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Holocene climate–fire–vegetation interactions at a subalpine watershed in southeastern British Columbia, Canada

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ABSTRACT

Vegetation assemblages and associated disturbance regimes are spatially heterogeneous in mountain ecosystems throughout the world due to the complex terrain and strong environmental gradients. Given this complexity, numerous sites describing postglacial vegetation and fire histories are needed to adequately understand forest development and ecosystem responses to varying climate and disturbance regimes. To gain insight into long-term historical climate–fire–vegetation interactions in southeastern British Columbia, Canada, sedimentological and paleoecological analyses were performed on a sediment core recovered from a small subalpine lake. The pollen assemblages, stomata, and macroremains indicate that from 9500 to 7500 cal yr BP, *Pinus*-dominated forests occurred within the catchment and *Alnus* was also present. Climate was an important control of fire and fire frequency was highest at this time, peaking at 8 fires 1000 yr^{-1}, yet charcoal accumulation rates were low, indicative of low terrestrial biomass abundance. From 7500 to 4600 cal yr BP, *Pinus* decreased as *Picea*, *Abies* and *Larix* increased and fire frequencies decreased to 3–6 fires 1000 yr^{-1}. Since 7500 cal yr BP the fire regime varied at a millennial scale, driven by forest biomass abundance and fuel accumulation changes. Local scale (bottom-up) controls of fire increased in relative importance since at least 6000 cal yr BP.

Introduction

Fire is one of the most important abiotic disturbances to forested ecosystems, influencing stand age and composition, biodiversity, soil stability, carbon flux and biogeochemical cycling. It is critical to examine historical rates of change and the mechanisms interacting at various spatial and temporal scales to understand how future climate and vegetation changes could affect fire regimes (Whitlock et al., 2010; Hessl, 2011). Fire occurrence in mountainous regions is a patchy phenomenon due to complex interactions between top-down (climatic) and bottom-up (local) controls over multiple temporal and spatial scales. The relative importance of various controls on fire regimes is not static through time (Gedalof, 2011), highlighting the need to develop numerous long-term paleoecological records to fully understand past climate–fire–vegetation interactions. Climate variability, a top-down control, varies on short and long time scales and interacts with vegetation, fuel abundance and conditions, topography and other bottom-up control factors that affect the energy and moisture regimes across the landscape. Over long time scales, in mesic ecosystems with substantial biomass growth and accumulation, temperature is the most important control of postglacial fire activity (Daniau et al., 2012). In western North America, fire activity has increased in recent decades due to climate warming (Westerling et al., 2006) in combination with the ecological impacts caused by effective fire suppression (Marlon et al., 2012). However, how these influences affect fire regimes at high-elevation forests is less certain. Bottom-up controls, such as vegetation type and density, topography, and aspect, have also been shown to be important influences in explaining the spatial variability of fire activity in western North America (Heyerdahl et al., 2001; Gavin et al., 2006; Heyerdahl et al., 2007; Courtney Mustaphi and Pisaric, 2013). Lake sediment records provide a long-term perspective permitting the examination of Holocene fire regime variability and the relative importance of abiotic and biotic controls on past biomass burning (Whitlock and Larsen, 2001).

Records of vegetation and fire regime variability are necessary to inform land management policy and decision making (Gavin et al., 2007). Long-term fire histories need to be examined to resolve the relative importance of the interactions between top-down and bottom-up influences of fire across the heterogeneous landscape of mountainous southern British Columbia, Canada. Multiple studies have examined the long-term disturbance histories of Engelmann spruce-subalpine fir (ESSF; *Picea engelmannii*, *Abies lasiocarpa*) forests in southern British Columbia, Canada (Wong et al., 2004; Gavin et al., 2006; Courtney Mustaphi and Pisaric, 2013). Previous dendroecological studies in this region have shown that pre-settlement disturbance regime intervals in wet cool ESSF forest stands range from 90 to 807 yr and 105 to

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508 yr and that fires are the most important disturbance causing mortality over large areas (Dorner, 2001; Wong et al., 2004). Lower elevation sites within 10–40 km of the study site experience mixed severity fire regimes (Nesbitt, 2010), with stand-replacing disturbance intervals of 150–350 yr for ESSF forests (Pollack et al., 1997; Wong et al., 2004). Additional Holocene records are necessary to examine the linked or cascading interrelationships between different types of disturbance, to capture the full variability of fire activity (Courtney Mustaphi and Pisaric, 2013), and to spatially-resolve the controls of fire regimes (Heyerdahl et al., 2001).

Regional synchrony of fire activity is caused by the top-down influences of climate across the Pacific Northwest (Gedalof et al., 2005). Large fires in the interior of British Columbia are associated with blocking circulation patterns resulting in prolonged high-pressure systems causing fuel drying and intermittent convective thunderstorms (Johnson et al., 1990; Johnson and Wowchuk, 1993). Forests of the Kootenay region of southeastern British Columbia tend to burn during warm and dry years with no significant relationship with previous growing season conditions (Heyerdahl et al., 2002; Da Silva, 2009). Fires can occur in any given year, but are more likely to occur in the Pacific Northwest during regional summer drought (Trouet et al., 2010) associated, at decadal and sub-decadal time scales, with El Niño conditions (Heyerdahl et al., 2002). Climatic influences of winter snow accumulation and the rate of spring melting both have implications for the timing of the fire season. At multi-decadal scales, ocean–atmosphere interactions over the Pacific and Atlantic Oceans modulate fire activity in the Pacific Northwest (Kitzberger et al., 2007). It has also been shown, however, that local site factors can override regional climate as the dominant influence on fire regimes at many montane study sites in the Pacific Northwest (Heyerdahl et al., 2002; Gavin et al., 2006; Heyerdahl et al., 2007) and that the relative influence of various top-down and bottom-up controls varies through time (Heyerdahl et al., 2001; Courtney Mustaphi and Pisaric, 2013). For example, Engelmann spruce–subalpine fir forests on north-facing slopes burn less frequently than those on south-facing slopes during the late Holocene (Steventon, 1997; Gavin et al., 2006; Courtney Mustaphi and Pisaric, 2013), with Weibull median fire return intervals of 226–241 yr across north-facing watersheds and 135–190 yr at south-facing sites (Courtney Mustaphi and Pisaric, 2013). These results suggest that bottom-up controls of fire are important regulators of fire regimes in this region; however, these controls may vary in relative importance through time and be mediated by top-down controls.

We present a postglacial fire and vegetation history study from a north-facing, subalpine watershed covered by a wet and cool ESSF forest. This study aims to investigate how fire regimes at a high-elevation ESSF-forested site have been influenced by Holocene climatic variability, changes to the vegetation assemblages, and the quantity of biomass within the catchment. To provide information on the local and regional vegetation cover variability, we present a pollen and stomata record established from a sediment core collected from the center of Lake NEL03 (unofficial name). We then integrate vegetation data with a high-resolution charcoal record representing historical forest fire activity to investigate Holocene climate–fire–vegetation interactions. Inferences on biomass are based on qualitative interpretations of dominant pollen types and total macroscopic charcoal accumulation rates. Sedimentological and other paleoecological information are also discussed to understand other lake-system changes and their relationships to terrestrial vegetation changes.

Study site

A large portion of the Nelson Range of the Selkirk Mountains is a managed forested land that is crucial to conservation efforts and the continued sustainability of environmental services provided by natural spaces in southeastern British Columbia, Canada. This region contains a large managed area of ESSF forest that has been minimally disturbed by anthropogenic activities and development, such as deforestation, fragmentation and grazing. It is therefore a useful region to examine the natural variability of past ecosystem dynamics; although, since AD 1945, these forests have been influenced by effective modern fire suppression (Nesbitt, 2010; Greene, 2011). A large portion of the population and infrastructure in the region is at the wildland–urban interface and it is one of the most ecologically diverse regions of British Columbia with notable managed lands, including the Harrop-Procter Community Forest, Midge Creek Wildlife Management Area, West Arm Provincial Park, and Kokanee Glacier Provincial Park.

NEL03 (unofficial name; 49°29′46″N, 116°54′17″W; Figs. 1 and 2) has a catchment area of 36 ha and is located near the head of a north–northeast trending, glacially and fluvially incised valley. It is a small (0.35 ha) subalpine, cirque lake (2074 m asl), with a subcircular

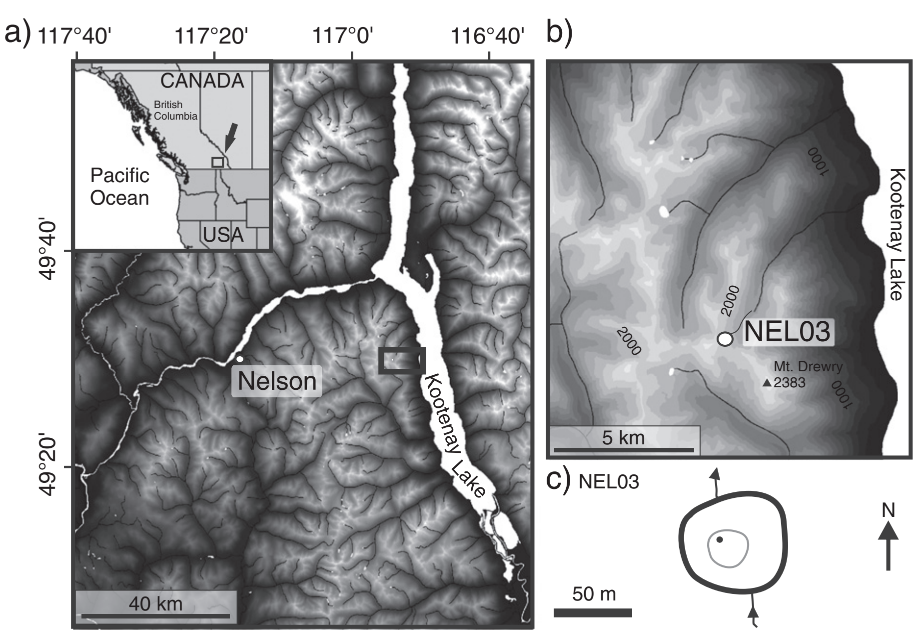

Figure 1. a) Location of study region, black box shows b) the lake site (100 m topographic contours), and c) depicts the lake bathymetry (2 m isobath) and coring location.

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Figure 2. South–southwest view of NEL03 (photo: Ze'ev Gedalof).

morphometry and maximum measured depth of ~2.4 m. Drainage into the lake is ephemeral and outflow, during periods of high water levels, drains north and northeast through Heather Creek to Kootenay Lake. NEL03 is situated within siltstone, argillite, quartzite, and dolomite rock of the Dutch Creek Formation (Mesoproterozoic) overlain with glacial debris from the bedrock and micaceous schists from the nearby Eagle Bay assemblage (Cambrian–Devonian; Journeay et al., 2000). Coarse-textured orthic humo-ferric podzols of the Bonner Association, 50–150 cm deep, cover most of the watershed (Jungen, 1980) and colluvium and exposed bedrock facies occur to the southwest of the lake. Late-lying snowpacks can persist into July, significantly decreasing the length of the fire season at this ridge-shaded site with a generally north-facing aspect.

The Columbia Mountains are located at the eastern reach of the Interior Wet Belt of southeastern British Columbia, where the rugged landscape causes westerly air to converge (Kendrew and Kerr, 1955; Oke, 1987), leading to high total annual precipitation (1400 mm), much of it (~60%) in the form of snow. Winters are cold (−8.6°C) and summers are cool (10.9°C; 1981–2009 baseline; Wang et al., 2012) and punctuated by westerly, episodic, convective thunderstorms (Jungen, 1980). Fires can occur as early as April and as late as October (Canadian Wildland Fire Information System, 1959–1999).

The lake is within the Engelmann spruce-subalpine fir wet cold forest zone, dominated by subalpine fir (*A. lasiocarpa*), subalpine larch (*Larix lyallii*), Engelmann spruce (*P. engelmannii*); and some whitebark pine (*Pinus albicaulis*). Arboreal coverage is moderately dense and there is a significant understory of herbs, forbs, sedges, grasses and shrubs. Above the lake, patchy ESSF forest extends ~200 m to the ridge. The stand age is estimated to be 190 yr old (BC Vegetation Resources Inventory) and is classed as experiencing rare stand-initiating events (NDT1). Recent disturbance patterns suggest a mean stand-replacing disturbance interval of 150–350 yr for ESSF forests (Pollack et al., 1997; Wong et al., 2004) and sedimentary charcoal-based fire reconstructions at this site suggest a Weibull median fire return interval (mFRI) of 226 yr (range 20–440 yr) over the past 5000 yr (Courtney Mustaphi and Pisaric, 2013). Recent observations showed no evidence of mountain pine beetle (MPB; *Dendroctonus ponderosae* Hopkins, 1902) outbreaks and estimates of windthrow are unknown (Wong et al., 2004); however, multiple large standing-dead larches proximal to the lake suggested low-incidence of windthrow over, at least, the recent past. Tree cores collected from the southeast section of the lake suggested a mixed-age stand of ESSF forest. Some larger subalpine fir had pith dates around the AD mid-1800s and a single sample dated to 1750, while larger subalpine larches dated to the mid-1600s with one dated to 1602. These dates suggest that the two most recent reconstructed fires (Courtney Mustaphi and Pisaric, 2013) did not kill all the trees surrounding the lake.

Methods

Field methods

During the summers of 2009 and 2010, surface and long cores were collected from the central deep basin of NEL03. Deployed from a boat, a large diameter Glew gravity corer (Glew et al., 2001) was used to collect the uppermost sediments with an intact sediment–water interface. The core was extruded into Whirl-Pak® bags using a vertical extruder (Glew, 1988) at contiguous 0.25-cm intervals, from 0 to 15 cm depth, and at 0.5-cm intervals to the base. Deeper sediments were collected in 1-m drives using a 5-cm diameter Livingstone piston corer (Wright et al., 1984) deployed from an anchored floating platform. Piston cores were extruded in the field, wrapped in plastic and aluminum foil, and shipped to the Carleton University Paleoecological Laboratory where they were refrigerated at ~4°C until being sectioned and bagged at contiguous 0.5-cm intervals. The lake bathymetry was estimated through 18 line depth soundings across the basin.

Chronostratigraphy

The surface core and Livingstone cores were overlapped using the presence of tephra shards as well as magnetic susceptibility, loss-on-ignition, and charcoal concentration data, to create a composite stratigraphy from the sediment–water interface to where coring ceased. Sediment ages were established using radiometric methods and an

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established regional tephrochronology (Foit et al., 2004; Gavin et al., 2006). The AMS radiocarbon dating results were previously published (Courtney Mustaphi and Pisaric, 2013). Felsic ashfall deposits from Cascadian eruptions were identified as the Mt St Helens Wn tephra (Mullineaux, 1996) dated to AD 1481–82 (468–469 cal yr BP; Yamaguchi, 1983, 1985) and the Mazama O tephra layer (Bacon, 1983) dated to 7627 ± 150 cal yr BP (Zdanowicz et al., 1999). A redeposited layer of Mazama O tephra occurred in the stratigraphy and all ash deposits were assumed to be near-instantaneous deposition events for age–depth modeling (Lowe, 2008). Radiocarbon dates were calibrated using IntCal09 (Reimer et al., 2009). The age–depth model was produced using the *R* script *Bacon* (Blaauw and Christen, 2011) using ~8 million *t*-walk MCMC algorithm iterations through the probability densities of the calibrated dates to establish a probable age–depth model within a 95% confidence envelope.

Sedimentological and paleoecological methods

A number of sedimentological methods were used to characterize the sediments. Magnetic susceptibility was measured at contiguous 0.25–0.5 cm intervals using Bartington Systems MS2B and MS2E sensors to gain insights about the history of erosion in the watershed and deposition in the lake (Dearing, 1999; Gedye et al., 2000). Loss-on-ignition analysis (Dean, 1974; Heiri et al., 2001) used 1 cm^3 subsamples taken at contiguous 0.25–0.5 cm intervals down core that were dried at 105°C for 24 h and then burned at 550°C for 4 h to estimate sediment organic content. The sample was then reheated to 950°C for 2 h to estimate carbonate content and the residual was assumed to represent the siliciclastic content. Particle size distributions were measured at ~10 cm resolution from 1 cm^3 subsamples that were pretreated with 30% hydrogen peroxide (H_2O_2) in a water bath (~75°C) to digest organic material while clastic material remained. Following this, 5 mL of 50 g L^{-1} sodium hexametaphosphate ($Na_6P_6O_{18}$) was added to disperse the sample before triplicate measurement runs of 60 s, at an obscuration of 10 ± 2%, using a Beckman Coulter LS 13 320 laser diffraction particle size analyzer (Syvitski, 1991).

Pollen preparation was done using 1 cm^3 subsamples taken at 4–10 cm resolution using standard digestion techniques (Fægri and Iversen, 1989) with *Lycopodium* tablet inoculation (Stockmarr, 1971). The samples were neither fine nor coarse sieved to preserve microscopic charcoal and stomata in the residues. Pollen was identified under 400–1000× magnification to a minimum terrestrial pollen count of 500 grains. Pollen identification was facilitated by the use of a reference collection at the Carleton University Paleoecological Laboratory and dichotomous keys (McAndrews et al., 1973; Bassett et al., 1978; Moore and Webb, 1978). The stomata were identified using keys by Hansen (1995) and Sweeney (2004) and represented local presence within the catchment of each taxon observed (MacDonald, 2001). Microscopic charcoal was also tallied during pollen counts. Pollen zones were established through constrained incremental sums of squares cluster analysis (CONISS) of the relative abundance of pollen taxa.

For macroscopic charcoal analysis, contiguous 1 cm^3 subsamples of homogeneous sediment were removed and soaked in a metaphosphate solution (Bamber, 1982) for at least 24 h and then wet sieved through a 150 μm mesh. The remaining material was transferred to a Petri dish and examined under a Nikon SMZ800 stereoscope at 6–40× magnification. Charcoal morphologies were classified into six morphotypes following Enache and Cumming (2006) and the total number of macroscopic charcoal pieces was tallied (Whitlock and Larsen, 2001). Other plant macro remains, such as conifer needles and seeds were counted as presence data during charcoal counts.

Fire episode reconstruction

Sediment macroscopic charcoal concentrations (pieces cm^{-3}) were converted to charcoal accumulation rates (CHAR; pieces cm^{-2} yr^{-1}) and analyzed using *CharAnalysis* software (Higuera et al., 2009) and the data was processed similarly to Moos and Cumming (2012) and Courtney Mustaphi and Pisaric (2013). The data were resampled to the global median sampling interval of 14 yr to create an interpolated CHAR series for peak analysis. The varying background charcoal accumulation rate (bCHAR) was estimated using a LOWESS smoother robust to outliers over a 500-yr window that was adequate for capturing centennial scale variability and will not be strongly biased by high-frequency peaks in CHAR (Gavin et al., 2006). Locally defined CHAR peaks within a 500-yr window, were obtained by subtracting the bCHAR component from the interpolated CHAR series (Higuera et al., 2010a) and identified through a 99% noise cut-off probability established by a 0- or 1-mean Gaussian mixture model (Gavin et al., 2006; Higuera et al., 2010a). The minimum charcoal concentration, within a 75-yr period prior to a peak, needed a probability <5% of coming from the same Poisson distribution as the associated peak to be retained as a significant peak (Higuera, 2009; Higuera et al., 2010a). Identified CHAR peaks were interpreted as fire episodes of ≥1 large fire(s) that occurred within the lake catchment area during a 14-yr window of the peak. Identified insignificant peaks could potentially be related to fire activity outside of the lake catchment (Higuera et al., 2010b). Fire return interval (FRI) distributions were calculated from the identified fire episode years and were then fitted with maximum likelihood estimate Weibull models. The FRI distribution passed a one-sample Kolmogorov–Smirnov goodness-of-fit test ($P > 0.05$) and Weibull median fire return interval with 95% confidence intervals were calculated (mFRI; Grissino–Mayer, 1999). Fire frequencies were smoothed using a LOWESS smoother with a window span of 1000 yr using the tally of number of fires during each 1000-yr period using K1D software (Long et al., 1998; Gavin et al., 2006; Higuera, 2009; Hallett and Anderson, 2010). Significant changes to biomass abundance

Table 1
Uncalibrated radiocarbon ages, 2σ calibrated ages (IntCal09 calibration curve; Reimer et al., 2009), and tephrochronology, used to develop the age–depth model.

Depth top (cm)	Depth bottom (cm)	Raw (^{14}C yr BP)	Raw ^{14}C error (yr)	^{13}C:^{12}C ratio	Calibrated age (cal yr BP)	Material	Reference
0					−59		Top of core
27.5	28				468–469	MSH Wn	Yamaguchi, 1985
90.5	95	2230	30	−26	2153–2335	Wood	Beta-301980
217.5	218.5	5690	40	−23.7	6399–6630	Wood	Beta-301981
244.5	253					Ash redeposit	
266.5	292.5				7477–7777	Mazama O	Zdanowicz et al., 1999
317	317.5	7920	40	−24.9	8606–8977	Plant material	Beta-301982
349.5	350	8600	50	−24.4	9495–9678	Wood	Beta-301983
	351.5						Base of core

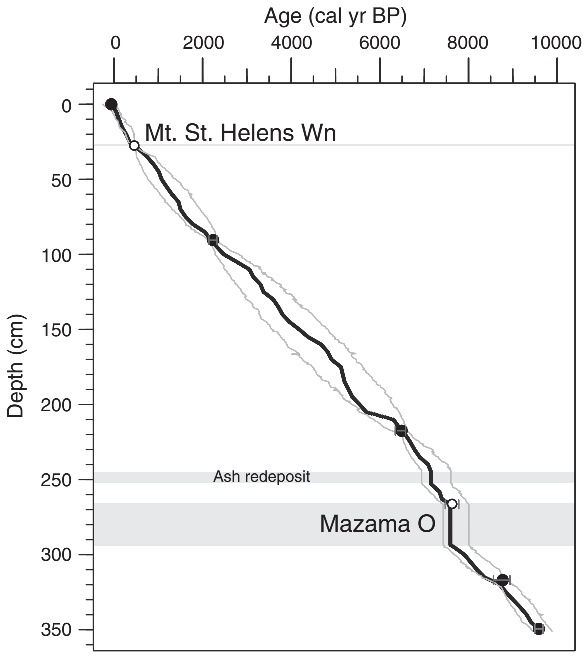

Figure 3. Age–depth model (black line) of the highest probability density of 8 million random walk iterations within the 95% confidence enveloped (gray lines) using calibrated radiocarbon dates (black circles; IntCal09, Reimer et al., 2009) and known tephra deposit dates (open circles) developed in the program *Bacon*. The 2σ error range of radiocarbon dates and the error estimate for the Mazama O eruption event are shown by dark gray error bars. Tephra thicknesses within the stratigraphy are illustrated by gray bands.

within the catchment, interpreted from bCHAR, were identified by a regime shift index (RSI; P = 0.0001, cut-off length = 100 yr, Huber's weight function = 5; Rodionov, 2004; Morris et al., 2013).

Results

Geochronology

The results from AMS radiocarbon dating of organic material preserved in the long core were presented by Courtney Mustaphi and Pisaric (2013) and are summarized in Table 1. In this study, an updated age–depth model was used that made full use of the probability densities of the calibrated radiocarbon ages (Fig. 3). This is a more realistic model that avoids the subjective nature of controlling the flexibility of a spline curve through an average of the radiocarbon dating errors, which is center-weighted and therefore limits the information within the probability densities of calibrated radiocarbon dates.

Lithology

Organic matter content of the sediment was highest (45–60%) during the early Holocene until the deposition of the Mazama O tephra (7627 ± 150 cal yr BP) and remained at 35–45% throughout the Holocene until steadily increasing to 55% after CE ~1915 (Fig. 4). Carbonate content remained low throughout the core (<10%) and few calcareous shell remains were found. The high (40–60%) siliciclastic input through small, steep, ephemeral streams resulted from the influx of a poorly sorted and coarse silty gyttja, containing subangular clasts with occasional sandier silt layers (Fig. 4). Conspicuous peaks in magnetic susceptibility were related to the thick tephra and sandy layers and the uppermost tephra layer was much less pronounced than earlier tephras. Clay content remained low, usually <6%, throughout the core, with the exception of the redeposited ash layer where it increased to ~10%, likely due to degradation of the thick Mazama ash in the soil (Jungen, 1980) prior to transport into the lake.

Vegetation record

The pollen record of NEL03 was divided into four zones using CONISS, 9600–7500, 7500–4600, 4600–1300, and 1300 cal yr BP–present (Fig. 5). Zone 1 (9600–7500 cal yr BP) was characterized by *Pinus*-dominated (50–60%) forests with abundant *Alnus* (15–30%), *Picea* (<20%) and *Abies* (<12%). A macroremain of a five-needle *Pinus* species indicated the local presence of a *Strobus* subgenus, likely *P. albicaulis*. The earliest, but not persistent detection of *Pseudotsuga*/*Larix* occurred at ~8200 cal yr BP. Deciduous shrub taxa, such as *Salix*, *Artemisia*, and *Betula*, were generally most abundant during this period (<8%). Herbaceous pollen abundances were also high (<2%) and Chenopodiaceae and pteridophytes were intermittently found (<1.5%). Aquatic pollen was most abundant during Zone 1 and *Pediastrum* was present with counts of <75 colonies.

Zone 2 (7500–4600 cal yr BP) remained *Pinus*-dominated; however, the abundance of *Pinus* pollen decreased from 55 to 35%, concomitant with increased *Picea* pollen (10–40%). A minor abundance of *Tsuga* (<1%) was found between 6000 and 5000 cal yr BP and *Pseudotsuga*/*Larix* (<1%) was found continuously from this zone to present. *Abies* remained moderately abundant (up to 15%) and deciduous trees and shrubs (*Populus*, *Betula*, *Alnus*, *Artemisia* and *Salix*) were present on the landscape. Asteraceae, and Caryophyllaceae remained <2%, Chenopodiaceae was absent, and Ericaceae, Poaceae and pteridophytes were intermittently present (<2%). *Pediastrum* colonies were consistently found with counts <10.

Pinus and *Picea* pollens were generally stable (30–35%) in Zone 3 (4600–1300 cal yr BP). *Abies* was variable (2–15%) and *Pseudotsuga*/*Larix* and Cupressaceae were present at low abundances (<2%). *Tsuga* increased to 2% and appeared continuously from 4000 cal yr BP. The shrub taxa were stable during Zone 3, generally 10–15%. Asteraceae, Poaceae and pteridophytes were present (<1.6%) and Caryophyllaceae, Ericaceae and Chenopodiaceae were rare. *Pediastrum* was found with counts <5 colonies and was occasionally absent.

Pinus decreased (~20%) and varied considerably in Zone 4 (1300 cal yr BP–present). *Picea* abundance peaked (40–70%) and *Abies* generally remained abundant (5–15%). *Pseudotsuga*/*Larix* increased to 2% and *L. lyallii* made up a large portion of the current forest immediately adjacent to the lake. Cupressaceae pollen, likely *Thuja plicata*, was consistently <1%. *Tsuga* increased to 2.5% and *Alnus* and *Artemisia* remained low (<10% and <5%, respectively). Artemisia decreased to <1.5%, *Betula*, *Salix*, and *Populus* remained <1%, and understory taxa decreased to <1% or was rare, as were *Pediastrum* remains.

Charcoal records and fire episode reconstruction

Microscopic charcoal accumulation rates, an indicator of regional biomass burning (Whitlock and Larsen, 2001; Conedera et al., 2009), averaged 1860 pieces cm^{-2} yr^{-1} and ranged from 265 to 7700 pieces cm^{-2} yr^{-1} (Fig. 5). Microscopic charcoal also varied between each pollen zone, averaging 2033 pieces cm^{-2} yr^{-1} in Zone 1, 1690 pieces cm^{-2} yr^{-1} in Zone 2, 1958 pieces cm^{-2} yr^{-1} in Zone 3 and 1786 pieces cm^{-2} yr^{-1} in Zone 4. The total macroscopic charcoal record had a global signal-to-noise index (SNI) value of 5.55 and local SNI values were all >3 (3.54–11.91), suggesting the charcoal record from NEL03 is robust for the reconstruction of the local fire history (Kelly et al., 2011). The median charcoal accumulation rate (CHAR) was 0.81 (range 0–12.2) pieces cm^{-2} yr^{-1}. Since 9650 cal yr BP, there had been 44 fire episodes in the watershed of NEL03 (Fig. 6). The Weibull median fire return interval (mFRI) was 216 yr (95% confidence interval 186–256 yr) within a range of fire return intervals of 42–630 yr. This fire history reconstruction represented an area fire return interval because the macroscopic charcoal influx represented the fire activity within the lake catchment (Clark, 1988; Higuera et al., 2007, 2010b).

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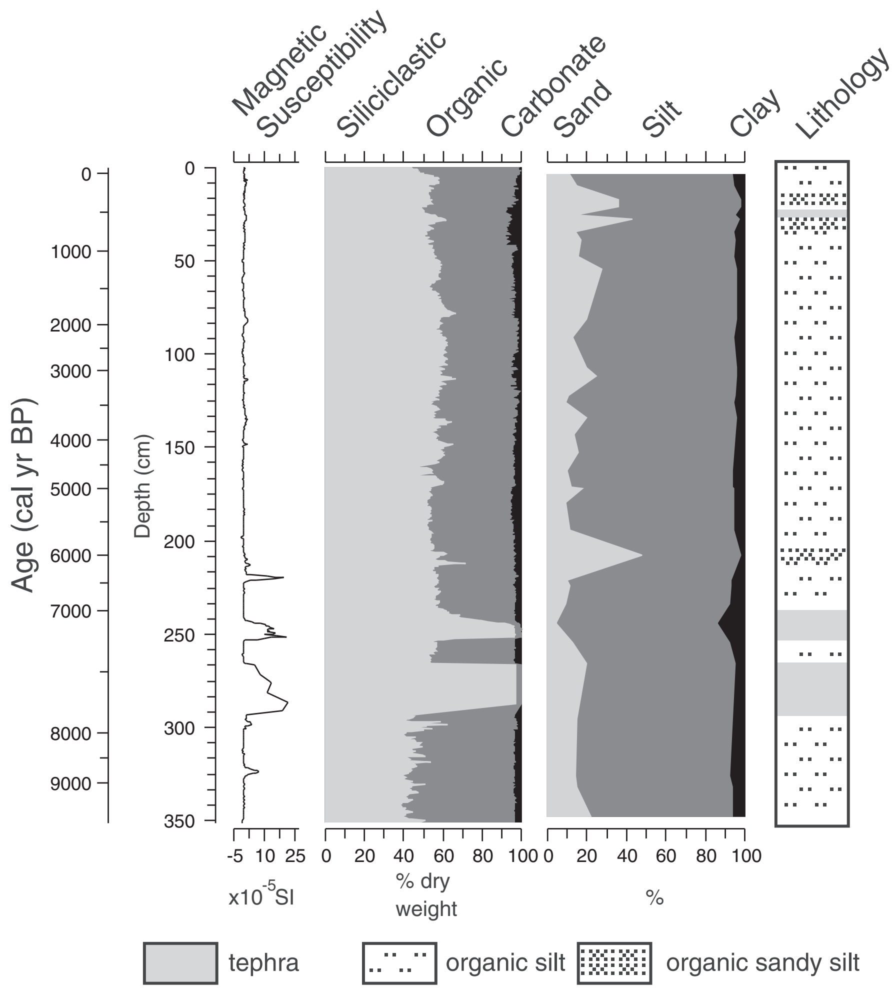

Figure 4. Physical sedimentology and general lithology of the composite stratigraphy collected from NEL03. Loss-on-ignition analysis was used to estimate the relative proportion of organic, carbonate, and siliciclastic content and particle size distributions were used to calculate the sand-, silt-, and clay-sized fractions of the siliciclastic sediments using *GRADISTAT* (Blott and Pye, 2001).

All morphotype CHARs were strongly correlated with total CHAR ($r = 0.69$–0.95, $p < 0.001$; Table 2). Morphotypes P, S, C, and F, shared the strongest similarity with total CHAR, having had much higher CHAR values and more CHAR peak dates that agreed with total CHAR peak fire events (Table 2; Fig. 6). Morphotype CHARs were also significantly correlated; the relationship was weakest between types M and D ($r = 0.46$, $p < 0.001$) and strongest between types C and P ($r = 0.84$, $p < 0.001$). Some insignificant peaks in morphotype CHARs occurred coeval with significant peaks in total CHAR. This occurred more often with charcoal morphotypes M, D, and F. Those morphotypes tended to have lower correlations with total CHAR and generally lower CHARs. Of the four insignificant peaks in total CHAR, 2–3 (50–75%) insignificant peaks of each morphotype CHARs agreed. Insignificant peaks identified in individual morphotype CHAR profiles were much more abundant than in total CHAR, notably with M and D morphotypes.

Variability of the fire regime

The fire frequency was highest during zone 1 (9600–7500 cal yr BP) with 4–8 fires 1000 yr^{-1} and a Weibull mFRI of 150 yr (12 fire events; Fig. 7). The microscopic charcoal input was also the highest of the record (Fig. 5). Fires occurred less frequently during zone 2, Weibull mFRI = 236 yr (10 fire events) and the period between 6500 and 5500 cal yr BP conspicuously showed very low bCHARs, decreased fire frequency (3–4 fires 1000 yr^{-1}; Fig. 7), and low microcharcoal inputs (Fig. 5). In zone 3, fire frequency increased slightly with a Weibull mFRI of 200 yr (17 fire events). There were five fire episodes during Zone 4 (1300 cal yr BP–present) with a raw median return interval of 287 yr. A Weibull mFRI was not calculated due to the low number of fires within the period. The occurrence of microcharcoal cenospheres in the uppermost sediments of Zone 4 is likely the result of recent transport and deposition of fossil fuel burning products to this remote subalpine location since CE ~1920 (Thevenon and Anselmetti, 2007; Fig. 7).

Discussion

Vegetation and fire regime interpretation

Zone 1: 9500–7500 cal yr BP

The low pollen influx (<8000 grains cm^{-2} yr^{-1}) and low bCHAR suggested the *Pinus* and *Alnus* forest of the early Holocene was not as dense as the mid- and late Holocene forests (Figs. 5 and 7). Background CHAR, an indicator of biomass on the terrestrial landscape (Carter et al., 2013; Marlon et al., 2006), was lowest during this early period (Fig. 7) and Poaceae pollen was high and suggested that the forest canopy was open and possibly patchy. Patchy open parkland forests are common to modern high-elevation sites in the region (Coupé et al., 1991). The RSI analysis of the bCHAR indicated a sustained period of low values during this zone. The lack of a tundra-type vegetation assemblage suggested that the immediate postglacial accumulation of sediments was not collected. The presence of conifer stomata and macroremains early in the record suggested the vegetation was subalpine (Fig. 5). Indeed,

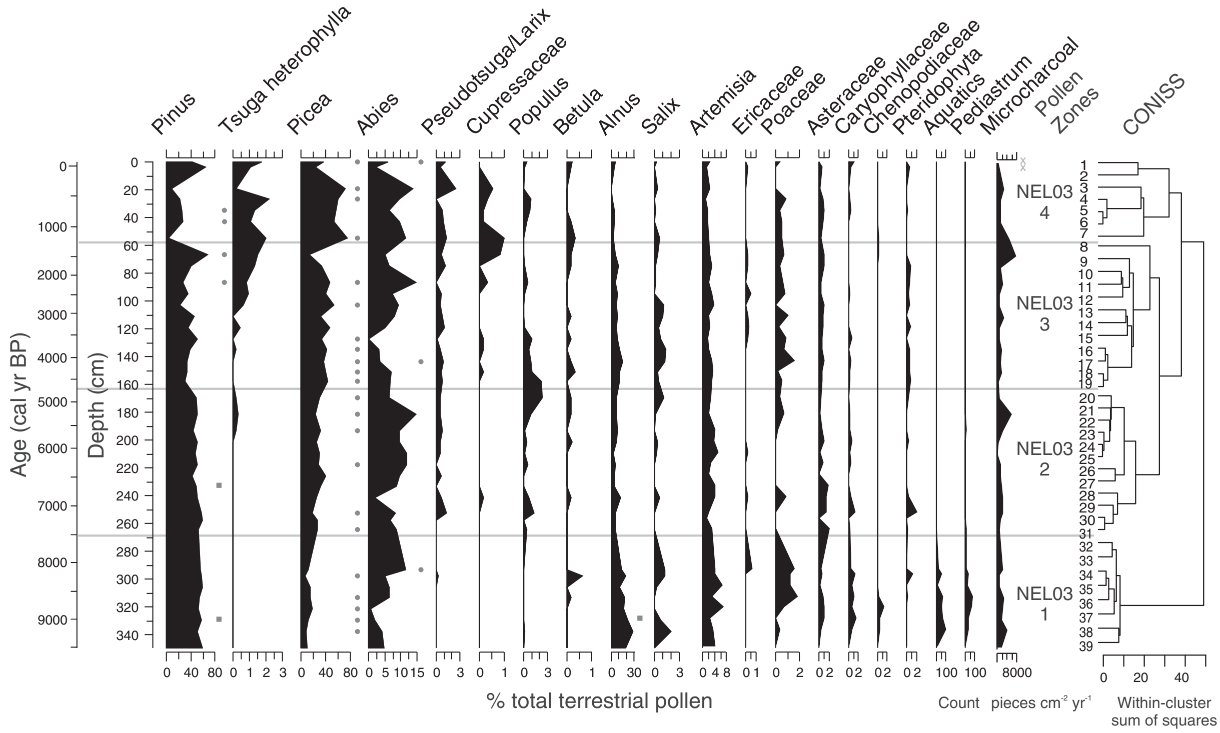

Figure 5. Pollen percentage data of selected taxa from NEL03, *Pediastrum* and aquatic pollen presented as counts and some rarer taxa are not shown. Stomata presence on pollen slides are shown by gray dots for coniferous taxa and macrofossil presence observed during macroscopic charcoal analysis is shown by a gray square. Microcharcoal, counted during pollen counts, is presented as accumulation rate and note the presence of cenospheres near the top (gray 'x' symbol). Pollen zones were established using CONISS.

the climatically controlled timberline was present at higher elevations across much of British Columbia during the early Holocene (Clague and Mathewes, 1989).

Ground fires would be necessary to spread fire between stands and the low accumulation rate of type S charcoal (Fig. 6), likely sourced from woody biomass, also suggested that tree density was low. The charcoal record of NEL03 during the early Holocene was consistent with a composite record of biomass burning from boreal regions of northwestern North America and global fire activity (Marlon et al., 2013; Fig. 7). This suggests broad-scale climatic controls influenced the fire regime of this region during the early Holocene. Summer temperatures at this time were 1–5 °C warmer than present in southeastern British Columbia (Rosenberg et al., 2004; Chase et al., 2008) and promoted a high diversity of vegetation types in *Pinus* forests that experienced a higher fire frequency at NEL03 (Figs. 5 and 6). Fire activity was highest during this period with a Weibull mFRI of 150 yr, due to increased summer insolation (Berger and Loutre, 1991) and warmer summer temperatures (Chase et al., 2008). The warmer than present conditions lengthened the fire season promoting fire ignition and spread and a longer fire season. Broad-scale climatic controls leading to warmer summer conditions were probably the dominant influence of the higher fire frequency during this period of low biomass *Pinus* forests.

Zone 2: 7500–4600 cal yr BP

ESSF forests began to establish closer to the lake between 7500 and 4600 cal yr BP (Figs. 5 and 7). The thick Mazama O ashfall of 7627 ± 150 cal yr BP may have been a significant disturbance at a time when climate-mediated vegetation changes were occurring in the catchment. A significant vegetation transition at the time of the Mazama O tephra was also found in the pollen record of Dog Lake, British Columbia (160 km to the northeast; Hallett and Hills, 2006). *Pseudotsuga*/*Larix* appeared continuously after 7000 cal yr BP and *L. lyallii* continued to be a significant proportion of the modern forest surrounding the lake. *L. lyallii* is commonly associated with ESSF forests, especially on slightly drier sites (Coupé et al., 1991). The first arrival of *Tsuga heterophylla* pollen in the record around 5200 cal yr BP was likely due to long range transport as its abundance increased within lower elevation forests (Fig. 5; Rosenberg et al., 2004; Gavin et al., 2006). Short diameter *Tsuga* pollen grains are known to be transported long distances (Gavin et al., 2005; Day et al., 2013). Deciduous taxa, including *Betula*, *Alnus*, *Artemisia*, and *Salix*, which are uncommon in ESSF forests (Coupé et al., 1991), decreased throughout the period and indicated the further establishment of ESSF forests (Fig. 5). During the early to mid Holocene (NEL03 zone 2), summer temperatures cooled as orbital geometry changes led to decreased summer insolation (Fig. 7; Berger and Loutre, 1991; Rosenberg et al., 2004; Chase et al., 2008).

The bCHAR and the arboreal pollen influx increased during this period; yet, varied considerably (Fig. 7). The bCHAR increased between 7600 and 6500 cal yr BP and fire frequency was largely unchanged with 3–4 fires 1000 yr^{-1}. A noticeable drop in bCHAR and the lowest arboreal pollen influx occurred between 6500 and 5500 cal yr BP and suggested the forest was much less productive than any other period before increasing again at 5500 cal yr BP (Figs. 6 and 7). Northwestern (boreal) North America regional biomass burning trends also exhibited an abrupt decrease at 6000 cal yr BP (Marlon et al., 2013), consistent with cold chironomid-inferred July temperatures in southeast British Columbia (Rosenberg et al., 2004; Chase et al., 2008). During this time there was a short period of increased siliciclastic content and particle sizes in the catchment that may have related to increased transport in the catchment (Fig. 4) due to the reduced forest cover and may have contributed to a reduced pollen influx (Fig. 7). The reduced pollen and charcoal accumulation rates were depressed over a 1000-yr period and suggested decreased biomass abundance caused by cooler summer temperatures ~6000 cal yr BP (Fig. 7).

Following this period of cooler summer temperatures that caused low forest production, the relative abundance of *Picea* and *Abies* pollen continued to increase, indicating further encroachment of ESSF forest into the upper mid-elevation *Pinus* forests. After 5500 cal yr BP, arboreal pollen influx and bCHAR increased and some of the largest peak

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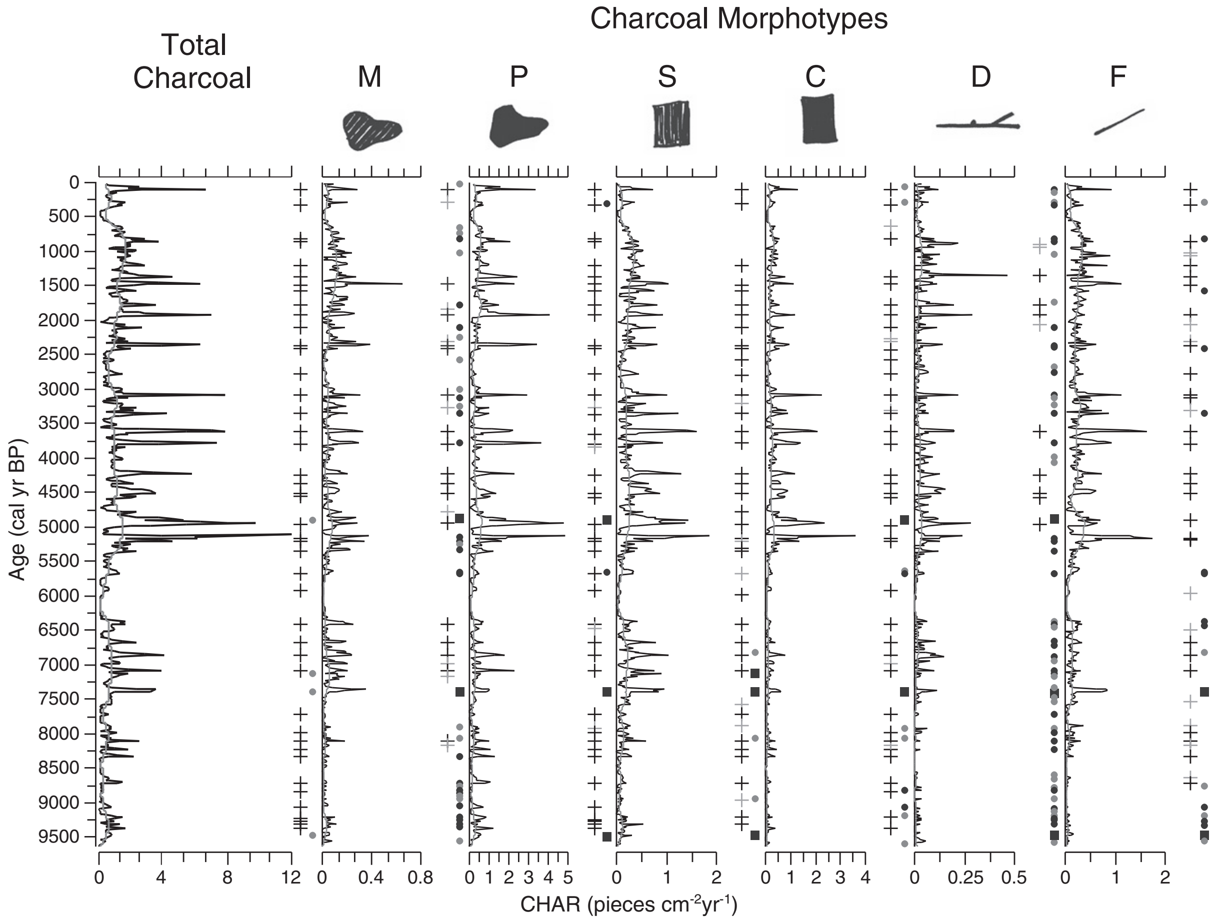

Figure 6. Fire episode reconstruction at NEL03 over the past 9650 cal yr BP using total macroscopic charcoal accumulation rates (black line) showing 44 significant peak events ('+' symbol) above the variable background rate (gray line). Gray circles represent insignificant peaks in CHAR defined by a cut-off probability of a minimum charcoal concentration value within 75 yr before a peak having ≥5% chance of coming from the same Poisson distribution as the maximum charcoal count associated with the peak (Higuera, 2009). CHAR values for each 222222macroscopic charcoal morphotype counted, using the morphotype classification of Enache and Cumming (2006), are also shown (black lines) with the associated variable background rates (gray lines). Significant peaks in morphotype CHAR that agree (≤±24 yr) with total charcoal peak events are shown by black '+' symbols, gray '+' symbols signify significant peaks that do not agree, black circles represent insignificant peaks in morphotype CHAR that agree with significant peaks in total CHAR, gray circles represent insignificant peaks in morphotype CHAR, and insignificant morphotype peaks that agree with total CHAR insignificant peaks are show by black squares.

CHARs of the record occurred (Figs. 6 and 7). The RSI of the bCHAR series indicated a significant 1000-yr period of increased biomass abundance within the watershed (Fig. 7). Conditions seemed to have promoted increased abundance of ESSF trees that supported much larger fires. Fire frequency also doubled to 6 fires 1000 yr^{-1} (Fig. 7), which suggested that the accumulation of biomass became an important bottom-up control of fire frequency. A long-term build-up of fuels, interacting with top-down fire controls, was required for fire occurrence in subalpine *Larix–Pinus–Betula* forests of the western Alps in Europe over the past 8000 yr (Blarquez and Carcaillet, 2010). During this period of cooling summer temperatures, long-term climatic controls became less important relative to the abundance of biomass fuels in the catchment.

Zone 3: 4600–1300 cal yr BP

Similar to the modern assemblage, the mid-to-late Holocene (NEL03 zone 3) forest of the catchment appeared to be dominated by ESSF species. The forests also contained *Larix*/*Pseudotsuga* (likely *L. lyallii*), *Pinus*, at least partially consisting of *P. albicaulis*, and a variety of shrubs (Fig. 5). This is consistent with earlier studies in the Columbia

Table 2
Summary of total and morphotype charcoal accumulation rates (CHAR) results.

	Total	M	P	S	C	D	F
Mean CHAR (pieces cm^{-2} yr^{-1})	0.81	0.03	0.28	0.16	0.15	0.014	0.16
Median peak return interval (years)	216	337	211	225	226	407	239
Correlation with total ($p < 0.001$)	1	0.69	0.95	0.84	0.92	0.69	0.77
Mean relative abundance of total CHAR (%)		5.3	36.4	19.7	19	2.3	17.3
Standard deviation of relative abundance of total CHAR (%)		5.8	16.2	11.9	10.3	3.6	11.3
Significant peaks	44	25	45	42	42	11	37
Insignificant peaks	4	37	5	6	12	57	18
Agreeing peaks (N and %)		17	41	37	36	8	26
		68%	91%	88%	86%	83%	70%
Agreeing insignificant morphotype CHAR peaks with significant total CHAR peaks		17	2	0	3	27	10

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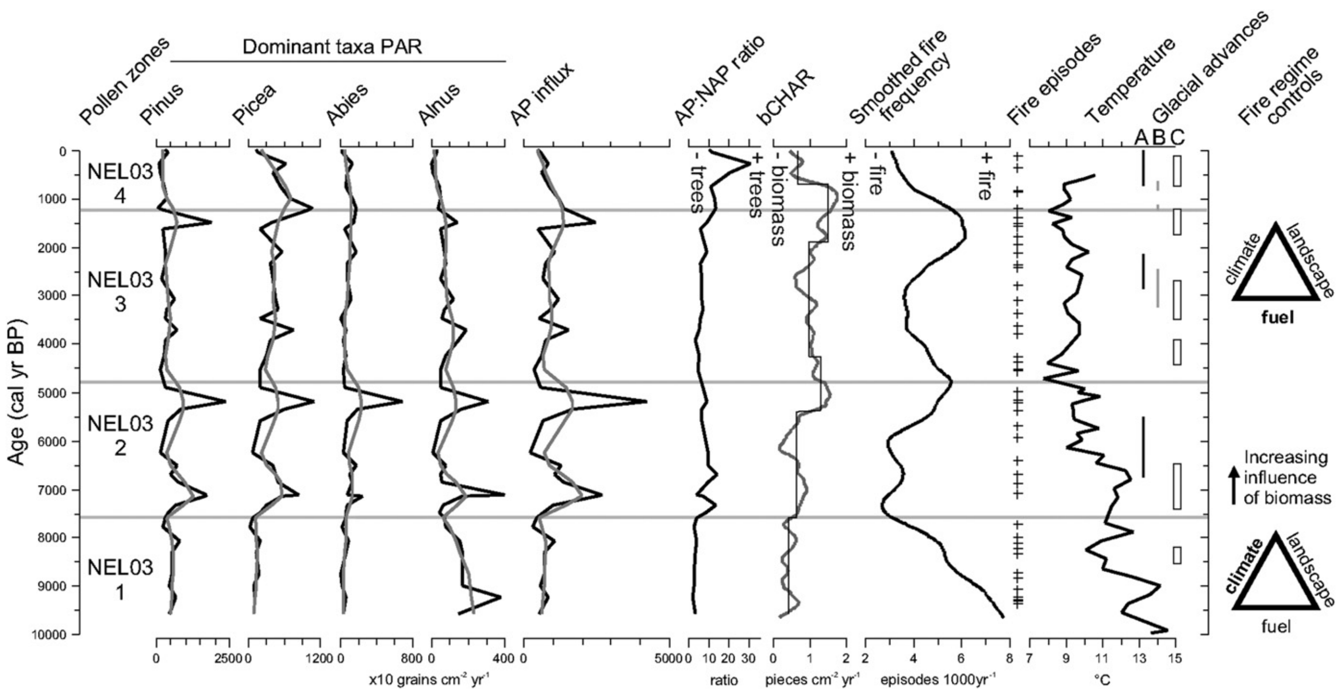

Figure 7. Holocene vegetation and disturbance history summary at the NEL03 catchment. Pollen zones represent significant changes to the terrestrial plant community. Arboreal pollen (AP) influx, smoothed with a LOWESS with a span of 0.2 (gray line), and the ratio of AP to non-arboreal pollen (NAP) are presented to show the changes to the density of trees in the area. Background charcoal accumulation rate (bCHAR) shows changes to abundance of biomass within the catchment and the solid black line shows the fire regime zones derived from a regime shift algorithm (RSI; $P = 0.0001$, cut-off length $=$ 100 yr, Huber's WF $=$ 5; Rodionov, 2004). Smoothed fire frequency shows the long-term changes to fire activity, '+' symbols represent reconstructed fire episodes. A chironomid-inferred mean July temperature reconstruction shows the general Holocene pattern in the southern interior of British Columbia (Chase et al., 2008). A) Black bars represent the maximum extents of glacier advances along the Coastal Ranges (Reyes and Clague, 2004), B) the gray bars show advances at the central Rocky Mountains (Luckman et al., 1993), and C) the white bars represent the general pattern of glacier advances across central and southern British Columbia, summarized by Menounos et al. (2009). The relative influence of regional climate, catchment landscape (topography, elevation, aspect), and fuel condition (biomass, fuel connectivity) controls of the decadal-to-centennial scale fire regime are emphasized by skewing of the fire-regime triangles based on the relative importance of each set of factors (Whitlock et al., 2010).

Mountains that found evidence of the establishment of modern forests by 4500–4000 cal yr BP (Fig. 5; Hebda, 1995; Gavin et al., 2006). *T. heterophylla* pollen was not consistently found until 4000 cal yr BP and its earliest presence was slightly earlier at NEL03 than at Eagle Lake which is located further north and at similar elevation (Rosenberg et al., 2003). The increase in *Tsuga* pollen has been attributed to decreased continentality and increased winter moisture in the Interior Wet Belt region (Gavin et al., 2011). Cupressaceae, including *Thuja plicata*, pollen increased during this period due to the continued establishment of interior cedar-hemlock (ICH) forests found at elevations below ESSF forests (Coupé et al., 1991). The forest composition was generally stable and the increased abundance of *Salix* and pteridophyte pollen (Fig. 5) suggested increased moisture at the site.

Fire frequency varied concomitant with changes to biomass (Fig. 7), which suggested that increased biomass production and fuel loads influenced fire activity. Arboreal pollen influx was relatively stable throughout the period, but increased at the beginning and ending of Zone 2 as did bCHAR values (Fig. 7). The RSI analysis also showed that these variations in bCHAR were significant. Fire frequency peaked during periods of limited glacier advances in southern British Columbia (Fig. 7), which suggested climate-mediated impacts on vegetation were an important control of the fire regime. Fire frequency seemed to peak during periods of reduced glacial activity as indicated by glacial dynamics on the coastal mountain systems, while there seemed to be a less clear pattern between fire frequency and reduced glacial advances in central British Columbia and the Rockies (Fig. 7). Climate-mediated influence of millennial-scale fire regimes was also shown in the *P. engelmannii*–*A. lasiocarpa* forests of the Markaganut plateau in Utah (Morris et al., 2013). During periods of increased biomass abundance, fire frequencies increased, doubling in frequency for a period around 2000 cal yr BP (Fig. 7).

Zone 4: 1300 cal yr BP–present

Dominance of *Picea* in these forests was reached during the most recent pollen zone. *Pinus* reached lowest influx and relative abundance, yet the forest remained similar to the previous zone (Figs. 5 and 7). The presence of *Pinus* stomata and current populations of few whitebark pine (*P. albicaulis*) suggested continued presence within the watershed throughout the Holocene. The increased Cupressaceae and *Tsuga* abundance was associated with the expansion of interior cedar-hemlock (ICH) forests found at low-to-mid elevations (Hebda, 1995; Gavin et al., 2006). The increased fire frequency at the start of this zone corresponded to periods of increased bCHAR that were related to the amount of biomass on the landscape (Fig. 7; Marlon et al., 2006). This increase occurred later than the climatically influenced regional and global increase in biomass burning identified between 3000 and 2000 cal yr BP (Marlon et al., 2013). This lag could be due to the elevation of this site, having a delayed climate-mediated vegetation change. This discrepancy with broad-scale patterns of biomass burning indicated the increased importance of local bottom-up controls of the fire regime. Fire activity between five different catchments within a 550 km^2 area of this region were largely independent, and since 2500 cal yr BP, fire activity became increasingly asynchronous between sites with a north-facing aspect versus sites with south-facing aspects (Courtney Mustaphi and Pisaric, 2013). Fire frequency decreased throughout Zone 4, possibly due to climate-mediated influences on the vegetation during the past millennium. Recent sediments since AD 1920 also contained cenospheres related to the burning of fossil fuels in the region; although, the input remained very small with only 4–10% of total microcharcoal being classified as spherules.

Comparison with other Holocene Engelmann spruce forest records

Earlier work compared the fire records from five sites from nearby watersheds, which included NEL03, and found that aspect was an important control of fire frequency and fire episode synchroneity over the past 5000 yr (Courtney Mustaphi and Pisaric, 2013). Both fire frequency and fire synchroneity were similar between NEL03 and Cooley Lake, located 50 km to the west and also situated in a north-facing ESSF forest. Forest composition records were similar over the past 7000 yr; the Cooley Lake record also showed a general decrease in *Pinus* and increasing *Picea*, with a significant, yet varied proportion of *Abies* (Gavin et al., 2006). The Cooley Lake charcoal record showed a short period of decreased charcoal concentration around 5600 cal yr BP, which was ~500 yr later than the conspicuous decrease at NEL03. This could possibly be due to the lower elevation of the Cooley Lake site relative to NEL03 (~500 m) or other spatial variability in the biomass abundance in ESSF forests.

The vegetation history of this catchment showed an initial forest that was *Pinus*-dominated with *Alnus* near the lake and relatively low terrestrial biomass within the catchment (Figs. 5 and 7). *Pinus* pollen decreased throughout the Holocene as *Picea*, *Abies*, and *Pseudotsuga*/*Larix*, increased as ESSF forest began to dominate at this elevation and became much more dense by 5500 cal yr BP. A similar Holocene vegetation sequence has been described at Morris Pond, a subalpine (3129 m asl) site in southwest Utah, having decreased *Pinus* pollen and concomitant increased *Picea* and *Abies* pollen. At this site, millennial-scale fire regimes were influenced by forest structure, biomass, and fuel connectivity (Morris et al., 2013). The millennial-scale fire history at Morris Pond was similar, with increased fire frequency during the early Holocene, variable mid Holocene fire frequency, and decreased fire frequency during the past millennium. However, the mid Holocene millennial-scale fire frequency was out of phase between the two sites by 1000 yr, having occurred earlier at the more northern NEL03 site. A conspicuous mid Holocene decrease in bCHAR occurred at Morris Pond between 6200 and 4200 cal yr BP, and overlapped with the bCHAR minima at NEL03 (6500–5500 cal yr BP; Fig. 7). The differences in the fire records suggest that climatic variability experienced at each site varied or that other local site factors may have become important controls of the fire regime. The similarly timed decrease in bCHAR was interpreted as a short-duration period where the interactions between broad-scale climatic control and local controls resulted in similar changes to biomass abundance in these ESSF forests.

Charcoal morphology record

The charcoal morphotype concentrations and accumulation rates were highly correlated throughout the Holocene, which differed from other studies that employed the same classification (Enache and Cumming, 2009). The high degree of correlation was likely due to the combination of the stationarity of the fuel types through the persistent dominance of a mixed-conifer forest and the very small catchment area of NEL03 that limited differential transport and sorting of macroscopic charcoal fragments. Enache and Cumming (2007) showed that precipitation within the watershed correlated with the robust morphotypes that were less susceptible to breakage, which suggested a link between transport and taphonomic processes and morphotype assemblages. Differential transport and sorting of charred plant parts have been observed in the field following wildfires (Scott et al., 2000) and could influence the morphotype assemblages deposited in lake sediments (Courtney Mustaphi, 2013).

Varying influence of the fire regime controls

The early Holocene had warmer summer temperatures, higher insolation, and maintained a *Pinus*-dominated forest with significant *Alnus* and produced more frequent fire events ($n = 11$, 25% of the total record); although, with much lower CHARs than the latter half of the record. By 5500 cal yr BP, the biomass of the conifer forest became dense enough to support high-severity fires (Fig. 7) and produced consistently higher peaks in CHAR, with more charcoal of all morphotype classes (Fig. 6). The interaction between a cooling climate and the change from *Pinus*-dominated to ESSF-dominated forest at the site, which remained dominant through the mid Holocene to present, led to a relatively homogeneous fire regime across the pollen zones with a mFRI of ~216 yr since 7600 cal yr BP. This suggested that vegetation composition was not a significant control of fire since the establishment of ESSF forests. There remained much variability in fire frequency that occurred across the pollen zones. At centennial-to-millennial scales, the accumulation of forest biomass controlled the fire regime. It is difficult to assess if these variations in biomass and forest production were mediated by slight variations in climate. Comparisons with regional isotope records of temperature and precipitation/evaporation do not show any clear patterns (not shown; Nelson et al., 2011; Ersek et al., 2012; Steinman et al., 2012), but peaks in fire frequency had occurred during periods of reduced glacial extents (Fig. 7; Luckman et al., 1993; Reyes and Clague, 2004; Menounos et al., 2009).

Although some influence of climate on fire regime variability was suggested by the relationship between fire frequency and glacial advances, further investigation of climate–fire interactions could be facilitated by development of high-resolution Holocene hydroclimatic records for the region (Wilson et al., 1994; Gedalof et al., 2004; Booth, 2008; Steinman et al., 2012). If hydroclimatic records could not be directly established using the NEL03 sediment archive, a high-resolution pollen record could be created to interpret centennial scale climate variability and would provide insight into post-disturbance forest successional changes. Establishing climate records from the same archive would reduce errors associated with examining other records with different chronologies and permit a detailed analysis of decadal-to-centennial scale climate influences on the fire regime.

Local site conditions seem to be important controls of the fire regime since the early Holocene climate cooled and ESSF forests began to dominate during the mid Holocene. Since the establishment of ESSF forests in the watershed, climate variability and vegetation composition changes do not fully explain the Holocene fire regime variability at NEL03 (Fig. 6); therefore, local site factors such as fuel conditions, lightning frequency, topography, and especially fuel abundance must have been important (Gavin et al., 2006; Hu et al., 2006; Gavin et al., 2007; Higuera et al., 2009; Courtney Mustaphi and Pisaric, 2013).

Conclusion

In this study, we have presented a Holocene vegetation record and a high-resolution fire history record from sedimentary macroscopic charcoal preserved in a lake sediment core from southeast British Columbia, Canada. We have shown that the early Holocene warm and dry, low biomass abundance, *Pinus*-dominated forests burned frequently. As summer climate cooled during the mid Holocene, an ESSF forest was established and was characterized by millennial-scale variability in forest biomass and related fire activity. It remains unclear to what degree slight variations in climate over the past 6000 yr influenced forest productivity, but this does support the idea that bottom-up local site factors have been important controls of the fire regime during this period. High-resolution analyses of vegetation variability may reveal more information regarding climate–vegetation–fire interactions. Additional sites in the region with Holocene-length records are needed to understand the varying importance of top-down and bottom-up controls of fire activity throughout the entire Holocene. Further high-resolution

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analyses aimed at building a mechanistic understanding of decadal-to-centennial scale climate–fire–vegetation interactions would be beneficial to land management and conservation planning in the region.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at http://dx.doi.org/10.1016/j.yqres.2013.12.002. These data include Google map of the most important areas described in this article.

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