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Effects of vegetation zones and climatic changes on fire-induced atmospheric carbon emissions: a model based on paleodata

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Abstract. An original method is proposed for estimating past carbon emissions from fires in order to understand long-term changes in the biomass burning that, together with vegetation cover, act on the global carbon cycle and climate. The past carbon release resulting from paleo-fires during the Holocene is examined using a simple linear model between measured carbon emissions from modern fires and sedimentary charcoal records of biomass burning within boreal and cold temperate forests in eastern Canada (Quebec, Ontario). Direct carbon emissions are estimated for each ecozone for the present period and the fire anomaly per kilo annum (ka) *v.* present day (0 ka) deduced from charcoal series of 46 lakes and peats. Over the postglacial, the Taiga Shield ecozone does not match the pattern of fire history and carbon release of Boreal Shield, Atlantic Maritime, and Mixedwood Plains ecozones. This feature results from different air mass influences and the timing of vegetation dynamics. Our estimations show, first, that the contribution of the Mixedwood Plains and the Atlantic Maritime ecozones on the total carbon emissions by fires remains negligible compared with the Boreal Shield. Second, the Taiga Shield plays a key role by maintaining important carbon emissions, given it is today a lower contributor.

Additional keywords: air masses, biomass burning, boreal forest, Canadian vegetation ecozone, charcoal database, modelling, paleo-fires.

Introduction

During the current interglacial period between 7000 and 1000 calibrated years before present (hereafter cal BP), the cumulative release of carbon (C) into the biosphere has increased by 195×10^9 t (195 000 Tg) (Indermühle *et al.* 1999), which far exceeds the natural range prevalent over the last 800 000 years (Lüthi *et al.* 2008). Several processes might explain this recent increase, e.g. transformations in tropical vegetation cover or changes in sea-surface temperatures (Indermühle *et al.* 1999; Ewen *et al.* 2004). However, even if the tropical forests of Africa have contracted significantly since 6000 cal BP, over the same period total forest biomass has significantly increased in the South American tropics in response to wetter conditions (Mayle *et al.* 2000; Behling 2002). Thus, any reduction of carbon sources due to the depletion of vegetation in Africa would be compensated by increases in South America (Mayle and Beerling 2004). Because biomass combustion is one of the most important elements in carbon flux processes (Seiler and Crutzen 1980; van der Werf *et al.* 2004), we hypothesise that millennial changes in fire activity might

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influence the global carbon cycle and thus the climate system itself (Carcaillet *et al.* 2002).

Boreal forest, which is the largest forest biome in mid- and high latitudes (Dixon *et al.* 1994), acts as sink or source of carbon depending principally on the occurrence of fire. Surprisingly, very few attempts have been made to estimate the past rate of carbon release into the atmosphere through paleo-biomass burning, either by mechanistic models or by proxy-based reconstructions at regional or global scales (Bowman *et al.* 2009).

Field observations have shown close interconnections to exist between fire, the global climate and the carbon cycle. For example, the Indonesian fires of 1997 and 1998 released an amount of carbon equivalent to ~13–40% of the average global annual load of fossil carbon normally released for energy supply (Page *et al.* 2002). Moreover, Patra *et al.* (2005) have shown that natural and anthropogenic biomass burning constitute the major component of the land–atmosphere carbon flux. Current global warming could increase the frequency of fire (Stocks *et al.* 1998; van der Werf *et al.* 2004; Flannigan *et al.* 2005), so adding further atmospheric CO_2 into the global carbon cycle, and so affect the climate in a system of positive feedbacks (Bowman *et al.* 2009).

In this paper, we present an original approach for estimating carbon emissions from fires into the atmosphere during the Holocene. The method is based on a dataset derived from sedimentary charcoal sampled from small lakes and peats (<<10 ha) in eastern Canada. The charcoal accumulation rate is an appropriate proxy for biomass burning (Clark *et al.* 1996). The amount of CO_2-C released by past fires is estimated by comparison with known rates of carbon release by modern fires, and known events of biomass burning in recent and past times. Eastern Canada is a region where fires occur frequently (Bergeron *et al.* 2004; Flannigan *et al.* 2005) and provides ideal conditions in the form of a high density of charcoal series covering these regional biomes (Carcaillet and Richard 2000). Furthermore, the high latitudes of this region are among those areas that are most likely to be significantly affected by global warming (Overpeck *et al.* 1997) and are the least affected by anthropogenic activities (Marlon *et al.* 2008). Globally, boreal forest has sequestrated 559 Gt C, i.e. ~1/3 of the total stored terrestrial carbon (Apps *et al.* 1993). The transfer of this carbon from soils to the atmosphere would trigger a final loss, given the millennia required to sequestrate such stocks.

We analyse here data from the last 6000 years, i.e. after or very close to the ultimate collapse of the Laurentide ice-sheet ~7000 years ago (Lauriol and Gray 1987; Dyke *et al.* 2003) since when lake sediments have been able to accumulate in eastern Canada. Moreover, pollen investigations have indicated that emplacement, physiognomy and structure of the major vegetation zones in the Quebec–Labrador peninsula have not changed over the last 6000 years (Richard 1995; Williams *et al.* 2000); they consequently provide a useful set of boundary conditions that are necessary when assessing the variability of carbon emissions. Thus, in the present paper the discussion focusses on carbon emissions at 6000 and 3000 cal BP, and recent, modern-day emissions, which can be considered as a reference point of current carbon emissions from wildfires.

Material and methods

Data sources

Lacustrine sedimentary charcoal is a good proxy for estimating the past occurrence of fires, which can be scaled to the surrounding local or regional environment depending on the size of particle measured: e.g. charcoal particles >>150 μm represent local fires (Higuera *et al.* 2007), whereas smaller charcoal (<100 μm) includes both local and regional fires (Tinner *et al.* 1998; Carcaillet *et al.* 2001*b*). Even where sedimentary charcoal data have not been recorded by the same method (e.g. in terms of parameters relating to fragment size, surface area or number), it has been shown that they still provide the same fire signal, and can therefore be used in large-scale analyses (Ali *et al.* 2009*b*).

The charcoal fragments are extracted from sediments by physicochemical methods, and then counted under a microscope. The data structure of the eastern Canadian charcoal database is described in Carcaillet and Richard (2000), together with other low- and high-resolution series from Fuller (1997), Carcaillet *et al.* (2002), Simard *et al.* (2006), Ali *et al.* (2009*a*), and some unpublished charcoal chronologies from McGinnis Lake in south-eastern Ontario, and Lac Amont and Lac Aval in northern Quebec. All ^{14}C measurements have been published in the original studies. However, the age–depth models were all recalculated using a simple smoothing equation or better, with a polynomial when the sets of ^{14}C measurements were of high quality. All data are available on the Global Charcoal Database (http://www.bridge.bris.ac.uk/resources/Databases, accessed 5 November 2010).

We used data series from 46 lakes distributed throughout Quebec and conterminous Ontario (Fig. 1). The data were sampled from sites that were representative of the four major vegetation types or ecozones that correspond to the Canadian vegetation classification edited by the Canadian Committee on Ecological Land Classification (Wiken 1986; SISCan 2008), and which are used in the fire carbon emission database (e.g. Amiro *et al.* 2001, 2009). The four ecozones (Table 1) that occur in our studied area were as follows: the Mixedwood Plains, which are a mixed forest zone dominated by broadleaf deciduous tree-species; the Atlantic Maritime, composed of mixed forests dominated by evergreen needle-leaf trees; the Boreal Shield ecozone, which is represented by closed forests of evergreen needle-leaf trees; and the eastern Taiga Shield, composed of lakes, wetlands and open evergreen or deciduous needle-leaf forests interwoven with shrublands (ericaceous, dwarf willows and birches) and meadows more typical of the Arctic tundra. The forest stands form lichen woodlands that merge into areas of open Arctic tundra. Broadleaf trees are more widely distributed towards the south; needle-leaf trees are more widely distributed towards the north.

The Quebec climate 6000 and 3000 cal BP

Six thousand years ago, temperatures in northern Quebec were colder than at present owing to the proximity of the remaining ice sheets (Kerwin *et al.* 2004). However, in southern Quebec temperatures were between 0.2° and 0.5°C higher than today, seasonality was more pronounced and mean annual

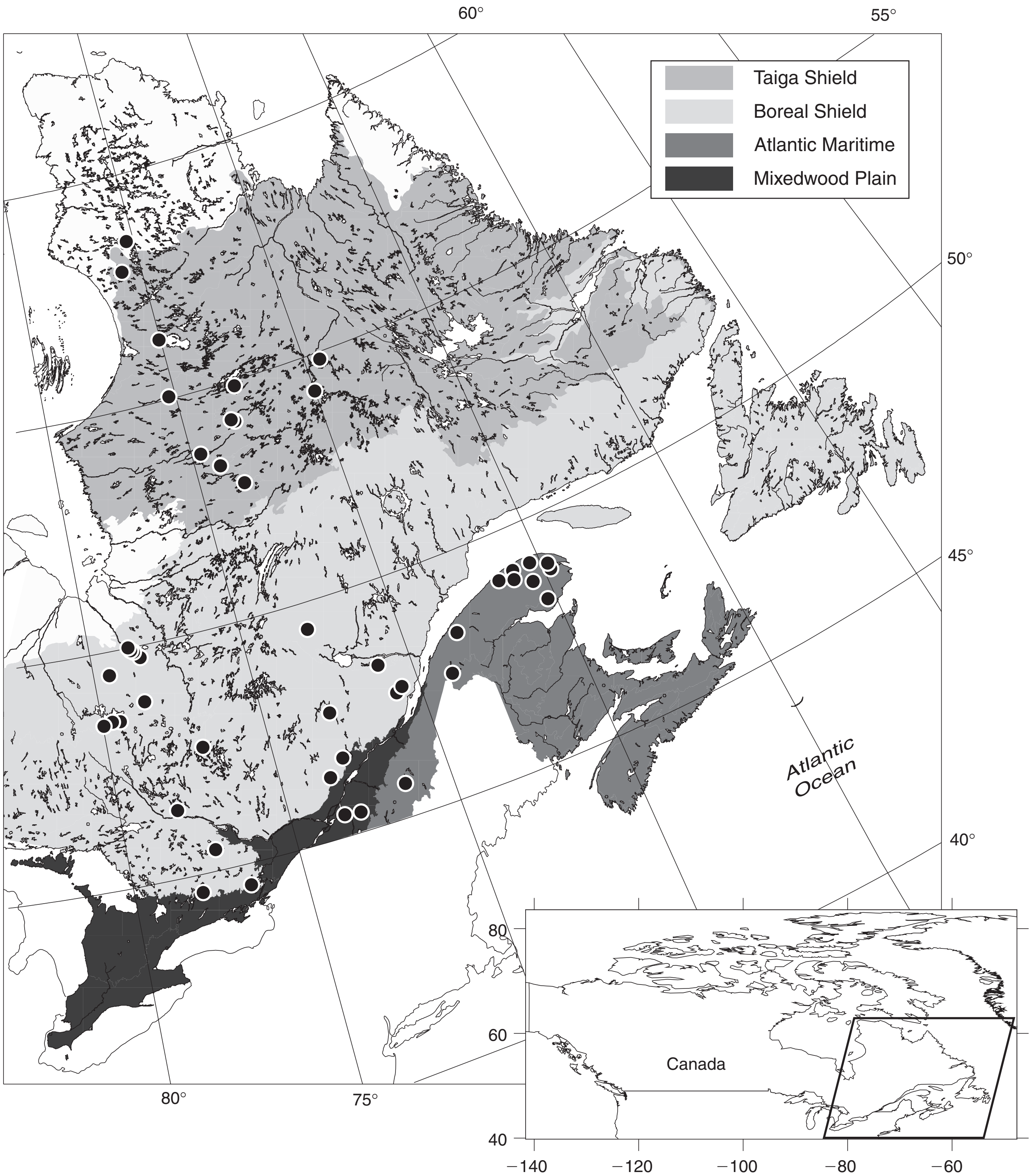

Fig. 1. Locations of the 46 lakes sampled for sedimentary charcoal, plotted on a map of the Canadian ecozone classification (modified from Ecozones of Canada, edited by the Department of Agriculture and Agri-Food Canada; Soil Landscapes of Canada Working Group 2007).

precipitation 6000 years ago was lower (<900 mm) than it is today, >1000 mm (Muller *et al.* 2003). During the late Holocene, summer cooling occurred in eastern Canada and the net precipitation budget was higher (Payette and Filion 1993; Yu *et al.* 1996; Lavoie and Richard 2000; Moos *et al.* 2009), but summer droughts were probably more frequent, resulting in increased fire occurrence (Payette and Gagnon 1985; Carcaillet and Richard 2000; Power *et al.* 2008).

Table 1. Description of vegetation zones, main tree composition, modern C emissions, number of lakes and kilo annum (ka) sequences

Vegetation zone (Richard 1995)	Ecozone (Wiken 1986) areas (SISCan 2008)	Forest description (functional physiognomy)	Dominant and characteristic tree species	Modern C emissions (Amiro *et al.* 2001, 2009) (Tg C $year^{-1}$)	Number of lakes investigated	Sequences (ka) potential/available during the first 6 ka
Treed-tundra and lichen woodland	Taiga Shield East (882 506 km^2)	Open evergreen Needle-leaf trees Shrubby woodland	*Picea mariana*, *Larix laricina*, *Pinus banksiana*	1.487	12	72/69
Boreal coniferous	Boreal Shield East (1 224 287 km^2)	Closed evergreen Needle-leaf trees Moss understorey	*Picea mariana*, *Betula papyrifera,* *Larix laricina*, *Pinus banksiana*		16	96/93
Mixed boreal		Evergreen closed Mixed needle-leaf Broadleaf trees	*Abies balsamea*, *Picea glauca*, *Picea mariana*, *Betula papyrifera*, *Populus tremuloides*, *Thuja occidentalis*, *Pinus banksiana*, *Pinus resinosa*, *Populus balsamifera*	1.640		
	Atlantic Maritime (287 703 km^2)	Evergreen closed Mixed needle-leaf and broadleaf trees	*Abies balsamea*, *Picea rubrum*, *Picea glauca*, *Betula alleghaniensis*, *Betula papyrifera*, *Acer saccharum*	0.186	11	66/66
Mixed temperate	Mixedwood (169 066 km^2)	Deciduous closed forest Broadleaf trees	*Acer saccharum*, *Fagus grandifolia*, *Betula alleghaniensis*, *Acer rubrum*, *Tilia americana*, *Carya cordiformis*, *Tsuga canadensis, Pinus resinosa*, *Pinus strobus*	0.016	7	42/42

Past carbon emission

Although some studies have estimated carbon emissions from fires in Canada, only one, Amiro *et al.* (2001), has given estimates for the four ecozones occurring in eastern Canada. The overall range in the mean annual carbon emission reported by Amiro *et al.* (2001) is between 22 and 33 Tg C year^{-1}. For Boreal North America, including Alaska, French *et al.* (2000) calculated an average weighted emission of 53 Tg C year $^{-1}$, whereas van der Werf *et al.* (2006) obtained an estimate of 44 Tg C year^{-1}.

Estimates of current carbon emissions from vegetation fires (Amiro *et al.* 2001, 2009 in Table 1) are based on a database of large fires in Canada, recorded over the period 1959–99 for all the Canadian ecozones. Amiro *et al.* (2001, 2009) report differences in the occurrence of fires across the Boreal Shield and Taiga Shield ecozones due to the strong east–west moisture gradient (east being wetter) that can affect certain properties of the forest environment. The Taiga Shield ecozone is composed of the Taiga Shield West and the Taiga Shield East with Hudson Bay as the divider, which also splits the Boreal Shield ecozone at the northern tip of Lake Superior into the Boreal Shield West and Boreal Shield East ecozones. These are henceforth used as the reference ecozones on which estimates of modern carbon emissions are based.

Standardising charcoal data and computing C emissions

Data with appropriate ^{14}C chronologies were converted into terms of charcoal influx (CHAR for CHarcoal Accumulation Rate). In the analyses, we only included data from lakes that have a sediment chronology based on at least three ^{14}C measurements dates per site (average of 6.2 ± 2.3). Detailed chronologies are presented in previous published data (see *Data sources*).

Each CHAR series is putatively related to the burned biomass and C emissions from fires within the catchment area of the lake. Because of the broad range of record types, site characteristics, and methodological and analytical techniques, the relationship between C release and CHAR values varies greatly among sites, both in scale and shape. For instance, a similar amount of CO_2 released per unit surface area may correspond to different CHAR values in two different records: this induces mean CHAR values ranging from 0.011 to 21.8 mm^2 cm^{-2} year^{-1} among the 46 records. Moreover, a doubling of C release indices is very likely to induce increases of different magnitudes among records. CHAR data therefore need homogenisation before comparisons are made. Power *et al.* (2008) addressed the problem of record comparison by performing two successive transformations independently on each record: a Box–Cox transformation to normalise CHAR frequency distributions, then a rescaling of the transformed data to a reference period. In our case, the underlying hypothesis is that a non-linear relationship, specific to each record, links a normally distributed C release value to CHAR values. Normalising CHAR data then induces a linear relationship between the transformed CHAR values and C emission. Rescaling these values then allows temporal changes in the transformed CHAR values to be identified with changes in C emissions. Mathematically, this is expressed as follows:

1. Normalisation of CHAR frequencies. Let c_i be the ith CHAR value of a particular site. The Box–Cox transformation of CHAR data involves creating a series of transformed CHAR values c_i^*:

$$c_i^* = \begin{cases} \frac{(c_i+\alpha)^\lambda - 1}{\lambda} & \text{if } \lambda \neq 1 \\ \log(c_j + \alpha) & \text{if } \lambda = 1 \end{cases}$$

where α is a small positive constant that avoids singularity when $c_i = 0$ and $\lambda = 0$ (here taken equal to 1% of the smallest non-null c_i), and the transformation parameter λ is selected so that $[c_i^*]$ frequency distribution is approximately normal. The λ value is estimated by maximum likelihood following Venables and Ripley (2002).

2. Rescaling of the transformed data. We consider that the transformed CHAR c_i^* is now commensurate with the amount of burned biomass and C emissions, and that the minimum value of c_i^*, $\min(c_i^*)$, corresponds to a period with approximately no emission. Then, a series e_i of rescaled transformed charcoal influx is defined as

$$e_i = \frac{c_i^* - \min(c_i^*)}{\overline{c_{1000-0}^*} - \min(c_i^*)}$$

and estimates the C emission as a proportion of C emitted in a standard period 1000–0 cal BP, where $\overline{c_{1000-0}^*}$ is the average value over the period 1000–0 cal BP. Including Power *et al.* (2008) intermediate computation of Z scores over a given reference period (i.e. 4000–0 cal BP) would not change the final result but was less appropriate with a shorter time period (e.g. 6000 cal BP) and less variation in the length of the charcoal records within eastern Canada than over world records.

For each site, mean e_i values were computed for 1000-year intervals (1500–500 cal BP, 2500–1500 cal BP, etc.). The average proportional increase or decrease of emitted CO_2 by each ecozone, compared with the standard period 1000–0 cal BP, was then expressed as the median value of all sites belonging to the particular ecozone in question.

Results

General trend

Biomass burning, as illustrated by the rescaled, transformed charcoal influx, showed similar temporal patterns over the last 11 000 years for the three southern ecozones of eastern Canada: i.e. the Mixedwood Plains, the Atlantic Maritime and the Boreal Shield (Fig. 2). During the ultimate regional deglaciation, fire activity increased between 11 000 and 9–8000 cal BP, with a long period of afforestation in the southern part of the study area. Biomass burning decreased from 10 000 to 7000 cal BP in the Atlantic Maritime ecozone, whereas a similar decrease in the Mixedwood Plains and the Boreal Shield East was delayed to a period from between 9000 and 6000 cal BP (Fig. 2). The lowest values were obtained for 6000 cal BP. The rescaled transformed charcoal influx then rose progressively from 5000 cal BP to the present day in the Atlantic Maritime, but was greater than modern values at 4000 and 3000 cal BP in the Mixedwood Plains and the Boreal Shield East. The Atlantic Maritime showed low rescaled transformed charcoal influx between 8000 and 1000 cal

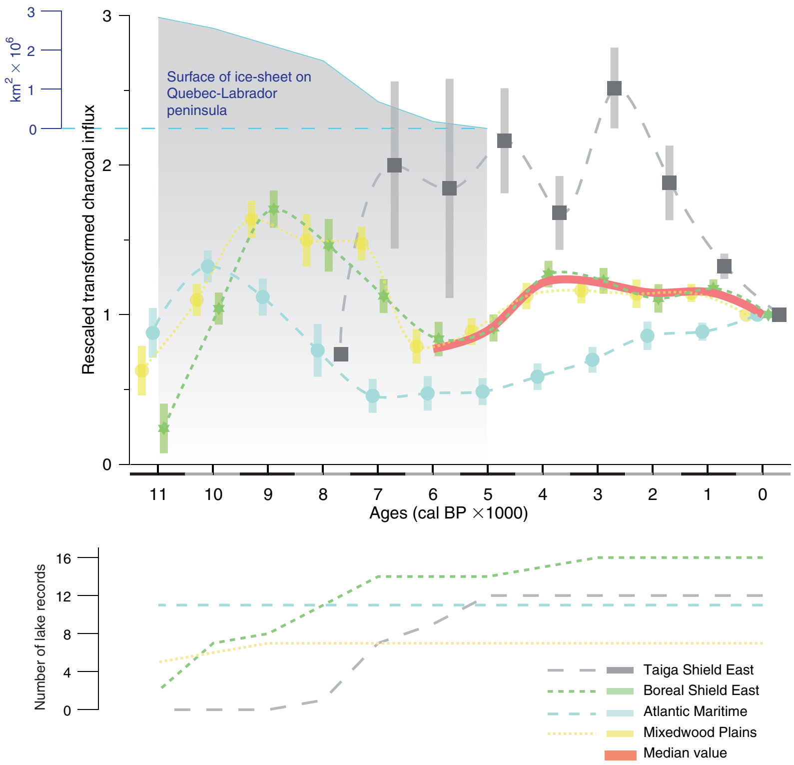

Fig. 2. Values of transformed charcoal influx for each ecozone of Quebec. Mean values are calculated at 1000-year intervals for each lake record and per ecozone, rescaled on the period 1000–0 cal BP (calibrated years before present). The median values are also calculated for the four ecozones. Error bars (vertical) display standard error. The grey area indicates the period when the major vegetation zones were different from the modern vegetation (Richard 1995; Williams *et al.* 2000) and when the melting of the Laurentide ice-sheet was not complete. The estimates of the ice-sheet areas for eastern Canada were assessed from geomorphological maps (Dyke *et al.* 2003) analysed with a GIS (Geographic Information System) and corrected in calibration years. The number of lakes used to calculate rescaled transformed influx through time is represented for each ecozone.

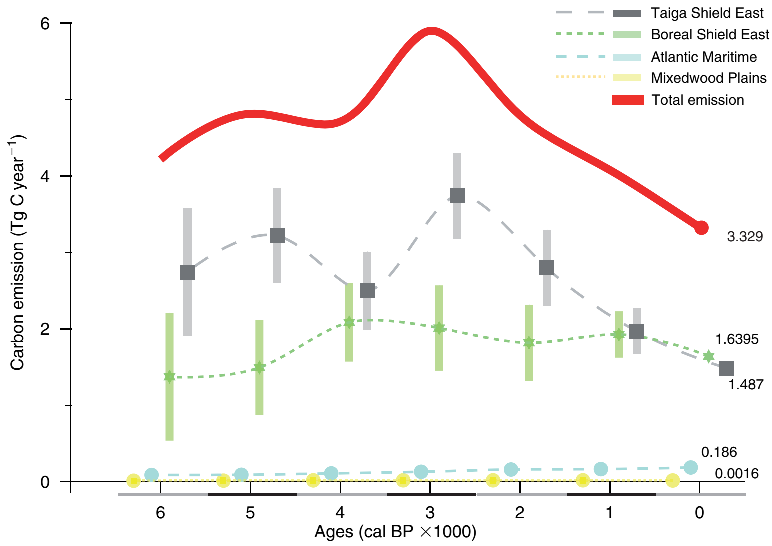

Fig. 3. Estimated proportion of emitted CO_2 (Tg C year^{-1}) by each ecozone compared with the standard period expressed as the median value for all sites belonging to the ecozone under consideration (modern C emissions are from Amiro *et al.* 2001, 2009). Error bars (vertical) display standard error. The red curve shows the total cumulative carbon emissions for the four ecozones.

BP, suggesting less fire activity in this ecozone during that period.

In the northern sites such as in the Taiga Shield, biomass burning displays a different pattern, including a time-lag with respect to the southern ecozones (Fig. 2). Biomass burning increased rapidly between 8000 and 7000 cal BP (data for only one lake at 8000 cal BP and there are no data available before 8000 cal BP owing to the glacial environment then prevalent). However, this increase occurred 2000 or 3000 years after the maximum observed in the three southern ecozones. In the Taiga Shield, the fire activity remained high before 3000 cal BP, whereas it was low at 6000 cal BP in the three southern ecozones: i.e. the Mixedwood Plains and the Boreal Shield. The pattern exhibited by the Taiga Shield data is characterised by a very high level of intersite variability, as evidenced by the 50% confidence interval.

The regional median trend calculated over the period between 6000 and 0 cal BP for the four ecozones does not vary considerably over that entire period, but the medians were below modern values 6000 cal BP, and above them between 4000 and 3000 cal BP (Fig. 2). No median value was calculated before 6000 cal BP because of the different environmental conditions that prevailed in the four ecozones from 11 000 to 6000 years ago, including the late deglaciation processes and the type of vegetation then present, which is not comparable with modern types. The trend exhibited by the median values matches the Boreal Shield East trend quite well (Fig. 2), although the number of investigated lakes and the number of available millennium sequences for the Boreal Shield do not dominate the whole dataset: viz. 16 lakes for the Boreal Shield East *v.* 30 cumulated lakes for the three other ecozones, and 96 millennium sequences available from 6000 cal BP to the present *v.* 180 cumulated millennium sequences respectively (Table 1).

Carbon emissions

Total carbon emissions are estimated only for the period 6000 cal BP to the present (Fig. 3), because the vegetation before 6000 cal BP differed owing to post-glacial dynamics (Richard 1995). In both the Atlantic Maritime and the Mixedwood Plains ecozones, the estimates of carbon emissions (Fig. 3) remained relatively constant throughout the Holocene from 6000 cal BP to the present, during which time only small amounts of carbon, similar to modern values (Table 1), were released from these ecozones. By contrast, carbon emissions released in the Taiga Shield and the Boreal Shield were significantly different from the modern rates. Today, the Taiga Shield and the Boreal Shield emit almost equal amounts of carbon, but between 6000 and 5000 cal BP and at 3000 cal BP, the Taiga Shield emitted only 2/3 of the carbon in modern emissions. The total C-emission curve shows a humped-back pattern with a peak at 3000 cal BP due to high estimates for that time for both the Taiga and the Boreal Shield ecozones. The Taiga Shield played an important role during the last 6000 years, maintaining strong C-emission rates that ranged between 2.0 and 3.7 Tg C year^{-1}, which are 32 and 152% above modern emission rates respectively. After the maximum of 3000 cal BP, the C-emission rate of the Taiga Shield has decreased to reach modern values. The equivalent C emissions for the Boreal Shield ranged between 1.6 and 2.0 Tg C year^{-1}, and exhibited an increasing trend between 6000 and 4000 cal BP, followed by a gentle decrease to reach modern values (Fig. 3). Between 6000 cal BP and the present day, the Boreal Shield minimum is estimated to have occurred 6000 years ago, whereas the Taiga Shield minimum is its modern rate, which has a strong influence on the total C-emission curve.

Discussion

Our study indicates that historical patterns of carbon released into the atmosphere by biomass burning can be assessed. This has been possible owing to: (i) the large set of charcoal data that has been gathered from across a wide territory, and that is publicly available and regularly enriched by new data (Power *et al.* 2008); and (ii) the modern large-scale studies of carbon released by fires that have been conducted over wide regions, and that have taken into account the different vegetation zones, and the variation in fuel quality and landscape structure (e.g. French *et al.* 2000; Amiro *et al.* 2001, 2009). The following discussion highlights the importance of the type of vegetation zone and climatic change on the millennial-scale dynamics of carbon transfer into the atmosphere, and concludes with a budget that compares our findings with global trends.

Biomass burning activity and vegetation pattern

Compared with previous reconstructions (Carcaillet and Richard 2000; Carcaillet *et al.* 2002; Power *et al.* 2008), the fire history reconstruction presented here indicates that the entire region cannot be considered as an homogeneous unit, but one that includes boreal biomes that are composed of the Atlantic Maritime, the Boreal Shield and the Taiga Shield (Table 1; Fig. 2). Previous studies have shown that biomass burning was lowest between 6000 and 4000 cal BP; our study similarly indicates activity to have been at a minimum between 7000 and 5000 cal BP. This slight difference in dates may be partly due to the fact that previous reconstructions combined data from the Quebec–Labrador peninsula, without taking into account the vegetation differences. Surprisingly, differentiating between the Boreal Shield and the Taiga Shield highlights some major functional differences. The Boreal Shield, the Mixedwood Plains and the Atlantic Maritime all show more or less the same pattern, which matches previous reconstructions, with a minimum between 6000 and 5000 cal BP, and a higher level of fire activity between 10 000 and 7000 cal BP or between 4000 cal BP and the present depending on the ecozone. The Boreal Shield East gave the highest amount of biomass burning in two periods: one at ~8000 cal BP and another between 4000 and 1000 cal BP (Fig. 2). This is partly linked to the post-glacial vegetation history, especially to the late establishment of vegetation in the Taiga Shield before 6000 cal BP. The residual Laurentide Ice Sheet that persisted in the central area of the Quebec–Labrador peninsula (Dyke *et al.* 2003) created unfavourable conditions for colonisation by woody species (Richard 1995).

The historical pattern of charcoal influx suggests that the Taiga Shield East ecozone would have been the main contributor of carbon into the atmosphere in the region during the early and the late Holocene, whereas the Boreal Shield would have been an important source only for the last 4000 years. However, this direct interpretation does not give sufficient weight to the irregular distribution and concentration of sites in each ecozone when computing charcoal values; neither does it fully account for differences in carbon released by fires in each ecozone, or the exact surface area of each ecozone through time. For instance, recent fire histories assessed by dendrochronology highlight that the fire return interval or the fire cycle can differ tremendously within an ecozone (Gauthier *et al.* 2000; Bergeron *et al.* 2001; Le Goff *et al.* 2007; Bouchard *et al.* 2008). More sedimentary charcoal needs to be investigated at high resolution in sites from the Labrador and the Maritime Provinces of New Brunswick, Nova Scotia, and Newfoundland in order to improve our estimates of the relative amounts of carbon linked to past sources of biomass burning in the Taiga Shield, the Boreal Shield and the Atlantic Maritime ecozones.

The history of biomass combustion in the Taiga Shield East is not well represented in our analyses because only a few sites in that ecozone were investigated at high resolution, all of which were located in the western part of the study area, with none in either central Quebec or Labrador (Fig. 1). A similar argument holds for the north-eastern part of the Boreal Shield. The lack of data in these regions might therefore result in an overestimation of the C flux into the atmosphere because the influence of moist air masses from the Atlantic Ocean (maritime Arctic) should decrease the frequency of fires (Bouchard *et al.* 2008), and thus the rescaled transformed CHAR values.

Carbon release and vegetation dynamics

During the last 6000 years, the maximum value of carbon emissions was reached at 3000 cal BP with ~5.9 Tg C year^{-1}, 77% more than the modern value, which stands at 3.3 Tg C year^{-1}. The minimum value, which was ~13% above the modern emission rate, occurred at 6000 cal BP (Fig. 3). These temporal variations are partly due to differences in biomass composition, different ecozone histories (Fig. 2), and differences in the rates of carbon released by fires (French *et al.* 2000; Amiro *et al.* 2001, 2009), which are in turn linked to diverse ecosystem properties such as fuel quality and quantity (Hély *et al.* 2001). The two northern ecozones (Boreal and Taiga Shield East) are more fire-prone, being mainly composed of needle-leaf tree species. They also cover the largest area, whereas the Atlantic Maritime is dominated by the same set of species as the Boreal Shield, and the Taiga Shield (Table 1) is limited to a very small area along the Atlantic coast from Quebec to the maritime provinces of Canada. This area is therefore wetter owing to the influence of the oceanic climate, which in turn affects vegetation composition and fuel quality. The area of the Mixedwood Plains is also very limited at the southern edge of the study area, being mainly composed of broadleaf species, which are not as flammable as coniferous vegetation (Hély *et al.* 2000). Furthermore, even if the vegetation zone were already established 6000 cal BP (Richard 1995; Williams *et al.* 2000), the vegetation within the different biomes would continue to be dynamic. For instance, the Boreal Shield experienced a gentle and progressive decrease in the abundance of both *Pinus strobus* and *Thuja occidentalis*, and an increase in the abundance of *Picea glauca*, *P. mariana*, *Pinus banksiana* and *Betula papyrifera*. This process occurred both in the southern (Richard 1980; Liu 1990) and the northern part of the Boreal Shield (Garralla and Gajewski 1992; Ali *et al.* 2008), and in the Atlantic Maritime (Marcoux and Richard 1995; Asnong and Richard 2003), but it varied in magnitude and timing depending on the specific latitude within an ecozone (Carcaillet *et al.* 2001*a*). There was a significant shift in vegetation cover ~3000 cal BP, with plant cover being more typical of the boreal after 2000 cal BP, and being richer in southern species before 4000 cal BP. The Taiga Shield vegetation was probably denser before 3000 cal BP and some fire-prone species might have expanded their distribution range when fire frequency increased (Payette 1993; Richard 1995). These few variations within the vegetation of the main ecozones during the last 6000 years could have generated some error in the estimation of carbon emissions but were difficult to take into account. Moreover, carbon emissions before 6000 cal BP cannot yet be estimated owing to highly variable boundaries to ecozones, and difficulties in their identification according to modern vegetation types. Estimates of the vegetation extant before 6000 cal BP will be addressed in future work using vegetation modelling techniques such as the modern analogue method based on squared-chord distances, which aim to quantify the probability that fossil pollen assemblages resemble modern assemblages from North American vegetation (e.g. Gavin *et al.* 2003).

Carbon release and climatic change

Although temperatures oscillated somewhat from 11 000 to 8000 cal BP, the overall climate warmed, and caused the boreal-type vegetation (*Picea*, *Pinus*, *Betula*, *Alnus*) in the southern part of the study area to expand, until broadleaf temperate species dominated 7000 years ago in the St Lawrence lowlands (Richard 1994; Anderson *et al.* 2007), an area that more or less corresponds to the area of the Mixedwood, the Atlantic Maritime and the southern part of the Boreal Shield ecozones. This climatic change largely explains the increasing biomass burning observed from 11 000 to 8000 or 7000 cal BP in the reconstruction of the Mixedwood, Atlantic Maritime and Boreal Shield ecozones (Fig. 2).

In response to the late warming over eastern Canada (Kerwin *et al.* 2004; Viau and Gajewski 2009), the northern part of the Boreal Shield ecozone reached its geographical limit ~6000 years ago, whereas the Taiga Shield continued to expand until 3000 cal BP (Payette and Lavoie 1994; Richard 1995). This late expansion of needle-leaf species (mainly *Picea mariana*, *Pinus banksiana* and *Betula*) largely explains the pattern of increasing biomass burning in the Taiga Shield ecozone from 7000 to 3000 cal BP, whereas the southern ecozones experienced their lowest activity between 7000 and 4000 cal BP (Fig. 2). The Taiga Shield climate is subarctic, which is colder and drier than the climate of southern ecozones, and is favourable to fire spread (Payette *et al.* 1989). The expansion of needle-leaf species, linked to regional warming within the northern part of the Quebec–Labrador peninsula (Viau and Gajewski 2009), thus contributed to the creation of fire-prone conditions and the large annual release of carbon during the middle Holocene in the Taiga Shield.

Three thousand years ago, carbon emissions reached a maximum (5.9 Tg C year^{-1}) in the Boreal Shield and Taiga Shield ecozones, which represents the highest level of historical carbon emissions reconstructed for eastern Canada (Fig. 3). This occurred at a time when the Taiga Shield experienced its warmest period during the Holocene (Viau and Gajewski

2009) and the Boreal Shield ecozone and Mixedwood Plains started to experience a wetter climate (Lavoie and Richard 2000; Muller *et al.* 2003). However, the increase in biomass burning (Fig. 2) cannot be explained only by changes in patterns of annual precipitation. Carcaillet and Richard (2000) suggested that although the mean annual climate was wetter, the late Holocene is characterised by more frequent summer droughts that favoured the ignition and spread of fire.

The minimum in past carbon release by fires (4.2 Tg C $year^{-1}$) occurred 6000 years ago. This period was considered as the best analogue for comparison with what is likely to occur at the end of the 21st century (Flannigan *et al.* 2001). This is based on the fact that, globally, orbital forcing delivers more heat to the ground during the summer because of variation in the level of insulation (Berger and Loutre 1991) and that this insolation effect, combined with a drier climate (Lavoie and Richard 2000; Viau and Gajewski 2009), should create suitable conditions for fires. However, owing to the late collapse of the Laurentide Ice Sheet in central Quebec (Dyke and Prest 1987), the regional climate was not significantly warmer 6000 cal BP (Kerwin *et al.* 2004; Viau *et al.* 2006), except in the southern lowlands covered by the Mixedwood Plains (Muller *et al.* 2003). Hence, although it was a time that experienced the most suitable orbital pattern for comparison with the end of the 21st century, it was not the most important time for carbon release by fires in eastern Canada.

Global contribution

Carbon release in eastern Canada has decreased over the last 2000 years (Fig. 3) following the decrease in large fires (Hély *et al.* 2010). This results from a long-term decline in biomass burning (Fig. 2) that has already been observed for the whole Northern hemisphere, combined with a general trend of global cooling (Marlon *et al.* 2008). Current global warming could therefore stimulate fire activity, even if precipitation increases (Girardin and Mudelsee 2008), and so increase the direct carbon flux into the atmosphere. However, future fire trends in boreal forests are difficult to predict globally owing to significant regional variation (Girardin *et al.* 2009).

The boreal North American zone, including Alaska, represents ~8% of the total area of the whole earth that is susceptible to burning, and currently contributes 9% of global carbon emissions per year (van der Werf *et al.* 2006). Although the modern Taiga Shield East is not a major source of carbon release, our study has shown that this ecozone played a major role during the late Holocene, especially as this ecozone underwent important structural changes during the last 2000 years (Payette and Lavoie 1994). It is thus likely that the North American boreal zone carbon emissions estimated for 3000 cal BP in Quebec, which were 80% higher than present emission levels, would have made a significant contribution to the global carbon budget. By comparison, the Taiga Shield East today contributes less than 5% of the carbon directly emitted by the Canadian forest, whereas the Taiga Shield West and the Taiga Plains represent 9 and 33% respectively (Amiro *et al.* 2009). Three thousand years ago, the Taiga Shield East would have released more carbon than the modern Taiga Shield West.

To improve the estimates presented in this paper, a better quantitative reconstruction of vegetation is needed. Furthermore, some areas such as Newfoundland, Labrador and east-central Quebec, where fires are strongly influenced by the oceanic climate, are currently poorly covered by charcoal series (Power *et al.* 2008). Similarly, central Canada from Ontario to Alberta is represented by too few high-resolution charcoal series with a good chronology to allow a large-scale reconstruction of the northern boreal biomes, similar to that carried out in the present study.

Conclusion

This study has proposed an original method based on sedimentary charcoal records for estimating past carbon emissions from biomass burning. However, our results may underestimate the true state of affairs because the data were compared with data of recent forest fires over the last 1000 years, which are characterised by a high level of variation of fire activity (Girardin 2007), and do not take into account anthropogenic effects on fire management during the 20th century. Additional charcoal data, which better represent recent decades, are needed to resolve this problem and to improve the calibration of the method. In the next phase of our research, we intend to extend our method to the rest of Canada and Alaska, and to include: (i) better estimates of the limits and extents of past ecozones, which can be derived from the extensive fossil pollen data available in this region of the world; and (ii) new high-resolution charcoal series to fill the gaps where series are currently missing.

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