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Abstract

The north-central Canadian boreal forest experienced increased occurrence of large and severe wildfires caused by unusually warm temperatures and drought events during the last decade. It is, however, difficult to assess the exceptional nature of this recent wildfire activity, as few long-term records are available in the area. We analyzed macroscopic sedimentary charcoal from four lakes and pollen grains from one of those lakes to reconstruct long-term fire regimes and vegetation histories in the boreal forest of central Northwest Territories. We used regional estimates of past temperature and hydrological changes to identify the climatic drivers of fire activity over the past 10,000 years. Fires were larger and more severe during warm periods (before *ca.* 5,000 cal. yrs. BP and during the last 500 years) and when the forest landscape was characterized by high fuel abundance, especially fire-prone spruce. In contrast, colder conditions combined with landscape opening (i.e., lower fuel abundance) during the Neoglacial (after *ca.* 5,000 cal. yrs. BP) were related with a decline in fire size and severity. Fire size and severity increased during the last five centuries, but remained within the Holocene range of variability. According to climatic projections, fire size and severity will likely continue to increase in central Northwest Territories in response to warmer conditions, but precipitation variability, combined with increased abundance of deciduous species or opening of the landscape, could limit fire risk in the future.

Keywords

large wildfires; charcoal; pollen; fire size; lake sediments; vegetation dynamics; climate change; Holocene

1. Introduction

The global climate is experiencing rapid warming conducive to increased occurrence of extreme weather events like heatwaves and droughts (IPCC, 2014; Mann et al., 2017). Such conditions may increase the occurrence of exceptionally large and severe wildfires due to the strong positive correlation between temperature and area burned (Abatzoglou and Kolden, 2013; Ali et al., 2012; Duffy et al., 2005; Gaboriau et al., under review; Kasischke and Turetsky, 2006; Turco et al., 2017). Particularly high temperatures and recurrent droughts during the last decade have led to severe wildfire seasons in forest ecosystems worldwide, for example in Australia (Boer et al., 2020; Nolan et al., 2020), Sweden (SFOR, 2018), Chile (De la Barrera et al., 2018), Greece and the USA (Smith et al., 2019), as well as in the Northwest Territories (NWT) in Canada where about 3.4 million ha burned during the summer of 2014 (Kochtubajda et al., 2019; NTENR, 2015).

Large wildfires in the NWT mainly occur from June to September. Lightning-induced fires are predominant and fire frequency and area burned are among the highest in the North American boreal forest (Erni et al., 2020; Veraverbeke et al., 2017). However, the 2014 wildfire season was exceptional in comparison with observations of the last decade (3.4 million ha burned in 2014, compared to a mean of 622 thousand hectares burned annually from 2009 to 2019; Gaboriau et al., under review). More than 380 fires were observed, some of which having burned more than 150,000 hectares. Moreover, firefighting costs reached more than CAD 56 million during this exceptional year, compared to an annual mean of CAD 7.5 million in the previous 20 years (NTENR, 2015).

Few remnant trees are left behind in areas burned during large and severe wildfires (Erni et al., 2017; Stephens et al., 2014). Recolonization is dominated by fast-growing pioneer deciduous tree species whose seeds disperse over long distances (Chapin et al., 2010; Walker et al., 2018). Large wildfires also reduce access to ecosystem services (Adams, 2013), cause health issues in human populations (Dodd et al., 2018), and decrease the capacity of Indigenous people to maintain traditional activities on the land (Berkes and Davidson-Hunt, 2006).

Since the 1950s, mean annual temperature increased by over 1.5°C in Canada, leading to an earlier onset of snowmelt across much of the boreal forest (Zhang et al., 2011). According to climate projections, temperature will continue to rise in the next decades, particularly in the north-central Canadian boreal forest (Price et al., 2013; Wang et al., 2014). Summer and fall precipitation are projected to increase in the boreal forest of Canada and annual precipitation could increase up to 20% by 2100. However, these conditions will not necessarily lead to increased soil and vegetation moisture, as projected warmer temperatures will lead to increased evapotranspiration (IPCC, 2014; Meehl et al., 2007). Hence, the drying effect of higher temperatures in the future will not necessarily be compensated by increased precipitation. Boreal forest landscapes will consequently be altered by a longer growing season (Jain et al., 2018), increased fire activity (Flannigan et al., 2013; Hassol, 2005; Wang et al., 2017; Wotton et al., 2017), higher proportion of deciduous tree species and more open forest canopies than currently (Baltzer et al., submitted; Boulanger et al., 2017; Chaste et al., 2019; Mekonnen et al., 2019). However, the robustness of the predictions of climate change impacts on wildfire regimes, particularly large wildfires, is limited by our lack of knowledge of the long-term interactions between climate, vegetation and fire (Ali et al., 2012; Conedera et al., 2009; Coogan et al., 2019). While the main drivers of fire activity have been documented in the NWT for the recent past (1965-2016; Gaboriau et al., submitted), few palaeoecological data are available at multi-millennial timescales.

To get insights into the long-term relationships between climate, vegetation and fire, we used high-resolution sedimentary proxies (pollen grains and charcoal particles) to reconstruct vegetation and fire histories for the past 10,000 years in the boreal forest of central NWT. We compared the reconstructed vegetation and fire histories with previous studies based on regional climatic reconstructions. Our objectives were (1) to determine whether the recent wildfire regime (last 500 years) is within or outside the Holocene range of variability and (2) to decipher the respective roles of climate and vegetation controls on past fire activity. We expected that the recent fire regime would be characterized by larger and/or more severe wildfires compared to the past 10,000 years in response to ongoing climate warming and recurrent development of severe drought episodes.

2. Material and methods

2.1. Study area and present-day vegetation

We sampled sediments from lakes *Emile*, *Izaac*, *Paradis* and *Saxon* (unofficial names) located north of Yellowknife, in central NWT, and 5-100 km distant from each other (Fig. 1). The sampled lakes have a mean elevation of 382 m (Table 1). According to the Canadian National Fire Database (CNFDB, http://cwfis.cfs.nrcan.gc.ca/ha/nfdb), time since the last fire varied between 4 and 39 years in the watersheds of the studied lakes (Table 1). The regional climate is dry continental, with long cold winters and short warm summers (Environnement Canada, 2017). The study area is located close to the treeline (Timoney et al., 2019) and is part of the Taiga Boreal Shield West ecoregion, characterized by rock outcrops and discontinuous permafrost (Olson et al., 2001). Vegetation is currently dominated by conifer tree species, mostly black spruce (*Picea mariana* (Mill.) B.S.P.). White spruce (*Picea glauca* (Moench) Voss.) is mainly found in the southwestern part of the study area, and jack pine (*Pinus banksiana* Lamb.) mostly occupies the southern part of the NWT, *i.e.* in the Taiga Plains. Broadleaf tree species are also present in younger stands having regenerated after fire, mostly paper birch (*Betula papyrifera* Marsh) and trembling aspen (*Populus tremuloides* Michx.), such as observed at lake *Emile* (Appendix S1A). The shrub layer is mainly composed of *Betula glandulosa* Michx., *Alnus alnobetula* (Ehrh.) K. Köch, *Andromeda polifolia* L., *Ledum palustre* L., and *Chamaedaphne calyculata* (L.) Moench. The ground layer is mostly composed of *Vaccinium vitis-idaea* L., *Rubus chamaemorus* L., *Kalmia* *angustifolia* L. and *Lycopodium annotinum* L., with abundant mosses (mostly *Sphagnum* spp.), while *Cladonia* lichens are found on xeric rock outcrops.

2.2. Past regional climate and vegetation

Previous studies have documented long-term regional climatic and vegetation changes in the northwestern and north-central Canadian boreal forest, which in turn had impacts on wildfire regimes. Before *ca.* 11,000 cal. yrs. BP (calibrated years before present), high solar radiation and summer insolation increased temperatures in the northwestern Canada (Viau et al., 2006), reduced snow cover, and led to the retreat of the Laurentide ice sheet (Dyke, 2005). Deglaciation occurred between *ca.* 10,000 – 9,000 cal. yrs. BP in the study area (Dalton et al., 2020) and left behind numerous lakes and swamps scattered across the landscape (Latifovic et al., 2017; MacDonald, 1995). Following ice retreat, the land was quickly colonized by tundra vegetation soon followed by trees once temperatures warmed during the early Holocene (Dyke, 2005). The Holocene Thermal Maximum (HTM) occurred in northern continental Canada during the mid-Holocene (between 7,300 and 4,300 cal. yrs. BP; Kaufman et al., 2004). During this period, major vegetation reorganization occurred in response to warmer temperature, such as treeline migration to the north and increased abundance of fire-adapted species (i.e. jack pine and black spruce), as observed in the NWT and southwestern Yukon (Gajewski et al., 2014; Pienitz et al., 1999; Sulphur et al., 2016). The decrease in temperature after 4,300 cal. yrs. BP, characterizing the Neoglacial period, favored coniferous species such as spruce as well as deciduous shade-tolerant tree species such as paper birch in permafrost-free areas (Moser and MacDonald, 1990; Sulphur et al., 2016). However, interactions between climate, vegetation and fire depend on regional atmospheric and oceanic oscillations, as observed in the eastern Canadian boreal forest where conifer species also increased in density after 3,500 cal. yrs. BP (Fréchette et al., 2018; Remy et al., 2017).

2.3. Sampling

We sampled lake sediments in June 2018 from the deepest point in each of the four lakes. The vegetation of the watersheds was mainly composed of recently burned conifer tree species and/or shrub birch having grown after the last fire (Appendix S1). The four lakes were small, deep, circular, and characterized by the absence of connections to surrounding watercourses (Table 1). We used a Kajak-Brinkhurst (KB) gravity corer to collect the most recently deposited material at the water-sediment interface (Glew 1991). We subsampled surface sediments on site in 0.5-cm thick sections stocked in plastic bags. We collected deeper sediments in 1-m cores using a 5-cm diameter Russian corer; we wrapped these cores in aluminum foil and placed them in hemicylindrical tubes for protection. We sliced the sediment cores in the laboratory into continuous 1-cm thick subsamples that we stocked in plastic bags and kept refrigerated at 4°C until analysis. Surface sediments were sampled at a shorter interval (0.5cm) than deeper sediments (1 cm) to ease comparison with recent (last 500 years) reconstructions of vegetation and fire history.

2.4. Radiocarbon dating and age-depth models

The basal 10 cm of the deepest sediment cores were mostly clay, corresponding to the beginning of lake sedimentation following deglaciation in the four sites, whereas the upper sediment was composed of a gyttja richer in organic matter. We constructed core chronologies based on radiocarbon dating of bulk gyttja samples by ^{14}C accelerator mass spectrometry (AMS). We used the '*WinBacon*' v.2.3.7 R package (Blaauw and Christen, 2011) to reconstruct Bayesian sediment accumulation histories and calibrate age-depth models constructed from seven dates for lakes *Emile*, *Paradis* and *Saxon* and nine dates for lake *Izaac* (Table S1). We used the IntCal13.14C calibration curve for terrestrial northern hemisphere material (Reimer et al., 2013). We interpolated ages at contiguous 0.5-cm depth intervals and all dates were expressed in calibrated years before present (cal. yrs. BP). We used linear interpolation between ^{14}C ages rather than best-fit curves, assuming that both methods yield equivalent chronologies (Blaauw, 2010; Trachsel and Telford, 2017).

2.5. Subsample chemical preparation and charcoal analysis

To distinguish charcoal from other biological material, we applied a common charcoal extraction protocol to each subsample using the 'chemical digestion' method (Winkler, 1985), based on charcoal resistance to chemicals (Mooney and Tinner, 2011). We removed a 1.4-cm^3 subsample from each liquid subsample of surface sediment (KB) and 1-cm^3 for the rest of the core (solid part), following Mustaphi et al. (2015). We took sediments in the central part of each core slice to limit the risk of contamination with modern material. We deflocculated and bleached the subsamples by placing them in a potassium hydroxide (KOH) solution combined with bleach and sodium hexametaphosphate – $(NaPO_3)_6$ – on a stirring table for 24 hours at room temperature; which later allowed us to differentiate black charcoal from bleached organic matter (Braadbaart and Poole, 2008; Swain, 1973). We wet-sieved the solution through a 160 μm mesh to collect larger charcoal particles produced by fire events having mostly occurred 0-3 km from the lakeshore (Oris et al., 2014). We sorted charcoal particles in a Petri dish, and measured their area using a camera (D-Moticam 1080, Motic Images, 2018) mounted on a binocular microscope and connected to an image-analysis software (MOTICAM IMAGE Plus 3.0).

2.6. Fire history reconstruction

We reconstructed burned biomass (hereafter *BB*; unitless) at each study site by measuring the cumulative area of charcoal particles per cm^3 for each subsample. Based on numerical age-depth models, we transformed this measure into charcoal accumulation rate (hereafter CHAR, i.e. $mm^2\ cm^{-2}\ yr^{-1}$), using the R package '*paleofire'* v.1.2.3 (Blarquez et al., 2014). We pooled and smoothed these series (using a 500-year window) by (1) rescaling initial CHAR values using min-max transformation, (2) homogenizing the variance using Box-Cox transformation, and (3) rescaling the values to Z-scores

(Power et al., 2008). The resulting values (cumulative charcoal area) are interpreted as representing the regional signature of fires having occurred in the watersheds of the studied lakes, and thus the regional burned biomass (hereafter *RegBB*; unitless).

We detected past fire events (Appendix S2) within each individual CHAR series using the CharAnalysis v.1.1 software (Higuera, 2009; available at https://github.com/phiguera/CharAnalysis). Following Brossier et al. (2014), we used the narrowest time window allowing us to obtain a median Signal-to-Noise Index greater than 3.0 (Table 1). We obtained the past regional fire frequency (hereafter *RegFF*; fire.year^{-1}) by combining the smoothed series using the R package '*paleofire*' v.1.2.3 (Blarquez et al., 2014). We assessed the significance of changes in *RegFF* and *RegBB* by using a bootstrap procedure with 999 iterations (BCI; 90%).

Following Ali et al. (2012), we added a constant equal to 1 to *RegBB* and *RegFF* and we computed the ratio (*FS* index), which we interpreted as indicating fire size and/or severity. The *FS* index corresponds to mean burned biomass per fire calculated for 500-year intervals, and thus, reflects fire size. Considering that large fires in boreal North America are known to cause high vegetation mortality, the *FS* index could also represent fire severity (Ali et al., 2012; Cansler and McKenzie, 2014; Hély et al., 2020; Hennebelle et al., 2020; Kelly et al., 2013). We assumed that high *FS* index values corresponded to periods when the study area experienced larger and more severe wildfires. Following Power et al. (2008), we used a reference period, here corresponding to 500-0 cal. yrs. BP (by convention, 0 cal. yr. BP corresponding to AD 1950), to calculate *RegBB*, *RegFF* and *FS* index anomalies.

2.7. Vegetation history reconstruction

We reconstructed the Holocene vegetation history using palynological analysis of subsamples of bulk sediment (1 cm^3) taken at regular 4 cm intervals along the core of lake *Emile*. We counted and determined pollen and spores in sediment subsamples using standard techniques described by Faegri and

Iversen, (1989). We counted a minimum of 300 pollen grains of terrestrial taxa for each subsample (Djamali and Cilleros, 2020), under a microscope with a × 200 or × 400 magnification factor. We identified pollen grains based on the Pollen Atlas of Arctic and Boreal Canada (Williams, 2006) and pollen keys from (Vincent, 1973) and Richard (1970). We identified tree pollen to the genus level. We also identified the green algae *Pediastrum* and spores of aquatic plants. We added exotic marker pollen (*Eucalyptus*) to each subsample to estimate pollen concentration (grains cm^3) and pollen accumulation rate (*n* grains cm^{-2} yr^{-1}) for taxa with an average percentage greater than 0.1% (Appendix S3), using the R package 'rioja' (Juggins, 2017). We measured the total pollen accumulation rate (PAR), based on all terrestrial taxa with percentages greater than 0.1%.

2.8. Temperature and drought

We used quantitative temperature reconstructions from four different proxies (described in Appendix S4) and originating from northwestern and north-central Canada (Lecavalier et al., 2017; Porter et al., 2019; Upiter et al., 2014; Viau et al., 2006). The reconstructions included (i) mean air temperature in July from 6,000 cal. yrs. BP to present estimated from chironomid assemblages sampled in lake sediments of the central Northwest Territories (Upiter et al., 2014), (ii) mean summer temperature anomalies relative to the 1961-1990 period in central Yukon, from 10,000 cal. yrs. BP to present, estimated using precipitation isotopes in syngenetic permafrost (Porter et al., 2019), (iii) mean Arctic summer air temperature anomalies relative to the preindustrial period reconstructed from ice melt in the Agassiz ice cap (Ellesmere Island, Northwest Territories) (Lecavalier et al., 2017), and (iv) mean July temperature estimated from pollen records for the past 10,000 years across northwestern North America (Viau et al., 2006). We assumed that these reconstructions captured temperature fluctuations during the summer months, i.e. the period of warm and dry weather conducive to wildfire activity in the boreal forest. For each reconstruction, we interpolated and standardized values for each year between 10,000 cal. yrs. BP and present, converting the data in anomalies relative to the mean for the entire period. We then averaged the four reconstructions. Then, we assessed the significance of temporal changes in temperature by bootstrapping the pooled means 999 times (BCI; 90%). We present temperature anomalies relative to the reference period 500-0 cal. yrs. BP.

To quantify the relationships among the fire, vegetation and temperature time series reconstructed at different temporal resolutions, and to avoid interpolation when time series were not sampled at identical points, we measured the Pearson correlation coefficient with 95% confidence interval between the automatically binned undetrended time series following Mudelsee (2013) and using the R package 'BINCOR' (Polanco-Martinez et al., 2019).

We used three indicators of hydrological variability to define Holocene dryness periods assumed to correspond to low vegetation productivity (inferred from pollen and plant macrofossil assemblages) and low lake water levels (inferred from diatoms) (Lauriol et al., 2009; Pienitz et al., 1999; Viau and Gajewski, 2009). We compared these records providing independent evidence of dryness periods with our fire reconstructions to detect possible temporal overlap of specific fire regimes and dryness periods.

3. Results

3.1. Age-depth models

Core length varied from 359 cm at lake *Izaac* to 456 cm at lake *Emile* (Table S1). According to the Bayesian age-depth models, basal sediments were dated between *ca.* 10,000 cal. yrs. BP at lake *Saxon* and *ca.* 9,530 cal. yrs. BP at lake *Izaac* (Fig. 2). We used 9,530 cal. yrs. BP as the earliest date for wildfire regime reconstructions, as it was included in all four reconstructions. The four age-depth models had similar sedimentation rates, varying between 0.036 cm and 0.049 cm year^{-1} (Table 1), similar to the results a previous study of other boreal lakes in the NWT (Crann et al., 2015).

3.2. Regional fire history

The regional fire activity remained fairly constant throughout the Holocene with approximately 5 fires per millennium (+2 anomaly; Fig. 3A). Mean *RegFF* very slightly increased from *ca.* 9,500 cal. yrs. BP to *ca.* 500 cal. yrs. BP, before decreasing during the reference period (500-0 cal. BP). *RegBB* steadily increased starting from *ca.* 9,500 cal. yrs. BP, peaked between *ca.* 7,000-5,000 cal. yrs. BP, before gradually decreasing to present values, the lowest of the entire series (Fig. 3B). Higher *FS* index values were recorded before 4,000 cal. yrs. BP (Fig. 3C), and, although to a lesser extent, during the reference period (500-0 cal. yrs. BP).

3.3. Vegetation history

We identified a total of 41 pollen taxa in the whole sequence extracted at lake *Emile*, covering the period from *ca.* 9,700 cal. yrs. BP to present. Eighteen taxa had a mean pollen percentage > 0.1% over the entire study period (Appendix S3). The PAR diagram is dominated by few tree and shrub taxa throughout the sequence (*Picea*, *Betula, Pinus* and *Alnus*; Fig. 4). *Artemisia*, Cyperaceae, Ericaceae, Poaceae, *Juniperus*, *Larix* and *Salix* were also present, but in low numbers.

PAR was low from *ca.* 9,700 to 7,800 cal. yrs. BP, concomitant with a stable and relatively low sedimentation rate (Fig. 4). Most trees and shrubs were relatively low, except *Populus, Myrica* and *Lycopodium*, which recorded their highest values at that time. A marked increase of *Picea* (likely black spruce, MacDonald et al., 1993) occurred *ca.* 7,800 cal. yrs. BP, along with a simultaneous expansion of *Alnus alnobetula* subsp. *crispa*. At the same time, the lake recorded a major decrease in *Populus*, *Myrica* and *Lycopodium*. PAR increased markedly, while the sedimentation rate remained stable. PAR, *Picea* and *Alnus alnobetula* subsp. *crispa* remained high until *ca.* 6,000 cal. yrs. BP, before decreasing, concomitant with an increase of the sedimentation rate ca. 5,700 cal. yrs. BP. *Pinus* (likely jack pine; Sulphur et al., 2016), *Betula* and *Alnus incana* subsp. *rugosa* increased somewhat later, while the sedimentation rate remained high. After *ca.* 4,000 cal. yrs. BP, the sedimentation rate decreased and remained stable for the last two millennia of the record, while *Pinus* remained relatively high, especially after *ca.* 2,500 cal. yrs. BP. A short-duration PAR peak occurred between *ca.* 3,000 and 2,200 cal. yrs. BP, although the sedimentation rate remained stable. *Pediastrum* and *Nuphar* increased simultaneously *ca.* 2,200 cal. yrs. BP. During the last two millennia, *Picea*, *Pinus*, *Betula* and *Alnus* sp. decreased, while Ericaceae and Cyperaceae remained stable or even slightly increased, such as for *Juniperus*. The sedimentation rate decreased after 500 cal. yrs. BP, following a slight increase between *ca.* 1,500 and 500 cal. yrs. BP.

3.4. Temperature and drought

We identified three main climatic periods over the Holocene, based on pooled summer temperature reconstructions (Fig. 5). Larger confidence intervals reveal more uncertainty in temperature estimations from 10,000 to 6,500 cal. yrs. BP, corresponding to the early Holocene period. Overall, maximum temperature values were recorded between *ca*. 6,500 and 4,500 cal. yrs. BP, before decreasing until *ca.* 1,500 cal. yrs. BP, and increasing again drastically during the last few centuries. According to inferred hydrological conditions (Fig. 6; Lauriol et al., 2009; Pienitz et al., 1999; Viau and Gajewski, 2009), the period between *ca.* 7,000-3,000 cal. yrs. BP was characterized by wetter conditions in central NWT and northern Yukon, compared to the rest of the study period. A major change in diatom-inferred dissolved organic carbon at *ca.* 5,800 cal. yrs. BP suggests a warm and humid interval during the mid-Holocene (Fig. 6a; Pienitz et al., 1999), resulting in high lake productivity (Fig. 6c; Lauriol et al., 2009). Conversely, high annual precipitation were recorded before *ca.* 7,500 cal. yrs. BP and around *ca.* 1,000 cal. yrs. BP (Fig. 6b; Viau and Gajewski, 2009).

3.5. Relationships between fire, temperature and vegetation

Warm periods (*ca*. 10,000 to 5,000 cal. yrs. BP and last 500 years) coincided with low *RegFF*, high *FS* index, and relatively high *RegBB* (Table 2). Higher temperatures corresponded with higher *Populus*, *Myrica* and *Lycopodium* abundance. A cooler period (*ca*. 5,000 to 500 cal. yrs. BP) was characterized by a low *FS* index and a landscape richer in *Pinus*, *Larix*, *Juniperus*, Poaceae, *Pediastrum* and *Nuphar* (Table 2). PAR, interpreted as fuel availability, coincided with high *RegFF* and high *RegBB*, especially for *Picea*, *Betula* and *Alnus alnobetula* subsp. *crispa* (Table 2).

4. Discussion

Our fire history reconstructions based on macroscopic sedimentary charcoal provide evidence that fire size and/or severity were higher during warmer periods (i.e. before *ca.* 5,000 cal. yrs. BP and during the last 500 years) than during the Neoglacial (after *ca.* 5,000 cal. yrs. BP). During the Holocene, the *RegBB* metric was positively correlated with the *FS* index because *RegFF* was relatively constant from the early to the late Holocene.

4.1. Fire, climate and vegetation interactions

Early Holocene (10,000-6,500 cal. yrs. BP)

The spatial variability of the climate data used in air temperature reconstructions shown by the large confidence interval around summer temperature during the early Holocene (i.e. before 6,500 cal. yrs. BP; Figure 5) is due to differences in time of deglaciation across northern Canada (Dyke, 2005; Kaufman et al., 2004). Summers were warmer in far northern NWT and in central Yukon (Lecavalier et al., 2017; Porter et al., 2019) but cooler in northwestern Canada (Viau et al., 2006), probably due to a later time of deglaciation. Conditions were also dry at this time in north-central Canada but not in northwestern Canada (i.e. Mackenzie region; Fig. 6), confirming the presence of the Laurentide ice sheet in the study area during the early Holocene. PAR indicates that the first stage of vegetation colonization following ice retreat was characterized by an open woodland, as observed in previous studies for the same period and in the same area (Conedera et al., 2009; Macumber et al., 2011; Sulphur et al., 2016). The landscape was dominated by pioneer taxa such as *Populus* and *Betula* (likely dwarf birch, Andrews et al., 1980). The relatively high percentages of *Picea* spp. (~ 40-50 %) was likely due to long-distance transport of pollen by wind, from populations located to the southwest of the study area (Campbell et al., 1999). Low tree abundance before 8,000 cal. yrs. BP can be explained by the time required for northward migration following deglaciation (Gajewski et al., 1993; Moser and MacDonald, 1990; Ritchie, 1985). Dry conditions also could have limited tree cover. During the early Holocene, deciduous shrubs, less inflammable than conifer trees, and low fuel abundance likely limited fire ignition and spread, but might have promoted large and/or severe fires in shrub areas during dry years.

Mid-Holocene (6,500-5,000 cal. yrs. BP)

Warmer and wetter climate during the mid-Holocene in northern continental Canada (Kaufman et al., 2004; Porter et al., 2019; Ritchie et al., 1983) favored tree growth, leading to gradual densification of the vegetation cover as suggested by the increase in PAR values despite a relatively stable sedimentation rate. This result is in line with previous reconstructions in southwestern Yukon (Cwynar and Spear, 1995; Gajewski et al., 2014), Alaska (Tinner et al., 2006) and Northwest Territories (Sulphur et al., 2016). Warm temperature and wet conditions of the mid-Holocene especially favored dense conifer forests (*Picea* and *Pinus spp.*), causing fuel accumulation conducive to large and severe fires, as previously observed in Alaska (Hoecker et al., 2020). During this period, wet conditions limited the increase in fire frequency (*RegFF*), which remained stable, while biomass burning (*RegBB*) was higher. Because the period was wet, it suggests that these non-frequent large wildfires occurred during episodic drought that dried the fuel and favored fire spread. High fire activity during the mid-Holocene promoted fire-prone coniferous species (*Picea* and *Pinus*) and pioneer trees and shrubs (*Betula* and *Alnus*), as previously observed in the study area (Parisien et al., 2020).

Neoglacial (5,000-1,000 cal. yrs. BP)

The Neoglacial was characterized by a gradual decrease in temperature, reaching an all-time low *ca.* 1,500 cal. yrs. BP. Cooling favored the expansion of *Pediastrum* and *Nuphar ca.* 3,000 cal. yrs. BP, as observed in Alaska (Edwards et al., 2000). The Neoglacial was also characterized by landscape opening, as shown by a decrease in *Picea* and increase in Poaceae, also observed in previous studies (MacDonald, 1995; Pienitz et al., 1999). Lower fuel abundance, likely due to cooler and drier conditions less favorable to shrub and tree productivity, hindered fire spread leading to lower biomass burned.

Last millennium (1,000-0 cal. yrs. BP)

During the last 1,000 years, temperature varied but remained low until *ca.* 500 cal. yrs. BP before a marked increase, especially during the last century. Vegetation density and fire frequency were as low as during the early Holocene, while *RegBB* reached its lowest levels for the entire time series. Contrary to our expectations, the recent fire size/severity (i.e. over the last 500 years) is below the maximum values observed during the mid-Holocene.

4.2 Implications for future fire risk

We provide evidence that both climate conditions and vegetation dynamics played a key role in shaping the wildfire regime over the past 10,000 years in central NWT. Fire size and/or severity were higher under the warmer and wetter climate of the mid-Holocene (7,000 to 5,000 cal. yrs. BP), which favored fuel availability, corroborating recent observations on large wildfires during the last decades in the study area (Gaboriau et al., under review). Our results can be used to anticipate future fire risk and to elaborate risk mitigation strategies including fuel management.

While temperature is expected to continue increasing in western Canada over the 21st century, dryness periods could be more severe than in the past (Price et al., 2013). Hence, fire frequency might increase (Wotton et al., 2017), but not necessarily biomass burning. Indeed, temperature increase could lead to the conversion of coniferous to deciduous forests (Hansen et al., in press; Mekonnen et al., 2019), or to a more open landscape (Asselin and Payette, 2005; Baltzer et al., submitted), which would negatively feedback on fire ignition and spread by reducing fuel flammability, combustibility, and/or connectivity. Hence, for large wildfires to be frequent in the future, the warming-induced increase in evapotranspiration will have to be compensated by increased precipitation to produce sufficient fuel (Flannigan et al., 2016).

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Data and R codes used in this manuscript are available here.

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Table 1. Characteristics of the four studied lakes and sediment records.

Lakes	**Emile**	**Izaac**	**Paradis**	**Saxon**

Characteristics				
Latitude (°N)	64°03'16.64"	64°05'56.21"	64°00'23.54"	63°48'26.06"
Longitude (°W)	114°06'21.67"	114°10'33.08"	114°59'19.24"	114°58'33.29"
Surface area (ha)	1.4	1.5	1.8	0.9
Elevation (m above sea level)	393	394	347	392
Maximum water depth (cm)	150	115	235	540
Maximum length (m)	203	162	271	177
Lakeshores (**F**lat / **A**brupt)	F	F	F	A
Fluvial input (**P**resent / **A**bsent)	A	A	A	A
Date of the last local fire (CNFD)	2014	2014	1979	1998
Sediment records				
Total length of sediment (cm)	456	359	450	395
Depth of the basal ^{14}C date (cm)	448	340	441	350
Sedimentation rate (cm year^{-1})	0.049	0.036	0.046	0.036
Median signal-to-noise index	5.4	4.2	4.1	3.7
Total number of significant fire events	57	49	55	46

Table 2. BINCOR Pearson correlation coefficient (with 95% confidence interval) between fire metrics, temperature, total pollen accumulation rate (PAR), and accumulation rates of taxa with an average percentage greater than 0.1% over the entire study period. *** p < 0.001, ** p < 0.01 and * p < 0.05.

	RegFF	***RegBB***	***FS* index**	**Temperature**
Temperature	-0.53 [-0.63, -0.42]***	0.21 [0.07, 0.35]**	0.66 [0.58, 0.74]***	-
PAR	0.50 [0.35, 0.62]***	0.58 [0.45, 0.68]***	0.14 [-0.04, 0.31]	-0.13 [-0.30, 0.04]
Picea	0.37 [0.21, 0.51]***	0.64 [0.53, 0.74]***	0.29 [0.12, 0.44]***	0.00 [-0.17, 0.17]
Pinus	0.63 [0.52, 0.72]***	-0.16 [-0.32, 0.00]	-0.56 [-0.67, -0.44]***	-0.52 [-0.63, -0.39]***
Betula	0.27 [0.12, 0.42]***	0.31 [0.16, 0.45]***	0.09 [-0.08, 0.25]	-0.06 [-0.22, 0.1]
Populus	-0.34 [-0.47, -0.18]***	0.04 [-0.13, 0.20]	0.28 [0.12, 0.43]***	0.18 [0.02, 0.34]*
Alnus alnobetula* subsp. *crispa	0.40 [0.25, 0.53]***	0.42 [0.28, 0.55]***	0.09 [-0.08, 0.25]	-0.02 [-0.18, 0.14]
Alnus incana* subsp. *rugosa	0.23 [0.07, 0.38]**	0.16 [0.00, 0.32]*	-0.02 [-0.18, 0.15]	-0.14 [-0.30, 0.02]
Juniperus	0.09 [-0.08, 0.25]	-0.30 [-0.45, -0.15]***	-0.33 [-0.47, -0.17]***	-0.38 [-0.51, -0.23]***
Salix	0.21 [0.05, 0.36]*	0.10 [-0.06, 0.26]	-0.04 [-0.21, 0.12]	-0.17 [-0.32, -0.01]*
Larix	0.21 [0.05, 0.36]*	-0.23 [-0.38, -0.07]**	-0.34 [-0.48, -0.19]***	-0.26 [-0.41, -0.1]**
Cyperaceae	0.20 [0.03, 0.35]*	-0.01 [-0.17, 0.15]	-0.13 [-0.29, 0.03]	-0.19 [-0.34, -0.03]*
Ericaceae	0.14 [-0.03, 0.29]	0.11 [-0.06, 0.27]	0.00 [-0.16, 0.17]	0.01 [-0.15, 0.17]
Artemisia	-0.05 [-0.22, 0.11]	0.10 [-0.06, 0.26]	0.13 [-0.04, 0.29]	0.15 [-0.01, 0.31]
Myrica	-0.46 [-0.58, -0.32]***	-0.01 [-0.17, 0.16]	0.33 [0.17, 0.47]***	0.38 [0.23, 0.51]***
Lycopodium	-0.31 [-0.45, -0.15]***	-0.05 [-0.21, 0.12]	0.19 [0.03, 0.35]*	0.2 [0.04, 0.35]*
Poaceae	0.23 [0.07, 0.38]**	-0.04 [-0.21, 0.12]	-0.19 [-0.34, -0.02]*	-0.24 [-0.39, -0.08]**
Pediastrum	0.28 [0.12, 0.43]***	-0.23 [-0.38, -0.07]**	-0.37 [-0.5, -0.22]***	-0.23 [-0.38, -0.07]**
Potamogeton	-0.14 [-0.30, 0.02]	0.14 [-0.02, 0.30]	0.22 [0.06, 0.37]**	0.10 [-0.06, 0.26]
Nuphar	0.38 [0.23, 0.51]***	0.01 [-0.15, 0.18]	-0.24 [-0.39, -0.08]**	-0.24 [-0.39, -0.08]**

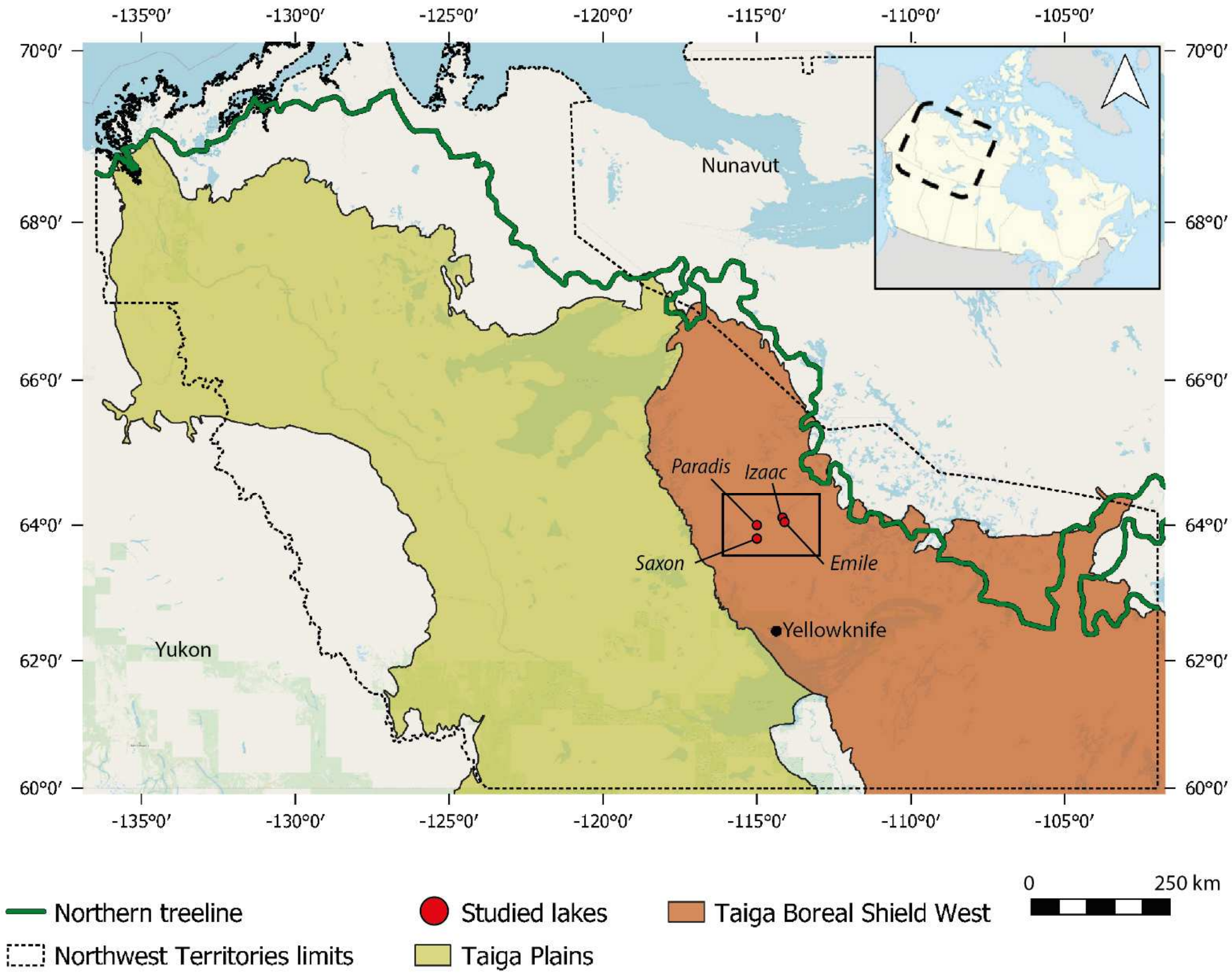

Figure 1. Locations of the studied lakes in the Northwest Territories (north-central Canada), showing ecozones (adapted from Olson et al., 2001) and treeline (adapted from Timoney et al., 2019).

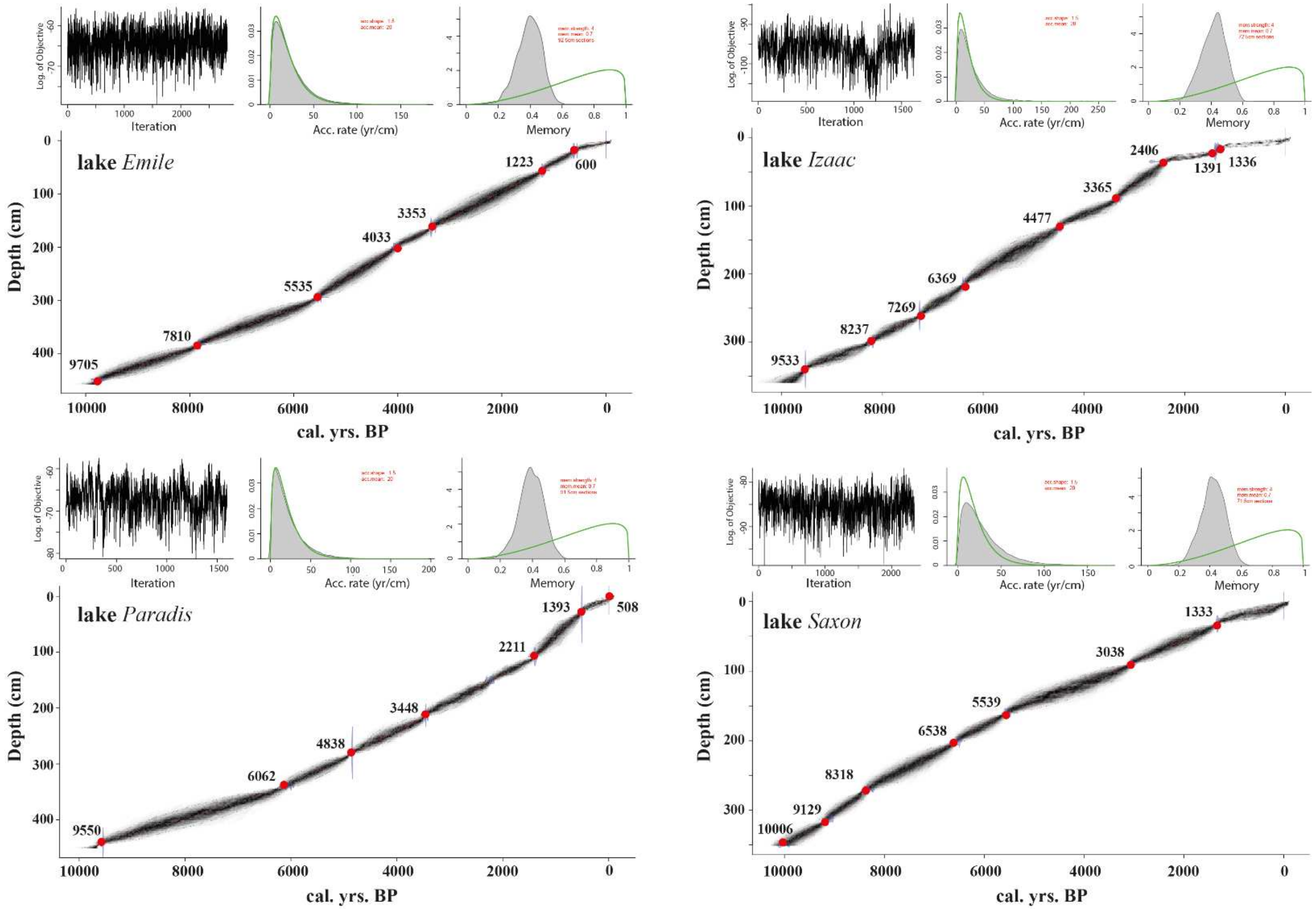

Figure 2. WinBacon outputs for lakes *Emile*, *Izaac*, *Paradis* and *Saxon*. Upper left panels describe the MCMC (Markov Chain Monte Carlo) iterations (the distribution is stationary with little structure among neighbouring iterations). Upper middle panels show the distribution of sediment accumulation rates. Upper right panels show the memory corresponding to the variation of sediment accumulation rate in time. Main panels show the calibrated ^{14}C dates (see Table S1 for full details on the chronology) and age-depth models (with 95% confidence intervals).

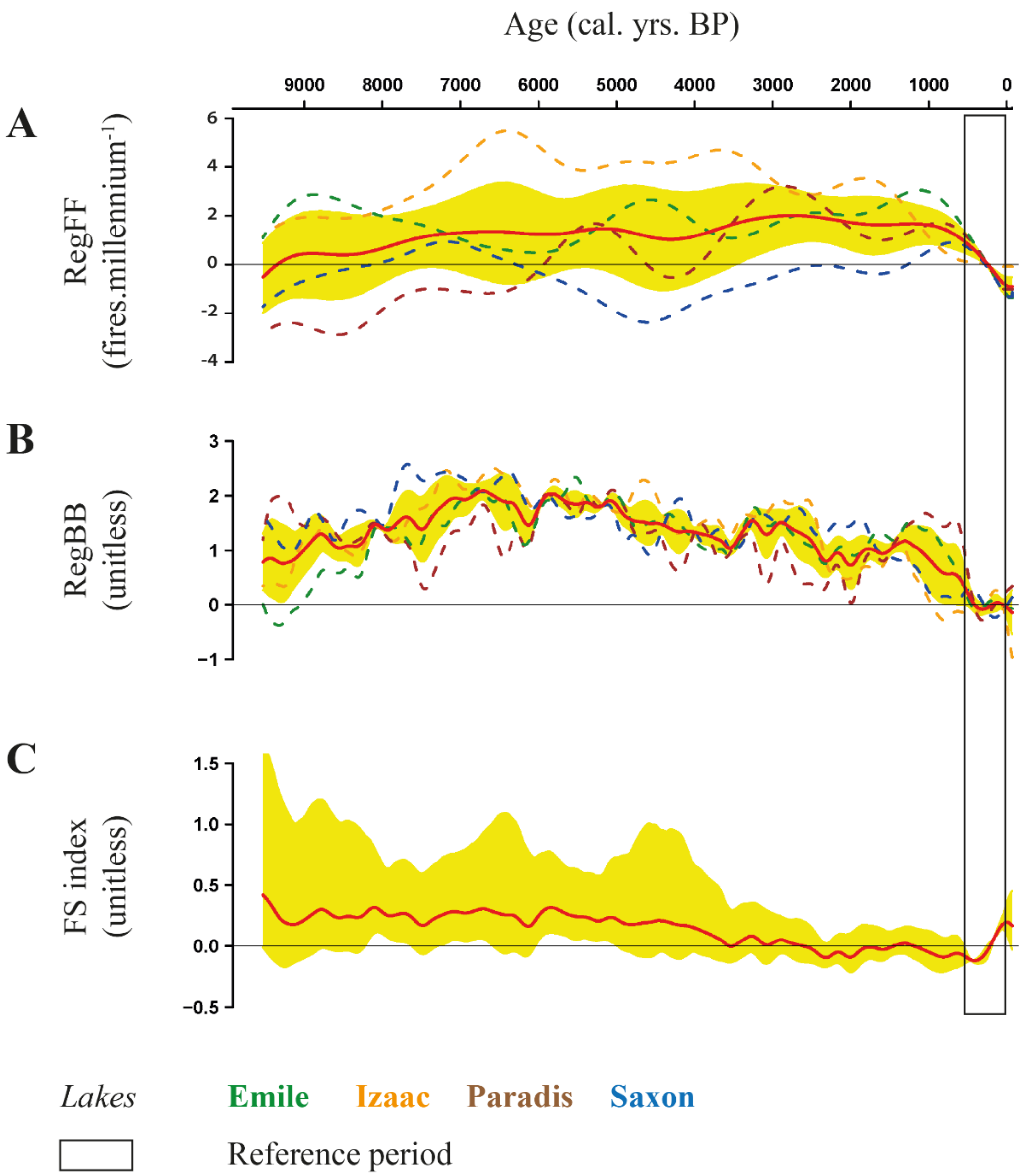

Figure 3. Holocene fire activity anomalies interpolated using a 500-year bandwidth smoothing, relative to the 500-0 cal. yrs. BP reference period (black horizontal line) for individual and regional fire-history reconstructions based on sediment charcoal records from lakes *Emile*, *Izaac*, *Paradis* and *Saxon*: (A) Regional Fire Frequency (*RegFF*), (B) Regional Biomass Burning (*RegBB*) and (C) Fire Size/Severity (*FS* index) based on regional biomass burning and fire frequency. Yellow shaded areas in each panel represent the 90% bootstrap confidence intervals.

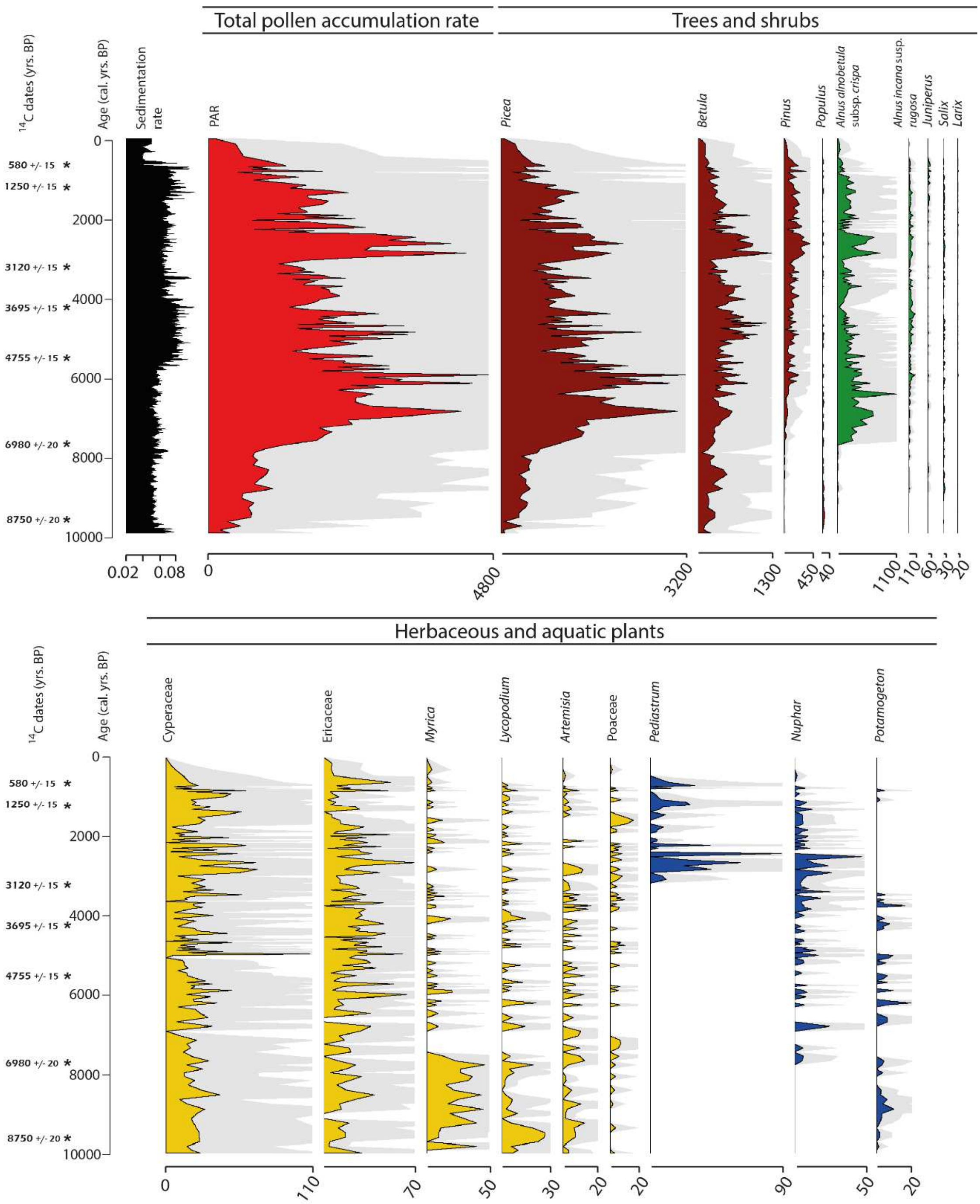

Figure 4. Sedimentation (cm^{-1} $year^{-1}$) and pollen accumulation rates (grains cm^{-2} $year^{-1}$) at lake *Emile* for total terrestrial pollen (PAR, in red) and for the main taxa: trees (in brown), shrubs (in green), herbaceous plants (in yellow) and green algae (*Coenobium*) and aquatic plants (in blue) all having an average percentage greater than 0.1% over the entire study period. Pale areas represent × 5 exaggeration. Note that the scale differs for each pollen or spore taxon.

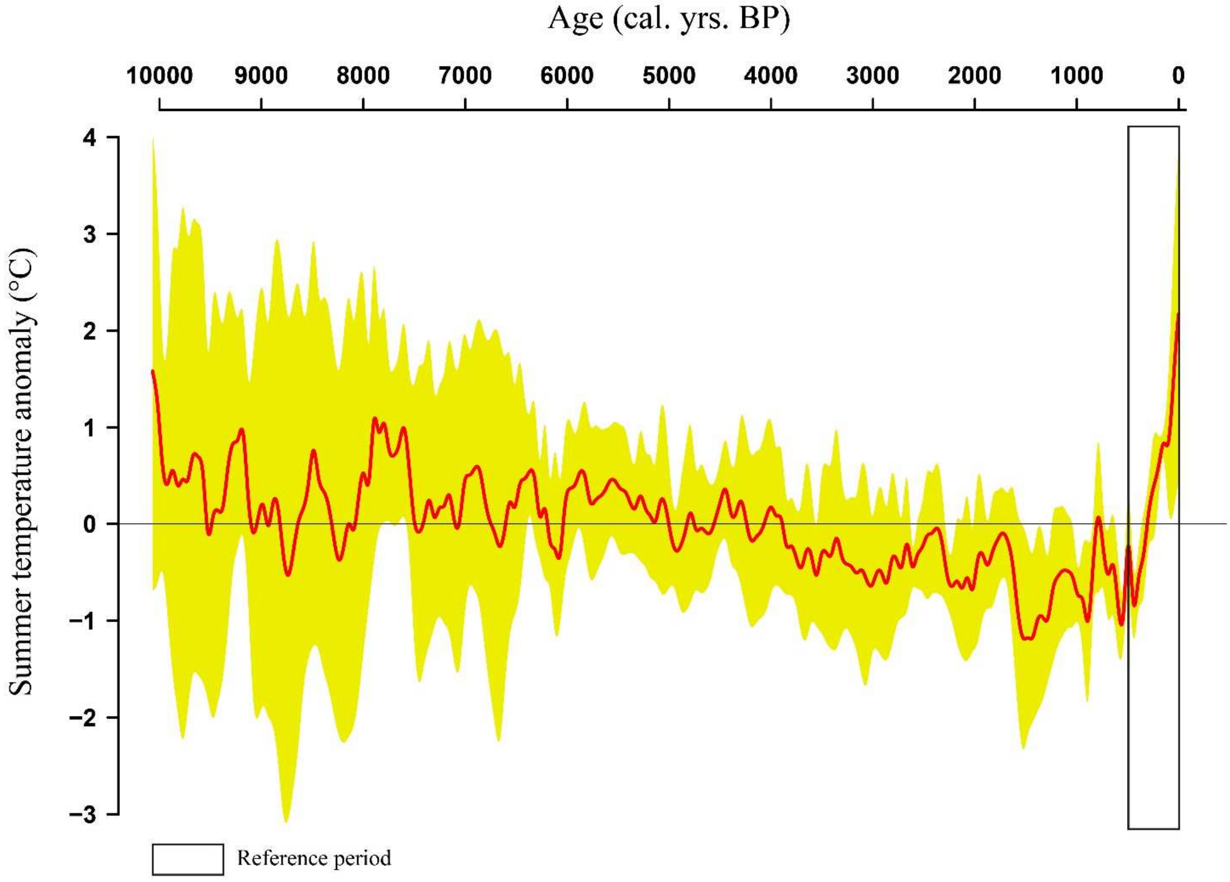

Figure 5. Mean summer temperature anomalies over the Holocene, relative to the 500-0 cal. yrs. BP reference period, obtained from the calculation of the means of standardized independent temperature datasets described in Appendix S4. The yellow shaded area indicates the 90% bootstrap confidence interval.

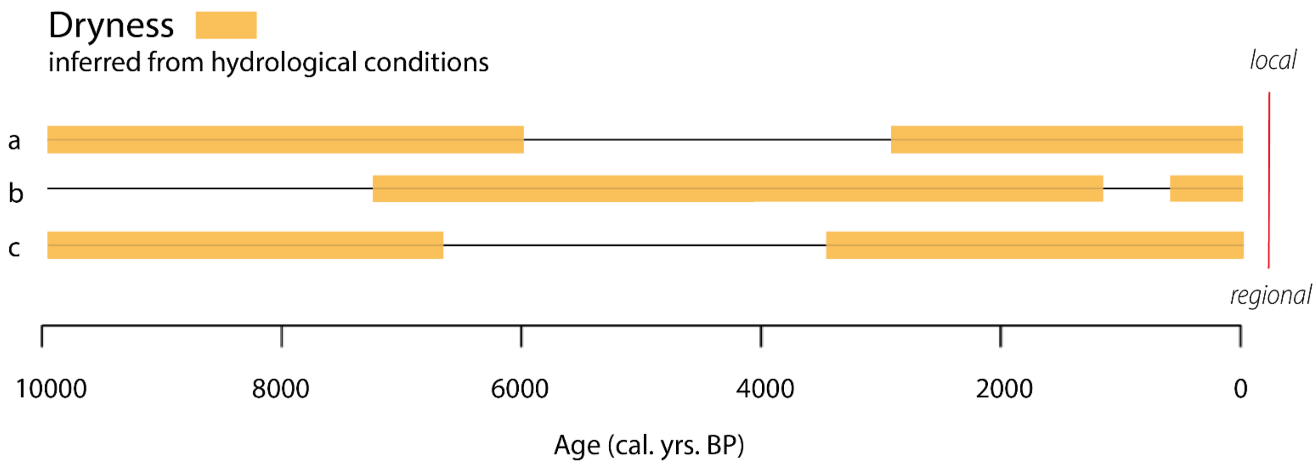

Figure 6. Dryness periods inferred from hydrological conditions from different locations in north-central and northwestern Canada. Distance from the study area increases from top to bottom. (*a*) Central Northwest Territories (Pienitz et al., 1999; dissolved organic carbon (DOC) inferred form diatom assemblages). (*b*) Mackenzie region (50°-70°N, 120°-140°W; Viau and Gajewski, 2009; annual precipitation inferred from pollen assemblages). (*c*) Northern Yukon (Lauriol et al., 2009; lake-level inferred from plant macrofossil analysis).