Holocene fire and paleoclimate history of a small lake in the Lower Seymour Valley, British Columbia

by

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BSc., University of Kansas, 2016

Thesis Submitted in Partial Fulfillment of the
Requirements for the Degree of
Master of Resource Management

in the
School of Resource and Environmental Management
Faculty of Environment

SIMON FRASER UNIVERSITY
Fall 2022

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Abstract

In the coming century climate variability is projected to increase along the Pacific Coast of Canada, increasing the need for land managers to understand how ecosystems change in response to new or enhanced disturbances. Southern British Columbia (BC) is thought to have experienced warm and dry climate conditions with higher than modern fire activity in the past, during the xerothermic interval (9500 - 7000 cal yr BP). In this study, I reconstructed past climate-fire-vegetation changes from a 13,000-year record from Lost Lake in Vancouver's Lower Seymour Conservation Reserve, BC. Contrary to other sites, the moist coastal western hemlock forest at this site remained cool and moist with low fire activity throughout the xerothermic period. Instead, peak fire frequencies were observed during the cool and moist Neoglacial period (4500 cal yr BP - present), when human activity became prevalent. These results have implications for the managed watershed's resilience to fire and response to future warming conditions.

Keywords: paleoclimate; paleoecology; wildfire; disturbance; novel ecosystem; palynology

Dedication

This thesis is dedicated to my brothers Mac and Cal, who are pretty good as far as brothers go.

Acknowledgements

A very big thank you to my supervisors Dr. Karen Kohfeld, Dr. Marlow Pellatt, and Dave Dunkley for their unwavering support, guidance, mentorship, and frequent pep-talks throughout this project. I appreciate your patience as I slowly shifted my brain from rocks to plants. Karen, thank you for instilling in me a deep respect for the Oxford comma and proper conjunction usage. I would also like to thank Dr. Rolf Mathewes for his assistance with all things pollen. Thank you to Metro Vancouver's Watershed Management and Environment Division and Mitacs for providing generous financial support, and to Parks Canada for providing laboratory space, equipment, and supplies to conduct my research despite the roadblocks caused by the COVID-19 Pandemic.

Hasini Basnayake, Jacqui Levy, Dylan Cunningham, and Cass Perreira supported me in the lab, throughout the writing process, and as good friends. My lovely parents provided consistent love and encouragement all along the way. I would also like to thank the following friends, family and colleagues for supporting me in various ways throughout this process: Matt Fagan, Daggery Folant, Jordan Helms, Dr. Nicholas Rabbit, Kasha Riddle, Herma Dings, Martin Sneeb, Tina "Tiny" O'Malley, Bart Harley Jarvis, Snela Poly, The Rev. Charlie Chuckl, Hillary Schartz-Cox, Logan Gregorchuk, Mrs. Snell, Eiry Loots, Jessica Klitt, Kelly Duotong, Betty Hostetler, Tony and Tricia Fillet, and Trampus Goodman.

And finally, thank you Emily D-H for being willing to visit the Lost Lake newts with me over and over again.

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List of Acronyms

^{14}C	Radioisotope of carbon (C)
^{210}Pb	Radioisotope of lead (Pb)
AMS	Accelerator mass spectrometry
B-A	Bølling-Allerød event
BCWS	British Columbia Wildfire Service
BGCZ	Biogeoclimatic zone
Cal yr BP	Calibrated years before present (present being 1950 AD)
CDF	Coastal Douglas Fir zone
CHAR	Charcoal accumulation rate (pieces/cm^2/yr)
CRS	Constant rate of supply model
CWHvm1	Coastal Western Hemlock, very wet maritime, submontane zone
ENSO	El Niño Southern Oscillation
FI	Fire incidence
FVFP	Fraser Valley Fire Period
GOF	Goodness-of-fit
LIA	Little Ice Age
LSCR	Lower Seymour Conservation Reserve
mFRI	Mean fire return interval
MWP	Medieval Warm Period
PDO	Pacific Decadal Oscillation
SFU	Simon Fraser University
SNI	Signal-to-Noise Ratio
SST	Sea Surface Temperature
TSLF	Time since last fire
WSA	Water Supply Area
YD	Younger Dryas

Chapter 1. Introduction

1.1. Introduction

Understanding historical changes in climate and ecology is a necessary component of managing the environmental shifts associated with climate change. By the year 2050, temperatures in British Columbia are estimated to increase by 1.3 – 2.7 °C (Government of British Columbia, 2019), and regional wildfire activity and severity are expected to intensify (Mote et al., 2010). While the impacts of wildfires on contemporary forest and water resources along the northwest coast of North America have been well documented (Green et al., 1999; Carignan et al., 2000; Zwolinski, 2000; Gavin et al., 2007; Sankey et al., 2017), understanding how climate-fire-vegetation interactions in South Coastal B.C. will change in response to increasing temperatures and wildfire severity are less clear. Paleoecology is a well-regarded method of understanding past interactions between climate and vegetation. Quantifying past disturbances and the associated ability of ecosystems to withstand change is a valuable means of identifying and evaluating disturbance regimes, understanding the timing and significance of ecosystem shifts, and planning for future land management and restoration projects (Landres et al., 1999; Swetnam et al., 1999, Davies & Bunting, 2010; Whitlock et al., 2010; Kidwell, 2015; Pellatt et al., 2015; Murphy et al., 2019).

In the Lower Mainland of British Columbia, Canada, the Metro Vancouver Regional District is responsible for managing much of the land on Vancouver's North Shore (Metro Vancouver, 2018). Drinking water is held in the Capilano, Seymour and Coquitlam watersheds, which together comprise the Water Supply Area (WSA) and provide drinking water for approximately 2.6 million people in Vancouver and the Lower Mainland. While these areas are currently protected with restricted access, climate change impacts are expected to affect these areas at rates exceeding the natural range of variability (Canadian Parks Council, 2013). A 2016 report by the Metro Vancouver Regional District states that a lack of fire science research focused on the wet ecosystems of Vancouver's WSA has resulted in uncertainty regarding future wildfire activity projections in the watersheds (van der Kamp, 2016). The cool, moist Coastal Western Hemlock zone, which occupies much of the water supply area, has a large geographic range but a narrow temperature range (Meidinger & Pojar, 1991), indicating that the rising temperatures projected within the next century may have detrimental effects. Coastal temperate rainforests generally contain high levels of biomass for burning but are considered climate- and ignition-limited systems because of their characteristic high moisture levels (Power et al., 2008; Bowman et al., 2009; Whitlock et al., 2015; Hoffman, Lertzman & Starzomski, 2017). However, few long-term climate or fire history studies have focused on reconstruction of coastal temperate rainforest zones, with most reconstructions being further inland (Wainman & Mathewes, 1987) in drier subzones (Brown & Hebda, 2002b; Brown et al., 2019) or not representative of the entire post-glacial record (Pellatt, Mathewes & Clague, 2001; Hoffman, Lertzman & Starzomski, 2017; Murphy et al., 2019).

The Holocene period (11,700 cal yr BP – present; Walker et al. (2009)) is of interest for understanding longer-term changes in climate and wildfire activity throughout western North America~~,~~ because the scale of temperature and precipitation variability may have been larger than that seen in the past 200 years. Two commonly used proxies that have been applied effectively in past regional studies are palynomorph data (Mathewes, 1973; Mathewes 1989; Hebda, 1983, Cwynar, 1987; Hebda 1995; Heusser, 1983; Allen, 1995; Barnosky, 1985; Pellatt et al., 2000; Gavin et al., 2001; Pellatt et al., 2001; Brown and Hebda, 2002a,b; Gavin and Brubaker, 1999; Brown et al., 2019) and charcoal records (Cwynar, 1987; McLachlan and Brubaker, 1995; Gavin et al., 2001; Gavin et al., 2003; Brown & Hebda, 2002a; Brown & Hebda, 2002b; Sugimura, Sprugel, Brubaker, & Higuera, 2008; Hallett et al., 2003; Derr, 2014; Murphy et al., 2019; Pellatt et al., 2015; Lucas & Lacourse, 2013; Brown et al., 2019; Prichard et al., 2009 (Appendix A; Figure 1).

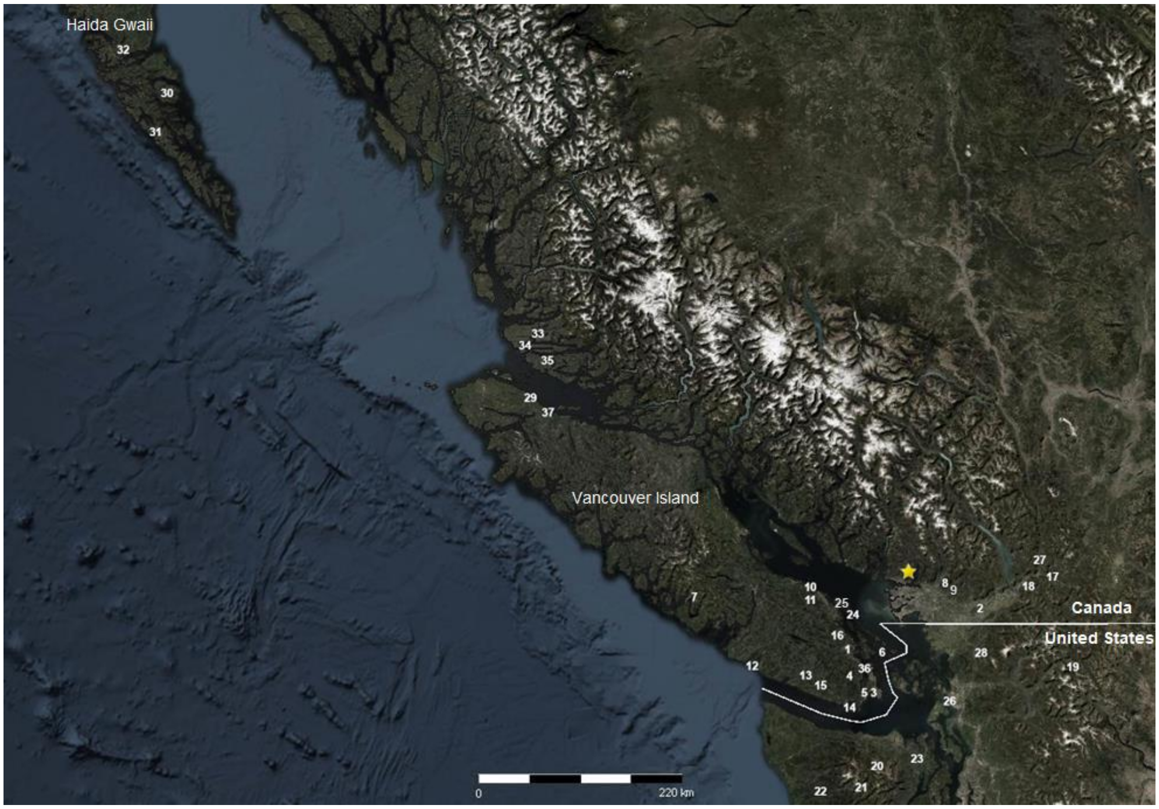


Figure 1. Study region of Western North America showing Lost Lake (yellow star) and previous paleoecological studies conducted in the Coastal Western Hemlock (CWH) and Coastal Douglas Fir (CDF) biogeoclimatic zones using pollen and/or charcoal as proxies (Appendix A).

Study sites include: (Star) Lost Lake (this study); (1) Somenos Lake and (2) Chadsey Lake (Murphy et al., 2019); (3) Quamichan Lake(Pellatt et al., 2015); (4) Begbie Lake (Brown et al., 2019); (5) Florence Lake (Pellatt et al., 2015); (6) Roe Lake (Lucas & Lacourse, 2013); (7) Clayoquot Lake (Gavin et al., 2003); (8) Marion Lake (Wainman & Mathewes, 1987); (9) Mike Lake (Pellatt, Mathewes & Clague, 2001); (10) Enos Lake (Brown & Hebda, 2002a); (11) Boomerang Lake (Brown & Hebda, 2002a); (12) Whyac Lake (Brown & Hebda, 2002b); (13) Pixie Lake (Brown & Hebda, 2002b); (14) East Sooke Fen (Brown & Hebda, 2002b); (15) Walker Lake (Brown & Hebda, 2003); (16) Porphyry Lake (Brown & Hebda, 2003); (17) Frozen Lake (Hallett et al., 2003); (18) Mt. Barr Cirque Lake (Hallett et al., 2003); (19) Panther Potholes (Prichard et al., 2009); (20) Moose Lake and (21) Martins Lake (Gavin et al., 2001); (22) Yahoo Lake (Gavin et al., 2013); (23) Crocker Lake (McLachlan and Brubaker, 1995); (24) and (25) Valdes Island Off and On Site Bogs (Derr, 2014); (26) Mt. Constitution (Sugimura, Sprugel, Brubaker, & Higuera, 2008); (27) Pinecrest Lake (Mathews, 1975); (28) Mosquito Lake Bog (Hansen and Easterbrook, 1974); (29) Bear Cove Bog (Hebda, 1983); (30) Louise Pond (Pellatt & Mathewes, 1994); (31) SC1 Pond and (32) Shangri-La Bog (Pellatt & Mathewes, 1997); (33) Tiny Lake (Galloway et al., 2007); (34) Two Frog Lake (Galloway et al., 2009); (35) Woods Lake (Stolze et al., 2007); (36) ODP Hole 1034B (Pellatt et al., 2001); and (37) Misty Lake (Lacourse, 2005). Base layer and inset map sources are Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, NOAA NGDC, and the GIS User Community.

Pollen percentage diagrams are used to infer long-term changes in vegetation by representing the percentages of individual pollen taxa relative to the total number of pollen grains counted at a given depth interval. When graphed together, taxa percentages show temporal shifts in assemblage. However, because the final percentages are interdependent, the results should be analyzed using prior understanding of vegetational regimes and regional climatic change. Similarly, pollen accumulation rates (PAR) represent changes in the influx of pollen over time. PAR is useful as a secondary analysis to determine the independent nature of pollen records, because pollen is plotted as grains/unit area/time.

The most common proxies for inferring fire history are sediment charcoal accumulation rates (CHAR), the mean fire return interval (mFRI) and fire frequency. CHAR is calculated by measuring the number of charcoal pieces (usually >125-150 μm) per square centimetre per year, and once detrended, provides a measure of fire events (indicated by number of charcoal accumulation peaks) relative to background changes in charcoal accumulation rates (Whitlock & Larson, 2001; Higuera, Gavin, Bartlein & Hallett, 2010). A complication to the interpretation of CHAR is the contribution of noise in the record due to disturbance (i.e., sediment mixing, bioturbation) and charcoal deposited from adjacent watersheds during non-fire years (Whitlock & Larson, 2001). However, recent methods of detrending CHAR results allow for reliable separation of signal and noise (Higuera et al.,2010). The mean fire return interval (mFRI) is most often calculated by averaging the time between fire episodes (Agee, 1993) and is useful in understanding historical fire regimes and determining a system's natural range of variability (Cyr, Gauthier, Bergeron, & Carcaillet, 2009). Similarly, fire frequency provides insight into continual changes in fire activity over time by defining a moving average of the number of fires within a 1000-year period (Higuera, 2009).

Natural ranges of variability are often discussed in resource management and restoration ecology (Swetnam et al., 1999; Jackson et al., 2009; Whitlock et al., 2003). Broadly speaking, a natural range of variability is the disturbance-driven spatial and temporal variability of an ecological system before anthropogenic influences, and management strategies often attempt to mimic these ranges (Doyon et al., 2008). Prior to European settlement, Indigenous peoples across North America used fire as a means of creating grazing areas, facilitating hunting and berry harvesting, and clearing paths for travel (Pyne, 1982; Boyd, 1999; Hessburg, 2003, Lake & Christianson, 2020). The arrival of settlers brought disease and caused a decline in controlled burns, as they instead tried to prevent fires from damaging land and settlements. Over the past century in North

America, fire suppression was the dominant tool used by forest managers for wildfire management (Rothman, 2005; Parks Canada, 2020). Land managers have since learned that the long-term adverse effects of suppression generally outweigh the short-term benefits (Ryan et al., 2013). A major negative effect of suppression is the extension of fire return intervals, causing unnaturally long breaks in the fire cycle during which fuels build up to dangerous levels, often causing uncontrollable high-intensity, stand replacing fires. Other impacts include changes to soil composition and forest health (Harmon, 1984; Hatten et al., 2005; Allen et al., 2002, Stephens & Ruth, 2002). In recent decades, management practices have shifted towards controlled or prescribed burns in order to restore ecosystems to their pre-colonial fire regimes as much as possible (Fernandes & Botelho, 2003). The goal is to re-establish the vegetation mosaic and biodiversity that were present prior to suppression. Many protected areas in North America have updated their fire management plans to utilize prescribed burning (Stephens & Ruth, 2005; U.S. National Park Service, 2014; Parks Canada, 2019).

The Greater Vancouver Regional District (GVRD) has favored wildfire management over suppression for decades, but they face a unique challenge within their watersheds. Burning has the potential to negatively affect water and air quality, which could have far-reaching and long-term effects on drinking water supply and public health. The proximity of the watersheds and the Lower Seymour Conservation Reserve (LSCR) to high-density population zones delivers added risks to property damage and public safety (Greater Vancouver Regional District, 2002). Fires can cause significant alteration to the hydrologic cycle within watersheds because of the potential disturbance to vegetation, litter, and soil health, affecting surface runoff, erosion, peak discharges, and ultimately water quality (Zwolinksi, 2001). Low-severity fires (i.e., in the case of prescribed burns) show little to no hydrologic impacts as long as mineral soil is not exposed (Zwolinksi, 2001). Nevertheless, fires in low elevation areas and areas close to recreation facilities are closely managed by the GVRD to minimize risks to human safety (GVRD, 2002).

South coastal British Columbia has already experienced significant summer temperature increases in the past ~75 years, regionally averaging a rise of 1.4°C during the 20th century and early 21st century (Mote, 2003; Vincent et al., 2007; Meyn et al., 2012; B.C. Ministry of Environment, 2016). Since mid-century, the average number of large fires (>200 ha) and annual area burned by wildfire activity have also increased in

British Columbia (Hanes et al., 2018), and are expected to continue rising (Metro Vancouver, 2016; Schoennagel et al., 2017). 2017 was the worst wildfire season in BC history (12,161 km^2 burned) until the following year, which again broke records with approximately 13,000 km^2 burned (B.C. Wildfire Service, 2021). Notably, during the 2018 fire season, several lightning-ignited fires burned large patches of Coastal Western Hemlock (CWH) forest on northern Vancouver Island, confirming that fire disturbance is a threat even to moist forests during dry summers (Brown et al., 2019). In the face of predicted increases in temperature and fire activity, land managers need to understand how ecosystems may change when faced with new or enhanced disturbances. Paleo fire and vegetation data provide information on past ecosystem behaviors that are location-specific, which can supplement larger-scale regional or global climate model projections to produce more robust predictions of future change at local scales.

This study uses paleoecology to understand past climate-fire-vegetation changes within the Seymour Watershed and the associated response of forests. Investigating the mechanisms behind ecosystem changes during the Holocene provides insight into how local watersheds may respond to climate change in the coming decades and can be used to inform adaptation strategies within the Metro Vancouver WSA. For this thesis I have collected lake sediment cores from Lost Lake (49.40°N 122.98°W; 235 masl), a small glacial lake within the Lower Seymour Conservation Reserve which lies in the very moist maritime variant of the Coastal Western Hemlock biogeoclimatic zone at an elevation of 235 masl. I have used palynological and charcoal analyses to examine changes in paleo-vegetation and fire dynamics over the past ~13,900 years, which covers the very end of the Pleistocene and the entire Holocene period to the present day. The goal of my work was to address the following research questions:

- How do the vegetation composition and fire regime change at Lost Lake during known periods of climatic change, including immediately following deglaciation (13,900 – 11,700 cal yr BP), the xerothermic period (9500 – 7000 cal yr BP), and the Neoglacial period (4500 cal yr BP to present)?

- Do changes in charcoal influx (which suggest changes in fire activity) correspond with pollen zonation changes within my record, suggesting a relationship between fire activity and forest composition?

- Can shifts in vegetation biomes and fire regimes be tied to changes in climate or human activity throughout the post-glacial period (11,500 cal yr BP – present)?

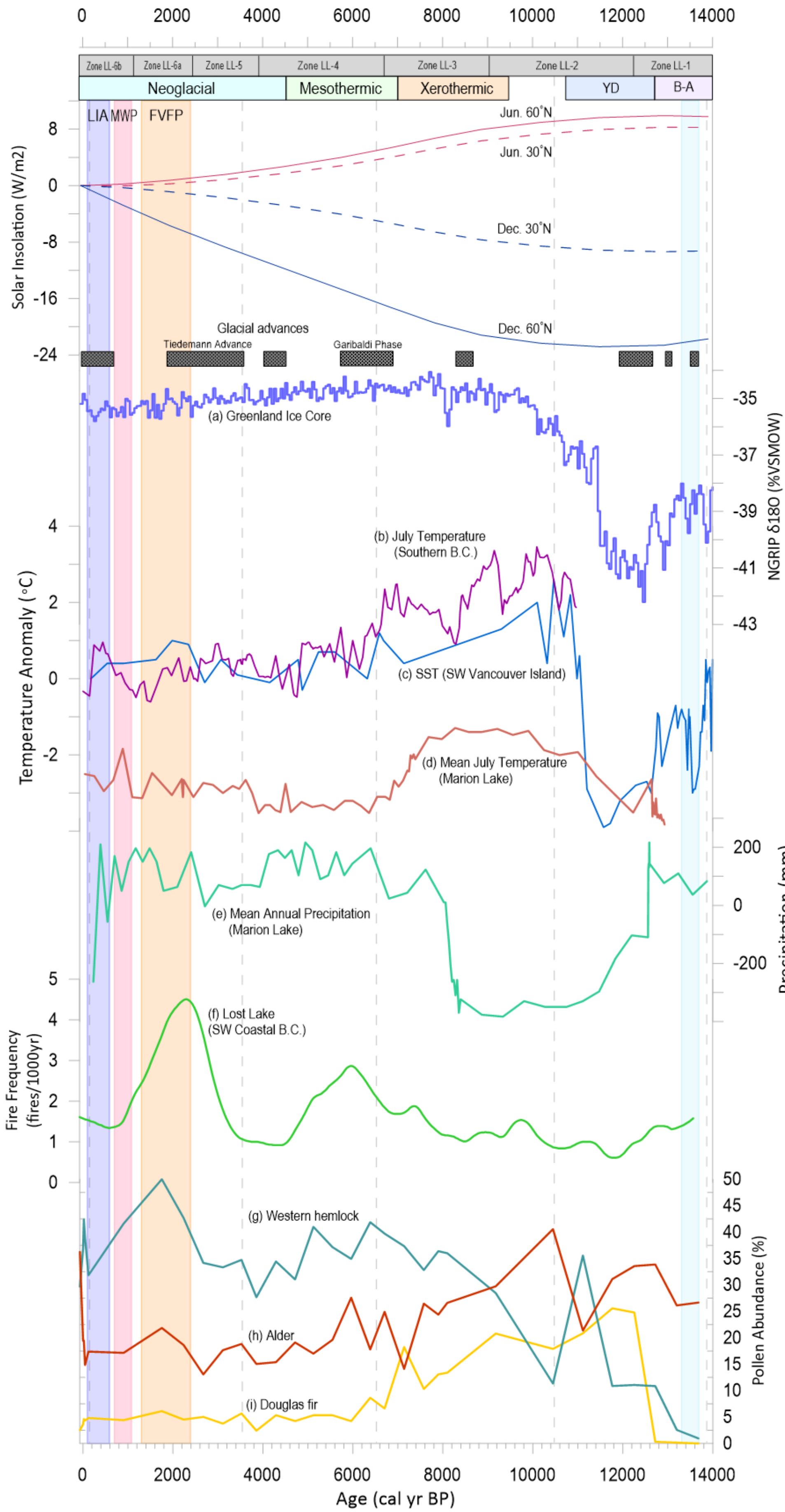
In this thesis I will use pollen and charcoal analysis to show the range of climatic changes in a coastal temperate site from the early Holocene to modern period. My work will produce a record of vegetation changes and charcoal fluxes over the past 13,000 years at Lost Lake, which can be placed in the context of regional climatic changes throughout the Holocene to interpret long-term ecosystem changes in the LSCR. This site shows some response to climate change but is muted in comparison to sites further inland, while fire frequency in my record is often not linked to the canonical representation of climate periods in the literature (Mathewes, 1973; Wainman & Mathewes, 1987; Walker & Pellatt, 2003; Walker & Pellatt, 2008; Walsh et al., 2008). I will discuss the importance of understanding the paleoclimate at Lost Lake to management of the watershed and planning for future climate change.

1.2. Paleoclimate History of Southwestern British Columbia

Regional Holocene records of the past ~12,000 years suggest that the forested ecosystems in and around the Fraser Valley experienced large disturbances over relatively short (centennial to several thousand year) timescales (Walker and Pellatt, 2003). Previous paleoclimatic studies of British Columbia have divided the Holocene into major periods of climatic change. The Younger Dryas, occurring approximately between 12,900 – 11,600 cal yr BP, marked a period of cooling in much of the Northern Hemisphere around the north Atlantic Ocean (Cheng et al., 2020), with moderate cooling effects observed on Canada's Pacific coast as well (Mathewes et al., 1993). The late-glacial (>11,500 cal yr BP) palynomorph record documented by Mathewes (1973) at Marion Lake, as well as macrofossil analysis from the Fraser Lowlands (Wainman and Matthews, 1987; Clague et al., 1997), suggest a pine-dominated open-canopy vegetation regime with fir (*Abies)* and spruce (*Picea*) abundances increasing soon after.

The xerothermic period followed the Younger Dryas, from approximately 9,500 cal yr BP to 7,000 cal yr BP, and palynological evidence suggests temperatures in southwestern British Columbia were as much as 2 °C warmer than present (Mathewes and Heusser, 1981). Specifically, low-elevation coastal sites have shown palynomorph evidence of a regime dominated by Douglas-fir (*Pseudotsuga menziesii*) (generally associated with warmer, drier climates), alder (*Alnus*) and bracken fern (*Pteridium aquilinum*) (Wainman & Mathewes, 1987; Walker & Pellatt, 2003). This, combined with little to no record of western hemlock (*Tsuga heterophylla*), western redcedar (*Thuja plicata*) and spruce (*Picea*) (Mathewes, 1973 (Marion Lake); Brown & Hebda, 2002 (Pixie Lake); Walker & Pellatt, 2003; Galloway et al., 2007 (Tiny Lake); Brown et al., 2019 (Begbie Lake)), as well as higher charcoal abundance in some sediment records (Hallett et al., 2003 (Frozen and Mount Barr Cirque Lakes); Brown et al., 2019 (Begbie Lake)), are indicative of the xerothermic's warm, dry climate. The eruption of Mount Mazama (Crater Lake, Oregon) at approximately 7600 cal yr BP (Zdanowski et al., 1999) deposited a tephra layer across over 100,000 km^2 of western North America including southern British Columbia (Sarna-Wojcicki & Davis, 1991), providing an important chronological control for the mid-Holocene.

The transition out of the xerothermic was experienced differently throughout the province. Some southern interior locations experience a warm, moist 'mesothermal' from ca. 7000 – 4500 cal yr BP (Hebda, 1995), whereas this interval was not well defined at wet coastal sites, with many instead exhibiting a gradual transition to cooler, moist conditions throughout the mid-Holocene (Pellatt & Mathewes, 1997; Brown & Hebda, 2002; Gavin et al., 2003; Lacourse, 2005; Galloway et al., 2007). Regional palynological evidence generally shows a gradual increase in precipitation and a transition to cooler temperatures beginning between 7500 and 6000 cal yr BP on the southern coast (Mathewes, 1973; Mathewes and Heusser, 1981; Pellatt et al., 2001), and even earlier at sites on the central and northern coast of BC (Spooner, 1997, 2002; Galloway et al., 2009). At ~5000 cal yr BP, summer solar insolation was in decline (Fig. 2) and regional records suggest that the intensity of the Aleutian low pressure system increased (Heusser et al., 1999), causing a decline in summer temperatures, wetter conditions, and mid-Holocene cooling in southwestern British Columbia (Walker & Pellatt, 2003). Fire activity during this period was correspondingly low in British Columbia and western Washington (Wainman & Mathewes, 1987; Gavin et al., 2001; Sugimura et al., 2008; Prichard et al., 2009; Gavin et al., 2013), as many fuels were likely too wet to ignite and burn. Far fewer sites, and dominantly those in drier biogeoclimatic zones, report an increase in fire incidences between ca. 5,000-3,500 cal yr BP (Brown and Hebda, 2002; Hallett et. al., 2003; Murphy et al., 2019). Most attribute the higher charcoal influx to increased anthropogenic land use, a rise in lightning-strike ignitions (Brown and Hebda, 2002; Hallett et al., 2003) or intensified summer droughts (Murphy et al., 2019), due to the fact that the cool, moist regional climate during this time period simply was not favorable to high fire activity.



Zone LL-6b
Zone LL-6a
Zone LL-5
Zone LL-4
Zone LL-3
Zone LL-2
Zone LL-1
Neoglacial
Mesothermic
Xerothermic
YD
B-A
LIA
MWP
FVFP
Solar Insolation (W/m2)
Jun. 60˚N
Jun. 30˚N
Dec. 30˚N
Dec. 60˚N
Glacial advances
Tiedemann Advance
Garibaldi Phase
(a) Greenland Ice Core
NGRIP δ18O (%VSMOW)
(b) July Temperature (Southern B.C.)
Temperature Anomaly (°C)
(c) SST (SW Vancouver Island)
(d) Mean July Temperature (Marion Lake)
(e) Mean Annual Precipitation (Marion Lake)
Precipitation (mm)
Fire Frequency (fires/1000yr)
(f) Lost Lake (SW Coastal B.C.)
(g) Western hemlock
(h) Alder
(i) Douglas fir
Pollen Abundance (%)
Age (cal yr BP)

Figure 2. Synthesis of regional climatic data in the Northern Hemisphere from 14,000 cal yr BP to present.

Grey bars represent timing of Lost Lake Pollen Zones LL-1 to LL-6b. Colored bars represent timing of major climatic events: Bolling-Allerod (B-A, purple, Ivanovic et al., 2016), Younger Dryas (YD, blue, McManus et al., 2004; Renssen et al., 2015), Xerothermic (orange, Hebda, 1995), Mesothermic (green, Hebda, 1995), Neoglacial (light blue, Walker & Pellatt, 2003). Vertical bars represent timing of the Fraser Valley Fire Period (FVFP, light orange, Hallett et al., 2003), Medieval Warm Period (MWP, pink, Mann et al., 2009), and Little Ice Age (LIA, dark blue, Grove, 2001). Dashed grey vertical lines represent drive breaks in the Lost Lake core. Solar insolation records are from Berger & Loutre (1991) and NOAA/NGDC Paleoclimatology Program. Red solid line represents June insolation at 60˚N; red dashed line represents June insolation at 30˚N; blue solid line represents December insolation at 60˚N; blue dashed line represents December insolation at 30˚N. Historic solar insolation values were subtracted from modern values. Glacial advance periods are from Ryder & Thomson (1986), Friele & Clague (2002), Menounos et al. (2009), and Gavin et al. (2011). NGRIP $\delta^{18}O$ record (a) is from NGRIP members (2007); "VSMOW" acronym is Vienna Standard Mean Ocean Water. Temperature anomaly records are from (b) a composite record of July temperature anomalies from four chironomid records in southern BC (Frozen Lake, Rosenberg et al. (2004); North Crater Lake and Lake of the Woods, Palmer et al. (2003); Windy Lake, Chase et al. (2008)), (c) alkenone-derived SSTs from core JT96-09PC off the southwestern coast of Vancouver Island (Kienast & McKay, 2001) and (d) a pollen-based transfer function from the Marion Lake record (Mathewes, 1973; Heusser, 1985). Modern mean annual SST temperature (acquired from DFO Amphitrite Point Lightstation SST Data Archives) was subtracted from SST values, and modern mean July temperature (acquired from Government of Canada's Haney UBC Research Forest Station) was subtracted from the Marion Lake record. Precipitation record (e) was derived from pollen-based transfer function from the Marion Lake record (Mathewes, 1973; Heusser, 1985). Modern mean annual precipitation (acquired from Government of Canada's Haney UBC Research Forest Station) was subtracted from all values. Fire frequency record (f) and CONISS-derived pollen abundance records ((g), Coastal western hemlock; (h), Alder; (i) Douglas fir) were taken from Lost Lake (this study).

Finally, the combined effects of multiple glacial advances across much of British Columbia and the continued intensified Aleutian low pressure system established the cool, moist conditions of the Neoglacial period (ca. 4500 cal yr BP – present) (Heusser, 1983; Hebda, 1995; Pellatt and Mathewes, 1994; Clague and Mathewes, 1996). The effects of the intensified Aleutian low pressure period are thought to have continued until 2400 – 2000 cal yr BP, causing low or moderately low fire frequency throughout much of coastal western North America (Gavin et al., 2001; Hallett et al., 2003; Walker & Pellatt, 2003; Murphy et al, 2019).

From ca. 2400 cal yr BP to present, multiple shorter, discrete climate intervals have been observed in coastal B.C. An increase in fire activity and drought frequency between 2400 and 1300 cal yr BP has been referred to as the Fraser Valley Fire Period (FVFP) (Hallett et al., 2003). Regional studies have also found increased charcoal presence indicating higher fire frequency at sites in southwestern British Columbia (Pellatt et al., 2001; Hallett et al., 2003), eastern British Columbia (Hallett and Walker, 2000), western Washington (Gavin et al., 2001), and coastal Oregon (Long et al., 1998). Observations of the FVFP in the paleo record are location-dependent, as various potential local-scale factors (i.e., an increase in lightning due to enhanced high-pressure circulation (Rorig & Ferguson, 1999), anthropogenic ignitions, summer drought conditions, aspect) influenced the regional occurrence of fire.

From 1300 cal yr BP onwards, the climate in the Fraser Valley remained relatively similar to modern conditions, with short intervals of climatic variability. (Walker & Pellatt, 2003; Hallett et al., 2003). The Medieval Warm Period (MWP; ~1000 – 600 cal yr BP), dominantly observed in northern Europe and parts of eastern North America, has been interpreted as representing a localized brief period of increased fire activity characterized by charcoal influx in the eastern Fraser Valley (Hallett et al., 2003) and Gulf Islands (Lucas and Lacourse, 2013). The Little Ice Age (LIA; 600-150 cal yr BP) marked a return to a moist and cool climate, with records from Vancouver Island (Brown and Hebda, 2002; Grove, 2013; Pellatt, 2015) displaying a decrease in fire frequency, and records from southwestern B.C. and the Coast Mountains indicating multiple glacial advances (Ryder, 1986; Luckman, 1995; Pitman and Smith, 2012). Similar to the FVFP, observations of the Medieval Warm Period and Little Ice age are location-dependent and not found in all regional records.

Chapter 2. Methods

2.1. Study Area: Lost Lake, North Vancouver, BC

The Lost Lake site (Fig. 3) sits on Vancouver's North Shore, in the lowlands of the southernmost range of the Coast Mountains. The area is characterized by steep coastal granite mountains carved and shaped by glacial action, which flowed southward through the valleys and fjords of the region before merging into a huge, singular ice sheet in the Strait of Georgia. Post-glaciation, isostatic rebound and sediment buildup formed lowlands and canyons at the base of the mountains (Jakob & Weatherly, 2003). Remnants of this activity can be seen in the glaciofluvial deposits and till visible along riverbanks and roadcuts on the North Shore.

Figure 3. Lost Lake study site (facing west) and surrounding coastal western hemlock forest.

Lost Lake is a 3.8-hectare water body located at 49 24’00” N, 122 58’00” W, (elevation 235 masl) within the Metro Vancouver-governed Lower Seymour Conservation Reserve (LSCR) in North Vancouver, British Columbia (Figure 1). The lake has a maximum measured depth of 12m, with one inflow at the south end of the lake and one main outflow (Lost Creek) located at the north end. A steep hillslope runs along the eastern shoreline of the lake. Large boulders along the edge and a visible trail of alder growth down the slope indicate landslide activity within the past 40 years. Lost Lake was likely scoured out during the Fraser Glaciation (beginning approximately 30,000 years ago), when many of the landform and geomorphological features of the Lower Mainland were formed (Acres, 1997).

British Columbia is divided into fourteen biogeoclimatic zones based on regional similarities in climate and soil that produce similar climax vegetation communities. These zones are further divided into subzones and variants based on relative precipitation (moisture), relative temperature, and continentality (Green & Klinka, 1994). The major biogeoclimatic zones in North Vancouver are the CWH and Mountain Hemlock (MH) zones. The Coastal Western Hemlock dry maritime (CWHdm) subzone occurs along the shoreline and at lowest elevations. The Coastal Western Hemlock very wet maritime submontane (CWHvm1) subzone is the most extensive unit in the Vancouver Forest Region (Green & Klinka, 1994) and occurs at elevations above the CWHdm but generally below 650 m. The Coastal Western Hemlock very wet maritime montane subzone (CWHvm2) occurs above the CWHvm1, with an elevation range of approximately 650 – 1000 masl. Above the CWHvm2, the Mountain Hemlock moist maritime windward (MHmm1) subzone occurs to a maximum of 1350 masl. Late fall and winter (October to March) in this region are characterized by northeast Pacific storms bearing heavy precipitation. Less commonly, Arctic outflow winds build up in the interior and blow southward through river valleys, bringing short bursts of cold and wind to the Puget-Georgia basin (Walker & Pellatt, 2003). Summers are generally less rainy, with warm, dry conditions more recently exacerbated by uncharacteristically long heat waves (Eyquem & Feltmate, 2022).

My study area sits within the Coastal Western Hemlock very wet maritime submontane (CWHvm1) biogeoclimatic zone (Meidinger & Pojar, 1991; British Columbia Ministry of Forests, 1999), and is characterized by a cool, moist climate, high rainfall, a long growing season, and a dominant western hemlock (*Tsuga heterophylla;* tree species code Hw) presence. Other common vegetation of the CWH zone includes western redcedar (*Thuja plicata*; Cw), Douglas fir (*Pseudotsuga menziesii*; Fd), and bigleaf maple (*Acer macrophyllum*; Mb). Amabilis fir (*Abies amabilis*; Ba) and yellow-cedar (*Cupressus nootkatensis*; Yc) are present in cooler, wetter regions, while red alder (*Alnus rubra*, Dr) is commonly seen in disturbed areas. Common understory surrounding Lost Lake includes red huckleberry (*Vaccinium parvifolium*), sphagnum moss (*sphagnum spp.*), sword fern (*Polystichum munitum*), bracken (*Pteridium*), foamflower (*Tiarella trifoliata*), and hardhack (*Spiraea douglassi*) (Johnson et al., 2013). The *CwSs – Skunk cabbage* site series is present near Lost Lake's south inflow, characterized by areas of nutrient-rich swamp forest with high cover of skunk cabbage.

The 5,668-hectare LSCR (formerly the “Seymour Demonstration Forest”) lies south of the upper WSA and the Seymour Falls Dam. It is mandated as a water supply area for Metro Vancouver, although for various logistical reasons (i.e., water quality, costs, and the presence of other more favourable sites) its usage as a water source would be unlikely (Dave Dunkley, personal communication). While public access was historically restricted to maintain water quality, the region – including Lost Lake – is now a popular recreation area for swimming, hiking, biking, and canoeing. Lost Lake and much of the lower valley and mid-slopes in the LSCR were logged in the 1920s and early 1930s until logging was ceased in the region in 1931. Large-scale logging in much of the watersheds, including the LSCR, began again in 1967 and continued until 1994 when concerns over drinking water quality halted operations (Walsh, 2017). Since this time, the GVRD has deactivated logging roads and placed the LSCR and surrounding water supply areas under full protection. Lost Lake was most recently logged between 1970 and 1974, resulting in the current second-growth forest characterized by an immature stand with limited understory and shrub cover (BCTWA, 2013). Visible evidence of slash burning can be seen along the edges of the lake in the form of burnt stumps.

2.2. Fieldwork

Fieldwork was conducted in Fall 2020 following all Simon Fraser University and Metro Vancouver COVID safety protocols. The Lost Lake sediments were collected in September 2020 using an inflatable zodiac boat and a temporary raft constructed of dock blocks and anchored at each corner. The raft was anchored near the northwestern edge of Lost Lake over its deepest point. A 3.64 m Livingston core was collected in six separate drives using a 5-cm diameter modified Livingstone piston corer (Wright, Mann & Glaser, 1984). Two surface cores measuring 27 cm and 41 cm were collected using a 7.6-cm diameter Glew gravity corer (Glew, 1988) and 7.6-cm diameter plexiglass tubes connected to the corer. The sediment-water interface remained intact for both surface cores during collection, but the cores were damaged after being frozen in-lab, which disturbed the positioning of the sediment-water interface and rendered both surface cores unusable.

A third 41-cm surface core was collected in November 2020 using a 7.6-cm Glew gravity corer and plexiglass tube. The core was extruded in the field at 1-cm intervals into Whirl-Pak™ sample bags using a sectioning device and stored in a refrigerator until ready for sampling. The Livingstone piston core was photographed on-site and wrapped in cellophane and tinfoil, then placed in 3-cm diameter PVC tubing for protection and storage.

2.3. Lithological Analysis

The cores were processed in the laboratory by cutting them longitudinally and separating the halves to be used for pollen and charcoal sampling. The cores were photographed, and their lithology was documented. A 2-cm layer of Mazama tephra, deposited during the eruption of Mount Mazama in southwestern Oregon ca. 7627 cal yr BP (Zdanowicz et al., 1999), was noted in the core at a depth of 207-209 cm.

To estimate the sediment mass accumulation rate and age of the core, an age-depth model was created using tie points based on ^{210}Pb and AMS-^{14}C dates as well as the accepted age of the Mazama tephra (ca. 7627 cal yr BP; Zdanowicz et al., 1999)). Eleven samples were selected from the surface core and sent to Flett Research Ltd. in Winnipeg, Manitoba, for ^{210}Pb analysis (Table 1). Additional ^{137}Cs analyses were conducted on four samples, and ^{226}Ra analyses were conducted on three samples (Table 1). Activity of ^{137}Cs was non-negligible between depths of 4 and 8 cm. The highest activity of 3.02 ± 0.6 DPM/g (disintegrations per minute per gram) was measured at the 6-7 cm interval and represents the maximum terrestrial inventory, which occurred in 1966 (Flett Research Ltd., 2020; Ritchie & McHenry, 1990). Five macrofossil samples from the Livingston core (Table 2) were sent to Beta Analytic in Marathon, Florida, for AMS-^{14}C dating. In preparation for dating, the macrofossil samples were washed with distilled water and dried in an oven overnight at 30°C.

2.4. Age-Depth Modelling

The AMS-^{14}C ages were converted to calendar years before present (cal yr BP) via the program CALIB 8.2.0 (Stuiver and Reimer, 1993) and the IntCal20 dataset (Stuiver et al., 2020). The final age-depth model uses the median ages produced by the program (Table 2). The deposition of the Mazama tephra is interpreted as a near-instantaneous event. Therefore, the 2 cm-thick ash layer at depth 204 – 206 cm has been subtracted from the total core depth for all calculations.

Table 1. ^{210}Pb age measurements, sediment mass accumulation rates, ^{137}Cs activity and ^{226}Ra activity for Lost Lake.

Site	Depth (cm)	Age (cal yr BP)	CRS Sediment Accumulation Rate (g/cm^2/yr)	^{137}Cs Activity DPM/g (dry wt.)	^{137}Cs approx. Error (DPM/g)	^{226}Ra Activity (DPM/g Dry Wt.)
Lost Lake	0.5	5.9	0.0152	—	—	—
Lost Lake	2.5	17.2	0.0134	—	—	—
Lost Lake	4.5	49.9	0.0114	3.05	0.36	—
Lost Lake	5.5	—	—	2.94	0.31	—
Lost Lake	6.5	64.4	0.0085	3.02	0.66	0.48
Lost Lake	7.5	—	—	0.95	0.32	—
Lost Lake	8.5	71.0	0.0143	—	—	—
Lost Lake	10.5	73.6	0.0407	—	—	—
Lost Lake	12.5	77.0	0.0733	—	—	0.45
Lost Lake	18.5	86.3	0.0488	—	—	—
Lost Lake	24.5	102.3	0.0353	—	—	—
Lost Lake	30.5	122.6	0.0336	—	—	—
Lost Lake	39.5	134.7	0.0410	—	—	0.34

Table 2. AMS ^{14}C ages for Lost Lake.

Site	Beta Laboratory Number	Drive Number	Total Core Depth (cm)	Drive Depth (cm)	Material	Conventional Radiocarbon Age (BP)	Calib 8.20 Age (cal yr BP) (2σ)
Lost Lake	598873	1	61.5	20.5	Plant	1840 ± 30	1763[c] (1699 – 1826)
Lost Lake	—	3	205	33	Mazama tephra	—	7627 ± 150[a]

Site	Beta Laboratory Number	Drive Number	Total Core Depth (cm)	Drive Depth (cm)	Material	Conventional Radiocarbon Age (BP)	Calib 8.20 Age (cal yr BP) (2σ)
Lost Lake	612302	3	229.5	52.5	Plant	9100 ± 30	10248 [c] (10197 – 10298)
Lost Lake	612303	5	284.5	7.5	Plant	9980 ± 40	11445 [c] (11265 – 11625)
Lost Lake	598874	5	326.5	48.5	Plant	11620 ± 30	13465 [c] (13409 – 13520)
Lost Lake	598875	6	386	43	Plant	10290 ± 30	11993 [b,c] (12107 – 11878)

[a] Zdanowicz et al., 1999.
[b] Date not used in final age model.
[c] Indicates midpoint.

The age model for the Lost Lake surface core was created based on a constant rate of supply (CRS) model (Appleby & Oldfield, 1977) applied to eleven ^{210}Pb age determinations. The age model for the composite core (surface core + piston core) was constructed using the Bacon modelling program in R (Blaauw & Christen, 2011), which uses Bayesian analysis to reconstruct historical accumulation rates by combining contemporary dating with prior information. Bayesian models simulate variable sediment deposition rates and can thus produce a more complex and realistic age-depth relationship when compared with classic linear interpolation models (Wang et al., 2019). The ^{210}Pb CRS model was locked at the base of the Bacon model, and the calibrated calendar ages of four AMS-^{14}C ages plus the date of the Mazama tephra were used to construct a representative age-depth model for the composite core (Fig. 4).

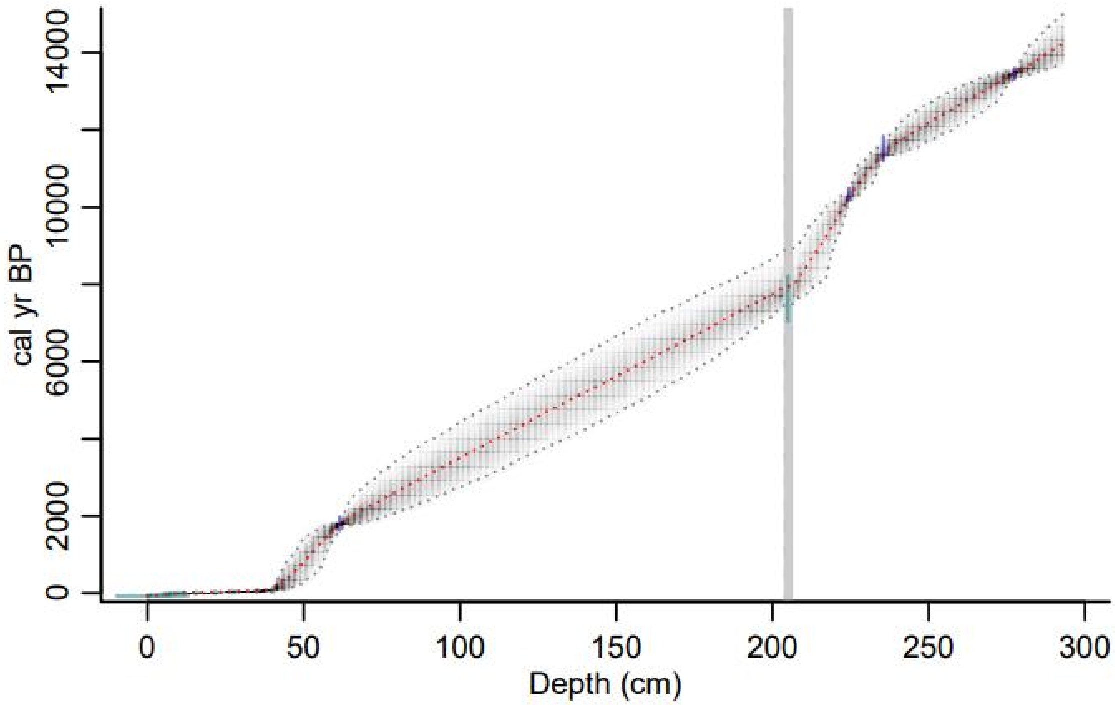


Figure 4. Bacon age model of the Lost Lake Core.
Grey dotted lines represent error envelope at 95% confidence interval. Grey fill represents all likely age-depth models calculated by Bacon. Red line represents mean age (the "best" model selected by Bacon). Blue points represent AMS-^{14}C samples (2σ probability distributions of calibrated ^{14}C ages). Green points represent locked in ^{210}Pb dates and Mazama ash date. Gray vertical line represents slump point at Mazama tephra to indicate instantaneous deposition.

A date inversion was observed between the AMS-^{14}C dated depths of 326.5 cm in Drive 5 (13,408 – 13,519 cal yr BP) and 386 cm in Drive 6 (11,923 – 12,106 cal yr BP). A light-gray basal clay layer is present in both drives (Appendix A), suggesting that Drive 6 is likely re-cored material, thus explaining the inverted radiocarbon date.

Based on the AMS-^{14}C dates at the end of Drive 3 and beginning of Drive 5 (Table 2) and comparisons of charcoal concentrations and pollen percentages between the fourth and fifth drives of the core (Appendix B), I concluded that drive 4 (228-272 cm) is also very likely re-cored material. The proximity in age between the two radiocarbon dates in Drive 3 (10,248 cal yr BP) and Drive 5 (11,445 cal yr BP) reflect the possibility of sediment loss between drives, and as such there is uncertainty in the ages and sedimentation rates for the 16 cm of sediment between the two dated points (See Appendix B, Figure B.5). For the purposes of this thesis, I will be analysing with the assumption that no significant sediment loss has occurred between Drives 3 and 5.

The mean sample resolution for the composite core was 48.8 yr/cm, with a much higher resolution (5.3 yr/cm) in the first 41 cm of the core. The mean sedimentation rate for the composite core was 0.086 cm/yr, with a maximum of 1.2 cm/yr between 12 and 15 cm of the composite core.

2.5. Pollen Analysis

2.5.1. Pollen Extraction

The composite core was sub-sectioned for pollen analysis at 10 cm intervals along the entire length of the core. Pollen isolation methods were adapted from standard laboratory techniques (Faegri and Iverson, 1989; Moore et. al.,1991). The key steps of pollen isolation are deflocculation, removal of extraneous matter, staining, and mounting. A 1-cm^3 volume of sediment was extracted at each 10-cm interval, and a *Lycopodium* (clubmoss) marker tablet (10,679 ± 191 spores/tablet; Batch No. 938934) was added to each sample as a control. A solution of 10% HCl (hydrochloric acid) was added to the samples to remove $CaCO_3$, and then organic material was removed with a 20-minute hot water bath in 6% KOH (potassium hydroxide). Ten ml of HF (hydrofluoric acid) were added to dissolve silicate mineral matter and the samples were cooked in a hot water bath for 10 minutes. Subsequently, five ml of 95% ethanol were added before centrifuging and decanting. The samples were mixed with 10 ml of 10% HCl and cooked in a hot water bath for an additional 10 minutes to increase acidity, then rinsed with distilled H_2O.

Subsequently, 20-25 ml of glacial acetic acid were added to dehydrate the samples in preparation for acetolysis, which removes all material except sporopollenin (which composes the outer pollen walls) from the samples. Acetolysis was performed using a 9:1 mixture of acetic anhydrite:sulfuric acid. Five ml of the mixture were added to each sample, and they were cooked in a hot water bath for 8-9 minutes with frequent stirring. After centrifuging and decanting the acetolysis mixture, glacial acetic acid was added to the samples, centrifuged, and decanted.

The samples were gradually dehydrated in three steps using progressively increasing concentrations of ethanol. Approximately 3 ml of tertiary butanol was added to each sample and left to settle, after which excess tertiary butanol was removed with a pipette. The samples were mixed with silicon oil and left to dry at room temperature for 24 hours before being placed on glass microscope slides for analysis.

A minimum of 500 grains per sample were counted using a Carl Zeiss binocular stereomicroscope at 400x to 1000x magnification. Taxa were identified using reference slides from the SFU pollen library (Dr. Rolf Mathewes, personal communication), published morphological keys (Faegri and Iverson, 1989; Moore et. al., 1991) (Fig. 5), and The Global Pollen Project webpage (globalpollenproject.org). Taxa were identified to the species level when possible. However, when no distinguishing features were visible (e.g., Rosaceae), they were identified to genus or family. Pine pollen were separated into haploxylon (white pine), diploxylon (yellow pine) and undifferentiated grains when possible by identifying a minimum of 50% of the pine grains from each sample and extrapolating a percentage to the total pine count.

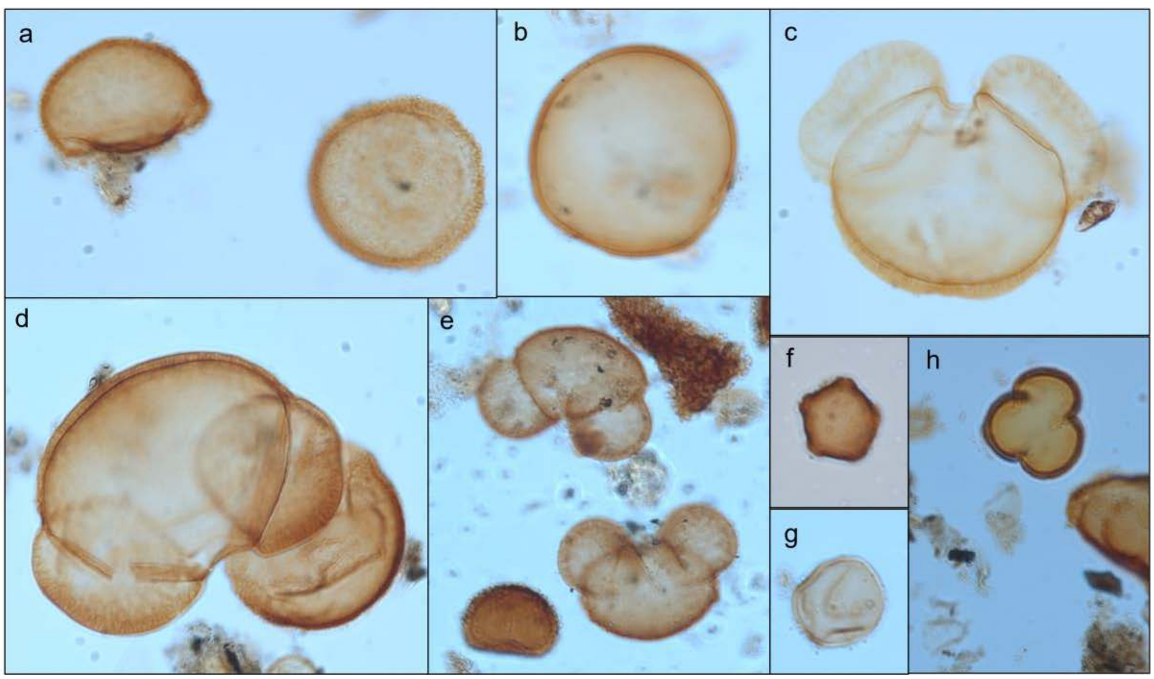


Figure 5. Pollen and spore grains from the Lost Lake core.
Pollen grains are identified as: (a) *Tsuga heterophylla* (western hemlock); (b) *Pseudotsuga menziesii* (Douglas fir); (c) *Abies* (true fir); (d) *Picea* (spruce) in foreground and *T. heterophylla* behind; (e) *Pinus* (diploxylon) (vesicular grains) and *Polypodium* (polypody fern); (f); *Alnus* (alder); (g) *Cupressaceae* (cedar); (h) *Acer* (Maple, top center) and folded *T. heterophylla*. All grains identified using keys from Faegri and Iverson (1989), Moore et. al. (1991), and type slides from the collection of Dr. Rolf Mathewes.

2.5.2. Pollen Interpretation

For the purposes of this study, I have assumed yellow pine to be *Pinus contorta* (Lodgepole pine) and white pine to be *Pinus monticola* (Western white pine). Lodgepole pine is presently the most widely distributed pine species in western Canada (Klinka et al., 1999). Macrofossils from nearby Marion and Surprise Lakes indicate that *P. contorta* and *P. monticola* are the representative species in their records (Mathewes, 1973; Mathewes & Wainman, 1987). Specifically, the high abundance of *P. contorta* macrofossils soon after deglaciation at Marion Lake supports the assertion that *P. contorta* was also dominant at Lost Lake during this time. Studies on southern Vancouver Island (Brown and Hebda, 2003; Brown et al., 2008) have also found *P. contorta* to be the dominant pine in early Holocene cores. *P. ponderosa* (Ponderosa pine) is another species of pine common in British Columbia, but it favors warm, dry conditions (i.e., those found in the interior of the province) and most likely did not grow in my study area. Thus, it has been eliminated as a possible source of *Pinus* pollen in this study.

Pseudotsuga and *Larix* pollen are very similar in morphology and I did not try to distinguish between them in this study. Given the present range and characteristics of each species (Klinka et al., 1999), I expect that *Pseudotsuga* is far more likely to be found in the Lost Lake core than alpine larch (*Larix lyallii*), which is found at higher elevations. Western larch (*Larix occidentalis*) does not presently grow naturally in the Coast Mountains (Hallett et al., 2003), although its range could have expanded westward during periods when the climate was warmer and drier. Given the current abundance of *P. menziesii* in coastal B.C. and the conflicting ranges of *Larix* spp., I have chosen to consider *Pseudotsuga/Larix* pollen as *P. menziesii* in the Lost Lake record.

Pollen and spore microfossil diagrams were created using the program TILIA version 2.6.1 (Grimm, 2004). The basic pollen sum used for percentage calculation includes all terrestrial pollen, while the sum used for spore percentage calculation includes all terrestrial pollen and all spores. The resulting percentages were used to plot a pollen percentage diagram (Fig. 6). Pollen concentrations (P_{conc}, pollen grains/cm^3) were calculated in TILIA for each sample by dividing the number of pollen grains counted per 1-cm^3 volume (ΣP) by the number of control grains counted and then multiplied by the total number of control grains added (Eqn 1):

$$P_{conc} = \frac{\Sigma\ P}{Marker\ spores\ counted} \times Total\ marker\ spores\ added \qquad (1)$$

The relative pollen abundances were then plotted against the age model to show pollen changes over time.

The Lost Lake core was separated into six subsections based on pollen assemblage zones, which were identified using a stratigraphically constrained cluster analysis conducted using incremental sum of squares (CONISS; Grimm, 1987) within the Tilia program. Samples were square root transformed prior to analysis, then clustered based on similarities to adjacent samples. The resulting zones were selected based on prior knowledge of historical climate and ecological changes combined with visual assessment of changes along the core's pollen assemblage. Square root transformed data are used for CONISS analysis to account for the skew of large versus small relative abundances between pollen types. The impact of overrepresented species (i.e., *Alnus* and *Pinus*, which release large amounts of pollen) are reduced while the impact of underrepresented or less abundant species are increased (Pellatt et al., 1997; Pellatt, Mathewes & Clague, 2001).

Pollen percent diagrams plot the percentage of each taxon relative to all other terrestrial pollen. The resulting data are interdependent, as changes in each record are proportional to other pollen types. A common secondary measure of changes in pollen over time is pollen accumulation rates (PAR; grains $cm^{-2}\ yr^{-1}$) which provide an estimate of the abundance of each taxon independently (Birks and Birks, 1980; Faegri et al., 1989):

$$PAR = P_{conc} \times v \qquad (2)$$

Where P_{conc} is estimated in Eqn (1), and the sediment accumulation rate (*v*; cm/year) was calculated using the Bacon age model. PARs were described along the length of the entire core rather than in subsections because large scale shifts were not as apparent in the PAR record.

2.6. Charcoal Analysis

Analysis of macroscopic charcoal followed methods modified from Hallett et al. (2003), Whitlock et al. (2008) and Murphy et al. (2019). Experiments of Murphy et. al. (2019) provide a process for charcoal extraction that most successfully produces a reliable representation of the charcoal record. First, a 1-cm^3 volume of material was taken at contiguous 1 cm intervals along the length of the composite core. Samples were then mixed with 20 ml of 5% $Na(PO_3)_6$ (sodium hexametaphosphate) and soaked for 24 hours to facilitate disaggregation. The samples were then soaked for 1 hour in a 6% H_2O_2 (hydrogen peroxide) solution to lighten any non-charcoal organic material. Samples were gently washed through a 125-µm sieve, backwashed into a petri dish and dried for 24 hours at 30-40 ºC.

The samples were placed on a 1-cm white gridded counting coaster, and charcoal particles were counted under a Leica® M205C stereomicroscope. Charcoal particles were identified based on the characteristics defined by Clark & Royall (1995), including reflectance, brittleness, evidence of cell structure, and color.

To obtain charcoal concentration values (particles/cm^3), the number of particles per sample were divided by the sample volume (Whitlock et al., 2008). My high-resolution charcoal analysis used the free software package CharAnalysis (Higuera et al., 2008) to measure (1) charcoal influx (CHAR, particles/cm^2/year); (2) background charcoal (particles/cm^2/year); (3) fire episode (peak) magnitude (charcoal pieces/cm^2/peak); (4) fire event frequency (fires/1000yrs); and (5) mean fire return interval (mFRI, measured in years) (Long et al., 2008; Whitlock et al., 2008; Murphy et al., 2019).

CHAR is frequently used as a reliable measure of local fire history (Brown & Hebda, 2002a,b; Gavin et al., 2003; Whitlock and Larsen, 2001; Gavin et al., 2007; Derr, 2014; Murphy et al., 2019; Rodengen et al., 2022). The general procedure involves decomposing CHAR into background and peak components to differentiate local fires from background noise (Clark et al.,1996; Long et al.,1998). The background component represents long-term (centennial-scale) changes in the overall magnitude of CHAR caused by fluctuation of variables including vegetation, fire behavior, area burned, and charcoal deposition to the lake centre. The peak component represents fires that occurred near the sample location, with high values indicating local fires and low values representing distant fires, noise in the data, and charcoal redeposition, either within the lake or within the watershed (Clark et al., 1996; Gavin et al., 2006).

The median temporal resolution of the Lost Lake composite core was calculated in CharAnalysis to be 44 years. I interpolated charcoal counts, sample volume, and sample depths to this resolution to create evenly spaced time intervals for sediment accumulation rates and charcoal concentrations. The resulting interpolated sediment accumulation rates and interpolated charcoal concentrations were multiplied to calculate CHAR along the length of the core and for each individual zone (i.e., LL-1 to LL-6b).

Low-frequency background CHAR ($C_{background}$) was estimated using a LOWESS smoother robust to outliers with a smoothing window of 900 years, which was determined to be the best fit via a sensitivity analysis run in CharAnalysis. The sensitivity analysis maximizes the sum of the median signal-to-noise index (SNI) and goodness-of-fit (GOF) at 900 years for Lost Lake. SNI is a ratio of the separation between fire-related peaks and non-fire related peaks. GOF is a measure of how well the noise of the peaks component fits with the noise distribution model.

The high-frequency CHAR (C_{peak}) component of the record was calculated as the positive residuals resulting when background CHAR was subtracted from interpolated CHAR ($C_{peak}=C_{interpolated} – C_{background}$). To separate fire-related peaks from non-fire related peaks, a Gaussian mixture model was used to define the noise distribution, with threshold values limited to the locally defined 99th percentile of the noise distribution. The cut-off probability for minimum counts was set to 0.05, indicating that the minimum charcoal count within 75 years before a given peak was required to have a less than 5% chance of coming from the same Poisson distribution as the maximum count associated with said peak, otherwise the peak was removed (Higuera et al., 2008). To determine whether samples came from the same probability distribution, a non-parametric Kolmogorov-Smirnov goodness-of-fit test was used (Higuera et al., 2008).

Fire frequency and mean fire return interval were estimated by smoothing the fire-related peaks component (C_{peak}) using a 1000-year moving smoothing window on the fire-related peaks component of CHAR. The mFRI was calculated by averaging time periods between fire episodes along the entire length of the composite core (Agee, 1993). The mFRI and fire frequencies were also calculated for each of the seven vegetation zones delineated in TILIA using the same methods.

Peak magnitude was estimated for the Lost Lake record by taking the sum of all charcoal counts within a given peak that exceed the previously determined CHAR threshold value. Peak magnitude can be related to fire size and severity, and taphonomic processes post-fire (Higuera et al., 2007).

Chapter 3. Results

3.1. Lithology

Most of the Lost Lake composite core is composed of gyttja, a fine-grained organic-rich sediment. Four lithological units were identified along the 2.87 m length of the core. From 0 to 204 cm, the sediment is an undifferentiated dark brown gyttja containing few macrofossils. Sediment for the depth interval of 204 – 206 cm consists of a light brown tephra from the Mount Mazama eruption. From 206 to 260 cm, the sediment is again composed dominantly of dark brown gyttja. A mottled, inconsistent layer of presumably displaced Mazama tephra is present from 228 – 233 cm. A five-cm layer of lighter brown gyttja intermingled with small (1 – 5 mm) angular clay inclusions is observed from 260 to 269 cm and underlain by more undifferentiated dark brown gyttja from 269 to 281 cm. The deepest section of the core, from 281 cm to its termination at 287 cm, is composed of a light gray clay. A basal clay layer is often found at the bottom of glacial lakes due to harsh scouring as glaciers moved over the landscape. The clay found at the bottom of the Lost Lake core is interpreted to represent this basal layer. The final seven cm of the core (287 – 294 cm) is again composed of gyttja. I determined it highly unlikely that an organic-rich sediment layer lay directly beneath the basal clay, and the final 7 cm is interpreted as redistributed younger sediment caused by shifting during extraction of the Livingston corer (Appendix B, figure B.6). After removal of the presumed re-cored sediments in Drive 4, Drive 6, and the final 7 cm of Drive 5, the total composite core length was 287 cm.

3.2. Lost Lake Pollen Record

3.2.1. Changes in pollen assemblages

Six pollen assemblage zones were identified in the Lost Lake composite core (LL-1 to LL-6, Fig. 6) based on a constrained cluster analysis using CONISS total sum of squares (Grimm, 1993). Pollen sums and percentages were calculated using total terrestrial pollen, while spore sums and percentages were calculated using total pollen and spores.

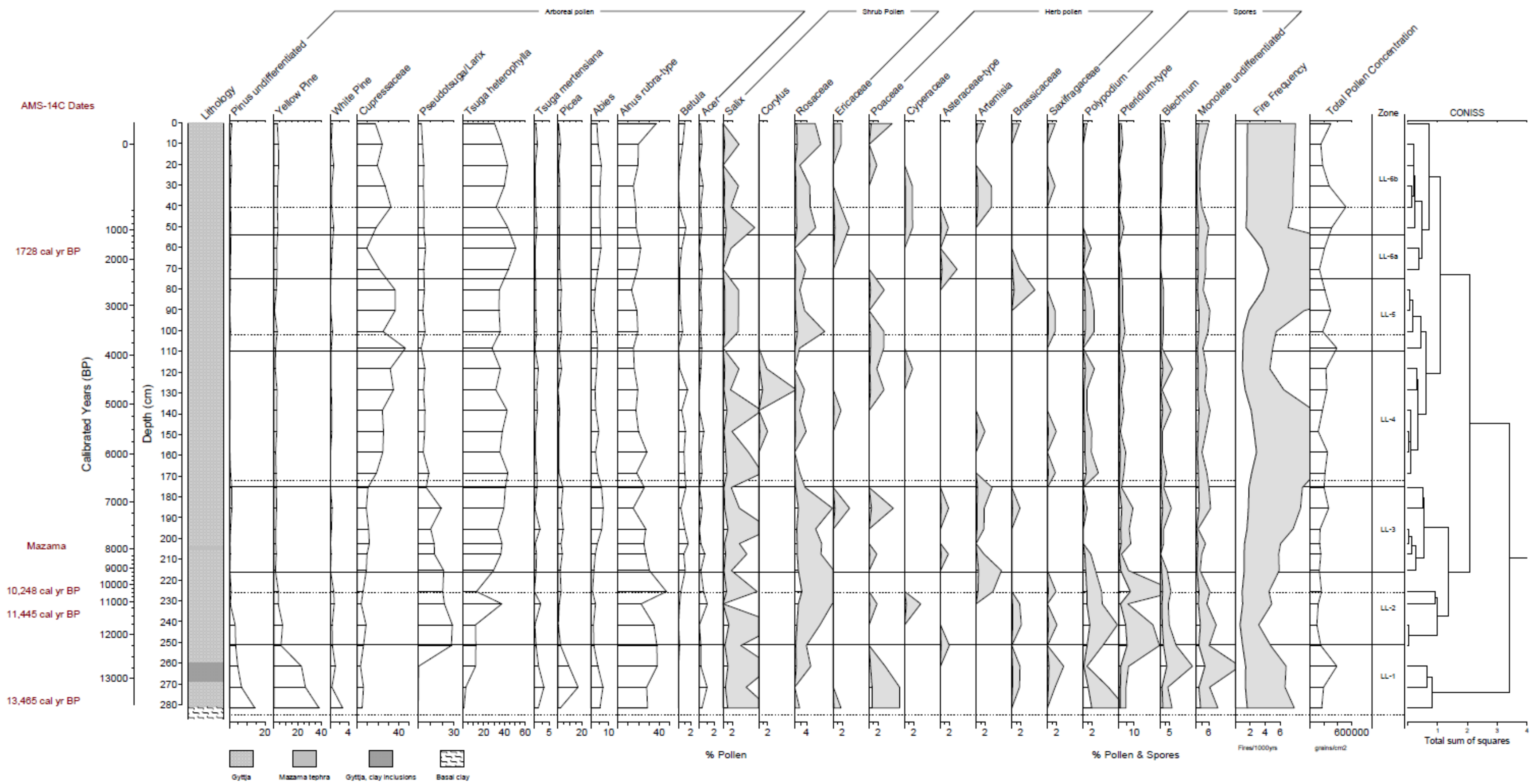


Figure 6. Pollen percent diagram of Lost Lake.
Radiocarbon dates and lithology are shown on the left, fire frequency and total terrestrial pollen concentration are shown on the right. Tenfold exaggeration curves (light grey) are shown to highlight abundance of infrequent pollen and spores. Zones were delineated using constrained cluster analysis (CONISS). Solid horizontal lines represent zonal breaks. Dotted horizontal lines represent core breaks.

Zone LL-1 (287 - 251 cm; >12,229 cal yr BP)

The oldest and deepest zone was dominated by decreasing *Pinus*. *Pinus* reached its highest pollen percentage (>50%) at the beginning of the zone (13,645 cal yr BP); *Pinus contorta* (yellow pine) was the dominant species during this time, composing 69% of the pine assemblage by approximately 13,600 cal yr BP. *Pinus* steadily decreased to 9% of the pollen assemblage by the end of Zone LL-1. *Tsuga mertensiana* and *Abies* (likely *Abies amabilis*) remained below 3% and 6% respectively for the duration of Zone LL-1. *Alnus rubra* pollen was common throughout this zone (27-38%), while *Cupressaceae* was present in lesser amounts (3-6%). *Pseudotsuga/Larix* pollen abundance remained at <1% from the base of the core until 12,700 cal yr BP, before quickly increasing to its maximum value (~28%) during the transition between LL-2 and LL-1 (ca. 12, 229 cal yr BP). *Picea* (inferred to be *Picea sitchensis*) pollen gradually increased to its maximum percentage of 16% at 13,160 cal yr BP, then steadily decreased to 2% by the end of Zone LL-1. *Tsuga heterophylla* abundance increased from <1% during the oldest part of the core to 12% at 12,650 cal yr BP.

Zone LL-2 (251 – 216 cm; 12,229 – 9,085 cal yr BP)

In Zone LL-2, *Tsuga heterophylla* increased from 12% at the beginning of the zone to ~29% by 9,100 cal yr BP and was accompanied by a corresponding decrease in *Alnus* from 37% to 30%. *Pinus* pollen continued to decline to ~1.5% of the assemblage by the end of the zone at 9,085 cal yr BP. *Tsuga mertensiana* percentages remained low at 0.5-1% of the assemblage. *Abies* and *Picea* pollen abundances remained steady but low between 0.5 and 2%, respectively.

Zone LL-3 (216 – 175 cm; 9,085 – 6,680 cal yr BP)

In Zone LL-3, *Pinus*, *Alnus* and *Picea* percentages stabilized. Pinus pollen percentages remained below 3.5% for the duration of the zone. Additionally, *Tsuga heterophylla* increased from 28% to 40%, and *Cupressaceae* increased from 3-5% at approximately 12,229 cal yr BP to 10% by the end of the zone LL-3 (6680 cal yr BP). *Pseudotsuga/Larix* decreased substantially from 21% at the beginning of zone LL-2 to 7% at the conclusion of this zone, while *Abies* fluctuated between 2 and 6.5%. *Picea* similarly fluctuated between 2 and 4%. *Tsuga mertensiana* remained low (<2%). Shrub and herb pollen species experienced their highest percentages during Zone LL-3, with *Rosaceae, Salix,* and *Artemisia* reaching 1.3%, 1% and 0.55%, respectively.

Zone LL-4 (175 – 110 cm; 6,680 – 3,940 cal yr BP)

Tsuga heterophylla abundances decreased from 40% at 6680 cal yr BP (the conclusion of Zone LL-3) and remained between 28 and 34% throughout most of Zone LL-4. *Cupressaceae* increased from 10% at the beginning of Zone LL-4 to nearly 40% by 3940 cal yr BP. *Alnus* pollen remained a dominant contributor to the pollen assemblage and fluctuated from 15-20%. In contrast, *Pseudotsuga/Larix* declined from 7-9% to <5% by the conclusion of Zone LL-4. *Pinus*, still dominated by *P. contorta*, remained below 3.5% of the pollen assemblage.

Zone LL-5 (110 – 75 cm; 3,940 – 2,417 cal yr BP)

Tsuga heterophylla remained unchanged for most of this zone, increasing to ~39% by ca. 2400 cal yr BP. *Cupressaceae* declined briefly to 24% before again rising to >30%, while *Alnus* continued to fluctuate between 13 and 19%. *Pinus* remained a small part of the assemblage in Zone LL-5, fluctuating between 0.5 and 4%, while *Pseudotsuga/Larix* remained between 2 and 5%.

Subzone LL-6a (75 – 54 cm; 2,417 – 1,173 cal yr BP)

Tsuga heterophylla reached its maximum contribution of 50% of the assemblage at 1,700 cal yr BP, at the expense of *Cupressaceae*, which decreased from above 30% to 9% during this time period. *Alnus* abundances remained largely unchanged, contributing between 17 and 22% to the pollen assemblage. *Pinus* percentage steadily increased from 2.5 to 4% by 1,173 cal yr BP, and *P. contorta* still comprised most of the *Pinus* assemblage (3%). *Pseudotsuga/Larix* contributed 4-6% of the pollen assemblage.

Subzone LL-6b (54 – 0 cm; 1,173 – -71 cal yr BP)

Alnus abundance remained unchanged until increasing to 36% at ca. -35 cal yr BP. *Cupressaceae* sharply increased to 32% at ca. 140 cal yr BP at the expense of *Tsuga heterophylla*, which decreased to 32% before climbing back to a peak of 42% at approximately 20 cal yr BP (1930 AD). *Pinus* abundance remained between 3 and 6%, while *Pseudotsuga/Larix* remained between 2.5 and 5%. *Tsuga heterophylla* and *Picea* pollen abundances both displayed a minor increase at approximately 1950 AD but remained below 3%. *Abies* decreased slightly to 3.4% by the end of the zone.

3.2.2. Pollen Accumulation Rate Diagram

The PARs for the Lost Lake core (Fig. 7) provide a different representation of historical ecological changes than the pollen percentage diagrams. To maintain a relationship with the pollen percentage diagram in Section 3.2.1, the same zones (Zone LL-1 to LL-6) are described from the base to the top of the core (i.e., oldest to youngest). The PAR patterns of *Pinus*, *Tsuga heterophylla, Abies* and *Alnus rubra* were substantially different from the pollen percentage diagram, whereas the PARs of *Cupressaceae, Picea, Betula* and *Tsuga mertensiana* followed the general patterns of their pollen percent diagrams. In all arboreal species and the majority of herb and shrub species PARs appear to emphasize rapid increases in influx in the final 200 years of the record. This pattern was not observed in the pollen percentage records of most species, but the percentages of *Alnus rubra* and *Poaceae* display moderate increases that follow the trend of their PARs.

Pinus PAR decreased from 1780 to 290 grains/cm^2/year in Zone LL-1 between approximately 13,600 and 12,229 cal yr BP, although the change is not as pronounced as in the pollen percentage diagram. Subsequently *Pinus* accumulation rates remained low throughout Zones LL-2 to LL-6b before increasing rapidly at 140 cal yr BP, during Zone LL-6a. This pattern differs from what is observed in the *Pinus* percentages, which did not increase as substantially in the final 200 years of the record.

Alternatively, while *Tsuga heterophylla* percentages rose to approximately 40% during LL-1 and LL-3 (13,600 – 6680 cal yr BP), the PARs remained between 1000 – 2500 grains/cm^2/yr until 45 cal yr BP (during Zone LL-6b) when they abruptly increased to nearly 50,000 grains/cm^2/yr. Pseudotsuga/Larix similarly displayed low PARs (0 to 1000 grains/cm^2/year) between Zones LL-1 and LL-5 (13,600 to 1173 cal yr BP) and then increased to 4500 grains/cm^2/yr at approximately 45 cal yr BP during Zone LL-6b.

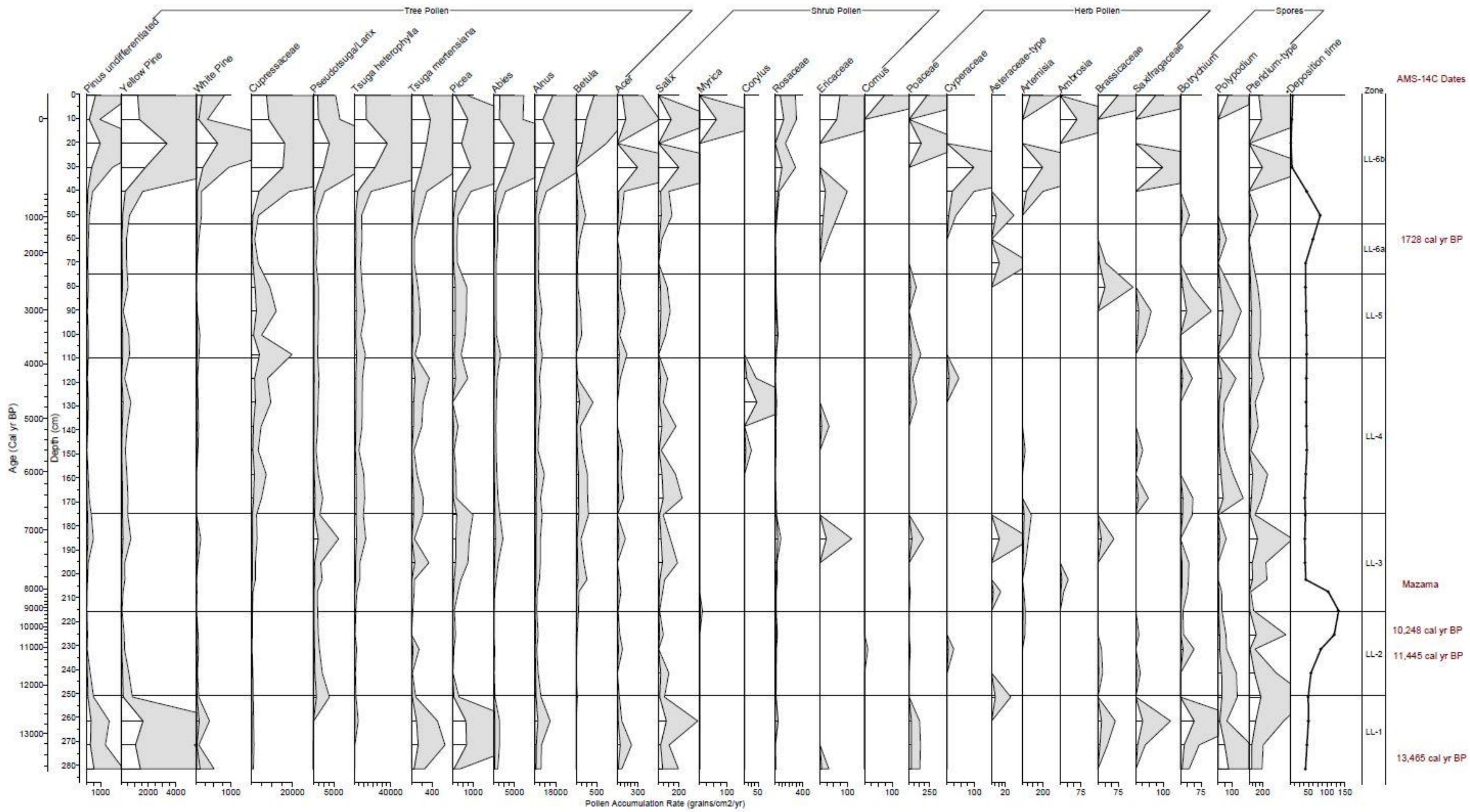


Figure 7. Pollen influx diagram of Lost Lake.

Radiocarbon dates are shown on the right. Fivefold exaggeration curves (light grey) are shown to highlight PAR curves. Solid horizontal lines represent zonal breaks.

Cupressaceae PARs did not fluctuate as much as the pollen percentages, but PAR increased to 4300 grains/cm^2/year at 3,850 cal yr BP during Zone LL-5. *Cupressaceae* PARs increased again at 45 cal yr BP (during subzone LL-6b) to approximately 27,000 grains/cm^2/year. Changes in *Alnus* PAR are similar to *Cupressaceae* and remained low until 45 cal yr BP (during Zone LL-6b) when PAR increased to approximately 15,300 grains/cm^2/year, following the pattern of *Alnus* percentage. Minor but noticeable increases in *Alnus* accumulation rate also occurred during Zones LL-1 (12,740 cal yr BP) and LL-4 (5790 cal yr BP), which align with increases in the pollen percentage diagram.

Pseudotsuga/Larix PARs generally remain low (below 500 grains/cm^2/year) until Zone LL-6b, differing from the percentage record which shows high *Pseudotsuga/Larix* percentage (10-27%) between 12,300 and 7130 cal yr BP. At approximately 45 cal yr BP *Pseudotsuga/Larix* PARs increase to 4500 grains/cm^2/year and remained elevated for the remainder of the record.

3.3. Lost Lake Charcoal Record

3.3.1. Composite Charcoal Record

CharAnalysis detected 23 significant fire episodes during the 13,900 years of the Lost Lake charcoal record (Fig. 8; Table 4). The current time since last fire (TSLF) is 144 years, with the most recent detected fire event occurring in 1845 AD. The longest interval of no recorded fire activity was approximately 1230 years between 12,469 and 11,237 cal yr BP in Zone LL-2, while the mean fire-free interval is 555 years. The mean fire return interval (mFRI) for the composite core is 598 years with a natural range of variability of 466 - 735 years.

The mean smoothed fire frequency along the length of the core is 1.7 fires/1000yrs. The lowest frequency (0.7 fires/1000yrs) occurred between 12,117 and 11,501 cal yr BP. Fire frequency reached a maximum of 4.5 fires/1000yrs at 2305 cal yr BP in Zone LL-6a.

The mean CHAR for the Lost Lake composite record, interpolated to 44-year sample intervals, is 14 pieces/cm^2/year. The period of highest CHAR occurred from -71 cal yr BP to 61 cal yr BP during Zone LL-6b (Fig. 8), during which the mean CHAR was

234 pieces/cm^2/year. Other periods of high CHAR occurred between ca. 7190 cal yr BP and 2660 cal yr BP during Zones LL-3 to LL-6a (mean CHAR = 16 pieces/cm^2/year, with intermittent peaks), between ca. 12,470 cal yr BP and 12,290 cal yr BP during Zone LL-1 (mean CHAR = 25 pieces/cm^2/year), and between ca. 13,260 cal yr BP and 13,170 cal yr BP during Zone LL-1 (mean CHAR = 20 pieces/cm^2/year).

The background component can be broadly separated into three sections. The first section occurred between ca. 13,940 to 7320 cal yr BP during Zones LL-1 to LL-3 and had a mean background of 8 pieces/cm^2/year. The second section, from 7320 to 2390 cal yr BP during Zones LL-3 to LL-6a, had a mean background of 15 pieces/cm^2/year. The third section, from 2390 to -71 cal yr BP during Zones LL-6a to LL-6b, had a mean background of 8 pieces/cm^2/year. The mean background component for the entire Lost Lake composite core is 10 pieces/cm^2/year.

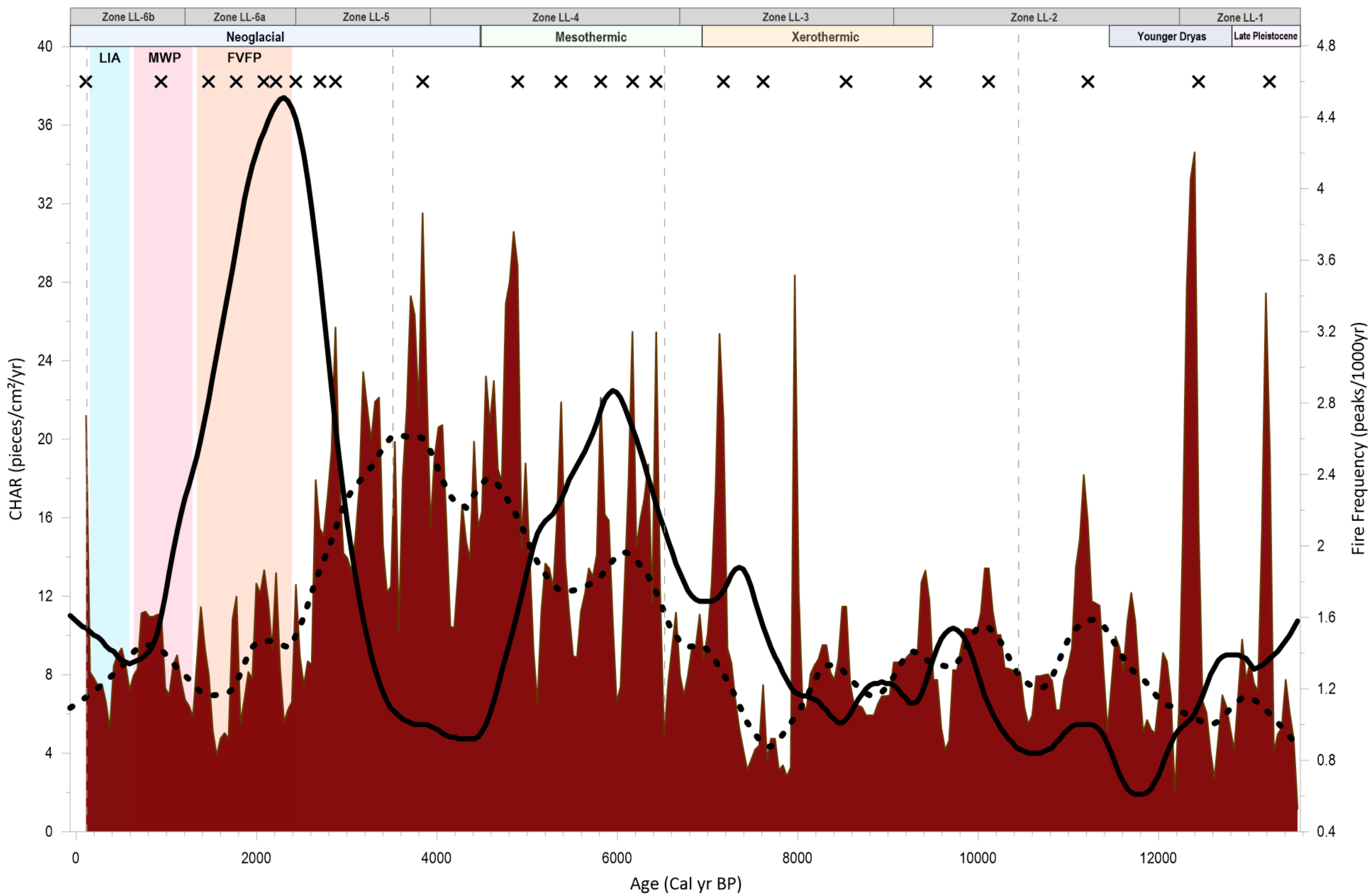

Zone LL-6b
Zone LL-6a
Zone LL-5
Zone LL-4
Zone LL-3
Zone LL-2
Zone LL-1
Neoglacial
Mesothermic
Xerothermic
Younger Dryas
Late Pleistocene
LIA
MWP
FVFP
CHAR (pieces/cm²/yr)
Fire Frequency (peaks/1000yr)
Age (Cal yr BP)
0
2000
4000
6000
8000
10000
12000
0
4
8
12
16
20
24
28
32
36
40
0.4
0.8
1.2
1.6
2
2.4
2.8
3.2
3.6
4
4.4
4.8

Figure 8. Synthesis of Lost Lake charcoal-inferred fire activity and major anthropogenic and climatic changes in southwestern British Columbia since the Late Glacial.

Red fill represents interpolated CHAR; black solid line represents fire frequency; dotted black line represents interpolated background CHAR; X symbols represent fire events; Vegetation Zones LL-1 – LL-6b and major climatic intervals are shown at the top of the figure. Vertical colored bars represent the FVFP (orange), MWP (pink) and LIA (blue). Dotted grey vertical lines represent breaks in the Lost Lake core.

Interpretation of these results are dependent on estimation of the Signal to Noise Index (SNI), a parameter often used to distinguish fire incidents in the charcoal record (Higuera et al., 2010). The SNI quantifies the division of the record into fire event peaks (signal) and background variability (noise). For a charcoal record to be confidently used in fire reconstruction it should have an SNI > 3; otherwise, distinct peaks may be over-interpreted (Kelly, Higuera, Barrett & Hu, 2011). The global SNI for the Lost Lake charcoal record is 3.61, comfortably meeting the minimum standard. The local SNI for Lost Lake also meets this SNI threshold (median local SNI = 3.06; mean local SNI = 9.00); however, six intermittent sections within the core do not meet the minimum threshold (Table 3). These sections may indicate periods of weak peak detection during which the signal is not well distinguished from the noise, and consequently fire episode characteristics should be cautiously inferred in these sections (Kelly et al., 2011).

Table 3. Sections in Lost Lake core below minimum SNI threshold.

Time Period (cal yr BP)	Mean Local SNI
853 – 985	2.11
1,909 – 2,305	2.85
3,054 – 4,417	2.68
5,693 – 6,397	2.35
9,345 – 10,533	0.67
11,193 – 11,677	2.51

Similar to estimates of PAR, CHAR measurements are dependent on sediment accumulation rates. For this reason, the PAR and CHAR results of the composite core were graphed side-by-side to allow for visualization of similarities between charcoal and pollen accumulation rates (Figure 9). The results exhibit similarities in accumulation rate along the length of the composite core, indicating that the major control on charcoal and pollen deposition was likely watershed-scale sedimentation processes. Minor differences in accumulation rates between the two proxies, such as at ca. 8000 cal yr BP and ca.

13,000 cal yr BP, may be due to the higher resolution of the charcoal samples (~44 years/sample) in comparison to those of pollen (~450 years/sample) providing a more sensitive record of CHAR fluctuations. It could also be caused by natural processes that affect charcoal deposition, such as post-fire debris flows (Brown et al., 2019).

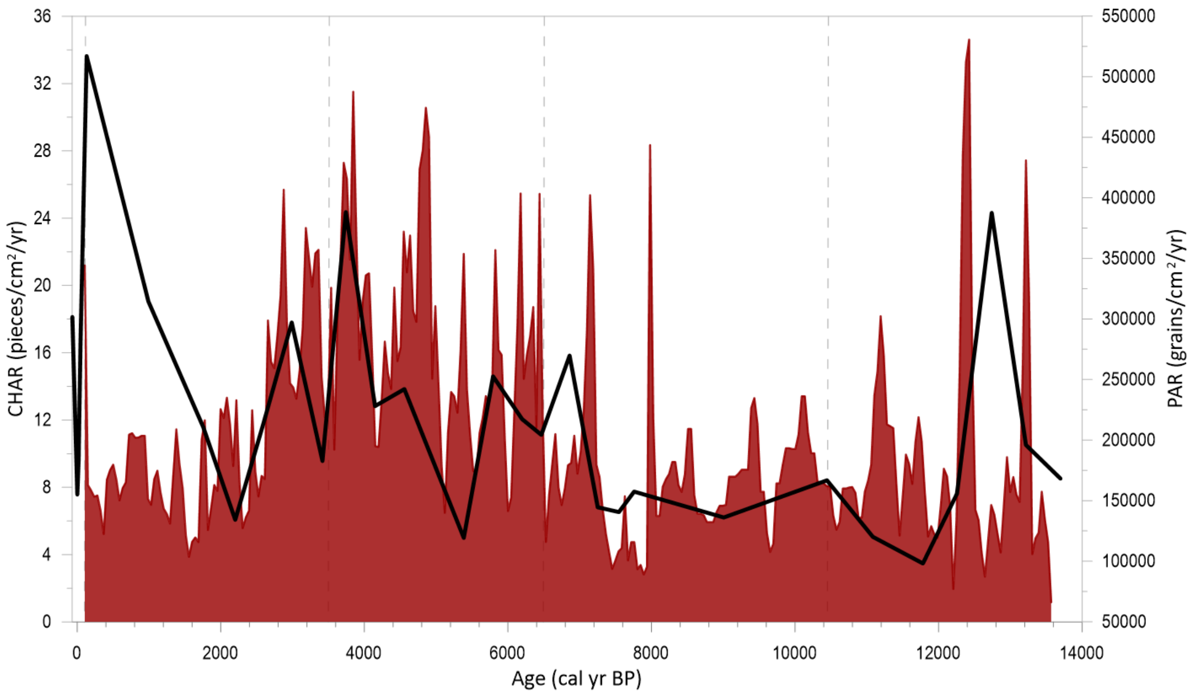


Figure 9. CHAR (red) and total PAR (black) of the Lost Lake composite core. Dotted grey vertical lines represent breaks in the Lost Lake core. Total PAR is all terrestrial pollen and spores. Initial 100 years of the CHAR record have been removed for ease of comparison between CHAR and PAR.

3.3.2. Zonal Charcoal Record

The charcoal record of Lost Lake was assessed in greater detail by using the CharAnalysis output data to calculate the fire characteristics for the individual vegetation Zones LL-1 to LL-6 (Table 4).

Fire frequency in the deepest section of the Lost Lake core is moderate, and remains low during the post-glacial pine-dominated assemblage of Zone LL-1. Frequency begins increasing again ca. 11,600 cal yr BP, displaying a generally increasing trend throughout Zones LL-2 to LL-4, and peaking at ca. 6000 cal yr BP when *T. heterophylla* and *Cupressaceae* dominated the vegetation assemblage. A rapid decrease is observed between ca. 6000 and 3400 cal yr BP in Zone LL-4, followed by a similarly rapid increase reaching the record maximum at 2300 cal yr BP. Fire frequency declines post-2300 cal yr BP with a minor increase from 550 cal yr BP to the present. CHAR values fluctuate significantly, however generally remain low to moderate in the first half of the core, with intermittent peaks throughout. CHAR begins noticeably increasing in Zone LL-3 at approximately 7300 cal yr BP, remaining elevated until a gradual decline begins ca. 4000 cal yr BP. CHAR exhibits a final small increase from approximately 550 cal yr BP to the present. Background CHAR generally aligns with the CHAR trend, remaining moderate to low from the deepest section of the Lost Lake core until approximately 7500 cal yr BP, at which point it increases significantly. Background CHAR levels peak at approximately 3580 cal yr BP, then decline and remain low for the remainder of the core.

Table 4. Major zonal charcoal trends and fire episodes of the Lost Lake core.

Zone	Time Period (cal yr BP)	Depth (cm)	Mean CHAR (pieces /cm^2 /year)	CHAR trend*	Mean Background CHAR (pieces/ cm^2/year)	Background CHAR trend	Mean Fire Frequency (fires/ 1000yrs)	Fire Frequency Trend*	Number of Fire Episodes	Fire Episodes (cal yr BP)	Fire Episode Peak Magnitudes (pieces/cm^2/ peak)
LL-6b	1,173 – -71	54 – 0	40.3	Decreasing then increase to present	8.2	Decreasing	1.54	Decreasing , then minor increase to present	2	105	42,871
										941	2
LL-6a	2,417 – 1,173	75 – 54	8.21	Decreasing	8.3	Minor increase, then decreasing	3.60	Peak at beginning, then decreasing	4	1469	299
										1777	230
										2085	48
										2217	26
LL-5	3,940 – 2,417	110 – 75	17.5	Decreasing	16.9	Peak at beginning, then decreasing	2.12	Rapidly increasing	4	2437	27
										2701	185
										2877	446
										3845	174
LL-4	6,680 – 3,940	175 – 110	15.6	Increasing	14.6	Increasing	1.90	Local peak, then decreasing	5	4901	1162
										5385	173
										5825	106
										6177	160
										6441	312
LL-3	9,085 – 6,680	216 – 175	8.3	Increasing	7.3	Decreasing, then increasing	1.40	Increasing	3	7189	1378
										7629	24
										8553	132

<table>
<tr><th>Zone</th><th>Time Period (cal yr BP)</th><th>Depth (cm)</th><th>Mean CHAR (pieces /cm^2 /year)</th><th>CHAR trend*</th><th>Mean Background CHAR (pieces/ cm^2/year)</th><th>Background CHAR trend</th><th>Mean Fire Frequency (fires/ 1000yrs)</th><th>Fire Frequency Trend*</th><th>Number of Fire Episodes</th><th>Fire Episodes (cal yr BP)</th><th>Fire Episode Peak Magnitudes (pieces/cm^2/ peak)</th></tr>
<tr><td rowspan="3">LL-2</td><td rowspan="3">12,229 – 9,085</td><td rowspan="3">251 – 216</td><td rowspan="3">9.05</td><td rowspan="3">Decreasing</td><td rowspan="3">8.9</td><td rowspan="3">Increasing</td><td rowspan="3">1.0</td><td rowspan="3">Low; Increasing</td><td rowspan="3">3</td><td>9443</td><td>45</td></tr>
<tr><td>10,137</td><td>34</td></tr>
<tr><td>11,237</td><td>227</td></tr>
<tr><td rowspan="2">LL-1</td><td rowspan="2">13,932 – 12,229</td><td rowspan="2">287 - 251</td><td rowspan="2">10.2</td><td rowspan="2">Increasing</td><td rowspan="2">5.9</td><td rowspan="2">Low, stable</td><td rowspan="2">1.30</td><td rowspan="2">Decreasing</td><td rowspan="2">2</td><td>12,469</td><td>2462</td></tr>
<tr><td>13,261</td><td>1336</td></tr>
</table>

*Trends in CHAR and fire frequency are relative to previous zone.

Chapter 4. Interpretation and Discussion

4.1. Zonal Interpretations

Zone LL-1 (287 - 251 cm; 13,932 – 12,229 cal yr BP)

With an estimated basal age of 13,932 cal yr BP, Zone LL-1 records the transition from the last glacial period into the Holocene period (11,700 cal yr BP–present) (Walker et al., 2018), covering multiple climatic events including the Bølling-Allerød event, the Sumas glacial event, and the Younger Dryas period (Mathewes et al., 1993; Clague et al., 1997; Ivanovic et al., 2016).

The Bølling-Allerød event (B-A; approximately 14,700 – 12,800 cal yr BP) occurred during the transition between the Last Glacial Maximum and the current interglacial period (Ivanovic et al., 2016) and was characterized by rapid sea level rise, temperature increases, and warm, moist conditions over much of the Northern Hemisphere. The B-A has been associated with Meltwater Pulse 1A, a large influx of fresh water into the Atlantic Ocean which triggered a reinvigoration of the Atlantic meridional overturning circulation (AMOC) causing warming oceans and abrupt global sea level rise (Clark et al., 2002; Weaver et al., 2003; Gregoire et al., 2012; Thiagarajan et al., 2014). The B-A was dominantly observed in Europe and eastern North America (Peteet, 2000), and there is no consensus in the literature regarding a B-A link to the northeastern Pacific and western North America. However, certain studies have documented a notable increase in precipitation in northwestern North America coinciding with moistening conditions in eastern North America during the same timeframe, suggesting a link to the B-A (Whitlock & Bartlein, 1997; Lora et al., 2016).

In general, low fire frequency and general vegetation characteristics indicate that warm climate conditions traditionally associated with the B-A period were not strongly apparent at Lost Lake. The presence of *Salix* and *Poaceae* in the deepest part of the Lost Lake core (>13,000 cal yr BP) along with abundant diploxylon *Pinus* (yellow pine) indicate that an open, shrubby environment existed directly following deglaciation, and it was quickly colonized by *Pinus*. I infer that the landscape was an open-canopy woodland dominated by shade-intolerant *P. contorta* until ca. 13,400 cal yr BP. I recognize that *Pinus* is often over-represented in the pollen record as it produces much more pollen than other species (Allen & Hebda, 1993; Pellatt, 1996; Pellatt et al., 1997), so it may have been less dominant than its pollen percentages imply. However, the accumulation rate of *Pinus* pollen is also high during the glacial-to-Holocene transition (ca. pre-13,000 cal yr BP), an observation that is supported by the increase in *Pinus* macrofossil evidence from nearby sites in coastal B.C. (Clague et al., 1997; Wainman and Mathewes, 1987).

Fire frequency and background CHAR data suggest that only two fire episodes occurred at around 13,261 and 12,469 cal yr BP, one of which occurred during the B-A period. The boreal vegetation composition during this time indicates that cool conditions presided, and the overall high pollen accumulation rate in the middle of Zone LL-1 (ca. 13,200 – 12,600 cal yr BP) may be indicative of a closed forest landscape (Ritchie & Lichti-Fedorovich, 1967; Mathewes, 1973; Brown & Hebda, 2002). However higher pollen and charcoal accumulation could also be a result of redeposition of old material as deglaciation occurred, or elevated influx during periods of high fluvial activity or erosion (Wainman & Mathewes, 1984; Pellatt, personal communication).

Somewhat coincident with the B-A period on the Pacific Coast of Canada was the Sumas glacial event, or (“Sumas Advance,” 13,700 – 13,200 cal yr BP) which occurred near the end of the Fraser Glaciation when the Cordilleran ice sheet had begun downwasting and retreat (Clague et al., 1997). During the Sumas Advance, a lobe of the Cordilleran sheet re-advanced into the Fraser Valley and northern Washington, covering recently established vegetation in low-lying areas and producing a short but noticeable return to glacial conditions. Previous literature estimates that the Sumas lobe extended through the Fraser Valley to the east and southeast of Lost Lake (Clague et al., 1997; Pellatt et al., 2001), and no evidence of the Sumas Drift has found in geomorphological field surveys of the Seymour Valley (Lian, 1991).

Following the B-A warming was the Younger Dryas period (YD; 12,800 – 11,500 cal yr BP), a period of cooling in the Northern Hemisphere which disrupted the warming trend in the North Atlantic (McManus et al., 2004; North Greenland Ice Core Project Members, 2007; Renssen et al., 2015). Palaeoecological evidence of abrupt cooling was first observed in northwestern Europe (Watts, 1980; Rind, 1986), and later linked to coeval cooling in eastern North America (Mott et al., 1986; Walker et al., 1991). Evidence of YD-related cooling in British Columbia is less abundant, but has been inferred via northeastern Pacific sea surface temperature (SST) records (Kienast & McKay, 2001; Fig. 2), concurrent glacial readvancement in British Columbia's Coast Mountains (Friele & Clague, 2002, Fig. 2), and palynological indicators along the Pacific coast (Mathewes, 1993; Lacourse, 2005; Galloway et al., 2009). Records from the NGRIP ice core in central Greenland show an abrupt shift in the Northern Hemisphere to cool, nearly glacial conditions during the YD (Fig. 2) despite mid-June solar insolation at 60°N nearly reaching its 12-ka maximum (Berger & Loutre, 1991; Fig. 2). Ecosystem-scale shifts in vegetation have been associated with the YD period in the Rocky Mountains (Reasoner et al., 1994; Reasoner and Jodry, 2000), western North America (Kennett et al., 2008) and coastal British Columbia (Mathewes, 1993, Mathewes et al., 1993; Walker & Pellatt, 2003; Galloway et al., 2009; Rodengen et al., 2022).

The late-glacial (>11,500 cal yr BP) pollen assemblages in the oldest part of in the Lost Lake core reveal a vegetation regime that was very similar to those described at Marion Lake in southwestern BC (Mathewes,1973), in the Fraser Lowlands at Surprise Lake (Wainman & Matthews, 1987), and in the CWHvm zone of northern Vancouver Island (Lacourse, 2005), which showed vegetation regimes that were dominated by pine with elevated fir (*Abies)* and spruce (*Picea*).

As the climate moistened ca. ~13,000 cal yr BP (Walker & Pellatt, 2003; Galloway et al., 2008; Hebda, 1995), the more shade-tolerant *Tsuga mertensiana, Picea,* and *Abies* increased in abundance at Lost Lake, suggesting that coniferous forests began overtaking the landscape. I suggest that the majority of *Picea* pollen observed in the Lost Lake core is likely that of *P. sitchensis* (Sitka spruce), while the majority of *Abies* pollen is likely *A. amabilis* (Amabilis fir). This interpretation is supported by works of Brown & Hebda (2002a) and Lacourse (2005) in the CHWvm zones of Vancouver Island and Mathewes & Wainman (1987) in the CWH zone of southwestern BC at Marion Lake. *Picea* and *Abies* macrofossils found within the Marion Lake core were those of *P. sitchensis* and *A. amabilis.* Sitka spruce is also a common pioneer species of open and disturbed landscapes, including recently deglaciated terrain (Harris, 1966).

The vegetation at Lost Lake continued to show signs of a boreal composition (*Pinus, Picea* and *Abies*) well into the YD until approximately 12,500 cal yr BP, supporting the inferred cool climate (Fig. 6). Fire frequency noticeably dipped at Lost Lake during the YD (1 fire/1000yrs; Fig. 8). Elevated levels of background CHAR during fire events may be indicative of fires that were occurring regionally outside the Lost Lake watershed (Clark et al., 1996, Long et al., 1998, Higuera et al., 2007), as the inferred boreal forest composition, cool climate and newly established forests at Lost Lake would not have been conducive to generating fuel loads capable of producing such high CHARs (Krawchuk & Moritz, 2011). Background CHAR influx may also be a result of redeposition events such as landslides or periods of high fluvial discharge, which are capable of affecting charcoal flux despite no major climatic changes.

Despite the two inferred fire events in Zone LL-1, fire frequency and CHAR remained low, indicating minimal fire activity during this zone and suggesting elevated CHAR was caused by regional burning outside the Lost Lake watershed or redeposition of charcoal particles. Summer insolation was increasing throughout Zone LL-1 and the YD period, but reconstructed precipitation estimates, northeast Pacific SST and inferred temperatures from the NGRIP ice core indicate a cool and dry regional climate (Fig. 2).

Zone LL-2 (251 – 216 cm; 12,229 – 9,085 cal yr BP)

The end of the Younger Dryas period was characterized by warming across much of the Northern Hemisphere (Alley et al., 2000; Jennings et al., 2006; Brauer et al., 2008; Lynch-Stieglitz, 2011). Although the YD did not officially end until approximately 11,500 cal yr BP, the Lost Lake record suggests that the region around Lost Lake began warming and drying earlier at approximately 12,000 cal yr BP (± 576 years). At this time, *Pseudtosuga menziesii* pollen grew in abundance while *Pinus, Picea* and *Abies* abundances were very low, indicating a transition to warmer, dry conditions. The notably high abundances of *Pteridium* and *Alnus* suggest increased disturbance. *Salix* and *Rosaceae* increased as well, suggesting moderately open forest conditions. Overall, in Zone LL-2 PARs were 32 percent lower than during the previous zone LL-1, but the elevated PARs of both *Pseudtosuga* and *Pteridium* during Zone LL-2 support their observed increases in pollen percentages and therefore also the interpretation of a transition to warm, dry conditions after 12,000 cal yr BP.

Despite the pollen abundance showing signs of warmer, dry conditions, fire frequency reached its minimum during this zone and remained at or below 1 fire/1000yrs until 10,270 cal yr BP. Throughout the rest of Zone LL-2 fire frequency increased slightly but did not surpass 2 fires/1000yrs. This indicates that the warming climate at Lost Lake was not accompanied by marked increases in fire within the watershed during Zone LL-2.

The inferred shift to warmer, drier conditions around Lost Lake after 12,000 cal yr BP is supported by regional data and is indicative of increased disturbance. The shift in vegetation assemblage and the increasing fire frequency align with rising solar insolation values which peaked at approximately 12,000 cal yr BP, and increasing SST in the northeast Pacific between ca. 12,000 and 10,000 cal yr BP (Fig. 2). Coincident increases in *Pteridium* and *Alnus* pollen abundances across this time period support an increase in disturbance in the understory. Similar results are found in studies conducted in the CWH zone in the Fraser Valley (Mathewes, 1973; Wainman & Mathewes, 1987: Pellatt et al., 2001) and on Vancouver Island (Brown & Hebda, 2003; Brown et al., 2019), which indicate increased disturbance and that precipitation began to decline ca. 11,500 cal yr BP. The coastal variety of *P. menziesii* is a largely shade intolerant species and is often one of the first species to establish after fire disturbances (Farrar, 1995; Nuszdorfer et al., 1991). Thus, the ecosystem during this time was most likely a *P. menziesii*-dominated forest with a shrubby understory, indicating a forest which was adapted to warmer conditions and higher disturbance than the present.

At the same time, the fire-sensitive species *Tsuga heterophylla* gradually began increasing to near-modern levels beginning at ca. 10,000 cal yr BP (Fig. 6). This increase aligns temporally with a decline in summer insolation values and increase in winter insolation values at 60°N (Fig. 2), and is supported by similarly timed *T. heterophylla* increases further north along the coast of BC (Lacourse, 2005; Galloway et al., 2009). A decrease in seasonality brought about by less extreme summer and winter solar insolation may have caused milder winter temperatures and a longer growing season, allowing T. *heterophylla* an opportunity to expand (Gavin et al., 2011; Pojar & Mackinnon, 1991), replacing the less shade-tolerant *Picea* (Fastie, 1995).

An alternative explanation for *T. heterophylla's* slow increase is that soil composition and moisture rather than climate was the controlling factor (Hansen, 1960; Mathewes, 1973). Because *T. heterophylla* favors podzolized soils (those which contain a deep organic layer and are typical of coniferous forests) and decaying conifer wood for growth (Krajina, 1969), the initial spread of the species may have been dependent on the formation of a satisfactory soil site series.

Zone LL-3 (216 – 175 cm; 9,085 – 6,680 cal yr BP)

Zone LL-3 began approximately at the commonly stated beginning of the xerothermic period. The xerothermic period likely resulted from multiple climatic factors, dominantly the still high solar radiation causing higher summer insolation in the Northern Hemisphere (Fig. 2; Brown et al., 1995; Imbrie & Imbrie, 1980; Walker & Pellatt, 2003). In the interior of the province, summers were up to 2-4°C warmer than present (Hebda, 1995; Walker & Pellatt, 2003; Rosenberg et al., 2004). Coastal areas likely experienced a dampening effect from the Pacific Ocean, and some studies suggest that warming on the coast was closer to 1-2°C warmer than present (Hebda, 1995).

However, the observed vegetation shifts suggest that the climate around Lost Lake shifted toward moister conditions during the xerothermic period and likely remained relatively cool rather than warming significantly. During Zone LL-3, *Pteridium* and *Alnus* declined but were still present in both the pollen percentages and in PARs. *Ericaceae, Artemisia, Poaceae and Brassicaceae* pollen appeared in significant amounts for the first time during this zone, suggesting both an abundant understory presence and continued disturbance. While still present, *P. menziesii* began to decline during Zone LL-3 between ca. 9000 and 8000 cal yr BP, at the same time that *T. heterophylla* began to increase, followed by *Cupressaceae* and *Picea.* These shifts in arboreal composition suggest moistening and likely cooling instead of warming. Both *T. heterophylla* (western hemlock) and *T. plicata* (western redcedar) favor wet conditions, while *P. sitchensis* (Sitka spruce) favors wet and cool conditions (Meidinger & Pojar, 1991).

Following the Mazama ash horizon at ca. 7600 cal yr BP, the assemblage at Lost Lake showed an increase in *Abies* pollen percentages. These increases are consistent in timing with *Abies amabilis* macrofossils and *Isothecium stoloniferum* (tree moss) subfossils found in the Marion and Surprise Lake cores (Mathewes 1973). Amabilis fir prefers cool growth conditions and has one of the highest water demands of evergreens in British Columbia (Klinka et al., 1995), which suggests continued wetness and the beginning of cooler conditions at Lost Lake. *Betula*, likely swamp birch, also increased in this zone, further suggesting a transition toward a wetter climate. These shifts are again very similar to those described by Mathewes (1973) and Wainman & Mathewes (1986) at Marion Lake, with the key difference being the earlier development of *T. heterophylla* at Lost Lake, which may be attributed to its low elevation and closer proximity to the Pacific coast.

My results suggest that the xerothermic interval was not felt strongly at Lost Lake in comparison to other sites in the Fraser Valley (Mathewes & Heusser, 1981; Wainman & Mathewes, 1987) on southern Vancouver Island (Brown et al., 2019), along the Pacific coast of Washington (Heusser et al., 1980) and at higher elevations in southwestern British Columbia (Hallett et al., 2003; Shea et al., 2022). The interpreted cool, moist climate and mixed-species forest during Zone LL-3 at Lost Lake appears more similar to sites on Haida Gwaii (Louise Pond, SC1 Pond, Shangri-La Bog, Pellatt & Mathewes, 1994, 1997), on the central coast of BC (Tiny Lake, Doherty, 2005; Galloway et al., 2009) and on northern Vancouver Island (Bear Cove Bog, Hebda, 1983; Misty Lake, Lacourse 2005), which tended to experience a phase of warming earlier in the Holocene (between ca. 11,000 and 7500 cal yr BP), followed by a longer phase of cooling throughout the mid-Holocene. In the final ~1700 years of Zone LL-3, fire frequency, interpolated CHAR and background CHAR increased steadily, suggesting an increase in frequent, local fires as well as more regional biomass burning. The persistence of western hemlock, spruce and birch during this time indicate that forests remained cool and moist despite the increase in fire activity.

It is feasible that the proximity of Lost Lake to the Pacific coast (<20 km) may have buffered the coastal site against the stronger Pacific High and weakened Aleutian Low pressure systems that are thought to have occurred during the early Holocene (Walker & Pellatt, 2008). This buffering allowed for relatively cooler, wetter conditions than were experienced further inland. A stronger resilience to climatic changes was also observed throughout the Holocene in the outer coastal zones on the west coast of Vancouver Island (Brown et al., 2006). The more pronounced seasonal amplitude of insolation during the early-mid Holocene (Fig. 2) likely did cause a more continental climate at my site and allowed for the expansion of western hemlock due to cool winters. However, the coastal nature of the site may have provided enough moisture to account for the decrease in overall biomass burning at Lost Lake between 10,100 and 7600 cal yr BP.

Zone LL-4 (175 – 110 cm; 6,680 – 3,940 cal yr BP)

The mid-late Holocene (ca. 7800 – 3800 cal yr BP) in southern British Columbia has been considered a climatic transition out of the warm, dry xerothermic period, first into warm, moist conditions and then into cooler moist conditions (Mathewes, 1985). This transition was termed the mesothermic period by Hebda (1995), who defined it as beginning when annual precipitation in the record rose to modern levels while temperature remained warmer than present and ending when temperatures decrease to those of the present day. While a warm, moist mesothermic period has been observed in some mid-Holocene records (Hebda, 1995, 1997), it is not well-defined on the coast, and many sites appear to transition from the warmer early Holocene directly into a cooler and moist neoglacial-type climate (Heusser, 1985; Mathewes, 1973, 1989; Mathewes & Heusser, 1981; Hebda, 1983, 1995; Wainman & Mathewes, 1987; Pellatt & Mathewes, 1996; Pellatt et al., 2001; Galloway et al., 2008; Brown et al., 2019). Some sites in northwestern Washington began showing signs of cooling as early as 7500 cal yr BP (Barnosky et al., 1987; McLachlan & Brubaker, 1995; Gavin et al., 2001), while those in southwestern BC generally exhibited a long and variable transitional period trending towards modern conditions between 7500 and ~4000 cal yr BP (Mathewes, 1973; Mathewes & Heusser, 1981; Heusser, 1985; Pellatt et al., 2001; Hallett et al., 2003; Brown et al., 2019). Sites along the north-central coast and on Haida Gwaii varied, with some experiencing a relatively warm, moist period at ca. 6000 cal yr BP (Quickfall, 1987; Hebda 1995) while others experienced a gradual transition from the warm, dry early Holocene into a cool, wet climate as early as 9500 cal yr BP, without a distinct mesothermic period (Hebda, 1983; Pellatt & Mathewes, 1994, 1996; Galloway et al., 2009). Reconstructed temperature records from the northeast Pacific Ocean (Fig. 2) and southern BC (Palmer et al., 2003; Rosenberg et al., 2004; Chase et al., 2008; Fig. 2) do not appear to cohesively demonstrate a mesothermic interval in the mid-Holocene, suggesting it was a site-specific phenomenon dependent on local factors rather than regional climatic change.

Following the mesothermic/mid-Holocene transition period, an interval of renewed glacial activity overtook much of the Northern Hemisphere. In western North

America following 4000 cal yr BP, solar insolation was nearing modern levels, which was reducing seasonality and changing ocean-atmosphere dynamics (Berger & Loutre, 1991; Shea, 2022). During this time the Aleutian low pressure system is inferred to have strengthened while the Pacific High pressure system weakened, resulting in more modern conditions associated with the El Niño-Southern Oscillation (ENSO) in the North Pacific Ocean (Walker & Pellatt, 2003; Barron & Anderson, 2011). These changes in the average atmospheric pressure conditions likely heightened the cool, moist conditions associated with the Neoglacial along the Pacific coast of Canada (Walker & Pellatt, 2003; Barron & Anderson, 2011). Major cooling associated with glacial advance occurred as early as 6800 cal yr BP throughout British Columbia (Garibaldi Phase advance (6840 - 5700 cal yr BP); Ryder and Thomson, 1986; Ryder, 1989; Walker & Pellatt, 2003), although Neoglacial cooling affected some southern coastal sites at later time periods of 5500 - 4000 cal yr BP (Gavin et al., 2006; Menounos et al., 2009) and sites on Haida Gwaii around ca. 3200 cal yr BP (Pellatt & Mathewes, 1994, 1996), aligning with the timing of glacial advances in the Coast and Rocky mountains (Ryder & Thompson, 1986; Menounos et al., 2009).

At Lost Lake, the inferred vegetation assemblage in Zone LL-4 continued to suggest that a moist climate dominated from 6,680 to 3,940 cal yr BP. *Tsuga heterophylla* remained the largest component of the pollen assemblage and *Cupressaceae* pollen increased in abundance throughout the zone. As with modern CWH forests, increasing abundances of *Cupressaceae* and *T. heterophylla* are likely indicative of high moisture during this time period. Other moisture indicators are the elevated abundances of *Betula* (swamp birch) and *Salix* (willow), which are both species common to very moist regions (Meidinger & Pojar, 1991; Klinka et al., 1999). *Pseudotsuga* declined to near modern levels by ca. 6000 cal yr BP, indicating that the climate was no longer dry enough to support growth. The influx of *Tsuga mertensiana* – a species which in the present day prefers the cool conditions of the subalpine – rose after fire frequency began to decline at ca. 5800 cal yr BP. Studies of modern pollen assemblages have found that while *T. mertensiana* is more abundant at higher elevations, moist valley bottoms may also receive pollen from upslope (Hebda & Allen, 1993; Heusser, 1978). This idea would signal the presence of cold, moist conditions at higher elevations surrounding Lost Lake, which has been observed in other subalpine coastal rainforests during this time period (Brown & Hebda, 2003).

As forest composition became dominated by western hemlock and western redcedar during Zone LL-4, *Picea* (likely Sitka spruce) pollen decreased from 4% of the composition to <1% between 6300 and 4700 cal yr BP. A possible reasoning could be that as mixed-forest stands of western hemlock and western redcedar increased, Sitka spruce – which is moderately shade-intolerant and has higher soil nutrient requirements – was simply outcompeted (Fastie, 1995; Klinka et al., 1999). A minor influx of *Poaceae* between zones LL-4 and LL-5 (approximately 4700 – 2595 cal yr BP) indicates that gaps in the forest canopy may have opened mid-zone as a result of the inferred increased fire activity at ca. 6000 cal yr BP. The inferred cool and wet winters throughout this interval and the rest of LL-4 likely produced ample snowpack that remained later into the summer, creating wetter summer environments and perhaps also contributing to the decreasing fire frequency between 6000 and 3500 cal yr BP (Gavin et al., 2009; Rodengen et al., 2022).

While pollen assemblages suggest that conditions were cool and moist during the onset of Zone LL-4 (ca. 6680 cal yr BP), a local peak in fire frequency to ~3 fires/1000yrs with an average mFRI of 308 years was observed in the record at ca. 5950 cal yr BP (Table 4, Fig. 8). Background CHAR was also elevated between 6500 and 4000 cal yr BP (Fig. 8), suggesting that biomass burning increased both locally around Lost Lake and regionally at more distant locations. *Betula* and *Alnus* are known to grow in response to fire, and the high percentage of both in this zone reflect the continued increases in inferred fire frequency around Lost Lake between 8500 and 5950 cal yr BP.

Declining temperatures were recorded during this time period in the northeast Pacific (Kienast & McKay, 2001; Fig. 2) and in southwestern BC (Heusser et al., 1985, Fig. 2). Midge data from southern BC (Frozen Lake, Rosenberg et al., 2004) also suggest that near-modern cool conditions were established at approximately 4500 cal yr BP, yet an earlier study at Frozen Lake suggests fire frequency was increasing until approximately 4000 cal yr BP. Several other fire history studies around southwestern BC also found increases in fire frequencies in the mid-Holocene (ca. 8000 – 4500) despite cool, moist conditions during this time. Fire frequency at Frozen Lake (southern Coast Mountains; Hallett et al., 2003) was elevated between 4000 and 5000 cal yr BP, while fire frequency at Somenos Lake (southeastern Vancouver Island, Murphy et al., 2019) was elevated from 5000-4500 cal yr BP. On southern Vancouver Island, Brown and Hebda (2002a,b) determined that fires were likely limited by regional moistening.

However, they also noted a continuous influx of charcoal at Pixie Lake (CWHvm) between approximately 6800 – 3000 cal yr BP, suggesting that the wetter climate did not deter fires from burning regularly. This site-specific and highly variable fire frequency data in southwestern BC during the mid-Holocene is indicative of location-dependent factors rather than broad climatic controls on fire activity.

Because the climate is inferred to have been unfavorable to natural fire ignitions during Lost Lake's local peak in fire frequency at 5900 cal yr BP, increases in fire activity may have instead been at least in part due to anthropogenic use of fire as a landscape modification tool, which is a well-documented phenomenon in northwestern North America (Boyd, 1999; Lepofsky et al., 2005; Walsh et al., 2018). Evidence of human habitation along the northwestern coast of North America dates to 11,000 cal yr BP in southern Alaska and on the central BC coast (Carlson, 1994, 1996). Human activity in coastal BC and Washington has been suggested in the post-Mazama interval by Walsh et al. (2015) and documented on Orcas Island in the southern Strait of Georgia as early as ca. 11,900 cal yr BP (Kenady et al., 2011). The Stó:lō-Coast Salish peoples likely occupied the Fraser Valley as early as ca. 8000 cal yr BP, but signs of residential structures did not begin until ca. 4800 cal yr BP (Schaepe, 1998; Lepofsky et al., 2009). Some of the earliest documentation of Indigenous peoples on the Strait of Georgia coast is evidenced by signs of small, single-family dwellings dated between 4000 – 3500 cal yr BP (Grier & Kim, 2012; Johnstone, 2003). The ample evidence that humans were well-established on the coast by the early-mid Holocene, in combination with little to no change in vegetation assemblage at Lost Lake, provides a basis for the theory that fire disturbance during this interval was not solely caused by climatic factors and human burning was likely a contributor.

As fire frequency decreased in the last half of Zone LL-4, CHAR remained high, which is generally an indicator of burns that occur either upwind of the lake or outside the watershed entirely (Gardner & Whitlock, 2001). Another possible cause for the high CHAR observed throughout Zone LL-4 is that increased rainfall during this time resulted in higher levels of sediment transport (Long et al., 1998), which resulting in additional reworked charcoal washing into the lake from other areas of the watershed (Hallett et al., 2003). This is thought to have occurred at other sites during the mid-Holocene (Brown & Hebda, 2003, 2019; Hallett et al., 2003), and would explain the presence of high CHAR despite the inferred cool, moist climate. Thus, it is likely that the fire behavior at Lost

Lake during Zone LL-4 is from a combination of increased anthropogenic burning and elevated charcoal influx due to increased precipitation relative to the early Holocene.

Zone LL-5 (110 – 75 cm; 3,940 – 2,417 cal yr BP)

Zone LL-5 occurred shortly after the beginning of the Neoglacial period of ca. 4500 cal yr BP, during a point of low fire frequency (1 fire/1000yr) at Lost Lake. During the Neoglacial period, several glacial advances occurred in the Coast Mountains north of the study area, including the Tiedemann Advance, comprised of several advances between 3500 and 1800 cal yr BP (Ryder & Thomson, 1986). Increased glacial activity is inferred to be a result of declining summer insolation in the Northern Hemisphere (Fig. 2; Clague et al., 2009). Clear evidence of these advances was not observed in the vegetation shifts at Lost Lake, likely because the glacial advances did not extend far enough south to reach the site and were relatively short in duration. While no evidence of re-glaciation is present in the sediments of the Lost Lake core because of its proximity to the south coast, the regional climate during the first portion of Zone LL-5 (3940 – 3500 cal yr BP) is still inferred to have been cold. At Lost Lake, moisture indicators *Cupressaceae* and *T. heterophylla* continued to dominate the vegetation assemblage. While *T. plicata* is the more likely producer of *Cupressaceae* pollen at Lost Lake, both red and yellow cedar macrofossils were found during this timeframe in southwestern BC at Surprise Lake (Mathewes, 1973), so I cannot entirely rule out *C. nootkatensis*, especially considering the inferred glacial climate of the time. *T. plicata* is very susceptible to fire damage due to its thin bark, shallow root system, and flammable foliage (Parminter, 1983; Fischer & Bradley, 1987), and consequently on the coast it is most often found at low to mid-elevation, wet sites.

The high abundance of *Cupressaceae* aligns with the very low fire frequency (1.0 fire/1000yrs) and long mFRI (1012 years) in the first portion of Zone LL-5. The continued abundance of *T. heterophylla* indicates that a hemlock-dominated assemblage was well established during Zone LL-5, and a wet moisture regime was present (Walker & Pellatt, 2003; Hallett & Hills, 2006). While it is not possible to directly correlate fire extent or severity to CHAR activity, the large charcoal peak (indicated by peak magnitude) observed at 3845 cal yr BP (174 pieces/cm^2/yr) in combination with high background CHAR (31 pieces/cm^2/yr) likely indicate that a large or high severity fire event occurred during this time. The long fire-free interval (1162 years) preceding the fire at 3845 cal yr

BP would have allowed ample time for fuels to build up. In the second part of the zone (3500 – 2400 cal yr BP), fire frequency and fire activity rapidly increased to reach their peaks at the transition between Zones LL-5 and LL-6a (Fire frequency = 4.40 fires/1000yrs; mFRI = 198 years). At the same time, background CHAR declined, suggesting that while local fires were occurring more frequently, biomass burning in the region was reduced in size or severity. These data could indicate that low-severity ground fires burned frequently while high-severity canopy fires were rarer (Brown & Hebda, 1998).

The increasing fire activity at Lost Lake beginning at ca. 3500 cal yr BP aligns with the intensification of the positive (El Niño) phase of El Niño Southern Oscillation (ENSO) activity that developed between 3500 and 2500 cal yr BP (Moy, Seltzer, Rodbell, & Anderson, 2002), and the enhanced positive PDO that occurred in the North Pacific Ocean after ca. 3200 cal yr BP (Barron & Anderson, 2011). ENSO activity occurs on three to seven year frequencies, and originates from changing SSTs in the eastern equatorial Pacific Ocean which affect ocean-atmosphere interactions throughout the Pacific (Barron & Anderson, 2011). In the modern ocean, El Niño intensification is associated with anomalously warm sea-surface temperatures in the central and eastern tropical Pacific Ocean, which push the Pacific jet stream further south than normal and cause warmer, drier conditions along the northwest Pacific coast of North America (NOAA 2021). Enhanced (positive) PDO (Pacific Decadal Oscillation) phases are associated with sea-surface temperature and circulation patterns that are similar to intensified El Niño activity. As the northeastern Pacific Ocean warms, it results in decreased winter precipitation in the Pacific Northwest and southwestern coastal BC (Barron & Anderson, 2011).

The SST warming associated with the positive phases of both climate oscillations in the late Holocene (post- ~4000 cal yr BP) would have reduced the effective moisture (precipitation-evaporation) on the southern BC coast and caused drier winter conditions. Diatom proxy data in southern BC indicate reduced effective moisture and lower lake levels after ~3600 cal yr BP (Bennett et al., 2001), while varve data from the Saanich Inlet became thinner after ca. 3200 cal yr BP (Nederbragt & Thurow, 2001), indicating a drier climate. Positive PDO and ENSO phases have been correlated to higher fire activity by causing reduced snowpack and thus longer fire seasons (Heyerdahl, Brubaker & Agee, 2002), which may have been a factor in the increased fire activity at Lost Lake.

Subzone LL-6a (75 – 54 cm; 2,417 – 1,173 cal yr BP)

The rapid increase in fire frequency observed at Lost Lake between ca. 3500 and 2300 cal yr BP has similarly been observed in other paleoclimate studies in southwestern BC (Hallett et al., 2003; Lepofsky et al., 2005) and northwestern Washington (Rorig & Ferguson, 1999; Prichard et al., 2009), for which multiple reasons have been proposed. A period of cultural change (the Marpole phase, 2400 – 1200 cal yr BP) occurred nearly simultaneously with a regional climatic drying (the Fraser Valley Fire Period, 2400 – 1300 cal yr BP) during Zone LL-6a, complicating interpretation of the possible causal factors of Lost Lake's high fire frequency.

Though the Coast Salish peoples have been living in and around the Vancouver region for thousands of years and likely since soon after deglaciation in the late Pleistocene (Hoffman, Lertzman & Starzomski, 2017; Fisher et al., 2019; McLaren et al., 2019; Fedje et al., 2021), the Marpole cultural phase was a time of growing interconnectedness and advancement of socioeconomic systems between Indigenous peoples living near and around the Fraser River system. It is "characterized by the widespread appearance of large houses, standardized art forms, and elaborate burials" (Lepofsky et al., 2005). The mid-Holocene establishment of *T. plicata* – which was observed in the Lost Lake core at approximately ca. 6400 cal yr BP – was important to the growth of cultures on the coast of southwestern BC and northwestern Washington (Hebda & Mathewes, 1984), and the high abundance of *T. plicata* around Lost Lake was likely a contributor to the value of the area. Previous researchers have linked the Marpole phase with the FVFP, with summer droughts and increased wildfire activity affecting the availability of resources and causing intensification of socioeconomic relations between inhabitants of the region (Lepofsky et al., 2005). Consequently, the observed rise in fire activity during this time could very well have been at least partially due to the coeval increase in anthropogenic activity.

The Fraser Valley Fire Period (FVFP) was a climatically distinct period concurrent with the Marpole Phase, during which wildfire incidences increased and climate was drier, and possibly warmer, than today (Hallett, 2001; Hallett et al., 2003). The primary inferred cause of increased fires in western North America are prolonged and frequent summer droughts during this time, which have been inferred via elevated charcoal records from southwestern BC (Hallett et al., 2003, Murphy et al., 2019). In addition to the possible contributions of stronger summer drought conditions and increased anthropogenic ignitions (Lepofsky et al., 2005), Pritchard et al. (2009) also suggest fires during the FVFP may have been caused by an increase in lightning ignitions (Rorig & Ferguson, 1999) or more productive forests producing more fuel for high-severity fires. However, most sites along the coast of BC and northwestern Washington suggest that climate was similar to modern for at least the past ~3000 years (Mathewes 1973; Mathewes and Rouse 1975; Mathewes and Heusser 1981; Mathewes and King 1989; Whitlock 1992; Thompson et al. 1994; Hebda 1995; Pellatt and Mathewes 1997; Pellatt et al. 1998, 2000, 2001; Gavin and Brubaker 1999; Gavin et al. 2001; Galloway et al., 2009; Derr, 2014; Leopold et al., 2016; Brown et al., 2019). This regional lack of response suggests (1) that the manifestation of the FVFP was highly dependent on the individual characteristics of each watershed; (2) that many palynological studies have sampling intervals too coarse to capture decade- to century-scale changes in climate; and (3) that ecosystems may have been resilient enough to withstand some level of increased seasonal dryness with little to no change in species composition (Agee, 1993).

Lost Lake's charcoal record rapidly increased beginning at ca. 3500 cal yr BP and reached maximum fire frequency (4.5 fires/1000yrs) and minimum mFRI (211 years) near the beginning of Zone LL-6a between ca. 2480 and 2260 cal yr BP (Fig. 8). These increases suggest a rapid rise in fire activity and magnitude occurred during this time, before seemingly declining equally as rapidly during the rest of the zone. Background CHAR decreased at the same time as the rise in fire frequency and mFRI, perhaps suggesting that the FVFP occurred locally around Lost Lake but not regionally in nearby watersheds. This apparent localized response is further confirmed by the lack of noticeable charcoal influx in of other regional records at this time.

Nearby sites with elevated fire activity similar to Lost Lake between 3000 and 1500 cal yr BP are rare and include Marion Lake (Wainman & Mathewes, 1986; Fig. 1), Pixie Lake (Brown & Hebda, 2002b; Fig. 1), Martins Lake (Gavin et al., 2001), Panther Potholes (Prichard et al., 2009), Frozen and Mount Barr Cirque Lakes (Hallett et al., 2003; Fig. 1) and Chadsey Lake (Murphy et al., 2019). Periods of drought between ca. 2400 and 1500 cal yr BP have also been observed eastward into the Kootenays (Hallett and Walker, 2000) and Rockies (Reasoner & Huber, 1999), and to the south in the Klamath Mountains of California and Oregon (Mohr et al., 2000). Diatom data from Big Lake in south-central BC indicate dry conditions and declining lake levels after 3600 cal yr BP (Bennett et al., 2001) while SST trends in in the northeast Pacific off Vancouver Island show an increase of up to 1°C between 2700 and 1600 cal yr BP (Kienast & McKay, 2001; Fig. 2).

Evidence of the FVFP/Marpole phase is lacking in the Lost Lake pollen record, with the only notable fluctuation being that of *T. plicata*. During this zone, both the percentage and PAR of *T. plicata* declined at ca. 1800 cal yr BP, which would be consistent with a slightly offset vegetational response to increased wildfire and drought conditions inferred between ca. 3000 and 2000 cal yr BP. The lack of relationship between fire frequency and vegetation patterns at this site (i.e., vegetation assemblage remained stable throughout the Neoglacial and FVFP) indicates that climate was not the main driver of fire activity during this time period.

Lost Lake's increase in fire activity beginning ca. 3500 cal yr BP pre-dates the Marpole phase and FVFP by nearly one thousand years, which suggests that increased human activity may not have been the main driver of fire. A potential cause for the disparity between Lost Lake fire activity and the established beginning of the Marpole phase is the accuracy of the age model. The age uncertainty associated with the Bacon age model is 576 years (95% confidence), meaning that fire frequency changes and/or pollen shifts could have occurred later in the record than stated. Another consideration to explain the lack of vegetational response to inferred drought is the potential for subsurface water storage to mediate plant water stress during periods of rainfall variability (Hahm et al., 2019). The subsurface storage capacity of the catchment area surrounding Lost Lake has not been quantified, but if it were to have a high maximum storage capacity, it could, in theory, have enough carryover storage to sustain forests through periods of summer drought (Rempe et al., 2019; Rempe et al., 2022, Preprint). This would be dependent on winter precipitation levels throughout the period, but is plausible given the maritime climate and near-valley bottom location of Lost Lake, which would receive runoff from higher elevation sources to the east.

From ca. 2000 cal yr BP until the end of Zone LL-6a at ca. 1170 cal yr BP, the pollen assemblage remained suggestive of cool, wet conditions while fire frequencies declined. Moisture indicators *Cupressaceae, Betula* and *Salix* increased at approximately ca. 1300 cal yr BP, while *T. heterophylla* and *P. menziesii* remained stable, signaling only minor change to the overall forest composition. During the transition between Subzones LL-6a and LL-6b fire frequencies continued to decline, aligning with the observed cooling in locales that experienced the FVFP, and the continued neoglacial conditions in locations that did not display evidence of a FVFP.

Subzone LL-6b (54 – 0 cm; 1,173 – -71 cal yr BP)

The final Zone LL-6b represents vegetation changes observed during the last millennium, during which time the Pacific coast may have experienced some of the climate changes associated with the Medieval Warm Period (MWP) and the Little Ice Age (LIA) (Walker & Pellatt, 2003; Larocque & Smith, 2004; Walsh et al., 2008; Gavin et al., 2008). Specifically, a short interval of warm, dry conditions occurred from ca. 1100 to 700 cal yr BP across northern Europe and eastern North America during the MWP (also called the Medieval Climate Anomaly) and may be linked to coeval warming seen in some records in western North America (Broecker, 2001; Mann et al. 2009). This warm interval was followed by a brief return to glacial climatic conditions in the North Atlantic region during the "Little Ice Age" (LIA) from 600 to 100 cal yr BP, which has been linked to glacial activity and cooling on the Pacific coast of North America (Ryder, 1987; Pitman et al., 2003; Wilkie & Clague, 2009).

Evidence of the MWP and LIA are strongly observed in Europe and parts of North America, but are less pervasive in other parts of the world and variably observed in records from western North America (Bradley et al., 2003; Ljungqvist, 2010; Neukom et al., 2019). While some evidence of these climate periods, especially the LIA, has been observed regionally in southwestern BC and western Washington (Ryder, 1989; Walker & Pellatt, 2003; Cumming et al., 2002; Arsenault et al., 2007; Menonous et al., 2009; Pitman & Smith, 2012), little evidence is observed in the pollen and charcoal records at Lost Lake. Notably, some oxygen isotope records from southwestern BC and northern Washington give evidence of a period of cooling and/or moistening during the MWP (Ljungqvist et al., 2016; Steinman et al., 2012), whereas warming during this period is noted in the records of Roe Lake (Lucas & Lacourse, 2003), Battle Ground Lake (Walsh et al., 2008), and Frozen and Mt. Barr Cirque Lakes (Hallett et al., 2003). These data suggest the observance of short-term climate fluctuations during the last millennium was highly location-dependent, which may explain both the lack of a regionally coherent response to the MWP or LIA, and the absence of evidence in the Lost Lake record.

At Lost Lake, evidence of the MWP is difficult to distinguish. Fire frequency at the beginning of Zone LL-6b continued to decline following the FVFP and was very low between 900 and 200 cal yr BP (mean = 1.0 fires/1000yrs). Within the MWP (i.e., 1100 – 700 cal yr BP) the mean resolution was 88 years/cm, and only one fire event was recorded (at ca. 940 cal yr BP) with a low peak magnitude of approximately 2.4 pieces/cm^2/peak. Based on the age model used in this study, Lost Lake's pollen record for the past millennium may not be high enough resolution (five pollen samples from 1000 cal yr BP to present) to capture any meaningful vegetation clues during the MWP. Overall, the vegetation and fire history data do not suggest that a noticeable increase in temperature was experienced during the MWP at Lost Lake.

The LIA is more commonly documented in paleoclimate studies in southwestern BC, Washington, and Oregon (Luckman et al., 1995; Gavin et al., 2003; Hallett et al., 2003; Lucas & Lacourse, 2003; Walsh et al., 2008; Pellatt et al., 2015; Rodengen et al., 2022). Dendrochronological dating at Robson and Bennington Glaciers documents a glacial advance that occurred between 500 – 100 cal yr BP (Luckman et al., 1995), while glacial advances and cooler temperatures were recorded at the Tiedemann and Lillooet Glaciers of the Coast Mountains during the LIA between approximately 150 and 800 cal yr BP (Ryder & Thomson, 1986; Larocque & Smith, 2003; Reyes & Clague, 2004; Arsenault et al., 2007). Similarly timed decreases in SST off the coast of Vancouver Island (-0.4°C; Keinast et al., 2001; Fig. 2) and reconstructed July mean temperature at Marion Lake in southwestern British Columbia (up to -1.1°C; Heusser, 1985, Fig. 2) support a regional cooling during the LIA.

Evidence of the LIA is not observed in the pollen record at Lost Lake but can be inferred via the drop in fire activity in Zone LL-6b. Fire frequencies reach a local low (1.3 fires/1000yrs at ca. 590 cal yr BP) immediately before the literature-established onset of the LIA. Interpolated CHAR decreased slightly between 760 and 200 cal yr BP, while background CHAR began dropping at 720 cal yr BP and did not increase again in the record. Mann et al. (2009) acknowledge that the timeframes of the observed climatic patterns of the LIA and MWP are not necessarily exact, and consequently, the inferred low fire activity at Lost Lake between ca. 800 – 200 cal yr BP could be evidence of the

LIA. Again, the core's low pollen resolution (5 samples from 1000 cal yr BP to present) and the short duration of the LIA makes it difficult to confidently confirm evidence in the pollen record.

Interestingly, $\delta^{18}O$ isotope-based winter precipitation reconstructions from lake sediments in northern Washington indicate that conditions in the Pacific Northwest were wetter during the MWP and drier during the LIA (Steinman et al., 2012; Ljungqvist et al., 2016; Shea et al., 2022), which is corroborated by Northern Hemisphere hydroclimate estimations of the past millennium based on multi-proxy reconstructions of temperature and precipitation (Ljungqvist et al., 2016). This oxygen isotope evidence may be suggestive of stronger seasonality in the past ~1000 years, as wetter winters manifest as low $\Delta\delta^{18}O$ due to higher catchment runoff. Stronger La Niña-like conditions in the tropical Pacific are thought to have been a cause of high precipitation in the Pacific Northwest during the MWP (Steinman et al., 2012). Large seasonal changes (i.e., dry summers and wet winters) are more likely to be observed through decadal to centennial-scale proxies like $\Delta\delta^{18}O$ and tree rings, explaining why many study sites including Lost Lake recorded limited or no evidence of centennial-scale climate changes in their fire and vegetation records.

In the past ~170 years (i.e., ca. 100 cal yr BP to present), fire activity at Lost Lake increased greatly, as evidenced by the large increases in CHAR, rising fire frequency, and high levels of disturbance in the pollen record. Recent evidence of fires in coastal BC is hard to establish because charred bark is difficult to see in the field, and fire scars can heal over quickly (Hemstrom and Franklin, 1982). However, the presence of *P. menziesii* in older, naturally regenerated stands is often used as an indicator of periodic fire because it is a shade-intolerant, early seral species that commonly regenerates after fire events (Green et al., 1999). The stands surrounding Lost Lake are overwhelmingly composed of even-aged *T. heterophylla* and *T. plicata* because of logging activity during the past two centuries. These stands are not useful for determining long-term fire histories because they are not sufficiently old. However, stands within the Capilano watershed that include *P. menziesii*, and younger stands which clearly established after fire, have both been used in the past to reconstruct recent watershed fire history (Green et al., 1998).

Green et al. (1999) indicated that the current fire incidence (FI; the interval between fires that have burned *somewhere* in a study area, but not necessarily in the same location (Kilgore, 1981)) is approximately 80 – 90 years within the Capilano watershed, meaning that a significant fire has burned within the watershed almost every century. This timing aligns with the current recorded fire incidences on Vancouver Island (Schmidt, 1970), in the North Shore Mountains (Eis, 1962), and in Mount Rainier National Park (Hemstrom & Franklin, 1982). The highest recorded fire frequency in the Capilano watershed during this zone is believed to have occurred between 1920 and 1930 AD, when extensive railway logging took place (Green et al., 1999).

Historically, experts have suggested that the main disturbance regime of CWH forests is high-severity, stand-replacing fires every 250 – 350 years (Agee, 1993, 1997; Green et al., 1999). Similarly, average mFRIs in the low-elevation western hemlock forests along the Oregon Coast Range are approximately 230 years for the past 3000 years (Long et al., 1998). However, more recently, it has been posited that canopy gap dynamics have been a larger driver of forest disturbance in wet coastal BC forests (Daniels & Gray, 1996) and that mFRIs may be longer. A key indicator is the high percentage of fire-intolerant western redcedar in these forests, which indicates stand-replacing fires may not have been as frequent as previously thought. Green et al. (1999) found that pre-colonial-settlement mFRIs in the Capilano watershed were approximately 350 years, while post-settlement time since last fire (TSLF, a measure of the years since the last recorded fire event) at individual sites ranged from 310 to 850 years, which is significantly longer than Lost Lake's TSLF of 177 years. Importantly, Green et al. (1999) focused predominantly on Douglas-fir dominated stands on southern aspect slopes, which topographically are at a greater risk of burning than the north-facing, hemlock-cedar-dominated Lost Lake (Gavin et al., 2003).

Fire return intervals are dependent on many different factors (temperature, precipitation, weather, fuel availability and conditions, slope), and thus even a moderate-sized watershed can have a range of mFRIs (Agee, 1993). Lost Lake's mFRI for the past ~3000 years is 416 years, which is longer than the generally agreed upon average, but still in the range of those found in the Capilano watershed (Green et al. 1999).

Logging and Wildfire in the Current Era

Land-use practices in the watersheds of North Vancouver were frequently altered throughout the 1800s and 1900s as private landowners claimed land that traditionally belonged to the Musqueam First Nation, logged the region, and clashed over land claims. Logging operations were permanently halted in 1994, although logging activity still occurs adjacent to the watersheds in the Indian River, Furry Creek, and Britannia Creek drainages (Walsh, 1999; Green et al., 1999). While these drainages are outside the boundaries of Metro Vancouver's water supply area, the use of industrial machinery is a potential concern for wildfire ignition.

Visible evidence of logging at Lost Lake is apparent in the road access, the burnt stumps surrounding the lake, and the log dump on the north end of the lake. Historical documents (and a large tree named Bigfoot near the Lost Lake entrance) indicate that pre-harvested stands were dominated by large *T. plicata*, which were highly sought after for logging (Morton, 1970). Despite the Seymour Watershed and WSA's generally wet climate, recent dry conditions and human activity have been the dominant cause of fire activity over the past hundred years during periods of logging, which caused large fires in the 1920s and 1930s (Green et al., 1999). However, most fire activity did not occur in the CWHvm zone that comprises much of the WSA.

No fires over 10 ha have been observed in the WSA since 1980, and in the past 40 years, most fires in the WSA have occurred in drier biogeoclimatic zones within the Coast Mountains (van der Kamp, 2016). The Elaho Valley Fire of 2015 burned a 125 km^2 area north of Squamish, and in the same year, the western hemlock forests of Olympic National Park experienced an 11 km^2 fire. Overall, however, fire activity in moist coastal forests has decreased over the course of the 20^{th} century, likely because summer precipitation in these areas has increased over this same time period (Meyn et al., 2012).

The top 40 cm (~200 years) of the Lost Lake core provides evidence of the greatly increased human activity in the region that has occurred since colonial settlement. Settler activity within the watershed intensified disturbance via logging, mining, slash burning, and sediment redeposition, which likely contributed substantially to the high CHAR beginning at 105 cal yr BP (AD 1845) and continuing to the present day. CHAR is, at its highest point, ~400 pieces/cm^2/year, at AD 1933, which falls directly within the reported high period of logging and wildfire activity within the watersheds (Green et al., 1999). The mFRI continued to increase throughout Zone LL-6b, with the final recorded fire episode occurring at 105 cal yr BP (mFRI = 836 years). The increasing mFRI aligns with the inferred LIA decrease in fire activity, but the final fire episode is clearly not entirely accurate, as Metro Vancouver has extensive evidence of logging-related fire activity more recently than 1845, as evidenced by the observed increases in CHAR since 1845 AD.

Pinus displayed a distinct increase between approximately AD 1880 and AD 1960, which may be partially attributed to increased fire frequency in the surrounding area (lodgepole pines rely on fire to regenerate, and are often a pioneer species after fire events because of the favorable soil and sunlight conditions (Meidinger & Pojar, 1991)). The Pinus PAR increase may also be due to the overall high amount of pollen produced by pines (Allen et al., 1999). Distinct increases in *Alnus* and *Pteridium* suggest that disturbance also increased in the forests around Lost Lake. Increases in the shrubs and herbs *Poaceae, Rosaceae,* and *Ericaceae* may represent increased understory growth due to gaps in the stand, either from logging, fire, or other disturbances. *Cupressaceae*, *P. menziesii* and *T. heterophylla* PARs and percent abundances decreased between ca. AD 1920 and AD 1940, most likely a direct result of logging activity.

4.1.2. Management Implications

Looking ahead to the next century, watershed management in the LSCR has the potential to change substantially as vegetation, fire activity, and natural ranges of variability are altered by climate change (Hallett & Walker, 2000). Future climate simulations for the Metro Vancouver region project temperature increases of up to 3°C by the 2050s with rates of warming between 0.1°C to 0.6°C per decade (Wang et al., 2016; Metro Vancouver, 2018). Summer months are projected to have the highest rates of warming and up to a 20% decrease in precipitation (Metro Vancouver, 2018). Even with the most conservative precipitation decrease estimates of -14% by the 2080s for northwestern North America, the associated reduction in cloud cover may have significant impacts on subsurface water storage, evaporative demand and, consequently, forest fires (Mote & Salathé Jr., 2010; McKenzie et al., 2004). Changes in temperature of these magnitudes don’t appear to have occurred at Lost Lake during the xerothermic period, but would be analogous to temperature and precipitation changes reconstructed for the xerothermic period at sites further inland in southwestern BC (Pellatt et al., 2000; Rosenberg et al., 2004).

One important implication of this warming and drying could be a vegetational shift away from the moisture-loving *T. plicata* and *T. heterophylla* and a potential increase in fire-adapted species such as *P. menziesii*, perhaps moving towards a composition more similar to the drier subvariants of the CWH zone or the coastal Douglas fir (CDF) zones that currently exist in the rain shadow of Vancouver Island. Sites in the cool, moist variants of the CWH zone, such as I currently observe near my site, became more Douglas-fir dominated during the xerothermic period (Hebda, 1995).

My pollen record suggests that the period of highest temperature at Lost Lake occurred when *P. menziesii* was at its highest abundance (18-26% between ca. 12,200 and 9200 cal yr BP; Fig. 6), which was several thousand years before the commonly stated xerothermic interval (9500-7000 cal yr BP). *P. menziesii* forest likely propagated at this time due to a combination of the inferred dry climate in the early Holocene (Walker & Pellatt, 2003) and the strong summer insolation (Fig. 2) causing very warm summer conditions suitable to growth of the species. The low fire frequency and CHAR values during this time (Fig. 8) can perhaps be explained by lower overall biomass post-deglaciation, or Lost Lake's proximity to the coast giving it a "moisture buffer" (Cwynar, 1987; Galloway et al., 2009) that prevented it from being as fire-prone as more inland sites. The appearance of *T. heterophylla* beginning ca. 12,000 cal yr BP, and its expansion beginning at ca. 11,500 cal yr BP, provides additional evidence of coastal moisture moderation, as it suggests that even during a time of high summer insolation the Lost Lake site remained able to sustain a species that requires cool and/or moist conditions (Meidinger & Pojar, 1991).

Considering the period of 12,200 - 9100 cal yr BP is inferred to be the warmest time in Lost Lake's Holocene history, if summers in coastal BC become warmer and drier in the coming century it can be inferred that forests may again progress towards an assemblage with a stronger *P. menziesii* presence than currently. However, it must be noted that this timespan of *P. menziesii* dominance was during a period of much higher summer solar insolation than present (Fig. 2), which was a dominant factor in early Holocene warmth in southwestern BC (Gavin et al., 2011).

A major difference between Lost Lake's *P. menziesii*-dominated assemblage in the early Holocene and a potential future assemblage is the absence of *T. plicata* in the early Holocene versus its current abundance in the LSCR and around Lost Lake. *T. plicata* prefers high-moisture environments and is sensitive to fire (Meidinger & Pojar, 1991), as evidenced by its decline at Lost Lake during the FVFP at ca. 2200 cal yr BP (Fig.8). While *T. plicata* establishes well in mineral soils post-fire, it may slowly decrease in abundance if summer precipitation declines in the Metro Vancouver area as expected (Metro Vancouver, 2018). Currently, western redcedar is abundant on southern Vancouver Island in the CWHdm, CWHxm and CDF zones (Meidinger & Pojar, 1991; Hebda, 1997; Brown et al., 2019), confirming that it fares well in mixed species stands with western hemlock and coastal Douglas fir. Thus, while western redcedar may experience a decrease in abundance if precipitation decreases and/or fire activity increases, it is unlikely to disappear from the assemblage altogether. The dominant species in the current assemblage, western hemlock, has been present at Lost Lake at near modern levels since ca. 11,000 cal yr BP, indicating it is highly resilient to the disturbances experienced in the watershed throughout the Holocene and is unlikely to shift a great amount.

Increases in temperature and dryness may affect understory and arboreal species that currently grow at Lost Lake. The modern understory, dominated by moisture- and shade-tolerant zonal species such as salal, red huckleberry, deer fern and sword fern, may be altered if the system becomes drier. Although salal and red huckleberry still thrive in dry maritime environments, the shrub layer could become more dominated by vine maple and bracken fern, as is typical of drier maritime subzones of the CWH zone (Pojar, Klinka & Demarchi, 1991). Bigleaf maple may become more common in light of warming and drying, as could alder if disturbance becomes more common. (Pojar, Klinka & Demarchi, 1991).

The current closed canopy forests surrounding Lost Lake indicate a higher fuel load and potential for high-severity, stand-replacing crown fires. Decreasing precipitation in the coming decades will likely dry fuels out, resulting in higher burn likelihood in the event of an ignition. Alternatively, a transition to intermediate precipitation and moderate to high fuel loads could result in mixed-severity fire regimes, causing a patchwork distribution of ground and crown fires that results in variable tree mortality, similar to what is seen in montane forests today (Agee, 1998).

I consider multiple hypotheses for the cause of Lost Lake's perceived resilience to fire during its warmest period. The first is that the coastal setting of the LSCR created a climatic buffering effect as the cool, moist air of the Pacific Ocean dampened the insolation-driven summer temperature variations, somewhat protecting the site from extreme temperature and drought (Cwynar, 1987; Renssen et al., 2005; Galloway et al., 2009). The resulting wetter conditions along the coast may have allowed the vegetation around Lost Lake to withstand warmer temperatures due to the continuous moist climate, and this maritime influence is likely the reason drought-intolerant taxa such as *T. heterophylla* and *Alnus* were able to grow during this period (Krajina, 1969). A second possibility is that high winter precipitation, coupled with the ability of root systems to access groundwater sources, provided a hydrologic buffer during periods of drought. As such, the vegetation around Lost Lake had ample groundwater sources even during warm, dry summers (Hahm et al., 2019; Rempe et al., 2022, Preprint). I also note that the timing of peak fires in my record occur during the Neoglacial period at ca. 2500 – 1500 cal yr BP, suggesting that human activity, which is known to have been extensive during this time, may have contributed to high fire frequency in an otherwise cool climate (Lepofsky et al., 2009).

Similarly to the inferred anthropogenic burning during the FVFP, the potentially intensive effects of human activity on fire frequency may have implications for watershed management as population density around the watersheds continues to increase. Housing developments in North Vancouver already border the southern boundary of the LSCR, and the nature of the COVID-19 pandemic has increased the number of people using trails and outdoor spaces (de Burger, 2020; Beery et al., 2021). With the population of the Lower Mainland projected to increase to 4.1 million by the 2040s (Ip & Lavoie, 2020), the risk of human-caused ignitions will increase around the watersheds.

4.2. Sources of Uncertainty

This study provides a first attempt at estimating the wildfire activity and vegetational changes throughout the Holocene period in a very moist coastal western hemlock forest. However, multiple sources of uncertainty are present, including model limitations, sampling resolution, and seasonal biases.

4.2.1. Age Model Uncertainties on the Lost Lake Core

Almost all species show increases in PAR in the top 40 cm (last 200 years) of the core, and CHAR is also at its highest during this time. These increases may be attributed to the younger sediments having less time to be reworked, and thus more individual grains being identifiable. However, similar results have also been observed in the surface cores of other studies (Murphy et al., 2019), and may be attributed to issues surrounding ^{210}Pb dating of surface sediments versus dating using only an AMS-^{14}C-based age model. The Bacon software was rerun twice, first using only AMS-^{14}C dates, and then using AMS-^{14}C dates and the 1966 maximum terrestrial inventory estimated from ^{137}Cs measurements (Flett Research Ltd., 2020). The resulting age models differed from the original (i.e., the model using both ^{210}Pb and AMS-^{14}C dates) most notably in the final 41 cm (approximately 150 years) of the core (Figure 10). These final 41 cm represent the surface core, from which ^{210}Pb dates (Table 1) were taken and modelled using a CRS model. The sediment accumulation rate and PAR in the original age model were noticeably elevated between ca. 1700 and 150 cal yr BP (Fig. 10), which is the period of time between the final ^{210}Pb-dated sample at ca. 152 cal yr BP and the first AMS-^{14}C at ca. 1700 cal yr BP. However, in the models using only AMS-^{14}C dates and ^{137}Cs + AMS-^{14}C dates, the sediment accumulation rates declined slightly at ca. 1700 cal yr BP and then remained stable for the remainder of the core. The timing of the Lost Lake CHAR increase is similar, following the break between ^{210}Pb and AMS-^{14}C dates at ca. 150 cal yr BP. The average PAR (terrestrial pollen + spores) was 29% higher in the final 150 cal yr BP in comparison to the rest of the core, while CHAR was 161% higher in the final 150 years (Fig. 10). Despite the difference in sampling resolution between pollen (mean ~450 years/sample) and charcoal (mean temporal resolution 44 years/sample) in the Lost Lake composite core, fluctuations of CHAR and PAR are still visually comparable (Fig. 10). Fluctuations in these proxies often correspond with changes in sedimentation rate, indicating that watershed processes were likely the dominant driver of sediment deposition. Minor differences in the timing of CHAR peaks versus PAR peaks are likely a combination of the higher resolution of charcoal sampling, and natural processes that affect charcoal and pollen deposition differently.

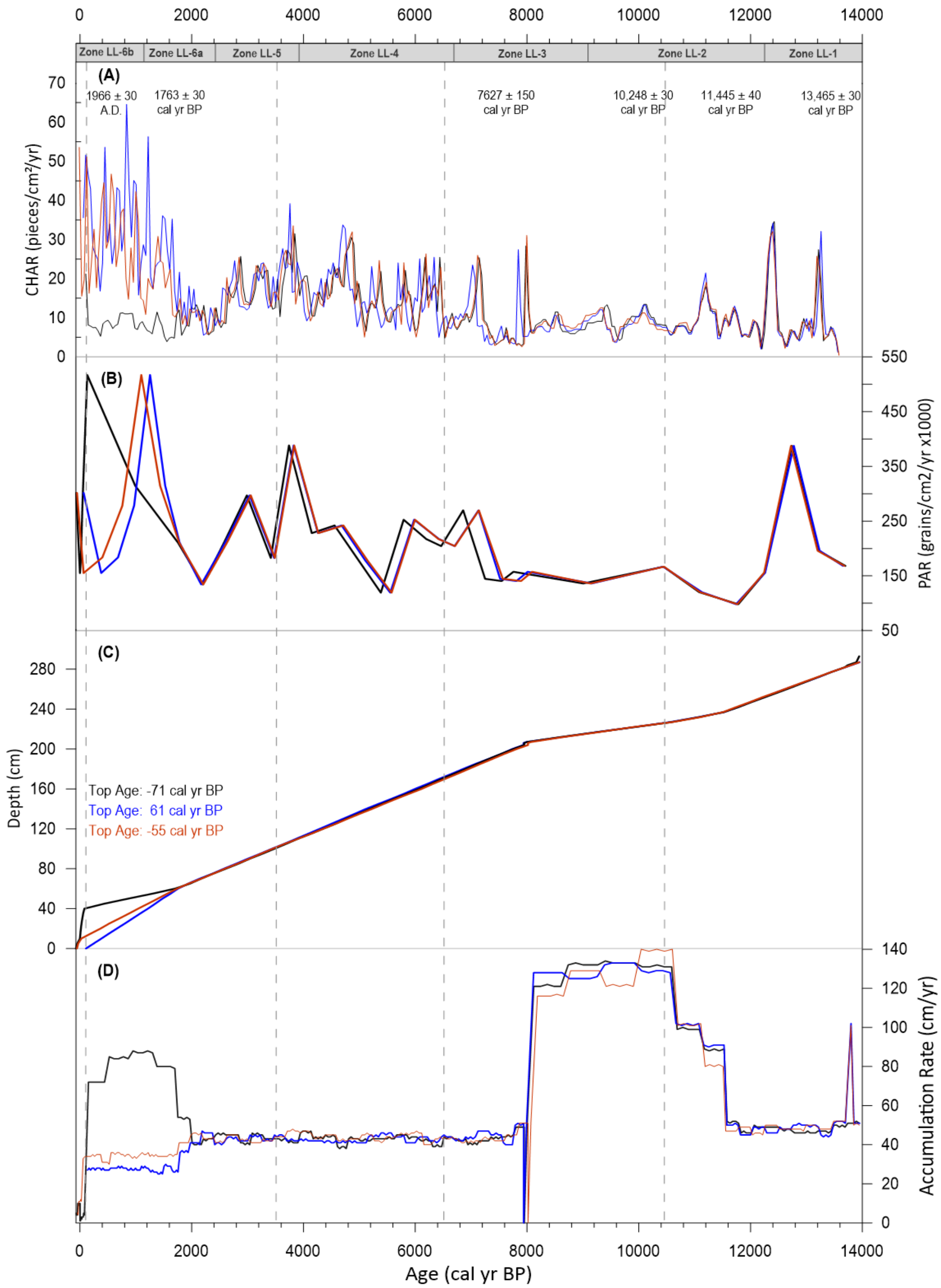


Figure 10. Comparison of Lost Lake CHAR, PAR, age model, and sedimentation rate between original age model (using ^{210}Pb and AMS-^{14}C), AMS-^{14}C-based age model, and age model using ^{137}Cs and AMS-^{14}C.

(A) Interpolated CHAR of the Lost Lake core using original age model (black), AMS-^{14}C-based age model (blue), and age model using ^{137}Cs and AMS-^{14}C (red). Note that the initial ~50-100 years have been removed from each record for easier visual comparison between CHAR and PAR records. (B) total PAR for all terrestrial pollen and spores in the Lost Lake core using original age model (black), AMS-^{14}C-based age model (blue), and age model using ^{137}Cs and AMS-^{14}C (red). (C) Black line depicts original age model using ^{210}Pb and AMS-^{14}C dates, blue line represents age model using only AMS-^{14}C dates, and red line represents age model using ^{137}Cs and AMS-^{14}C dates. (D) Black line depicts sediment accumulation rate for original age model, blue line depicts sediment accumulation rate for age model using only AMS-^{14}C dates, red line depicts sediment accumulation rate for age model using ^{137}Cs and AMS-^{14}C dates. Vertical dotted grey lines represent core breaks. Bacon-calculated top ages of each age model, coded by color, are listed in panel (C).

Modern CHAR values calculated using the original age model could be interpreted as high-severity fire events or as overall higher fire activity from 150 cal yr BP to present. Similarly, PAR increases from 150 cal yr BP to present could be interpreted as high levels of pollen rain or increased disturbance in the watershed. However, these high rates are most likely a product of the high sediment accumulation rate calculated using the original age model. Despite the discrepancies between the three age models in the final 200 years of the core, the remainder of the age models (i.e., 150 cal yr BP to approximately 13,900 cal yr BP), and thus, sedimentation rates, were nearly identical (Fig. 10).

To further understand the effects of various age models on fire history characteristics, the full CharAnalysis results from each age model were compared (Fig. 11, Table 5). While CHAR and background CHAR remained similar between models, fire frequency results were more varied, indicating that choice of age model has a significant effect on certain fire history characteristics. These age model differences could affect attempts to use not only fire frequency, but mFRI, TSLF and fire episodes to understand past fire behavior, as previously seen in the work of Murphy et al. (2019). Pollen results would also most likely be affected by age model choice, as PAR is directly influenced by accumulation rate, which is in turn chronologically linked to an age model.

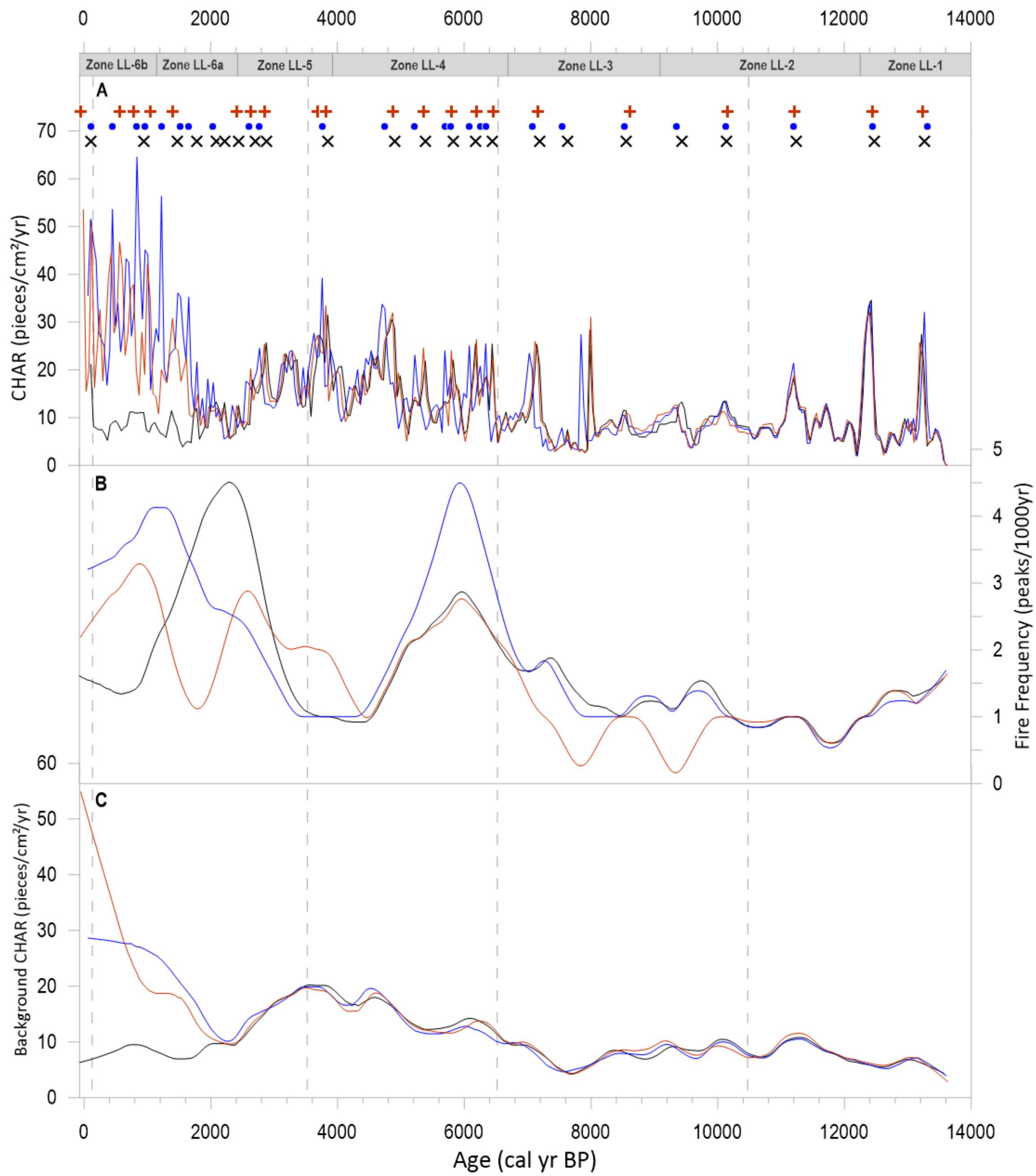


Figure 11. Comparison of Lost Lake CHAR, fire frequency, background CHAR, and fire events between original age model (using ^{210}Pb and AMS-^{14}C), AMS-^{14}C-based age model, and age model using ^{137}Cs and AMS-^{14}C.

(A) Black X symbols mark fire episodes in original age model; blue dots mark fire episodes in AMS-^{14}C-based age model; red crosshairs mark fire episodes in age model using ^{137}Cs and AMS-^{14}C; black line represents interpolated CHAR of the Lost Lake core using original age model; blue line represents AMS-^{14}C-based age model; and red line depicts age model using ^{137}Cs and AMS-^{14}C. (B) Fire frequency of the Lost Lake composite core using original age model (black line), AMS-^{14}C-based age model (blue line), and age model using ^{137}Cs and AMS-^{14}C (red line). (C) Background CHAR of the Lost Lake composite core using original age model (black line), AMS-^{14}C-based age model (blue line), and age model using ^{137}Cs and AMS-^{14}C (red line). Vertical dotted grey lines represent core breaks.

Table 5. Comparison of fire history characteristics between the original age-depth model, the AMS-^{14}C-based age-depth model, and the ^{137}Cs-based age-depth model in the Lost Lake composite core.

Metric	Original composite core record	AMS-^{14}C-based composite core record	^{137}Cs-based composite core record
Fire Episodes	23	26	21
TSLF	144 years	174 years	15 years
mFRI[a]	598 (466 to 735)	528 (390 to 660)	664 (490 to 860)
CHAR[b]	14.1 (1.2 to 397.4)	14.3 (1.2 to 64.6)	14.1 (0.4 to 184.9)
Background CHAR[b]	10.2 (4.3 to 20.2)	12.4 (4.3 to 28.6)	12.9 (2.8 to 54.9)
Fire Frequency[c]	1.7 (0.7 to 4.5)	1.9 (0.5 to 4.5)	1.5 (0.2 to 3.3)

[a] Mean and range of fire return interval (years).
[b] Mean and range of charcoal accumulation rate (CHAR; pieces/cm^2/yr).
[c] Mean and range of fire frequency (fires/1000yrs).

The use of ^{210}Pb in pollen and/or charcoal-based paleoclimatology reconstructions appears to be a matter of preference, as many studies use AMS-^{14}C-based age models (Wainman & Mathewes, 1987; Brown & Hebda, 2002a,b; Gavin et al., 2001, 2013; Leopold et al., 2016) while others also include ^{210}Pb dating (Gavin et al., 2003; Sugimura et al., 2008; Walsh et al., 2008; Brown et al., 2019; Murphy et al., 2019; Appendix A). My results suggest that modern fire and pollen behavior should be interpreted with caution if choosing to analyze results with dates of stronger temporal resolution at the top of the core (i.e., including ^{210}Pb dates in the surface core with comparatively few radiocarbon dates in the composite core).

4.2.2. Other Sources of Uncertainty

Coarse sampling intervals of pollen in many studies including ours (median 434 years) complicates the determination of decade-to century-scale changes, as many old rainforest trees can live hundreds of years and are able to withstand short periods of climatic change (Meidinger & Pojar, 1991; Agee, 1993). As a result, an atmospheric disturbance like fire is more likely to catch the short-term pulse of drought conditions (for example, during the FVFP), which may not be reflected as clearly in the pollen record

(Hallett et al., 2003). A higher-resolution analysis of core sections reflective of short periods of climatic change (i.e., the FVFP, MWP and LIA in the past ~1000 years) could provide further insight into what, if any, vegetative changes occurred during these periods. Regional climatic changes can also be obscured in the pollen record because of natural vegetational successional patterns, local topography and soil development, which can obscure climatic effects seen in the pollen and charcoal records (Mathewes & Rouse, 1975). The nature of my dating method creates challenges with pinpointing individual fires or fire episodes as well; however, my data for the past ~200 years align with similar studies along the coast which observe abrupt changes to fire and vegetation records shortly after colonial habitation, when settlement-related land clearing likely began increasing (Heusser, 1983; Pellatt et al., 2015; Murphy et al., 2019).

Another factor which affects charcoal results is the initial input parameters of the CharAnalysis program. Notably, smoothing of the fire-related peaks component of CHAR (C_{peak}) is used to estimate fire frequency and, conversely, mFRI. The model used in the final charcoal results applied a C_{peak} smoothing window of 1000 years. The potential for different sized C_{peak} smoothing windows to affect results was investigated by modelling fire frequencies under various window widths (Figure 11). While clear distinctions are evident between various window widths, it appears that the chosen 1000-year smoothing window provides an acceptable middle ground between the noisy data produced by smaller windows and the overly smoothed data produced by larger smoothing windows. Initial parameter selection has important effects on final results and should be approached with an understanding of the potential impacts on model robustness.

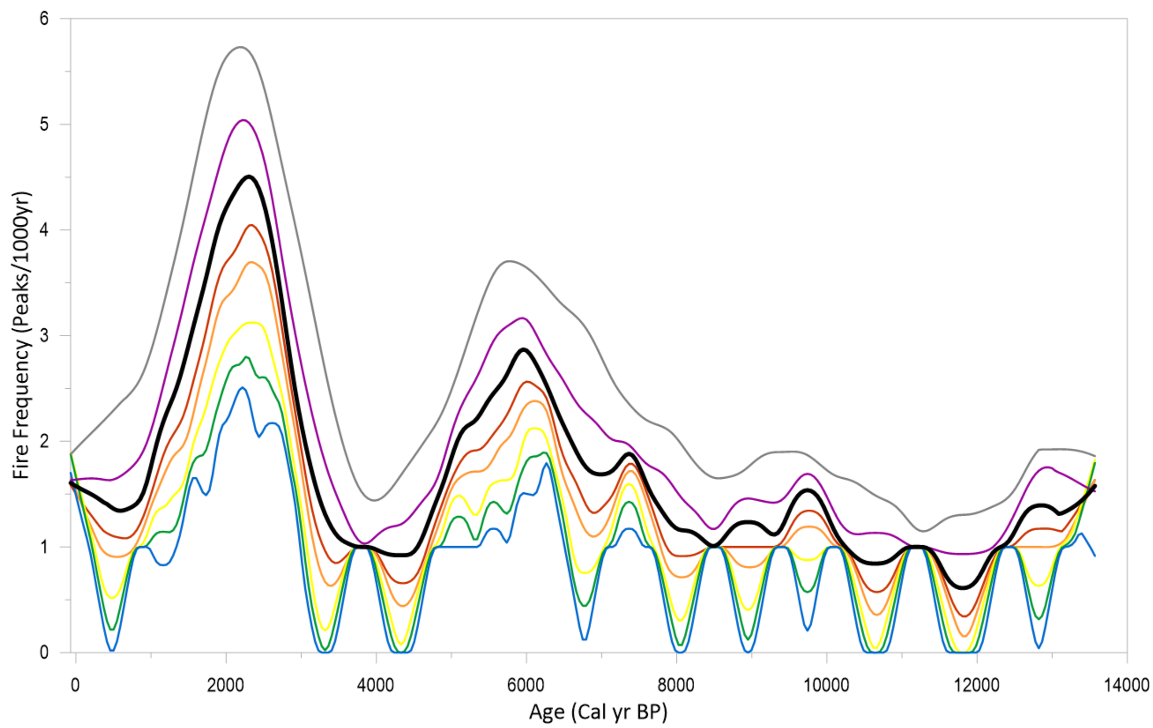


Figure 12. Fire Frequency of the Lost Lake composite core under various C_{peak} smoothing window widths.
Window widths used are 1500 years (grey); 1200 years (purple); 1000 years (black; used in final results); 900 years (red); 800 years (orange); 700 years (yellow); 600 years (green); and 500 years (blue).

A consideration when consulting paleoproxy reconstructions is the seasonal sensitivities of the proxies used. For example, tree ring growth rate data are thought to be more strongly influenced by summer and spring seasonality (i.e., temperature and precipitation affecting soil moisture content (Barron & Anderson, 2011)), whereas lake oxygen isotope data are primarily influenced by precipitation during winter and fall (Steinman et al. 2012). Similarly, pollen is a seasonally-sensitive proxy due to its interannual variability in production, and annual variations in the record can be hidden in lower-resolution cores (Pellatt et al., 2001). Charcoal is a seasonally sensitive proxy because wildfires are more likely to occur during the drier periods of summer and early fall, resulting in proxy reconstructions that may be biased towards summer-fall precipitation despite the possibility of continued moist conditions reflected in the vegetation record (Sikes et al., 2013). The importance of below-ground root water storage capacity on ecosystem drought resilience is still not fully understood, but the ability of my site to access subsurface water even during dry summer conditions could be a driver of the observed disconnect between pollen suggesting moist conditions and charcoal records suggesting dry conditions (Hahm et al., 2019, 2022; Rempe et al., 2022, Preprint).This concept of subsurface storage capacity has not been explored in the LSCR, and could provide further insights into the resilience of ecosystems in the region to future climatic change.

This study fills a gap in the regional paleoclimate record by focusing on a wet, cool site in south coastal BC. However, considering the management implications of my findings requires an understanding of site-specific characteristics that are not covered by the broad-scale BGCZ system. Even when studying a site at the subzone level (i.e., the very wet maritime subvariant of the Coastal Western Hemlock zone), local attributes such as topography, elevation, aspect, and soil composition can affect vegetation growth and create microclimates that require specialized management. Evidence of these local, site-specific controls can be seen in the Lost Lake record during the xerothermic (ca. 9000-7500 cal yr BP) when Lost Lake was cooler than other regional sites, and at ca. 6000 cal yr BP when fire frequency at Lost Lake reached a peak that was not observed at other sites in southwestern BC. Thus, applying these findings to other sites in the CWHvm1 zone should be done with the understanding that local attributes can strongly control the response of systems to large-scale changes in climate.

Chapter 5. Conclusion

This study presents new sedimentary pollen and charcoal records from Lost Lake to fill an information gap in the paleoclimate history of CWHvm1 forests in the coastal Lower Mainland of British Columbia. My reconstructions were created using pollen analysis, strong chronological controls (^{210}Pb and AMS-^{14}C age constraints) and high-resolution charcoal analysis (~44-year intervals). My record differs from previously proposed climatic change over the past 13,000 years in southwestern British Columbia. My primary hypothesis was that a distinct xerothermic period, signaled by high fire frequency and potential associated vegetation changes, would be visible in the record between ca. 9500 and 7000 cal yr BP, similar to other regional sites on the Pacific coast of North America (Heusser et al., 1980, Hebda, 1983, 1995; Pellatt & Mathewes, 1997; Pellatt et al., 2000; Brown & Hebda, 2002, 2003; Walker & Pellatt, 2003; Rosenberg et al., 2004; Lacourse, 2005; Brown et al., 2019). Instead, my site displayed signs of warming in the pollen assemblage between ca. 11,500 and 9000 cal yr BP, while fire activity did not substantially increase. I present this interpretation as an addition to the regional understanding of post-glacial climatic, vegetation and wildfire history studies in coastal British Columbia and Washington.

Table 6. **Summarized post-glacial vegetation, climate and fire activity at Lost Lake.**

Lost Lake Vegetation and Climate Summary		
Vegetation Relative to Modern	**Climate Relative to Modern**	**Fire Activity Relative to Modern**
Modern	Modern	Modern
Modern	Modern winters, warm dry summers	High
Western hemlock-Cedar-Spruce	Cool, moist	Low
Western hemlock-Cedar-Alder	Cool, moist	Moderate-high
Western hemlock-Fir-Spruce	Cool, moist	Moderate
Douglas fir-Alder-Western hemlock	Warming, dry	Low-moderate
Spruce-Fir-Mountain hemlock	Cool, dry	Low
Lodgepole pine, shrubs	Cool continental, dry	Low; isolated fires

Cool — Warm
Less fire — More fire

Temperature and fire activity scales are relative to modern.

During the transition from the last glacial period into the Holocene period (13,900 – 11,700 cal yr BP) the area around Lost Lake had been recently deglaciated, and had an open, shrubby landscape that was quickly colonized by *Pinus contorta*. Evidence of a period of warming associated with the Bølling-Allerød event (~14,700 – 12,800 cal yr BP) was not observed in the Lost Lake record. However, a Younger Dryas (12,800 – 11,500 cal yr BP) climate signal may have been experienced, as the climate was cool and dry and fire activity was moderate to low, perhaps with isolated and infrequent summertime fires. Coniferous forests composed of *T. mertensiana, Picea* and *Abies* soon overtook the landscape, forming closed-canopy forests within a boreal climate.

The end of the Younger Dryas may have occurred earlier at Lost Lake (ca. 12,000 cal yr BP ± 576 years) than in other regional studies, with a warming, drier climate at this coastal site causing a rise in moderately open *P. menziesii*-dominated forests. Disturbance was high, and fire activity increased. This time period is inferred to have been the warmest point in Lost Lake's record, predating the xerothermic interval (ca. 9000-7500 cal yr BP) by up to 1500 years. *T. heterophylla* presence began increasing ca. 11,000 cal yr BP, perhaps signifying moist winters despite the higher fire frequency during summer.

The xerothermic interval (ca. 9000-7500 cal yr BP) is not strongly documented in the Lost Lake record. Disturbance remained high and fire activity rose slightly (reaching a maximum of 1.77 fires/1000yrs during this period) but did not reach peak levels observed in other studies from the region (Heusser et al., 1980, Hebda, 1983, 1995; Pellatt & Mathewes, 1997; Pellatt et al., 2000; Brown & Hebda, 2002, 2003; Walker & Pellatt, 2003; Rosenberg et al., 2004; Lacourse, 2005; Brown et al., 2019). *P. menziesii* was overtaken by forests of western hemlock, cedar, spruce, and later, true fir. Precipitation was inferred via the pollen assemblage to be moderate to high, suggesting that while summers were warm, dry, and fire-prone, winters remained moist.

Fire activity at Lost Lake reached a local peak of ~3 fires/1000yrs at ca. 6000 cal yr BP, approximately 1500 years after the xerothermic period. Climate at Lost Lake during the mid-Holocene was cool and moist, with higher fire frequency results likely produced by a combination of increased anthropogenic landscape modification and elevated charcoal influx due to higher precipitation relative to the early Holocene.

Cooling ca. 7000-6000 cal yr BP, several thousand years before the beginning of the Neoglacial interval (ca. 4500 cal yr BP), brought glacial readvances throughout western North America. The vegetation assemblage at Lost Lake reflected the cooling climate during this time interval, and western hemlock-western redcedar closed forests dominated. Fire frequency declined following ca. 6000 cal yr BP at Lost Lake, but regional fire activity may have remained high. In the late Holocene (3000 cal yr BP onward), the vegetation assemblage at Lost Lake signaled a continuous temperate and moist climate, but fire frequency peaked during the Fraser Valley Fire Period at ca. 2400 – 1300 cal yr BP. Plant assemblages were likely similar to modern day (possibly with slightly lower western redcedar presence), but prolonged and frequent droughts may have occurred in summer due to a stronger summer Pacific High and weakened winter Aleutian Low (Fritz, 1996). This disconnect between fire and vegetation assemblage suggests that non-climatic factors such as increased anthropogenic burning were influencing fire activity, or that subsurface water storage allowed for increased resilience to summer drought conditions (Lepofsky et al., 2005; Hahm et al., 2022, preprint).

During the Neoglacial and after the FVFP (post-1300 cal yr BP), the region around Lost Lake may have experienced a brief period of moderate warming, during which time regional biomass burning increased. Vegetation assemblages likely did not experience any substantial changes due to the short-lived nature of the warming. The Little Ice Age (~500 – 100 cal yr BP) is observable in the Lost Lake record via noticeable dips in CHAR and fire frequency, suggesting a brief return to cool conditions, although again the dominant arboreal species likely withstood this interval with little change.

The final 170 years display high rates of disturbance and regional fire, almost certainly because of increased logging and wildfire in the watersheds during this time. Assemblage is inferred to have still been western hemlock-cedar dominated closed forests. A distinct decrease in both species as well as Douglas fir coincides with an increase in alder pollen in the past ~100 years, representing the effects of logging in the watersheds.

The interpreted climatic changes of this study indicate that natural vegetational succession, disturbance, broad-scale climatic changes, and human impact all contributed to the development of the current conditions found at Lost Lake. Ample evidence of human habitation since at least the late Pleistocene has been found along the north and central coast of British Columbia (Hoffman, Lertzman & Starzomski, 2017; Fisher et al., 2019; McLaren et al., 2019; Fedje et al., 2021); however, fewer data exist regarding human habitation on the southern coast. Indigenous peoples were likely living on the land surrounding Lost Lake for much of the record (Hoffman, Lertzman & Starzomski, 2017; Fisher et al., 2019; McLaren et al., 2019; Fedje et al., 2021), and consequently, fire activity was likely influenced by their behavior. However, until further archaeological evidence of human habitation and fire use in the LSCR is found and quantified, I can only infer the impacts of humans on paleo fire regimes at my site. The magnitude of human versus climate influence in shaping vegetation regimes is notably difficult to quantify (Ryan et al., 2013). However, at broad scales, the geographic localization of human disturbances generally results in human effects falling secondary to those of climate (Whitlock et al., 2015). The observed disparities between inferred vegetation and fire activity at the Lost Lake site suggests that local ecosystem shifts were not always directly related to large-scale climatic changes. From approximately 6000 cal yr BP to present, the vegetation surrounding Lost Lake has reflected a near-modern assemblage despite multiple climatic events (The Neoglacial period (4500 cal yr BP to present), FVFP (2400 – 1300 cal yr BP), MWP (1100 – 700 cal yr BP) and LIA (600 – 100 cal yr BP)) and significant changes in fire frequency during the same time period.

The differences between pollen percentage, pollen influx, and charcoal results also show the importance of considering multiple data sources when reconstructing paleoclimate. Fire history characteristics can provide valuable data on the spatial and temporal variability of historic wildfire. Fire frequency data provide insights into continuous trends in fire occurrence over time, while mFRI demonstrates that fire return intervals are not static (Whitlock et al., 2003), and can help managers assess the natural range of variability of fire disturbances within their lands (Swetnam et al., 1999). The pre- and post-settlement mFRIs at Lost Lake for the past 3000 years are higher than what was previously suggested for coastal western hemlock forests (Green et al., 1999; Long et al.,1998), suggesting that severe fire disturbances in this region may not have been as frequent as originally thought. Adding vegetation reconstructions from pollen analysis provides land managers with a means of comparing fire and vegetation responses to climatic changes throughout time. Significant fluxes in fire frequency or background charcoal influx can be deceptively suggestive of large-scale or abrupt ecosystem changes. However, consideration of the pollen percentage and influx records can confirm whether vegetation was equally affected or minimally impacted, which is crucial when considering overall effects of wildfire and disturbance on the ecosystem as a whole.

Based on my pollen and charcoal data and the province's projected shift towards drier conditions within the century, the future vegetation and fire regime at Lost Lake may change in the following ways: (1) The fire season will lengthen, potentially with less available moisture during summers; (2) Fire-adapted, drought-tolerant species such as *P. menziesii* and *Pteridium* will increase in abundance, and the ecosystem could shift towards a drier variant of the CWH zone; (3) Fuel availability will rise due to increased mortality of arboreal species, potentially increasing fire risk if not managed; (4) Canopy gaps may increase due to fire disturbance, creating growth opportunities for disturbance-tolerant and herbaceous/shrubby species such as *Alnus, Brassicaceae* or *Poaceae.* (5)

Continued population growth in the Greater Vancouver Area will increase the risk of ignitions in the WSA.

These data can assist land managers in understanding past climatic changes and ranges of variability when considering future changes to the watershed that may occur as climate change progresses. Local-scale controls (i.e., topography/elevation, weather, human influence, fuel availability) undoubtedly play a part in Lost Lake's paleoclimate history and should be considered when applying these results broadly to CWH forests on British Columbia's southern coast. The apparent resilience of the forests around Lost Lake suggests that they are able to withstand large amounts of disturbance without major changes in assemblage, which bodes well for the future of the watershed. This study should also be considered in the context of its limitations, including the uncertainty of AMS-^{14}C dates, the constraints of the CharAnalysis program, the resolution of my pollen analysis, and the fact that pollen grains were identified by a very careful geologist. Despite these limitations, my study provides a strong basis for understanding the variability of fire in relation to vegetational and climatic changes in the CWH zone of the Water Supply Area of Metro Vancouver.

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Appendix A.

Table of Previous Paleoclimatic Studies in Southwestern British Columbia and Western Washington

Table A.1. Previous studies of pollen and charcoal paleoclimatic reconstructions in southwestern British Columbia and western Washington.

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Lost Lake[1]	49.40°N 122.98° W	235	CWHvm1	Both	13400	11 ^{210}Pb + 5 AMS-^{14}C + Mazama, Bacon	Pollen: 10 Charcoal: 1	10% KOH, 10% HCL, HF, acetolysis	5% $(NaPO_3)_6$ & 6% H_2O_2, >125 um	mFRI, Fire Frequency (#/1000 yr)	This study
Marion Lake	49.18ºN 122.32º W	305	CWHdm	Both	12350	^{14}C Lia	5 & 10		70% alcohol, >250µm	Charcoal pieces	Wainman & Mathewes, 1987
Whyac Lake	48.40ºN 124.50º W	15	CWHvh	Both	10860	5 ^{14}C Lia	1	(Moore et al., 1991)	10% $(NaPO_3)_6$, 6% H_2O_2, >150µm & 150-500µm	CHAR	Brown & Hebda, 2002b

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Clayoquot Lake	49.12ºN 125.30º W	17	CWHvm	Charcoal	1800	20 ^{210}Pb cubic spline & 10 AMS-^{14}C Linear regression	1		10% $(NaPO_3)_6$, 6% H_2O_2, 150-500µm	Frequency (#/200 yrs)	Gavin et al., 2003
Pixie Lake	48.35ºN 124.11º W	70	CWHvm	Both	12990	7 ^{14}C Lia + Mazama	1	(Moore et al., 1991)	10% $(NaPO_3)_6$, 6% H_2O_2, >150µm	CHAR	Brown & Hebda, 2002b
Misty Lake	50.365 ºN 127.16 ºW	70	CWHvm	Pollen	14,150	6 ^{14}C	5-10	Following Faegri & Iversen, 1989			Lacourse, 2005
Two Frog Lake		3.59	CWHvh1	Pollen	13030	5 AMS-^{14}C + Lia	--	Following Faegri & Iversen, 1989, excluding HF			Galloway et al., 2007
Tiny Lake	51.117ºN 127.221º W	475	CWHvh1	Pollen	13815	4 AMS-^{14}C + Lia	8-12	Following Faegri & Iversen, 1989, excluding HF			Galloway et al., 2009

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Woods Lake	51.00ºN 127.161º W	2	CWHVH1	Pollen	13950	4 AMS-^{14}C + cubic spline	5-10	Following Faegri & Iversen, 1989			Stolze et al., 2007
Begbie Lake[1]	48.59ºN 123.68º W	187	CWHxm	Both	13800	7 ^{210}Pb + 10 ^{14}C + Mazama, Clam 2.1	Pollen: 10; Charcoal: 1	Standard (Moore et al., 1991), excluding HF	>150µm	mFRI, Fire Frequency (#/1000 yr)	Brown et al., 2019
Mike Lake[1]	49.17ºN 122.32º W	225	CWHdm	Pollen	2500	11 AMS-^{14}C	2-5	Following Faegri & Iversen, 1989			Pellatt, Mathewes & Clague, 2001
Surprise Lake[1]	49.32ºN 122.56º W	540	CWHmm	Pollen	13100	5 ^{14}C + Mazama, Li	5-10	5% KOH, HF, acetolysis			Mathewes, 1973
Chadsey Lake	49.07ºN 122.08º W	620	CWHdm	Charcoal	4258	9 ^{210}Pb CRS & 5 AMS-^{14}C & Mazama Bacon	1		5% $(NaPO_3)_6$ & 6% H_2O_2, >125 um	mFRI, Fire Frequency (#/1000 yr)	Murphy et al., 2019
East Sooke Fen	48.21ºN 123.40º W	155	CWHxm	Both	11700	4 ^{14}C Lia + Mazama	1	(Moore et al., 1991)	10% $(NaPO_3)_6$, 6% H_2O_2, >150µm & 150-500µm	CHAR	Brown & Hebda, 2002b

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Mosquito Lake Bog[1]	48.77ºN 122.12º W	~400	Western Hemlock	Pollen	11400	^{14}C + Mazama Li	10-20	10% KOH, 10% HCL, HF, acetolysis			Hansen and Easter-brook, 1974
Killebrew Lake Fen[1]	48.36ºN 122.54º W	88	Western Hemlock	Pollen	13700	5 ^{14}C + Mazama, Bacon	10	Following Faegri et al. (1995)			Leopold et al., 2016
Mt. Constitution C11	48.39ºN 122.50º W	660	Western Hemlock	Both	~7650	^{210}Pb polynomial & 4 AMS-^{14}C Li/ regression + mazama	Charcoal: 0.25 & 0.5: Pollen: 0.25 & 1	Following Faegri & Iversen, 1989	10% $(NaPO_3)_6$, 6% H_2O_2, 150-500µm	CHAR	Sugimura, Sprugel, Brubaker, & Higuera, 2008
Mt. Constitution C32	48.39ºN 122.50º W	660	Western Hemlock	Both	~3800	^{210}Pb polynomial & 1 AMS-^{14}C Li/ regression	Charcoal: 0.25 & 0.5; Pollen: 1 & 2	Following Faegri & Iversen, 1989	10% $(NaPO_3)_6$, 6% H_2O_2, 150-500µm	CHAR	Sugimura, Sprugel, Brubaker, & Higuera, 2008
Mt. Constitution C38	48.39ºN 122.50º W	685	Western Hemlock	Both	~7200	^{210}Pb polynomial & 4 AMS-^{14}C Li/ regression	Charcoal: 0.25 & 0.5; Pollen: 1 & 2	Following Faegri & Iversen, 1989	10% $(NaPO_3)_6$, 6% H_2O_2, 150-500µm	CHAR	Sugimura, Sprugel, Brubaker, & Higuera, 2008

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Battle Ground Lake	45.08ºN 122.49º W	154	Western Hemlock	Both	14300	18 ^{210}Pb + 7 AMS-^{14}C + 4 tephra 4th order polynomial	Charcoal: 1 & 0.5; Pollen: 5	Faegri & Iversen, 1989	5% $(NaPO_3)_6$ >24 hrs, bleach 1 hr, > 125 µm	Fire Frequency (#/1000 yr)	Walsh, Whitlock, & Bartlein, 2008
Crocker Lake[1]	47.94ºN 122.88º W	60	Western Hemlock	Both	12300	5 14C + 1 AMS-^{14}C date + Mazama	Charcoal: 500 yr; Pollen: 10-30 cm	Faegri & Iversen, 1989 + sieve to 7µm	10% KOH overnight, < 425 µm	CHAR	McLachlan & Brubaker, 1995
Cedar Swamp[1]	47.91ºN 122.87º W	60	Western Hemlock	Both	13200	3 14C + 3 AMS-^{14}C dates + Mazama	Both: 10-30	Faegri & Iversen, 1989 + sieving to 7µm	10% KOH overnight, < 425 µm	CHAR	McLachlan & Brubaker, 1996
Kirk Lake[1]	48.14ºN 121.37º W	190	Western Hemlock	Both	13300	7 ^{14}C + Mazama Li	Both: 5, 10 or 20	Faegri and Iversen, 1975	10% KOH, acetolysis, HF	CHAR	Cwynar, 1987
Boomerang Lake	49.18ºN 124.15º W	373	CWHxm	Charcoal	10270	^{14}C Lia	1		10% $(NaPO_3)_6$, 6% H_2O_2, >150µm	CHAR	Brown & Hebda, 2002a
Bear Cove Bog	50.43ºN 127.272 ºW	30	CWH	Pollen	16500	4 ^{14}C + Mazama Li	--	Faegri & Iversen, 1975			Hebda, 1983

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Pinecrest Lake[1]	49.49ºN 121.43º W	320	CWH, close to IDF transition	Pollen	12900	^{14}C + Mazama Li	10	5% KOH, HF, acetolysis			Mathews, 1975
Somenos Lake	48.48ºN 123.42º W	16	CDFmm	Charcoal	4855	12 ^{210}Pb CRS & 4 AMS-^{14}C & Mazama Bacon	1		5% $(NaPO_3)_6$ & 6% H_2O_2, >125 um	mFRI, Fire Frequency (#/1000 yr)	Murphy et al., 2019
ODP Hole 1034B	48.384ºN 123.30º W	0	CDFmm	Pollen	11450	8 ^{14}C, Lia	25 years	Berglund & Ralska-Jasiewiczowa (1986)			Pellatt, Mathewes & Clague, 2001
Grant's Bog[1]	49.47ºN 125.07º W	80	Coastal Douglas Fir	Pollen	13400	9 AMS-^{14}C + Mazama, Lia	8	10% KOH 8 minutes, < 150 µm, acetolysis			Lacourse et al., 2019
Crater Lake[1]	49.11ºN 120.05º W	2110	ESSFxc and AT transition	Pollen	12000	7 ^{14}C + 2 tephra, Li	1-5	Faegri & Iversen, 1989			Heinrichs et al., 2002
Enos Lake	49.28ºN 124.15º W	47	CDFmm	Charcoal	12840	^{14}C Lia	1		10% $(NaPO_3)_6$, 6% H_2O_2, >150µm	CHAR	Brown & Hebda, 2002a

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Valdes Island On-site Bog (Shingle Point)	49.20ºN 123.38º W	140	CDFmm	Charcoal	6671	2 AMS-^{14}C Lia	1		10% $(NaPO_3)_6$, 5% bleach for 24 hrs, 150-500μm	CHAR	Derr, 2014
Quamichan Lake	48.46ºN 123.40º W	33	CDFmm	Both	250	22 ^{210}Pb CRS	1	Standard (Faegri & Iversen, 1989)	30% KOH for 24 hrs & dilute H_2O_2, >125 μm	mFRI	Pellatt et al., 2015
Florence Lake	48.27ºN 123.30º W	81	CDFmm	Both	564	14 ^{210}Pb CRS & one AMS-^{14}C	2	Standard (Faegri & Iversen, 1989)	30% KOH for 24 hrs & dilute H_2O_2, >125 μm	mFRI	Pellatt et al., 2015
Roe Lake 1	48.46ºN 123.18º W	117	CDFmm	Both	250	18 ^{210}Pb CRS	0.5	Standard (Faegri & Iversen, 1989)	30% KOH for 24 hrs & dilute H_2O_2, >125 μm	mFRI	Lucas & Lacourse, 2013; Pellatt et al., 2015
Wentworth Lake[1]	48.01ºN 124.53º W	37	Sitka Spruce	Both	10600	9 AMS-^{14}C +Mazama, smoothing spline	Pollen: 10 Charcoal: 7-20	Following Faegri & Iversen, 1964	10% KOH overnight, > 150 & > 500 μm	Charcoal concentration	Heusser, 1973; Gavin & Brubaker, 2014
Kalaloch Lake[1]	47.63ºN 124.38º W	35	Sitka Spruce	Pollen	70000	14 ^{14}C Li	10-20	10% KOH, HF, acetolysis			Heusser, 1972

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Davis Lake[1]	46.35ºN 122.15º W	282	Mountain Hemlock	pollen	26000	15 ^{14}C + Linear regression	10	Standard (Faegri & Iversen, 1975) + $Na_4P_2O_7$			Barnosky, 1981
Louise Pond	53.25ºN 131.452º W	650	MHvh	Pollen	12690	5 ^{14}C + 2 AMS-^{14}C	5-10	Standard (Moore et al., 1991)			Pellatt & Mathewes, 1994
SC1 Pond	54.25ºN 131.544º W	550	MH	Pollen	8040	3 ^{14}C Lia	10	Standard (Moore et al., 1991)			Pellatt & Mathewes, 1997
Shangri-La Bog	53.16ºN 132.24º W	595	MH	Pollen	8060	4 ^{14}C Lia	5	Standard (Moore et al., 1991)			Pellatt & Mathewes, 1997
Mt. Barr Cirque Lake	49.16ºN 121.31º W	1376	MHmm	Charcoal	7500	AMS-^{14}C Lia + Tephra	1		10% $(NaPO_3)_6$, for 24 hrs, >125μm	Fire Frequency (#/1000 yr)	Hallett et al., 2003

Site Name	Latitude (ºN) & Longitude (ºW)	Ele-vation (masl)	BC Biogeo-climatic Zone/US General Forest Zone	Paleo-proxy Used	Record Length (yrs)	Age Model	Sample Interval (cm)	Pollen Extraction Method	Charcoal Extraction Method	Primary Measure of Fire Activity	Reference
Frozen Lake	49.36ºN 121.28º W	1180	MHmm	Charcoal	11400	AMS-^{14}C Lia + Tephra	1		10% $(NaPO_3)_6$, for 24 hrs, >125µm	Fire Frequency (#/1000 yr)	Hallett et al., 2003
Walker Lake 1	48.31ºN 124.00º W	950	MHmm	Both	12240	4 ^{14}C Lia + Mazama	5	Standard (Moore et al., 1991)	10% $(NaPO_3)_6$, 6% H_2O_2, >150µm	CHAR	Brown & Hebda, 2003
Porphyry Lake 1	48.54ºN 123.50º W	1100	MHmm	Both	12540	4 ^{14}C Lia	5	Standard (Moore et al., 1991)	10% $(NaPO_3)_6$, 6% H_2O_2, >150µm	CHAR	Brown & Hebda, 2003
Moose Lake	47.53ºN 123.21º W	1508	Sub Alpine Forest	Both	13100	5 ^{14}C & AMS-14C Lia +Mazama	Charcoal: : 15; Pollen: 3-30	Standard (Faegri & Iversen, 1992)	10% KOH overnight, > 150 & > 500 µm	Charcoal concentration	Gavin et al., 2001
Martins Lake	47.42ºN 123.32º W	1415	Sub Alpine Forest	Both	11200	3 ^{14}C & AMS-14C Lia & Mazama	Charcoal: 5; Pollen: 3-30	Standard (Faegri & Iversen, 1992)	10% KOH overnight, > 150 & > 500 µm	Charcoal concentration	Gavin et al., 2001
Yahoo Lake	47.84ºN 124.80º W	717	Pacific Silver Fir	Both	13700	5 AMS-^{14}C cubic spline + Mazama	Charcoal: 1; Pollen: 8-9	Standard (Faegri & Iversen, 2000), 7µm sieve for clay	5% KOH at 40C for 20 min, 150-500µm	Fire Frequency (#/1000 yr)	Gavin et al., 2013

Adapted from Murphy et al. (2019)
[1]Additional site data added for this study. Note that all pollen data was added for this study.
Li = Linear interpolation
Biogeoclimatic zones: Coastal Douglas Fir (CDF), Coastal Western Hemlock (CWH), Mountain Hemlock (MH), Englemann-Spruce-Subalpine-Fir (ESSF), and Alpine Tundra (AT)
Biogeoclimatic subzones: Very Wet Hypermaritime (vh), Very Wet Maritime (vm), Moist Maritime (mm), Dry Maritime (dm), Very Dry Maritime (xm), and Very Dry Cold (xc)

Appendix B.

Wiggle Matching Core Drives D3, D4, and D5

To further investigate my hypothesis of core drives LLC2D4 and LLC2D6 being re-cored material, I used images (Figures B.1, B.2, B.5), pollen percentages (Figure B.3) and charcoal concentrations (Figure B.4) of the drives to wiggle-match data. By visually positioning images of each drive, I determined that the Mazama tephra observed at the top of drive 4 (image C in Figure B.1) most likely aligns with the Mazama tephra observed in drive 3 (image A in Figure B.1). Similarly, the basal clay observed at 19 cm in drive 6 (image D in Figure B.1) is presumed to align with the clay at the end of drive 5 (image B in Figure B.1), suggesting it too is re-cored. This is further confirmed by the inverted date below the basal clay in drive 6.

Pollen percentages of *T. heterophylla, Pinus undiff., P. menziesii* and *Cupressaceae* were found to align nearly identically between drive 4 and drive 5 (Figure B.2). Charcoal concentrations between drives 4 and 5 did not align as clearly, however the approximate locations of peaks are similar (Figure B.3).

Figure B.1. Core drives 3, 4, 5 and 6 aligned at suggested overlap points.
From top: Drive 3, Drive 4, Drive 5, Drive 6.

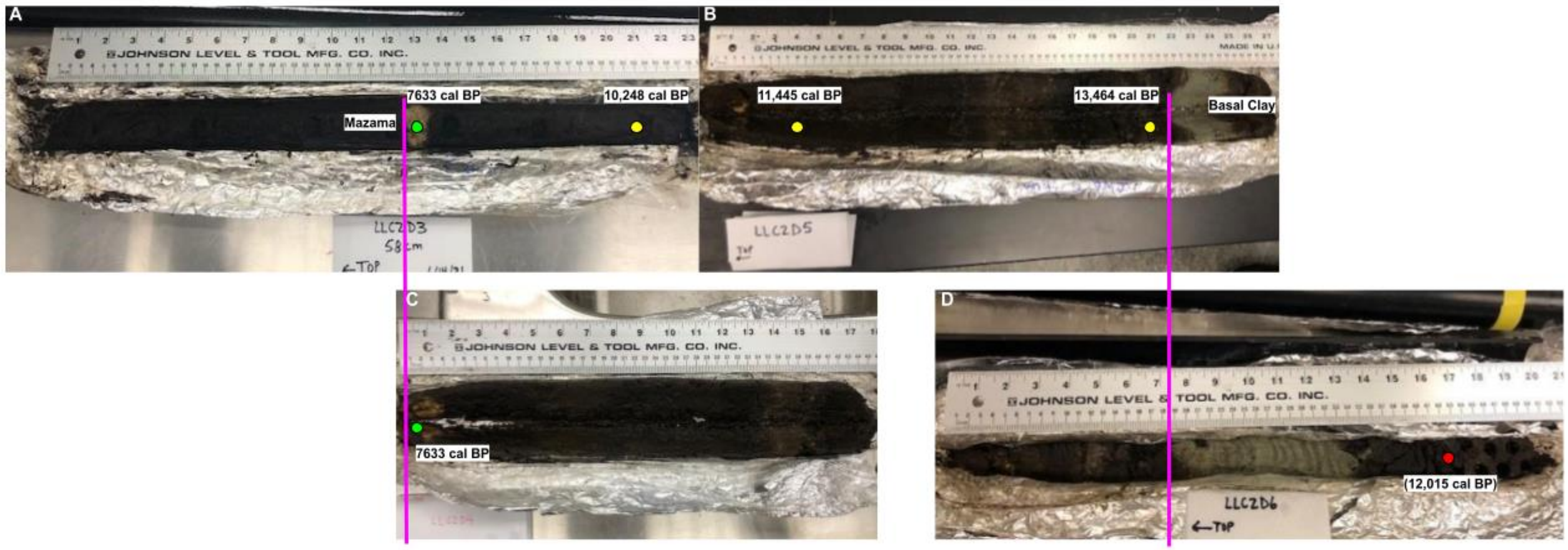

Figure B.2. Core drives 3, 4, 5 and 6 showing hypothesized overlaps.
Green circles represent locations of Mazama tephra. Yellow circles represent AMS-^{14}C-dated points used in final age model. Red circle represents AMS-^{14}C-dated point assumed to be re-cored material and not used in final age model. Pink lines represent hypothesized tie points between core drives.
Image A: Drive LLC2D3; image B: Drive LLC2D4; image C: Drive LLC2D5; image D: Drive LLC2D6.

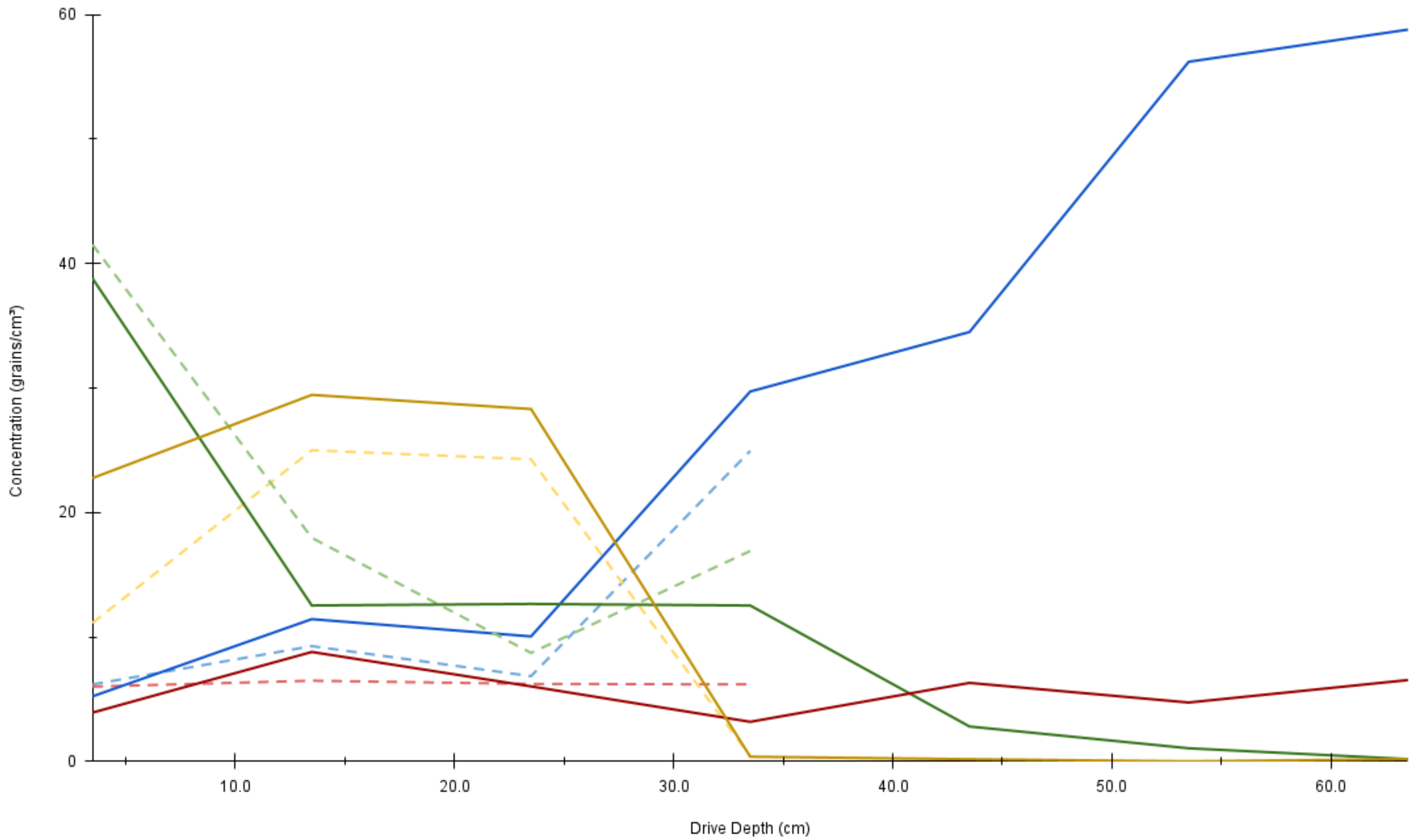

Figure B.3. Select pollen percentages of Lost Lake Core Drive 4 and Drive 5.
Light dotted green represents Drive 4 *T. heterophylla*; light dotted blue represents Drive 4 *Pinus*; light dotted yellow represents Drive 4 *P. menziesii*; light dotted red represents Drive 4 *Cupressaceae*.
Dark solid green represents Drive 5 *T. heterophylla*; dark solid blue represents Drive 5 *Pinus*; dark solid yellow represents Drive 5 *P. menziesii*; dark solid red represents Drive 5 *Cupressaceae*.

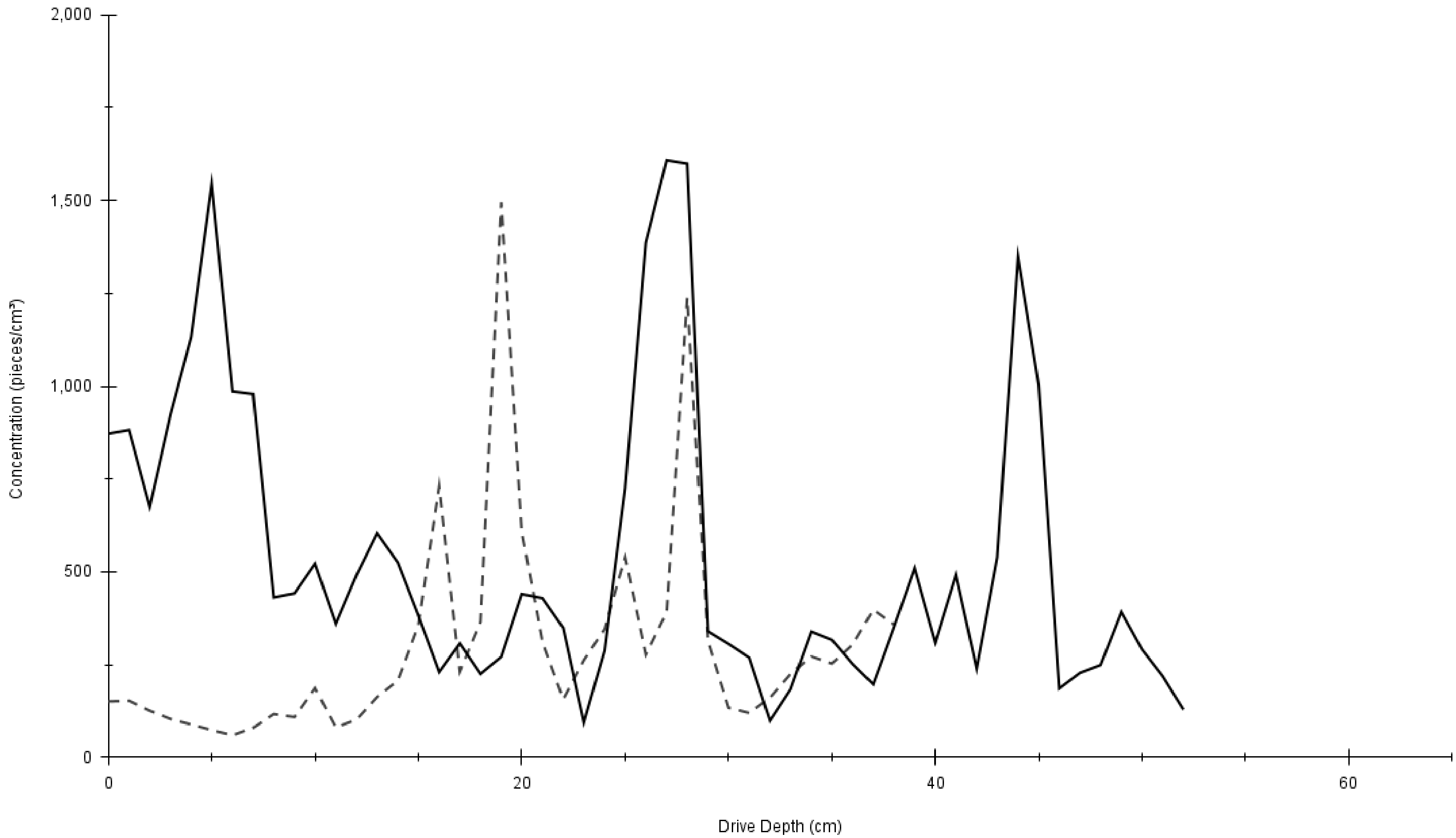

Figure B.4. Charcoal concentrations of Lost Lake core Drive 4 and Drive 5.
Light dotted black represents Drive 4 charcoal concentrations.
Solid black represents Drive 5 charcoal concentrations.
Note top 4cm of Mazama tephra has been removed from Drive 4; bottom 13cm (basal clay and post-clay sediment) has been removed from Drive 5.

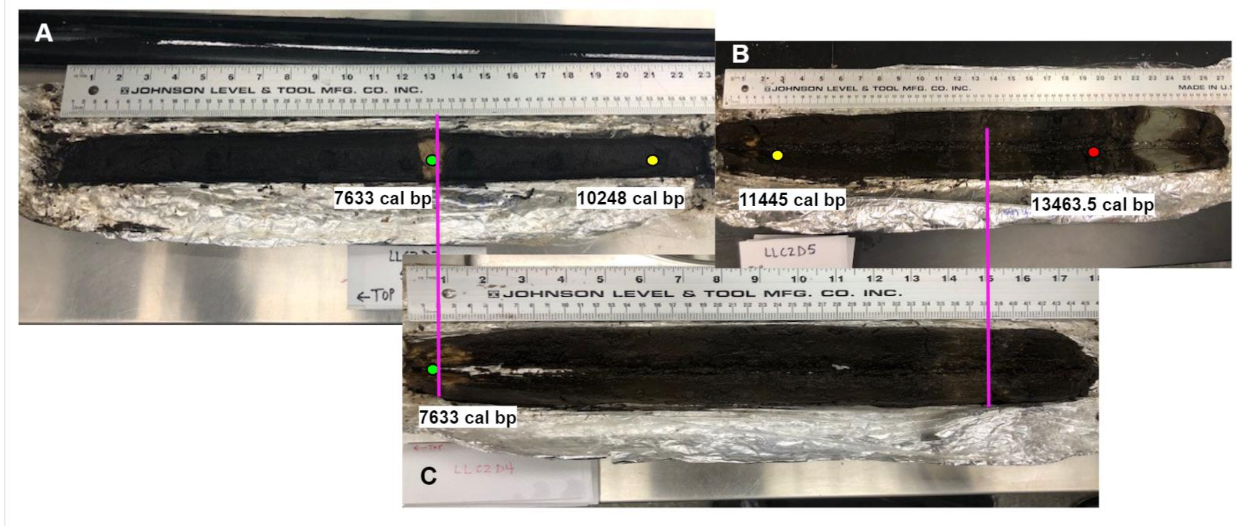

Figure B.5. Cores 3, 4 and 5 showing interpreted re-core of Drive 4.
Green circles represent locations of Mazama tephra. Yellow circles represent AMS-^{14}C-dated points used in final age model. Red circle represents AMS-^{14}C-dated point assumed to be re-cored material and not used in final age model. Pink lines represent hypothesized tie points between core drives.
Image A: Drive LLC2D3; image B: Drive LLC2D5; image C: Drive LLC2D4.

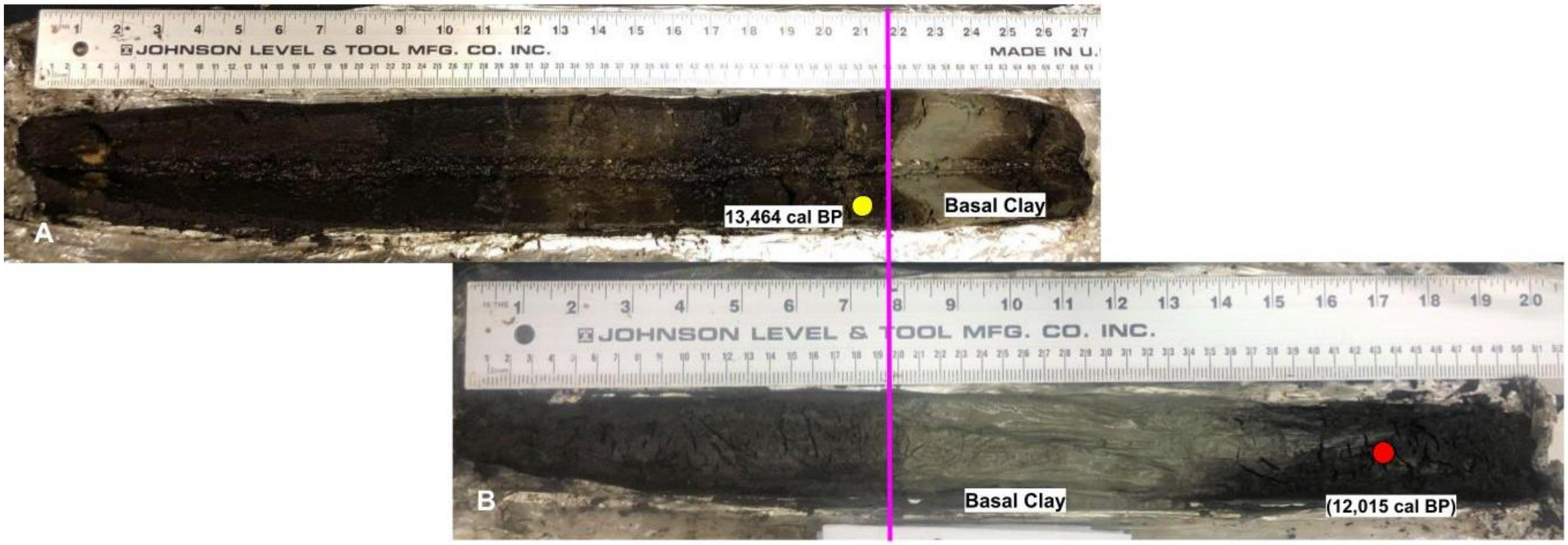

Figure B.6. Cores 5 and 6 showing interpreted re-core of Drive 6.
Yellow circle represents AMS-^{14}C-dated point used in final age model. Red circle represents AMS-^{14}C-dated point assumed to be re-cored material and not used in final age model. Pink line represents hypothesized tie point between core drives.
Image A: Drive LLC2D5; image B: Drive LLC2D6

Appendix C.

Graphs with Unaltered CHAR Data

The large increase in charcoal influx between ca. 61 and -71 cal yr BP (1889 AD – present) affects the y-axis scale of my graphs and decreases overall readability of data trends. To provide a better visual of post-glacial trends, the top 152 years (~41 cm) of CHAR data were removed from graphs in the main document. Graphs with unaltered CHAR (i.e,. including the top 152 years) are shown here.

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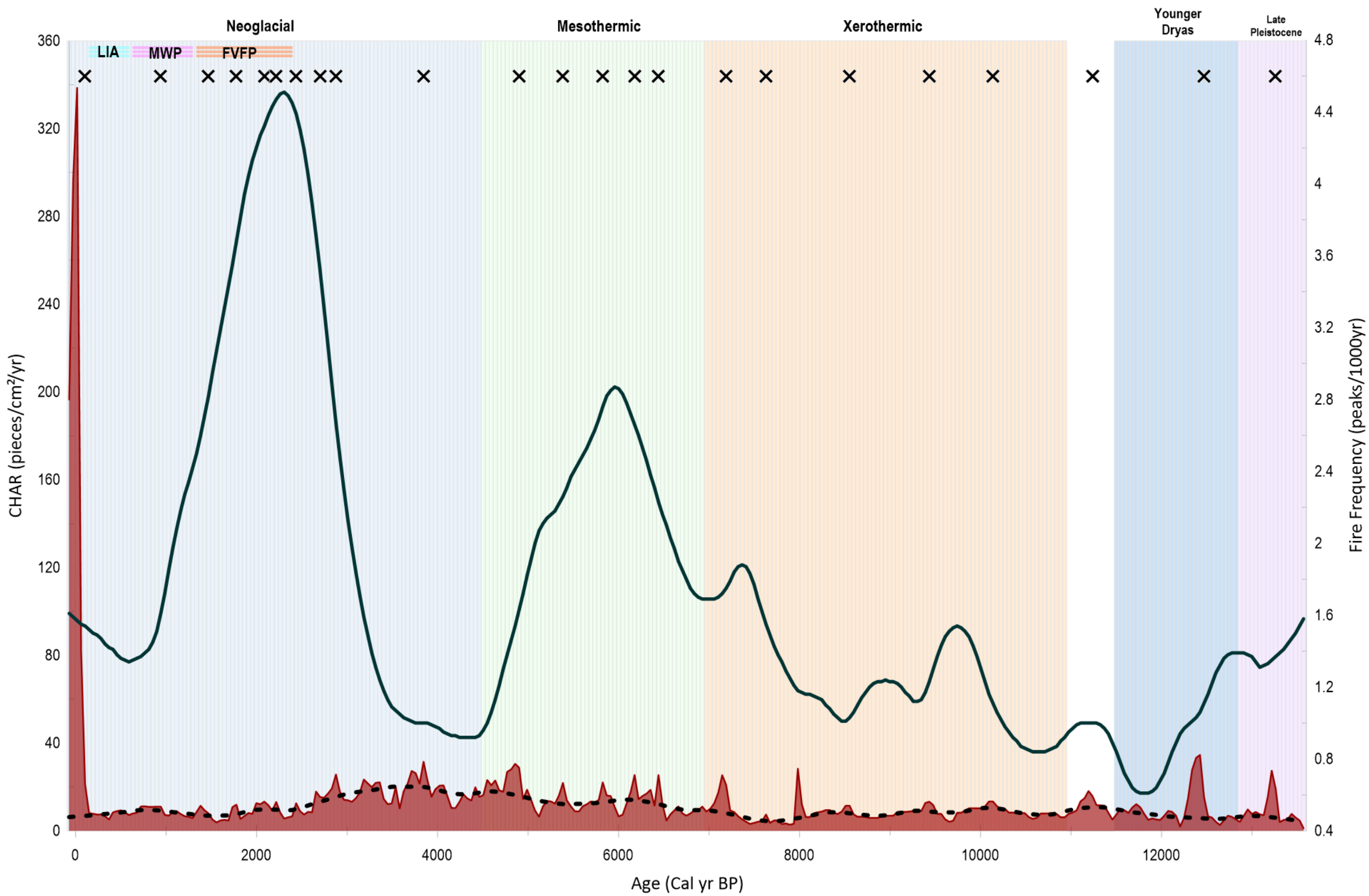

Figure. C.2. Synthesis of Lost Lake charcoal-inferred fire activity and major anthropogenic and climatic changes in southwestern British Columbia since the Late Glacial.

Red fill represents interpolated CHAR; solid black line represents fire frequency; dotted black line represents interpolated background CHAR; X symbols represent fire events.