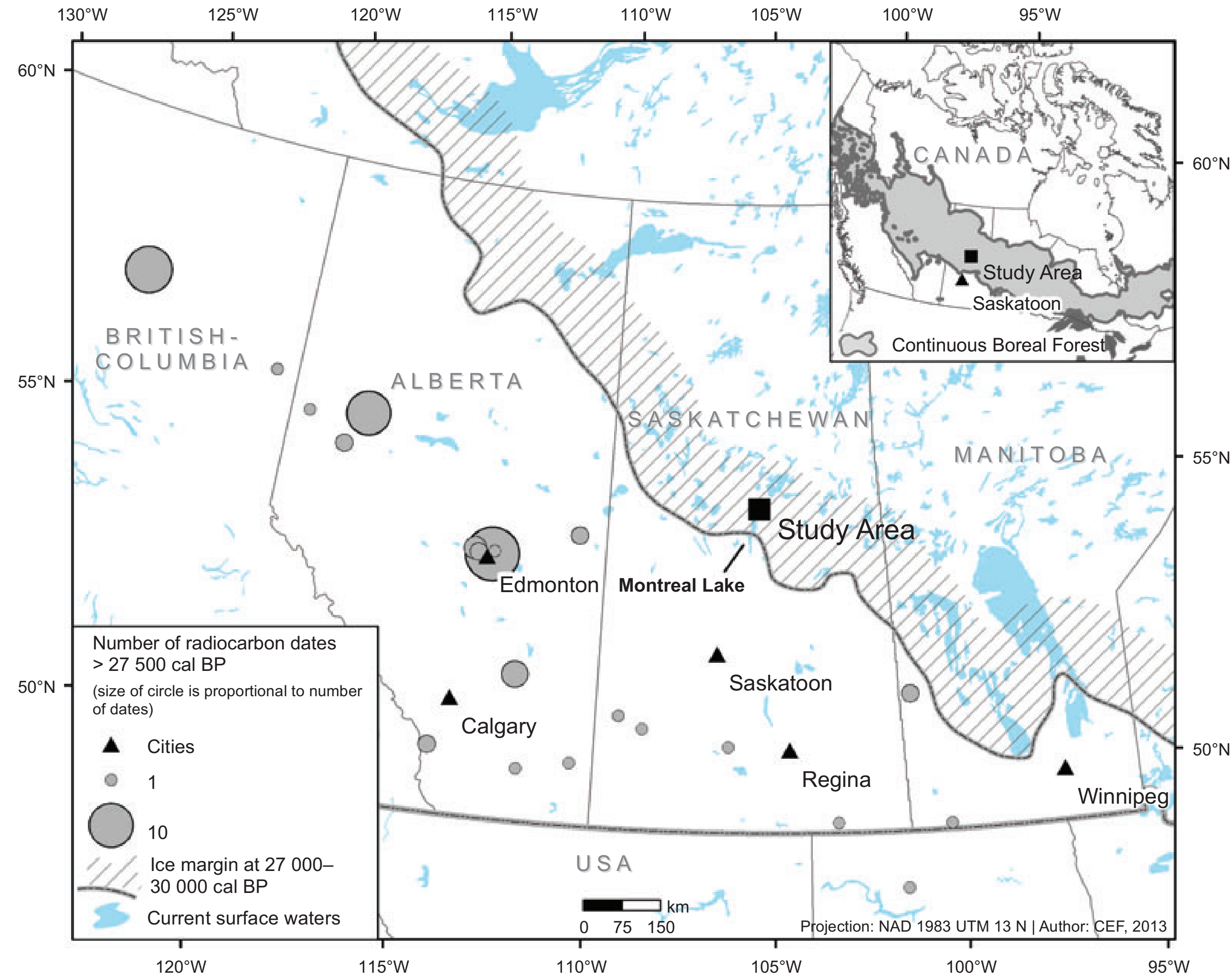

Fig. 1 Location of study area and pre-Last Glacial Maximum radiocarbon dates (50 000–26 600 cal BP) collated in Central Canada, east of Rockies. The interstadial ice margin (30 000–27 000 cal BP) in the Canadian prairies (Dyke *et al.*, 2002) is interpreted based on radiocarbon dates of faunal and plant remains (grey dots) (Burns, 1996; Dyke *et al*. 2001). The figure shows that the data from central Saskatchewan are the largest collection of ^{14}C dates during the pre-LGM (bulk of MIS 3) and effectively constrains the best approximated margin of ice recession but removes it another 150 km to the northeast (i.e. hatched area starting from the interstadial ice margin). The insert map (top-right) locates the study area (54°40′53″N, 105°29′55″W), north of Saskatoon (52°08′N, 106°38′W), whereas the large grey polygon illustrates the modern extent of the boreal forest.

Charcoal layers in paleosoil 1 were collected to a depth of 2 m from the bottom to the surface to avoid contamination of fragments falling from the upper soil horizons. Figure 2 provides the details of the sampling scheme for paleosoil 1. Soils in this profile were also sampled based on diagnostic horizons (i.e. E, B1, B2, BC, C). In paleosoil 2, charcoal layers were sampled to a depth of 1.4 m, again from the bottom to the surface, using the same approach as paleosoil 1. The shallowest layer of charcoal deposition was sampled in the other three paleosoils, which corresponds to a depth of approximately 50 cm for paleosoils 3 and 4 and 80 cm for paleosoil 5 (see Fig. 2 for the sampling scheme of paleosoil 5).

Radiocarbon (^{14}C) measurements were carried out on charcoal fragments using accelerator mass spectrometry (AMS) at the LMC14 laboratory (Saclay, France). Numerous ^{14}C dates were possible for each charcoal layer as all the layers contained several charcoal fragments of >10 mg sample (threshold mass for AMS dating). Thus, two ^{14}C dates on different charcoal samples were obtained for three layers of paleosoil 1, whereas one ^{14}C date per layer was carried out in paleosoils 2–5. As a whole, 17 ^{14}C measurements were carried out for this study. Prior to AMS dating, charcoal fragments were cleaned under a binocular (×40) to remove the small roots, the fungi hyphae and the mineral particles that could alter charcoal age. Further, each fragment was treated for more than 24 h with a solution of $Na_4P_2O_7$ to extract organics adsorbed within the charcoal porosity. The solution was changed daily until no more release of coloured organics was observed, which

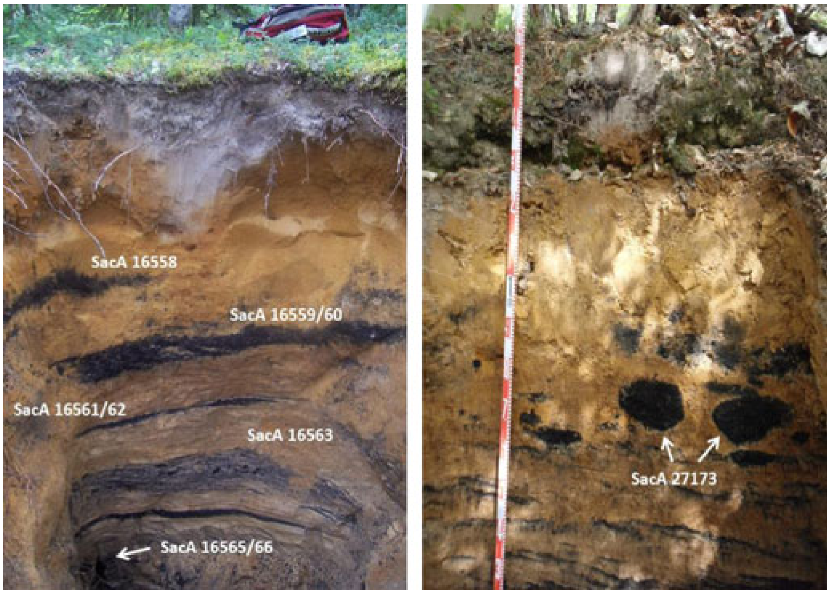

Fig. 2 Charcoal depositions in paleosoils. Paleosoils 1 (left) and 5 (right) at Montreal Lake, central Saskatchewan. Correspondence with laboratory sample codes for radiocarbon dating (Table 1) is provided, which indicates sampling intervals within the profiles. Note that sample SacA 16567 (178–183 cm) in paleosoil 1 falls outside of the picture.

normally takes 5–7 days. The ^{14}C measurements were calibrated with CALIB 6.1.0 (Stuiver & Reimer, 1993) using the Intcal09 data set. Also, the tool 'sum probabilities' of CALIB 6.1.0 was used with all measurements to compute a cumulated probability distribution, except for samples of >50 000 ^{14}C years BP (i.e. SacA16565, SacA16567 and SacA27177) because the Intcal09 calibration data set is valid only from 0 to 46 400 year BP. Two sigma ranges also impinged on the end of the calibration data set for two older dates (i.e. SacA 16562 and SacA 16566), and consequently, one sigma ranges were interpreted instead.

Only charcoal fragments >400 µm were identified from paleosoils 2–5 based on anatomical structure using an incident light microscope (×200, ×500, ×1000) and were compared with descriptions of wood in atlases (e.g. Jacquiot, 1955; Schweingruber, 1990) or with wood charcoal in reference collections. The two genus *Larix* and *Picea* were not distinguished due to their close anatomy (Marguerie *et al.*, 2000), resulting in the lumped charcoal taxa *Larix-Picea*.

RESULTS

In total, 194 fragments could be microscopically examined. Among them, only 93 fragments could be taxonomically identified. Needleleaf gymnosperm identifications in paleosoil 2 are dominant relative to broadleaf angiosperms (Table 1). Among gymnosperms, *Pinus* is identified in the first three layers, whereas *Juniperus* is in the second layer only (70–80 cm). Charcoal fragments in paleosoils 3–5 are all gymnosperms, with most of them being *Pinus* and few *Larix-Picea* types in paleosoil 3, and being either *Pinus* or *Abies* in paleosoil 5 (Table 1). About half of charcoal fragments in paleosoil 2 and most in paleosoil 4 were not identified because of an advanced stage of degradation altering the wood charcoal surfaces, which is likely related to the activity of soil micro-organisms (Nocentini *et al.*, 2010) and to the environmental conditions of combustion that control the final oxidation rate of resistant carbon (Naisse *et al.*, 2013). Furthermore, the mechanical breakage of charcoal fragments was also possible, in time, due to frost-thaw and plant roots. These mechanisms can reduce the size of fragments according to time, albeit they do not alter the charcoal surfaces (Carcaillet & Talon, 1996).

Radiocarbon dates of the charcoal range from >50 000 to 22 330 ^{14}C BP (Table 1), which is between >50 000 and 26 600 cal BP (Fig. 3). This set of ^{14}C measurements (n = 17) is the largest record in Central Canada during the Upper Pleistocene. The cumulated probability distribution of ^{14}C dates generated by the CALIB program indicates a high fire probability between 50 000 and 42 000 cal BP, and a lower probability at around 36 000, 33 000 and 27 000 ^{14}C BP, which illustrates the density of calibrated charcoal dates presented in Fig. 3.

Charcoal accumulation is time-stratified in paleosoil 2 with the shallowest (youngest) depth at 22 330 ^{14}C BP and the deepest (oldest) layer at more than 50 000 ^{14}C BP (Table 1). Time stratification of charcoal in paleosoil 1 is less clear: ages are between 44 000 and 40 000 ^{14}C BP in the upper 120 cm, >50 000 and 44 500 ^{14}C BP in the lower profile (150–200 cm), but 31 470 ^{14}C BP at 130 cm (Table 1, Fig. 2). This lack of pure stratification suggests more a problem of secondary accumulation by wind than the result of bioturbation. Indeed, millimetre size charcoal fragments could have been transported by wind to a few hundred metres from their source area (Clark *et al.*, 1998; Lynch *et al.*, 2004; Higuera *et al.*, 2007).

Table 1 Charcoal sample information, radiocarbon dates and botanical identifications of paleosoils 1–5. Each radiocarbon date within a layer was measured on a distinct fragment

Paleosoil/Depth (cm)	Laboratory code	Dates (^{14}C year BP)	Identified charcoal (*n*)	Charcoal taxa (×*n*)	†Sand–silt–‡carbon (%)
Paleosoil 1					
48–53	SacA 16558	41 200 ± 1200	Not available		94.8–5.2–3.8
66–77	SacA 16559	39 430 ± 940	Not available		84.9–15.1–5.7
66–77 (duplicate)	SacA 16560	43 800 ± 1600			idem
90–92	SacA 16561	39 030 ± 890	Not available		88.6–11.4–3.6
90–92 (duplicate)	SacA 16562	44 200 ± 1700			idem
108–117	SacA 16563	41 000 ± 1100	Not available		98.7–1.3–1.4
129–131	SacA 16564	31 470 ± 360	Not available		90.8–9.2–7.2
149–163	SacA 16565*	> 50 000	Not available		94.9–5.1–14.6
149–163 (duplicate)	SacA 16566	43 800 ± 1600			idem
178–183	SacA 16567*	50 000 ± 3500	Not available		96.2–3.8–1.4
Paleosoil 2					
50–60	SacA 27174	22 330 ± 130	30	Gymnosperm (×14) *Pinus* (×5) Angiosperm (×1) Unidentified (×10)	Not available
70–80	SacA 27175	38 250 ± 790	30	Gymnosperm (×10) *Pinus* (×2) *Juniperus* (×5) Unidentified (×13)	Not available
80–100	SacA 27176	40 000 ± 1000	30	Gymnosperm (×14) *Pinus* (×2) Unidentified (×14)	Not available
100–120	SacA 27177*	50 100 ± 3300	30	Gymnosperm (×5) Unidentified (×25)	Not available
Paleosoil 3					
50–60	SacA 27178	38 960 ± 850	24	Gymnosperm (×3) *Pinus* type *banksiana* (×15) Type *Larix-Picea* (×2) Unidentified (×4)	Not available
Paleosoil 4					
50–60	SacA 27179	29 030 ± 260	30	Gymnosperm (×6) Unidentified (×24)	Not available
Paleosoil 5					
80–100	SacA 27173	28 290 ± 260	17	Gymnosperm (×11) *Pinus* (×1) cf. *Abies* (×2) Unidentified (×3)	Not available

*Dates not calibrated because the Intcal09 calibration data set is not appropriate for dates >46 400 ^{14}C year BP.†Samples were first treated with sodium hexametaphosphate and sonication to disperse the samples before particle size distribution measurement on the Horiba Partica LA-950 Laser Particle Size Analyzer. None of the samples contained clay.‡Subsamples were also ground for determination of total C by dry combustion and infrared detection using the Leco CNS-2000 Analyzer.

Some parallel can be made between fire dates and DO warming events (Fig. 3). In the case of >40 000 cal BP dates, three dates were predominantly centred between the DO13 and DO12 events but overlapped with DO12. The centres of these dates are approximately 2000 years after the onset of DO13. Three dates were distinctly centred between the DO12 and DO11 events. The centres of these dates are 2000–3000 years after the onset of DO12. Three more dates overlapped with DO11. A final date lagged the onset of DO11 by ~400 years and was more closely centred during the cooling phase between the DO11 and DO10 events. For dates <35 000 cal BP, fires are strongly associated with DO7, DO6, DO5 and DO3. Fire dates strongly overlap DO events except with DO3 where fire appeared to have been delayed by ~1000 years. Despite of the uncertainties with ~50 000 cal BP dates, it is safe to suggest that fires occurred between the DO14 and DO13 events.

DISCUSSION

Plant species identification

In total, five gymnosperm taxa are identified in the Montreal Lake paleosoils. In central Alberta, *Picea* wood was also dated as MIS 3 (Burns, 1996). The presence of

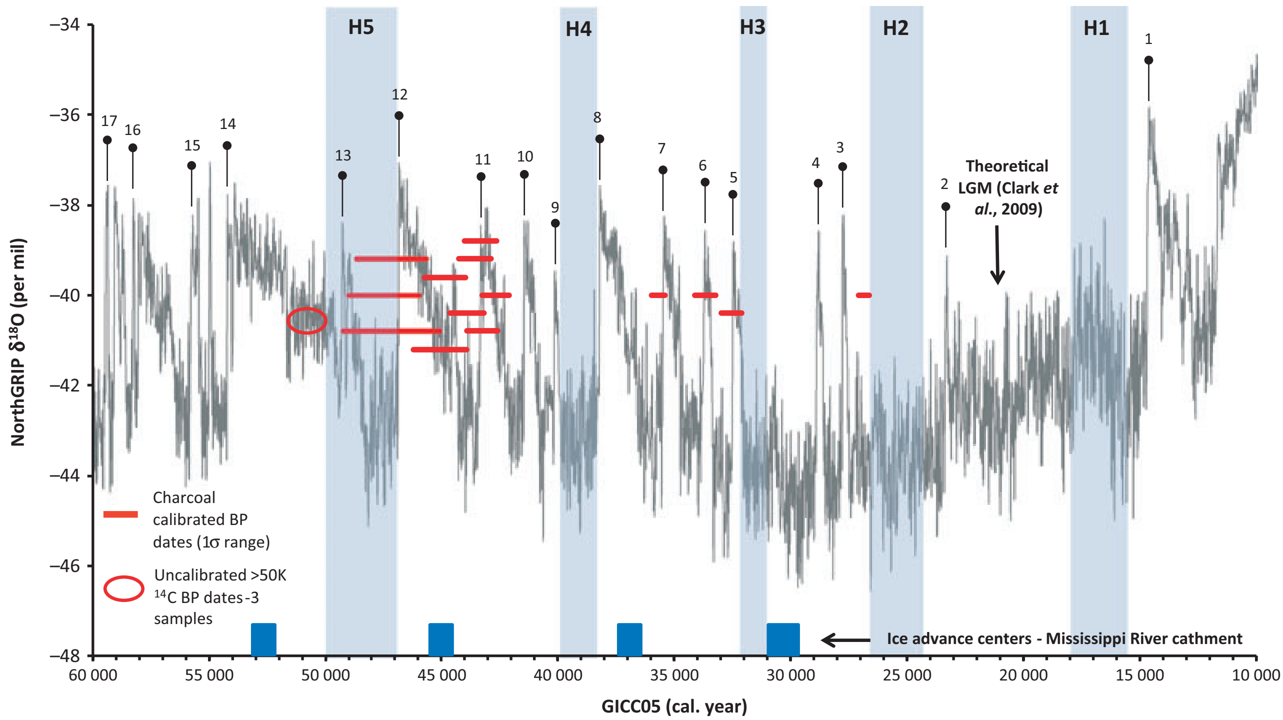

Fig. 3 Comparison between NGRIP oxygen isotopes and calibrated charcoal dates from the Montreal Lake study region, Central Canada, between ~50 000 and 26 600 cal BP. The fire history (red marks) is based on charcoal radiocarbon dating. Dansgaard–Oeschger (DO) events are labelled with black dots based on the most up-to-date NorthernGRIP $\delta^{18}O$ climate record plotted against GICC05 calibrated years (Svensson *et al.*, 2008). Heinrich (H) stadials 1–5 are labelled as vertical grey shaded areas (Sanchez Goni & Harrison, 2010). Centres of ice advance episodes into the Mississippi River catchment, as defined by melt water spikes into the Gulf of Mexico, are labelled as blue boxes (Tripsanas *et al.*, 2007).

permafrost at ice margins possibly hinders the establishment and growth of vascular plants, especially woody species. Our identifications demonstrate, however, that woody plants, eventually erected trees, grew in periglacial environments of MIS 3 in Central Canada. It was argued that *Picea* and *Pinus* could have dominated south of the ice margin not because of cold conditions, but rather because of low atmospheric CO_2 levels which would have favoured gymnosperms over angiosperms (Loehle, 2007). The very large dominance of gymnosperms from charcoal identifications near the LIS margin in Montreal Lake during MIS 3 is consistent with boreal trees prevailing in periglacial environments, close to the ice margin, and not taking refuge only in southern US areas. Our observations converge with those from Scandinavia showing that tree species (*Picea*, *Larix*, *Pinus*) survived in the glaciated mountains during the LGM (Kullman, 2002) probably on nunataks (Paus *et al.*, 2006) and that afforestation occurred at the edge of the retreating ice sheet (Carcaillet *et al.*, 2012). Matching with these north European patterns, our finding of close-to-ice presence of trees in Central Canada explains why afforestation in eastern Canada occurred immediately following ice recession without a tundra phase (Richard, 1980; Liu, 1990; Genries *et al.*, 2012).

Radiocarbon charcoal dates and extent and timing of ice advance and recession

The radiocarbon dates collected in this study (Table 1) along with other dates of faunal and plant remains (Burns, 1996; Dyke *et al.*, 2001, 2002) suggest that ice recession during the MIS 3 warming events left ice-free parts of the Upper Midwest and areas of the Great Plains. The most northern indication of ice-free conditions during MIS 3 is in central Alberta (Burns, 1996; Dyke *et al.*, 2002), whereas most other dates (Dyke *et al.*, 2001) suggest ice-free conditions in southern Alberta, Saskatchewan and Manitoba (Fig. 1). During the LGM, the ice margins were extrapolated south of the Canadian-US boundary in northern areas of Montana and North Dakota at 18 000 cal BP, with cold steppes at the ice sheet margin (Dyke, 2005). However, during MIS 3, the margins were extrapolated further north by Dyke *et al.* (2002) at about 30 000 cal BP from 300 to 500 km northeast of the highest latitudes of radiocarbon dates (Fig. 1). If the ice margins were relatively well constrained for Manitoba at 30 000 cal BP, they were not westward for Alberta and Saskatchewan as only a few dates are available. Our new fire evidence in central Saskatchewan during the bulk of MIS 3 effectively constrains the best approximated margin of ice recession but removes it at least another 150 km to the northeast (Fig. 1). This line of maximum ice recession during MIS 3, however, is built from ^{14}C dates that encompass not only a series of DO warming events (i.e. DO14 to DO2), but also cold stadials such as the H5 and H2 cooling events and episodes of continental ice advance (Fig. 3). The line in Fig. 1 is therefore an interpretation of maximum ice recession due to interstadials as a whole during MIS 3 and lower MIS 2 in Central Canada and was certainly not permanent between 50 000 and 26 000 cal BP.

Time series of melt water spikes in the Gulf of Mexico, a proxy of Mississippi River discharge, provided an indirect proxy for LIS advance in the Mississippi River catchment (Tripsanas *et al.*, 2007). Four major episodes of ice advance in the Mississippi River catchment were centred at 53 000, 45 000, 37 000 and 31 200–29 400 cal BP (Fig. 3), all of which clearly corresponding to cold periods (stadials) according to NGRIP and GICC05 years (Svensson *et al.*, 2008). In northern Europe, similar activity was recorded at 54 000–46 000 and 35 000–30 000 cal BP (Houmark-Nielsen, 2010). We recorded high cumulated probabilities of fires from >50 000 cal BP until 42 000 cal BP (Fig. 3). Our fire reconstruction therefore refutes the general idea that ice covered all of Saskatchewan during these periods. Due to the small sampling area covered in this study, however, further investigations are needed in the region and elsewhere in Central Canada (Alberta, Saskatchewan and Manitoba) to ascertain how much area was covered by ice and to possibly find new evidence of plant refugium during these glacial periods.

The LGM is suggested to have peaked at about 21 000 cal BP (Clark *et al.*, 2009), whereas Svensson *et al.* (2008) suggest an onset at about 27 200 cal BP and the end at 23 500 cal BP, between DO3 and DO2. The general scheme is a large and thick sheet of ice during lower MIS 2 covering western Canada and parts of the United States, from the Rockies to the Great Lakes region (Dyke, 2005). The absence of charcoal ^{14}C dates associated with the DO2 warming event (23 340 cal BP) would corroborate this general LGM scheme, unless the hazard of charcoal sampling and dating explains this pattern. The least dated fire is approximately 1000 years following the DO3 event, that is, ca. 27 000 cal BP. This is the only charcoal sample dated close to the onset of the theoretical LGM.

It is hypothesized that the cooling periods between 50 000 and 32 000 cal BP, which capture DO events from DO14 to DO5, might have been too short for the LIS to progress south of Montreal Lake. The area would therefore have been ice-free for a prolonged period, allowing for gymnosperms to produce ample fuel. Although our charcoal sample is relatively small for that time period (Table 1), there is a solid density of similar dates of boreal plant remains in Michigan (Schaetzl & Forman, 2008) and Alberta (Dyke *et al.*, 2001). While boreal trees were reported in the northwestern US during MIS 3 (Jiménez-Moreno *et al.*, 2010), they lack for the Great Plains. We therefore believe that this period between 50 000 and 32 000 cal BP, interspersed by warmer events (DO) and two cold stadials, reveals sustained ice-free conditions. This created suitable conditions for tree growth and for fires in central Saskatchewan.

Parallels between DO events and fire dates

The charcoal data can only be used to partially infer the response of fire to climate variability across glacial periods in central Saskatchewan because the lack of fire dates at a certain period does not mean the absence of fire disturbance during that time. Another caveat is that the fire dates (especially the older dates) have a large range of uncertainty, which makes it difficult to correlate with specific climatic events (stadials vs. interstadials). Furthermore, the inbuilt age (Gavin, 2001) increased this uncertainty, based on the fact that ^{14}C ages correspond to the wood age and not to the fire age (Carcaillet, 1998). The problem is amplified when we consider that dead woody debris can remain centuries on the ground in periglacial areas (Payette *et al.*, 1985) and that fires generally consume such debris (Begin & Marguerie, 2000).

The asynchronous patterns between fire and the DO warming and cooling events of MIS 3 offer too much complexity to be fully elucidated. Two general models can, however, be considered.

First, our data suggest that more than half of the older fire dates (i.e. >40 000 cal BP), despite large sigma ranges for some dates, correspond to cooling after the DO events 14, 13 and 12 (Fig. 3). This is also the case for one younger fire date (i.e. <35 000 cal BP) following DO3. The lagged response of fire to rapid climatic change could be explained several ways. The DO onsets have often been associated with an increase in effective moisture (Wagner *et al.*, 2010), which would have been conducive to the establishment of vascular plants after ice removal. But fires were not likely active at that time as fuel (biomass) was still low in abundance. The abundance of fuel and low effective moisture required to dry the fuel during fire season, and thus the appropriate conditions to promote these fires, seem to have occurred in most cases only hundreds or thousands of years following the DO onsets, during stadials. These cooling periods, which are normally linked to a greater dominance of high-pressure systems and low effective moisture, were likely central to generating sufficient fuels and landscape flammability, lightning being the ignition source. At that point, lower moisture conditions and the abundance of fuels could have overshadowed the effects of lower temperatures (Higuera *et al.*, 2008). During the last glacial episode, MIS 3 had the most variability in fires compared to MIS 2 and MIS 4, and fire increased due to DO warming events and decreased during cooling with a lag effect of ca. 100–200 years (Daniau *et al.*, 2010). The much longer lag effect in central Saskatchewan could likely be explained by the need for the ice to decay completely, whereas sites collated by Daniau *et al.* (2010) were not at the ice margin, and thus, fuels built quicker because vegetation could respond immediately to the rapid DO warming events.

Second, most younger fire dates (i.e. <35 000 cal BP) strongly overlap with DO events 7, 6 and 5. There are older fire dates overlapping the DO11 onset, whereas the oldest calibrated dates also overlap with the onset of DO12 due to their large sigma ranges that likely result from a measurement bias on the oldest material poor in residual ^{14}C. Because DO8, DO7, DO6 and DO5 are relatively close, it could be argued that the fire dates overlapping DO7 to DO5 events are associated with the onset of the previous DO events (e.g. DO8 vs. the fire date overlapping DO7). In such cases, the centres of these fire dates lag from 1000 to 2500 years the previous DO events, which would lead to similar conclusions as with fire dates falling within stadials. If we apply the same logic for the dates overlapping DO events 11 and 12, than the lag response of fire after DO12 is the longest at 3500–4000 years. Such a prolonged period for fire to appear on the landscape is more of a stretch, and thus, an alternative explanation is likely. Indeed, the uncertainty associated both to the ^{14}C dating and to the calibration process could result in a lag time in the chronology of events. Fire could have occurred few decades to few centuries after the age of the charcoal – cf. inbuilt age (Gavin, 2001) and burned woody debris (Begin & Marguerie, 2000). The age of charcoal would correspond to DO warming events that are wetter, thus favouring vegetation growth. Fires could have occurred either at the end of DO events or during the beginning of cold stadials. Further sampling and analysis of the charcoal in the Montreal Lake area will possibly allow to produce more robust inferences in regard to the response of fire to specific climatic events such as DO12.

CONCLUSION

We can conclude that Central Canada has experienced fires during MIS 3, from >50 000 until 26 600 cal BP. This evidence reveals fire-prone environments with long-lasting warm and moist periods promoting fuel build-up, notably needleleaf taxa of the boreal forest, and short dry events (season) allowing fire ignition and spread. Contrary to former glacier maps, the Montreal Lake area in central Saskatchewan was not covered by the LIS throughout the Upper Pleistocene. There is evidence that many periods were ice-free. MIS 3 in Central Canada was not a periglacial continental desert, but may be a long-lasting period that experienced fluctuations of the northern tree limits, where biogeochemical processes occurred under fire effects related to interstadial climate fluctuations. The ongoing tree expansion into the tundra (Sturm *et al.*, 2001; Hallinger & Wilmking, 2011; Naito & Cairns, 2011) associated with the current global warming could trigger fire occurrences in periglacial areas and thus the release of GHG (Schuur *et al.*, 2013).

Although the causes of the millennial-scale climatic oscillations during MIS 3 are not yet fully elucidated, we provided evidence that half of the charcoal dates followed DO warming events, whereas the other half of charcoal dates preceded the following DO events. The mechanisms leading to rapid or delayed fires following the onset of DO events are thus not elucidated. If this fire evidence during these pre-DO warmings is indicative of more large-scale processes in cold treed-tundra or treed-steppe biomes, DO warming events could be linked to the release of GHG such as N_2O, CO_2 and CH_4 (Flückiger *et al.*, 2004). This biogeochemical hypothesis contributing to the DO warming onset should be investigated based on the fact that modern tundra fires in Alaska are important processes of carbon emissions towards the atmosphere (Mack *et al.*, 2011; Schuur *et al.*, 2013) and that taiga fires situated near the tree line was the most important contributor of carbon emissions during the Holocene in eastern Canada (Taylor *et al.*, 1996; Bremond *et al.*, 2010).

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