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Did fires drive Holocene carbon sequestration in boreal ombrotrophic peatlands of eastern Canada?

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ABSTRACT

Wildfire is an important factor on carbon sequestration in the North American boreal biomes. Being globally important stocks of organic carbon, peatlands may be less sensitive to burning in comparison with upland forests, especially wet unforested ombrotrophic ecosystems as found in northeastern Canada. We aimed to determine if peatland fires have driven carbon accumulation patterns during the Holocene. To cover spatial variability, six cores from three peatlands in the Eastmain region of Quebec were analyzed for stratigraphic charcoal accumulation. Results show that regional Holocene peatland fire frequency was ~2.4 fires 1000 yr^{-1}, showing a gradually declining trend since 4000 cal yr BP, although inter- and intra-peatland variability was very high. Charcoal peak magnitudes, however, were significantly higher between 1400 and 400 cal yr BP, possibly reflecting higher charcoal production driven by differential climatic forcing aspects. Carbon accumulation rates generally declined towards the late-Holocene with minimum values of ~10 g m^{-2} yr^{-1} around 1500 cal yr BP. The absence of a clear correlation between peatland fire regimes and carbon accumulation indicates that fire regimes have not been a driving factor on carbon sequestration at the millennial time scale.

Introduction

Ongoing climate changes affect the ecological dynamics of boreal ecosystems where wildfires play a key role (Bergeron et al., 2010), including the carbon (C) balance of peatlands (Turetsky et al., 2002). Peatlands are frequent in topographic depressions of the boreal and subarctic regions, globally covering 3–4 million km^2 (MacDonald et al., 2006; Yu et al., 2010). Due to cool, humid, nutrient-poor and acidic conditions that restrict decomposition, ombrotrophic peatlands sequester organic C and expanded vertically and laterally over millennia (Korhola, 1994; Tolonen and Turunen, 1996; Korhola et al., 2010). As a result, the global northern peatland C stock, which started accumulating after the Last Glacial Maximum, presently contains ~547 Pg (range: 473–621 Pg), constituting approximately a third of global soil C (Yu et al., 2010). Peatland fires influence local C dynamics by a release of C to the atmosphere through combustion, estimated at 2.5–3.2 kg m^{-2} per fire (Pitkänen et al., 1999; Turetsky et al., 2002). In addition, postfire C loss is generally important due to a delay in vegetation re-establishment (Wieder et al., 2009).

In Canadian boreal zones, a potential increase of fire frequency and area burned during the next decades (Flannigan et al., 2005; Bergeron et al., 2010) may force peatlands and upland areas to switch from net C sinks to sources to the atmosphere. The negative effect of recurrent fires on long-term peat and C accumulation has been established for different peatland types and regions around the globe, yet research has been concentrated in boreal western and central Canada (e.g. Kuhry, 1994; Robinson and Moore, 2000; Turetsky et al., 2002; Camill et al., 2009). In these regions, climate-driven increases in fire frequency or severity may force peatlands to switch from net sinks to sources in the course of the 21st century (Turetsky et al., 2002; Wieder et al., 2009). Whereas in western Canada forested peatlands have developed under a dry continental climatic regime, dominance of open bogs with relatively high water tables in the more humid climate of eastern boreal Canada may inhibit differential fire and C accumulation dynamics (Payette et al., 1989; Payette and Rochefort, 2001). In the Eastmain region, peatlands have accumulated considerable amounts of C during the last ~7000 yr, regionally averaging 91 kg m^{-2} at a mean rate of 16.2 g m^{-2} yr^{-1} (van Bellen et al., 2011a). C accumulation rates are given here as "apparent" rates,

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representing the remaining accumulated C actually present in the peat sequence, instead of "real-time" historical accumulation rates; the difference between these two rates is due to long-term decomposition of peat under anoxic conditions.

Peatland hydrology and vegetation strongly influence the potential for fire propagation, whereas the amount of peat consumed varies with local and regional humidity conditions (Zoltai et al., 1998). The highly variable spatial patterns of moisture conditions within and among peatlands are closely linked to surface microtopography (i.e. hummocks and hollows) and vegetation composition. Hence, interactions between vegetation and hydrology are important factors on burning potential (Higuera et al., 2009). The presence of trees and shrubs in peatlands positively influences fuel continuity and it is therefore associated with more frequent fires (Camill et al., 2009), and forested bogs may be more susceptible to burn as water tables are generally low. Although burning in wet, open peatlands is evident under drought conditions (Fig. 1) open peatlands often remain unaffected by fire, especially when local water tables are high and trees are sparse (Hellberg et al., 2004). As a result, fire frequencies in open peatlands are generally lower than those of adjacent forest stands (Kuhry, 1994; Zoltai et al., 1998; Camill et al., 2009).

Here, we present a study on the Holocene patterns of fire and C sequestration based on the quantification of stratigraphic charcoal from three ombrotrophic peatlands in the Eastmain region in boreal Quebec, northeastern Canada. Specifically, we aimed to estimate if peat fires have driven long-term variations in C accumulation. Ombrotrophic peatland fire regimes, ecohydrological change and C sequestration are directly or indirectly linked to climate conditions, therefore we considered variations in climate regimes during the Holocene. Given the evident negative influence of individual fires on peatland C sequestration (Wieder et al., 2009), we hypothesized that if Holocene fire regimes had driven long-term C sequestration, this would be manifested by a negative relationship between fire regimes and C accumulation rates (cf. Kuhry, 1994). As shown in the study by Kuhry (1994) in western boreal Canada, we verified correlations between fire frequency and C accumulation rates, as well as between peatland fire-associated cumulative charcoal and C accumulation rates. This study provides an understanding of Holocene dynamics in peatland fire regimes in a relatively unexplored region. Besides, additional knowledge on the influence of historical fire regimes on C sequestration may also clarify the role of unforested boreal peatlands in future C cycling.

Study region

Three pristine peatlands were studied, Lac Le Caron (LLC), Mosaik (MOS) and Sterne (STE), located in the Eastmain river watershed (51°50′–52°20′N/75°00′–76°00′W) (Fig. 2). Regional mean annual temperature is −2.1 ± 0.2°C (January: −22.0 ± 0.5°C; July: 14.6 ± 0.2°C) and mean precipitation is 735 ± 12 mm, of which about one third falls as snow (interpolated means and standard errors of 1971–2003 NLWIS data; Hutchinson et al., 2009). Forest fire regimes are among the most important in eastern Canada with actual fire cycles (i.e. the time required to burn an area equivalent to the area studied) of 90–100 yr (Payette et al., 1989; Mansuy et al., 2010). Regionally, all large fires may be assumed wildfire as human activity is highly restricted. Figure 1 shows an example of an Eastmain peatland burning pattern. A complete description of peatland characteristics can be found in van Bellen et al. (2011a).

Methods

Fieldwork

From each of the three peatlands two coring sites were selected at opposing sections within each peatland (Fig. 2). To obtain records with a sufficiently high temporal resolution, we aimed to extract cores of at least 1.5 m in length, nearby the forest–peatland boundary, which was identified by absence/presence of a surface *Sphagnum* cover and an organic horizon thickness of >40 cm (Commission Canadienne de Pédologie, 1998). As the slope of the peatland basin was highly variable between sites, the distance between coring location and peatland–forest limit varied between 12 and 132 m. Coring was performed using a box corer (10 × 10 cm width) to sample the upper 1 m and Russian peat samplers (4.5- to 7.5-cm diameter) for deeper horizons. Monoliths were wrapped in plastic, transferred to polyvinyl chloride tubes and stored at 4°C until analysis.

Charcoal fragment and peat C quantifications

Prior to specific treatment, cores were sliced into contiguous 1 cm subsamples in the laboratory. From each slice, 2 cm^3 was retained for macrocharcoal analysis, assumed large enough to provide replicable data (Carcaillet et al., 2001). The subsample was soaked for at least 14 h in 5% KOH and carefully rinsed through a 355-μm sieve. Material

Figure 1. Eastmain region peatland after fire. This peatland was not a study site. Photo by Hydro-Quebec.

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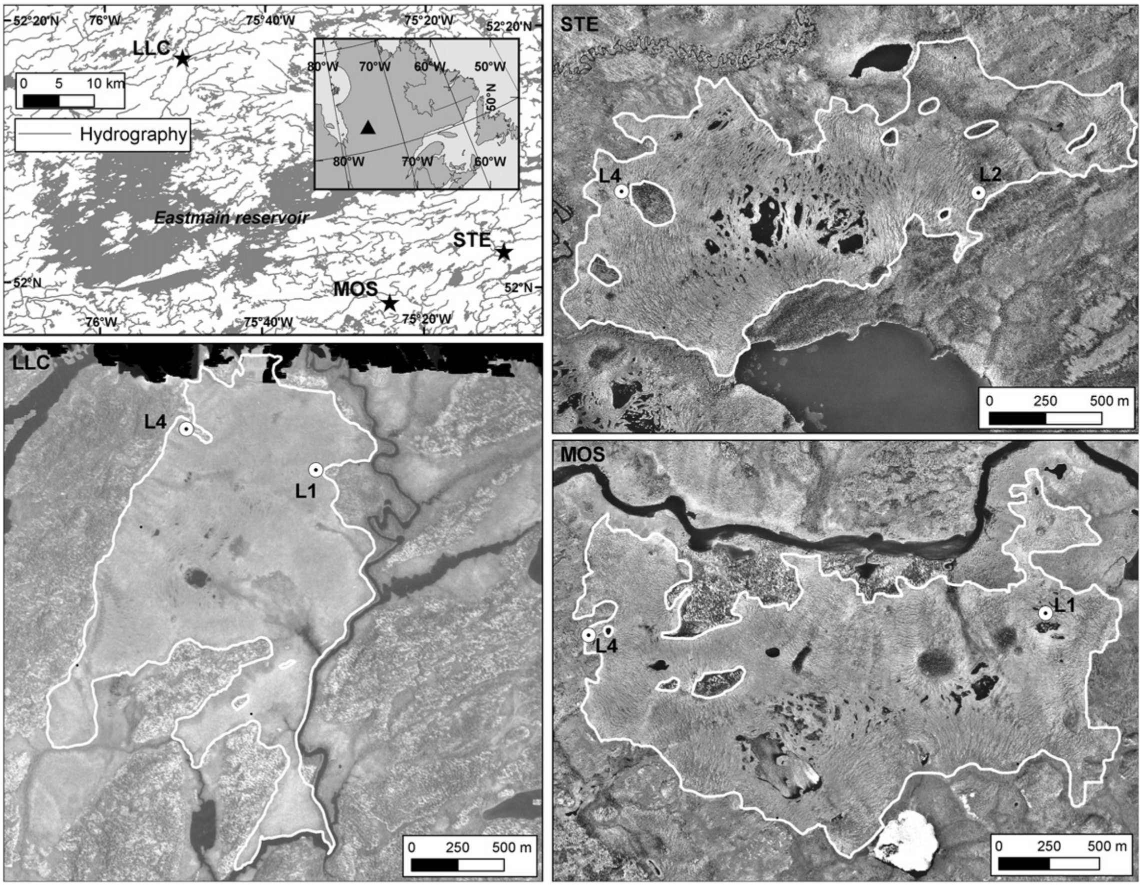

Figure 2. Study region, peatlands and coring locations.

larger than 355 μm was transferred to a dish for dry analysis. Macrocharcoal fragments were counted using a binocular microscope (×25).

Peat C contents were calculated from bulk density and loss-on-ignition (LOI) analyses. Bulk density was determined from contiguous 1 cm^3 subsamples after drying for 16 h at 105°C. Subsequently, LOI analysis was performed at 550°C for 3.5 h (Heiri et al., 2001). The amount of organic matter (OM) was defined as the product of bulk density and LOI (Dean, 1974). The resulting OM was converted to organic C assuming a mean of 0.5 g C g^{-1} OM (Turunen et al., 2002).

Dating and age-depth models

A total of 40 samples was submitted to Keck-CCAMS facility (Irvine, USA) for ^{14}C accelerator mass spectrometry (AMS) dating. Dated samples were selected either at levels of apparent charcoal peaks or, when peaks were less conspicuous, at the boundaries of zones with abundant fragments. Radiocarbon ages were calibrated using the IntCal04 calibration curve (Reimer et al., 2004) within the *Bchron* software package (Haslett and Parnell, 2008). Age-depth models were constructed assuming vertical accumulation as a continuous monotonic process applying piecewise linear interpolation. All ages were expressed as calendar years before present (BP = before AD 1950). Age-depth models were based on these calibrated ages and the age of the peatland surface when cores were extracted, established at −56 cal yr BP (i.e. AD 2006). C accumulation rates were calculated for different sections of each core by dividing the C mass by the period of accumulation and correcting for a uniform surface area (Clymo et al., 1998).

Origin of peat charcoal deposits

Long-term reconstructions of fire regimes from sediment records in lakes and peatlands are generally inferred from macroscopic charcoal quantification from sediments, with peaks in the number of charcoal fragments reflecting "local" fires, i.e. at a 500–1000 m distance (Ali et al., 2009b). Stratigraphic charcoal presence is generally quantified by charcoal accumulation rates (CHAR; expressed as pieces cm^{-2} yr^{-1}). Here, we assumed that fires originate from the upland forest with some of these potentially spreading into peatlands, leaving patchy patterns in case spatial variability in ecohydrology is important (e.g. Fig. 1). Depending on injection height, fire intensity, forest type, wind speed, and local topography, charcoal fragments are dispersed over large distances; therefore some of the charcoal fragments found in peatlands may be originating from the upland forest. In general, small particles tend to travel farther than large ones (Clark, 1988; Ohlson and Tryterud, 2000; Lynch et al., 2004; Peters and Higuera, 2007). Thus, we assumed that quantification of large charcoal fragments (> 355 μm) could be a means of establishing a peatland fire history. As taphonomic processes controlling the accumulation of charcoal in peatland and lacustrine environments are different, we needed to adjust the methodology to identify fires from charcoal records.

Peat fire identification

Prior to the analysis of charcoal records, all cores were rescaled to a constant sample age resolution, defined by the median value, to reduce biases resulting from changes in sediment accumulation rate. Data manipulation was performed using *CharAnalysis* (Higuera et al., 2009) To identify peatland fire events, we hypothesized that the

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CHAR records were composed of a slowly varying background value (C_{back}) and a peak component (C_{peak}), comprising two populations of CHAR values (C_{noise} and C_{fire}). Variations in C_{back} are the result of long-term variations in charcoal production, charcoal deposited after fire, occasional charcoal movement through the acrotelm and possible slight contamination during sampling and preparations. We applied a LOWESS smoothing (robust to outliers) to calculate C_{back} to detrend the CHAR records. The C_{peak} component was assumed to be composed of Gaussian distributions that were modeled with a mixture model (Gavin et al., 2006; Ali et al., 2009a; Higuera et al., 2010), the lower-mean population C_{noise} reflecting long-distance charcoal input and random variability (cf. Higuera et al., 2010), while the higher-mean population C_{fire} representing charcoal from local peat vegetation. We used a local approach in determining the threshold separating C_{fire} and C_{noise}, i.e. the threshold was determined for sections of the C_{peak} record rather than being uniform for the entire sequence, as this approach takes account of changes in variability in the record (Higuera et al., 2010). We consider the 99th percentile of each C_{noise} distribution as a threshold, and C_{fire} values exceeding this threshold are interpreted as peat fires. Finally, a minimum count screening was applied to distinguish significant peaks (Gavin et al., 2006). To verify the sensitivity of the reconstructions to different methods, results for different C_{back} smoothing window widths and the local vs. global approach were compared.

Temporal variations in fire frequency were synthesized to obtain an image of historical changes at the scale of the region. Secondly, the peak magnitude was defined as the number of charcoal fragments associated with a single peak exceeding the final threshold. Previous studies interpreted peak magnitude as a proxy for fire size, fuel consumption (Higuera et al., 2009), fire proximity or intensity (Whitlock et al., 2006; Hély et al., 2010; Higuera et al., 2011). Regional trends in peak magnitudes were reconstructed since 4500 cal yr BP, which was the maximum age of core LLC_L4, to obtain an unbiased history (i.e. uniform sample size). To quantify the relationship between fire frequency and C accumulation rates and between cumulative peak magnitudes and C accumulation rates, regression analyses were performed on 1000-yr mean values for each record. Cumulative peak magnitudes are defined as the sum of the peak magnitudes that lie within the same 1000-yr interval.

Results

Chronologies

Radiocarbon dating showed ages of peat inception varying between 7321 cal yr BP (MOS_L1) and 4586 cal yr BP (LLC_L4) (Table 1). Age-depth models show generally convex shapes, with lowest rates of accumulation between 2000 and 1000 cal yr BP and apparent high rates towards the surface (Fig. 3), an effect of the incomplete decomposition of recently deposited organic matter. Median sample resolutions, which were used to rescale each record, varied between 22 and 48 yr sample^{-1} (Table 2).

Table 1
Radiocarbon dates selected by peatland and core. Age is defined by the median of the 2σrange. Sph = *Sphagnum*, Eric = Ericaceae, Cyp = Cyperaceae.

Site	Core	Sample depth (cm)	UCIAMS laboratory number	Material	^{14}C age (^{14}C yr BP)	2σ range (cal yr BP)	Age (cal yr BP)
LLC	L1	45–46	58637	Sph stems	630 ± 15	551–670	601
	L1	58–59	54956	Sph stems	1250 ± 30	1078–1289	1207
	L1	68–69	58639	Sph stems, Larix/Picea leaf frs	1940 ± 20	1810–1975	1889
	L1	99–100	57423	Sph stems	3125 ± 15	3255–3390	3355
	L1	112–113	58638	Sph stems	3395 ± 15	3586–3702	3656
	L1	130–131	54955	Sph stems	3780 ± 25	4056–4269	4162
	L1	170–171	57418	Sph stems	4625 ± 15	5295–5447	5418
	L1	210–211	64581	Sph stems	5035 ± 20	5721–5903	5841
	L1	249–254	40365	Sph stems, Eric leaf frs	6055 ± 20	6673–7218	6908
	L4	39–40	57417	Sph stems, Eric/Picea leaf frs	190 ± 15	151–294	198
	L4	52–53	58632	Sph stems	1070 ± 15	933–1075	978
	L4	61–63	58633	Sph stems, Eric leaf frs	1405 ± 20	1216–1427	1301
	L4	78–79	57415	Sph stems, Larix/Eric leaf frs	2170 ± 15	2129–2317	2152
	L4	127–128	57422	Sph stems	3135 ± 15	3342–3435	3351
	L4	147–148	58635	Sph stems, Eric/Larix leaf frs	3495 ± 20	3670–3826	3769
	L4	188–189	40368	Sph stems	4120 ± 20	4520–4788	4586
MOS	L1	55–56	65378	Charcoal frs	190 ± 15	147–379	243
	L1	77–78	65385	Charcoal frs	1260 ± 20	1071–1308	1218
	L1	89–90	65389	Charcoal frs	1840 ± 20	1678–2045	1787
	L1	117–119	65386	Charcoal, Cyp seeds	3625 ± 20	3626–4312	3933
	L1	167–170	43474	Eric leaf frs; Cyp seeds	6420 ± 20	6694–7657	7321
	L4	51–52	57425	Sph stems	455 ± 15	502–550	545
	L4	73–74	58642	Sph stems	2165 ± 20	2097–2285	2130
	L4	97–99	58641	Eric/Larix leaf frs	2750 ± 20	2790–2911	2852
	L4	136–137	58640	Eric/Larix/Picea leaf frs	3835 ± 15	4162–4326	4322
	L4	169–170	43476	Sph stems	4670 ± 20	5318–5457	5323
STE	L2	44–45	67506	Sph stems	265 ± 25	269–436	316
	L2	68–69	67507	Sph stems	1195 ± 25	1031–1232	1123
	L2	98–99	67508	Sph stems	2380 ± 25	2329–2588	2399
	L2	142–143	67509	Sph stems	3200 ± 25	3360–3471	3415
	L2	180–181	67510	Sph stems	3820 ± 25	4108–4363	4207
	L2	212–213	67511	Sph stems	4465 ± 25	4977–5288	5182
	L2	244–246	40362	Sph stems	5760 ± 20	6412–6725	6550
	L4	35–36	65384	Sph stems	135 ± 20	70–287	224
	L4	48–49	67512	Charcoal frs	945 ± 25	778–967	867
	L4	63–64	65380	Charcoal; Picea leaf frs	2455 ± 20	2343–2708	2524
	L4	84–85	67513	Charcoal frs	3540 ± 25	3703–3948	3826
	L4	125–126	65376	Charcoal, Picea/Eric leaf frs	5490 ± 20	6189–6339	6288
	L4	174–176	40363	Sph stems; Larix leaf frs; Cyp seeds	6185 ± 20	6976–7345	7090

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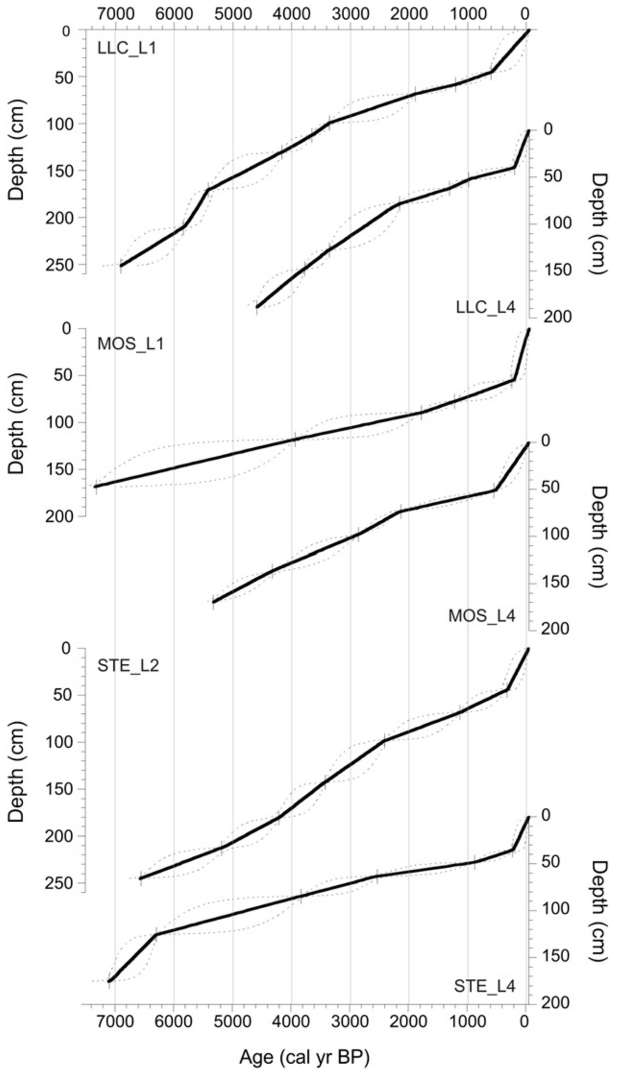

Figure 3. Age-depth model (black) and 95% age confidence intervals (gray, dashed) for each record.

Peat fire histories

Following the recommendations of Higuera et al. (2010), window size for determination of C_{back} was equivalent to $>30*$median sample resolution to obtain sufficient CHAR values for calculation of local thresholds. Median C_{back} values for each core varied between 0.021 and 0.159 pieces $cm^{-2}\,yr^{-1}$ and were consistently higher in cores from the western margin of each peatland (LLC_L4, MOS_L4 and STE_L4) compared to those from the eastern margin (Table 2). As fires are generally associated with western winds (Bergeron et al., 2004), variations in C_{back} may thus be partly the result of variations in charcoal transport, and therefore the location of the coring site relative to the adjacent forested upland may be a factor. The use of local thresholds resulted in a good separation of C_{fire} and C_{noise}, as median SNI values were >4.5 (Table 2; Kelly et al., 2011). As the mean number of identified fires was not significantly different for the various methods (varying smoothing windows and local vs. global thresholds), we assumed our local threshold reconstructions to be suitable for the identification of fires.

The resulting total number of fires detected per core varied between 9 and 20 individual events since local peat inception (Table 3; Fig. 4), representing a mean ($\pm\sigma$) Holocene frequency of 2.4 ($\pm$0.8) fires 1000 yr^{-1}. Regionally, fire frequency slightly diminished over the last 4000 yr, but temporal trends were highly variable as shown by the large whiskers (Fig. 5a).

Peak magnitude z-scores increased markedly after 1400 cal yr BP and remained high until 400 cal yr BP (Fig. 5b). Pooled peak magnitudes z-scores were significantly higher during the 1400–400 cal yr BP period compared to the previous and succeeding periods (t test; $p = 0.0046$; Fig. 5b).

Table 3
Peatland core fire regime characteristics.

Peatland	Core	Fires (#)	Holocene fire frequency (# 1000 yr^{-1})	Holocene CAR ($g\,m^{-2}\,yr^{-1}$)
LLC	L1	15	2.2	15.3
	L4	16	3.4	19.4
MOS	L1	15	2.0	10.3
	L4	14	2.6	15.7
STE	L2	20	3.0	15.9
	L4	9	1.3	13.3

C sequestration and fire-frequency patterns

Holocene apparent rates of C accumulation varied between 10.3 and 19.4 $g\,m^{-2}\,yr^{-1}$ (Table 3). Temporal variations in C accumulation rates were reconstructed within all cores, and showed a minimum mean value of ~10 $g\,m^{-2}\,yr^{-1}$ around 1500 cal yr BP (Fig. 5c). Regression analyses show a significant but weak positive relationship between the mean fire frequency and C accumulation rates (Fig. 6a), but no relationship between the cumulative peak magnitudes and C accumulation rates (Fig. 6b).

Discussion

Peatland fire detection using stratigraphic charcoal

Reconstructing peatland fire regimes is a challenging issue because taphonomic processes controlling charcoal sequestration are likely to be more variable in peatlands than in lacustrine environments. Despite these complications, understanding the

Table 2
Peatland core and analyses characteristics.

Peatland	Core	Distance to forest (m)	Core length (cm)	Basal age (cal yr BP)	Median sample resolution (yr $sample^{-1}$)	C_{back} smoothing window (yr)	Median C_{back} (pieces $cm^{-2}\,yr^{-1}$)	Median SNI
LLC	L1	26	252	6908	26	800	0.021	4.9
	L4	12	189	4586	22	1000	0.159	4.5
MOS	L1	132	169	7321	47	1500	0.101	12.8
	L4	39	170	5323	30	900	0.120	10.2
STE	L2	57	246	6550	24	1000	0.059	5.9
	L4	34	176	7090	48	1500	0.091	10.5

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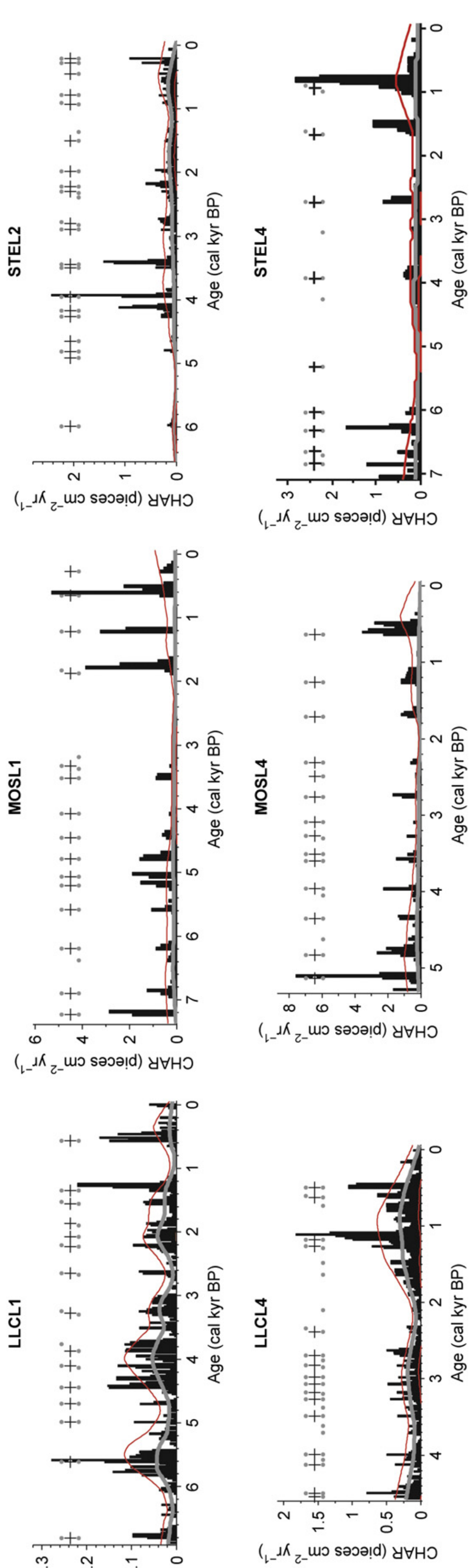

Figure 4. Holocene CHAR records showing C_{back} (thick gray line) and threshold (thin red line) values. Identified fires are indicated for each core by "+" symbols.

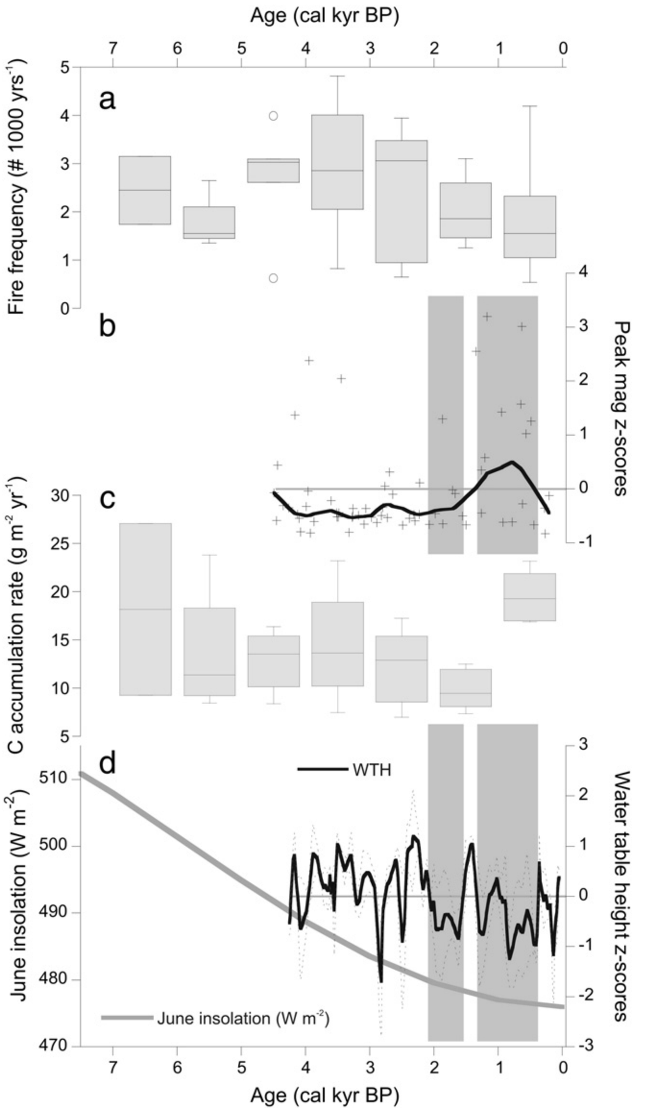

Figure 5. a) Box plot of all fire frequencies for each millennium, showing the median, upper and lower quartile, whiskers indicate the maximum values (excepting outliers); b) pooled peak magnitude z-scores, showing significantly higher values between 1400 and 400 cal yr BP (gray zone). c) box plot of C accumulation rates for each millennium. d) Holocene variation in June insolation at 60°N (Berger and Loutre, 1991) and Eastmain peatland mean water table height z-scores (±SE) from three cores. Adapted from van Bellen et al., 2011b.

resilience of peatland C stocks to variations in fire regimes is of major importance considering the feedback on climate change and future greenhouse gas budgets. The SNI values of >4.5 for all cores show that charcoal peaks were clearly separated from noise, even if the relative contribution of charcoal influx from uplands remains unclear. Although charcoal peak identification could not be validated as no independent regional fire records were available, the charcoal deposition pattern of a recent fire aided interpretation. In 1995, the adjacent forest along the northwestern section of MOS bog burned, as indicated by fire maps (Ministère des Ressources naturelles et de la Faune, 2010), although this fire did not affect the coring locations. Considering the apparent rapid accumulation and some potential for small fragments to move vertically through unconsolidated peat, this fire may have caused charcoal deposition in the upper 20 cm of the peat. Some concentrated charcoal fragments were found in MOS_L1 and MOS_L4 around 18 and 14 cm, respectively, yet these charcoal peaks were not significant. Thus, these fires were correctly omitted in the reconstructed fire record. Another indication for the detection of strictly local fires is the fact that both cores of each peatland did not register the same patterns in frequency, showing that strictly local conditions, as peatland microtopography, may play an important role in both fire potential and taphonomic processes.

In the Eastmain peatlands, charcoal horizons were visually indistinct throughout the sequences, probably because of relatively low charcoal fragment quantities and poor peat preservation resulting in a general dark appearance. The absence of charred *Sphagnum* in these horizons implies that most burning was surficial, only charring the standing biomass, i.e. (small) trees and shrubs. The apparent regional increase in peak magnitude during the late-Holocene could theoretically be due to higher decomposition and thus loss of charcoal fragments towards older deposits. Nevertheless, we have indications that differential decomposition was of minor importance here as decomposed organic matter did not show a distinct trend downward in detailed stratigraphic analyses (van Bellen et al., 2011a) and charcoal fragments were generally large (>0.5 mm).

Considering the important intra-peatland variability and poor synchroneity in reconstructed fire events, peatland fire frequency in the region appears to have been influenced principally by strictly local factors, possibly dominated by local peatland hydrology and microtopography. These results confirm that, in order to obtain complete understanding of the complexity of peatland fire regimes, multiple records from multiple peatlands are essential to cover spatial variability.

The reconstructed mean Holocene fire frequency of 2.4 fires 1000 yr^{-1}, comparable to a fire interval of ~400 yr, implies that mid- to late-Holocene peatland fires were less frequent than the present-day upland fires as the actual mean regional forest-fire cycle in the Eastmain region was quantified at 90–100 yr (Payette et al., 1989; Mansuy et al., 2010). This difference in fire frequency is in accordance with the hypothesis that fires are ignited in the upland and only occasionally proceed into peatlands, although forest-fire intervals have been highly variable since the mid-Holocene (Cyr et al., 2009), complicating comparisons between actual fire cycles and Holocene averages. The Eastmain region peatland fire interval falls within estimates from other boreal regions, even though spatial and temporal variability in fire activity is high. Zoltai et al. (1998) estimated peat fire return intervals around 150 yr in continental boreal bogs and 800 yr in humid boreal bogs. In Manitoba, fire return intervals were quantified between 624 and 2930 yr, depending on the criterion for local fire identification (Camill et al., 2009). In western Canada, reconstructions in *Sphagnum* peatlands show temporally and spatially highly varying frequencies, from mid-Holocene frequency of 5.3 fires 1000 yr^{-1} to no recorded fires over the entire accumulation in other sections (Kuhry, 1994).

Late-Holocene climate and fire frequency

We showed that mid- to late-Holocene peatland fire frequencies have been highly variable spatially and temporally in the Eastmain region, while charcoal peaks were significantly higher between 1400 and 400 cal yr BP. Previous research has indicated that boreal western Quebec forest fire frequency was generally high during the warmer mid-Holocene before 3500 cal yr BP, characterized by higher summer insolation and higher temperatures (Fig. 5d) whereas fire frequencies were lower during the cooler Neoglacial (Ali et al., 2009a; Hély et al., 2010). Comparisons between forest fire regime aspects and those of peatlands are complicated not only by taphonomic differences between lake and peat records, but also by a possible differential sensitivity of fire regime aspects to climate conditions. In forest stands, climate primarily regulates large-scale fire activity by the occurrence of summer drought events with variations in temperature, precipitation and wind speed (Bergeron and Archambault, 1993; Senici et al., 2010). However, in open peatlands, microtopography may be important, complicating the potential for propagation of fire as microforms have differential sensitivity to drought conditions (Benscoter and Wieder, 2003). Compared with forest stands, drainage and insolation dynamics, and thus summer water-table fluctuations, are linked to climate in a different manner. Summer precipitation and to a lesser extent temperature are likely to influence summer water tables (Charman et al., 2009; Booth, 2010). Furthermore, indirect climate effects and climate–vegetation interaction may exert an additional influence on peatland surface drought. Snow cover, for instance, varies with vegetation type and microtopography (Camill and Clark, 2000), and the combination of hummock *Sphagnum* and thin snow covers may allow permafrost aggradation (Zoltai, 1993), which further complicates the climate–hydrology relationship. Thus, even if regional climate has been a principal factor driving long-term variations in water tables and moisture contents, upland forests and peatlands could potentially show differential trends as they are driven by specific aspects of climate regimes, which in turn are mediated by specific ecological feedbacks on climate conditions (van der Molen and Wijmstra, 1994; Wotton and Beverly, 2007).

Corresponding to the trends obtained for uplands by Ali et al. (2009a) and Hély et al. (2010), late-Holocene regional trends of decreasing fire frequency are, although subtle, observed in the Eastmain peatlands. Nevertheless, we showed that intra-peatland variability is the dominant characteristic of the Eastmain peatland fire regimes, emphasizing the importance of local ecohydrology and microtopography on peatland fire regimes (see also Benscoter and Wieder, 2003). This importance may be reflected by the higher charcoal peak magnitudes between 1400 and 400 cal yr BP (Fig. 5b), corresponding to generally drier conditions in an independent record of Eastmain region water table heights (Fig. 5d; details in van Bellen et al., 2011b).

Differing trends in fire frequency and peak magnitudes may be explained by differences in controlling factors: peatlands may be more frequently affected by fire in absence of an important microtopography, whereas peak magnitude, which may reflect charcoal production, may be higher when dry microtopes are abundant or during periods of drought resulting in a low peat humidity.

Fire frequency and C sequestration

We hypothesized that if fires were important millennial-scale drivers for C sequestration, the studied fire regime aspects and C accumulation rates would show a strong negative correlation. However, as regression showed only a very weak, positive relationship between fire frequency and C accumulation rates (Fig. 6a), we conclude that Holocene C sequestration has not been primarily driven by the studied aspects of the fire regime. As fires generally have a negative effect on C sequestration through direct combustion and post-fire net emissions (Turetsky et al., 2002; Wieder et al., 2009), short-term fire effects have apparently been masked by other, more important autogenic or allogenic factors. Relatively low fire frequencies in the Eastmain region, combined with a dominance of surface fires, may be the principal cause for the absence of a significant influence of fire regimes on long-term C accumulation rates.

Long-term C accumulation patterns in peatlands are the result of combined effects of climate variations (e.g. temperature, precipitation and moisture balance effects) and internal dynamics (large-scale peatland surface topography, microtopography and basin morphology) (Belyea and Clymo, 2001; Yu et al., 2009; van Bellen et al., 2011a). Eastmain peatland long-term C accumulation has been affected by late-Holocene climate change (van Bellen et al., 2011b) that was mediated by local geomorphology (basin morphology) and surface topography factors (van Bellen et al., 2011a). Climatic cooling may have been the principal driving factor of C sequestration, as limited rates have been identified between 2000 and 1000 cal yr BP in multiple cores from the Eastmain region (van Bellen et al., 2011a, 2011b). As fires generally originate from uplands, burning should be

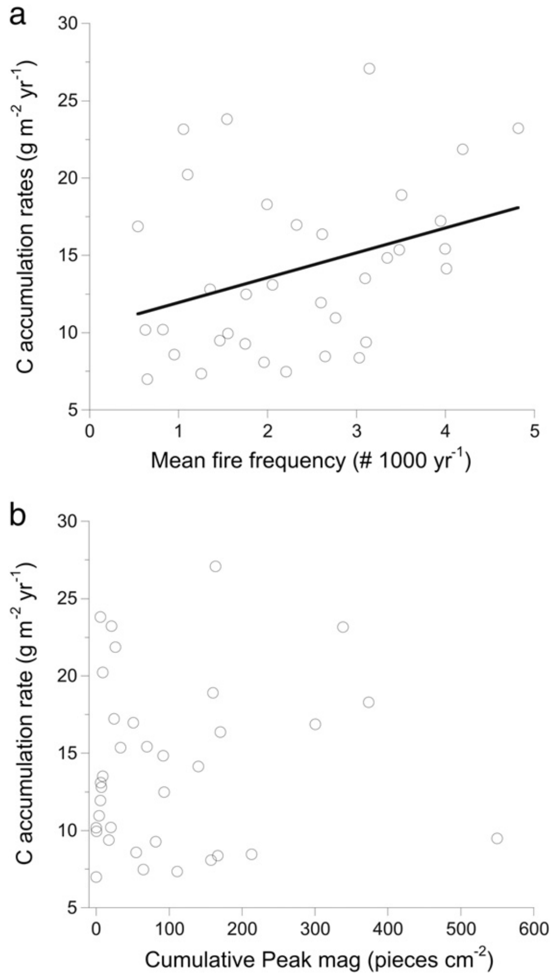

Figure 6. Regression analyses between fire regimes and C accumulation rates for 1000-yr mean values for each of the cores. a) C accumulation rates vs. mean fire frequency showing a weak positive relationship; $n=35$, $r^2=0.12$, $p=0.04$. b) C accumulation rates vs. cumulative peak magnitude, lacking a significant relationship.

more frequent near the forest–peatland boundary compared to the central parts (e.g. as visible in Fig. 1). Although lateral coring positions may thus be advantageous for fire reconstructions, these sections are also more sensitive to autogenic change because vegetation may be less resilient here than in central sections of the peatland (Bhatti et al., 2006; Bauer et al., 2009). In addition, the forest–peatland boundary, forming an ecotone of varying width, may be more frequently exposed to (episodic) minerotrophic input, paludification and concomitant changes in vegetation. Thus, fire effects on C accumulation may well have been minimized by a more important local effect of autogenic factors.

Future perspectives on Eastmain peatland fire potential

Holocene peatland fire regimes have not been a determinant factor in long-term C sequestration in the Eastmain region, and this may well be valid for most open peatlands in boreal eastern Canada. Nevertheless, climate–fire–C cycling interactions may be variable in the light of present and ongoing climate change (Bergeron et al., 2010). Peatland fire potential is linked to fire occurrence in the adjacent uplands and therefore forest-fire regime projections should be considered for estimations of future fire regimes. Boreal Quebec climate projections, as modeled by the Canadian Regional Climate Model (A2 scenario), show increases in summer temperature of 2.0–2.5°C and ~10%

increases in summer precipitation around 2050 relative to 1980 in boreal Quebec (Plummer et al., 2006). As a result, fire regime scenarios for the Waswanipi region (~200 km south of the Eastmain region) indicate an increase in annual area burned of 7% and an increase in monthly fire risk of 30%, attaining 70% in July and 100% in August by 2100 (Le Goff et al., 2009). These increases in upland fire activity suggest a higher potential for peatland fire (Flannigan et al., 2009), yet climate may have a different influence on burning potential in large, wet and open peatlands. A trend of higher water tables and reduced drought frequency since the end of the last Little Ice Age episode (~AD 1850) in both peatlands and forest stands has been observed in northern Quebec (Bergeron and Archambault, 1993; Lesieur et al., 2002; Payette and Delwaide, 2004; Payette et al., 2004; Arlen-Pouliot and Bhiry, 2005; Loisel and Garneau, 2010; van Bellen et al., 2011b). Following these results, climate projections for peatland fire dynamics remain uncertain in northeastern Canada. This contrasts with scenarios for western continental Canada, where peatlands presently persist at the dry climatic end of their global distribution. A projected major increase in fire activity may cause western Canadian peatlands to switch from net C sinks to sources (Wieder et al., 2009).

Conclusion

Holocene reconstructions show that the studied fire regime aspects probably have not been important enough to have driven variations in long-term C sequestration in boreal peatlands of the Eastmain region. The mid- and late-Holocene show important spatial and temporal variability in fire-frequency trends, implying a strong intra-peatland control on fire regimes, while rates of C accumulation decreased to a minimum of ~10 g m^{-2} yr^{-1}. Charcoal peak magnitudes increased significantly after 1400 cal yr BP, possibly the result of the short-term occurrence of dry peatland surface conditions. In order to cover the important variability in fire records, the use of multiple records from multiple peatlands appears essential for a complete comprehension on driving factors of long-term fire regimes.

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