# 11 000 years of fire history and climate in the mountain hemlock rain forests of southwestern British Columbia based on sedimentary charcoal

**Douglas J. Hallett, Dana S. Lepofsky, Rolf W. Mathewes, and Ken P. Lertzman**

**Abstract:** Little is known about the role of fire in the mountain hemlock (*Tsuga mertensiana* (Bong.) Carrière) rain forests of southern British Columbia. High-resolution analysis of macroscopic charcoal from lake sediment cores, along with 102 accelerator mass spectrometry (AMS) ages on soil charcoal, was used to reconstruct the long-term fire history around two subalpine lakes in the southern Coast and North Cascade Mountains. AMS ages on soil charcoal provide independent evidence of local fire around a lake and support the interpretation of peaks in lake sediment charcoal as distinct fire events during the Holocene. Local fires are rare, with intervals ranging from centuries to several millennia at some sites. Overall fire frequency varied continuously throughout the Holocene, suggesting that fire regimes are linked to climate via large-scale atmospheric circulation patterns. Fires were frequent between 11 000 and 8800 calendar years BP during the warm and dry early Holocene. The onset of humid conditions in the mid-Holocene, as rain forest taxa established in the region, produced a variable fire period until 3500 calendar years BP. A synchronous decrease in fire frequency from 3500 to 2400 calendar years BP corresponds to Neoglacial advances in the region and cool humid climate. A return of frequent fire between 2400 and 1300 calendar years BP suggests that prolonged summer drought occurred more often during this interval, which we name the Fraser Valley Fire Period. The present-day fire regime was established after 1300 calendar years BP.

**Résumé :** On connaît peu de chose du rôle du feu dans les forêts ombrophiles de pruche subalpine (*Tsuga mertensiana* (Bong.) Carrière) du Sud de la Colombie-Britannique. L'analyse à haute résolution du charbon de bois macroscopique provenant de carottes de sédiments lacustres ainsi que 102 mesures de l'âge du charbon de bois présent dans le sol au moyen de la technique de spectrométrie de masse par accélérateur (SMA) ont été utilisées pour reconstituer l'historique des feux à long terme autour de deux lacs subalpins sur la côte sud et le nord de la chaîne des Cascades. L'âge du charbon de bois dans le sol déterminé par la technique SMA fournit une preuve indépendante qu'un feu local est survenu autour d'un lac et supporte l'interprétation que les points observés dans le charbon de bois des sédiments lacustres correspondent à des feux distincts survenus durant l'Holocène. Les feux locaux sont rares avec des intervalles allant de quelques siècles à plusieurs millénaires dans certains sites. La fréquence générale des feux a continuellement varié durant l'holocène indiquant que le régime des feux est lié au climat via les patrons de circulation atmosphérique à grande échelle. Les feux étaient fréquents il y a 8800 à 11 000 ans pendant la période chaude et sèche du début de l'holocène. L'apparition de conditions humides au milieu de l'Holocène, lorsque se sont établis dans la région des taxons caractéristiques de la forêt ombrophile, a été à l'origine d'une période où la fréquence des feux était variable jusqu'à il y a 3500 ans. Une diminution synchrone de la fréquence des feux il y a 3500 à 2400 ans correspond à une avancée néo-glaciaire dans la région et à un climat frais et humide. Un retour à une fréquence élevée des feux il y a 2400 à 1300 ans indique que de longues périodes de sécheresse estivale sont survenues pendant cet intervalle qu'on désigne sous le nom de Période des feux de la vallée du Fraser. Le régime de feu actuel remonte à moins de 1300 ans.

[Traduit par la Rédaction]

## Introduction

Immediately after a forest fire, airborne macroscopic charcoal particles begin to accumulate in nearby lake sediments, creating a pulse of charcoal deposition indicative of a local fire event. The amount of charcoal that enters a lake depends on fire severity or intensity, fuel type and quantity, and the prevailing winds at the time of a fire (Clark 1988; Whitlock and Millspaugh 1996; Clark and Patterson 1997; Clark et al. 1998; Gardner and Whitlock 2001). Charcoal may continue to accumulate in lake sediments due to local erosion or be redeposited from the littoral zone for decades after an actual

Received 26 February 2002. Accepted 4 August 2002. Published on the NRC Research Press Web site at http://cjfr.nrc.ca on 30 January 2003.

**D.J. Hallett[1,2] and R.W. Mathewes.** Department of Biological Sciences, Simon Fraser University, Burnaby, BC V5A 1S6, Canada.
**D.S. Lepofsky.** Department of Archaeology, Simon Fraser University, Burnaby, BC V5A 1S6, Canada.
**K.P. Lertzman.** School of Resource and Environmental Management, Simon Fraser University, Burnaby, BC V5A 1S6, Canada.

[1] Corresponding author (e-mail: Douglas.Hallett@nau.edu).

[2] Present address: Center for Environmental Sciences and Quaternary Sciences Program, Box 5694, Northern Arizona University, Flagstaff, AZ 86011, U.S.A.

Can. J. For. Res. **33**: 292–312 (2003) doi: 10.1139/X02-177

fire event (Whitlock and Millspaugh 1996; Whitlock et al. 1997). This slow accumulation of charcoal in deepwater sediments contributes to a smooth background component in the charcoal record. Charcoal accumulation rates (CHAR) in lake sediment appear as a time series of high-frequency peaks (or fire events) against a slowly varying background component (Clark and Royall 1995; Long et al. 1998). One challenge with reconstructing fire frequency from CHAR records is being able to calibrate a peak event with local evidence of fire such as fire-scarred trees, historical fire dates, and stand ages from the watershed (e.g., Clark 1990; Millspaugh and Whitlock 1995). Calibration of recent CHAR peaks allows for reconstruction of fire frequency over millennial time scales.

A growing number of CHAR records in western North America document a link between variations in fire frequency and climate. Most records in the Pacific Northwest region show increased fire frequencies during the dry and warm early Holocene period (Long et al. 1998; Hallett and Walker 2000; Millspaugh et al. 2000; Mohr et al. 2000; Hallett 2001), a variable phase through the mid-Holocene, and at some sites a slight increase in fire frequency during the late Holocene (Hallett and Walker 2000; Mohr et al. 2000; Hallett 2001). More specifically, over the last 17 000 years, long-term variations in fire frequency in Yellowstone National Park show a millennial-scale response to summer insolation changes (Millspaugh et al. 2000), and in the southern Canadian Rockies, fire frequency decreased during Neoglacial advances (Hallett and Walker 2000).

In this study, we reconstruct fire history using CHAR peaks from lake sediment and calibrate them with accelerator mass spectrometry (AMS) ages on soil charcoal over the entire Holocene (Hallett 2001). Radiocarbon-dated soil charcoal from around a lake basin can be used to calibrate a specific CHAR peak as a local fire event and provide information on the location and extent of past fires (Gavin 2000). This calibration step is limited by the error associated with radiocarbon measurement and the in-built age of a charred wood sample. The in-built age is the potential for dating the inner rings of a tree or coarse woody debris, thus giving a date older than the actual fire event (Gavin 2001). The soil charcoal method may be limited by the erasure effect of recent fires eliminating information of past events as in dendrochronological records (e.g., Clark 1990; Millspaugh and Whitlock 1995), but in some temperate forests, soil charcoal can provide evidence of local fire spanning several millennia (Carcaillet and Talon 1996; Gavin 2000; Hallett 2001). This additional temporal and spatial information strengthens the interpretation of CHAR peaks as real fire events during the Holocene. Here, we use these methods in wet mountain hemlock (*Tsuga mertensiana* (Bong.) Carrière) forests where fire occurs infrequently (Brooke et al. 1970; Lertzman 1992). Long-term fire frequency records in high-elevation rain forest enhance our knowledge of Holocene climate because fire events must be preceded by summer drought (Agee 1993; Veblen and Alaback 1996). We analyze temporal variations in fire frequency over the Holocene and use the connection between drought and fire to interpret climate at our sites in southwestern British Columbia.

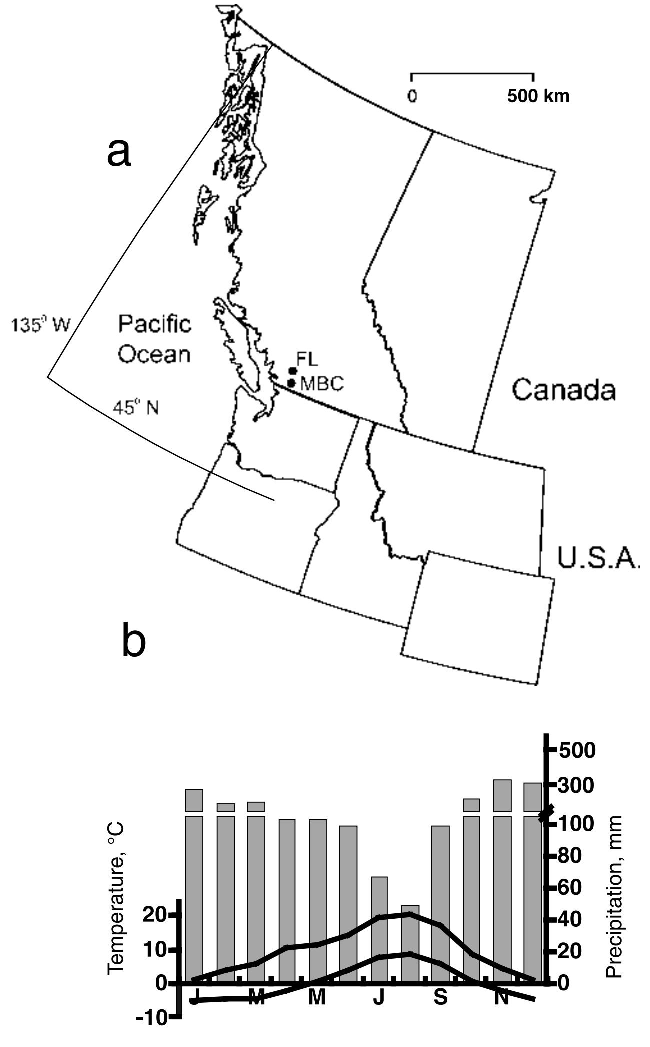
## Site description

Our study sites are located in the Mountain Hemlock (MH) zone, which is found at elevations above 900 m in the Coast and Cascade Mountains of southern British Columbia (Fig. 1). This zone can be divided into the forested subzone from 900 to 1400 m and the parkland subzone above 1400 m, although small areas of parkland can exist at lower elevations. Tree species characteristic of the forested zone include mountain hemlock, Pacific silver fir (*Abies amabilis* (Dougl. ex Loud.)), and yellow-cedar (*Chamaecyparis nootkatensis* (D. Don) Spach) (Brooke et al. 1970). Other tree species, such as subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.), whitebark pine (*Pinus albicaulis* Engelm.), and lodgepole pine (*Pinus contorta* Dougl. ex Loud.), are scattered at high elevations or on the eastern transition of the MH zone with communities more characteristic of continental conditions, while trees such as western hemlock (*Tsuga heterophylla* (Raf.) Sarg.), Douglas-fir (*Pseudotsuga menziesii* (Mirb.) Franco), and western redcedar (*Thuja plicata* Donn ex D. Don) are found at lower elevations in the Coastal Western Hemlock (CWH) zone (Meidinger and Pojar 1991).

The MH zone is characterized by cool, wet climate with significant precipitation and deep snow cover for up to 9 months (Fig. 1). Late-winter snowpack can be up to 3 m with most snow accumulation occurring in open areas between tree islands. The forested subzone is usually snow free for less than 150 days a year. Snow cover begins in October, creating an insulating layer that protects soils from freezing and cryoturbation (Brooke et al. 1970). Organic deposition and slow decomposition rates result in thick layers of mor humus that overlie Podzol soils. Little or no bioturbation occurs in these mor humus profiles because soils are cold and anaerobic due to wet snow cover for most of the year. Only in forested areas do roots cause some bioturbation in humus profiles and penetrate deep into mineral soils. In the parkland subzone, subalpine meadows are less affected by root disturbance because they are covered with herbs such as black sedge (*Carex nigricans*) and low shrubs such as pink heather (*Phyllodoce empetriformis*), white heather (*Cassiope mertensiana*), and dwarf blueberry (*Vaccinium deliciosum*). Humus deposits in late-snowmelt basins with black sedge often show signs of stratification and may include distinct layers of Mazama tephra dated at ~7585 calendar years BP (Hallett et al. 1997) and the Bridge River tephra dated at ~2388 calendar years BP (Clague et al. 1995). These two stratigraphic markers are useful in revealing rates of organic accumulation during humification.

Frozen Lake (49°36′N, 121°28′W, elevation 1180 m, depth 17 m, 3 ha), in the Coast Mountains, is located on a subalpine bench with steep slopes on the east and west sides. Water drains from upper Frozen Lake (elevation 1360 m, depth 8 m, 2 ha) through an open meadow into lower Frozen Lake and eventually the Fraser River. Mount Barr Cirque Lake (49°16′N, 121°31′W, elevation 1376 m, depth 9 m, 2 ha) is located 32 km south of Frozen Lake in the Cascade Mountains. This lake sits in a steep cirque at the base of Mount Barr with water draining into Sowerby Creek and the Fraser River. Mount Barr Cirque Lake has a steep-sided watershed with continuous MH forest on south aspects and parkland with avalanche slopes on north aspects. East Hunter Creek is a neighbouring drainage located 1.5 km northwest of Mount Barr Cirque Lake. The East Hunter Creek site supports a

**Fig. 1.** (*a*) Map of southern British Columbia with the location of the Frozen Lake (FL) and Mount Barr Cirque Lake (MBC) study sites. (*b*) MH zone climate data from Wells Creek, Mount Baker, U.S.A. (1270 m elevation and approximately 60–90 km southwest of our sites) for the period AD 1996–2000 (data from http://wrcc.dri.edu/snotel/). Bars represent monthly precipitation and lines represent maximum and minimum temperature over the year.



southeast aspect of subalpine forest from 1140 m to open parkland at 1650 m elevation. Our last site at 1300 m elevation is Mount Cheam, located 30 km west of Mount Barr Cirque Lake in a parkland environment below the steep avalanche slopes of Mount Cheam.

Present-day climate of southern British Columbia is influenced by major air masses associated with the semi-permanent pressure systems over the Pacific Ocean. The Aleutian Low dominates during winter when westerly storm tracks deliver abundant moisture to the coast (Lydolph 1985; McBean 1996). Winter precipitation usually falls as snow at higher elevations, producing large accumulations (Brooke et al. 1970). The eastern Pacific subtropical high pressure system (Pacific High) strengthens each spring as it extends north and dominates during the summer months. The anticyclonic flow of the Pacific High brings dry air masses to western North America and blocks storm fronts from the Aleutian Low. The blocking Pacific High produces warm and dry conditions in summer and increases the potential for extended drought and fire at subalpine elevations (Agee 1993).

Periods with large areas burned in western Canada correlate with anomalous 500-hPa geopotential height values over the region (Johnson and Wowchuk 1993; Skinner et al. 1999; Skinner et al. 2002). These persistent ridges of high pressure create the conditions needed for large forest fires, which include dry convective storms, lightning, and high winds (Agee 1993; Rorig and Ferguson 1999). In coastal forests, large stand-destroying fires are often associated with dry gradient winds from the east that behave as foehn winds originating from inland high pressure cells (Agee 1993). A short fire season is limited to July through September, provided snow has melted and forest fuels have dried. Detailed historic records of fire in MH forests are rare because tree ring based fire history methods are not effective due to a scarcity of fire-scarred trees and even-aged stands used to date fire (Agee 1993). Stands are typically multi- to all-aged and tree age is not an indicator of stand age because the time between stand-initiating fires is often longer than the life span of individual trees. Fine-scale gap processes tend to dominate MH forest structure and dynamics (Lertzman and Krebs 1991; Lertzman 1992). Therefore, a charcoal-based fire history is the best method for determining the role of fire in these infrequently burned environments.

## Materials and methods

### Field and laboratory method

#### *Soil charcoal*

We sampled soil charcoal by collecting profiles of deep organic humus (up to 50 cm thick) overlying mineral horizons. Large rectangular sections (e.g., 10 × 10 × ~30 cm) were excavated using a shovel, wrapped in aluminum foil, and transported back to the laboratory. We sampled in areas with distinct depositional sequences and evidence of charcoal, which is mainly in open parkland and late snowmelt basins with minimal tree rooting. At East Hunter Creek, we sampled humus profiles from forested slopes and in open parkland using a 5-cm-diameter tube. A total of 92 humus profiles were processed from subalpine forest, parkland, and late-snowmelt basins at Frozen Lake (40), Mount Barr Cirque Lake (33), East Hunter Creek (12), and Mount Cheam (7). Humus profiles were described and subsampled in contiguous 1-cm intervals to construct a record of charcoal presence (or absence) throughout the depositional sequence. Subsamples were soaked in warm 6% KOH to loosen humic material and then wet-sieved for charcoal fragments larger than 0.5 mm. After sieving, charcoal deposits were noted and compared in relation to the age and depth of tephra layers.

#### *Lake sediments*

We extracted a continuous 300-cm-long core from Frozen Lake, a 220-cm-long core from upper Frozen Lake, and a 290-cm-long core from Mount Barr Cirque Lake using a percussion corer. Cores were subsampled contiguously in

1-cm increments using 10-cm$^3$ volumes of sediment for their entire length. Magnetic susceptibility measurements were performed on each sample using a Sapphire Instruments meter at the University of Oregon (Eugene, Oreg.). Sediment samples were soaked for 24 h in 10% sodium hexametaphosphate to deflocculate organic matter. The treated samples were gently washed through a 125-μm-mesh sieve and residues placed in a gridded Petri dish for counting macroscopic charcoal using a dissecting microscope at 400× magnification.

### Data analysis

#### *Soil charcoal*

We identified and dated as many separate charcoal layers as possible using a limited number of AMS dates. The presence of volcanic tephra in numerous profiles allowed us to identify charcoal-rich (or charcoal-poor) periods without radiocarbon dating. The presence or absence of charcoal at a site was noted for specific time portions of the humus profiles. We dated only discrete charcoal layers from stratified humus profiles and tried to avoid replication of previously dated fire events. We submitted 10–15 charcoal samples at a time for AMS radiocarbon dating and used results from subsequent samples to decide on further submissions. The lake sediment charcoal records provided information on the potential timing of fire events around a site. At Frozen Lake, Mount Barr Cirque Lake, and East Hunter Creek we feel confident that most or all local fires were dated in the humus profiles we collected (Tables 1 and 2). Six charcoal dates from a single humus profile do not represent a complete reconstruction of local fire at Mount Cheam.

#### *CHAR in lake sediment*

To detect individual fires, sediment sampling resolution must be shorter than 600 to 1500 year fire intervals expected for subalpine MH forests (Lertzman 1992; Agee 1993). CHAR was calculated by dividing particle concentration values by the average sediment deposition time between AMS ages in the sediment core (Table 3). We used the decomposition methods of Long et al. (1998) to separate the CHAR time series into a slowly varying background component and a higher frequency peak component.

Background charcoal is strongly related to factors that affect charcoal production, such as standing biomass, fuel load and fire severity, and the effects of sedimentation processes (Millspaugh and Whitlock 1995; Clark and Patterson 1997; Long et al. 1998). Deposition of macroscopic charcoal in sediments depends on local wind, erosion, and fluvial processes that transport charcoal into a basin (Clark 1988). Sedimentary processes within a lake then influence where and when charcoal will deposit in the deepest portion of the basin. Background charcoal may reflect the contribution of material from the littoral zone over decades after a fire or the addition of charcoal from nonlocal or regional fires (Whitlock and Millspaugh 1996; Whitlock et al. 1997; Clark et al. 1998; Gardner and Whitlock 2001). The high-frequency peak component in a CHAR series represents charcoal from a fire within the watershed. Peaks may also result from minor noise in the series related to analytical error in charcoal determination and random variations in CHAR due to sedimentary processes.

CHAR time series were produced by interpolating the raw charcoal concentration values and deposition times to evenly spaced 10-year intervals. This resampling approach allows us to analyze the data at equally spaced intervals and retains the features of the series. A locally weighted 400-year moving average was used to calculate the background component in the CHAR series based on past studies (Long et al. 1998; Hallett and Walker 2000; Millspaugh et al. 2000; Mohr et al. 2000) and numerous tests with these data. Window widths greater than 400 years did not detect all local fires identified by soil charcoal ages. The continuously varying background was used as a CHAR value threshold to determine which peaks (fire events) rise above the noise subcomponent of the record. The threshold is defined as a ratio of CHAR divided by local background value. A single fire event is identified when the peak exceeds the background value for the first time. A CHAR value threshold ratio of 1.00 was needed to detect all of the local fires around our lake sites throughout the Holocene and higher values tended to underestimate real fire events. A binary series of fire events (1 = fire event, 0 = no fire event) was produced and plotted as a "+" beside the original CHAR series. To produce a fire frequency summary, we used a locally weighted mean to smooth the binary series and present the frequency of fire events (number of fire events per 1000 years) for the length of the record (Long et al. 1998). Mean fire return intervals along with their standard errors were calculated using the binary series of fire peaks for the entire length of the record, for fire frequency zones that we defined visually, and for peaks linked to local soil charcoal evidence.

## Results

### Soil charcoal

Most of the 102 AMS soil charcoal ages and their calibrated age ranges (Stuiver et al. 1998) are derived from open parkland and late-snowmelt basins because these environments had the best charcoal stratigraphy (Tables 1 and 2). Fifteen of our radiocarbon-dated profiles and many of our undated sections had visible tephra layers (Tables 1 and 2). The most abundant charcoal deposits in our humus profiles were located just above the Bridge River tephra and below the Mazama tephra. We dated mainly late and mid-Holocene charcoal deposits because early Holocene charcoal was found in mineral soil that was crumbly and nonstratified. Charcoal found in humus accumulation began at ~9500 calendar years BP. We suspect that warmer summers and colder winters due to increased seasonality led to higher decomposition rates and cryoturbation in the earliest Holocene soils (Mathewes and Heusser 1981; Mathewes 1985; Thompson et al. 1993).

No dated charcoal samples were inverted in age compared with the positions of the two tephras. Ten charcoal samples near the Bridge River tephra dated to within a few hundred years of its accepted age (~2388 calendar years BP; Table 3) (Clague et al. 1995). For example, charcoal ages located 1, 2, and 3 cm above the tephra dated at ~2355, 2176, and 2218 calendar years BP (Tables 1 and 2), suggesting that the

**Table 1.** Calibrated and uncalibrated AMS $^{14}$C ages for soil charcoal from the Frozen Lake area.

| Location code | Sample code | Monolith site code | Terrain type[a] | Vegetation type[b] | Organic horizon[c] | Soil horizon | Proximal tephra[d] | Depth below surface (cm) | CAMS laboratory No. | Uncalibrated $^{14}$C age ($^{14}$C years BP) | $\delta^{13}$C | Calibrated age with 2-sigma age range (calendar years BP)[e] |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| fl1 | fl3 | FL1-B | OP | Vd, Pe, Cm | Hh | | | 9 | 44126 | 1670±50 | –23.02 | 1706 (1553) 1418 |
| fl1 | fl4 | FL1-H | HF | Va, Tm, Aa | | Ae | | 18 | 44127 | 4480±50 | –25.26 | 5307 (5212, 5188, 5116, 5113, 5050) 4873 |
| fl1 | fl6 | FL1-I | HF | Va, Tm, Aa | | Ah | | 5 | 44129 | 920±50 | –21.55 | 932 (889, 864, 827, 812, 794) 728 |
| fl1 | fl32 | FL1-2 | OP | Vd, Pe, Cm | Hh | | | 13 | 44155 | 1320±50 | –22.85 | 1308 (1267) 1170 |
| fl1 | fl33 | FL1-2 | OP | Vd, Pe, Cm | Hh | | | 15 | 44156 | 1300±50 | –22.84 | 1301 (1261) 1088 |
| fl1 | fl34 | FL1-2 | OP | Vd, Pe, Cm | Hh | | | 19 | 44157 | 5410±60 | –28.89 | 6306 (6269, 6243, 6201) 5996 |
| fl1 | fl35 | FL1-3 | OP | Vd, Pe, Cm | Hh | | | 16 | 44158 | 4440±50 | –22.20 | 5295 (5039, 5006, 4993) 4866 |
| fl1 | fl36 | FL1-5 | OP | Vd, Pe, Cm | Hh | | | 8 | 44159 | 4520±50 | –24.68 | 5317 (5284, 5159, 5140, 5101, 5086) 4976 |
| fl1 | fl37 | FL1-5 | OP | Vd, Pe, Cm | Hh | | | 10 | 44160 | 4390±50 | –27.88 | 5261 (4966, 4926, 4920, 4896, 4889) 4845 |
| m | fl19 | MD-1 | OP | Vd, Pe, Cm | Hh | | MZ, 18 cm | 14 | 44142 | 5490±60 | –24.13 | 6406 (6289) 6123 |
| m | fl1 | MD-4 | OP | Vd, Pe, Cm | Fm | | | 3 | 44124 | 30±50 | –24.23 | 256 (0) 0 |
| m | fl7 | M-2 | OP | Vd, Pe, Cm | Hh | | | 6 | 44130 | 800±40 | –25.14 | 787 (694) 665 |
| m | fl61 | M-2 | OP | Vd, Pe, Cm | Hh | | | 9 | 48560 | 3160±40 | –23.93 | 3467 (3378) 3269 |
| m | fl8 | M-2 | OP | Vd, Pe, Cm | Hh | | | 12 | 44131 | 3150±50 | –23.40 | 3469 (3375, 3369, 3363) 3262 |
| m | fl9 | M-4 | OP | Vd, Pe, Cm | Hh | | MZ, 27 cm | 8 | 44132 | 840±50 | –23.90 | 911 (734) 669 |
| m | fl10 | M-6 | OP | Vd, Pe, Cm | Hh | | | 16 | 44133 | 2140±50 | –23.92 | 2310 (2144, 2144, 2122) 1953 |
| m | fl11 | M-6 | OP | Vd, Pe, Cm | Hh | | | 23 | 44134 | 5980±50 | –23.42 | 6937 (6845, 6843, 6795, 6767, 6760) 6671 |
| m | fl12 | M-6 | OP | Vd, Pe, Cm | Hh | | | 25 | 44135 | 5070±50 | –27.23 | 5924 (5886, 5825, 5820, 5811, 5756) 5661 |
| m | fl13 | M-6 | OP | Vd, Pe, Cm | Hh | | | 28 | 44136 | 6330±50 | –24.73 | 7414 (7253) 7099 |
| m | fl51 | M-6 | OP | Vd, Pe, Cm | | Bh | | 34 | 48550 | 9510±50 | –23.89 | 11090 (10740) 10582 |
| m | fl15 | M-14 | OP | Vd, Pe, Cm | Fm | | | 4 | 44138 | 230±40 | –26.18 | 420 (291) 3 |
| m | fl60 | M-14 | OP | Vd, Pe, Cm | Hh | | | 5 | 48559 | 100±40 | –24.86 | 272 (61, 43, 0) 0 |
| m | fl59 | M-14 | OP | Vd, Pe, Cm | Hh | | | 6 | 48558 | 130±40 | –26.53 | 281 (255, 225, 134, 28, 0) 0 |
| m | fl58 | M-14 | OP | Vd, Pe, Cm | Hh | | | 7 | 48557 | 140±40 | –26.20 | 284 (261, 137, 25, 0) 0 |
| m | fl14 | M-14 | OP | Vd, Pe, Cm | Hh | | | 8 | 44137 | 930±40 | –25.02 | 931 (907, 859, 831, 810, 795) 736 |
| m | fl57 | M-14 | OP | Vd, Pe, Cm | Hh | | | 9 | 48556 | 200±40 | –26.50 | 303 (278, 169, 155, 4, 0) 0 |
| m | fl56 | M-14 | OP | Vd, Pe, Cm | | Bh | | 14 | 48555 | 8080±50 | –23.31 | 9238 (9011) 8782 |
| m | fl16 | M-15 | OP | Vd, Pe, Cm | Fm | | | 3 | 44139 | 200±50 | –24.46 | 313 (282, 168, 155) 0 |
| m | fl17 | M-16 | OP | Vd, Pe, Cm | Fm | | | 3 | 44140 | 860±50 | –23.44 | 918 (758, 751, 742) 672 |
| m | fl18 | M-17 | OP | Vd, Pe, Cm | Hh | | | 14 | 44141 | 5110±60 | –25.22 | 5988 (5905) 5720 |
| fl2 | fl2 | FL2-E | HF | Va, Tm, Aa | | Ae | | 10 | 44125 | 1700±50 | –23.59 | 1713 (1606, 1580, 1571) 1518 |
| fl2 | fl5 | FL2-L | HF | Va, Tm, Aa | Hh | | | 13 | 44128 | 3540±50 | –23.90 | 3974 (3831, 3785, 3783) 3689 |
| fl2 | fl20 | FL2-1 | OP | Vd, Pe, Cm | Hh | | | 11 | 44143 | 1930±130 | –22.72 | 2297 (1875) 1550 |
| fl2 | fl22 | FL2-9 | OP | Vd, Pe, Cm | Hh | | | 5 | 44145 | 190±50 | –23.55 | 306 (275, 173, 152, 7, 0) 0 |
| fl2 | fl62 | FL2-9 | OP | Vd, Pe, Cm | | Ah | | 7 | 48561 | 1410±40 | –22.91 | 1387 (1304) 1269 |
| fl2 | fl21 | FL2-9 | OP | Vd, Pe, Cm | | Ae | MZ, 13 cm | 9 | 44144 | 1490±40 | –23.44 | 1512 (1352) 1301 |
| fl2 | fl26 | FL2-10 | LSB | Cn | Hh | | BR, 7 cm | 5 | 44149 | 2190±50 | –22.16 | 2337 (2296, 2270, 2176, 2172, 2153) 2042 |
| fl2 | fl25 | FL2-10 | LSB | Cn | Hh | | MZ, 22 cm | 12 | 44148 | 6390±60 | –22.09 | 7427 (7311, 7294, 7294) 7165 |
| fl2 | fl24 | FL2-10 | LSB | Cn | | Ah | | 19 | 44147 | 8120±50 | –23.20 | 9258 (9027) 8995 |
| fl2 | fl23 | FL2-12 | LSB | Cn | Hh | | BR, 9 cm | 5 | 44146 | 660±60 | –24.34 | 688 (651, 575, 575) 540 |
| fl2 | fl55 | FL2-14 | LSB | Cn | Fm | | | 2 | 48554 | 360±50 | –25.64 | 512 (459, 347, 341) 301 |

| | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| f12 | f131 | FL2-14 | LSB | Cn | Fm | | | 3 | 44154 | 640±60 | –24.95 | 675 (648, 581, 568) 532 |
| f12 | f154 | FL2-14 | LSB | Cn | Hh | | | 4 | 48553 | 1290±40 | –24.76 | 1291 (1259, 1247, 1242, 1197, 1192) 1142 |
| f12 | f130 | FL2-14 | LSB | Cn | Hh | | BR, 9 cm | 6 | 44153 | 1600±50 | –24.73 | 1608 (1520) 1353 |
| f12 | f129 | FL2-14 | LSB | Cn | Hh | | | 12 | 44152 | 4990±50 | –25.51 | 5892 (5721) 5605 |
| f12 | f153 | FL2-14 | LSB | Cn | Hh | | | 13 | 48552 | 4960±40 | –25.75 | 5855 (5659) 5602 |
| f12 | f152 | FL2-14 | LSB | Cn | Hh | | | 14 | 48551 | 4860±40 | –23.05 | 5656 (5596) 5488 |
| f12 | f128 | FL2-14 | LSB | Cn | Hh | | MZ, 19 cm | 15 | 44151 | 5780±70 | –23.21 | 6748 (6620, 6608, 6600, 6588, 6567) 6409 |
| f12 | f127 | FL2-14 | LSB | Cn | | Bh | | 22 | 44150 | 8100±40 | –27.56 | 9240 (9014) 8994 |
| f12 | f150 | FL2-14 | LSB | Cn | | Bh | | 23 | 48549 | 8100±40 | –24.17 | 9240 (9014) 8994 |

[a] OP, open Parkland; HF, hillslope forest; LSB, late-snowmelt basin.

[b] Vd, *Vaccinium deliciosum*; Pe, *Phyllodoce empetriformis*; Cm, *Cassiope mertensiana*; Va, *Vaccinium alaskense*; Tm, *Tsuga mertensiana*; Aa, *Abies amabilis*; Yc, *Chamaecyparis nootkatensis*; Cn, *Carex nigricans*.

[c] Humus form classification.

[d] MZ, Mazama tephra layer (7585 calendar years BP); BR, Bridge River tephra layer (2388 calendar years BP).

[e] Intercept ages within parentheses along with upper and lower age range are based on Stuiver et al. (1998).

in-built age of charcoal may be smaller than at lower elevation coastal western hemlock sites (Gavin 2001). There were 15 (15%) inverted ages of soil charcoal in humus profiles with only four (4%) older by more than 200 years and just two older by several thousand years. The two large inversions were most likely the result of mass erosion at East Hunter Creek in a late-snowmelt basin site. The 11 minor inversions were well within the 2-sigma measurement error of radiocarbon and calibration. In these cases, the in-built age of wood charcoal may be responsible (Gavin 2001), or a mixing event may have caused the inversion. In general, most of the ages (87 or 85% of the ages) were in agreement with their depositional sequence and provided information on fire return intervals and time since fire at specific sites in the watershed (Lertzman et al. 2002).

The cumulative probability distributions of 102 AMS ages show a gap or fire-free interval older than 9500 calendar years BP, which is the result of not dating samples from older mineral soil horizons (Fig. 2). Gaps from ~9000 to 8400 and from ~8000 to 7400 calendar years BP are intervals without fire. Local fire evidence indicates that fire-free intervals can last for many centuries and even millennia at some sites. The mapped midintercept of the AMS ages at Frozen Lake shows that most charcoal ages were located near the upper lake and in the meadow between the two lakes (Fig. 3). Open parkland areas and southeast- and southwest-facing areas at the base of forested slopes contained the best soil charcoal deposits and reflect most of the dated sites. At Mount Barr Cirque Lake, most soil charcoal deposits were also found at the base of south-facing forested slopes in an open parkland environment. North-facing avalanche slopes with isolated tree islands are largely free of charcoal and suggest that fires were rare on wetter slopes (Fig. 4).

## Age–depth relationship for Frozen Lake and Mount Barr Cirque Lake cores

We used a simple linear interpolation between AMS dates on conifer needles, plant macrofossils, and tephra layers to infer age from depth in each core (Fig. 5). The AMS $^{14}$C dates were converted to calendar ages (calendar years BP) using the calibration program (CALIB 4.3) of Stuiver et al. (1998) (Table 3). The accepted ages of the Bridge River (Clague et al. 1995) and Mazama (Hallett et al. 1997) tephras were used in the chronologies. The average deposition time of Frozen Lake is ~45 years/cm with 1-cm sampling times ranging from 19 to 71 years/cm. We disregarded the AMS date at 52 cm because it was older than the Bridge River tephra layer located at 60 cm. Mount Barr Cirque Lake sedimentation averages ~15 years/cm for the entire core with sampling times ranging from 9 to 59 years/cm. The period from 2400 to 1200 calendar years BP shows high sedimentation rates in both cores, but particularly in the Mount Barr Cirque Lake core, where five AMS ages indicate a very high sedimentation rate (Fig. 5). This is most likely due to erosion after frequent or severe fire events around the lakes (Fig. 2).

## Lithology of lake sediments

Frozen Lake sediments consist of fine laminated dark-brown gyttja in the core from 0–238 cm depth. The Bridge

**Table 2.** Calibrated and uncalibrated AMS $^{14}C$ ages for soil charcoal from the Mount Barr Cirque Lake, East Hunter Creek, and Mount Cheam sites.

| Location code | Sample code | Monolith site code | Terrain type$^{a}$ | Vegetation type$^{b}$ | Organic horizon$^{c}$ | Soil horizon | Proximal tephra$^{d}$ | Depth below surface (cm) | CAMS laboratory No. | Uncalibrated $^{14}C$ age ($^{14}C$ years BP) | $\delta^{13}C$ | Calibrated age with 2-sigma age range (calendar years BP)$^{e}$ |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| mbc | mbc18 | MBC-B | LSB | Cn | Fm | | | 5 | 57043 | 1490±40 | –24.60 | 1512 (1352) 1301 |
| mbc | mbc30 | MBC-B | LSB | Cn | Hh | | | 20 | 61981 | 2150±40 | –24.91 | 2307 (2146, 2139, 2125) 2002 |
| mbc | mbc31 | MBC-B | LSB | Cn | Hh | | | 36 | 61982 | 2230±40 | –24.59 | 2343 (2306, 2235, 2207, 2192, 2183) 2123 |
| mbc | mbc32 | MBC-B | LSB | Cn | Hh | | | 48 | 61983 | 2300±40 | –25.22 | 2354 (2339) 2161 |
| mbc | mbc33 | MBC-B | LSB | Cn | Hh | | | 61 | 61984 | 2250±40 | –22.90 | 2346 (2312, 2218, 2211) 2149 |
| mbc | mbc1 | MBC-D | LSB | Cn | Hh | | | 7 | 53972 | 2320±40 | –22.91 | 2357 (2344) 2184 |
| mbc | mbc34 | MBC-D | LSB | Cn | Hh | | BR, 22 cm | 25 | 61985 | 2520±40 | –25.07 | 2747 (2712, 2625, 2323) 2363 |
| mbc | mbc2 | MBC-D | LSB | Cn | Hh | | | 30 | 53973 | 4140±40 | –24.41 | 4828 (4805, 4763, 4644, 4641, 4629) 4526 |
| mbc | mbc19 | MBC-D | LSB | Cn | Hh | | | 36 | 57044 | 5390±40 | –23.10 | 6286 (6197) 5998 |
| mbc | mbc3 | MBC-D | LSB | Cn | Hh | | | 39 | 53974 | 7310±50 | –24.59 | 8190 (8153, 8138, 8129, 8121, 8111, 8082, 8059) 7979 |
| mbc | mbc21 | MBC-E | LSB | Cn | Hh | | MZ, 30 cm | 15 | 57046 | 1660±40 | –25.00 | 1691 (1543) 1422 |
| mbc | mbc20 | MBC-F | LSB | Cn | Hh | | BR, 27 cm | 49 | 57045 | 8210±40 | –28.20 | 9397 (9243, 9218, 9204, 9204, 9189, 9174, 9132) 9027 |
| mbc | mbc25 | MBC-4 | OP | Vd, Pe, Cm | Hh | | BR, 27 cm | 17 | 57050 | 1590±40 | –24.10 | 1557 (1517) 1355 |
| mbc | mbc15 | MBC-8 | OP | Vd, Pe, Cm | Hh | | BR, 15 cm | 10 | 57040 | 1500±40 | –25.20 | 1516 (1388, 1358, 1354) 1306 |
| mbc | mbc4 | MBC-9 | OP | Vd, Pe, Cm | Hh | | | 6 | 53975 | 1710±40 | –25.49 | 1711 (1685, 1683, 1609, 1575, 1575) 1527 |
| mbc | mbc5 | MBC-9 | OP | Vd, Pe, Cm | Hh | | | 9 | 53976 | 1810±40 | –23.73 | 1862 (1714) 1612 |
| mbc | mbc16 | MBC-9 | OP | Vd, Pe, Cm | Hh | | BR, 12 cm | 11 | 57041 | 2390±60 | –24.60 | 2713 (2355) 2329 |
| mbc | mbc17 | MBC-9 | OP | Vd, Pe, Cm | Hh | | MZ, 22 cm | 23 | 57042 | 8190±40 | –26.60 | 9396 (9233, 9221, 9184, 9183, 9129, 9098, 9091) 9025 |
| mbc | mbc6 | MBC-9 | OP | Vd, Pe, Cm | | Ah | | 25 | 53977 | 8280±40 | –23.73 | 9466 (9395, 9390, 9365, 9363, 9278) 9092 |
| mbc | mbc26 | MBC-10 | OP | Vd, Pe, Cm | Hh | | BR, 16 cm | 22 | 57051 | 5340±50 | –22.90 | 6279 (6171, 6135, 6114, 6065, 6063) 5949 |
| mbc | mbc22 | MBC-14 | LSB | Cn | Hh | | BR, 12 cm | 9 | 57047 | 2250±30 | –24.20 | 2343 (2312, 2218, 2211) 2152 |
| mbc | mbc23 | MBC-14 | LSB | Cn | Hh | | | 21 | 57048 | 4700±40 | –21.70 | 5582 (5462, 5361, 5332) 5316 |
| mbc | mbc24 | MBC-14 | LSB | Cn | Hh | | MZ, 39 cm | 28 | 57049 | 3950±50 | –25.50 | 4526 (4415) 4244 |
| ehc | sft6 | EHSFT | HF | Va, Aa, Yc, Tm | Hh | | | 13 | 34885 | 1380±60 | –23.30 | 1390 (1292) 1178 |
| ehc | sft4 | EHSFT | HF | Va, Aa, Yc, Tm | Hh | | | 14 | 35100 | 1440±60 | –24.20 | 1477 (1327, 1322, 1312) 1264 |
| ehc | sft3 | EHSFT | HF | Va, Aa, Yc, Tm | Hh | | | 17 | 34890 | 1220±50 | –28.10 | 1273 (1171) 991 |
| ehc | sft5 | EHSFT | HF | Va, Aa, Yc, Tm | | Bh | | 37 | 35099 | 7530±60 | –23.90 | 8412 (8356) 8183 |
| ehc | uf5 | EHUF-1 | HF | Aa, Tm, Yc | Hh | | | 11 | 35101 | 1670±60 | –26.20 | 1710 (1553) 1413 |
| ehc | uf1 | EHUF-1 | HF | Aa, Tm, Yc | Hh | | | 12 | 35102 | 1740±50 | –25.20 | 1816 (1690, 1669, 1659, 1653, 1628) 1531 |
| ehc | uf2 | EHUF-1 | HF | Aa, Tm, Yc | Hh | | | 14 | 34883 | 2140±60 | –26.70 | 2304 (2144, 2144, 2122) 1998 |
| ehc | uf3 | EHUF-1 | HF | Aa, Tm, Yc | Hh | | | 16 | 34882 | 2200±60 | –26.60 | 2346 (2298, 2267, 2177, 2170, 2156) 2010 |
| ehc | uf7 | EHUF-2 | HF | Tm, Aa, Yc | Hh | | | 24 | 34884 | 1730±60 | –23.40 | 1818 (1689, 1672, 1626, 1619, 1615) 1521 |
| ehc | uf8 | EHUF-2 | HF | Tm, Aa, Yc | Hh | | | 25 | 34889 | 1660±50 | –25.30 | 1694 (1543) 1415 |
| ehc | uf9 | EHUF-2 | HF | Tm, Aa, Yc | Hh | | | 27 | 34880 | 1940±60 | –26.20 | 2001 (1881) 1722 |
| ehc | uf10 | EHUF-2 | HF | Tm, Aa, Yc | Hh | | | 31 | 34891 | 2010±60 | –25.40 | 2121 (1985, 1983, 1967, 1962, 1949) 1824 |
| ehc | uf11 | EHUF-2 | HF | Tm, Aa, Yc | | Bh | | 38 | 34888 | 8010±60 | –23.90 | 9029 (8993) 8639 |
| ehc | eh1 | EHC-1 | HF | Aa, Yc, Tm | | Ae | | 35 | 44123 | 1680±40 | –26.24 | 1694 (1562) 1518 |
| ehc | rt7 | EHRT-4 | LSB | Cn | Fm | | | 4 | 34879 | 6170±60 | –24.80 | 7248 (7153, 7128, 7086, 7080, 7022) 6810 |
| ehc | rt6 | EHRT-4 | LSB | Cn | Hh | | | 6 | 35103 | 8300±60 | –25.60 | 9473 (9397, 9385, 9370, 9360, 9346, 9343, 9297) 9035 |
| ehc | rt5 | EHRT-4 | LSB | Cn | Hh | | | 13 | 34887 | 3310±60 | –24.20 | 3690 (3551, 3508, 3483) 3390 |

| | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| ehc | rt4 | EHRT-4 | LSB | Cn | Hh | | | 16 | 34881 | 3670±50 | −23.70 | 4148 (3982) 3839 |
| ehc | rt3 | EHRT-4 | LSB | Cn | Hh | | | 19 | 34878 | 6220±60 | −25.30 | 7267 (7177, 7172, 7159, 7110, 7097) 6947 |
| ehc | rt2 | EHRT-4 | LSB | Cn | | Ah | | 21 | 34886 | 7470±60 | −25.00 | 8390 (8329, 8256, 8251, 8232, 8218) 8169 |
| ehc | rt1 | EHRT-4 | LSB | Cn | | Ah | | 23 | 34877 | 8060±50 | −23.90 | 9228 (9007) 8776 |
| ehc | eh2 | EHR-1 | OP | Vd, Pe, Cm | Hh | | | 2 | 48562 | 2140±40 | −24.39 | 2304 (2144, 2144, 2122) 1998 |
| ehc | eh3 | EHR-1 | OP | Vd, Pe, Cm | | Ah | | 7 | 48563 | 3030±40 | −24.69 | 3355 (3242, 3226, 3213) 3078 |
| chm | cheam 2 | Cheam-5a.6 | OP | Cn, Cm, Pe | Hh | | | 6 | 61990 | 200±40 | −24.10 | 308 (282, 168, 155) 1 |
| chm | cheam 4 | Cheam-5a.8 | OP | Cn, Cm, Pe | Hh | | | 8 | 63438 | 1170±40 | −25.00 | 1221 (1063) 970 |
| chm | cheam 7 | Cheam-5a.10 | OP | Cn, Cm, Pe | Hh | | | 10 | 63440 | 1540±50 | −25.00 | 1535 (1412) 1311 |
| chm | cheam 3 | Cheam-5a.12 | OP | Cn, Cm, Pe | Hh | | | 12 | 61991 | 1220±40 | −22.10 | 1263 (1171) 1011 |
| chm | cheam 6 | Cheam-5a.17 | OP | Cn, Cm, Pe | Hh | | | 17 | 63439 | 4320±40 | −25.00 | 5025 (4860) 4830 |
| chm | cheam 5 | Cheam-5a.22 | OP | Cn, Cm, Pe | Hh | | MZ, 25 cm | 22 | 61992 | 4180±40 | −23.40 | 4834 (4815, 4754, 4730, 4718, 4711, 4666, 4659) 4550 |

[a] LSB, late-snowmelt basin; OP, open parkland; HF, hillslope forest.

[b] Cn, *Carex nigricans*; Vd, *Vaccinium deliciosum*; Pe, *Phyllodoce empetriformis*; Cm, *Cassiope mertensiana*; Va, *Vaccinium alaskense*; Aa, *Abies amabilis*; Yc, *Chamaecyparis nootkatensis*; Tm, *Tsuga mertensiana*.

[c] Humus form classification.

[d] BR, Bridge River tephra layer (2388 calendar years BP); MZ, Mazama tephra layer (7585 calendar years BP).

[e] Intercept ages within parentheses along with upper and lower age range are based on Stuiver et al. (1998).

River and Mazama tephra layers are located at 60 and 155–162 cm depth, respectively. Very thin 2-mm layers of silty clay are found at 36, 52, 89, 121, 122, 142, 143, and 226 cm depth. A hidden tephra layer, which erupted from Glacier Peak during the mid-Holocene, was revealed at 127 cm depth (Hallett et al. 2001). A 1-cm layer of silt and sand was located at 221 cm depth and a sharp transition from brown gyttja to greyish brown clay occurred at 238 cm depth. Grey clay was noted from 241 to 265 cm depth, and below 265 cm depth, a sharp transition to sand occurred until the base at 300 cm depth (Fig. 6).

The Mount Barr Cirque Lake core consists of dark-brown gyttja for the entire length of the core with a discrete 2-mm layer of sand found at 184 cm depth. The Bridge River tephra was located at 140 cm depth and remnants of the Mazama tephra were found in the base of the core tube upon extraction (Fig. 7). The same mid-Holocene Glacier Peak tephra layer was located at 239 cm depth in the Mount Barr Cirque Lake core (Hallett et al. 2001).

## Magnetic susceptibility

Magnetic susceptibility measurements detect allochthonous minerogenic input in the sediments and provide a proxy measure of erosion resulting from a fire or flood event or the deposition of a volcanic tephra (Thompson and Oldfield 1986). The magnetic susceptibility concentration data (emu·cm$^{-3}$) for Frozen Lake and Mount Barr Cirque Lake (Hallett et al. 2001) were transformed to electromagnetic accumulation rates (EMAR, emu·cm$^{-2}$·year$^{-1}$) in a similar way to the charcoal data (Figs. 6 and 7). EMAR records show infrequent peak events and a gradual decreasing trend over the length of each record, suggesting that erosion decreased with continued vegetation development and humus formation during the Holocene. For Frozen Lake, EMAR and CHAR are negatively correlated ($r = -0.29$, $p < 0.0005$) for the last 11 500 calendar years BP, suggesting that fires are not directly related to erosion events. In contrast, at Mount Barr Cirque Lake, EMAR and CHAR are positively correlated ($r = 0.24$, $p < 0.0005$), suggesting that fires are related to erosion events in the last 7500 calendar years BP. Steep avalanche slopes on the south side of Mount Barr Cirque Lake may be the cause of this relationship. Both EMAR records indicate increased erosion after the deposition of the Bridge River tephra when the local soil charcoal records indicate numerous local fires between 2400 and 1300 calendar years BP (Fig. 2). At Mount Barr Cirque Lake, a similar increase in erosion is seen from 4300 to 3600 calendar years BP.

## CHAR and fire frequency reconstruction

CHAR from Frozen Lake has background values ranging from 0.1 to 0.3 particle·cm$^{-2}$·year$^{-1}$ throughout most of the record with high values of 0.4–0.6 particle·cm$^{-2}$·year$^{-1}$ in the late Holocene. Peak amplitudes vary along with background values. Some CHAR peaks tend to extend over many centuries; however, we define a fire event at the initial increase in CHAR beyond the threshold ratio. Local soil charcoal ages are assumed to be a maximum age for a similarly aged fire recorded in the lake sediment (Gavin 2001). At Frozen Lake, we identified 18 of 40 fires (45%) as local events in the watershed over the last 9000 years using 49 soil charcoal AMS

**Table 3.** Calibrated and uncalibrated AMS $^{14}$C ages used in the lake sediment cores from Frozen Lake and Mount Barr Cirque Lake.

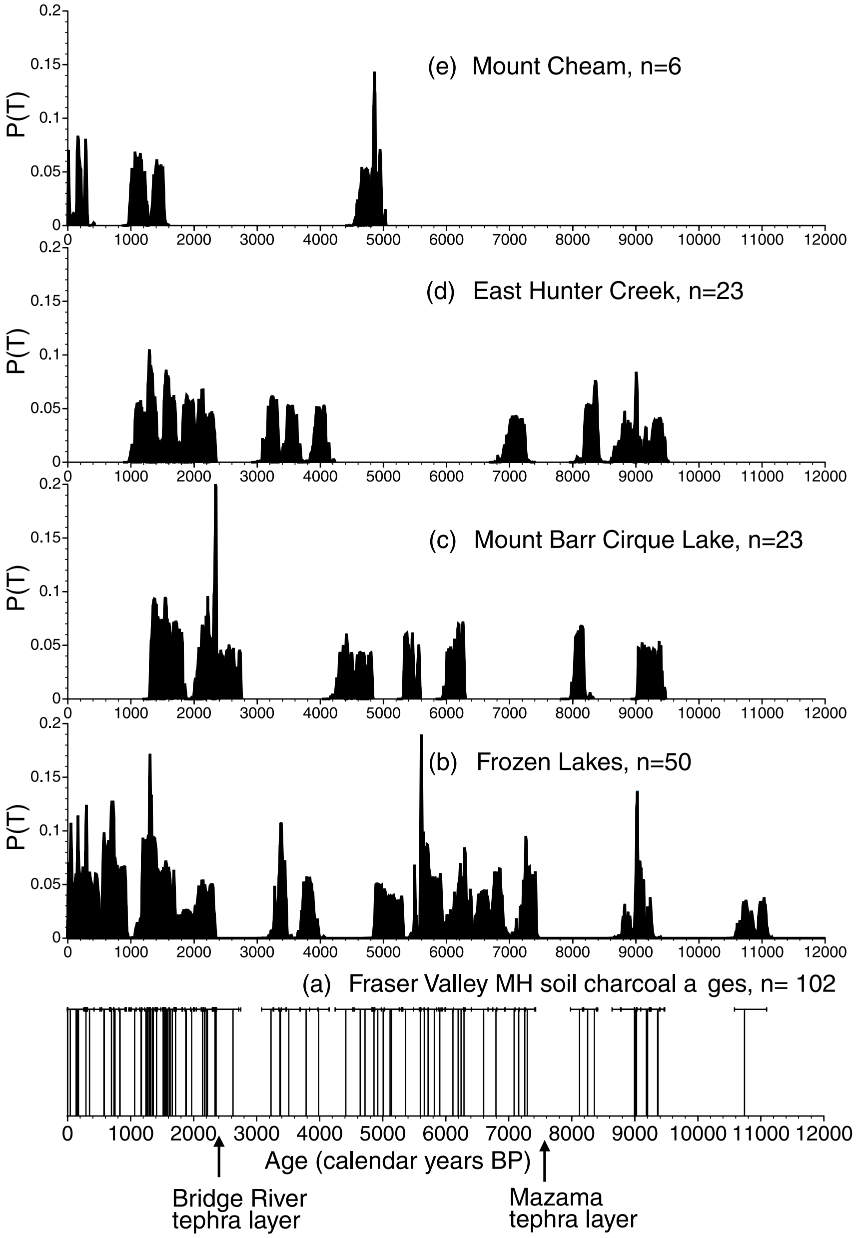
| Site | Depth (cm) | CAMS laboratory No. | Uncalibrated $^{14}$C age ($^{14}$C years BP) | $\delta^{13}$C | Calibrated age with 2-sigma age range (calendar years BP)[a] | Material dated | Reference |
|---|---|---|---|---|---|---|---|
| Frozen Lake | 25 | 45980 | 1570±60 | –26.40 | 1593 (1508, 1502, 1482,[b] 1468, 1420) 1314 | Conifer needle *Abies amabilis* | |
| | 35 | 45981 | 1950±40 | –25.00 | 1990 (1916, 1914,[b] 1889) 1820 | Conifer needle *Tsuga mertensiana* | |
| | 52 | 45982 | 2470±70[c] | –25.00 | 2749 (2706, 2644, 2490) 2346 | Conifer needle *Tsuga mertensiana* | |
| | 60 | | 2435±26 | –25.00 | 2710 (2465, 2411, 2400, 2388,[b] 2378, 2369, 2363) 2353 | Bridge River tephra | Clague et al. 1995 |
| | 89 | 45983 | 3560±40 | –25.00 | 3975 (3835[b]) 3703 | Conifer needle *Tsuga mertensiana* | |
| | 119 | 45984 | 4530±50 | –24.56 | 5435 (5289, 5155, 5150,[b] 5097, 5091) 4981 | Conifer needle *Tsuga mertensiana* | |
| | 147 | 45985 | 6170±40 | –25.91 | 7226 (7153, 7128, 7086,[b] 7080, 7022) 6911 | Twig fragment | |
| | 155–163 | | 6730±40 | –25.00 | 7671 (7585[b]) 7509 | Mazama tephra | Hallett et al. 1997 |
| | 188 | 45986 | 8180±50 | –25.62 | 9396 (9228, 9223, 9127,[b] 9101, 9090) 9010 | Conifer needle *Abies amabilis* | |
| | 221 | 45987 | 9390±70 | –25.69 | 11 055 (10 636, 10 613,[b] 10 580) 10 405 | Twig fragment | |
| | 235 | 45988 | 10020±50 | –25.58 | 11 937 (11 548, 11 505, 11 479, 11 473, 11 413,[b] 11 388, 11 354, 11 352, 11 343) 11 256 | Conifer needle *Abies lasiocarpa* | |
| Mount Barr Cirque Lake | 11 | 57052 | 210±40 | –25.00 | 310 (285, 166,[b] 161) 2 | Conifer needle *Tsuga mertensiana* | |
| | 44 | 53978 | 1540±50 | –25.00 | 1535 (1412[b]) 1311 | Conifer needle *Abies amabilis* | |
| | 73 | 53979 | 1620±50 | –25.00 | 1688 (1525[b]) 1393 | Conifer needle *Abies amabilis* | |
| | 99 | 53980 | 1970±50 | –23.88 | 2037 (1922, 1909,[b] 1900) 1820 | Conifer needle *Tsuga mertensiana* | |
| | 121 | 53981 | 2220±50 | –25.00 | 2345 (2303, 2240, 2205, 2200,[b] 2181, 2166, 2162) 2073 | Conifer needle *Tsuga mertensiana* | |
| | 142 | | 2435±26 | –25.00 | 2710 (2465, 2411, 2400, 2388,[b] 2378, 2369, 2363) 2353 | Bridge River tephra | Clague et al. 1995 |
| | 164 | 53982 | 3420±40 | –22.86 | 3825 (3686, 3660,[b] 3642) 3570 | Conifer needle *Tsuga mertensiana* | |
| | 196 | 53983 | 3890±40 | –25.01 | 4420 (4350, 4327,[b] 4299) 4154 | Conifer needle *Abies amabilis* | |
| | 247 | 53984 | 5310±50 | –25.00 | 6271 (6168, 6147, 6108, 6097, 6092, 6073,[b] 6056, 6049, 6016, 6013, 6000) 5933 | Conifer needle *Abies amabilis* | |
| | 289 | 53985 | 6480±50 | –25.00 | 7463 (7422[b]) 7272 | Conifer needle *Tsuga mertensiana* | |

[a] Intercept ages within parentheses along with upper and lower age range are based on Stuiver et al. (1998).

[b] Calibrated ages used in age versus depth determination.

[c] Rejected age due to inversion when compared with the age and position of the Bridge River tephra layer.

**Fig. 2.** (*a*) Composite soil charcoal AMS ages showing all fires (midintercept ages are represented by bars with their with 2-sigma age range) for the Fraser Valley mountain hemlock sites. Total cumulative probability distributions (2-sigma range) for all the soil charcoal ages are shown for each site and represent fires at (*b*) Frozen Lake, (*c*) Mount Barr Cirque Lake, (*d*) East Hunter Creek, and (*e*) Mount Cheam. Distinct gaps with no charcoal ages and low probabilities are contrasted by periods of higher probability. This indicates a periodic fire regime and, at times, long fire-free intervals in mountain hemlock forest.



ages (Fig. 6). The period from 11 500 to 9000 calendar years BP had one more AMS age dated at 10 740 calendar years BP, but our sampling bias against charcoal in mineral soil limits our interpretation of local fire older than 9000 calendar years BP.

The CHAR-based fire frequency reconstruction for Frozen Lake shows a continuously varying fire frequency over the Holocene (Fig. 6). The highest fire frequencies of 7 or 8 events/1000 years occur during the early Holocene from 11 000 to 9000 calendar years BP. After 9000 calendar years BP, CHAR peak frequency declines rapidly and reaches its minimum value of 2 events/1000 years around 6000 calendar years BP. Two high fire frequency periods occur from 5000 to 3500 and from 2400 to 1300 calendar years BP. There is a sharp decline in fire frequency from 1300 calendar years BP to present day. The mean fire return interval (MFI) for peak (or fire) events in the 11 400 year Frozen Lake record is 204 ± 12 years (mean ± 1 SE). The MFI for local peak (or fire) events calibrated by soil charcoal ages is 555 ± 100 years.

CHAR from Mount Barr Cirque Lake has a fairly constant background of 0.1–0.3 particle·cm$^{-2}$·year$^{-1}$ throughout most of the record until 2400–1300 calendar years BP when background values rise an order of magnitude from 0.8 to

**Fig. 3.** Location of soil charcoal ages (midintercept age only, calendar years BP) for the Frozen Lake watershed. Most ages are located above Frozen Lake, which lies downstream on a subalpine bench system that is contained by steep slopes on the west and east sides. Distinct charcoal deposits, with known stratigraphic markers such as volcanic tephra, were most common in the late-snowmelt depressions of areas with open aspect. This figure represents a crude time-since-fire map for local fire in the watershed. Contour intervals are in 20-m increments.

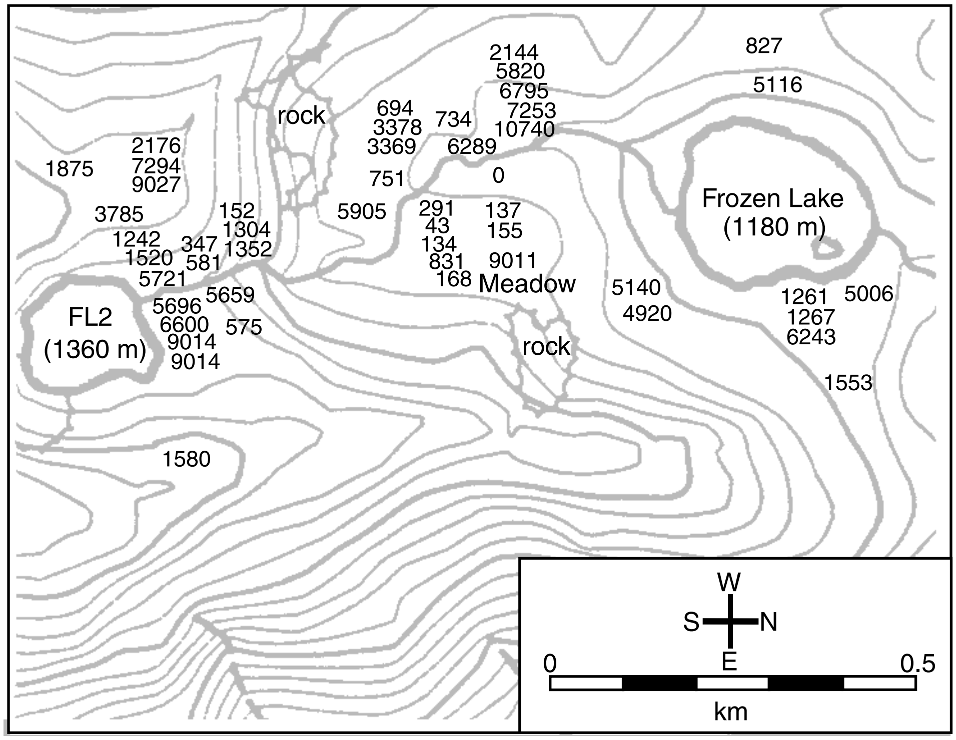


**Fig. 4.** Location of soil charcoal ages (midintercept age only, calendar years BP) for the Mount Barr Cirque Lake watershed. Most ages are located on south-facing slopes below continuous forest on this time-since-fire map for local fire. Sampling for charcoal was done all around the lake, and the open fuel-limited areas on the north and northeast facing slopes had the least amount of soil charcoal and few radiocarbon ages. Contour intervals are in 20-m increments.

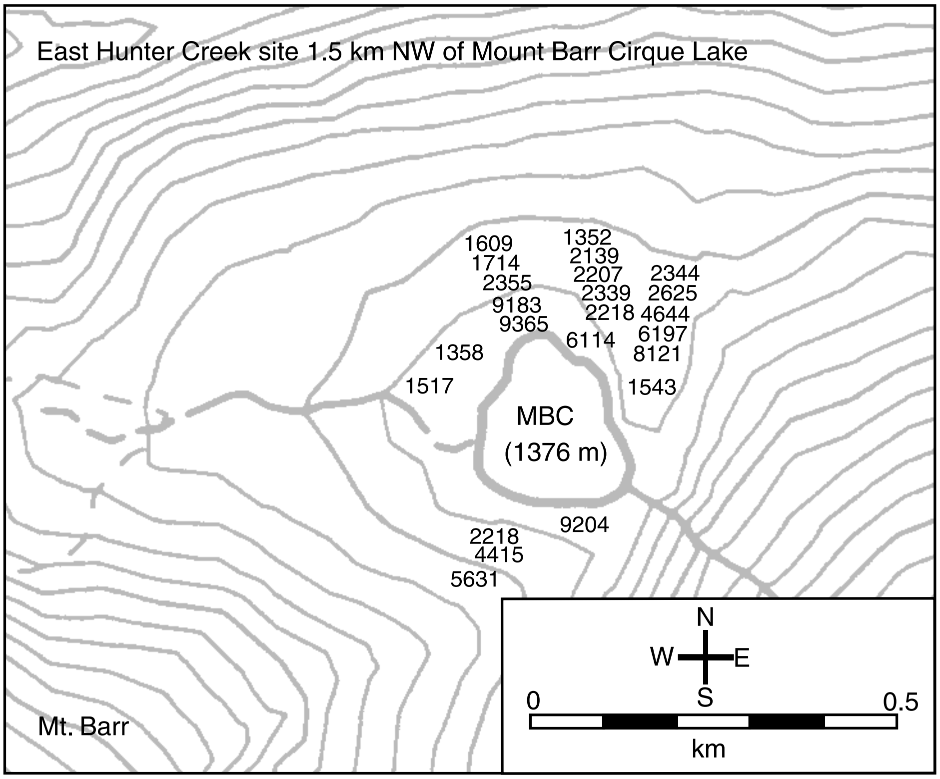
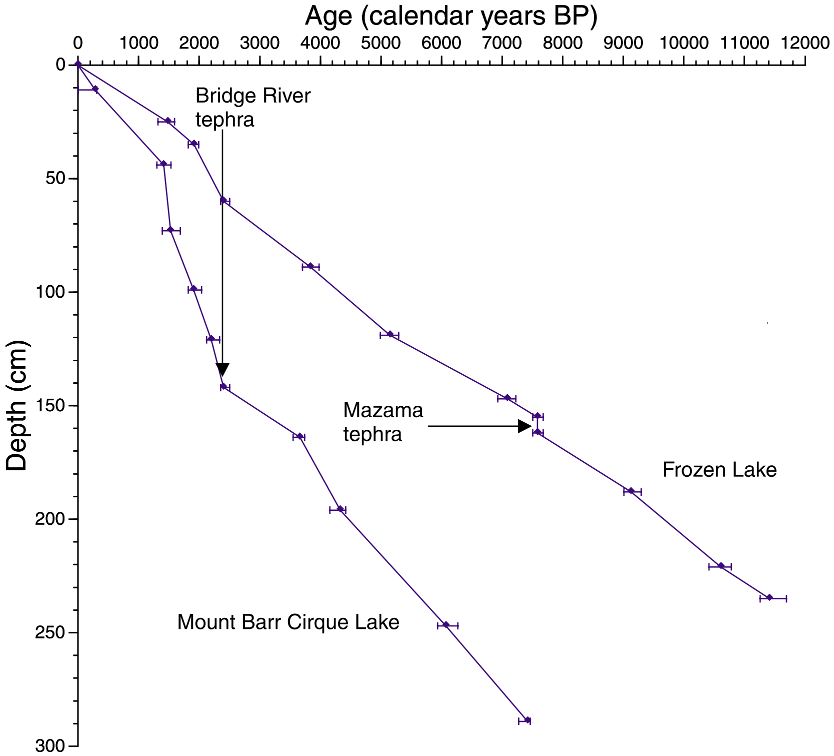


**Fig. 5.** Age versus depth relationship for Frozen Lake and Mount Barr Cirque Lake sediment cores. Calibrated AMS radiocarbon ages from Table 3 are plotted along with their 2-sigma age range. A simple linear interpolation between ages was used to construct an age–depth model.



4.0 particles·cm$^{-2}$·year$^{-1}$. Peak values also increase dramatically during this interval and require a log scale. In the Mount Barr Cirque Lake watershed, 11 of 46 fires (24%) in the last 7500 years were identified as local events using 16 of 23 AMS ages (Fig. 7). Frequent fire periods were identified between 6500 and 3500 and between 2400 and 1300 calendar years BP. Low fire frequency occurs around 7000, between 3500 and 2400, and after 1000 calendar years BP according to the CHAR record. Evidence of synchronous fire occurs in the neighbouring East Hunter Creek watershed from 2400 to 1000 calendar years BP. Individual fires at East Hunter Creek around 2000 calendar years BP, during the mid-Holocene, and at 7000 calendar years BP may have contributed airborne charcoal to the Mount Barr Cirque Lake basin (Fig. 7). Overall MFI for the 7500-year Mount Barr Cirque Lake record is 161 ± 12 years. The MFI for local peak (or fire) events calibrated by soil charcoal ages is 455 ± 100 years.

We separated the two MH fire frequency reconstructions into six zones with transitions at ~8800, 6400, 3500, 2400, and 1300 calendar years BP, with synchronous response over the last ~6000 years (Fig. 8). A variable fire frequency at both sites during the mid-Holocene becomes locked in phase after 4500 calendar years BP and continues to present day. A decrease occurs from 3500 to 2400 calendar years BP followed by increases to early Holocene levels of 6–8 events/1000 years from 2400 to 1300 calendar years BP and a decline after 1300 calendar years BP. MFI estimates for individual zones 1–6 complement the fire frequency plot but are best used in a comparison with one another rather than their actual value (Fig. 8) (Long et al. 1998; Hallett and Walker 2000).

## Discussion

### Reconstructing fire frequency with CHAR records and soil charcoal ages

This study is a comparative analysis of local fire history that combines lake sediment and soil charcoal methods over the entire Holocene. These methods provide independent evidence of fire around a site (Gavin 2000) and allow us to assess variations in fire frequency under changing climate. Most charcoal-based fire history studies use dendrochronological evidence, such as fire scars, and stand ages or historical fire records to calibrate CHAR peaks with a fire of known age, size, and proximity to the lake (Swain 1973, 1978; Cwynar 1978; Clark 1990; MacDonald et al. 1991; Millspaugh and Whitlock 1995; Larsen and MacDonald 1998; Long et al. 1998; Tinner et al. 1998; Gavin 2000; Hallett and Walker 2000). This calibration method is limited by short time scales and the “erasure effect”. Theoretical (Patterson et al. 1987; Clark 1988) and empirical studies of charcoal deposition in lakes after a fire event (Whitlock and Millspaugh 1996; Whitlock et al. 1997; Clark et al. 1998; Gardner and Whitlock 2001) support the interpretation of CHAR peaks as past fires. Addition of AMS-dated soil charcoal allows calibration of CHAR peaks as far back as the early Holocene. With comprehensive sampling of soil charcoal in a watershed and a generous radiocarbon dating budget, detailed time-since-fire maps can be constructed to analyze the spatial and temporal variations in fire regimes (Gavin 2000) and the impacts of climate or humans on fire regimes (Meyer et al. 1992; Carcaillet 1998). Not all soil charcoal sites are suitable for detailed fire history reconstruction because frequent freezing in winters may cause

**Fig. 6.** Charcoal accumulation rates (CHAR), peak events, soil charcoal data, fire frequency reconstruction, magnetic susceptibility, and lithology for Frozen Lake in the last 12 000 calendar years BP. Peak events corresponding to local fire evidence from Fig. 3 are marked with an arrow (fire directly around Frozen Lake), M (fire in meadow), and FL2 (fire at upper Frozen Lake). Soil charcoal dates are used as a maximum age to calibrate a local CHAR peak event.

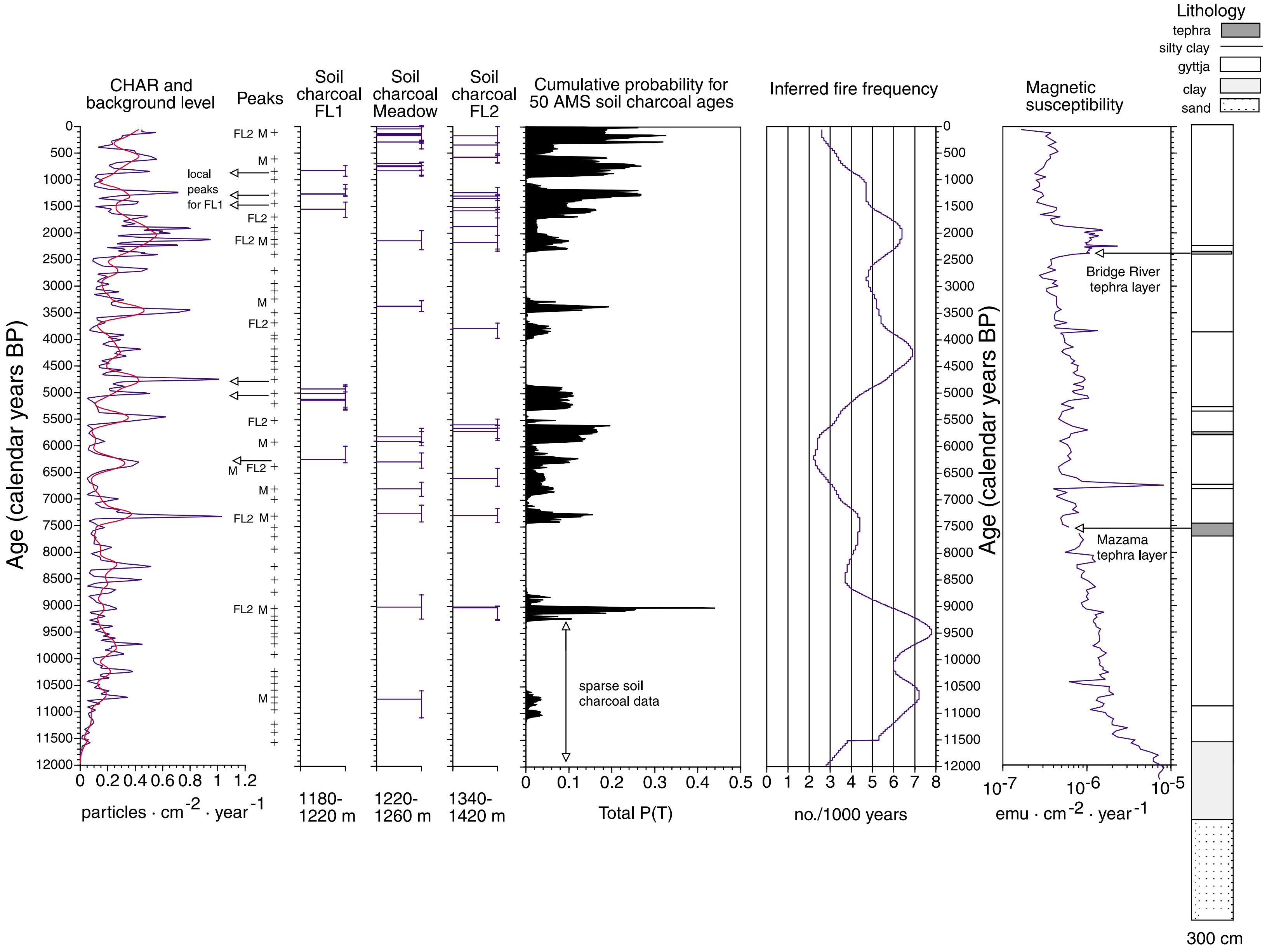


**Fig. 7.** Charcoal accumulation rates (CHAR), peak events, soil charcoal data, fire frequency reconstruction, magnetic susceptibility, and lithology for Mount Barr Cirque Lake in the last 7500 calendar years BP. Peak events corresponding to local fire evidence from Fig. 4 are marked with an arrow. The total cumulative probability of soil charcoal ages from neighbouring East Hunter Creek (EHC) are shown and peak events that may correspond to distant fires (>1.5 km) outside the watershed are marked with EHC.

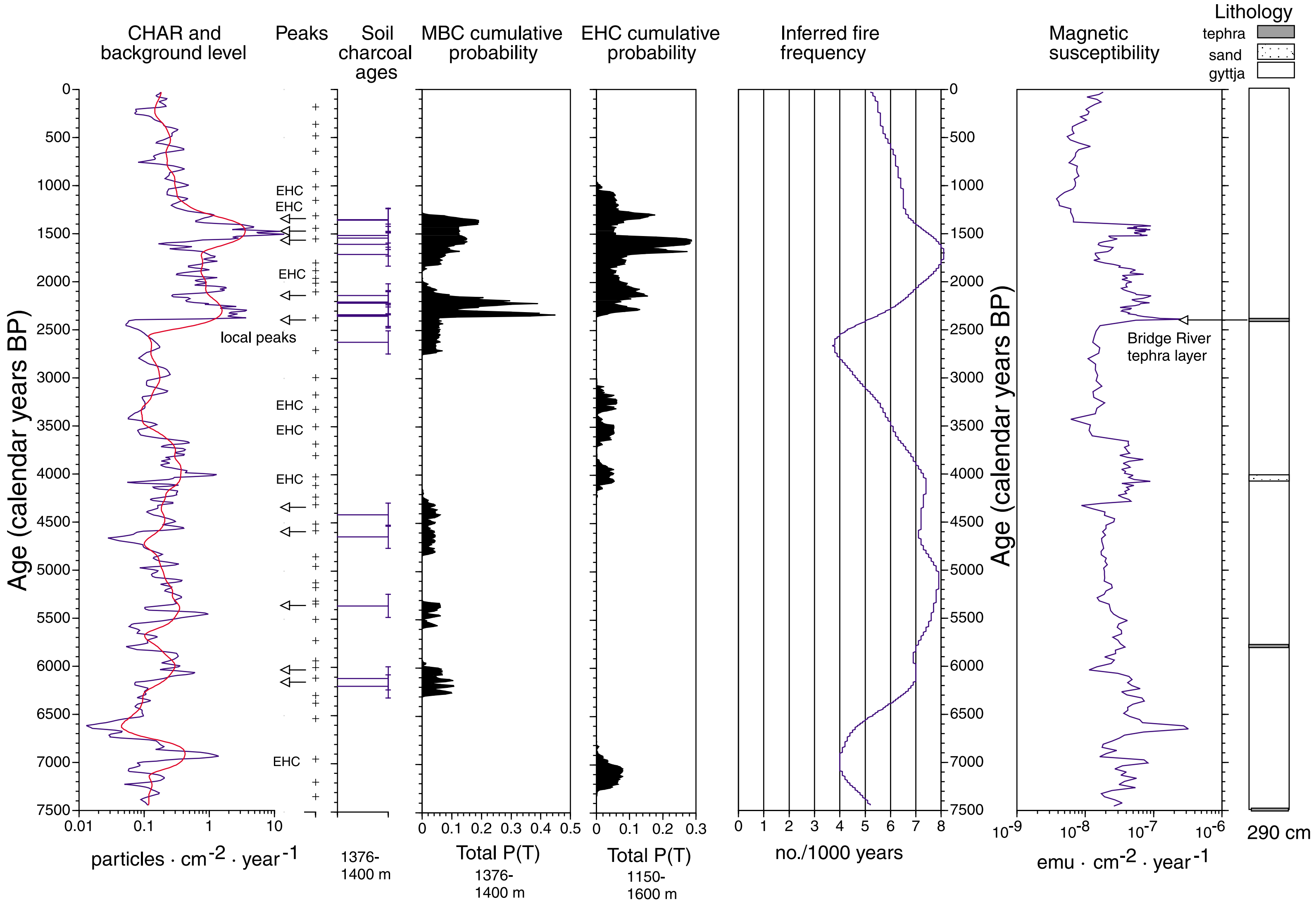
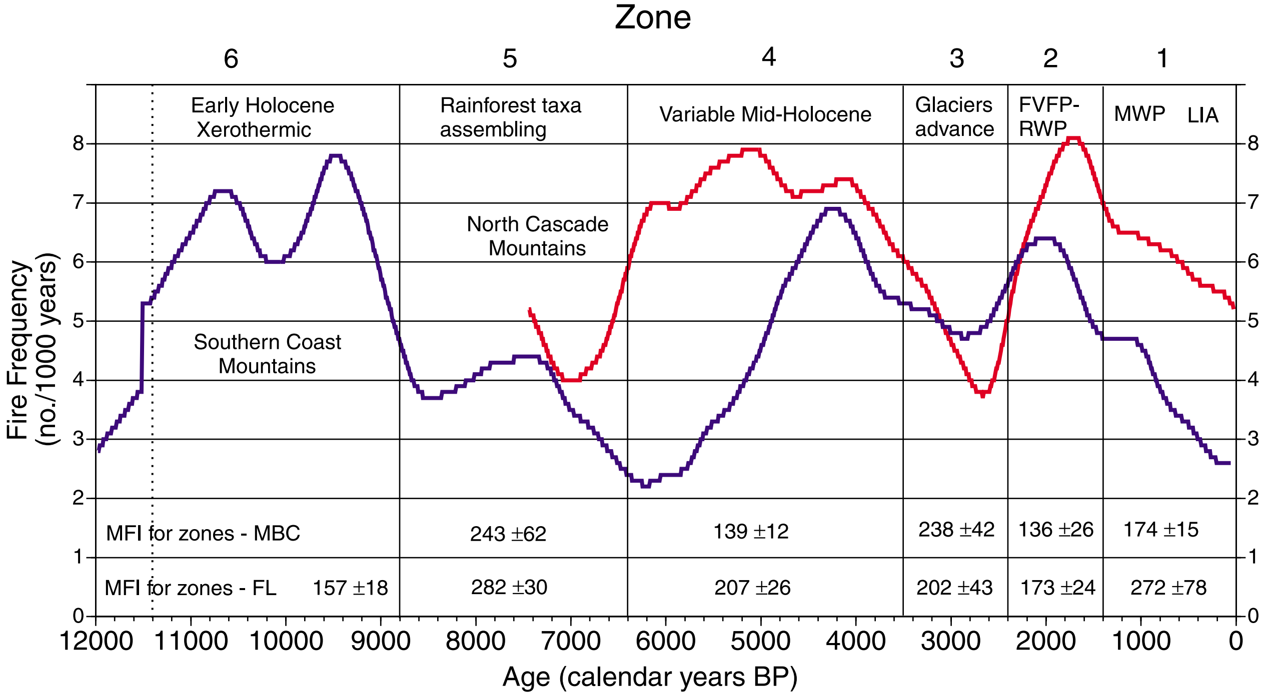


**Fig. 8.** Summary diagram for fire frequency changes in mountain hemlock forests of the southern Coast Mountains (lower line) and North Cascade Mountains (upper line) throughout the Holocene. Mean fire return intervals (MFI) along with their standard errors (±1 SE) are shown for each of zones 1–6 in the bottom portion of the diagram. Zones 1–6 are labelled with a short description of climate, vegetation, or environmental history that characterizes each zone. For zone 2, Fraser Valley Fire Period – Roman Warm Period (FVFP–RWP), and for zone 1, Medieval Warm Period (MWP) and Little Ice Age (LIA), are used to denote recent climate periods.



cryoturbation, and bioturbation during warm summers may disturb the stratigraphy and size fraction of charcoal deposits (Carcaillet and Talon 1996; Carcaillet 2001).

CHAR peaks lacking evidence of local fire likely represent burns outside the watershed or upwind of the lake (Gardner and Whitlock 2001), although most macroscopic charcoal does not travel far from its source (Clark et al. 1998; Ohlson and Tryterud 2000). Our results at Mount Barr Cirque Lake suggest that CHAR peaks represent fires near the lake but charcoal from adjacent watersheds such as East Hunter Creek may travel in plumes as much as 1.5 km. Wind is clearly a critical factor affecting charcoal distribution in high-elevation areas (Gardner and Whitlock 2001; Benedict 2002). Small CHAR peaks may represent some large but distant (>1 km) fires or small local events (<1 km). Most large peaks, however, are usually associated with nearby fires within the watershed (Millspaugh and Whitlock 1995). Magnetic susceptibility measurements helped to identify some local fires with erosion at Mount Barr Cirque Lake, but this was not the case at Frozen Lake were magnetic susceptibility and CHAR were negatively correlated. The deeper and larger Frozen Lake had a lower sedimentation rate, and erosion of minerogenic material delivered from gentler slopes may not have reached the deepest parts of the basin as at Mount Barr Cirque Lake. Mount Barr Cirque Lake with its steep avalanche slopes next to the lake probably has minerogenic material delivered with higher energy, allowing it to reach the deepest portions of the basin. In summary, we feel that soil charcoal is a better indicator of local fire around a lake basin.

The CHAR-based fire frequency reconstruction revealed shorter than expected MFI values of 204 and 161 ± 12 years for wet mountain hemlock forests. There are periods, such as fire frequency zones 5, 3, and 1, when MFI increases slightly (Fig. 8), but in general, we expected much longer intervals for this forest type. The local MFI values (fires with soil charcoal evidence) of 555 and 455 ± 100 years are more consistent with modern fire history and ecological studies that indicate MFI values of 600–1500 years for mountain hemlock forests (Lertzman 1992; Agee 1993). CHAR-based MFI estimates for drier montane spruce forests in southeastern British Columbia (~240 years) (Hallett and Walker 2000) and lower elevation western hemlock forests in the Oregon Coast Range (~230 years) (Long et al. 1998) were similar over the last 3000 years. In contrast, estimated fire intervals of 1200 and 2380 years for mountain and western hemlock rain forests based solely on soil charcoal AMS ages suggest the rarity of fire as local disturbance process (Lertzman et al. 2002). The soil charcoal method alone, however, may underestimate fire intervals because of taphonomic problems, radiocarbon dating limitations, or the fact that severe fires can consume earlier charcoal deposits (Gavin 2000). The relatively short MFI estimates suggest that the CHAR peak method may overestimate the number of local (in our case, this means within 1 km of the lake) fire events because we identified only 24–45% of peaks as local fires. More extensive sampling of soil and more AMS dates may resolve a greater number of fires. CHAR represents a complex spatial aggregation of fire events occurring across a charcoal source area that is specific for each site. Our sites had small ridgetop watersheds, but fires burning at lower elevations or in neighbouring watersheds likely contributed charcoal to the lakes.

The ~200-year MFI estimates are very similar to significant ~210- and 218-year periodicities in CHAR records from western hemlock and montane spruce forests over the last 1000 years and from the Frozen Lake record over the last 11 000 years. Shorter than expected MFIs for mountain hemlock forest and a similarity to values from drier forest types suggest that large-scale climate forcing may be causing synchronous fire events at century-scale frequencies (Hallett et al., in revision[3]).

### Comparison of mountain hemlock charcoal records and Holocene climate

Fire frequency around both Frozen and Mount Barr Cirque lakes has varied with climate throughout the Holocene. Fires were often more frequent at Mount Barr Cirque Lake, which may be explained by higher resolution sampling and a higher rate of sedimentation (Fig. 1) or its proximity to drier forest types on the east slopes of the Cascades where fires are more common (Agee et al. 1990; Agee 1993). In fire frequency zone 1, a range of 6–8 events/1000 years at Frozen Lake occurs during dry and warm early Holocene climate (Mathewes and Heusser 1981; Mathewes 1985; Clague and Mathewes 1989; Hebda and Whitlock 1997) and corroborates other charcoal studies in western North America (Long et al. 1998; Hallett and Walker 2000; Millspaugh et al. 2000; Mohr et al. 2000). During the early Holocene, the eastern Pacific High expanded due to greater than present summer insolation. This created increased summer temperatures and decreased effective precipitation across much of western North America (Thompson et al. 1993), promoting more frequent fires across the region. Infrequent CHAR peaks at Frozen Lake from 8200 to 7400 calendar years BP match well with a gap in soil charcoal evidence (Fig. 6) and may be a response to global cooling after the widespread 8200-year BP cold event (Stager and Mayewski 1997). Fires are also absent during this cold period around Dog Lake in the Rocky Mountains (Hallett and Walker 2000).

Fires decreased in zone 5 around both lakes as rain forest became established in the region (Mathewes 1973; Mathewes and Rouse 1975; Hebda and Mathewes 1984; Wainman and Mathewes 1987; Whitlock 1992; Sea and Whitlock 1995; Worona and Whitlock 1995; Hebda and Whitlock 1997; Pellatt and Mathewes 1997; Pellatt et al. 2000; Gavin et al. 2001). A few needle fragments of *A. amabilis* occur in the early Holocene, but drier subalpine taxa, such as *A. lasiocarpa*, tend to dominate the macrofossil record (in association with *T. mertensiana*) from the lakes. After 8500 calendar years BP, *A. amabilis* needles increase and *A. lasiocarpa* are absent after 6000 calendar years BP (K. Farquharson and D. Hallett, unpublished data). Pollen, plant macrofossil, and charcoal evidence from similar subalpine forests on the Olympic Peninsula, Washington (Gavin et al. 2001), and drier forest on nearby Mount Stoyama, British Columbia (Pellatt et al. 1998, 2000), support this interpretation. Soil charcoal evidence is sparse from 9000 to 7400 calendar years BP, suggesting that climate was moist enough to limit local fire at our sites (Fig. 2). Frozen Lake CHAR indicates a lower fire frequency; however, several peaks in the mid-Holocene appear to lag over many centuries, suggesting that secondary charcoal deposition may have occurred for long periods after a fire event (Whitlock and Millspaugh 1996; Whitlock et al. 1997).

The initial overlapping portions of the mountain hemlock records show a variable fire period in zone 4 with a broad range of 2–8 events/1000 years (Fig. 8). Local fire evidence is found around Frozen Lake and Mount Barr Cirque Lake (Figs. 6 and 7) during this period, but East Hunter Creek was fire free for millennia. There is little coherence between the records until after 4500 calendar years BP when fire frequency becomes locked in phase for the remainder of the Holocene. A synchronous decrease in fire frequency from 3500 to 2400 calendar years BP corresponds to Neoglacial advances in the region (Porter and Denton 1967) and changes in atmospheric circulation of the Northern Hemisphere (O'Brien et al. 1995). The Tiedemann advance in the southern Coast Mountains of British Columbia (Ryder and Thomson 1986) as well as the Peyto (Luckman et al. 1993) and Stutfield Glacier (Osborn et al. 2001) advances in the Canadian Rockies date to this interval. Glacier advances and infrequent fire events support a cooler moister climate regime most likely dominated by the Aleutian Low and frequent westerly storm tracks. A period of decreased fire frequency during the Peyto advance also occurs in the montane spruce forests surrounding Dog Lake in southeastern British Columbia (Hallett and Walker 2000).

A return to more frequent fire between 2400 and 1300 calendar years BP suggests that prolonged drought occurred more often during this interval, which we name the Fraser Valley Fire Period. This frequent fire period is coeval with the Roman Warm Period in North Atlantic records (Lamb 1995) when climate forcing such as enhanced Azores high pressure circulation, increased North Atlantic flow and ocean circulation (Bianchi and McCave 1999), and more active sun occurred (Stuiver et al. 1997, 1998; Bond et al. 2002). Records of longer and more frequent drought episodes in North America during the last 2000 years also correspond to the timing of the Fraser Valley Fire Period (Laird et al. 1996; Woodhouse and Overpeck 1998). Some pollen studies in the interior west of North America indicate increased aridity during this period (Thompson et al. 1993), but many studies in the Pacific coast region report modern climate conditions for the last 3000 years (Mathewes 1973, 1985; Mathewes and Rouse 1975; Mathewes and Heusser 1981; Barnosky et al. 1987; Mathewes and King 1989; Whitlock 1992; Thompson et al. 1993; Hebda 1995; Sea and Whitlock 1995; Hebda and Whitlock 1997; Pellatt and Mathewes 1997; Pellatt et al. 1998, 2000, 2001; Gavin and Brubaker 1999; Gavin et al. 2001). The relatively coarse sampling used in most pollen studies limits the ability to resolve decade- to century-scale climate changes such as drought events. Long-lived trees in the rainforest have eco-

---

[3] D.J. Hallett, D.S. Lepofsky, D.G. Gavin, R.W. Mathewes, and K.P. Lertzman. Century-scale solar forcing of drought and forest fire in southern British Columbia. Clim. Change. In revision.

logical inertia that allows them to survive through short-term climate changes (Agee 1993). Pollen data from these sites may or may not show a response to frequent and (or) severe summer drought. Atmospheric driven disturbances, such as fire, are more likely to track climate in the late Holocene. There is growing evidence of drier summer climate in the late Holocene because several charcoal records show increased fire activity during 2400–1300 calendar years BP in western North America (Wainman and Mathewes 1987; Long et al. 1998; Reasoner and Huber 1999; Hallett and Walker 2000; Mohr et al. 2000; Pellatt et al. 2000; Gavin et al. 2001). Forest fires in temperate rainforest indicate summer drought and can help to elucidate rapid climate change on the scale of decades to centuries. Enhanced high pressure circulation may have dried fuels and led to an increase in dry lightning storm tracks across the Pacific Northwest during the Fraser Valley Fire Period (Rorig and Ferguson 1999). Lightning is rare in coastal temperate rainforest but increases as you move into the mid- to upper Fraser Valley, which is closer to drier, more continental forests on the east side of the Cascade and Coast Mountains (Lertzman et al. 2002).

More frequent drought and dry fuel conditions during the Fraser Valley Fire Period may have increased the probability of human-lit fires. Modern studies suggest that human-lit fires may have been important in the lower parts of the valley and on adjacent mountainsides in enhancing berry crops and clearing vegetation for hunting (Turner 1991; Lepofsky et al. 2003). An active cultural period known as the Marpole in the Fraser Valley coincides roughly with the Fraser Valley Fire Period, although very little archaeological evidence exists for elevations in the MH zone (Matson and Coupland 1995). Human-lit fires are documented as frequent (yearly to decadal) events and are thought to occur more or less continuously from the historic period into the past, but this is not the case at our MH sites (Lepofsky et al. 2003). Fires tend to occur every two centuries or more at our sites throughout the Holocene and appear to be a background signal of natural fire (Fig. 8). Our charcoal data suggest that climate forcing (enhanced high pressure circulation leading to drier fuels and more lightning) is the most parsimonious explanation for increased fire at subalpine elevations where fuels are often wet even in summer. There may have been periods where weather and fuels were conducive for fire but no ignition sources, either lightning or human, existed. Long fire-free intervals during Neoglacial advances also support a climate-based explanation for fire variability. To better address the variability of fire and climate in the region, a comprehensive analysis of charcoal records is needed to increase our knowledge of the synchrony of fire response across different forest types and climate zones. Forest fire and its importance to paleoclimatology may help us to better understand the variability of late Holocene climate and future changes in response to global warming (Overpeck et al. 1990; Price and Rind 1994).

Modern fire frequencies of 2–5 events/1000 years were established after 1300 calendar years BP. With the exception of Frozen Lake, and to a lesser extent Mount Cheam, there is little evidence of local fire at our MH sites over the last 1000 years, although CHAR peaks indicate burning during the Medieval Warm Period (~1000–600 calendar years BP) around our subalpine lake sites (Figs. 2, 6, and 7). Fires appear to be rare during the cooler climate of the Little Ice Age (600–150 calendar years BP) (Luckman et al. 1993; Lamb 1995; Clague and Mathewes 1996) at various sites across the province (Gavin 2000; Hallett and Walker 2000; Hallett 2001). The last fires recorded at our MH sites correspond to a dry period around 150 calendar years BP (~AD 1800) (Watson and Luckman 2001) when large fires occurred at many sites in the northwestern United States (Hemstrom and Franklin 1982; Dickman and Cook 1989; Agee 1991, 1993; Huff 1995; Heyerdahl et al. 2001) and western Canada (Johnson and Fryer 1987; Masters 1990; Gavin 2000; Hallett 2001).

Modern ecological (Lertzman and Krebs 1991; Lertzman 1992) and fire history studies (Hemstrom and Franklin 1982; Agee and Smith 1984; Agee 1993) use tree-ring records to define the disturbance history of MH forests and the role of fire. From these records, we find that most modern fires are small and patchy, especially in reduced fuels such as subalpine heaths in open parkland (Douglas 1972; Potash and Agee 1998). Larger stand-destroying fires tend to affect continuous forest on drier south-facing slopes (Fahnestock and Agee 1983; Agee and Smith 1984; Agee 1993), which is also where we found most of the soil charcoal deposits at our sites. Our calibrated CHAR records reflect the frequency of stand-destroying fire in mountain hemlock forest, and smaller uncalibrated peaks may represent local patchy fires or large fires from nearby watersheds. These subalpine CHAR records undoubtly contain plume-transported charcoal from a broad source area that may include fires started at lower elevations.

In summary, our long-term charcoal study suggests that fires were more frequent than previously thought in the MH zone of the mid-Fraser Valley. These MH zone sites are located in transitional areas close to drier, more continental forests at lower elevations and just a few kilometres away in the subalpine. MH forest located closer to the coast or farther north should be studied to determine the range of potential MFI values for this forest type. At our sites, a variable fire frequency appears to be in phase with shifts in regional climate and may reflect the influence of inland high pressure cells as well as the strength and position of the Pacific High. The timing and extent of fire events can greatly influence the structure of these high-elevation old-growth forests by creating meadows and open areas in subalpine areas and altering sensitive ecotones near tree lines (Agee and Smith 1984; Rochefort et al. 1994; Rochefort and Peterson 1994; Huff 1995). For example, during the low fire frequency periods from 3500 to 2400 and from 600 to 0 calendar years BP, the structure of MH zone forest probably displayed more old-growth characteristics and greater biomass accumulation. Frequent fire periods such as the Fraser Valley Fire Period and Medieval Warm Period may have caused an increase in younger age-classes and more open areas. The temporal variability of fire regimes on century-to-millennial time scales and its potential impact on forest structure in old-growth subalpine forests may be important for global warming scenarios in western Canada. Future wildfire in circumboreal forests under global warming scenarios show a variable response across regions including expected increases and some decreases in fire frequency (Flannigan et al. 1998). Long-term charcoal records are useful for testing general circulation model and regional model simulations of fire climate and investigating links between forest dynamics and the global carbon cycle (Clark et al. 1997; Flannigan et al. 2001).

### Conclusions

The macroscopic charcoal records from the MH zone provide a unique window on the long-term fire history in the wet subalpine forests of the southern Coast and northern Cascade Mountains. These high-resolution records show synchronous responses to climate change on century-to-millennial time scales and provide insight into the variability of late Holocene climate and fire regimes across the region. Several conclusions follow from this study and should be tested with additional long-term charcoal records from other forest types.

1. AMS dating of soil charcoal deposits around a lake are useful for calibrating CHAR records over the Holocene. This method is most effective in wet forest types with little evidence of past fire, such as fire-scarred trees and stand age information. CHAR records may overestimate local fire frequency and produce shorter MFI values because CHAR represents a complex spatial aggregation of fire events from a charcoal source area. Soil charcoal records may underestimate fire frequency records because they provide site-specific (point) data, and extensive radiocarbon dating may be not be possible.
2. Decomposition of the components of macroscopic charcoal records derived from lake sediments is an effective method for reconstructing the local fire history on century-to-millennial time scales. These methods show a nonstationary fire response around Frozen Lake (last 11 500 years) and Mount Barr Cirque Lake (last 7500 years) over the Holocene. The fire frequency records are in phase from the mid-Holocene, suggesting that fire regimes since ~5000 calendar years BP have been synchronized by climate changes.
3. Fire events were frequent at Frozen Lake in the early Holocene when warm and dry summer conditions existed. The synchronous increases in fire frequency at both lakes between 2400 and 1300 and between 4500 and 3500 calendar years BP indicate more frequent summer drought and a strong relationship with climate forcing via blocking high pressure circulation. Synchronous decreases in fire frequency between 3500 and 2400 and between 600 and 0 calendar years BP in MH forests correspond to Neoglacial advances such as the Tiedemann and Little Ice Age, suggesting dominance of the Aleutian Low and frequent westerly storm tracks.

## Acknowledgements

This work was supported by a Forest Renewal British Columbia grant administered by the Science Council of British Columbia, a Natural Sciences and Engineering Research Council of Canada research grant to R.W.M., graduate fellowships from Simon Fraser University, and Global Forest contribution 18-2000-100. We thank B. Blackwell, B. Brett, D. Brown, J. Fall, L. Fraser, D. Gavin, R. George, E. Heyerdahl, M. Konradi, D. MacKenzie, M. Mayhew, A. McHalsie, S. McCuaig, N. Oakes, M. Plotnikoff, D. Schaepe, L. Tulk, A. Walton, K. Washbrook, and C. Wong for assistance with soil sampling and lake coring. We thank Cathy Whitlock and Pat Bartlein for use of CHAPS software and the magnetic susceptibility meter at the University of Oregon. We thank Cheryl Takahashi and Anne Walton for preparing numerous charcoal samples for radiocarbon dating and the Sto:lo Nation for assisting with research on traditional lands. AMS radiocarbon ages were provided by the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, Calif., and we thank Tom Brown for his assistance.

## References

Agee, J.K. 1991. Fire history along an elevational gradient in the Siskiyou Mountains, Oregon. Northwest Sci. **65**: 189–199.

Agee, J. 1993. Fire ecology of Pacific Northwest forests. Island Press, Washington, D.C.

Agee, J.K., and Smith, L. 1984. Subalpine tree establishment after fire in the Olympic Mountains, Washington. Ecology, **65**: 810–819.

Agee, J.K., Finney, M., and de Gouvenain, R. 1990. Forest fire history of Desolation Peak, Washington. Can. J. For. Res. **20**: 350–356.

Barnosky, C.W., Anderson, P.M., and Bartlein, P.J. 1987. The northwestern U.S. during deglaciation: vegetational history and paleoclimatic implications. *In* North America and adjacent oceans during the last deglaciation. *Edited by* W.F. Ruddiman and H.E. Wright. Geological Society of America, Boulder, Colo. pp. 289–321.

Benedict, J.B. 2002. Eolian deposition of forest-fire charcoal above tree limit, Colorado Front Range, U.S.A.: potential contamination of AMS radiocarbon samples. Arct. Antarct. Alp. Res. **34**: 33–37.

Bianchi, G.G., and McCave, I.N. 1999. Holocene periodicity in North Atlantic climate and deep ocean flow south of Iceland. Nature (London), **397**: 515–517.

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I., and Bonani, G. 2002. Persistent solar influence on North Atlantic climate during the Holocene. Science (Washington, D.C.), **294**: 2130–2136.

Brooke, R.C., Peterson, E.B., and Krajina, V.J. 1970. The subalpine mountain hemlock zone. *In* Ecology of western North America. *Edited by* V.J. Krajina and R.C. Brooke. Department of Botany, The University of British Columbia, Vancouver, B.C. pp. 147–349.

Carcaillet, C. 1998. A spatially precise study of Holocene fire history, climate and human impact within the Maurienne valley, North French Alps. J. Ecol. **86**: 384–396.

Carcaillet, C. 2001. Are Holocene wood-charcoal fragments stratified in alpine and subalpine soils? Evidence from the Alps based on AMS $^{14}$C dates. Holocene, **11**: 231–242.

Carcaillet, C., and Talon, B. 1996. A view of the wood charcoal stratigraphy and dating in soil: a case study of some soils from the French Alps. Geogr. Phys. Quat. **50**: 233–244.

Clague, J.J., and Mathewes, R.W. 1989. Early Holocene thermal maximum in western North America: new evidence from Castle Peak, British Columbia. Geology, **17**: 277–280.

Clague, J.J., and Mathewes, R.W. 1996. Neoglaciation, glacier-dammed lakes, and vegetation change in northwestern British Columbia, Canada. Arct. Alp. Res. **28**: 10–24.

Clague, J.J., Evans, S.G., Rampton, V.N., and Woodsworth, G.J. 1995. Improved age estimates for the White River and Bridge River tephra, western Canada. Can. J. Earth Sci. **32**: 1172–1179.

Clark, J.S. 1988. Particle motion and the theory of charcoal analysis: source area, transport, deposition, and sampling. Quat. Res. **30**: 67–80.

Clark, J.S. 1990. Fire and climate change during the last 750 yr in northwestern Minnesota. Ecol. Monogr. **60**: 135–159.

Clark, J.S., and Patterson, W.A.I. 1997. Background and local charcoal in sediments: scales of fire evidence in the paleorecord. *In* Sediment records of biomass burning and global change. *Edited by* J.S. Clark, H. Cachier, J.G. Goldammer, and B. Stocks. Springer-Verlag, Berlin. pp. 23–48.

Clark, J.S., and Royall, P.D. 1995. Particle-size evidence for source areas of charcoal accumulation in late Holocene sediments of eastern North American lakes. Quat. Res. **43**: 80–89.

Clark, J.S., Cachier, H., Goldammer, J.G., and Stocks, B. (*Editors*). 1997. Sediment records of biomass burning and global change. Springer-Verlag, Berlin.

Clark, J.S., Lynch, J., Stocks, B.J., and Goldammer, J.G. 1998. Relationships between charcoal particles in air and sediments in west-central Siberia. Holocene, **8**: 19–29.

Cwynar, L.C. 1978. Recent history of fire and vegetation from laminated sediment of Greenleaf Lake, Algonquin Park, Ontario. Can. J. Bot. **56**: 10–21.

Dickman, A., and Cook, S. 1989. Fire and fungus in a mountain hemlock forest. Can. J. Bot. **67**: 2005–2016.

Douglas, G.W. 1972. Subalpine plant communities of the western North Cascades, Washington. Arct. Alp. Res. **4**: 147–166.

Fahnestock, G.R., and Agee, J.K. 1983. Biomass consumption amd smoke production by prehistoric and modern forest fires in western Washington. J. For. **81**: 653–657.

Flannigan, M.D., Bergeron, Y., Engelmark, O., and Wotton, B.M. 1998. Future wildfire in circumboreal forests in relation to global warming. J. Veg. Sci. **9**: 469–476.

Flannigan, M., Campbell, I., Wotton, M., Carcaillet, C., Richard, P., and Bergeron, Y. 2001. Future fire in Canada's broeal forest: paleoecology results and general circulation model – regional climate model simulations. Can. J. For. Res. **31**: 854–864.

Gardner, J.J., and Whitlock, C. 2001. Charcoal accumulation following a recent fire in the Cascade Range, northwestern USA, and its relevance for fire-history studies. Holocene, **11**: 541–549.

Gavin, D.G. 2000. Fire history of a coastal temperate rainforest, Vancouver Island, British Columbia, Canada. Ph.D. thesis, University of Washington, Seattle, Wash.

Gavin, D.G. 2001. Estimation of inbuilt age in radiocarbon-derived ages of soil charcoal for fire history studies. Radiocarbon, **43**: 27–44.

Gavin, D.G., and Brubaker, L.B. 1999. A 6000 year soil pollen record of subalpine meadow vegetation in the Olympic Mountains, Washington, USA. J. Ecol. **87**: 106–122.

Gavin, D.G., McLachlan, J.S., Brubaker, L.B., and Young, K.A. 2001. Postglacial history of subalpine forests, Olympic Peninsula, Washington, USA. Holocene, **11**: 177–188.

Hallett, D.J. 2001. Holocene fire history and climate change in southern British Columbia, based on high-resolution analyses of sedimentary charcoal. Ph.D. thesis, Simon Fraser University, Burnaby, B.C.

Hallett, D.J., and Walker, R.C. 2000. Paleoecology and its application to fire and vegetation management in Kootenay National Park, British Columbia. J. Paleolimnol. **24**: 401–414.

Hallett, D.J., Hills, L.V., and Clague, J.J. 1997. New accelerator mass spectrometry radiocarbon ages for the Mazama tephra layer from Kootenay National Park, British Columbia, Canada. Can. J. Earth Sci. **34**: 1202–1209.

Hallett, D.J., Mathewes, R.W., and Foit, F.F. 2001. Mid-Holocene Glacier Peak and Mount St. Helens We tephra layers detected in lake sediments from southern British Columbia using high-resolution techniques. Quat. Res. **55**: 284–292.

Hebda, R.J. 1995. British Columbia vegetation and climate history with focus on 6 ka BP. Geogr. Phys. Quat. **49**: 55–79.

Hebda, R.J., and Mathewes, R.W. 1984. Holocene history of cedar and native Indian cultures of the North American Pacific Coast. Science (Washington, D.C.), **225**: 711–712.

Hebda, R.J., and Whitlock, C. 1997. Environmental history. Chap. 9. *In* The rain forests of home. *Edited by* P.K. Schoonmaker, B. von Hagen, and E.C. Wolf. Island Press, Washington, D.C. pp. 227–254.

Hemstrom, M.A., and Franklin, J.F. 1982. Fire and other disturbances of the forests in Mount Rainer National Park. Quat. Res. **18**: 32–51.

Heyerdahl, E.K., Brubaker, L.B., and Agee, J.K. 2001. Spatial controls of historical fire regimes: a multiscale example from the interior west, USA. Ecology, **82**: 660–678.

Huff, M.H. 1995. Forest age structure and development following wildfires in the western Olympic Mountains, Washington. Ecol. Appl. **5**: 471–483.

Johnson, E.A., and Fryer, G.I. 1987. Historical vegetation change in the Kananaskis Valley, Canadian Rockies. Can. J. Bot. **65**: 853–858.

Johnson, E.A., and Wowchuk, D.R. 1993. Wildfires in the southern Canadian Rocky Mountains and their relationship to mid-tropospheric anomalies. Can. J. For. Res. **23**: 1213–1222.

Laird, K.R., Fritz, S.C., Maasch, K.A., and Cumming, B.F. 1996. Greater drought intensity and frequency before AD 1200 in the Northern Great Plains, USA. Nature (London), **384**: 552–554.

Lamb, H.H. 1995. Climate, history and the modern world. 2nd ed. Routledge, London, U.K.

Larsen, C.P.S., and MacDonald, G.M. 1998. An 840-year record of fire and vegetation in a boreal white spruce forest. Ecology, **79**: 106–118.

Lepofsky, D.S., Hallett, D.J., Washbrook, K., McHalsie, S., Lertzman, K.P., and Mathewes, R.W. 2003. Documenting precontact plant management on the Northwest Coast: an example of prescribed burning in the central and upper Fraser Valley, British Columbia. *In* Keeping it living: traditional plant tending and cultivation on the northwest coast. *Edited by* D.D. Turner and N.J. Turner. University of Washington Press, Seattle, Ill. In press.

Lertzman, K.P. 1992. Patterns of gap-phase replacement in a subalpine, old-growth forest. Ecology, **73**: 657–669.

Lertzman, K.P., and Krebs, C.J. 1991. Gap-phase structure of a subalpine old-growth forest. Can. J. For. Res. **21**: 1730–1741.

Lertzman, K.P., Gavin, D.G., Hallett, D.J., Brubaker, L.B., Lepofsky, D.S., and Mathewes, R.W. 2002. Long-term fire regime from soil charcoal in coastal temperate rainforests. Conserv. Ecol. [serial online], **6**(2): 5. Available from http://www.consecol.org/vol6/iss2/art5.

Long, C.J., Whitlock, C., Bartlein, P.J., and Millspaugh, S.H. 1998. A 9000-year fire history from the Oregon Coast Range, based on a high-resolution charcoal study. Can. J. For. Res. **28**: 774–787.

Luckman, B.H., Holdsworth, G., and Osborn, G.D. 1993. Neoglacial glacier advances in the Canadian Rockies. Quat. Res. **39**: 144–153.

Lydolph, P.E. 1985. The climate of the earth. Rowman and Allanheld Publishers, Totowa, N.J.

MacDonald, G.M., Larsen, C.P.S., Szeicz, J.M., and Moser, K.A. 1991. The reconstruction of boreal forest fire history from lake sediments: a comparison of charcoal, pollen sedimentological, and geochemical indices. Quat. Sci. Rev. **10**: 53–74.

Masters, A. 1990. Changes in forest fire frequency in Kootenay National Park, Canadian Rockies. Can. J. Bot. **68**: 1763–1767.

Mathewes, R.W. 1973. A palynological study of post-glacial vegetation changes in the University Research Forest, southwestern British Columbia. Can. J. Earth Sci. **19**: 1185–1195.

Mathewes, R.W. 1985. Paleobotanical evidence for climatic change in southern British Columbia during Late-Glacial and Holocene time. *In* Climatic change in Canada 5: critical periods in the Quanternary climatic history of northern North America. *Edited by* C.R. Harington. National Museum of Natural Sciences, National Museums of Canada, Ottawa, Ont. pp. 397–422.

Mathewes, R.W., and Heusser, L.E. 1981. A 12,000 year palynological record of temperature and precipitation trends in southwestern British Columbia. Can. J. Bot. **59**: 707–710.

Mathewes, R.W., and King, M. 1989. Holocene vegetation, climate, and lake-level changes in the interior Douglas-fir biogeoclimatic zone, British Columbia. Can. J. Earth Sci. **26**: 1811–1825.

Mathewes, R.W., and Rouse, G.E. 1975. Palynology and paleoecology of postglacial sediments from the Lower Fraser River Canyon of British Columbia. Can. J. Earth Sci. **12**: 745–756.

Matson, R.G., and Coupland, G. 1995. The prehistory of the northwest coast. Academic Press, San Diego, Calif.

McBean, G.A. 1996. Factors controlling climate of the west coast of North America. *In* high-latitude rainforests and associated ecosystems of the west coast of the Americas. *Edited by* R.G. Lawford, P.B. Alaback, and E. Fuentes. Springer-Verlag, New York. pp. 27–41.

Meidinger, D., and Pojar, J. 1991. Ecosystems of British Columbia. B.C. Ministry of Forests, Victoria, B.C. Spec. Rep. 6.

Meyer, G.A., Wells, S.G., Balling, R.C.J., and Jull, A.J.T. 1992. Response of alluvial systems to fire and climate change in Yellowstone National Park. Nature (London), **357**: 147–150.

Millspaugh, S.H., and Whitlock, C. 1995. A 750-year fire history on lake sediment records in central Yellowstone National Park, USA. Holocene, **5**: 283–292.

Millspaugh, S.H., Whitlock, C., and Bartlein, P.J. 2000. Variations in fire frequency and climate over the past 17,000 yr in central Yellowstone National Park. Geology, **28**: 211–214.

Mohr, J.A., Whitlock, C., and Skinner, C.N. 2000. Postglacial vegetation and fire history, eastern Klamath Mountains, California, USA. Holocene, **10**: 587–601.

O'Brien, S.R., Mayewski, P.A., Meeker, L.D., Meese, D.A., Twickler, M.S., and Whitlow, S.I. 1995. Complexity of Holocene climate as reconstructed from a Greenland ice core. Science (Washington, D.C.), **270**: 1962–1965.

Ohlson, M., and Tryterud, E. 2000. Interpretation of the charcoal record in forest soils: forest fires and their production and deposition of macroscopic charcoal. Holocene, **10**: 519–525.

Osborn, G.D., Robinson, B.J., and Luckman, B.H. 2001. Holocene and latest Pleistocene fluctuations of Stutfield Glacier, Canadian Rockies. Can. J. Earth Sci. **38**: 1141–1155.

Overpeck, J.T., Rind, D., and Goldberg, R. 1990. Climate-induced changes in forest disturbance and vegetation. Nature (London), **343**: 51–53.

Patterson, W.A., Edwards, K.J., and Maguire, D.J. 1987. Microscopic charcoal as a fossil indicator of fire. Quat. Sci. Rev. **6**: 3–23.

Pellatt, M.G., and Mathewes, R.W. 1997. Holocene tree line and climate change in the Queen Charlotte Islands, Canada. Quat. Res. **48**: 88–99.

Pellatt, M.G., Smith, M.J., Mathewes, R.W., and Walker, I.R. 1998. Paleoecology and postglacial treeline shifts in the northern Cascade Mountains, Canada. Palaeogeogr. Palaeoclimatol. Palaeoecol. **141**: 123–138.

Pellatt, M.G., Smith, M.J., Mathewes, R.W., Walker, I.R., and Palmer, S.L. 2000. Holocene treeline and climate change in the subalpine zone near Stoyama Mountain, Cascade Mountains, southwestern British Columbia, Canada. Arct. Antarct. Alp. Res. **32**: 73–83.

Pellatt, M.G., Hebda, R.J., and Mathewes, R.W. 2001. High-resolution Holocene vegetation history and climate from Hole 1034B, ODP leg 169S, Saanich Inlet, Canada. Mar. Geol. **174**: 211–226.

Porter, S.C., and Denton, G.H. 1967. Chronology of neoglaciation in the North American Cordillera. Am. J. Sci. **265**: 177–210.

Potash, L.L., and Agee, J.K. 1998. The effect of fire on red heather (*Phyllodoce empetriformis*). Can. J. Bot. **76**: 428–433.

Price, C., and Rind, D. 1994. The impact of 2 × $CO_2$ climate on lightning-caused fires. J. Clim. **7**: 1484–1494.

Reasoner, M.A., and Huber, U.M. 1999. Postglacial palaeoenvironments of the upper Bow Valley, Banff National Park, Alberta, Canada. Quat. Sci. Rev. **18**: 475–492.

Rochefort, R.M., and Peterson, D.L. 1994. Temporal and spatial distribution of trees in subalpine meadows of Mount Rainier National Park, Washington, U.S.A. Arct. Alp. Res. **28**: 52–59.

Rochefort, R.M., Little, R.L., Woodward, A., and Peterson, D.L. 1994. Changes in sub-alpine tree distribution in western North America: a review of climatic and other causal factors. Holocene, **4**: 89–100.

Rorig, M.L., and Ferguson, S.A. 1999. Characteristics of lightning and wildland fire ignition in the Pacific Northwest. J. Appl. Meteorol. **38**: 1565–1575.

Ryder, J.M., and Thomson, B. 1986. Neoglaciation in the southern Coast Mountains of British Columbia: chronology prior to the late Neoglacial maximum. Can. J. Earth Sci. **23**: 273–287.

Sea, D.S., and Whitlock, C. 1995. Postglacial vegetation and climate of the Cascade Range, central Oregon. Quat. Res. **43**: 370–381.

Skinner, W.R., Stocks, B.J., Martell, D.L., Bonsal, B., and Shabbar, A. 1999. The association between circulation anomalies in the mid-troposphere and area burned by wildland fire in Canada. Theor. Appl. Climatol. **63**: 89–105.

Skinner, W.R., Flannigan, M.D., Stocks, B.J., Martell, D.L., Wotton, B.M., Todd, J.B., Mason, J.A., Logan, K.A., and Bosch, E.M. 2002. A 500 hPa synoptic wildland fire climatology for large Canadian forest fires, 1959–1996. Theor. Appl. Climatol. **71**: 157–169.

Stager, J.C., and Mayewski, P.A. 1997. Abrupt early to mid-Holocene climatic transition registered at the equator and the poles. Science (Washington, D.C.), **276**: 1834–1836.

Stuiver, M., Braziunas, T.F., and Grootes, P.M. 1997. Is there evidence for solar forcing of climate in the GISP2 oxygen isotope record? Quat. Res. **48**: 259–266.

Stuiver, M., Reimer, P.J., Bard, E., Beck, J.W., Burr, G.S., Hughen, K.A., Kromer, B., McCormac, G., Van Der Plicht, J., and Spurk, M. 1998. INTCAL98 radiocarbon age calibration, 24,000–0 cal BP. Radiocarbon, **40**: 1041–1083.

Swain, A.M. 1973. A history of fire and vegetation in northwestern Minnesota as recorded in lake sediments. Quat. Res. **3**: 383–396.

Swain, A.M. 1978. Environmental changes during the past 2000 years in north-central Wisconsin: analysis of pollen, charcoal, and seeds from varved lake sediments. Quat. Res. **10**: 55–68.

Thompson, R.S., and Oldfield, F. 1986. Environmental magnetism. Allen and Unwin Ltd., London, U.K.

Thompson, R.S., Whitlock, C., Bartlein, P.J., Harrison, S.P., and Spaulding, W.G. 1993. Climatic changes in the western United States since 18,000 yr BP. *In* Global climates since the last glacial maximum. *Edited by* H.E. Wright, Jr., J.E. Kutzbach, T. Webb, III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein. University of Minnesota Press, Minneapolis, Minn. pp. 468–513.

Tinner, W., Conedera, M., Ammann, B., Gaggeler, H.W., Gedye, S., Jones, R., and Sagesser, B. 1998. Pollen and charcoal in lake sediments compared with historically documented forest fires in southern Switzerland since AD 1920. Holocene, **8**: 31–42.

Turner, N.J. 1991. Burning mountainsides for better crops: aboriginal burning in British Columbia. Archeol. Mont. **32**: 57–93.

Veblen, T.T., and Alaback, P.B. 1996. A comparative review of forest dynamics and disturbance in the temperate rainforests of North and South America. *In* High-latitude rainforests and associated ecosystems of the west coast of the Americas. *Edited by* R.G. Lawford, P.B. Alaback, and E. Fuentes. Springer-Verlag, New York. pp. 173–213.

Wainman, N., and Mathewes, R.W. 1987. Forest history of the last 12 000 years based on plant macrofossil analysis of sediment from Marion Lake, southwestern British Columbia. Can. J. Bot. **65**: 2179–2187.

Watson, E., and Luckman, B.H. 2001. Dendroclimatic reconstruction of precipitation for sites in the southern Canadian Rockies. Holocene, **11**: 203–213.

Whitlock, C. 1992. Vegetational and climatic history of the Pacific Northwest during the last 20,000 years: implications for understanding present day biodiversity. Northwest Environ. J. **8**: 5–28.

Whitlock, C., and Millspaugh, S.H. 1996. Testing the assumptions of fire history studies: an examination of modern charcoal accumulation in Yellowstone National Park, USA. Holocene, **6**: 7–15.

Whitlock, C., Bradbury, J.P., and Millspaugh, S.H. 1997. Controls on charcoal distribution in lake sediments: case studies from Yellowstone National Park and northwestern Minnesota. *In* Sediment records of biomass burning and global change. *Edited by* J.S. Clark, H. Cachier, J.G. Goldammer, and B. Stocks. Springer-Verlag, Heidelberg. pp. 367–386.

Woodhouse, C.A., and Overpeck, J.T. 1998. 2000 years of drought variability in the central United States. Bull. Am. Meteorol. Soc. **79**: 2693–2714.

Worona, M.A., and Whitlock, C. 1995. Late Quaternary vegetation and climate history near Little Lake, central Coast Range, Oregon. Geol. Sci. Assoc. Bull. **107**: 867–876.