# Late Pleistocene palaeoenvironments, archaeology, and indicators of a glacial refugium on northern Vancouver Island, Canada

by

Christopher Franklin George Hebda
B.A., University of Victoria, 2014

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We acknowledge with respect the Lekwungen peoples on whose territory the university stands and the Songhees, Esquimalt and W̱SÁNEĆ peoples whose historical relationships with the land continue to this day.

# Supervisory Committee

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## Abstract

Recent research has revealed human settlement on the Pacific coast of Canada extending back nearly 14,000 years, but much of the late Pleistocene record is unknown due to shifting sea levels, poor understanding of Cordilleran ice extent, and limited research on the biota of the coast during this time. This study, undertaken in Quatsino First Nation and ‘Namgis First Nation territories as part of the Northern Vancouver Island Archaeology and Palaeoecology Project, employs modern multi-proxy analysis of lake sediment cores from two sites on northern Vancouver Island to reconstruct palaeoenvironments during and immediately following the Fraser Glaciation in coastal British Columbia. Evidence from radiocarbon samples, pollen, ancient environmental DNA, plant macrofossils, and diatoms indicates that Topknot Lake on the outer coast of Vancouver Island has remained unglaciated through most of the local Last Glacial Maximum since ca. 18,000 cal BP. A non-arboreal herb-shrub tundra assemblage prevailed from ca. 17,500-16,000 cal BP with taxa including willows (*Salix*), grasses, sedges (Cyperaceae), heathers (Ericaceae), and sagewort (*Artemisia*). After ca. 16,000 and into the terminal Pleistocene, Topknot Lake was dominated by pine, alder (*Alnus*), ferns, and aquatic plant species. In the Nimpkish River Valley deep in the Vancouver Island Ranges, Little Woss Lake also demonstrates a record extending to the late Pleistocene (ca. 14,300 cal BP). The environment comprised dry and cool conifer woodland dominated first by fir (*Abies*) until ca. 14,000 cal BP, then by pine, alder, and ferns from ca. 14,000-12,000 cal BP. eDNA evidence from ca. 14,000 cal BP corroborates these plant taxa as well as indicating brown bear and Chinook salmon in and around the basin at that time. A mixed-conifer assemblage consisting of pine, western hemlock, and alder followed from ca. 12,000-11,100 cal BP into the early Holocene. Collectively, these indicators demonstrate an open environment on the outer coast of northern Vancouver Island since ca. 18,000-17,500 cal BP and well-established biotic communities across the region throughout the late Pleistocene. These results inform future archaeological research for early human habitation in coastal British Columbia and provide key evidence to support the viability of the coastal migration route for the first peopling of the Americas.

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# 1. Introduction

## 1.1 Context and Research Questions

For many decades, the academic understanding of the first peopling of the Americas involved hunter-gatherer peoples bearing distinctive fluted Clovis points migrating across Beringia and into North America through an ice-free corridor between the Cordilleran and Laurentide ice sheets in the heart of the continent (e.g. Goebel et al. 2008; Meltzer 2009:4-5). After the global Last Glacial Maximum, these megafaunal hunters moved through the Upper Mackenzie River drainage along the eastern side of the Rocky Mountains before spreading southward to populate the Americas. The oldest Clovis-bearing archaeological sites south of the ice sheets were initially dated to ca. 13,600 calibrated years before present (cal BP), but improved dates on these original sites and new evidence from sites like El Fin del Mundo have constrained the oldest known use of the technology to ca. 13,400 cal BP (Waters and Stafford 2007; Sánchez-Morales 2018; Waters 2019). Due to the widespread occurrence of Clovis points and the alluring idea of early peoples pursuing mastodons and ground sloths across the postglacial plains, the late Pleistocene and early Holocene archaeological record of coastal British Columbia has long been peripheral to the debate surrounding the first peopling of the Americas.

Knut Fladmark (1979), having conducted early period archaeology in both the continental interior and along the coast, was one of the first researchers to present a robust argument for the arrival of early peoples in the Americas via the Pacific coastal route. Attributing the inception of the coastal migration theory to Heusser (1960) with encouragement by Krieger (1961), Fladmark (1979) acknowledged a lack of data across much of the coast and the need for additional archaeological and palaeoenvironmental research. However, he argued that a poorly drained interior migration route with little vegetation and large proglacial lakes was less conducive to human habitation than a long stretch of coastline rich in marine and littoral resources, even with extensive glaciation from the Pacific Coastal Ranges (Fladmark 1979). William Mathews (1979) also considered a coastal route possible, though he characterized the Pacific coast during the Fraser Glaciation as ‘a precipitous ice-front calving bergs directly into the sea,’ noting that a migration of early peoples through the region at this time would be ‘such a formidable undertaking, by virtue of both physical and psychological barriers, as to be a highly unlikely event’ (1979:150,153). Fladmark (1979), Mathews (1979), and others since—most recently Braje et al. (2019a)—all argue that early peoples may have moved up and down the interior ice-free corridor area since the terminal Pleistocene, but that it was not the first route by which humans entered the Americas.

The broad acceptance of the lower dates (ca. 14,500-14,000 cal BP) from the site of Monte Verde in southern Chile is often cited as the turning point in this debate that challenged the supremacy of the Clovis-first model (Dillehay 1997; Meltzer et al. 1997). Many late Pleistocene sites pre-dating Clovis that were previously dismissed as ‘too old’ began to resurface throughout the Americas, and new pre-Clovis sites began to be documented (Adovasio et al. 1990; Gilbert et al. 2008; Kenady et al. 2011; Waters et al. 2011; Dillehay et al. 2012; Dillehay et al. 2015; Halligan et al. 2016; Politis et al. 2016; Dillehay et al. 2017; Williams et al. 2018; Davis et al. 2019; Waters 2019; Williams and Madsen 2019). Increasing skepticism fueled by the slow trickle of credible pre-Clovis sites and DNA analyses of Siberian and American haplogroups (Moreno-Mayar et al. 2018a;2018b; Raff 2019)—punctuated by events such as the acceptance of Monte Verde’s early dates—has now turned the old theory on its head. Especially over the past 20 years, many researchers have disputed the ability for the ice-free corridor to be able to support such early migrations, and a renewed interest in the idea of early migrations along the northwest Pacific coast has emerged (Heintzman et al. 2016; Pedersen et al. 2016; Froese et al. 2019).

Unfortunately, many of the earliest sites that could attest to these migrations along the western edge of the Cordilleran ice are ephemeral, obscured by the movement of glaciers and the rising and falling of regional relative sea levels over thousands of years (Braje et al. 2019a; 2019b). In order to find evidence of sites shrouded by deep time, we must understand the geological and ecological conditions that both constrained and aided early peoples. To this end, over the past two decades researchers have begun to address the coastal corridor hypothesis with new insights drawing from archaeological, geological, and palaeoecological studies (Josenhans et al. 1997; Fedje et al. 2004; Ramsey et al. 2004; Fedje and Mathewes 2005; McLaren 2008; Fedje et al. 2011; Mackie et al. 2011; McLaren et al. 2014; Carlson and Baichtal 2015; Gauvreau and McLaren 2017; Fedje et al. 2018; Mackie et al. 2018; McLaren et al. 2018; Braje et al. 2019a). Due to the complex interplay between glacial expansion and retreat, relative sea level change, and the development of the thick coastal temperate rainforest, palaeoenvironmental information is paramount to effective archaeological investigations of the late Pleistocene in northwestern North America.

Some regions, like Haida Gwaii, have seen multidisciplinary research focusing on palaeoenvironments and early archaeology (Warner et al. 1982; Warner 1984; Clague 1983; Clague 1989a; Heusser 1989; Josenhans et al. 1997; Barrie and Conway 1999; Fedje and Mathewes 2005; Lacourse et al. 2005; Fedje et al. 2011; Cohen 2014; Mathewes and Clague 2017; Shaw et al. 2019). Other regions, like northern Vancouver Island, are potentially important portions of the coastal corridor, but have not been investigated as thoroughly. I seek to address some of these late Pleistocene palaeoenvironmental data gaps in this thesis by analyzing multi-proxy palaeoecological data from two lakes located on northern Vancouver Island: Little Woss Lake and Topknot Lake. In doing so I address two major questions, each with supporting lines of enquiry. The first major question is theoretical, the second methodological:

1. At what time during the late Pleistocene did northwestern and north-central Vancouver Island become viable environments for human habitation?
    a) What are the long-term trends in the makeup of biotic communities through the late Pleistocene and early Holocene on northwestern and north-central Vancouver Island, as recorded in the sediments of Topknot Lake and Little Woss Lake?
    b) Using the presence or absence of plant and animal species at these sites as proxies for glacial proximity, what can these records tell us about late glacial Cordilleran ice advance and retreat in these regions?
    c) Can palaeoecological records like those recovered from the two study sites reveal the presence of refugia in these regions during the LGM along the coast of British Columbia?

2. Is the use of ancient environmental DNA (eDNA) analysis valuable for reconstructing the palaeoenvironments of coastal British Columbia?
    a) Can fragmentary ancient eDNA be successfully extracted and analyzed from samples collected in coastal BC?
    b) Is eDNA analysis complementary to established approaches (including pollen and macrofossils) when analysing the presence or absence of species on the coast?

I do not expect to resolve the question of the peopling of the Americas in the following pages. By necessity, the results of this study speak to the regional and site-specific conditions of northern Vancouver Island. However, considering these records in the broader context of Cordilleran glaciation and coastal archaeology in northwestern North America, I believe this thesis contributes meaningful data to the discussion of the theorized late Pleistocene migration route into the Americas along the Pacific coast.

## 1.2 Chapter Organization

The chapters of this thesis are organized as follows:

Chapter 2, the Background, outlines the physiographic and biotic setting of the study sites, Little Woss Lake and Topknot Lake, on northern Vancouver Island. It also outlines the recent histories of glaciation, relative sea level change, archaeology, and reconstructions of floral and faunal habitats on the Pacific northwest coast of North America since the late Pleistocene.

Chapter 3, the Materials and Methods, outlines the Program of Research as well as the theoretical methods and procedures carried out in the field and in the lab during this project.

Chapter 4, the Results and Discussion, also known as the ‘Article’ chapter, was prepared as a draft for a standalone manuscript intended for publication. Because of this, it is somewhat repetitive of the earlier chapters and contains summaries of the Introduction, Background, and Materials and Methods. This is followed by the Results for each study site and then a Discussion of the evidence and its implications. Finally, this chapter contains a section of concluding thoughts, though these are also expanded upon in Chapter 5. I am the primary driver of the work in this thesis and the work that will be subsequently published. However, many others have contributed in significant ways to this study, and during the preparation of this material for publication several of them will be included as co-authors. I have outlined the contributions of the various co-authors at the beginning of Chapter 4 along with a disclaimer similar to this one.

Chapter 5, the Conclusions and Future Directions, considers the evidence and the contents of the discussion and relates it to the initial research questions I have laid out in Chapter 1. I also indicate what I believe are the important next steps for palaeoenvironmental research in coastal British Columbia and how we might use this knowledge to better understand the first peopling of the Americas.

As a general note, the first time a species, genus, or family of organisms is mentioned in this thesis, it will always be accompanied by both a scientific name and a common name. Thereafter, the classification will generally be referred to by its common name. The exception to this is in the Results section of Chapter 4, where I use scientific names preferentially over common names in keeping with the conventions of other palaeoecological studies. All scientific and common names are consistent with those documented in e-Flora BC for plants (Klinkenberg 2019a) and in e-Fauna BC for animals (Klinkenberg 2019b). Tables depicting all the scientific and common names referred to in this document can be found in Appendix A.

# 2. Background

## 2.1 Environmental Setting

### 2.1.1 Physiographic Regions of Vancouver Island

What is today the province of British Columbia in western Canada is home to an impressively diverse array of cultures, geology, and biology, stretching from the stormy shores of the Pacific Ocean to the mighty peaks of the Rocky Mountains. Largely blanketed with coniferous and boreal forest, the province is a mosaic of mountain ranges, fjords, plateaus, deserts, alpine tundra, and rocky coastlines (Holland 1976; Mathews 1986).

Located off the southwestern edge of BC and 450 km long, Vancouver Island demonstrates highly varied physiography (see Figure 1). The Vancouver Island Ranges—the spine of mountains that run along much of the island's length—are the result of hundreds of millions of years of tectonic activity that formed the Insular Belt, the westernmost of the five belts that constitute the Canadian Cordillera (Holland 1976). These high mountains create numerous ecosystems and microclimates along the island's flanks. Along Vancouver Island's western margin, the West Vancouver Island Fjordland carves deep into the island's interior as the mountains descend toward the Estevan Lowlands and the rugged Pacific Ocean (Yorath 2005). The exceptions to this rise and fall in elevation are the lowlands that span the northern tip of Vancouver Island: Suquash Basin and Nahwitti Plateau (Holland 1976). On the northeastern side of the island, Suquash Basin consists of rolling terrain and rivers that snake their way toward Queen Charlotte Strait and across to the Coast Mountains beyond. Nahwitti Plateau, covering the remainder of the northern and western sides of the island's tip, is characterized by broad valleys and low, rounded hills below 600 m in elevation between the northern side of Quatsino Sound and Cape Scott (Holland 1976). Nestled along the eastern side of the island, the Nanaimo Lowlands rise from the Salish Sea before transitioning to the hilly regions of the Victoria

Highlands, Nanaimo Lakes Highlands, and the Quinsam Plateau, with the Vancouver Island Ranges beyond (Yorath 2005).

***Figure 1.*** *Map depicting the physiographic regions of Vancouver Island. Boundaries and names have been derived from Yorath (2005). Map data: Natural Earth, NOAA, Viewfinder DEM, GeoBC.*

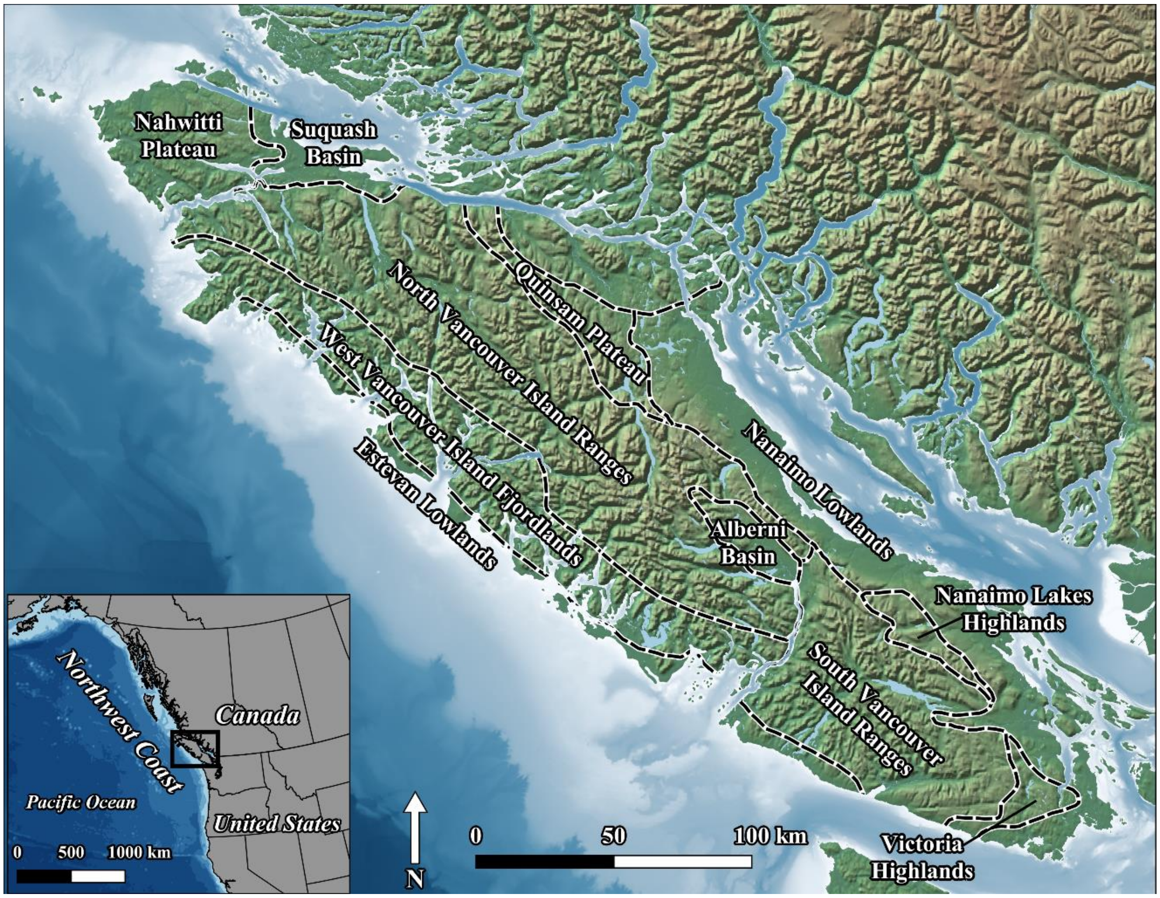
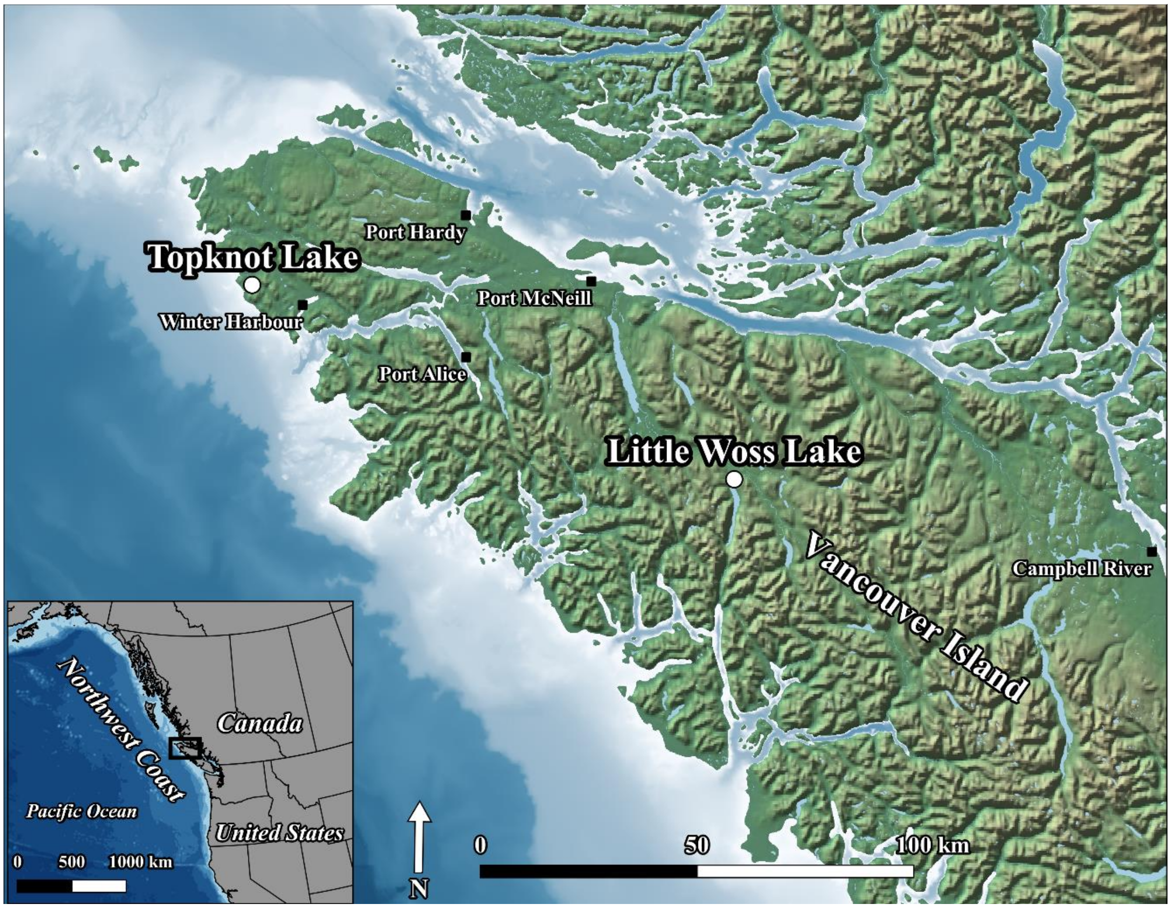


***Figure 2.*** *Study sites at Little Woss Lake and Topknot Lake on northern Vancouver Island. Nearby settlements are labeled with black squares. Map data: Natural Earth, NOAA, Viewfinder DEM, GeoBC.*



<u>2.1.2 Study Sites</u>

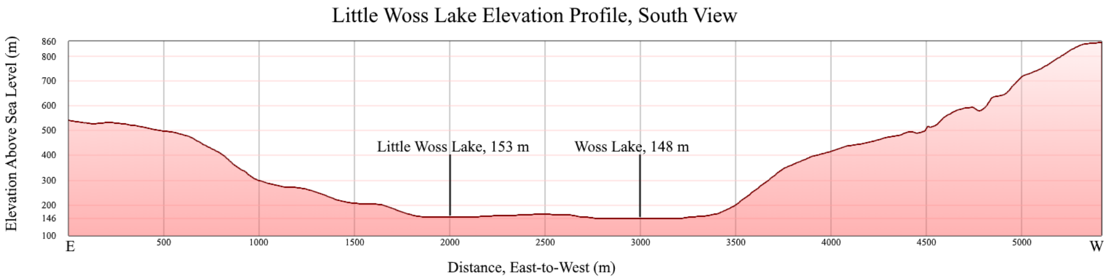
This study focuses on two sites: Little Woss Lake within ‘Namgis First Nation territory on north-central Vancouver Island and Topknot Lake within Quatsino First Nation territory on the outer west coast of the island’s northern tip (see Figure 2). Little Woss Lake is located at the northern end of Woss Lake, approximately 3 km south-southwest of the community of Woss and 26 km southeast of the southern end of Nimpkish Lake. Topknot Lake is located approximately 12 km west-northwest of the community of Winter Harbour at the mouth of Quatsino Sound, 3 km southeast of Raft Cove.

*Little Woss Lake*

Little Woss Lake (50.180729°, -126.611854°; Figure 3) is a small, shallow lake located approximately 550 m to the northeast of the northern end of Woss Lake, deep within the Northern Vancouver Island Ranges within ‘Namgis First Nation territory. The lake receives sediment runoff from nearby slopes to the east (see Figure 4) and it drains southward into the main body of Woss Lake. The lake measures approximately 700 m north-south and 250 m east-west, with an area of 10 hectares (GeoBC 2019). The underlying bedrock geology around Little Woss Lake is a mosaic of Triassic and Jurassic igneous and sedimentary formations (BC Geological Survey 2019). Little Woss Lake, the northern end of Woss Lake, and parts of the Nimpkish River Valley rest upon intrusive granodioritic rocks of the Island Plutonic Suite. However, much of the southern two-thirds of Woss Lake and the southern and western sides of Nimpkish Lake are underlain by dark grey-green Karmutsen Formation basalt pillow lava flows. Small lenses of limestone formations from the Sicker Formation-Buttle Lake Group and the Quatsino Formation are also located nearby in pockets to the east and the west (Nixon et al. 2011b).

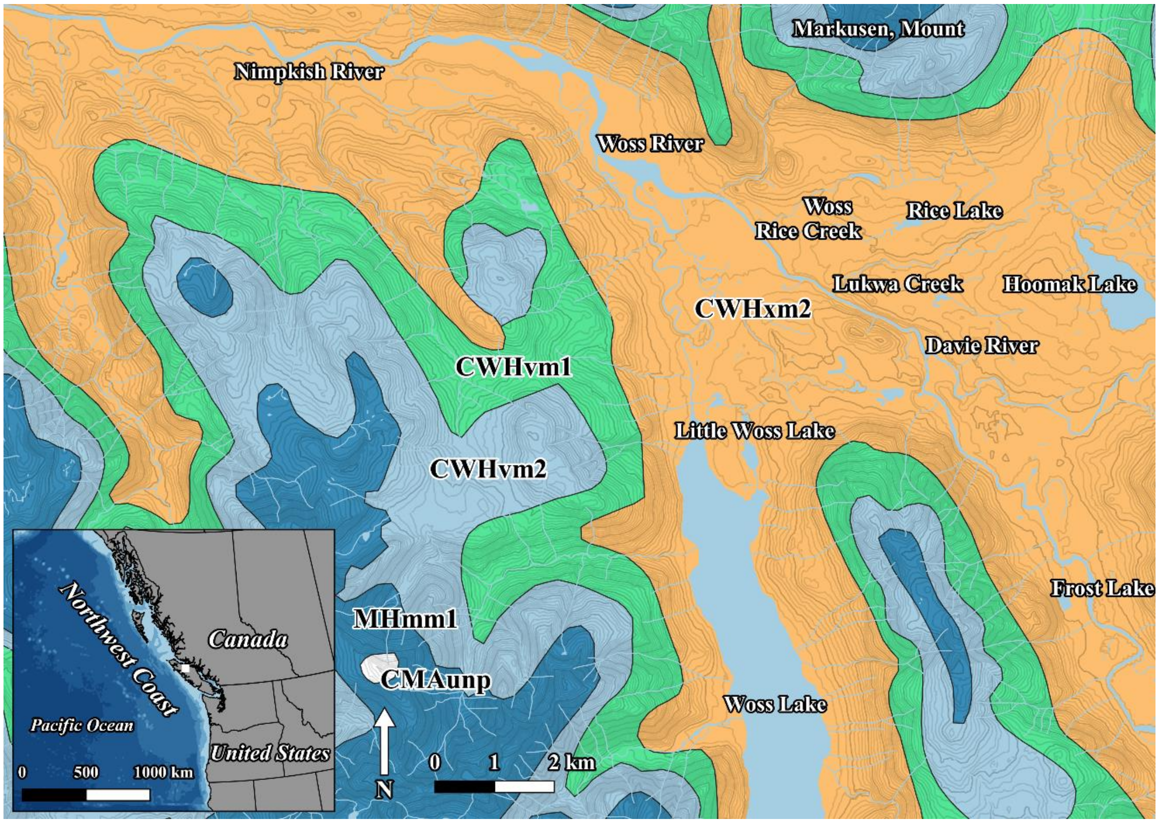
***Figure 3.*** *Satellite imagery of Little Woss Lake. Map data: Google, DigitalGlobe.*

***Figure 4.*** *Elevation profile and topographic setting of Little Woss Lake and the northern end of Woss Lake, facing south. Data: Google, DigitalGlobe.*



Today, at 153 m above sea level, Little Woss Lake is within the Western Very Dry Maritime subzone of the Coastal Western Hemlock biogeoclimatic ecosystem classification (CWHxm2; BC Forest Analysis and Inventory Branch 2018) (see Figure 5). This subzone extends through the region on a northwest-southeast axis along the valley bottoms around Nimpkish Lake, the Nimpkish River, Woss Lake, and Vernon Lake. The vegetation regime of the CWHxm2 subzone is characterized by Douglas-fir (*Pseudotsuga menziesii*), western hemlock (*Tsuga heterophylla*), and western redcedar (*Thuja plicata*) with an understory of salal (*Gaultheria shallon*), dull Oregon-grape (*Mahonia nervosa*), red huckleberry (*Vaccinium parviflorum*), vanilla-leaf (*Achlys triphylla*), sword fern (*Polystichum munitum*), and twinflower (*Linnaea borealis*) (Green and Klinka 1994; see Appendix A of this thesis for a list of scientific and common names discussed in the text). In wetland and bog/fen environments such as that

***Figure 5.*** *Context of Little Woss Lake, including biogeoclimatic ecosystem classification of nearby areas.* ***CWHxm2****-Coastal Western Hemlock Western Very Dry Maritime;* ***CWHvm1****-Coastal Western Hemlock Submontane Very Wet Maritime;* ***CWHvm2****-Coastal Western Hemlock Montane Very Wet Maritime.* ***MHmm1****-Mountain Hemlock Windward Moist Maritime;* ***CMAunp****-Coastal Mountain-heather Alpine Undifferentiated and Parkland. Map data: Natural Earth, GeoBC.*



surrounding Little Woss Lake, the subzone is characterized by Labrador tea (*Rhododendron groenlandicum*), sweet gale (*Myrica gale*), Sitka sedge (*Carex sitchensis*), bog cranberry (*Vaccinium oxycoccos*), western bog-laurel (*Kalmia microphylla*), and peat-moss (*Sphagnum*) among other species (MacKenzie and Moran 2004). The dry valley bottoms transition to somewhat wetter areas upslope, first into the Submontane and Montane Very Wet Maritime subzones of the Coastal Western Hemlock zone (CWHvm1/2). These subzones are characterized by western hemlock, amabilis fir (*Abies amabilis*), and western redcedar, as well as mountain hemlock (*Tsuga mertensiana*) and yellow-cedar (*Xanthocyparis nootkatensis*) at greater elevation. The highest elevations of these deeply incised valleys are classified under the Windward Moist Maritime subzone of the Mountain Hemlock zone (MHmm1), characterized by many of the same tree species (mountain hemlock, amabilis fir, yellow-cedar, and western hemlock) but lacking in lower-elevation species like western redcedar and shore pine (*Pinus contorta*) (Green and Klinka 1994).

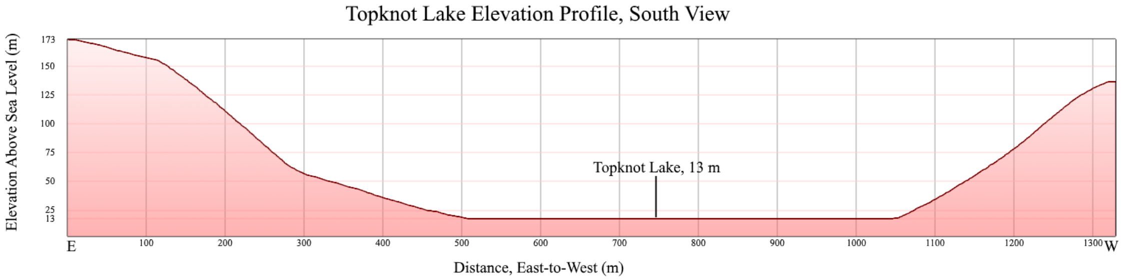
Species observed around the margins of Little Woss Lake in May 2018 include western hemlock, western redcedar, sweet gale, black twinberry (*Lonicera involucrata*), and sedge (Cyperaceae).

### *Topknot Lake*

Topknot Lake (50.563752°, -128.189976°; Figure 6) is a mid-sized lake located approximately 2 km inland from the outer west coast of Vancouver Island on Nahwitti Plateau within Quatsino First Nation territory. The drainage area of the lake is relatively small, including runoff from low hillsides to the east and west (see Figure 7) and a low boggy wetland at its southern end. Topknot Lake's outflow emerges at its north end from which a stream extends for approximately 1.8 km, joining the Macjack River before draining into the eastern side of Raft Cove and the Pacific Ocean. The lake measures approximately 1.1 km north-south and 600 m east-west, with an area of 38.5 hectares (GeoBC 2019). The underlying bedrock geology around Topknot Lake consists largely of Bonanza Group volcanic rocks running north-south along the outer west coast between Cape Scott and the mouth of Quatsino Sound, mostly dark grey-green basaltic to andesitic flows. Other areas farther inland include some Queen Charlotte Group sandstone and conglomerate formations, and the eastern side of northern Vancouver Island consists largely of dark grey-green Karmutsen Formation basalt flows (Nixon et al. 2011a).

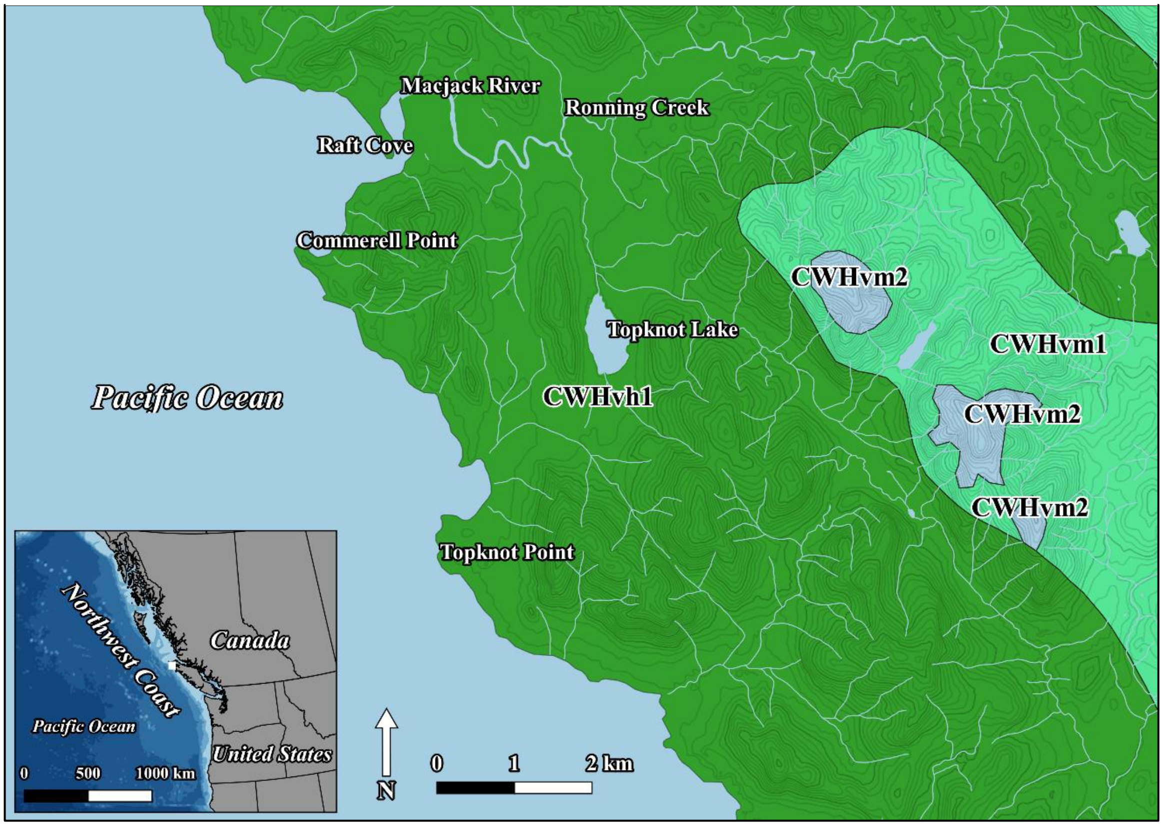
***Figure 6.*** *Satellite imagery of Topknot Lake. Map data: Google, DigitalGlobe.*

***Figure 7.*** *Elevation profile and topographic setting of Topknot Lake Valley and nearby slopes, facing south. Data: Google, DigitalGlobe.*



Topknot Lake is located at 13 m above modern sea level and falls within the Southern Very Wet Hypermaritime subzone of the Coastal Western Hemlock biogeoclimatic ecosystem classification (CWHvh1; BC Forest Analysis and Inventory Branch 2018) (see Figure 8). This subzone covers most of Nahwitti Plateau, wrapping around the northern tip of Vancouver Island from Winter Harbour to Cape Scott, around Cape Sutil and past Shushartie nearly to Hardy Bay.

***Figure 8.*** *Context of Topknot Lake, including biogeoclimatic ecosystem classification of nearby areas.* ***CWHvh1****-Coastal Western Hemlock Southern Very Wet Hypermaritime;* ***CWHvm1****-Coastal Western Hemlock Submontane Very Wet Maritime;* ***CWHvm2****-Coastal Western Hemlock Montane Very Wet Maritime. Map data: Natural Earth, GeoBC.*



The forest vegetation of the CWHvh1 subzone is characterized by western hemlock, western redcedar, and amabilis fir, with lesser amounts of yellow-cedar and mountain hemlock. The understory is composed of diverse shrubs and herbs including salal, deer fern (*Blechnum spicant*), red huckleberry, Alaskan blueberry (*Vaccinium alaskaense*), false azalea (*Menziesia ferruginea*), bunchberry (*Cornus canadensis*), and twinflower, among other species (Green and Klinka 1994). Due to the low relief of much of Nahwitti Plateau, the climatic regime around Topknot Lake changes little with vertical topography. The nearby interior portions of Nahwitti Plateau and those around the shores of Holberg Inlet are classified as Submontane and Montane Very Wet Maritime subzones of the Coastal Western Hemlock zone (CWHvm1/2), and so are characterized by slightly different vegetation assemblages. These areas have similar tree species including western hemlock, amabilis fir, and western redcedar, with infrequent mountain hemlock and yellow-cedar except at higher elevation (Green and Klinka 1994). These subzones are also distinguished from the CWHvh1 by having less salal and deer fern than the hypermaritime coastal assemblages. Extensive bog forest covers the lowlands and some of the upland areas throughout the Nahwitti Plateau (Green and Klinka 1994; BC Forest Analysis and Inventory Branch 2018).

Species observed around the margins of Topknot Lake in May 2018 include western redcedar, Sitka spruce (*Picea sitchensis*), mountain hemlock, red alder (*Alnus rubra*), salmonberry (*Rubus spectabilis*), Labrador tea, sedges, false azalea, salal, and Nootka rose (*Rosa nutkana*). The occurrence of some of these species, especially red alder, is the result of forest succession following recent extensive logging activity around the lake.

## 2.2 Glacial History and Relative Sea Level Change on the Northwest Coast

### 2.2.1 Late Pleistocene Glaciation on the Northwest Coast

The late Pleistocene of northwestern North America can be divided into two major components: the Olympia Nonglacial Interval (ca. 57,000 to 30,000 cal BP) and the Fraser Glaciation (ca. 30,000 to 14,800 cal BP) (Armstrong et al. 1965; Armstrong and Clague 1977; Clague 1981; Ryder and Clague 1989; Hebda et al. 2016). The Olympia Nonglacial Interval, also referred to as the Olympia Interglaciation (Armstrong et al. 1965) or recently as the Olympia Interstade (Hebda et al. 2016), represents a period of relative climatic stability with temperatures comparable to or slightly cooler than the present and limited glacial coverage of the region (Clague and MacDonald 1989). The Olympia Nonglacial Interval is generally correlated with the globally recognized marine isotope stage 3 (MIS 3) (Hebda et al. 2016). The Fraser Glaciation represents a cold period characterized by the advance and coalescence of mountain glaciers into an ice mass known as the Cordilleran ice sheet (CIS) covering much of northwestern North America including British Columbia, southern Yukon Territory, southern Alaska, and parts of the northwestern contiguous United States (Clague 1989b; Clague and James 2002). The Fraser Glaciation in the Pacific Northwest can be correlated broadly to the maximum of the Late Wisconsin glaciation of the Laurentide ice sheet in northern North America and will be referred to by preference over its eastern counterpart. The Fraser Glaciation also generally correlates with the globally recognized marine isotope stage 2 (MIS 2) (Hebda et al. 2016; Mathewes and Clague 2017).

The Olympia Nonglacial Interval consists of several sub-components depending on the region being discussed. For example, in the Georgia Depression (generally including the then-exposed lowlands of what is now the Salish Sea), these may include (1) glaciomarine, glaciolacustrine, or glaciofluvial deposits that represent deglacial sedimentation following the previous glacial period (the Dashwood and Semiahmoo Drifts); overlain by (2) organic silts, sands, and gravel of fluvial, estuarine, and marine origin (the Cowichan Head Formation); followed by (3) a thick upper unit of sand and silt (the Quadra Sand) (Armstrong et al. 1965; Armstrong and Clague 1977; Clague 1981; Ryder and Clague 1989; Miskelly 2012). The distribution of Olympia Nonglacial Interval deposits across the northwest coast of North America varies based on the geomorphic character of each region, but can be characterized as nonglacial sediments accumulating in valleys, lowlands, and off the continental shelf under climatic conditions similar to or somewhat cooler than the present (Clague 1981; Clague and MacDonald 1989; Hebda et al. 2016). Though not every region demonstrates the presence of deposits equivalent to the Quadra Sand, these sediments generally represent the outwash of approaching glaciers during the end of the Olympia Nonglacial Interval and the beginning of the Fraser Glaciation and were eventually overridden by ice (Ryder and Clague 1989).

The Fraser Glaciation is split into several sub-components depending on the region being discussed (Armstrong et al. 1965; Ryder and Clague 1989). All of these are characterized by bedded glacial sediments in periods variously referred to as stades, interstades, or nonglacial intervals based on the oscillation of glacial growth and decay and the influence of regional geomorphic and climatic processes (Hicock et al. 1982; Ryder and Clague 1989; Lian et al. 2001; Hebda et al. 2016). For example, evidence of the Fraser Glaciation (either physical deposits or depositional periods) is known as the Vashon Drift and Sumas Drift in southwestern British Columbia (Armstrong 1981), the Evans Creek Stade, Vashon Stade, Everson Interstade, and Sumas Stade in Puget Lowland (Armstrong et al. 1965), the Coquitlam Stade, Port Moody Interstade, Vashon Stade, and Sumas Stade in Fraser Lowland (Hicock and Lian 1995; Lian et al. 2001), the Gold River Drift on north-central Vancouver Island (Howes 1981a), and the Port McNeill Drift on northern Vancouver Island (Howes 1983). These differing names reflect varied glacial histories and the impacts of many geomorphic and climatic processes on the coast as well as the ideas and work of numerous researchers across many decades and institutions. Collectively, these sediments represent the growth, maximum, and retreat of massive glaciation events across northwestern North America. Globally, the coldest climatic period during the Last Glacial Maximum lasted from approximately 26,500 cal BP to 19,000 cal BP (Clark et al. 2009), but the height of glacial cover was variable across the coast from Alaska to Puget Sound, ranging anywhere from nearly 27,000 cal BP to nearly 17,000 cal BP (see Figure 9). As detailed below, the diachronous nature of these glacial processes demonstrates that local and regional influences have sometimes been equally as important as large-scale continental and global climatic changes in the distribution of life, including humans, across the continent during the last ice age. For the remainder of this thesis, I will distinguish the global Last Glacial Maximum as the GLGM, and regional/local glacial maxima as LLGM.

### *Recent Glacial History by Region*

Studies and maps that depict the most extensive advance of the Cordilleran ice during the Fraser Glaciation often show a dome of ice extending out over coastal BC and Alaska and across Vancouver Island to the edge of the continental shelf (for example Clague 1983, Figure 1; Fulton 1991, Figure 5; Clague and James 2002, Figures 2 and 4; Fulton et al. 2004, Figure 3; Dyke 2004; and others). However, recent studies demonstrate that the extent of the CIS and timing of deglaciation along the north Pacific margin of North America during and after the GLGM was highly regionally specific (see Figure 10) (for example Al-Suwaidi et al. 2006; Carrara et al. 2007; Seguinot et al. 2016; Briner et al. 2017; Darvill et al. 2018; Lesnek et al. 2018; Shaw et al. 2019). During the coldest climatic conditions on the coast, the CIS may have extended to the continental shelf margin in some places with localized smaller mountain glaciers contributing to a patchwork of on- and off-shore ice at the edge of the sheet (Barrie and Conway 1999; Lesnek et al. 2018). However, some outer coastal glacio-distal environments as well as plains exposed by isostatic rebound and crustal forebulge following deglaciation (see Clague 1983) may have served as refugia for plant and animal species during and immediately following the LGM (Hebda 1985; Kondzela et al. 1994; Hansen and Engstrom 1996; Heaton et al. 1996; Soltis et al. 1997; Barrie and Conway 1999; Brown and Hebda 2002; Conroy and Cook 2000; Fleming and Cook 2002; Lacourse et al. 2003; Gapare and Aitken 2005; Reimchen and Byun 2005; Carrara et al. 2007; Godbout et al. 2008; Shafer et al. 2010; Mathewes and Clague 2017; and others). The combination of these isostatic and eustatic effects led to many coastal areas experiencing significantly lower relative sea levels during maximum ice loading, including the Alexander Archipelago of southeastern Alaska (Carrara et al. 2007; Carlson and Baichtal 2015; Lesnek et al. 2018), Haida Gwaii (Warner et al. 1982; Fedje et al. 2005c; Shugar et al. 2014; Mathewes and Clague 2017), and Goose Bank and Cook Bank off the northern end of Vancouver Island (Luternauer et al. 1989b; Lacourse et al. 2003; Hetherington et al. 2004; Shugar et al. 2014; Shaw et al. 2019).

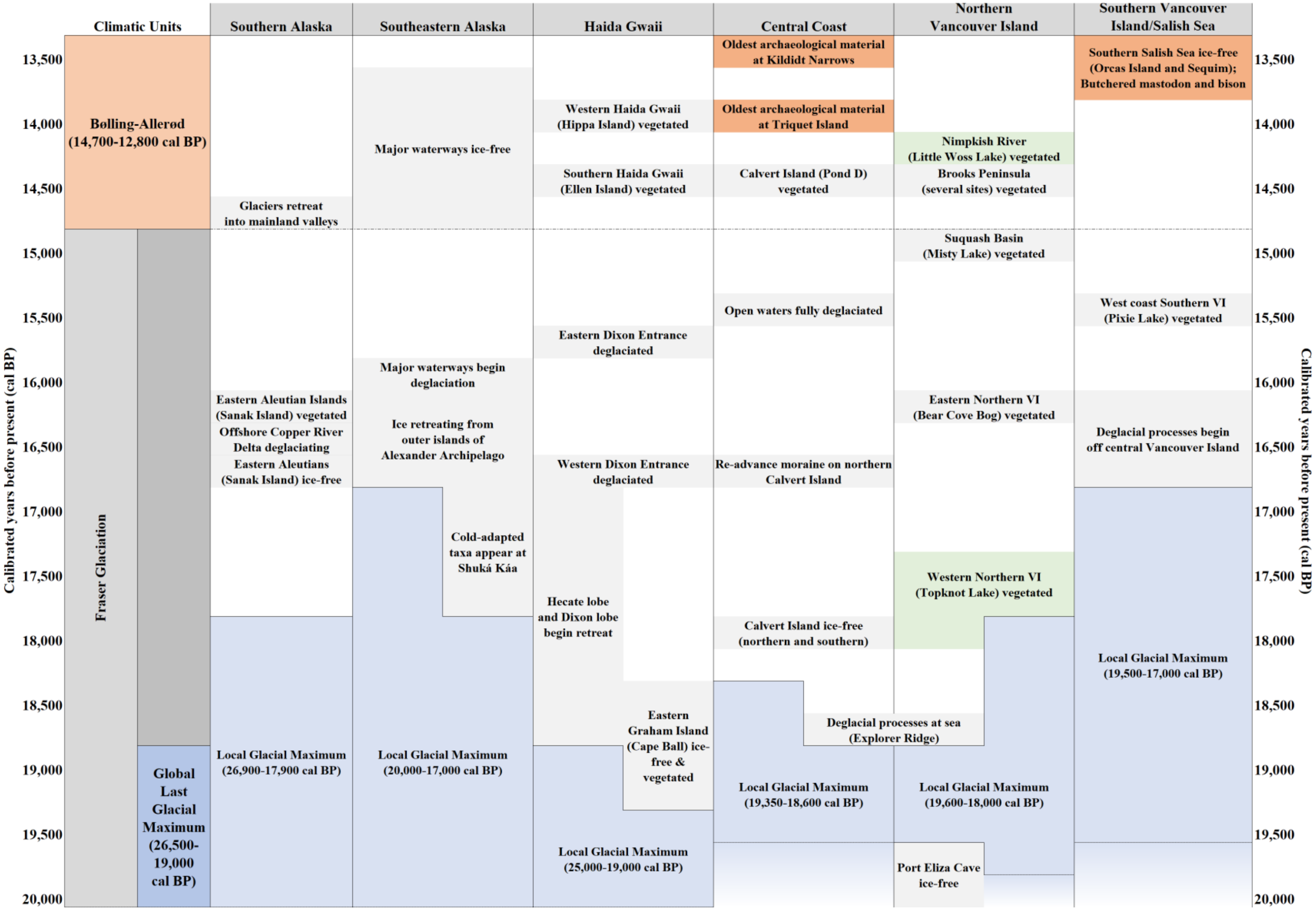
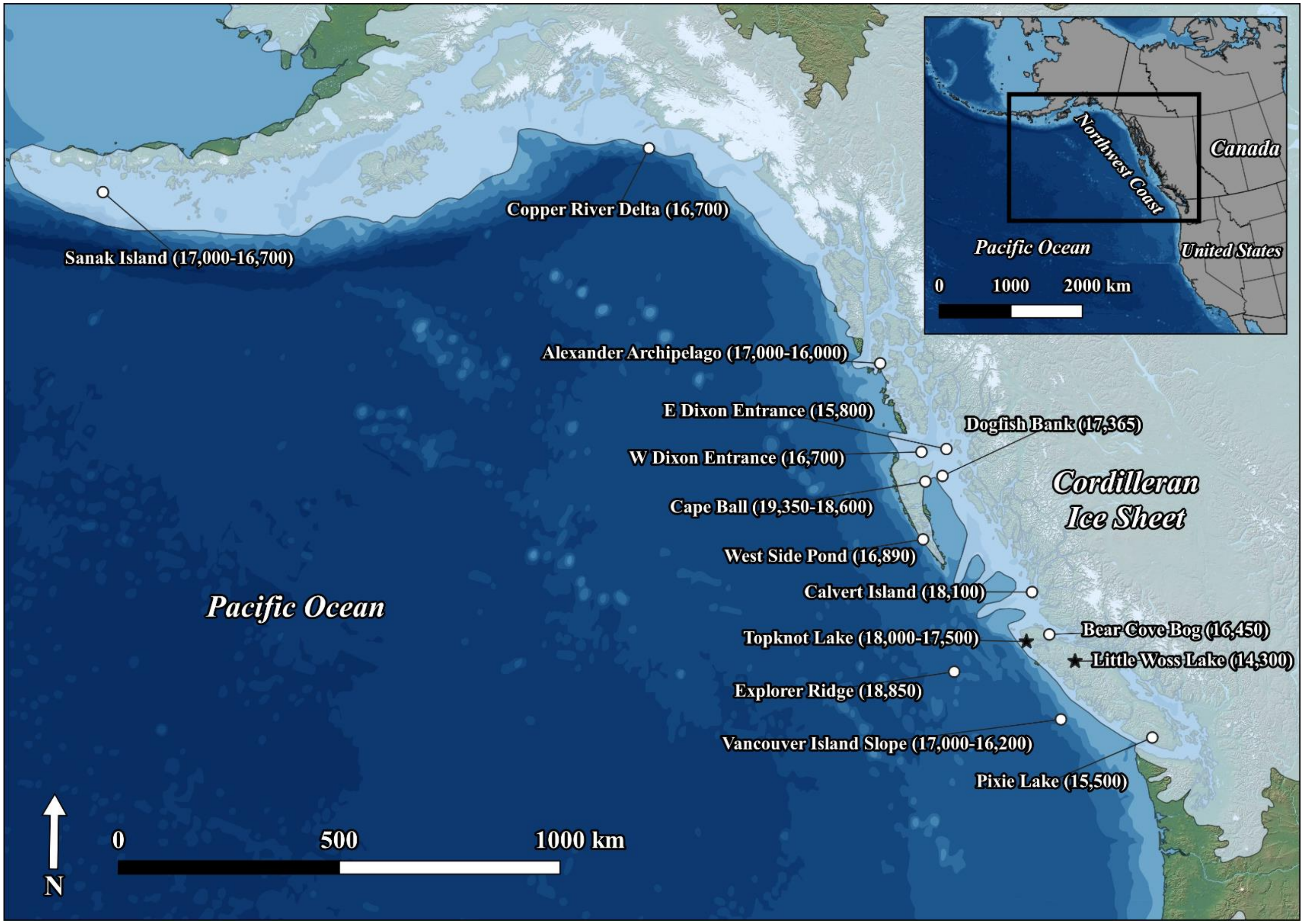


***Figure 9.*** *Stages of the late Pleistocene Fraser Glaciation on the Pacific northwest coast of North America by region, generally from northwest to southeast. Specific dates and citations noted in the text.*

***Figure 10.*** *Maximum glacial cover (ca. 20,000-17,500 cal BP; after Dyke 2004) and deglaciation chronology for selected sites on the Pacific northwest coast of North America. Glacial cover is depicted in light blue. Ages are noted after the site name and are expressed in calibrated years before present (cal BP). Citations for each site noted in the text. Map data: Natural Earth, Viewfinder DEM, GeoBC.*



### *Southern and Southeastern Alaska*

Mann and Peteet (1994) date the LLGM on Kodiak Island in southern Alaska to between 26,900 and 17,900 cal BP, with deglacial processes beginning toward the latter end of that period. Radiocarbon dates on cores recovered from sites in the Aleutian Islands and Gulf of Alaska also demonstrate deglacial processes with ice-free areas by at least 17,000-16,700 cal BP (Addison et al. 2012; Misarti et al. 2012). Lacustrine sediments recovered from Sanak Island off the southern coast of the Alaska Peninsula indicate that the eastern edge of the Aleutian Islands was deglaciated by at least 17,000 cal BP, and limited ice-override features with thin till deposits on the island suggest that glacial cover was constrained in both depth and duration through the LLGM. Furthermore, pollen data suggest an arid terrestrial ecosystem on the island by at least 16,300 cal BP (Misarti et al. 2012).

To the east, a marine core collected from the waters off the Copper River delta indicates that glacial ice had retreated from the central part of the Gulf of Alaska by approximately 16,700 cal BP (Davies et al. 2011; Addison et al. 2012). However, high sedimentation rates in the core through this period suggest outwash from a land-based or near-shore glacial source (Addison et al. 2012). The persistence of this ice is potentially related to several brief glacial re-advances in the region as documented on the Kenai Peninsula by Reger et al. (2007). Using a $\delta^{18}$O reduction as a proxy for meltwater input, Davies et al. (2011) note that regional glaciers in the Gulf of Alaska began retreating into the mainland valleys by at least 14,790 cal BP.

Glacial records from southeastern Alaska indicate a somewhat more complicated deglacial history due to regional isostasy and the dynamics of local ice, though generally the LLGM in southeastern Alaska stretches from ca. 20,000 to 17,000 cal BP (Lesnek et al. 2018). Carrara et al. (2007) identified several key areas spanning the length of the Alexander Archipelago that demonstrate evidence for possible glacial refugia (though see a more recent analysis by Lesnek et al. [2018], discussed below). The hypothesized refugia include areas on (or around the now-submerged continental shelf adjacent to) many of the outer coastal portions of the islands from the Fairweather Ground in the north down to Dall Island and Forrester Island in the south (Carrara et al. 2007). Potential refugia were noted by their lack of diagnostic glacial flow or depositional features including U-shaped valleys, smoothed ridgelines, cirques, arêtes, moraines, eskers, drumlins, or roches moutonées.

Marine cores and bathymetric data from the continental shelf off the Alexander Archipelago also suggest the presence of a 30- to 60-m crustal forebulge leading to lower relative sea level—and therefore exposed outer coastal plains—during the late glacial and early post-glacial period (Carrara et al. 2007; Carlson and Baichtal 2015). However, cosmogenic $^{10}$Be exposure-dating conducted by Lesnek et al. (2018) suggests that several of the refugia theorized by Carrara et al. (2007) that are now above sea level were likely covered by the CIS for at least part of the last glaciation from approximately 20,000 to 17,000 cal BP. Lesnek et al. do not, however, address glacial coverage of now-submerged refugia along the continental shelf. Several ecological and DNA-based studies further corroborate the presence of isolated populations of brown bears (*Ursus arctos*) (Heaton et al. 1996), long-tailed vole (*Microtus longicaudus*) (Conroy and Cook 2000), ermine (*Mustela erminea*) (Fleming and Cook 2002), chum salmon (*Oncorhynchus keta*) (Kondzela et al. 1994), shore pine (Hansen and Engstrom 1996; Godbout et al. 2008; Ager 2019), and other species in southeast Alaska either throughout or immediately following the LLGM (see Soltis et al. 1997 and Shafer et al. 2010). In addition to these possible refugial outer coastal areas, radiocarbon ages derived from shells in raised beach deposits indicate that all the major waterways of southeastern Alaska were ice-free by no later than 14,800-13,600 cal BP and were likely deglaciated even earlier, by 16,000 cal BP (Briner et al. 2017; Lesnek et al. 2018).

### *Haida Gwaii*

The late Pleistocene glacial history of Haida Gwaii was heavily influenced by the dynamics of local mountain glaciers in addition to the advance and retreat of the main mass of the Cordilleran ice (Clague 1989a; Barrie and Conway 1999; Mathewes and Clague 2017). Though parts of the archipelago such as Hecate Lowlands escaped glaciation, the LLGM on Haida Gwaii extends from ca. 25,000 to 19,000 cal BP (Warner et al. 1982; Barrie and Conway 1999; Shaw et al. 2019). Recent research by Mathewes and Clague (2017) indicates that the Hecate lobe of the CIS emerged from the Skeena River area and approached the eastern coast of Graham Island by 31,000-30,000 cal BP. The Hecate lobe slowly combined with the Dixon lobe of the CIS emerging from the Nass River area as well as with local ice flowing from mountains of Graham Island before flowing northwest out Dixon Entrance along the northern edge of Haida Gwaii and south along the outside of Hecate Lowlands before beginning rapid deglaciation between 19,000 and 17,000 cal BP (Warner et al. 1982; Mathewes and Clague 2017; Shaw et al. 2019). Palaeoecological evidence indicates that Dogfish Bank in centre Hecate Strait was ice-free by at least ca. 17,000 cal BP (Lacourse et al. 2005). The local Haida Gwaii Ice Cap sourced from the Queen Charlotte Mountains developed slower and with less overall coverage than the main Cordilleran ice, likely growing to a maximum thickness of approximately 500 metres over the central portion of Haida Gwaii (Clague 1983; Barrie et al. 2005).

However, as in southeastern Alaska, the presence of many endemic or genetically distinct populations of plant and animal species suggest that parts of Haida Gwaii remained unglaciated throughout this period, likely in the lowlands of Hecate Strait that were affected by eustatic sea level lowering or by isostatic glacio-distal forebulge effects (Clague 1983; Heusser 1989; Barrie et al. 2005; Mathewes and Clague 2017). Species that demonstrate survival in genetic refugia in Haida Gwaii include chum, sockeye (*Oncorhynchus nerka*), and coho salmon (*O. kisutch*) (Kondzela et al. 1994; Smith et al. 2001; Beacham et al. 2006), Haida Gwaii slug (*Staala gwaii*)

(Ovaska and Sopuck 2013), black bear (*Ursus americanus*) (Byun et al. 1997; Reimchen and Byun 2005), ermine (Fleming and Cook 2002), Steller's jay (*Cyanocitta stelleri*) (Burg et al. 2005), shore pine (Godbout et al. 2008), Sitka spruce (Gapare and Aitken 2005; Gapare et al. 2005), and others (see Soltis et al. 1997, Reimchen and Byun 2005, and Shafer et al. 2010).

Barrie and Conway (1999) collected marine cores in Dixon Entrance between the Alexander Archipelago of southeastern Alaska and the northern coast of Graham Island in Haida Gwaii, inferring the presence of ice-contact and ice-proximal deposits based on core lithology. Radiocarbon dates derived from these deposits suggest full deglaciation at the western end of Dixon Entrance by ca. 16,700 cal BP (13,770 ± 100 $^{14}$C BP, TO-3489) and at the eastern end of the entrance by 15,800 cal BP (13,140 ± 70 $^{14}$C BP, CAMS-33806). As these times represent open-water conditions, it is possible that seasonal ice rafting mixed with more open periods existed much earlier (Barrie and Conway 1999). Further south, the area around Cape Ball on the eastern side of Graham Island was deglaciated earlier, some time before approximately 18,600 cal BP (15,400 ± 190 $^{14}$C BP, GSC-3319), potentially beginning by approximately 19,350 cal BP (16,000 ± 570 $^{14}$C BP, GSC-3370) (Warner et al.1982; Warner 1984; Dyke 2004). This area would likely have been under the effects of more limited ice cover from the Haida Gwaii Ice Cap rather than the main CIS, which flowed to the northwest and south but left parts of Hecate Lowlands uncovered (Clague 1983; Barrie et al. 2005; Shaw et al. 2019).

### *Central Coast*

The timing of the LLGM on the central coast of BC has been understudied when compared with Haida Gwaii and the Alexander Archipelago to the north. Blaise et al. (1990) associate the LLGM on both Haida Gwaii and the central coast with approximately 19,350 to 18,600 cal BP, largely based on the radiocarbon dates collected at Cape Ball by Warner et al.

(1982). Cosmogenic $^{10}$Be dating has recently corroborated these early deglacial dates (Darvill et al. 2018).

Several deep-sea cores collected from the troughs of Queen Charlotte Sound and analyzed by Luternauer et al. (1989a) give a general chronology of the deglaciation of the central coast. Evidence for ice-rafted debris in the lower stratigraphic components of these cores suggests rapid deglaciation of the troughs in Queen Charlotte Sound beginning by at least 16,400 cal BP (13,600 ± 150 $^{14}$C BP, GSC-3711), with the ice having retreated fully from the open waters sometime before approximately 15,400 cal BP (12,910 ± 90 $^{14}$C BP, TO-175) (Luternauer et al. 1989a). A deep-sea core (PAR85-01) taken along the western side of Explorer Ridge off southern Queen Charlotte Sound indicates deglacial processes even earlier, with high sedimentation rates associated with glacial melt off the continental shelf of BC beginning 18,850 cal BP (15,570 ± 170 $^{14}$C BP, RIDDL-808) (Blaise et al. 1990).

Recent research based on numerous $^{10}$Be exposure dates indicates that western margin of the CIS on the central coast of BC had retreated to the approximate location of the modern coastline by at least 18,100 cal BP, making the exposed coastal islands ice-free at this time (Darvill et al. 2018). These deglaciation dates are older than those proposed by previous models of the CIS (e.g. Seguinot et al. 2016) but fall within the general timeframe indicated by some of the deep-sea cores collected in Queen Charlotte Strait (for example Blaise et al. 1990). However, moraine deposits at locations such as Mt. Buxton on Calvert Island dating to approximately 16,600 cal BP reveal that regional re-advances or still-stands may have occurred during the deglacial window (between 18,850 and 13,800 cal BP) indicated by previous studies (Luternauer et al. 1989a; Blaise et al. 1990; Eamer 2017; Darvill et al. 2018). Shaw et al. (2019) use palaeogeography on the ocean floor of Hecate Strait and Queen Charlotte Sound to clarify these processes further, suggesting that ice moving south along the eastern edge of Hecate Lowlands in

Haida Gwaii flowed out Moresby Trough, along with other lobes emerging from the Coast Mountains that flowed out Mitchell's Trough and Goose Island Trough. They argue that the surfaces between these troughs were exposed by lowered eustatic sea level during the LLGM and remained unglaciated islands as ice retreated after ca. 18,100 cal BP (Darvill et al. 2018; Shaw et al. 2019).

### *Northern Vancouver Island*

Until the present study, the ice limit of the Fraser Glaciation on the outer coast of northern Vancouver Island has been poorly constrained. Based on a radiocarbon date from sediments overlain by recent glacial till, the eastern end of Quatsino Sound was glaciated sometime after approximately 24,300 cal BP (20,600 ± 330 $^{14}$C BP, GSC-2505) (Clague et al. 1980). Maximum glacial cover of northern Vancouver Island probably occurred at the peak of the Fraser Glaciation between ca. 19,600 and 18,000 cal BP (Clague et al. 1980; Howes 1983; Al-Suwaidi et al. 2006). To the north, marine cores indicate that calving and glacial retreat from the open waters of Queen Charlotte Sound into mainland fjords had completed by sometime before 15,400 cal BP (Luternauer et al. 1989a).

Specific site histories at locations including Bear Cove Bog (Hebda 1983), Brooks Peninsula (Howes 1997; Hebda 1997), Misty Lake (Lacourse 2005), and Port Eliza Cave (Ward et al. 2003; Al-Suwaidi et al. 2006) provide local constraints on ice limits, but a general history of the Fraser Glaciation on northern Vancouver Island is difficult to construct with available data due to variable isostatic, eustatic, and tectonic factors in the region. Basal radiocarbon dates at Bear Cove Bog indicate that the CIS retreated from northern Vancouver Island and into Queen Charlotte Strait by at least 16,450 cal BP (13,630 ± 130 $^{14}$C BP, WAT-721) (Hebda 1983). Approximately 20 km to the southeast of Bear Cove Bog, basal radiocarbon dates from Misty Lake extend to at least 14,900 cal BP (Lacourse 2005). Both Bear Cove Bog and Misty Lake are located on the glacio-proximal eastern side of northern Vancouver Island, and therefore would likely have undergone deglaciation later than outer coastal sites.

On Brooks Peninsula on northern Vancouver Island, Howes (1997) notes that based on a suite of geomorphic features, at least 7 $km^2$ to 9.5 $km^2$ of the higher elevations and ridgetops of the landform were not overridden by ice during the Fraser Glaciation. Hebda (1997) reconstructed the palaeoecology of several sites on the peninsula extending back to approximately 14,500 cal BP (12,250 ± 790 $^{14}C$ BP, WAT-924) based on fossil pollen assemblages. The extensive geological and palaeoecological record of Brooks Peninsula suggests that at least part of the outer coast of northern Vancouver Island remained unglaciated through the peak of the Fraser Glaciation as nunataks (Howes 1997). The discovery of several plant species on Brooks Peninsula previously thought to be endemic to Haida Gwaii further suggests the presence of refugial environments on northern Vancouver Island (Ogilvie 1997). Radiocarbon dates from animal bones recovered from Port Eliza Cave indicate an interglacial environment on the western coast of Vancouver Island between Brooks Peninsula and Nootka Sound until approximately 19,600 cal BP (16,270 ± 170 $^{14}C$ BP, CAMS-88275), above which laminated silt and clay deposits and a lack of dated faunal remains suggest glacial ice proximal to the site (Al-Suwaidi et al. 2006). Deposition resumes in the cave by approximately 14,350 cal BP (12,340 ± 50 $^{14}C$ BP, CAMS-97342). Port Eliza Cave may have undergone a somewhat slower deglacial process than outer coastal areas farther north because it is closer to local glacial sources in the higher elevations of the Vancouver Island Ranges.

### *Central/Southern Vancouver Island and the Lower Mainland*

Some late Pleistocene geological and palaeoecological work has been conducted on the central and southern portions of the west coast of Vancouver Island, but the spatial resolution of these data is relatively low (Clague et al. 1980; Brown and Hebda 2002; Cosma et al. 2008;

Hendy and Cosma 2008). A radiocarbon date derived from a sample of shore pine wood embedded in glacial outwash gravels near Tofino suggests that the LLGM occurred sometime after approximately 20,150 cal BP (16,700 ± 150 $^{14}$C BP, GSC-2768) (Clague et al. 1980). Ice expansion after this time is corroborated by increased offshore glaciomarine sedimentation at ca. 19,500 cal BP described by Cosma et al. (2008) from core MD02-2496. Dates obtained from this same core are associated with regional deglacial processes, including the presence of ice-rafted debris, between 17,000 and 16,200 cal BP (Cosma et al. 2008; Hendy and Cosma 2008). These data indicate that the LLGM occurred between ca. 19,500 and 17,000 cal BP. This post-17,000 cal BP deglaciation window is corroborated by a similar core collected from Effingham Inlet in Barkley Sound (MD02-2494), which suggests that deposition following the retreat of Cordilleran ice began at the site by ca. 17,300 cal BP (Dallimore et al. 2008).

Lacustrine records from Whyac Lake (basal date of 12,800 cal BP [10,860 ± 130 $^{14}$C BP]), Pixie Lake (basal date of 15,550 cal BP [12,990 ± 180 $^{14}$C BP, BETA-81084]), and East Sooke Fen (basal date of 13,500 cal BP [11,700 ± 140 $^{14}$C BP, BETA-86809]) indicate that the western side of southern Vancouver Island was deglaciated and vegetated by at least 15,550 cal BP with a well-established open woodland ecosystem of *Pinus*, *Alnus*, and ferns (Brown and Hebda 2002).

The glacial history of the southern tip of Vancouver Island, Fraser Lowland, and the Salish Sea is well known. Initial glaciation of southwestern BC occurred during the Coquitlam-Evans Creek Stade, which began by approximately 26,000 cal BP (21,700 ± 130 $^{14}$C BP, GSC-2416) and reached its maximum by approximately 25,400 cal BP (Armstrong et al. 1965; Hicock and Armstrong 1981; Clague and James 2002). This glacial event was likely to have been mostly confined to the mountain valleys and lowlands of what is now the Lower Mainland (Hicock and Armstrong 1981). Following this, from approximately 22,600 cal BP (18,700 ± 170 $^{14}$C BP, GSC-

2344) to 22,150 cal BP (18,300 ± 170 $^{14}$C BP, GSC-2322), a brief climatic amelioration is expressed as the Port Moody Interstade (Hicock et al. 1982; Hicock and Armstrong 1985). The Port Moody Interstade then descends into the period of most extensive glaciation on the coast of BC: the Vashon Stade, which began by approximately 22,150 cal BP and reached its maximum extent by approximately 18,200 to 17,600 cal BP (15,000 to 14,500 $^{14}$C BP) (Clague et al. 1980; Hicock and Armstrong 1985). During the deepest glaciation of the coast, Cordilleran ice flowing from the Coast Mountains of BC and extended south through the Strait of Georgia before splitting into two lobes at the far southeastern corner of Vancouver Island, with one lobe flowing south into Puget Sound and another flowing west into Juan de Fuca Strait (Armstrong et al. 1965; Clague 1989b; Clague and James 2002). Following the LLGM, ice began to retreat from Puget Lowland and the southern portion of the Georgia Depression during the Everson Interstade, beginning after approximately 15,600 cal BP (13,500 ± 220 $^{14}$C BP, GSC-3124) (Armstrong et al. 1965; Fulton 1971; Hicock and Armstrong 1985). Smaller glacial re-advances, collectively termed the Sumas Stade in southwestern BC, lasted from approximately 13,600 to 13,200 cal BP, after which minor fluctuations in ice loading but no major glaciations have occurred in the Georgia Depression (Fulton 1971; Armstrong 1981; Clague and James 2002).

### 2.2.2 Global Eustatic Sea Level Change

During the global Last Glacial Maximum at approximately 21,000 cal BP, global mean sea level was at least as low as 120 m below present as a result of eustatic sea level change with massive volumes of water locked up in continental ice sheets (Peltier and Fairbanks 2006). Other factors likely also contributed to global sea level change, including reduced sedimentation on the sea floor and increased seawater density, though these had a lesser impact than eustasy resulting from glacial melt (Shugar et al. 2014). Following this lowstand, interplay between punctuated events of rapid global sea level rise (Khanna et al. 2017) and regional isostatic factors (Clague 1983; McLaren et al. 2014; Shugar et al. 2014; and others) influenced local biogeography, including human migration. Depressed global sea levels through the GLGM and during earlier Pleistocene glaciations facilitated the dispersal of modern humans into various regions, including migrations into Arabia by crossing the Red Sea basin from the Horn of Africa (Lambeck et al. 2011), into Sahul by crossing from the palaeo-landforms and islands of Sunda (O'Connell et al. 2010; O'Connor 2010; Clarkson et al. 2017), and into the Ryukyu Islands of the Japanese archipelago (Kaifu et al. 2015). Despite increased land area, however, some of these crossings would certainly have required the use of watercraft by early peoples (O'Connor 2010; Ikeya 2015; O'Connor 2015). The complex, rocky shorelines of the northwest coast of North America are no exception. Even during periods of lower sea level on the outer coast, the first people navigating the region would have required extensive seafaring ability as well as knowledge of both pelagic and littoral resources (see Ames 2002; Erlandson et al. 2007; Erlandson et al. 2011).

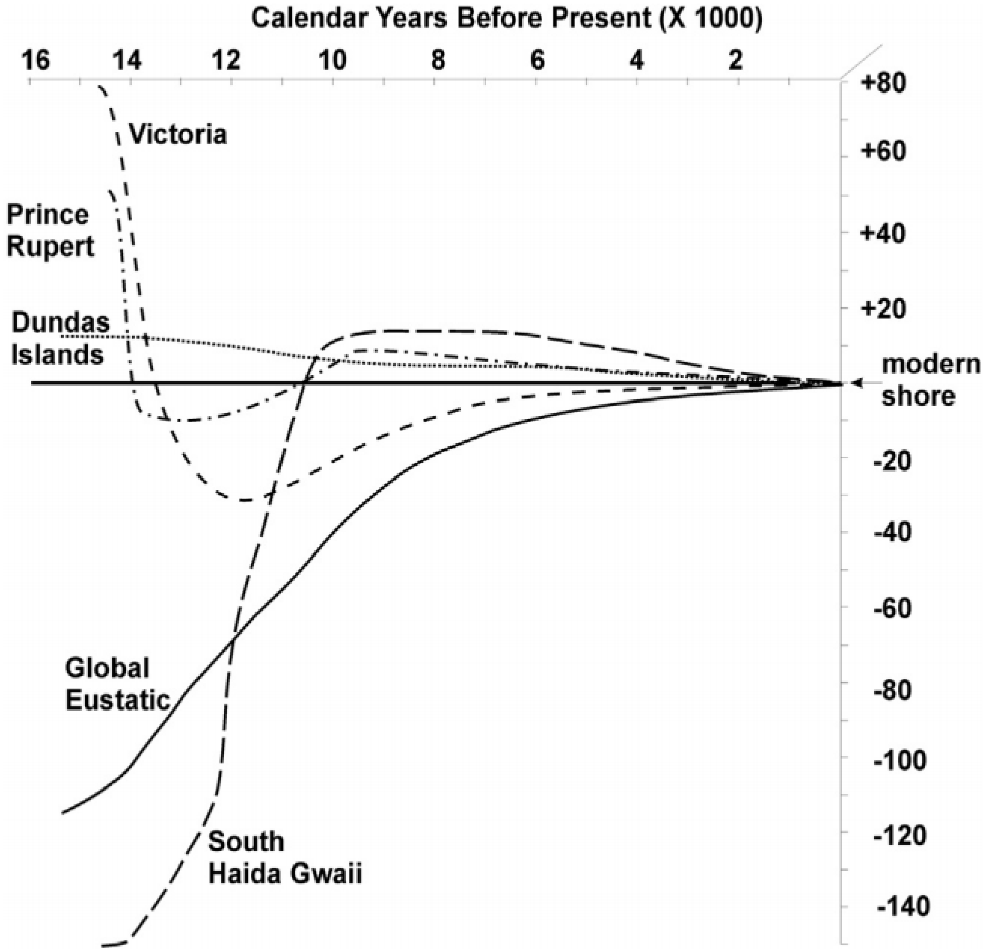
### 2.2.3 Documenting Glacio-Isostatic Effects on Sea Level History

In addition to global eustatic effects, glacial activity produced much more regionally specific sea level histories along the northwest coast of North America. Isostatic depression and rebound by Cordilleran ice resting on the continental edge as well as related crustal forebulge effects at some distance from the ice mass led to much higher relative sea levels on some parts of the coast, and relative sea levels even lower than the eustatic lowstand elsewhere (Clague 1983; Josenhans et al. 1995; McLaren et al. 2014; Shugar et al. 2014; Mathewes and Clague 2017; Fedje et al. 2018).

Changes in relative sea level across coastal British Columbia and southeastern Alaska (see Figure 11) have been demonstrated using various methods, the most common of which include analyzing diatom assemblages from isolation basin cores, dating marine and terrestrial indicators in archaeological deposits and sedimentary exposures, and identifying palaeo-landforms (Clague et al. 1982; Luternauer et al. 1989b; Josenhans et al. 1995; Fedje et al. 2005c;

James et al. 2009; McLaren et al. 2011; McLaren et al. 2014; Carlson and Baichtal 2015; Letham et al. 2016; Fedje et al. 2018).

***Figure 11.*** *Late Pleistocene and Holocene relative sea levels for selected regions on the Pacific northwest coast of North America (after Fedje et al. 2018, Figure 2).*



Isolation basin cores, usually extracted from modern or palaeo ponds, lakes, and lagoons, are examined for the presence of specific biological or geological proxies in undisturbed stratigraphic context (see Fedje et al. 2018 for an overview of methods and interpretation). Marine remains including shells, foraminifera, or diatoms can indicate sea level transgression at a site, and terrestrial deposits including peat layers, tsunami wash layers, or volcanic tephra can provide other dateable events or geomorphic processes. The presence of certain diatom species is a good indicator of relative sea level, as many species are constrained by the salinity of the basin (Hustedt 1953; Fedje et al. 2018). The ratio of diatom species and their salinity affiliations indicates whether the water at the time of deposition was fresh, brackish, or marine. By determining the age of the brackish transitional layer between fresh and saltwater components, relative sea level at that time can be said to be above, around the same level as, or below the elevation of the basin sill represented in the core. The presence or absence of other marine deposits (such as shells) or terrestrial deposits (such as a palaeosol) in the stratigraphy of the core can also help constrain relative sea level (for example, Luternauer et al. 1989b; Lacourse et al. 2003). See Fedje et al. (2005c), McLaren et al. (2014), Letham et al. (2016), and Fedje et al. (2018) for examples of using diatoms and other marine and terrestrial indicators to construct relative sea level curves for regions across coastal BC.

Though more difficult to pinpoint on the landscape than isolation basin targets, the presence of archaeological sites and features can also frame relative sea level histories with their associated dates (McLaren et al. 2014; Carlson and Baichtal 2015; Letham et al. 2016). If a hearth feature, for example, is dated to 10,000 cal BP and is now located at +15 m elevation, it provides a maximum possible sea level at that time in that area, as a hearth feature would not be built underwater (Fedje and Christensen 1999; Carlson and Baichtal 2015). Often, associated sediments (beach cobbles, midden, windblown dunes, etc.) can indicate the depositional environment for the feature, which may inform further investigation to help constrain relative sea level in the area at that time (Eamer 2017; Fedje et al. 2018; Lausanne 2018).

Sea level histories can also be interpreted through examining palaeo-landforms such as terraces, river courses, and deltas that do not appear to reflect current geomorphic processes, either above or below current sea level. Some researchers on the northwest coast of North America have included analysis of palaeo-landforms in their discussion of sea level histories

(Luternauer et al. 1989b; Josenhans et al. 1995; Fedje et al. 2005c; Fedje et al. 2018; Mackie et al. 2018; and see Mackie et al. 2011 for a review of similar studies).

In some regions including Haida Gwaii and portions of the central coast and the northern and southern Salish Sea, the relative effects of isostasy are known (see Clague 1983; Hutchinson et al. 2004; James et al. 2009; McLaren et al. 2014; Shugar et al. 2014; Fedje et al. 2018). However, glacial coverage on northern Vancouver Island during the LLGM and its impact on relative sea level is less well-documented, especially for outer coastal portions of the island. Glacial coverage has been inferred by some palaeoecological studies (e.g. Hebda 1983; Hebda et al. 1997; Howes 1997; Al-Suwaidi et al. 2006), but this has not generally been used to address late-glacial isostatic sea level change.

### 2.2.4 Relative Sea Level Change on the Northwest Coast

#### *Southeastern Alaska*

Glacial and postglacial sea level curves for the islands and mainland of far southeastern Alaska demonstrate a distinct dichotomy: sea levels higher than modern near the continent; and sea levels significantly lower than modern among the outer islands. Along the mainland of southeastern Alaska, higher relative sea level ca. 13,000 cal BP resulted from isostatic depression by glacial ice, while among the islands of the outer Alexander Archipelago significantly lower relative sea level (~122 m below modern) may have been the result of combined global eustatic lowering and isostatic crustal forebulge effects, also seen elsewhere on the coast such as in Haida Gwaii (Shugar et al. 2014; Carlson and Baichtal 2015).

Using radiocarbon dates derived from raised marine deposits and archaeological sites, Mobley (1988) constructed an initial model of Holocene sea level change for southeastern Alaska focused on Heceta Island and Prince of Wales Island. Carlson and Baichtal (2015) have since expanded on this earlier curve and have created a predictive archaeological model for Prince of Wales Island as well as smaller islands to the west and northwest. As with Mobley's (1988) earlier model, the updated curve relies on dates from both raised marine sediments and archaeological deposits to constrain relative sea level through time (Carlson and Baichtal 2015). These dates suggest that combined global eustatic sea level drop and isostatic crustal forebulge effects led to much lower relative sea levels on Prince of Wales Island in the late Pleistocene, which lasted until at least ca. 11,200 cal BP (9,730 $^{14}$C BP) when natural shell-bearing deposits appear at 18 m above modern sea level (Carlson and Baichtal 2015). Closer to the continental ice mass, relative sea level was much higher, reflecting between 52 and 256 m of isostatic depression and/or tectonic uplift at approximately 14,800 cal BP (12,500 $^{14}$C BP) (Shugar et al 2014; Carlson and Baichtal 2015). Following the collapse of the forebulge, relative sea level rose quickly to approximately 17.5 m above modern by 10,200-9,500 cal BP (9,000-8,500 $^{14}$C BP) before falling slowly to modern levels by approximately 5,700 cal BP (5,000 $^{14}$C BP). Based on the output of their predictive model, Carlson and Baichtal (2015) surveyed an elevation range of 16-22 +/-1 m for early period archaeological remains and discovered 70 new sites, including 11 early Holocene sites. However, while such relative sea level modeling has been successful for archaeological exploration targeting the early Holocene when sea levels were higher than present, the archaeological record of the exposed outer coastal plain in southeastern Alaska prior to 11,200 cal BP remains poorly known (but see Monteleone et al. 2012 for a model targeting submerged late Pleistocene archaeological sites in this region).

### *Haida Gwaii*

Sea levels on Haida Gwaii have shifted immensely since the end of the Pleistocene. During the early postglacial period, Haida Gwaii experienced very low relative sea levels due to crustal forebulge and global eustatic effects on the far outer coast of BC (Clague et al. 1982; Josenhans et al. 1997; Hetherington et al. 2004; Barrie et al. 2005). As ice retreated, the forebulge eventually migrated towards the mainland, submerging previously exposed areas of Haida Gwaii and Hecate Strait.

While under the uplifting effects of the forebulge, much of Hecate Strait to the east of Haida Gwaii would have been an open terrestrial landscape, with relative sea levels at least 30 m below modern on Dogfish Bank between approximately 17,365 and 14,570 cal BP (Barrie et al. 1993; Lacourse et al. 2005). By ca. 14,150 cal BP (12,500 $^{14}$C BP), relative sea level in southeastern Haida Gwaii and Hecate Strait was 150 m below modern as a result of eustatic and isostatic factors (Josenhans et al. 1995; Fedje et al. 2005c). Sea levels rose sharply following deglaciation as the forebulge migrated away from Haida Gwaii and toward the mainland—especially after 12,800 cal BP (11,000 $^{14}$C BP)—to a maximum of 15 m above modern by 10,000 cal BP (8,900 $^{14}$C BP) (Clague et al. 1982; Fedje et al. 2005c). Sea level remained around this height for the next several thousand years, eventually beginning a slow decline towards present levels around 5,700 cal BP (5,000 $^{14}$C BP) as a result of tectonic factors (Fedje et al. 2005c). Due to the shallow continental shelf to the east of Haida Gwaii in Hecate Strait, this provides a somewhat different archaeological picture from other regions, even southeastern Alaska which also saw the effects of a crustal forebulge (Carlson and Baichtal 2015). The earliest sites on Haida Gwaii, then, may be found far underwater. For example, Fedje and Josenhans (2000) recovered a single lithic from 53 m below modern sea level in Juan Perez Sound, east of Moresby Island in southern Haida Gwaii. More recently, preliminary research based on side-scan sonar conducted by autonomous underwater vehicle over the sea floor in the same area indicates the presence of submerged stone alignments and a possible house depression (Mackie et al. 2018). However, subsequent rapidly rising sea levels in the region have led to early Holocene sites being found above modern sea level, as demonstrated by finds at Kilgii Gwaay and on Richardson Island (Fedje et al. 2005a; Storey 2008; Cohen 2014; Cohen 2019). See Table 1, Appendix A, and Mackie et al. (2011) for a review of early archaeology on Haida Gwaii.

### *Outer North Coast – Dundas Islands*

Located between regions that experienced higher relative sea levels due to isostatic depression (Cordilleran ice on the mainland to the east) and regions that experienced lower relative sea levels due to crustal forebulge and eustatic effects (the outer coast and Haida Gwaii to the southwest), the Dundas Islands on the outer north coast have seen a comparatively small change in sea level since the late Pleistocene. Data derived from lake basin coring and archaeological excavation in the Dundas Islands indicate a drop in relative sea level as little as 13-14 m since ca. 15,000 cal BP (McLaren 2008). This suggests the presence of a 'sea level hinge' where the shorelines of these islands have been mostly consistent over thousands of years as a result of balanced eustatic and isostatic influences on relative sea level in the region (McLaren 2008; McLaren et al. 2011; McLaren et al. 2014). Sea levels descended slowly through the terminal Pleistocene and the Holocene, to 10 m above modern by ca. 12,130 cal BP (10,320 ± 40 $^{14}$C BP), to 6.5 m above modern by ca. 6,070 cal BP (5,280 ± 100 $^{14}$C BP), and to 3.5 m above modern by ca. 4,330 cal BP (3,885 ± 20 $^{14}$C BP) (McLaren et al. 2011). The remaining drop has occurred slowly since that time.

Farther to the east and closer to the sources of Cordilleran ice, the postglacial sea level for nearby Prince Rupert Harbour was not as consistent as the 'hinge' on the Dundas Islands. Between 14,500 and 13,500 cal BP, relative sea level dropped from 50 m above to 6 m below current sea level (Letham et al. 2016). By the end of the Younger Dryas, relative sea level had risen once more to 6 m above modern before slowly dropping through the Holocene to the level of the present shoreline by ca. 2,000-1,500 cal BP (Letham et al. 2016).

### *Central Coast*

To north of Vancouver Island, parts of the Hakai Passage region of the central coast of British Columbia have also been identified as a ‘hinge’ where relative sea level has remained mostly stable since at least the late Pleistocene (McLaren et al. 2014). However, due to some variance in relative sea level histories across the central coast, McLaren et al. (2014) split the region into four separate curves: Hakai West, Hakai East, Cook Bank, and Goose Bank. Two of these, Hakai West (including the hinge) and Hakai East, demonstrate limited change in sea level since the late Pleistocene, while the remaining two, Cook Bank and Goose Bank, represent areas of the outer coast with greater susceptibility to global eustatic and isostatic forebulge effects, and thus demonstrate much lower relative sea levels in the early postglacial period (McLaren et al. 2014). The lowstand for relative sea level on Goose Bank, a shallow portion of the continental shelf between Haida Gwaii and the central coast, was as deep as 133 m below modern between 14,600 and 13,980 cal BP, forming a low-lying island (Hetherington et al. 2004; McLaren et al. 2014). The island then began to shrink, with relative sea levels rising to 122 m below modern between 11,600 and 11,250 cal BP. Data for the timing of the subsequent sea level transgression and inundation of Goose Bank are unavailable.

### *Northern Vancouver Island*

Differing portions of northern Vancouver Island likely saw the effects of isostatic depression, the ‘hinge,’ and the outer coastal forebulge, though not as extreme as elsewhere on the coast. Like Goose Bank, Cook Bank off the northern tip of Vancouver Island consists of a relatively shallow expanse of the continental shelf which also experienced lower relative sea level as a result of crustal forebulge effects (McLaren et al. 2014). Submarine sediment cores demonstrate a relative sea level lowstand of 135 m below modern on Cook Bank between 15,400 and 14,100 cal BP (Hetherington et al. 2004). Submerged conifer roots and a paleosol recovered from this area indicate sea level least 95 m below modern between 12,931 and 12,000 cal BP

(Luternauer et al. 1989b; Lacourse et al. 2003), with palaeogeography including wave-cut terraces at 100 m below modern sea level (McLaren et al. 2014). These data indicate the presence of a broad outer coastal plain connected to the northern tip of the island at this time. By approximately 11,400 cal BP, however, this plain was drowned by weakening isostasy and rising global eustatic sea level resulting from glacial retreat and climatic amelioration (Lacourse et al. 2003; McLaren et al. 2014).

The outer coastal portions of northern Vancouver Island itself probably experienced relatively little isostatic depression compared with areas closer to the source areas of the CIS in the Coast Mountains. Minor isostatic depression (and correlated higher relative sea level) may be the result of the combined weight of the distant CIS and local ice in the Vancouver Island Ranges (Howes 1983). However, like Haida Gwaii and Cook Bank to the north, the region was likely most affected by global eustatic and isostatic forebulge effects resulting from its distance from the Cordilleran ice. However, the cumulative effects of eustatic and isostatic influences along the west coast of Vancouver Island south of Cook Bank are poorly known. Maximum late Pleistocene/early Holocene marine transgression has been inferred up to 92 m asl on the eastern side of northern Vancouver Island based on the presence of raised delta deposits composed of sandy gravels, likely laid down as glaciers retreated from the region (Howes 1983). This inundation may have been the result of isostatic depression and subsequent melt of combined mainland glaciers and Vancouver Island ice flowing north out of the Nimpkish Valley. Deposits indicating a notably less extensive early Holocene marine transgression have been observed at several locations on the outer west coast of Vancouver Island, including 20 m above modern at Cape Scott, 19-20 m above modern on Brooks Peninsula, and 32-34 m above modern at Hesquiat Harbour (Howes 1981b; Clague et al. 1982; Howes 1983; Howes 1997). Currently, research based on isolation basin coring, archaeological deposits, and sedimentary exposures is ongoing in

Quatsino Sound, focused on constructing a robust history of relative sea level change on the outer coast of northern Vancouver Island (D. McLaren, personal communication 2019).

### *Southern Vancouver Island – Barkley Sound*

Dallimore et al. (2008) constructed a relative sea level curve spanning the last 14,000 years for Barkley Sound on the west coast of Vancouver Island based on previous research (Hutchinson 1992) as well as sedimentology and palaeoenvironmental indicators found within a 41-metre marine core recovered from Effingham Inlet. At approximately 13,500 cal BP, the relative sea level lowstand was 46 m below modern as a result of eustatic factors combined with isostatic depression during the terminal Pleistocene. Shorelines remained relatively stable—at least 40 m below modern—until approximately 11,125 cal BP, after which relative sea level rose steadily over the Holocene, surpassing modern sea level at approximately 6,000 cal BP before reaching a highstand of approximately 3 m above modern. Relative sea level in Barkley Sound has slowly regressed to the modern level since that time (Dallimore et al. 2008). This sea level change appears to have been influenced largely by global eustasy, though Dallimore et al. (2008) suggest that isostatic depression of the crust from glacial coverage of southern Vancouver Island led to shorelines in Barkley Sound being consistently at least a few metres higher than the global average throughout the Holocene.

### *Northern Salish Sea –Quadra Island*

Unlike most of the regions mentioned previously, the northern Salish Sea—between the eastern side of Vancouver Island and the mainland of British Columbia—experienced significantly higher relative sea levels during the late Pleistocene largely due to proximity to the Cordilleran ice and resulting isostatic depression of the crust (Fedje et al. 2018). This crustal depression led to a relative sea level of at least 197 m above modern at Assu Bog on Quadra Island at 14,300 cal BP (Fedje et al. 2018). As the ice retreated, however, relative sea level regressed rapidly in the region. In various locations around Quadra Island, relative sea level regressed from the late Pleistocene highstand to 144 m above modern by 13,500 cal BP, 75 m above modern by 13,100 cal BP, 26 m above modern by 12,800 cal BP, 7 m above modern by 12,300 cal BP, and 2-3 m above modern around 12,000 cal BP (Fedje et al. 2018).

### *Southern Salish Sea –Victoria*

Numerous studies have included data on relative sea level histories for southeastern Vancouver Island and the southern Salish Sea (Mathews et al. 1970; Clague et al. 1982; Linden and Schurer 1988; James et al. 2002; Mosher and Hewitt 2004). Combining these data with the results from eight sediment cores collected from isolation basins throughout the Greater Victoria area, James et al. (2009) constructed a robust sea level curve covering more than 14,000 years. Relative sea level was at its highest on southeastern Vancouver Island, 75 m above modern, at approximately 14,250 cal BP. This can largely be attributed to isostatic depression of the crust by glacial ice sitting over southern Vancouver Island and in the Georgia Depression, though the isostatic influence in this region was not as strong as areas proximal to the sources of Cordilleran ice to the north such as Quadra Island (Fedje et al. 2018). Following the 75-metre highstand, glacial retreat and isostatic rebound led to the rapid regression of shorelines in the region. Relative sea levels dropped to 70 m above modern at 14,000 cal BP and 30 m above modern at 13,700 cal BP, falling below modern sea level at approximately 13,200 cal BP (James et al. 2009). By 12,600 cal BP, relative sea levels had fallen to at least 15 m below modern, likely reaching a lowstand of approximately 30 m below modern by 11,200 cal BP. Shorelines began to rise again over the next several thousand years to approximately 3 m below modern by 6,000 cal BP, reaching modern sea level by 4,000 cal BP (James et al. 2009).

## 2.3 Late Pleistocene Palaeoecology of the Northwest Coast

What follows is a summary of palaeoenvironmental conditions, including the presence of flora and fauna through time on the Pacific northwest coast of North America, including southeastern Alaska, Haida Gwaii and the north coast of BC, the central coast of BC, northern Vancouver Island, and southern Vancouver Island. Vegetation histories for several sites in coastal BC from Haida Gwaii to northern Vancouver Island are presented in Figure 12. The locations of these and other sites where late Pleistocene palaeoecological studies have been conducted on the coast are depicted in Figure 13.

***Figure 12.*** *Vegetation histories for selected palaeoecological sites on the Pacific northwest coast of North America, including Little Woss Lake and Topknot Lake. Citations for each site noted in the text.*

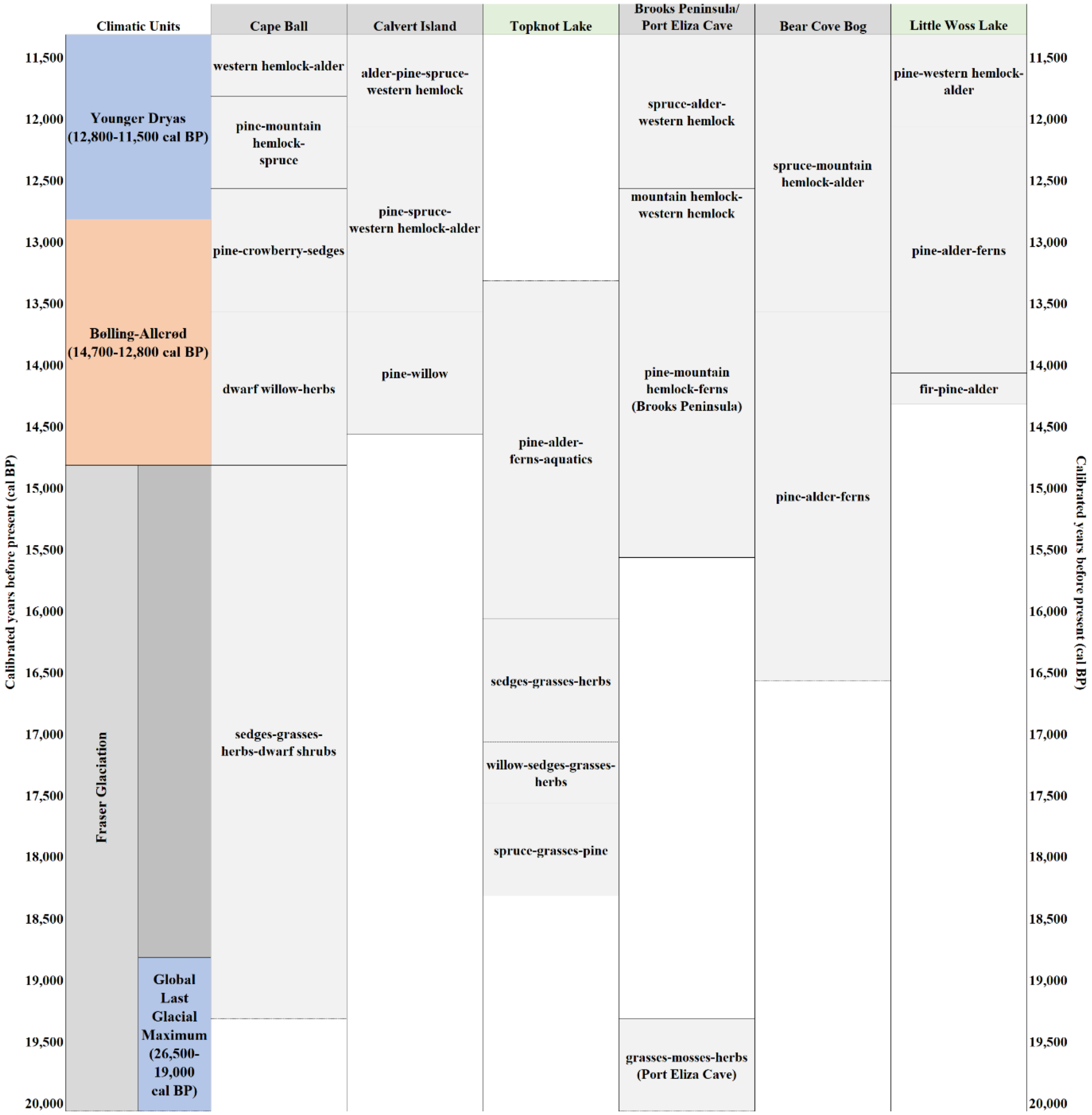


***Figure 13.*** *Selected palaeoecological sites on the Pacific northwest coast of North America with late Pleistocene deposits. Citations for each site noted in the text. Map data: Natural Earth, Viewfinder DEM, GeoBC.*

### *Southeastern Alaska*

#### *Pre-28,000 cal BP*

Prior to the advance of ice during the Fraser Glaciation, the Alexander Archipelago of Alaska was home to mammals of varying sizes, from heather vole (*Phenacomys intermedius*), brown lemming (*Lemmus trimucronatus*), hoary marmot (*Marmota caligata*), red fox (*Vulpes vulpes*), and arctic fox (*Vulpes lagopus*) to black bears, brown bears, and caribou (*Rangifer tarandus*) (Wigen 2005). Marine mammals including harbour seal (*Phoca vitulina*) and Steller sea lion (*Eumetopias jubatus*) populated the waters, and puffins (*Fratercula*), common murre (*Uria aalge*), and long-tailed duck (*Clangula hyemalis*) nested on the shorelines of the coast's many islands (Heaton and Grady 2003; Wigen 2005).

#### *ca. 28,000-16,000 cal BP*

A specimen of cold-adapted ringed seal (*Pusa hispida*) has been recovered from Shuká Káa (On Your Knees Cave) dating to ca. 28,200 cal BP, suggesting the degradation of pre-glacial conditions during the onset of the Fraser Glaciation (Heaton and Grady 2003). By the height of glaciation, and certainly between approximately 20,000 and 17,000 cal BP when most of southeastern Alaska was covered in ice, few of the pre-glacial species remained on the islands of the Alexander Archipelago (Wigen 2005; Lesnek et al. 2018). Some, however, including brown bears, voles, ermine, and chum salmon may have survived in coastal refugia (Kondzela et al. 1994; Heaton et al. 1996; Conroy and Cook 2000; Fleming and Cook 2002). Faunal records from the glacial period demonstrate the presence of cold-adapted species such as ringed seal, arctic fox, Steller sea lion, and numerous shorebirds in southeastern Alaska, though none of these appear to be present at Shuká Káa between ca. 20,500-17,600 cal BP (Heaton and Grady 2003). This may be due to inhospitable glacial conditions, or it may represent a period where the entrance to the cave itself was blocked by ice and was inaccessible to these animals.

### *ca. 16,000-11,500 cal BP*

Following the harsh conditions of the peak of glaciation in southeastern Alaska, surviving small animals as well as returning species such as bear (black and brown), caribou, river otter (*Lontra canadensis*), and wolverine (*Gulo gulo*) reappear in late Pleistocene palaeontological records in the Alexander Archipelago (Heaton and Grady 2003; Wigen 2005; Fedje et al. 2011). After this initial retreat of the ice from southeastern Alaska, vegetation also began to recolonize previously glaciated areas of the outer islands. Ager (2019) has analyzed the postglacial record from Hummingbird Lake on Baranof Island in the Alexander Archipelago extending back to ca. 15,240 cal BP. In this basal zone, which lasts until approximately 14,040 cal BP, the pollen assemblage at Hummingbird Lake is dominated by conifer parkland, including shore pine, willows (*Salix*), heathers (Ericaceae), sedges, grasses (Poaceae), and other herbs and ferns (Ager 2019). By ca. 14,040 cal BP, alder succeeds the earlier pine forest (Ager 2019). On Mitkof Island, much closer to the mainland of the Alaska Panhandle, the earliest postglacial assemblage emerges later by ca. 12,900 cal BP and is characterized by a mixed assemblage of pine and alder accompanied by sedges and ferns (Polypodiaceae) (Ager et al. 2010). Similar pine-dominated conifer parkland emerged slightly later, ca. 12,600 cal BP (10,690 ± 210 $^{14}$C BP, BETA-8735) on the Chilkat Peninsula, likely due to cooler temperatures and proximity to continental ice (Cwynar 1990). Following this postglacial biotic expansion, cool temperatures during the Younger Dryas ca. 12,800-11,500 cal BP may have affected the range of some of these species on the coast (Fedje et al. 2011). Despite this colder period, by ca. 12,240 cal BP Sitka spruce had spread to Mitkof Island along the inner portion of southeastern Alaska, extending its range to Baranof Island on the outer coast by 11,400 cal BP (Ager et al. 2010; Ager 2019). Climate amelioration following the Younger Dryas appears to have been rapid, with diverse early Holocene faunal assemblages in the Alexander Archipelago at Shuká Káa (On Your Knees Cave), Nautilus Cave, El Capitan Cave, and Bumper Cave (Heaton et al. 1996; Wigen 2005).

### *Post-11,500 cal BP*

In the early Holocene by 11,200 cal BP, mountain hemlock began to spread across both inner and outer islands to create forests of spruce and hemlock, and by ca. 9,200 cal BP, western hemlock became the dominant species on the Alexander Archipelago (Ager et al. 2010; Ager 2019). On the Chilkat Peninsula, western hemlock became established slightly later, by ca. 9,000 to 8,400 cal BP (7,870 ± 130 $^{14}$C BP, BETA-8732) (Cwynar 1990). Faunal records are rich throughout the Holocene in southeastern Alaska (Heaton et al. 1996; Wigen 2005).

## *Haida Gwaii*

### *Pre-28,000 cal BP*

On Haida Gwaii and the north coast of BC, few definite pre-glacial vertebrate faunal remains have been recovered, though the record does include remains of caribou and possibly either American mastodon (*Mammut americanum*) or woolly mammoth (*Mammuthus primigenius*)—species generally associated with herb/shrub tundra or open woodlands (Wigen 2005). At Cape Ball on the eastern side of Graham Island, Mathewes and Clague (2017) document arboreal species, including pine, as a major component of pre-glacial pollen assemblages and extending into the early period of the Fraser Glaciation ca. 31,400 to 30,000 cal BP. However, by ca. 28,600 to 27,500 cal BP at Mary Point on northern Graham Island the assemblage indicates a much colder, drier herb tundra-like landscape dominated by non-arboreal pollen including grasses and sedges, possibly the result of climatic deterioration and the approach of glaciers from Dixon Entrance (Mathewes and Clague 2017; Shaw et al. 2019).

### *ca. 26,000-16,000 cal BP*

As in the Alexander Archipelago to the north, few pre-glacial species remained on Haida Gwaii during the depths of the Fraser Glaciation until at least ca. 19,400 cal BP (Warner et al. 1982; Warner 1984; Wigen 2005; Mathewes and Clague 2017). During the LLGM there is little evidence of the caribou and proboscideans that may have previously inhabited the islands (Wigen 2005). There is, however, genetic evidence suggesting that species including shore pine, Sitka spruce, brown bear, black bear, ermine, Steller's jay, and chum, sockeye, and coho salmon remained on Haida Gwaii through this period, possibly indicating the presence of one or more refugia on the outer coast (Heusser 1989; Kondzela et al. 1994; Byun et al. 1997; Smith et al. 2001; Fleming and Cook 2002; Burg et al. 2005; Gapare et al. 2005; Reimchen and Byun 2005; Beacham et al. 2006; Godbout et al. 2008). Data from Cape Ball on Graham Island and Dogfish Bank in Hecate Strait suggest that the exposed land on the east coast of Haida Gwaii was covered in herb/shrub tundra consisting of small, stunted vegetation, including dwarf willow, crowberry (*Empetrum nigrum*), bunchberry (*Cornus*), grasses, sedges, heathers, and other herbaceous species by ca. 17,080 to 16,200 cal BP (Warner 1984; Barrie et al. 1993; Lacourse and Mathewes 2005; Lacourse et al. 2005). Similar herb tundra dominated by sedges, grasses, heathers, sagewort (*Artemisia*), Jacob's ladder (*Polemonium*), and other low-lying plants was established at West Side Pond on western Moresby Island in southern Haida Gwaii by ca. 16,890 cal BP (13,780 $^{14}$C BP; Lacourse et al. 2005) and on Langara Island off northwestern Haida Gwaii by ca. 16,250 cal BP (13,500 $^{14}$C BP; Heusser 1995).

### *ca. 16,000-11,500 cal BP*

Low relative sea level and shallow relief across Hecate Strait to the east of Haida Gwaii may have allowed for the temporary expansion of tundra environments as the ice retreated rapidly from 17,000-16,000 cal BP (Josenhans et al. 1995; Barrie and Conway 1999). Black and brown bears as well as ungulates are attested from K1 Cave on the west coast of Graham Island after ca. 14,000 cal BP, though many other animal species appear in the record during or after the Younger Dryas (Ramsey et al. 2004; Wigen 2005; Fedje et al. 2011). Herbaceous plants likely existed in exposed refugia until the CIS and Haida Gwaii ice began to retreat in the region, or they colonized the area quickly after the fact (Heusser 1989; Lacourse et al. 2005; Mathewes and

Clague 2017; Shaw et al. 2019). Herb and shrub vegetation along with open conifer parkland began to expand out from refugia and along the coast over the next few thousand years. The basal records from Kilgii Gwaay (ca. 14,540-13,850 cal BP) on Ellen Island and Hippa Island (ca. 14,000-13,230 cal BP) off the western side of Graham Island demonstrate continuity with these early herb/shrub tundra environments, though with the addition of some conifers like shore pine and Sitka spruce at the base of the Hippa Island record (Lacourse et al. 2012; Mathewes et al. 2019). On western Moresby Island, the record at West Side Pond demonstrates a transition from this early herb-shrub tundra to a pine woodland-type environment by ca. 14,600 cal BP (12,500 $^{14}$C BP) with numerous non-arboreal taxa still present (Lacourse et al. 2005). In eastern and southern Haida Gwaii (Logan Inlet and SG̱ang Gwaay [Anthony Island]), conifers such as pine began to appear first in open woodlands along with mountain hemlock and alder around 14,100 cal BP (12,240 $^{14}$C BP). This was followed by the expansion of spruce and western hemlock to form dense mixed-conifer forests after ca. 13,260-12,800 cal BP (11,000 $^{14}$C BP) (Hebda et al. 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Lacourse et al. 2012; Mathewes et al. 2019).

### *Post-11,500 cal BP*

After the Younger Dryas, many of the animals that appeared to survive through the glacial period in refugia were still present, as well as returning species including black bear and deer (Cervidae). By ca. 11,500-10,700 cal BP (10,000-9,450 $^{14}$C BP) larger species including harbour seal, Steller sea lion, river otter, sea otter (*Enhydra lutris*), and mule deer (*Odocoileus hemionus*) were present in the record on Haida Gwaii (at Gaadu Din and Kilgii Gwaay) (Wigen 2005). Additionally, these two sites indicate a variety of fish in the region by this time including salmon (*Oncorhynchus* spp.), Dolly Varden (*Salvelinus malma*), lingcod (*Ophiodon elongatus*), Pacific spiny dogfish (*Squalus suckleyi*), rockfish (*Sebastes* sp.), greenling (Hexagrammidae), and many others, along with birds such as short-tailed albatross (*Phoebastria albatrus*), snow goose

(*Chen caerulescens*), double-crested cormorant (*Phalacrocorax auritus*), common murre, rhinoceros auklet (*Cerorhinca monocerata*), and raven (*Corvus corax*) (Wigen 2005; Fedje et al. 2011).

As the climate ameliorated following Younger Dryas cooling, mixed-conifer and alder forests remained mostly consistent on the north coast, with the notable introduction of western redcedar by 8,700 cal BP and its subsequent dominance of the warmer and wetter coastal rainforest after ca. 6,800 cal BP (6,000 $^{14}$C BP) (Hebda et al. 2005; Lacourse and Mathewes 2005; Lacourse et al. 2012; Mathewes et al. 2019). Cedar has remained a significant part of the landscape for the remainder of the Holocene and forms a vital part of the material culture of Indigenous peoples up and down the coast (Hebda and Mathewes 1984; Hebda et al. 2005).

### *Central Coast*

The palaeoecology of the glacial period on the central coast of British Columbia is little studied when compared with the record on Haida Gwaii. As such, the palynological and palaeontological record does not begin until after the LLGM in the region.

#### *ca. 16,000-11,500 cal BP*

The oldest palynological record on the central coast of BC is that from Pond D on Calvert Island (Eamer 2015). The lowest pollen zone recorded at Pond D (beginning ca. 14,500 cal BP) indicates an assemblage dominated by shore pine, as seen elsewhere on the coast. However, this earliest zone also contains herb and shrub species including willow, grasses, sedges, and heathers (Eamer 2015). Galloway et al. (2008) present similar pollen assemblages extending to ca. 13,800 cal BP (12,000 to 11,900 $^{14}$C BP) from Seymour Inlet on the central coast. These basal deposits indicate a vegetation regime consisting of pine, alder, juniper (Cupressaceae-type pollen), and haircap mosses (*Polytrichum*-type). After ca. 13,600 cal BP (11,763 ± 87 $^{14}$C BP, SUERC-3090), the vegetation in Seymour Inlet consists of a more diverse assemblage of arboreal pollen, with decreases in pine pollen coupled with increases in spruce, fir, alder, and western hemlock (Galloway et al. 2008). Nearing the end of the Younger Dryas by ca. 11,800 cal BP, alder increases in abundance at these two sites (Galloway et al. 2008; Eamer 2015). Farther out from the mainland and closer to the north end of Vancouver Island, the Younger Dryas record at Cook Bank demonstrates less diverse arboreal pollen, with the record dominated by shore pine (Luternauer et al. 1989b; Lacourse et al. 2003).

### *Post-11,500 cal BP*

Over the next few thousand years of the early Holocene, evidence for many species of plants and animals becomes apparent across the central coast, leaving evidence in records on Triquet Island and King Island as well as on the inner coast at Namu (Cannon 1996; Cannon and Yang 2006; Fargo 2013; McLaren et al. 2015). Evidence of Douglas-fir appears on the central coast prior to ca. 9,740 cal BP (8,740 $^{14}$C BP, TO-12568) (Galloway et al. 2008). After ca. 7,700 cal BP (6,860 ± 50 $^{14}$C BP, BETA-206929), the pollen assemblage at Seymour Inlet indicates a massive increase in Cupressaceae-type pollen and a comparative decline in western hemlock, with low counts of spruce, fir, and non-arboreal pollen (Galloway et al. 2008). This likely represents a shift to the cool and wet-adapted assemblage that characterizes modern coastal temperate rainforests in the region (Hebda and Whitlock 1997).

Many terrestrial and marine mammals including North American beaver (*Castor canadensis*), porcupine (*Erethizon dorsatum*), river otter, mink (*Neovison vison*), marten (*Martes americana*), deer, harbour seal, northern fur seal (*Callorhinus ursinus*), Steller sea lion, Pacific white-sided dolphin (*Lagenorhynchus obliquidens*), and porpoise (Phocoenidae) appear in deposits from Namu from at least 6,800 cal BP (6,000 $^{14}$C BP), as well as birds including ducks and geese (Anatidae), loons (Gaviidae), gulls (Laridae), and bald eagle (*Haliaeetus* *leucocephalus*) (Cannon 1996; McLaren et al. 2015). Fish are also attested throughout this period from Namu and elsewhere on the central coast, including salmon (pink, chum, sockeye, and coho), eulachon (*Thaleichthys pacificus*), Pacific herring (*Clupea pallasi*), rockfish, cod (Gadidae), and flatfish (Pleuronectidae) (Cannon 1996; Cannon and Yang 2006; Fargo 2013; McLaren et al. 2015). By the late Holocene, most modern species appear in assemblages from Namu at ca. 1,950 cal BP (ca. 2,000 $^{14}$C BP) (Cannon 1996; Cannon and Yang 2006; McLaren et al. 2015).

### *Northern Vancouver Island*

#### *ca. 26,000-16,000 cal BP*

As on the north coast, small animals including Townsend's vole (*Microtus townsendii*), long-tailed vole, marmot (*Marmota* sp.), and American marten were present at Port Eliza Cave on the outer coast of western Vancouver Island ca. 21,800-19,300 cal BP (ca. 18,000-16,000 $^{14}$C BP), prior to the LLGM in the region (Al-Suwaidi et al. 2006). Evidence from Port Eliza Cave also indicates the presence of mountain goat (*Oreamnos americanus*) in the region at this time, suggesting nearby refugia large enough to support ungulate populations (Al-Suwaidi et al. 2006). Bird species including ducks, loons, cormorants (*Phalacrocorax* sp.), and savannah sparrow (*Passerculus sandwichensis*) are also part of the faunal record from this period on the western side of northern Vancouver Island, as well as fish species including salmon, cod, greenling, flatfish, and various kinds of shellfish (*Mytilus* sp., *Balanus* sp.) (Al-Suwaidi et al. 2006). The mammal species from this assemblage suggest an open, non-forested landscape. This is corroborated by the pollen record for the same period (ca. 21,800-19,300 cal BP), which is dominated by non-arboreal pollen including grasses, sedges, spikemoss (*Selaginella*), asters (Asteraceae), cow-parsnip (*Heracleum maximum*), and harebell (*Campanula*). The lesser arboreal component consists of pine and western hemlock (Al-Suwaidi et al. 2006).

### *ca. 16,000-11,500 cal BP*

As the ice sheets began to retreat from northern Vancouver Island, conifers, ferns, and open-canopy species began to spread across the landscape (Hebda 1983; Lacourse et al. 2003; Lacourse 2005). Pine, alder, and western/mountain hemlock arrived first and formed conifer parklands (Hebda 1983; Lacourse et al. 2003; Lacourse 2005). This initial pine-dominated assemblage appears first at Bear Cove Bog (ca. 16,450 cal BP) and in the offshore marine record off Cook Bank (ca. 15,500 cal BP), and slightly later in the record from Misty Lake near Port McNeill, ca. 14,900 cal BP (Hebda 1983; Lacourse et al. 2003; Lacourse 2005). Species including amabilis fir, birch (*Betula*), and western yew (*Taxus brevifolia*), appear in northern Vancouver Island records between ca. 13,800-12,800 cal BP to create increasingly thick coniferous forests (Lacourse et al. 2003; Lacourse 2005). Evidence from caves on northern Vancouver Island, including both Port Eliza Cave on the west side of the island and Pellucidar Cave and others in the Nimpkish Valley, indicates the presence of mountain goats as well as black, brown, and short-faced bear (*Arctodus simus*) from 14,500 cal BP (Nagorsen and Keddie 2000; Al-Suwaidi et al. 2006; Steffen and McLaren 2008; Steffen and Fulton 2018). Cooling during the Younger Dryas (ca. 12,800-11,500 cal BP) may have led to the temporary retreat of some of these species (Fedje et al. 2011).

### *Post-11,500 cal BP*

Post-Younger Dryas records from Bear Cove Bog (Hebda 1983) and Misty Lake (Lacourse 2005) both demonstrate an initial increase in alder and spruce through the early Holocene. By ca. 8,600 cal BP at both sites, western hemlock begins to dominate the assemblage, with lesser amounts of spruce and alder present (Hebda 1983; Lacourse 2005). Cupressaceae-type pollen (cedar) establishes itself on northern Vancouver Island by the mid-Holocene, ca. 5,100 cal BP (Hebda 1983; Lacourse 2005).

### *Southern Vancouver Island*

The modern climate and vegetation of southern Vancouver Island is characterized by a declining precipitation gradient from west to east. Modern forests on the west side of the island consist of western hemlock, western redcedar, and Sitka spruce, while the east side of the island is dominated by drier forests of Douglas-fir in the rain shadow of the Olympic Mountains and the Vancouver Island Ranges (Brown and Hebda 2002). However, this gradient was not established until approximately the last 2,000 years, following increasing precipitation through the mid- and late Holocene. Prior to the stabilization of this gradient, late Pleistocene and early Holocene climatic and vegetation shifts mostly mirror the vegetation succession of other sites on the coast (Brown and Hebda 2002).

#### *Pre-28,000 cal BP*

At Skutz Falls and at Qualicum Beach on the eastern side of southern Vancouver Island, Miskelly (2012) documents a cold, dry tundra environment with grasses, sedges, willow, and asters during the Olympia Nonglacial Interval (beginning at each site ca. 35,845 and ca. 33,890 cal BP, respectively). Toward the end of this period (ca. 28,800 cal BP at Osborne Bay), some of these sites on eastern Vancouver Island became dry mixed-conifer parklands composed of spruce, fir (Miskelly 2012).

#### *ca. 28,000-16,000 cal BP*

As on the central coast of BC, there is relatively little evidence for the maximum Fraser Glaciation in the palynological record on southern Vancouver Island. Most records either pre-date the advance of Cordilleran ice (e.g. Miskelly 2012) or begin following the retreat of glaciers from the south end of the island (e.g. Brown and Hebda 2002; Lucas and Lacourse 2013; Leopold et al. 2016). The early portion of this period, from ca. 28,000-21,200 cal BP appears to have been cold and dry at several sites on the eastern side of southern Vancouver Island (Miskelly 2012). Cold, dry tundra or parkland dominate after ca. 26,000 cal BP at Cordova Bay, McKenzie Bight, and at Skutz Falls. These are mostly non-arboreal assemblages consisting of grasses, sedges, asters, and sagewort, with arboreal components mostly consisting of pine or spruce (Miskelly 2012). Such herb/shrub tundra assemblages are comparable with refugial and post-LLGM records from the outer coast further north in Haida Gwaii (Barrie et al. 1993; Heusser 1995; Lacourse and Mathewes 2005; Mathewes and Clague 2017). However, with the approach of the local maximum of the Fraser Glaciation, there are few data available for the region after ca. 20,000 cal BP (Clague et al. 1980).

### *ca. 16,000-11,500 cal BP*

Following the deglaciation of southern Vancouver Island, conifer parkland dominated by pine spread across the landscape, appearing before ca. 15,500 cal BP (12,990 ± 180 $^{14}$C BP, BETA-81084) at Pixie Lake near the San Juan River, by ca. 13,500 cal BP (11,700 ± 140 $^{14}$C BP, BETA-86809) at East Sooke Fen west of Victoria, and before ca. 12,800 cal BP (10,860 ± 130 $^{14}$C BP) at Whyac Lake at the southern end of Nitinat Lake (Brown and Hebda 2002). In addition to pine, herb, shrub, and grass species including willow, soapberry (*Shepherdia canadensis*) and sagewort occupied open areas. Conifer parkland may have spread to both Pixie Lake and Whyac Lake significantly earlier than these dates, as they were not retrieved from the basal section of the cores (Brown and Hebda 2002). A similar pine-dominated assemblage is present in the Salish Sea from ca. 13,700 cal BP at Killebrew Lake Fen on Orcas Island, WA (Leopold et al. 2016). This initial pine parkland assemblage was replaced by mixed-conifer forests including pine, spruce, fir, western hemlock, mountain hemlock, and alder on the west coast of Vancouver Island, notably earlier at west coast sites than in the Salish Sea. This mixed-conifer assemblage appears first at Pixie Lake ca. 13,600 cal BP (11,750 ± 100 $^{14}$C BP, BETA-81085) (Brown and Hebda 2002) but is not documented at Killebrew Lake Fen until ca. 12,000 cal BP (Leopold et al. 2016). The onset of cooling during the Younger Dryas may be represented by the rise of mountain hemlock following pine dominance and mixed-conifer forests during the Bølling-Allerød in many coastal sites (Mathewes 1993). Rise in mountain hemlock pollen is demonstrated during the Younger Dryas chronozone at both the west coast sites and at Killebrew Lake Fen (Brown and Hebda 2002; Leopold et al. 2016).

*<u>Post-11,500 cal BP</u>*

During the warm early Holocene—beginning ca. 9,300 cal BP (8,330 ± 100 $^{14}$C BP, BETA-85177) at Pixie Lake—western hemlock is dominant, mixed with spruce, fir, and alder with very little pine (Brown and Hebda 2002). The early Holocene assemblage from Roe Lake on Pender Island also demonstrates a similar shift away from pine, though it occurs slightly later (after ca. 8,000 cal BP, to be replaced by dominant Douglas-fir and alder (Lucas and Lacourse 2013). Finally, by ca. 6,250 cal BP (5,450 ± 70 $^{14}$C BP, BETA-85176), Cupressaceae-type pollen—likely western redcedar—begins to increase in the record at west coast sites like Pixie Lake. This lasts until the current late Holocene vegetation and climate regime on the west coast is established after ca. 2,900 cal BP (2,770 ± 80 $^{14}$C BP, BETA-86805), dominated by western redcedar and western hemlock with lesser components of spruce and alder (Brown and Hebda 2002).

## 2.4 Late Pleistocene/Early Holocene Archaeology of the Northwest Coast

This review describes late Pleistocene and early Holocene archaeological sites ranging from southeastern Alaska to northern Washington and Puget Sound. These sites demonstrate the spread and interconnection between diverse peoples throughout the earliest period of human occupation on the northwest coast of North America. Figure 14 contains a geographic orientation of these sites, and Table 1 summarizes the site names, ages, and key details. The flow of this summary map and table is generally chronological from oldest to most recent based on currently published radiocarbon dates. Each number on Figure 14 corresponds to the number noted beside the name of the site in Table 1.

Several late Pleistocene archaeological sites on the northwest coast of North America are older than 13,500 cal BP, including Triquet Island (Gauvreau and McLaren 2017), Ayer Pond (Kenady et al. 2011), Manis Mastodon (Waters et al. 2011), and possibly Kildidt Narrows (McLaren et al. 2015). Even more are older than 12,000 cal BP, including Meay Channel I (McLaren et al. 2018), K1 Cave, Gaadu Din 1 and 2 (Fedje et al. 2011), the Bear Creek site (Kopperl et al. 2015), the Stave Watershed sites (McLaren 2017), Werner Bay (Fedje and Josenhans 2000), and Shuká Káa (Dixon 2013). Lithic assemblages from many of these early sites include stemmed projectile points, which have been found at late Pleistocene sites elsewhere in western North America and are contemporaneous with Clovis technology or possibly older (Waters 2019; McLaren et al. 2019b). Additionally, many late Pleistocene and early Holocene sites on the Pacific northwest coast demonstrate adaptations to maritime environments based on faunal remains, tool technologies, or implied use of watercraft (for example Chapman 1982; McLaren 2008; Fedje et al. 2011; Cohen 2014; Gauvreau and McLaren 2017). More detailed summaries of each site presented here can be found in Appendix B and a more expanded consideration of early archaeological sites in the Americas can be found in the Discussion in Chapter 4.

***Figure 14.*** *Late Pleistocene and early Holocene archaeological sites on the Pacific northwest coast of North America. See Table 1 for site names and information associated with each number. Map data: Natural Earth, Viewfinder DEM, GeoBC.*

***Table 1.*** *Late Pleistocene and early Holocene archaeological sites on the Pacific northwest coast of North America as represented in Figure 14, ordered chronologically based on the earliest date from each site and with a brief summary of the oldest cultural component.*

| **Site name** | **#** | **Age of oldest component (cal BP)** | **Region** | **Key features of oldest component** | **References** |
|---|---|---|---|---|---|
| Triquet Island (EkTb-9) | 1 | 14,086-13,613 | Central Coast | Lithic cache below extensive habitation layers; island implies watercraft | Gauvreau and McLaren 2017 |
| Ayer Pond | 2 | 13,950-13,750 | Salish Sea | Butchered remains of *Bison antiquus* | Wilson et al. 2009; Kenady et al. 2011 |
| Manis Mastodon | 3 | 13,860-13,763 | Salish Sea | Remains of American mastodon with embedded bone projectile point | Gustafson et al. 1979; Waters et al. 2011 |
| Kildidt Narrows (ElTa-18) | 4 | 13,673-10,692 | Central Coast | Lithics including biface tip and retouched flake | Cannon 2000; McLaren et al. 2015 |
| Meay Channel I (EjTa-4) | 5 | 13,317-13,241 | Central Coast | Intertidal footprints of at least three individuals and stone tools | McLaren et al. 2018 |
| K1 Cave (FgUc-6) | 6 | 13,150-12,150 | Haida Gwaii | Cave with foliate spearpoint bases and fauna indicating bear hunting | Fedje et al. 2004; Ramsey et al. 2004; Fedje et al. 2011 |
| Puget Sound/Stave Lake Clovis | - | 13,300-12,700 | Salish Sea | Fluted Clovis points around Puget Sound and Stave Lake | Croes et al. 2008; McLaren 2017 |
| Bear Creek Site (45K1839) | 7 | 12,770-11,274 | Salish Sea | Stemmed points with extensive debitage on ancient riverbank | Kopperl et al. 2015 |
| Gaadu Din 1 | 8 | 12,600-11,500 | Haida Gwaii | Cave with flaked stone spearpoints and fauna indicating bear hunting | Fedje et al. 2011 |
| Gaadu Din 2 | 8 | 12,500-10,700 | Haida Gwaii | Cave used as hunting camp or shelter with bipoint tips, flakes, and hammerstones | Fedje et al. 2011 |
| Stave Watershed Sites | 9 | 12,404-12,051 | Fraser Valley | Biface preform and flakes; stemmed and foliate points nearby | McLaren 2017; McLaren et al. 2019b |
| Werner Bay | 10 | 12,110-11,700 | Haida Gwaii | Utilized flake from 53 m below modern sea level in Juan Perez Sound | Fedje and Josenhans 2000; Fedje et al. 2005a |

| Site name | # | Age of oldest component (cal BP) | Region | Key features of oldest component | References |
|---|---|---|---|---|---|
| Shuká Káa (On Your Knees Cave) | 11 | 12,000 | Southeastern Alaska | Cave with worked bone tool; later (ca. 10,350 cal BP) remains of young man | Dixon et al. 1997; Lee 2001; Kemp et al. 2007; Dixon 2013 |
| Namu (ElSx-1) | 12 | 11,600-11,550 | Central Coast | Foliate bifaces and cobble tools at base of long-inhabited village | R. Carlson 1996; McLaren et al. 2015; Porter 2015 |
| Far West Point (GcTr-6) | 13 | 11,200-11,050 | North Coast | Cobble core, spall tool, and flake in charcoal-rich black silt | McLaren 2008; Martindale et al. 2009 |
| Kilgii Gwaay | 14 | 10,700 | Haida Gwaii | Coastal site with large faunal assemblage and waterlogged twine, split spruce-root, wedges, and stakes | Fedje et al. 2005a; Fedje et al. 2011; Cohen 2014; Cohen 2019 |
| Ground Hog Bay 2 | 15 | 10,683-10,303 | Southeastern Alaska | Lithics including bifaces, scrapers, choppers, gravers, and flakes | Ackerman 1996; Chatters et al. 2012 |
| Richardson Island | 16 | 10,600-10,400 | Haida Gwaii | Features including post-moulds and hearths, with extensive faunal assemblage and lithics | McLaren and Smith 2008; Storey 2008; Waber 2011 |
| Hidden Falls | 17 | 10,553-9,785 | Southeastern Alaska | Lithics including bifaces, scrapers, choppers, gravers, and flakes | Ackerman 1996; Chatters et al. 2012 |
| Arrow Creek 1 and 2 | 18 | 10,500-10,200 | Haida Gwaii | Stone tools in intertidal and on raised alluvial fan; dated by barnacle scars | Fedje and Christensen 1999 |
| Milliken (DjRi-3) | 19 | 10,500-9,650 | Fraser Canyon | East side of Fraser River; post-holes and hearths with foliate points and tools implying salmon fishery | Mitchell and Pokotylo 1996 |
| Rice Creek | 20 | 10,250-10,100 | Southeastern Alaska | Flakes recovered from tree throws and surface exposures along stream banks | Ackerman 1996 |
| Pruth Bay (EjTa-15) | 21 | 10,151-9,924 | Central Coast | Coastal site with hearth feature; above is charcoal with basalt, chert, andesite, and obsidian tools and flakes | McLaren et al. 2015 |
| Lyell Bay South and East | 22 | 10,000-9,450 | Haida Gwaii | Extensive lithic assemblages from nearby sites including microblades, cobble choppers, and biface fragment | Fedje and Christensen 1999 |

| Site name | # | Age of oldest component (cal BP) | Region | Key features of oldest component | References |
|---|---|---|---|---|---|
| Millard Creek (DkSf-2) | 23 | 9,650-8,650 | Vancouver Island | Coastal site with stemmed point, choppers, and worked bone beneath extensive midden deposits | Capes 1970; Grant 2004; BC Archaeology Branch 2019 |
| Glenrose Cannery (DgRr-6) | 24 | 9,550-8,450 | Fraser Valley | Palaeo-peninsula on Fraser River; extensive lithic assemblage with foliate bifaces, choppers, and hammerstones | Matson 1996 |
| Chuck Lake | 20 | 9,500-9,000 | Southeastern Alaska | Microblades and microblade cores, scrapers, flakes, hammerstones, and bone point; extensive marine fauna | Ackerman 1996 |
| Two-Flake Terrace (GbTo-185) | 25 | 9,304-9,028 | North Coast | Raised terrace above modern sea level; hearth features and debitage | Letham et al. 2016 |
| Bear Cove (EeSu-8) | 26 | 8,900 | Vancouver Island | Coastal site with chipped stone artifacts in black and brown silts and gravels | C. Carlson 1979; C. Carlson 2003 |
| Lawn Point (FiTx-3) | 27 | 8,450-7,950 | Haida Gwaii | Raised beach with extensive assemblage of microblades, microblade cores, pebble tools, and flakes | Fladmark 1986; Fladmark 1989 |
| Thorne River | 28 | 8,400-8,200 | Southeastern Alaska | Obsidian microblades, microblade cores, and bifaces | Ackerman 1996 |
| Argonaut Hill Forks (GaTw-9) | 29 | 8,000-7,900 | Haida Gwaii | Raised terraces with lithic scatters including choppers, spall tools, and flakes | Sanders 2009 |

# 3. Materials and Methods

## 3.1 Program of Research

### 3.1.1 Established Methods

Pollen analysis has formed the core of palaeoenvironmental reconstruction for many areas of British Columbia (Hebda 1983; Hebda 1995; Brown and Hebda 2002; Lacourse et al. 2003; Hebda et al. 2005; Lacourse 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Al-Suwaidi et al. 2006; Galloway et al. 2008; Brown et al. 2017; Mathewes and Clague 2017; and many others). When coupled with other sources, including palaeontological records and faunal remains from archaeological sites, these data have provided regionally specific reconstructions of the environment in coastal British Columbia stretching from the present day to the late Pleistocene (Heaton et al. 1996; Nagorsen and Keddie 2000; Heaton and Grady 2003; Wigen 2005; Steffen and McLaren 2008; Fargo 2013; Steffen et al. 2018; and others).

While these methods can provide good indicators of past environmental conditions, each also relies on specific assumptions and biases that can colour the interpretation of the evidence. Some studies, including those conducted by Hebda et al. (2005), Al-Suwaidi et al. (2006), Mathewes and Clague (2017) and others attempt to circumvent this by applying multi-proxy approaches (often some combination of analyzing pollen, plant macrofossils, sedimentology, and palaeontology) to cover the methodological gaps introduced by each approach alone. For example, pollen analysis can yield high-quality reconstructions on a regional scale and with broader taxonomic identifications (family, genus), while macrofossil analysis can provide details about local vegetation conditions with higher-resolution identification (genus, species), but lacks as much regional applicability (Hebda 1981; Moore et al. 1995; Bennett and Willis 2001).

By introducing new methods into these established research workflows, we can further strengthen our interpretation of past environments. Analysis of ancient environmental DNA (eDNA) from sediments has been used successfully to detect plant and animal species extending back into deep time (Hofreiter et al. 2003; Willerslev et al. 2003; Matisoo-Smith et al. 2008; Haile et al. 2009; Anderson-Carpenter et al. 2011; Epp et al. 2012; Jørgensen et al. 2012a;2012b; Parducci et al. 2012; Stoof-Leichsenring et al. 2012; Pedersen et al. 2013; Pedersen et al. 2015; Pedersen et al. 2016; Bremond et al. 2017). Several studies, including those by Jørgensen et al. (2012b) and Pedersen et al. (2013) have also demonstrated that the addition of ancient eDNA analysis to multi-proxy pollen and plant macrofossil-based research is often complementary, providing expanded taxonomic identifications compared with conventional methods. However, ancient eDNA analysis has yet to be applied to research questions in coastal British Columbia. This study applies several proxies—including pollen, diatom, macrofossil, and ancient eDNA analyses—to reconstructing terminal Pleistocene environments on northern Vancouver Island.

### 3.1.2 Program of Research

This program of study progressed through several stages, including:

1. Review literature on geological history, palaeoecology, and archaeology of coastal British Columbia with a focus on northern Vancouver Island, especially on late Pleistocene records.
2. Review literature on lacustrine coring methods, eDNA extraction and analysis, and pollen analysis.
3. Determine appropriate target lakes on northern Vancouver Island from which to extract cores.
4. Extract cores from Topknot Lake and Little Woss Lake on northern Vancouver Island (see Figure 2), using both Livingstone corer and a modified version of the Reasoner method.
5. Subsample cores for eDNA samples using clean Hakai Institute lab space in Victoria.
6. Perform eDNA extraction and sequencing on selected samples from both study sites at the Centre for GeoGenetics in Copenhagen, Denmark.

7. Subsample cores for pollen, diatoms, and radiocarbon samples using lab space at the Pacific Forestry Centre in Victoria.
8. Submit radiocarbon samples for dating to A.E. Lalonde AMS Laboratory, University of Ottawa.
9. Submit diatom samples for analysis by University of Victoria researcher (D. Fedje).
10. Submit pollen samples for preparation by Pacific Forestry Centre lab tech (N. Conder).
11. Analyze pollen samples, focusing on samples from lower portions of cores as well as transitional layers and segments dated by $^{14}C$.
12. Select specific samples (including the lower samples from Topknot Lake) for additional processing, including treatment with hydrofluoric acid and screening out fine sediments.
13. Submit additional radiocarbon samples for dating to A.E. Lalonde AMS Laboratory, University of Ottawa, in order to fill chronological gaps in previous date series.
14. Check over pollen counts and resolve unknown species identifications.
15. Interpret results of pollen analysis, diatom analysis, and eDNA sequencing.
16. Construct age-depth model for each site.
17. Build pollen diagrams for each site incorporating $^{14}C$ dates, pollen counts, eDNA results, and core lithology.
18. Collate literature review, final methods, results, discussion, and future directions into thesis.

## 3.2 Methods

Based on the Program of Research, several broad categories are expanded upon below, including a) Fieldwork, including core extraction and transport; b) Subsampling, including sampling for eDNA, pollen, diatoms, radiocarbon, and macrofossils; c) eDNA lab work, including sediment extraction, sample purification, library building, and sequencing; d) Pollen lab work, including sample preparation, acetolysis, hydrofluoric acid treatment, and identification and counting procedures; e) Other lab work, including diatom and macrofossil identification and quantification; and f) Data analysis and presentation, including the construction of age-depth models, pollen diagrams, and associated material derived from the primary data.

### 3.2.1 Fieldwork

Following initial discussions with my committee members, we agreed that multiple study sites would facilitate more robust palaeoenvironmental reconstructions of northern Vancouver Island, and that such research would fall within the limits of a master's-level project. However, to maintain the scope of the work, my committee members recommended that I focus on analyzing the late Pleistocene components of the cores collected as part of this project. We then selected targets for palaeolimnological coring with several major criteria in mind. We determined that each lake should:

- be accessible by vehicle such that the coring equipment can be carried to the lake without relying on prohibitively expensive transportation methods;
- be at least marginally above the theoretical late Pleistocene maximum relative sea level transgression for the study area (~18-20 m above modern on the outer west coast of northern Vancouver Island);
- be no more than approximately 1-1.5 km in length to limit basin size and water depth; and

- have a relatively small catchment area to limit basin sedimentation rate and maintain reasonable core length for recovery.

Given these criteria, we selected Little Woss Lake (153 m asl) and Topknot Lake (13 m asl) on north-central and north-western Vancouver Island, respectively, as the two study sites for this research (see Figure 2). Based on the location and elevation of each study site, we believed that they could produce extensive late Pleistocene records and possibly demonstrate the presence of glacial refugia on Vancouver Island.

Little Woss Lake was selected because it is located within a rainshadow in the mountains on the dry leeward side of the Vancouver Island Ranges, which may have reduced precipitation and limited glacial extent in the nearby valleys. It is also located near a large salmon-bearing stream—the Woss River—which makes the area a particularly good target for archaeological research. Little Woss Lake is located within approximately 1-2 km of recently discovered early Holocene (and possibly late Pleistocene) cultural deposits on the shores of Woss Lake and along the Woss River, allowing direct comparisons to be drawn between the palaeoenvironmental data of Little Woss Lake and nearby archaeological data.

Topknot Lake was selected because of its location on the outer west coast of northern Vancouver Island. The lake is nearly as far as possible from the sources of ice during the Fraser Glaciation (Cordilleran ice or local ice from the Vancouver Island Ranges) and is 13 m above modern sea level—this is at or above the maximum theoretical late Pleistocene/early Holocene sea level transgression in the region (Howes 1983; Howes 1997). Additionally, the low topography of the Nahwitti Plateau on which Topknot Lake is located means that there is less potential for local glacial growth and for mass wasting and other geological events that could complicate the depositional record.

In May 2018, we cored the lakes as part of fieldwork for the Northern Vancouver Island Archaeology and Palaeoecology Project (NVIAPP), co-funded by the University of Victoria and the Hakai Institute. Both basins were cored using a combination of a Livingstone corer (Livingstone 1955; Vallentyne 1955; Wright 1967) and a modified version of the Reasoner coring method (Reasoner 1986;1993). At Little Woss Lake, we retrieved the cores from a plywood platform resting on the surface of shallow water on the lake's eastern shoulder. At Topknot Lake, we retrieved the cores from the lake's southeastern quadrant using a plywood board suspended between two floating canoes as a platform.

Livingstone core sections were collected in 90 cm to 1 m drives using an aluminum core barrel and were extruded in the field into 5-cm diameter ABS core boxes that had been cut in half lengthwise. Plastic wrap was used to cover the core sections and pack the ends, and the core box was sealed along the sides and ends with tape for transport.

Modified Reasoner-method cores were collected continuously using 3.5-m sections of 7.5-cm diameter solid-core ABS tubing. Prior to driving, the flat bottom end of the tube was whittled down to sharpen it to decrease resistance and to help cut through organics while being pushed through the sediment. Metal core-catching segments were cut from aluminum and attached to the inside of the bottom of the core tube using duct tape. The tube was driven into the sediment by dropping a heavy weight repeatedly onto its top end. As the upper portion of each 3.5-m section of tubing approached the surface of the lake, a coupling was attached to its top using ABS glue, after which an additional 3.5-m section of tubing was attached above the coupling, again using ABS glue, before continuing the drive. The continuous cores were then retrieved using a hand winch mounted to an aluminum step ladder to draw them out of the sediment. The continuous Reasoner-method cores were split into manageable lengths in the field and were capped for transport.

### *Little Woss Lake*

Approximately 10 m of total sediment was recovered from Little Woss Lake, most of it collected in parallel using a Livingstone corer (0-475 cm dbs [depth below sediment-water interface]) and the modified Reasoner coring method discussed previously (0-532 cm dbs). The Livingstone core sections were extracted in 90 cm to 1 m drives, and the continuous core resulting from the modified Reasoner method was split into two equal sections in the field and capped. Prior to being transported to Victoria, the cores from Little Woss Lake were refrigerated at the Western Forest Products, Inc. Englewood Forest Operation office in Woss, British Columbia. All core sections were then transported to a walk-in sample cooler at the Pacific Forestry Centre for storage at 4°C until they were removed in late June 2018 for subsampling at Hakai Institute lab facilities in Victoria.

Researchers affiliated with the Pacific Forestry Centre, including K. Brown and R. Hebda, also collected a parallel core from Little Woss Lake in July 2018. I did not subsample this parallel core for pollen or eDNA, but I did collect samples for radiocarbon dating from the lower Livingstone sections to compare the basal dates with those from the initial core collected as part of NVIAPP.

### *Topknot Lake*

Approximately 10 m of sediment was recovered from Topknot Lake, including approximately 6.5 m of sediment extracted in 90 cm to 1 m drives using a Livingstone corer (475-1125 cm depth below the surface of the coring platform [dbp]) and 3.5 m of continuous sediment extracted using the modified Reasoner coring method described earlier (0-360 cm dbp). Just over

1 m of the sequence at Topknot Lake (360-475 cm dbp) was not collected by either method due to limitations in the field. The continuous core resulting from the modified Reasoner method was split into two equal sections, capped, and stored in a large 10-cm diameter solid-core ABS tube with ice. All seven Livingstone core box sections were stored in a 15-cm diameter PVC tube with ice to preserve ancient eDNA. All core sections were then transported to the archaeology lab at the Hakai Institute on Quadra Island for storage at 4°C until they were removed in late June 2018 for subsampling at Hakai lab facilities and the Pacific Forestry Centre in Victoria, British Columbia.

### 3.2.2 Subsampling

Between subsampling sessions, all core segments were stored as close to 4°C as possible either in a refrigerator/freezer provided by the Hakai Institute or in the walk-in cooler at the Pacific Forestry Centre.

#### *eDNA Subsampling*

The first round of subsampling on the cores from both study sites was conducted in late June 2018 with the purpose of collecting samples for ancient eDNA analysis at the Centre for GeoGenetics in Copenhagen, Denmark. All core sections were transported from storage to the Hakai Institute modern eDNA lab in Victoria for subsampling. All lab technicians, including the author, A. Gauvreau, L. Dixon, and D. McLaren followed protocol to limit sample contamination, including wearing full-body coveralls, shoe covers, a facemask, gloves, and safety goggles when in the lab space and working with the exposed core sections. Surfaces in the Hakai Institute lab space were regularly sterilized during the subsampling procedure using a 10% bleach solution.

Sampling focused on the lower sections of the cores from each study site, with intervals between sample locations ranging from 2 cm to 15 cm at my discretion depending on the presence of stratigraphic transitions. All samples were taken in duplicate from each depth. Samples were taken from the entirety of the continuous Reasoner method core from Little Woss Lake (n=22 samples) and from all Livingstone core sections from Topknot Lake (n=52 samples). At each sample location, a sterile scalpel blade was used to remove the top 0.5 cm of sediment from the outer edge of the core section, exposing a 2 cm-wide window for sampling. A sterile syringe tube was then used to extract 2-4 $cm^3$ of sediment from the sample location, transferring it to a sterile 15 ml centrifuge tube which was labeled with the core identifier, sample number, depth, date, and the initials of the technicians present. A duplicate sample was then extracted and transferred to a second sterile 15 ml centrifuge tube and labeled appropriately. Gloves, scalpel blades, and syringes were discarded between sampling from each location on the core. The resulting samples and duplicates were stored at 4°C or lower until they were transported to Copenhagen, Denmark in early July 2018. Following subsampling, the cores were sealed and transported to the Pacific Forestry Centre for storage in the walk-in cooler at 4°C.

### *Other Subsampling (Pollen, Diatoms, Radiocarbon, Macrofossils)*

Subsequent subsampling on cores from both study sites was conducted in several rounds beginning in late July 2018 after I returned from conducting initial eDNA lab work at the Centre for GeoGenetics in Copenhagen, Denmark. I collected pollen, diatom, and radiocarbon samples in laboratory facilities at the Pacific Forestry Centre. Subsampling for these proxies included the lower section of the modified Reasoner method core from Little Woss Lake as well as all sections of the Livingstone cores from Topknot Lake.

Before sampling commenced, the outer surface of each core segment was carefully scraped with a small metal spatula to expose the stratigraphy, which was then described visually. Test smears of sediment were collected for preliminary assessment throughout the cores and were mounted on glass slides with a cover slip. The test slides were examined under a microscope at up to 1000x to identify the presence of pollen grains, diatoms, and other organic debris, as well as to note the composition of the fine fraction of the sediment. These tests informed the more extensive pollen and diatom subsampling to follow.

### *Pollen and Diatom Subsampling*

Pollen samples were collected using clean syringes to extract between 2-5 $cm^3$ of material from each sample location and were transferred to sterile 50 ml centrifuge tubes and labeled with the core identifier, sample number, depth, date, and initials of the technicians present. Subsamples of approximately 0.75 $cm^3$ of sediment were taken from several of these samples for diatom analysis by D. Fedje at the University of Victoria, with the remaining material retained for pollen analysis.

Due to the relatively uniform stratigraphy of the Little Woss Lake lower core section, the unused eDNA samples taken at 15 cm-resolution were deemed sufficient for use in pollen and diatom analyses. Four additional samples were taken from this core section, one of which was collected from the top of the core section (172 cm LD) and three of which were taken at 2-cm intervals over the transition from fibrous gyttja to finer, silty gyttja and clays at the base of the core (310-314 cm LD).

The complex stratigraphy of the much longer core recovered from Topknot Lake required a more robust sampling procedure. Above 845 cm dbp in the sequence at Topknot Lake, pollen subsampling proceeded at a resolution of 5 cm throughout all core sections, with the samples extracted either using the method described above or where eDNA samples had been collected during the first round of subsampling but had not been used in the previous analysis at the Centre for GeoGenetics. Due to significantly reduced organic matter in the lower Livingstone core sections from Topknot Lake (~845-1120 cm dbp), pollen samples were collected at lower resolution across these sections focusing on stratigraphic transitions. In total, I collected 34 pollen samples from the Livingstone core sections from Topknot Lake in addition to the unused eDNA samples.

A subset of the pollen samples from both study sites (Little Woss Lake n=14; Topknot Lake n=18) were prepared at the Pacific Forestry Centre by lab technician N. Conder using standard methods including treatment with potassium hydroxide (KOH), hydrochloric acid (HCl), and acetolysis. Some samples were later treated with hydrofluoric acid (HF) to remove excess silica (Faegri and Iversen 1989; Moore et al. 1995).

### *Radiocarbon and Macrofossil Subsampling*

Initial radiocarbon samples were collected opportunistically based on the presence of macrofossils such as needles, leaves, seeds, and stems in the core sections from both study sites. These macrofossils include the preserved remains of leafy plants including sedge as well as aquatic plants (e.g. yellow pond-lily) and larger twigs, branches, and needles (e.g. shore pine). Samples were removed from the cores using forceps and placed in small glass vials which were then labeled with the study site, sample number, depth, and date.

In the lower section of the core from Little Woss Lake, I took radiocarbon samples on either side of the transition from fibrous gyttja to finer, silty gyttja and clays at the base of the core (307 cm LD and 315 cm LD), as well as a few more opportunistically further up the core (298.5 cm LD, 240 cm LD, 201 cm LD, and 171 cm LD) based on the presence of obvious macrofossils (n=6 total from the core). The basal dates were later compared with two radiocarbon samples I collected from a parallel core that was retrieved by researchers from the Pacific Forestry Centre within the same basin.

From the Topknot Lake Livingstone core sections, a total of 12 radiocarbon samples were collected and successfully dated (from 509-920 cm dbp). Organic matter was much rarer in the lower Livingstone core segments from Topknot Lake, so I collected radiocarbon samples from these sections later as bulk sediment samples and screened them (minimum 75 μm mesh) in the archaeology lab at the University of Victoria for identifiable fragmentary plant remains.

### 3.2.3 Lab Work

#### *eDNA Lab Work*

The first portion of the lab work for this project, conducted in early July 2018, involved the extraction and purification of ancient environmental DNA (eDNA) from the sediment cores collected from both study sites at Little Woss Lake and Topknot Lake. The eDNA samples (n=16 in duplicate, n=32 total) were transported in 15 ml centrifuge tubes by air from Victoria to Copenhagen, Denmark. During air travel, the samples were maintained as close to 4°C as possible using coolers and ice packs and were immediately stored at 4°C upon arrival in Copenhagen.

Along with A. Gauvreau, a PhD candidate in the Department of Anthropology at the University of Victoria, I prepared a detailed lab work protocol for eDNA extraction based on procedures followed by researchers at the Centre for GeoGenetics at the University of Copenhagen. A simplified version of the protocol is described below.

#### *eDNA Extraction and Sequencing*

All work in the ancient DNA lab at the Centre for GeoGenetics was carried out in a positive pressure laboratory environment with researchers wearing full-body coveralls, shoe covers, lab shoes, a facemask, two pairs of gloves, and sleeve extenders. All surfaces were sterilized using a 5% bleach solution, and equipment was sterilized using a 5% bleach solution and ethanol. Sample tubes were opened one at a time and only within a laminar flow hood. In total, 14 samples from Topknot Lake (n=9) and Little Woss Lake (n=5) underwent ancient eDNA extraction in the first round.

The method used for DNA extraction on the samples from Topknot Lake and Little Woss Lake was developed by Wales et al. (2014) and was modified by M.W. Pedersen at the Centre for GeoGenetics. The 15 sediment samples were weighed, and an empty 15 ml control tube was added to the batch to account for contamination introduced during the extraction. Every subsequent step and reagent addition performed on the samples was also performed on the control tube. The extraction was a multi-day process involving the separation of proteins and other larger particles from the liquid containing the free-floating DNA. This was facilitated in several stages through the addition of enzymes and a phenol-chloroform mixture. Several filtering stages, centrifugations, and supernatant transfers followed until the DNA was distilled and concentrated into a transparent liquid with no extraneous particles remaining.

The extraction was followed by a test using a fluorometer on each of the samples for the presence of sufficient DNA for amplification. The samples then underwent DNA ‘library building,’ during which the fragmented ends of the DNA were repaired by a targeted enzyme and had adaptors added to them to allow the identification of specific DNA strand lengths during the sequencing process. During this stage, an additional control sample was added to the batch of samples to account for contamination occurring during the library building process.

Once the index adaptors were attached to the DNA fragments, the DNA in each sample was amplified using polymerase chain reaction (PCR). Following PCR, the samples were once again purified using an ethanol wash before being quantified through gel electrophoresis by an Agilent 2100 Bioanalyzer. This process produced graphs of DNA base-pair lengths in each sample, providing an indicator as to whether the samples would yield identifiable DNA sequence fragments (100-1000 base-pairs) for downstream sequencing and bioinformatics. Five samples, including one from Topknot Lake, two from Little Woss Lake, and the two control samples demonstrated appropriate Bioanalyzer results to push for sequencing. The samples were then pooled together and submitted for sequencing on one high-throughput sequencing lane on the Centre for GeoGenetics' Illumina HiSeq machines.

### *eDNA Bioinformatics*

Following sequencing, M.W. Pedersen performed bioinformatic processing of the samples. Despite initial emergent incompatibilities with software utilities and the sequencing data, the samples were successfully processed using a modified version of 'Holi,' the bioinformatic workflow described by Pedersen et al. (2016). Metagenomic reads were trimmed of their adaptor sequences, with resulting segments of >30 bp retained. Reads with poly-A/T tails ≥4 were removed from the samples. Duplicate reads resulting from PCR were removed from each sample, and low-complexity sequences were removed using String Graph Assembler (SGA) in order to decrease computational time.

Following these quality control procedures, sequences were aligned against genomic reference databases including NCBI Nucleotide, RefSeq, and the Genome Taxonomy Database (GTDB) using Bowtie2. Matches aligned to these databases were passed through a lowest common ancestor (LCA) algorithm to determine the most parsimonious identification for each read. The results presented are those with an edit distance of 0 nucleobases—sequences that exactly match the reference genome for each taxon.

### *Pollen Lab Work*

Having received the first set of radiocarbon dates from both study sites, I selected a subset of the pollen samples discussed above for preparation and counting based on a) greatest depth, and presumably age; b) sufficient coverage of overall age and depth span; and c) transitions in core sedimentology. In total, I selected 14 samples from the Little Woss Lake core and 18 samples from the Topknot Lake core. These samples underwent initial processing in the analytical chemistry lab at the Pacific Forestry Centre in October 2018 by lab technician N. Conder with my observation and notation. Each sample was processed following standard methods (see Faegri and Iversen 1989; Moore et al. 1995) in three stages, including treatment with potassium hydroxide (KOH) followed by hydrochloric acid (HCl), then acetolysis. A summary of each stage of the protocol is described below. Except where noted, each centrifugation step was performed at 3,500 rpm for 5 minutes.

#### *Pollen Sample Preparation*

For initial treatment with potassium hydroxide, 1 $cm^3$ of each pollen sample was added to a 75 ml glass tube in a block digester at 120°C along with 30 ml of 5% KOH solution. Each KOH-sample solution was then homogenized with a VWR Digital Vortex Mixer and allowed to boil for 20 minutes in a block digester to disaggregate humic material. Following digestion, 3-5 ml of 5% potassium carbonate was added to the solution and vortexed to neutralize the KOH. We then sieved and rinsed the solution through 150 μm mesh and transferred the fine fraction to a 50 ml centrifuge tube with an exotic marker tablet (20848 ± 1,546 *Lycopodium* spores/tablet; Dept. Quaternary Geology, University of Lund, batch no. 1031). After vortexing the sample with the exotic marker spores, each tube was centrifuged, decanted, washed with 10 ml of distilled water ($dH_2O$), centrifuged, and decanted again. Each sample was stored in 10 ml $dH_2O$.

Following KOH treatment, we treated each sample with 5% hydrochloric acid (HCl) in order to remove carbonates. Each sample was centrifuged and the storage water decanted. Hydrochloric acid (10 ml of 10% solution) was then added to each sample and vortexed before being centrifuged and decanted. We did not observe significant effervescence in any of the samples during the reaction, which indicates that there was little carbonate present. We then washed the HCl from each sample by vortexing with 10 ml $dH_2O$, centrifuging, and decanting. Each sample was stored in 10 ml $dH_2O$.

The next pollen processing step involved acetolysis, which removes cellulose and other organics. Each sample was centrifuged and the storage water decanted, after which it was washed twice with glacial acetic acid ($CH_3COOH$) to dehydrate the residue. For each wash we added 10 ml of 99% glacial acetic acid, then vortexed, centrifuged, and decanted. To perform acetolysis we worked in small batches of four samples to control the length of the reaction. One sample at a time, we added 18 ml of 98% acetic anhydride and 2 ml of 96% sulphuric acid ($H_2SO_4$), swirled the sample-acid solution, and placed it in a heating block water bath at 100°C. After 2.5 minutes in the bath, each sample was immediately removed and the batch was centrifuged for 3 minutes at 3,500 rpm. Each sample was then decanted and neutralized with two washes of 10 ml 99% glacial acetic acid, as described above. Finally, each sample was washed twice with 10 ml $dH_2O$—vortexed, centrifuged, and decanted. Each sample was stored in 10 ml $dH_2O$ for slide mounting and/or further preparation steps. Some samples that were too difficult to count due to excess silica content on the prepared slides were subsequently treated with HF.

The concentrated and washed pollen samples were then mounted by Pacific Forestry Centre lab technician N. Conder with my observation and notation. A glass slide was prepared for each sample and heated on a warming block, to which a glycerine jelly mounting medium was added. Depending on the apparent concentration of organic material in the pellet for each sample, we added approximately 15-25 μl of sample material to the glycerine jelly. We then mixed the jelly and sample and mounted with a glass coverslip. Over the course of several tests we balanced the ratio of glycerine jelly to sample material for ease of viewing, which varied by study site and the depth of each sample. Two slides (labeled A and B) were prepared for each sample. Some samples from the lower core sections of Topknot Lake had a particularly low density of pollen and other organic material and required two additional slides (labeled C and D) to be prepared to achieve desired pollen counts.

### *Pollen Slide Counting*

For many of the samples from both Little Woss Lake and Topknot Lake, the process described above was sufficient for me to count a minimum of 300 pollen grains and spores per slide. However, the lowest samples from Little Woss Lake (n=2) and many of the lower samples from Topknot Lake (n=10) contained enough silica to impede my view of the pollen grains. In December 2018, R. Hebda and V. Pospelova treated the selected silty samples at the University of Victoria School of Earth and Ocean Sciences acid lab facility using 48% HF for three hours. The treatment was carried out in 50 ml polyethylene test tubes and the residue was stirred repeatedly to facilitate removal of minerals. After HF treatment, samples were washed and centrifuged three times with reverse osmosis water before slides were made. Many of the samples from Topknot Lake, both HF and non-HF (n=15), also contained significant fine-fraction siliceous material that obscured the pollen grains. To remove these particulates, I screened each of the affected samples through a fine 10 μm Nitex mesh and retained the coarse fraction containing the pollen grains. Following HF treatment and/or fine screening, the samples were once again mounted on slides at the Pacific Forestry Centre by lab technician N. Conder, as described above.

I performed pollen and spore counts on a Leica DM4000 B LED compound light microscope in K. Brown's palaeoecology lab at the Pacific Forestry Centre from October 2018 to

August 2019. I aimed to count each sample to a minimum of 300 grains to maintain statistical significance (see Hebda 1981; Moore et al. 1995; Bennett and Willis 2001). Most sample counts exceed 300 grains (n=21; Little Woss Lake n=14; Topknot Lake n=7). However, individual counts vary from 409 to as few as 26 identified grains in the deepest, most palynomorph-poor minerogenic sediments from Topknot Lake. Pollen identifications relied on reference materials including Kapp (1969), Moore et al. (1995), Demske et al. (2013), Martin and Harvey (2017), Klinkenberg (2019a), an unpublished key by R. Hebda, M. North and G.E. Rouse (2002), and the invaluable coaching and support of K. Brown, R. Hebda, and N. Conder. In total, I counted 14 samples from Little Woss Lake and 13 samples from Topknot Lake. Counts were checked by R. Hebda and K. Brown using test transects to confirm proportions of relevant taxa and through extended discussions on pollen morphology.

### *Other Lab Work (Radiocarbon, Diatoms, Macrofossils)*

#### Radiocarbon Samples

Radiocarbon samples were collected opportunistically throughout the cores from both study sites. However, wherever possible the samples were collected from near sedimentological transitions and from eDNA, pollen, and macrofossil sampling locations. The first round of radiocarbon samples (Little Woss Lake n=6; Topknot Lake n=6) were submitted to A.E. Lalonde AMS Laboratory at the University of Ottawa (UOC) in late August 2018, with results received in early October 2019. A second round of radiocarbon samples (Little Woss Lake n=2 from an alternate core for comparison with initial dates; Topknot Lake n=6) were submitted to UOC in early February 2019, with results received in mid-April 2019.

Most of the radiocarbon samples submitted for dating were macrobotanical remains, either monocot leaves and stems (Cyperaceae) or seeds and leaves of yellow pond-lily (*Nuphar polysepala*). However, the lower core sections from Topknot Lake (below approximately 865 cm dbp) lacked obvious macrobotanical remains that could be sampled for dating. Instead, I collected bulk sediment samples from between 881.5 and 1105 cm dbp and screened them in the archaeology lab at the University of Victoria for identifiable microscopic organic remains that would provide reliable dates. I washed these samples through nested mesh screens of 710 µm, 500 µm, and 75 µm. No obvious organic material was retained in any of the samples from the 710 µm or 500 µm screens. The 75 µm fine fraction from each sample was collected in a hexagonal dish and suspended in water. I then examined the fine fraction at 50x magnification under a Nikon SMZ-2B dissecting microscope, using forceps and a narrow-tipped glass pipette to collect organic material.

I classified the material into categories including arthropod chitin, monocot leaves, water flea (*Daphnia*) resting eggs, or other miscellaneous material depending on what was present in the sample. For every sample, each class of material was stored in separate 1.5 ml graduated microcentrifuge tubes and labeled with the core identifier, depth, date, and material category. Most of the samples I collected from the lower sections of the Topknot Lake Livingstone core did not contain enough material to produce reliable radiocarbon dates. Depending on the sample, this was either due to a simple lack of material or due to concerns over obtaining spuriously old dates from organisms that draw their carbon from the dissolved inorganic carbon of the lake basin itself rather than from atmospheric carbon stores (see Birks 2001; Marty and Myrbo 2014). To mitigate this, I included only material that could be confidently identified as the leaves, stems, and rhizomes of monocot plants (Cyperaceae), mosses (such as *Sphagnum*), or arthropod chitin when amalgamating samples to send for dating. Given these limitations, only one of the samples contained enough organic material for dating: 920 cm dbp, including monocot leaf fragments and arthropod chitin. I observed and collected small fragments of mosses, aquatic plants, and arthropod chitin even in the deepest sample (1100-1105 cm dbp), but this did not contain enough material for a successful radiocarbon assay.

### *Diatoms*

In total, 28 diatom samples of approximately 0.75 $cm^3$ each were collected from the cores from both study sites (Little Woss Lake n=8; Topknot Lake n=20). These were prepared on glass slides by lab technicians C. Celeste and A. Dyck in the archaeology lab at the University of Victoria and selected slides were analyzed by D. Fedje through Fall 2018 and Spring 2019. Diatom analysis focused on samples at sedimentological transitions near the base of the record from each site. Slide preparation and identification of the diatom samples followed standard protocol as described by Fedje et al. (2018), following Battarbee (1986). Each sample was sieved through a 500 μm mesh screen with the fine fraction retained. Where necessary, the clay fraction of each sample was decanted to improve visibility of the resulting slides. Each sample was then mounted on glass slides using Naphrax or Norland Optical Adhesive #61. Observed diatom assemblages were then characterized by salinity as either fresh, fresh-indeterminate, fresh-brackish, brackish, or marine, generalized from the halobian classes described by Hustedt (1953).

### *Macrofossils*

I did not explicitly collect macrofossil plant or animal remains as part of this study. However, through the process of subsampling and performing lab work on other proxies including pollen, diatoms, and radiocarbon samples, I observed and recorded the presence of diverse macrofossil and microfossil remains. These include leaves, needles, rhizomes, seeds, reproductive structures and fungal fruiting bodies, chitin, arthropod mouth parts, and eggs. I noted most of these while screening the lower Livingstone core sections from Topknot Lake for radiocarbon samples and while performing pollen counts.

### 3.2.4 Data Analysis and Presentation

#### *Age-Depth Models*

In order to produce a continuous late Pleistocene chronology, I created age-depth models based on the radiocarbon dates obtained from each site. Both age-depth models were created using the Bacon software package for R x64 3.6.1 (R Core Team 2019). Bacon uses Bayesian statistics to reconstruct continuous accumulation histories for lacustrine and peat deposits using intermittent dates from radiocarbon and other sources in combination with assumed prior parameters of sedimentation rate and 'memory' of accumulation across the core (Blaauw and Christen 2011). The benefit to this approach is that it can handle complicated deposition sequences that include age reversals without producing a model that implies the backward accumulation of sediments. However, because Bacon requires the input of prior assumptions such as sedimentation rate, dates that are projected forward or backward from the outermost dates in the sequence are not well-constrained (Blaauw and Christen 2011).

The sequence of radiocarbon dates from Little Woss Lake is stratigraphically consistent. Due to the apparent steady deposition through the radiocarbon sequence, my priors included the accumulation rate calculated between the highest and lowest samples along with assumed consistency in deposition across the core. There are no age reversals in the sequence and the model is well-constrained, with all six of the radiocarbon dates falling within the predicted ranges.

Age modelling for Topknot Lake is more complicated. The upper section of the core I analyzed from Topknot Lake has a much lower deposition rate than the several meters at the base of the core, which accumulated rapidly. Additionally, the radiocarbon dates from the lower sections of Topknot Lake include several reversals, though they all fall within the same general age range, many of them overlapping when their errors are expressed to two standard deviations

(95% confidence). To help accommodate this inconsistency I used two different sets of priors for the core, split by a change in sediment characteristics. The upper section of the core (509-591 cm dbp) is characterized by interbedded mats of organic matter with silty gyttja, and the radiocarbon dates collected through this section suggest a consistent but relatively low accumulation rate. The lower section of the core (591-920 cm dbp and below) consists of silty gyttja in the upper portion and massive fine silts in the lower portion with little apparent organic material. The radiocarbon dates suggest a much higher rate of sediment accumulation in the lake through this early time. Using these priors in Bacon allows the creation of an age-depth model that accounts reasonably well for differential deposition histories in the lake while also maintaining consistency with the observed stratigraphy.

### *Pollen Zonation and Diagrams*

After completing pollen counts for both Little Woss Lake and Topknot Lake, I digitized the results into spreadsheet format and imported the data into Tilia 2.1.1 (Grimm 1990). The counts from both sites were treated the same during pollen zonation and analysis.

Counts of each species were translated into percentages and the age-depth chronology generated for each site by Bacon was imported into Tilia. Based on the percentage of each species counted in each sample, I used the CONISS (Constrained Incremental Sum of Squares) program in Tilia to generate statistically significant pollen zones for the study sites (Grimm 1987). Zonation was stratigraphically constrained to maintain chronological consistency across samples and to preserve transitions in vegetation communities. I performed a square-root transformation of the data based on Edwards and Cavalli-Sforza’s chord distance, which places higher weight on rare taxa (Grimm 1987). Up-weighting rare taxa is beneficial for late Pleistocene records because infrequently occurring herbaceous pollen types are useful in characterizing periglacial environments with little vegetation cover (Grimm 1987).

# 4. Results and Discussion – ‘Article’

This chapter has been prepared as a draft for a standalone manuscript intended for publication. Sections 4.1 to 4.6 are summaries of sections contained within the preceding chapters 1 to 3 of this thesis, including the Introduction, Background, and Methods. Sections 4.7 and 4.8 contain the Results and Discussion sections of the thesis, and section 4.9 summarizes the Conclusion, which is expanded upon in chapter 5. Containing the Results and Discussion sections within this more concise ‘sandwich’ chapter is intended to allow for efficient transfer of these data to a publication following my oral defense. I will be the primary author on this publication, though many others have contributed their knowledge, expertise, and physical effort to this project as well as the broader Northern Vancouver Island Archaeology and Palaeoecology Project. I prepared and analyzed the pollen and eDNA samples, Duncan McLaren led the coring effort and is directing the overall project, Daryl Fedje analyzed the diatom samples, Mikkel Winther Pedersen and Eske Willerslev ran bioinformatics for the eDNA samples, Kendrick Brown provided facilities for pollen preparation and analysis, and Quentin Mackie and Richard Hebda provided in-depth guidance on archaeology and palaeoecology, respectively.

## 4.1 Introduction

Though comparatively few data were available at the time, Knut Fladmark (1979) was among the first to compose a robust argument for the peopling of the Americas via the Pacific coast. Fladmark (1979) attributed this hypothesis to the work of palaeoecologists like Calvin Heusser (1960), who had been developing ideas of glacial refugia on the northwest coast of North America for decades. William Mathews (1979) also proposed the possibility of a coastal migration route, though he was less favourable, describing the ice age coast as ‘a precipitous ice-front calving bergs directly into the sea,’ where migration during the coldest period would have been ‘such a formidable undertaking, by virtue of both physical and psychological barriers, as to be a highly unlikely event’ (Mathews 1979:150,153).

Prior to the publication of those two studies and for many years since, the archaeological consensus has been that the first peoples in the Americas were megafauna hunters who arrived through an ice-free corridor between the retreating Cordilleran and Laurentide ice sheets by ca. 13,400 cal BP—the Clovis people (e.g. Goebel et al. 2008; Meltzer 2009:4-5). The Clovis archaeological culture no doubt represents widespread adaptation of Indigenous peoples to the dynamic postglacial climate of the Americas. However, excavations at sites throughout both North America (Adovasio et al. 1990; Gilbert et al. 2008; Kenady et al. 2011; Waters et al. 2011; Halligan et al. 2016; Williams et al. 2018; Davis et al. 2019) and South America (Dillehay et al. 2012; Dillehay et al. 2015; Politis et al. 2016; Dillehay et al. 2017) have demonstrated that Clovis does not represent the earliest habitation of the continents. Especially following broad acceptance of the lower dates at Monte Verde (ca. 14,500-14,000 cal BP) (Dillehay 1997; Meltzer et al. 1997), researchers have called into question the ability for the interior ice-free corridor to support earlier and earlier movements of people to fit radiocarbon chronologies south of the ice sheets (Heintzman et al. 2016; Pedersen et al. 2016; Froese et al. 2019).

Returning to the ideas proposed by Fladmark (1979) and Mathews (1979), researchers began to address the idea of a coastal route for the peopling of the Americas with new archaeological, geological, and palaeoecological data in hand (Josenhans et al. 1997; Fedje et al. 2004; Ramsey et al. 2004; Fedje and Mathewes 2005; McLaren 2008; Fedje et al. 2011; Mackie et al. 2011; McLaren et al. 2014; Carlson and Baichtal 2015; Gauvreau and McLaren 2017; Fedje et al. 2018; Mackie et al. 2018; McLaren et al. 2018; Braje et al. 2019a; McLaren et al. 2019b). These research programs have been successful in unearthing late Pleistocene coastal archaeological sites. However, evidence remains elusive for sites along the coastal corridor that account for habitation as early as that documented at sites such as Monte Verde (Dillehay et al. 2008; Dillehay et al. 2015), Huaca Prieta (Dillehay et al. 2017), Arroyo Seco (Politis et al. 2016), Cooper's Ferry (Davis et al. 2019), the Debra L. Friedkin site (Waters et al. 2018), the Gault site (Williams et al. 2018), and Meadowcroft Rockshelter (Adovasio et al. 1990). These seven sites south of the ice sheets (and possibly more, see Waters 2019) all have archaeological components older than 14,000 cal BP. However, the oldest known sites on the Pacific northwest coast date to more recent than 14,000 cal BP, with the possible exception of the basal component at Triquet Island which is reported to span 14,086-13,613 cal BP (Gauvreau and McLaren 2017). The coastal corridor hypothesis indicates that the oldest sites in the Americas outside Beringia and interior Alaska and Yukon should be found on the Pacific coast. However, the vagaries of inconsistent glaciation, shifting sea levels, and unstable plant and animal communities through the late Pleistocene have stymied the search for sites that represent the first peopling process (Braje et al. 2019b). In order to help archaeologists to build better models and to make more informed research decisions, I seek to address some of these palaeoenvironmental knowledge gaps with the current study.

Two lakes located on northern Vancouver Island—Little Woss Lake and Topknot Lake—were chosen as study sites to better characterize these glacially influenced late Pleistocene environments. Bringing the records from these lakes into the broader discussion, I consider the challenges and affordances this dynamic coast may have provided to early peoples as well as the implications of this environmental history in searching for early archaeological sites on the Pacific coast of North America.

## 4.2 Regional Setting

Located off the southwestern coast of British Columbia, Canada, and 450 km long, Vancouver Island exhibits highly varied physiography. Mountains of the Vancouver Island

Ranges rise to over 2000 metres along the island's axis with quiet lowlands against their eastern slopes and storm-lashed Pacific fjords on the western coast. The island's north end is capped by the low, rolling terrain of Suquash Basin and Nahwitti Plateau (Holland 1976; Mathews 1986; Yorath 2005).

This study focuses on northern Vancouver Island, a relatively unknown portion of the theorized late Pleistocene coastal migration route of Indigenous peoples into the Americas (Braje et al. 2017; Froese et al. 2019; Braje et al. 2019a). The high, glaciated peaks of the Coast Mountains to the east and open Pacific Ocean to the west constrain the movement of people through this region. Today, as in the past, navigation of the open coastal environment of northern Vancouver Island requires well-constructed watercraft and skilful mariners.

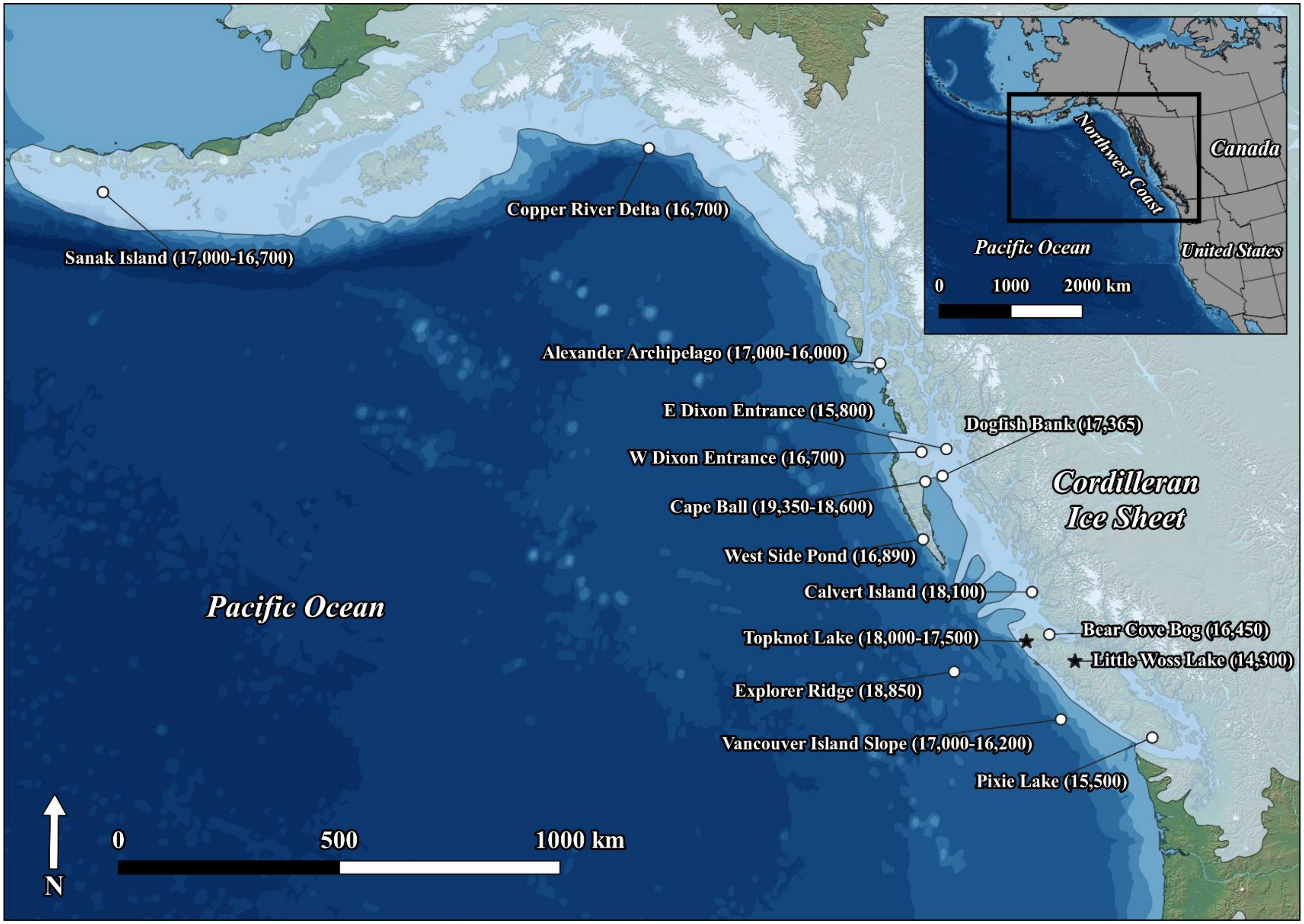
## 4.3 Glacial History

During climatic conditions much colder than today, the landscape of Vancouver Island was carved, cut, and ground away by the action of glaciers flowing from the high Coast Mountains of BC across the narrow channels and straits to the east (Clague 1989b) as well as from local island centres (Howes 1983). The most recent of these major glacial periods is referred to as the Fraser Glaciation in coastal BC (corresponding with global marine isotope stage 2), and the period preceding it as the Olympia Nonglacial Interval (corresponding with global marine isotope stage 3) (Armstrong et al. 1965; Armstrong and Clague 1977; Clague 1981; Ryder and Clague 1989; Hebda et al. 2016). The temporal extent of and relationship between the end of the Olympia Nonglacial Interval and the beginning of the Fraser Glaciation are regionally specific and poorly known in many areas. Globally, the coldest climatic period during the Last Glacial Maximum (LGM) lasted from approximately 26,500 cal BP to 19,000 cal BP (Clark et al. 2009). However, the height of glacial cover by the Cordilleran ice sheet (CIS)—corresponding to the peak of the Fraser Glaciation—varied across the coast of northwestern North America from

Alaska to Puget Sound, ranging anywhere from ca. 27,000 cal BP to nearly 17,000 cal BP. This is largely due to the influence of global and regional geomorphic and climatic processes on the oscillation of glacial growth and decay across the coast (Hays et al. 1976; Clague et al. 1980; Hicock et al. 1982; Clague 1983; Barnosky et al. 1987; Thompson et al. 1993; Hebda and Whitlock 1997; Brown and Hebda 2002; Clague and James 2002; Hendy and Cosma 2008; Nisancioglu 2009; Shaw et al. 2019).

Recent research has begun to demonstrate that in addition to the diachronous nature of Pacific coastal deglaciation, the maximum extent of ice cover during and after the LGM was also regionally specific (see Figure 15) (for example Al-Suwaidi et al. 2006; Carrara et al. 2007; Carlson and Baichtal 2015; Seguinot et al. 2016; Briner et al. 2017; Mathewes and Clague 2017; Darvill et al. 2018; Lesnek et al. 2018; Shaw et al. 2019). During the coldest climatic conditions on the coast, the CIS may have extended to the continental shelf margin in some places with localized smaller mountain glaciers contributing to a patchwork of on- and off-shore ice at the edge of the sheet (Barrie and Conway 1999; Lesnek et al. 2018). However, some outer coastal glacio-distal environments as well as plains exposed by isostatic rebound and crustal forebulge following deglaciation (see Clague 1983) may have served as refugia for plant and animal species during and immediately following the LGM (Heusser 1989; Hebda and Haggarty 1997; Soltis et al. 1997; Lacourse et al. 2003; Reimchen and Byun 2005; Shafer et al. 2010; Mathewes and Clague 2017; and others).

***Figure 15.*** *Maximum glacial cover (ca. 20,000-17,500 cal BP; after Dyke 2004) and deglaciation chronology for selected sites on the Pacific northwest coast of North America. Glacial cover is depicted in light blue. Ages are noted after the site name and are expressed in calibrated years before present (cal BP). Citations for each site noted in the text. Map data: Natural Earth, Viewfinder DEM, GeoBC.*



## 4.4 Palaeoecology

Plants and animals were continually displaced by the ebb and flow of ice and associated climatic conditions across the coast of northwestern North America during the Fraser Glaciation, sometimes into a patchwork of coastal refugia and sometimes entirely out of their previously established ranges. The subsequent recolonization of coastal landscapes was heavily dependent on local refugial conditions, deglacial history, relative sea level change, and other factors.

### *Pre-28,000 cal BP*

Prior to the onset of the Fraser Glaciation (generally >28,000 cal BP), palaeoecological records across the coast indicate cool and dry interglacial conditions with varied ecological communities. Southeastern Alaska, Haida Gwaii, northern Vancouver Island, and southern Vancouver Island have all yielded records demonstrating herb-shrub tundra or conifer parkland with diverse mammals, birds, and fish (Heaton and Grady 2003; Wigen 2005; Al-Suwaidi et al. 2006; Miskelly 2012; Mathewes and Clague 2017).

### *ca. 28,000-16,000 cal BP*

As the coast began to descend into glacial conditions after ca. 28,000 cal BP, climatic changes and encroaching ice eventually led to the local extirpation of many species (Heaton and Grady 2003; Wigen 2005; Al-Suwaidi et al. 2006; Lesnek et al. 2018). The specific timing of local glacial maxima and associated gap in biotic records vary across the coast, but are generally documented between ca. 20,000 and 17,000 cal BP (Barrie and Conway 1999; Wigen 2005; Al-Suwaidi et al. 2006; Mathewes and Clague 2017; Darvill et al. 2018; Lesnek et al. 2018; Shaw et al. 2019). Cold-adapted marine and terrestrial animals including ringed seal, northern sea lion, arctic fox, and some species of waterbirds survived this climatically harsh interval in places such as Haida Gwaii and southeastern Alaska (Heaton and Grady 2003; Wigen 2005). Other, less coldtolerant species including brown bear, voles, ermine, and chum salmon retreated to refugia in Hecate Lowlands and elsewhere on the coast (Heusser 1989; Kondzela et al. 1994; Heaton et al. 1996; Conroy and Cook 2000; Fleming and Cook 2002; Heaton and Grady 2003; Wigen 2005; Lesnek et al. 2018).

### *ca. 16,000-11,500 cal BP*

Early plant and animal communities that emerged as environments ameliorated following local glacial maxima (ca. 16,000-11,500 cal BP) generally indicate cold, dry herb/shrub tundra environments, especially in Haida Gwaii (Barrie et al. 1993; Heusser 1995; Lacourse and Mathewes 2005; Lacourse et al. 2005; Lacourse et al. 2012; Mathewes et al. 2019). As climate continued to warm and ice retreated, pine-dominated parkland emerged across the coast, often with lingering cold-adapted herb and shrub species. These environments have been recorded generally after ca. 15,500-14,000 cal BP in southeastern Alaska (Ager et al. 2010; Ager 2019), Haida Gwaii (Hebda et al. 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Lacourse et al. 2012; Mathewes et al. 2019), the north and central coasts of BC (Lacourse et al. 2003; Galloway et al. 2008; McLaren 2008; Eamer 2015), northern Vancouver Island (Hebda 1983; Lacourse 2005; Al-Suwaidi et al. 2006), and southern Vancouver Island (Brown and Hebda 2002; Lucas and Lacourse 2013; Leopold et al. 2016). Alongside these pioneering communities, evidence from caves in southeastern Alaska (Heaton and Grady 2003), Haida Gwaii (Ramsey et al. 2004; Wigen 2005; Fedje et al. 2011), and northern Vancouver Island (Nagorsen and Keddie 2000; Al-Suwaidi et al. 2006; Steffen and McLaren 2008; Steffen and Fulton 2018) also indicates the presence of large mammals including black, brown, and short-faced bear after ca. 14,500 cal BP. Palaeontological records are not extensive from most places on the coast until after the Younger Dryas, but in southeastern Alaska, caribou, river otter, wolverine, and small mammals have been identified from terminal Pleistocene deposits at Shuká Káa (Heaton and Grady 2003). Remains of megafauna that has been hunted by humans, including ancient bison and American mastodon, have been collected in the Salish Sea that date to at least ca. 13,800 cal BP (Kenady et al. 2011; Waters et al. 2011), and evidence for Younger Dryas bear hunting has been documented on Haida Gwaii dating to ca. 12,600 cal BP (Fedje et al. 2011).

*Post-11,500 cal BP*

By the end of the Younger Dryas (ca. 11,500 cal BP) and especially through the early Holocene thick mixed-conifer forests and diverse animal species appear in records across the coast (Hebda 1983; Brown and Hebda 2002; Hebda et al. 2005; Lacourse 2005; Lacourse and Mathewes 2005; Galloway et al. 2008; Eamer 2015; Ager 2019; Mathewes et al. 2019). At the same time, post-Younger Dryas palaeontological records indicate numerous species across the coast from southeastern Alaska to Vancouver Island in the early Holocene (Cannon 1996; Heaton and Grady 2003; Wigen 2005; Cannon and Yang 2006; Fedje et al. 2011; Fargo 2013; McLaren et al. 2015). Eventually through the mid- and late-Holocene, mixed-conifer forests transitioned to the widespread temperate rainforests seen today, dominated by western redcedar and western hemlock (Hebda and Mathewes 1984; Hebda 1995; Hebda and Whitlock 1997).

## 4.5 Late Pleistocene Archaeology

Early peoples undoubtedly occupied the dynamic glacial and postglacial landscapes of the Pacific northwest coast, though their signature is difficult to detect beneath deep seas and in thick rainforests which are resistant to erosion and exposure (Fedje et al. 2011; Mackie et al. 2011; Fedje et al. 2018; Mackie et al. 2018; McLaren et al. 2018; Braje et al. 2019a; 2019b; McLaren et al. 2019b).

The earliest archaeological sites on the coast are known from areas in which relative sea level during the late Pleistocene was either comparable to or higher than modern sea level (Fedje et al. 2011; McLaren et al. 2014; Shugar et al. 2014; McLaren et al. 2015; Gauvreau and

McLaren 2017; McLaren et al. 2018). Evidence of early human occupation on the central coast of BC comes from Triquet Island, where a cache of stone tools and a hearth has been dated to ca. 14,086 to 13,613 cal BP (Gauvreau and McLaren 2017). The evidence at Triquet Island along with remains from other nearby early sites, including lithics from Kildidt Narrows (McLaren et al. 2015) and ancient footprints at Meay Channel I (McLaren et al. 2018), demonstrate human presence and consistent sea levels on this part of the coast between ca. 14,000-13,000 cal BP.

Many archaeological sites in Haida Gwaii have also yielded early dates, ranging from approximately 13,000 to 10,000 cal BP across the archipelago, including at K1 Cave, Gaadu Din 1 and 2, Werner Bay, Kilgii Gwaay, Richardson Island, Arrow Creek 1 and 2, and Lyell Bay (Fedje and Christensen 1999; Fedje et al. 2005; Storey 2008; Fedje et al. 2011; Mackie et al. 2011; Cohen 2014; Cohen 2019). These earliest sites also contain extensive faunal remains, including birds, fish, and small and large mammals such as deer, black bear, and brown bear, with artifacts including foliate and stemmed bifaces, flaked stone points, and bone points in addition to smaller flake tools and debitage (Fedje et al. 2011). Kilgii Gwaay, though relatively younger than some of the sites in Haida Gwaii (ca. 10,700 cal BP) yielded an extensive assemblage of water-saturated artifacts, including wooden stakes, twine, and other organic remains in addition to marine-oriented faunal and stone tool assemblages (Fedje et al. 2005; Cohen 2014; Cohen 2019). Werner Bay is also noteworthy—the site consists of a single utilized stone flake tool recovered by bucket dredge from the sea floor at 53 m below modern sea level in the waters of Juan Perez Sound off Haida Gwaii (Fedje and Josenhans 2000). Sea level reconstructions indicate that this palaeo-landform would have been above sea level ca. 12,100 to 11,700 cal BP (Fedje et al. 2005). Additionally, late Pleistocene deposits from Shuká Káa in southeastern Alaska (ca. 12,000 cal BP) have yielded evidence of human habitation, including a worked bone tool (Dixon 2013).

On the south coast of BC few late Pleistocene archaeological sites have been documented, likely due to limited research effort in the area and the challenges of developing effective archaeological potential models for raised beach landscapes (but see Vogelaar 2017 and Lausanne 2018). The earliest archaeological sites that have been documented in the Salish Sea date to ca. 13,900-13,750 and include Ayer Pond (Kenady et al. 2011) and Manis Mastodon (Waters et al. 2011), both of which rely on modified faunal material to demonstrate human agency rather than extensive artifact assemblages. However, some recent finds of artifacts at Bear Creek in Puget Sound including bifaces, expedient flake tools, and apparent fluted stemmed points have helped to expand the early record in the area (ca. 12,770 to 11,274 cal BP; Kopperl et al. 2015). Stemmed and foliate points have also been collected from the Stave Watershed on the nearby mainland where the associated archaeological component dates to ca. 12,404-12,051 cal BP (McLaren 2017; McLaren et al. 2019b). Despite the numerous constraints of the glacial, biotic, and archaeological history of the region, sites like these demonstrate early human occupation of the submerged ancient plains. However, these late Pleistocene sites are not early enough to account for the antiquity of sites elsewhere in the Americas (Braje et al. 2019a; Waters 2019). Understanding the complex local and regional processes of deglaciation, relative sea level change, and ecological succession across the coast is paramount to uncovering this early archaeological record.

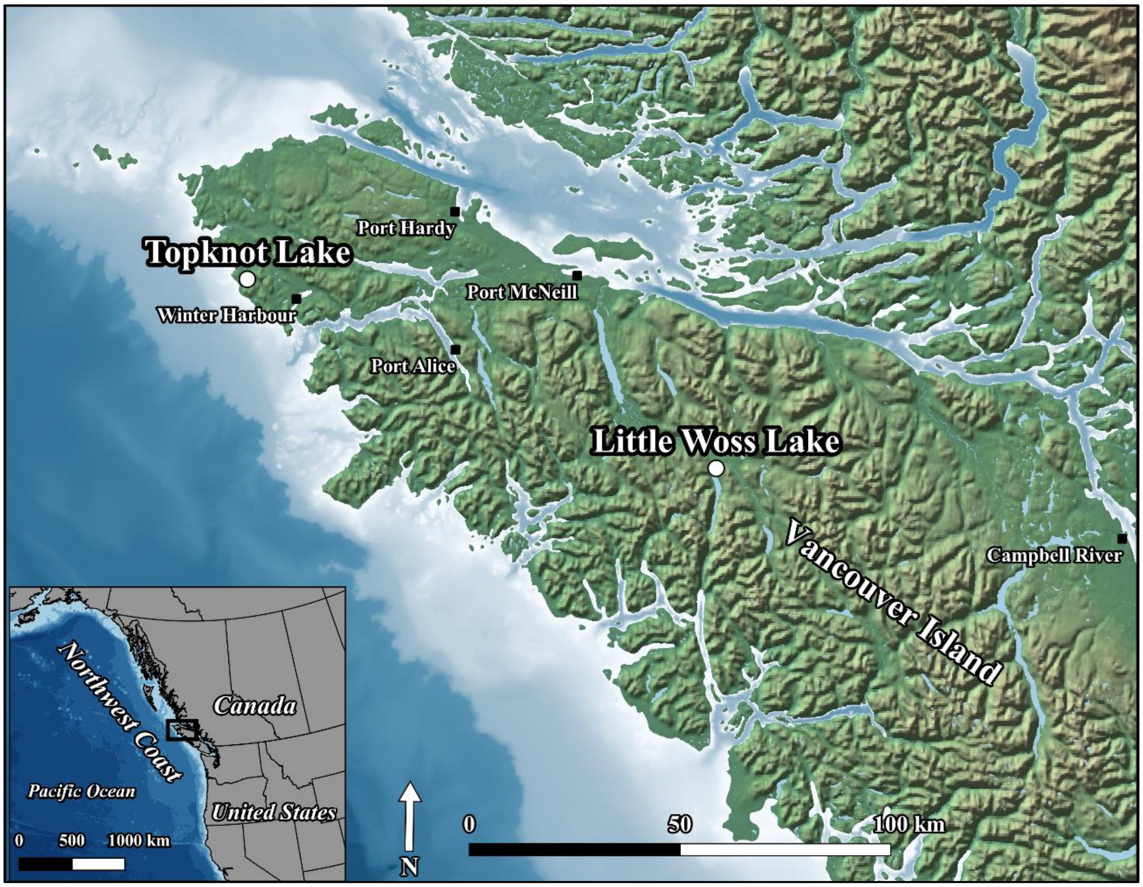
## 4.6 Materials and Methods

Pollen analysis has formed the core of palaeoenvironmental reconstruction for many areas of British Columbia (Heusser 1960; Hebda 1983; Hebda 1995; Heusser 1995; Brown and Hebda 2002; Lacourse et al. 2003; Hebda et al. 2005; Lacourse 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Al-Suwaidi et al. 2006; Galloway et al. 2008; Brown et al. 2017; Mathewes and Clague 2017; and many others). When coupled with other sources, including palaeontological records and faunal remains from archaeological sites (Heaton et al. 1996;

Nagorsen and Keddie 2000; Heaton and Grady 2003; Wigen 2005; Steffen and McLaren 2008; Fargo 2013; Steffen and Fulton 2018; and others), these data can provide regionally specific reconstructions of the environment in coastal British Columbia stretching from the present day to the late Pleistocene. While these methods can provide good indicators of past conditions, each also relies on specific assumptions and biases that can colour the interpretation of the evidence (Hebda 1981; Moore et al. 1995; Bennett and Willis 2001). By applying multi-proxy approaches and introducing new methods into these established research workflows, we can mitigate certain methodological weaknesses to further strengthen the interpretation of past environments (Jørgensen et al. 2012b; Pedersen et al. 2013). This study applies several proxies—including pollen, diatom, macrofossil, and ancient eDNA analyses—to reconstruct terminal Pleistocene environments on northern Vancouver Island.

I selected Little Woss Lake (153 m asl) and Topknot Lake (13 m asl) on north-central and northwestern Vancouver Island, respectively, as the two study sites for this research (see Figure 16). I believed that these lakes could produce extensive late Pleistocene records and possibly demonstrate the presence of glacial refugia. I chose Little Woss Lake due to its location in the sheltered rainshadow of the Nimpkish River drainage and its proximity to Woss River, a large salmon-bearing stream with high late Pleistocene archaeological potential. I chose Topknot Lake due to its location on the outer coast and distance from the source areas of ice during the Fraser Glaciation, as well as its elevation near the theoretical late Pleistocene/early Holocene relative sea level highstand (Howes 1983).

***Figure 16.*** *Study sites at Little Woss Lake and Topknot Lake on northern Vancouver Island. Nearby settlements are labeled with black squares. Map data: Natural Earth, Viewfinder DEM, GeoBC.*



### 4.6.1 Coring

In May 2018, we cored Little Woss Lake and Topknot Lake as part of fieldwork for the Northern Vancouver Island Archaeology and Palaeoecology Project (NVIAPP) using a combination of a Livingstone corer (Livingstone 1955; Vallentyne 1955; Wright 1967) and a modified version of the Reasoner coring method (Reasoner 1986;1993). Livingstone core sections were collected in 90 cm to 1 m drives using a 5-cm diameter aluminum core barrel and were extruded and packed for transport in the field. Modified Reasoner-method cores were collected continuously using 3.5-m sections of 7.5-cm diameter solid-core ABS tubing. Additional 3.5-m sections of tubing were attached as necessary until we reached basal deposits or were otherwise unable to continue the drive. The continuous Reasoner-method cores were split into manageable lengths in the field and capped for transport. All cores were stored at 4°C at Natural Resources Canada's Pacific Forestry Centre in Victoria, BC, between sampling.

### 4.6.2 Subsampling and Preparation

Cores from both sites were sampled at several different times to extract material for analysis by several proxies. First, the cores were sampled for ancient eDNA analysis at Hakai Institute eDNA lab facilities in Victoria, BC, using fully-body coveralls, shoe covers, facemasks, gloves, and safety goggles to limit contamination. eDNA sample handling and preparation for this study followed protocols in Pedersen et al. (2016). I sampled the lower sections of the cores from each study site, focusing on sediment transitions. Samples were taken from the entirety of the continuous Reasoner-method core from Little Woss Lake (22 samples collected; 5 processed) and from all Livingstone core sections from Topknot Lake (52 samples collected; 9 processed).

Subsampling for pollen, diatoms, and radiocarbon samples was carried out in laboratory facilities at the Pacific Forestry Centre. Subsampling for these proxies included the lower section of the continuous Reasoner-method core from Little Woss Lake as well as all sections of the Livingstone core sections from Topknot Lake. The resolution of pollen (Little Woss Lake, 14 samples collected and processed; Topknot Lake, 34 samples collected, 18 processed) and diatom sampling (Little Woss Lake, 8 samples collected and processed; Topknot Lake, 20 samples collected and processed) was dependent on the presence of sediment transitions and obtaining representative core coverage. Both pollen and diatom samples were prepared and mounted for analysis following standard methods (see Battarbee 1986; Faegri and Iversen 1989; Moore et al. 1995; Fedje et al. 2018). Pollen samples were spiked with an exotic marker tablet to determine pollen and spore concentrations (20848 ± 1,546 *Lycopodium* spores/tablet; Dept. Quaternary Geology, University of Lund, batch no. 1031). Some of the lowest samples from both study sites required treatment with HF to remove excess silica, and some required screening with 10 μm Nitex mesh to remove fine particulates.

Radiocarbon samples were collected opportunistically from both cores based on the presence of macrofossils such as needles, leaves, and other preserved foliage in the core sections from both study sites. These macrofossils include the preserved remains of leafy plants including sedges as well as aquatic vascular plants (e.g. yellow pond-lily) and larger twigs, branches, and needles (e.g. shore pine). Where larger organic material was unavailable, bulk samples of sediment were screened and picked for fragments of the remains described above. Throughout this process, macrofossil remains were recorded and photographed where possible. These macrofossils are described in the results that follow.

### 4.6.3 Lab Work and Analysis

Ancient environmental DNA lab work was conducted at the Centre for GeoGenetics at the University of Copenhagen, Denmark, in July 2018. Extractions were performed using methods described by Wales et al. (2014) and Pedersen et al. (2016). Following PCR amplification and evaluation using gel electrophoresis, three samples (Little Woss Lake, 2 samples; Topknot Lake, 1 sample) were successfully sequenced. Bioinformatic analysis was performed by M.W. Pedersen at the Centre for GeoGenetics using a modified version of ‘Holi,’ the workflow developed by Pedersen et al. (2016). Metagenomic reads were trimmed of their adaptor sequences, with resulting segments of >30 bp retained. Reads with poly-A/T tails ≥4 were removed from the samples. Duplicate reads resulting from PCR were removed from each sample, and low-complexity sequences were removed using String Graph Assembler (SGA) in order to decrease computational time. These trimmed and cleaned sequences were compared against several international genomic databases including NCBI Nucleotide, RefSeq, and the Genome Taxonomy Database (GTDB) using BowTie2 and were assigned to matching records based on

LCA algorithms. The results presented here include the most parsimonious identification with an edit distance of 0 nucleobases—sequences that exactly match the reference genome for each taxon.

Pollen and spore counts were made at the Pacific Forestry Centre to exceed 300 grains (Little Woss Lake, 14 samples; Topknot Lake, 7 samples). However, individual counts vary from as high as 409 to as few as 26 identified grains in the deepest, most palynomorph-poor minerogenic sediments of Topknot Lake. Raw counts for each site were transferred into the computer program Tilia and were translated into percentages (Grimm 1990). These were subjected to stratigraphically constrained cluster analysis using CONISS (Grimm 1987) to establish pollen zones based on Edwards and Cavalli-Sforza’s chord distance, which places higher weight on rare taxa. Up-weighting rare taxa is beneficial for late Pleistocene records because infrequent herbaceous taxa are useful in characterizing transitions in periglacial environments with little vegetation cover (Grimm 1987).

Diatom counts were performed by D. Fedje at the University of Victoria based on presence or absence of taxa. Observed assemblages were then characterized as either fresh, fresh-brackish/alkaline, brackish, or marine, generalized from the halobian classes described by Hustedt (1953).

## 4.7 Results

### 4.7.1 Little Woss Lake

*Site Description*

Little Woss Lake (50.180729°, -126.611854°) is located within ‘Namgis First Nation territory on north-central Vancouver Island, British Columbia, Canada. At 153 m above modern sea level, the lake is nestled in a valley bottom in the northern Vancouver Island Ranges at the northern end of the much larger Woss Lake, which is approximately 550 m to the southwest. Both Woss Lake and Little Woss Lake are part of an elongated valley watershed feeding northwestward into the Nimpkish River Valley and Nimpkish Lake. Little Woss Lake is located approximately 3 km south-southwest of the community of Woss and 26 km southeast of the southern end of Nimpkish Lake. On the eastern side of the valley, the lake receives sediment runoff from slopes to the east which rise to approximately 550 m above modern sea level. From an outlet at the basin’s southern end, Little Woss Lake drains south into Woss Lake. The lake measures approximately 700 m north-south and 250 m east-west with an area of 10 hectares (GeoBC 2019).

The underlying bedrock geology around Little Woss Lake is a mosaic of Triassic and Jurassic igneous and sedimentary formations (BC Geological Survey 2019). Little Woss Lake, the northern end of Woss Lake, and parts of the Nimpkish River Valley rest upon intrusive granodioritic rocks of the Island Plutonic Suite. However, much of the southern two-thirds of Woss Lake and the southern and western sides of Nimpkish Lake are underlain by dark grey-green Karmutsen Formation basalt pillow lava flows, with small limestone formations from the Sicker Formation-Buttle Lake Group and the Quatsino Formation nearby in pockets to the east and the west (Nixon et al. 2011b).

Today, Little Woss Lake falls within the Western Very Dry Maritime subzone of the Coastal Western Hemlock biogeoclimatic ecosystem classification (CWHxm2; BC Forest Analysis and Inventory Branch 2018). The vegetation regime of the CWHxm2 subzone is characterized by tree species including Douglas-fir, western hemlock, and western redcedar with an understory of salal, dull Oregon-grape, red huckleberry, vanilla leaf, sword fern, and twinflower (Green and Klinka 1994). In wetland and bog/fen environments such as those surrounding Little Woss Lake, the subzone is characterized by Labrador tea, sweet gale, Sitka sedge, bog cranberry, western bog-laurel, and peat-moss among other species (MacKenzie and Moran 2004). The dry valley bottoms transition to somewhat wetter and cooler areas upslope that also include stands of amabilis fir, mountain hemlock, and yellow-cedar (Green and Klinka 1994). Species observed around the margins of Little Woss Lake in May 2018 include western hemlock, western redcedar, sweet gale, black twinberry, and sedge.

## *Stratigraphy and Chronology*

### *Radiocarbon Dates*

Six samples were successfully dated in the lower portion of the core from Little Woss Lake. The samples range in depth from 315 to 171 cm LD with median calibrated ages in stratigraphic order between 14,285 and 10,703 cal BP (Table 2). The dates in the table below were calibrated using CALIB 7.1.0 (Stuiver et al. 2019) with IntCal13, the terrestrial northern hemisphere calibration curve (Reimer et al. 2013).

***Table 2.*** *Radiocarbon assay results from Little Woss Lake including sample collection depth, type of material dated, lab ID from UOC (A.E. Lalonde AMS Laboratory) as well as conventional $^{14}$C ages and calibrated age ranges derived from CALIB 7.1.0 using IntCal13 (Stuiver et al. 2019).*

| **Depth (cm LD)** | **Material** | **Lab ID** | **Radiocarbon age ($^{14}$C BP)** | **Calibrated age range (cal BP, ±2σ)** | **Median calibrated age (cal BP)** |
|---|---|---|---|---|---|
| 171 | matted leaves in fibrous gyttja | UOC-7374 | 9464 ± 36 | 10,787-10,582 (95.7%)<br>10,996-10,976 (1.7%)<br>11,059-11,034 (2.5%) | 10,703 |
| 201 | twig fragment in fibrous gyttja | UOC-7373 | 9995 ± 31 | 11,618-11,286 (100.0%) | 11,459 |
| 240 | *Nuphar* head in fibrous gyttja | UOC-7375 | 10,356 ± 33 | 12,389-12,046 (100.0%) | 12,218 |
| 298.5 | twig fragment in fibrous gyttja | UOC-7372 | 11,637 ± 38 | 13,568-13,396 (100.0%) | 13,469 |
| 307 | twig fragment in silty gyttja | UOC-7371 | 12,034 ± 38 | 14,017-13,759 (100.0%) | 13,879 |
| 315 | *Pinus* needle in silty gyttja | UOC-7370 | 12,332 ± 35 | 14,609-14,102 (100.0%) | 14,285 |

### Stratigraphy

In the field, we sampled to a depth of 532 cm below the sediment-water interface. This sediment was compressed during the percussion action of the core drive to 318 cm of total lab depth (LD) (see Figure 17).

Stratigraphy for the core analyzed from Little Woss Lake consists of the following units in relation to their measured LD, beginning at the base of the sequence:

(1) **LWL Unit 1, >318 cm LD**: This portion of the record consists of impenetrable pebble gravels and pea gravels that were not retained in the core tube save for a few clasts embedded at the base of the gyttja above. These gravels were not analyzed as part of this study.

(2) **LWL Unit 2, 318-311 cm LD (7 cm thick)**: This unit consists of fine silty gyttja that is light brownish-grey with a limited organic component (<5%). One $^{14}$C macrofossil

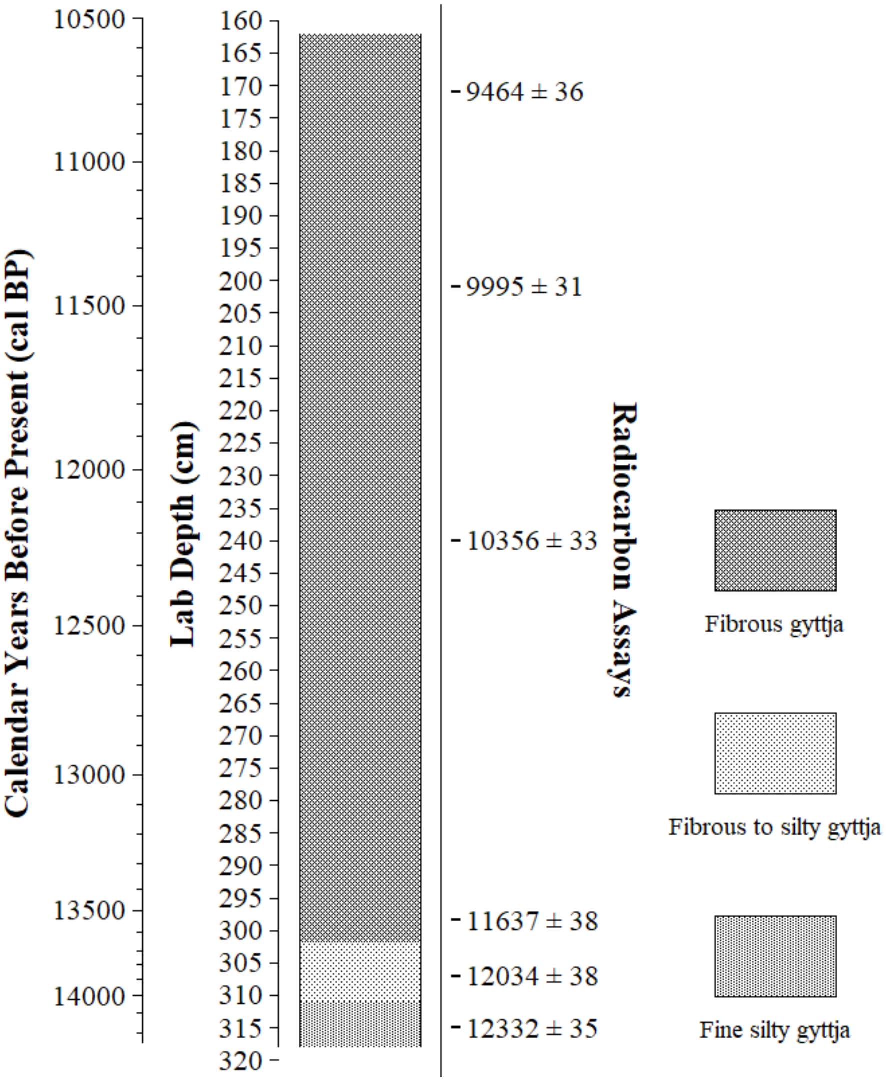
sample (*Pinus* needle) was found *in situ* within the centre of the core at 315 cm LD (UOC-7370, 12,332 ± 35 $^{14}$C BP).

(3) **LWL Unit 3, 311-302 cm LD (9 cm thick)**: This unit is transitional from the fibrous gyttja above to silty gyttja below and is progressively finer and siltier with depth. It has a medium-brown organic component (generally <20-30% organic, especially toward bottom of transition). One $^{14}$C macrofossil sample (twig fragment) was dated from this component at 307 cm LD (UOC-7371, 12,034 ± 38 $^{14}$C BP).

(4) **LWL Unit 4, 302-162 cm LD (140 cm thick)**: This unit consists of dark brown fibrous gyttja with leafy organic lenses throughout and is highly organic (generally ~30-50%). Four $^{14}$C macrofossil samples were dated from this component at 298.5 cm LD (twig fragment, UOC-7372, 11,637 ± 38 $^{14}$C BP), 240 cm LD (*Nuphar* flower, UOC-7375, 10,356 ± 33 $^{14}$C BP), 201 cm LD (twig fragment, UOC-7373, 9995 ± 31 $^{14}$C BP), and 171 cm LD (matted leaves, UOC-7374, 9464 ± 36 $^{14}$C BP). This unit includes most of the late Pleistocene and early Holocene sequence (ca. 13,600 to 10,500 cal BP).

(5) **LWL Unit 5, 162-0 cm LD**: These sediments were collected but not analyzed as part of this study. This material consists of dark brown fibrous gyttja and lake peat and is highly organic (>50%) to the lake surface. The unit includes most of the Holocene record from this site, from ca. 10,500 cal BP to the present.

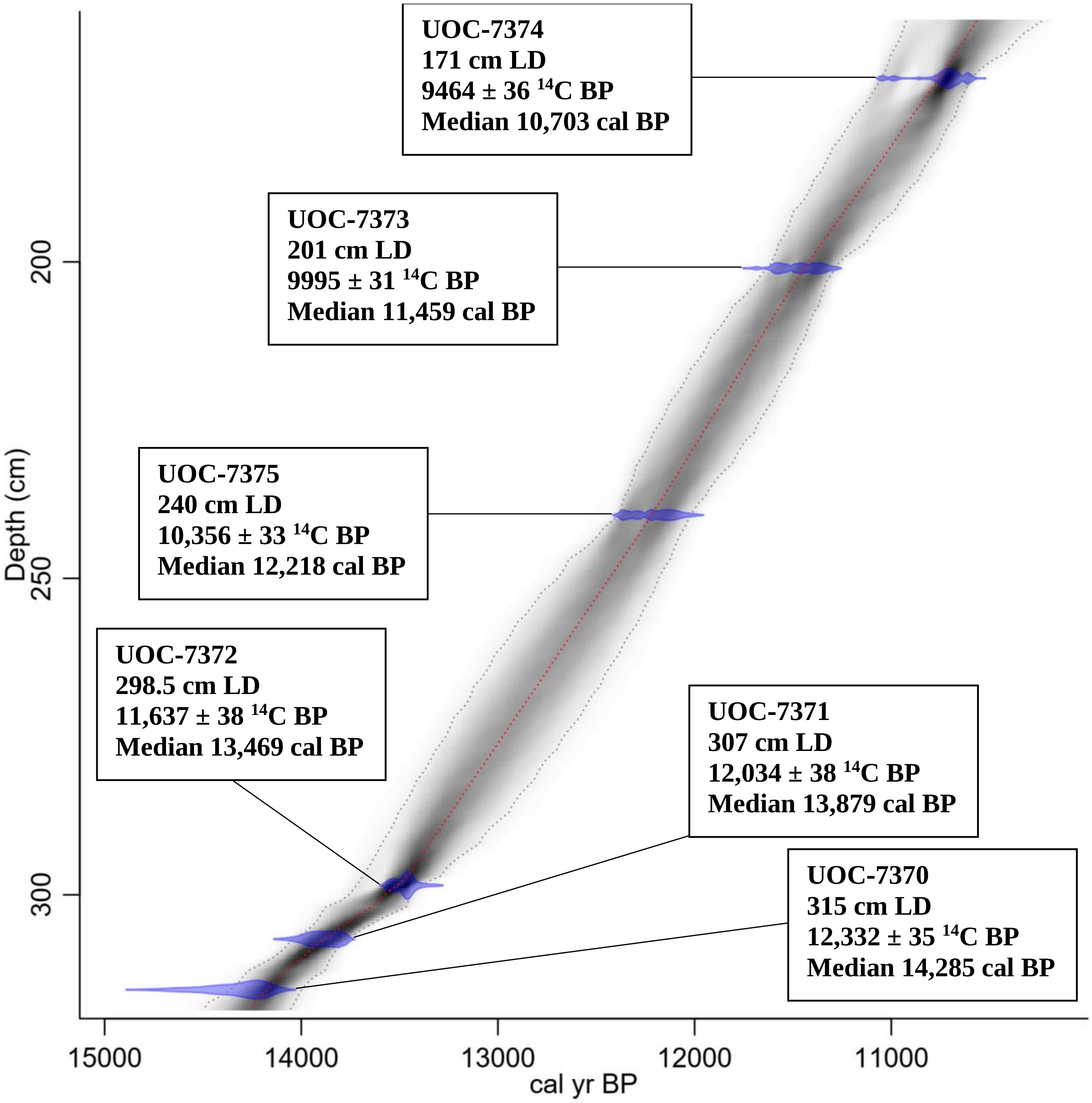
***Figure 17.*** *Diagram depicting stratigraphy of analyzed sediments from Little Woss Lake. Calibrated radiocarbon ages are plotted on the left and are based on the results of the age-depth model shown in Figure 18. Uncalibrated* $^{14}$*C ages are plotted on the right.*



## <u>Age-Depth Model</u>

The uncalibrated $^{14}$C dates listed in Table 2 were used to create a continuous age-depth model (Figure 18) for the lower deposits at Little Woss Lake using the Bacon software package for R x64 3.6.1 (Blaauw and Christen 2011; R Core Team 2019). Bacon uses the IntCal13 terrestrial northern hemisphere calibration curve in its analysis (Reimer et al. 2013). The ages from this model were used to create the chronology for the pollen diagram and zone descriptions for Little Woss Lake.

***Figure 18.*** *Bacon age-depth model for Little Woss Lake generated based on radiocarbon results. Individual radiocarbon assays are labeled. Darker grey sections of the graph indicate higher likelihood of the model passing through a particular age at that depth.*



*Pollen Analysis*

CONISS cluster analysis (Grimm 1987) identified three major pollen zones with >0.25 total sum of squares distance (LWL-1 at the base of the core up to LWL-3b at the top of the section). The third zone, LWL-3, is split into two subzones, LWL-3a and LWL-3b due to similarities in the assemblage. Arboreal pollen (AP) assemblages dominate throughout all three zones but vary in their composition.

The deepest zone, LWL-1 (14,137-14,077 cal BP; two samples), consists of a mixed-conifer assemblage dominated by *Abies* and *Pinus* with notable *Alnus* and ferns but relatively few herbaceous taxa. Pollen accumulation rates in zone LWL-1 are moderate when compared to other zones, ranging from ~4560 to 5240 grains/cm$^2$/cal yr. The second-lowest zone, LWL-2 (14,077-12,055 cal BP; eight samples), is dominated by *Pinus* and *Alnus* with much lower values for *Abies*, *Picea*, and the appearance of *Tsuga* species. This zone also contains notably more and varied non-arboreal pollen and spores (NAP), including herbaceous, aquatic, and fern taxa. PAR values in LWL-2 begin quite low (~2160 grains/cm$^2$/cal yr) but with an upward trend driven by increasing *Alnus* and *Pinus* pollen toward the end of the zone, to a maximum of ~10430 grains/cm$^2$/cal yr. The next zone, LWL-3a (12,055-11,118 cal BP; three samples), exhibits less *Alnus* with a notable increase in both *Tsuga heterophylla* and *T. mertensiana* pollen. There is relatively little NAP present, and PAR values decrease through the zone from a high value of ~8630 grains/cm$^2$/cal yr to a low of ~2620 grains/cm$^2$/cal yr. Reduced accumulation of *Alnus*, *Pinus*, and *T. heterophylla* through time is responsible for this declining trend. The uppermost zone considered in my analysis, LWL-3b (11,118-10,744 cal BP; one sample), is characterized by higher counts of *Pinus* and *Alnus* along with reduced presence of both *T. heterophylla* and *T. mertensiana* pollen when compared with zone LWL-3a below it. Pollen from herbaceous species is infrequent in this zone, but PAR values increase notably from LWL-3b to ~8,010 grains/cm$^2$/cal yr, driven once more by increasing influx of *Alnus* and *Pinus* pollen along with shrub and fern species including *Salix* and *Equisetum*. More detailed descriptions of these zones are provided below, and relative pollen and spore contributions are graphed in Figure 19.

### *LWL-1: 314-312 cm LD; 14,137-14,077 cal BP*

In this thin basal section, the assemblage at Little Woss Lake is dominated by arboreal (tree and shrub) pollen, ranging from 93.6% to 97.8% of the total. The remainder consists of pteridophytes (ferns; 0.6% to 4.2%), herbaceous taxa (0.9% to 1.9%), and aquatic vascular plants (0.3% to 0.6%). The most common taxa in this zone are *Abies* (29.9% to 39.4%), *Pinus* (31.1% to 34.4%), and *Alnus* (20.5% to 27.7%). Other arboreal taxa are present in small amounts, including *Picea* (0.3% to 1.2%), *Tsuga mertensiana* (0% to 1.2%), and unidentifiable coniferous pollen (0% to 2.8%), with trace amounts (<1%) of *T. heterophylla*, Cupressaceae (possibly *Juniperus*), and *Populus*. Shrub taxa, including *Salix* and Ericaceae, are present in trace amounts (<1%), and herbaceous species are rare with mainly Cyperaceae represented (0.9% to 1.6%). Pollen originating from aquatic vascular plants occurs in similarly small amounts, with <1% of *Typha* and *Nuphar* present in this zone. The incidence of pteridophyte taxa is slightly higher, with monolete-type ferns (0% to 2.9%) and trace amounts (<1%) of *Cryptogramma*, *Equisetum*, *Pentagramma triangularis*, and unspecified trilete-type ferns.

### *LWL-2: 312-232 cm LD; 14,077-12,055 cal BP*

This zone, representing the greatest amount of deposition in the analyzed core section (80 cm), is also dominated by trees and shrubs (86.0% to 96.9% of the total). The incidence of herbaceous taxa varies through this section (0% to 8.1%), with modestly abundant pteridophytes (2.0% to 7.0%) and low levels of aquatic vascular plants (0.3% to 2.4%). *Pinus* is the dominant pollen type in this zone (26.0% to 65.4%) followed by *Alnus* (18.0% to 39.0%). Percentages of *Abies* (2.1% to 14.6%) decrease markedly when compared with LWL-1 but other tree taxa see modest increases, including *Picea* (ranging from 0% to 4.3%), *Tsuga mertensiana* (2.6% to 7.6%), and *T. heterophylla* (0.5% to 19.4%). Other tree species, including *Populus*, occur at relatively low amounts (0% to 1.4%), with trace occurrences (<1%) of *Pseudotsuga menziesii* and

Cupressaceae. Shrubs including *Salix* (0.3% to 1.6%) and trace amounts (<1%) of Ericaceae, *Myrica gale*, and *Cornus* are infrequent in this zone.

Graminoid and herb taxa are more common and varied in this zone than in LWL-1, including Poaceae (0% to 1.6%), Cyperaceae (0% to 5.9%), and trace amounts (<1%) of Rosaceae, Asteraceae, and *Artemisia*. Pollen derived from aquatic vascular plants is somewhat more common than in the lowest component but still infrequent, including *Potamogeton* (0% to 1.8%), *Nuphar* (0.3% to 1.5%), and trace amounts (<1%) of *Typha* and *Isoëtes*. The occurrence of pteridophytes is variable, but generally these spore types are more common through this period than in any of the other zones, including monolete-type ferns (0.6% to 4.6%), *Equisetum* (0% to 1.5%), *Pteridium aquilinum* (0% to 1.1%), unspecified trilete-type ferns (0% to 1.5%), and trace amounts (<1%) of *Cryptogramma* spores.

### *LWL-3a: 232-187 cm LD; 12,055-11,118 cal BP*

Tree and shrub taxa remain most common through this zone (92.6% to 97.3%), but the diverse assemblage of herbs (0.7% to 2.1%), pteridophytes (1.7% to 4.2%), and aquatic vascular plants (0% to 2.2%) present in earlier zones is reduced. *Pinus* remains the dominant taxa in the arboreal assemblage (22.4% to 41.1%); however, the occurrence of other tree taxa shifts. Especially in the early part of the zone, *Alnus* appears less frequently (19.0% to 22.7%, though its proportion in the transitional sample with the zone above rises to 34.9%). The incidence of *Tsuga heterophylla* (18.2% to 22.0%) increases, with relatively unchanged percentages of other tree species including *Picea* (3.7% to 9.8%), *T. mertensiana* (1.0% to 8.6%), *Abies* (2.4% to 7.1%), and *Populus* (0.6% to 1.5%). Trace amounts (<1%) of *Pseudotsuga menziesii* and *Betula* were also observed in this zone. Shrub species are rare in LWL-3a, with only *Salix* (0.2% to 1.5%) and trace amounts (<1%) of Ericaceae.

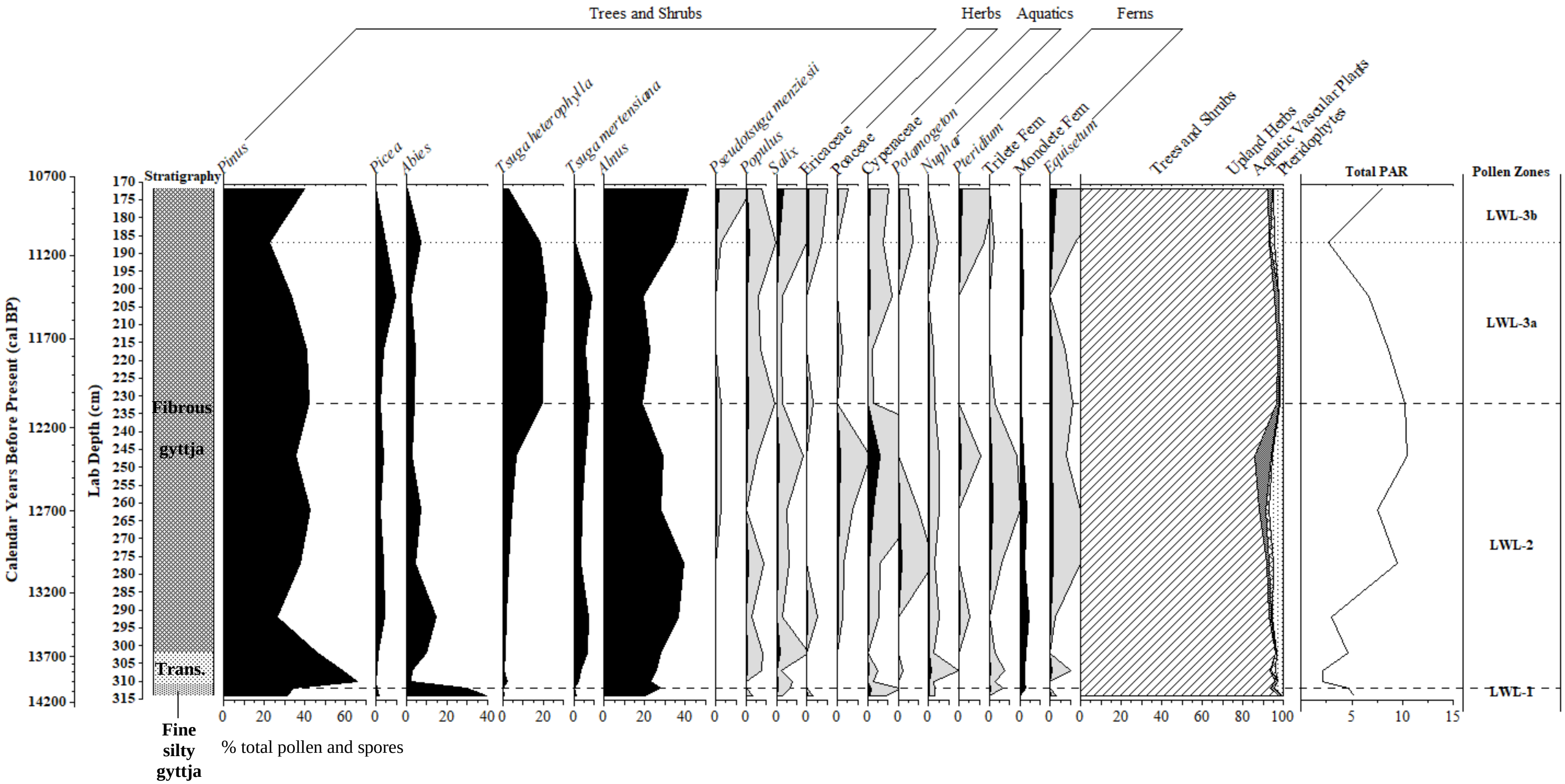
Graminoid, herb, and aquatic taxa are less plentiful in this zone than in LWL-2. Cyperaceae pollen was observed at low levels (0.2% to 1.2%) with other groups appearing in trace amounts (<1%) including Poaceae, Rosaceae, Asteraceae, and *Plantago*. Aquatic vascular plants also only appear in trace amounts (<1%) in this zone, though are represented by several species including *Typha*, *Menyanthes trifoliata*, *Potamogeton*, and *Nuphar*. Pteridophytes, though varied, are uncommon in this zone, including *Pteridium aquilinum* (0% to 1.2%), *Equisetum* (0% to 1.2%), monolete-type ferns (0.7% to 1.8%), and trace amounts (<1%) of *Cryptogramma* and unspecified trilete-type fern spores.

### *LWL-3b: 187-172 cm LD; 11,118-10,744 cal BP*

The uppermost component of the pollen and spore record is similar in composition to LWL3a below, but with notable increases in *Pinus* and *Alnus* as observed in zone LWL-2. As with the other zones, LWL-3b is dominated by tree and shrub species (92.6%), with lesser amounts of pteridophytes (4.8%), herbs (2.0%), and aquatic vascular plants (0.5%). The arboreal assemblage is dominated by *Alnus* (41.3%) followed closely by *Pinus* (39.8%), with limited indications of *Tsuga heterophylla* (2.6%) and *Pseudotsuga menziesii* (1.8%) and trace amounts (<1%) of *Abies*, *T. mertensiana*, Cupressaceae, *Populus*, and *Betula*. *Salix* (3.3%), Ericaceae (1.0%), and trace amounts (<1%) of *Sambucus* represent the shrub taxa in this assemblage.

Cyperaceae pollen (1.0%) is the largest proportion of any herbaceous group in LWL-3b, with trace amounts (<1%) of Poaceae, *Ambrosia*, and *Lysichiton americanus*. Aquatic vascular taxa are even less plentiful, represented only by trace amounts (<1%) of *Potamogeton*. Low incidence of pteridophytes continue through this zone, including *Equisetum* (3.1%) and *Pteridium aquilinum* (1.8%).

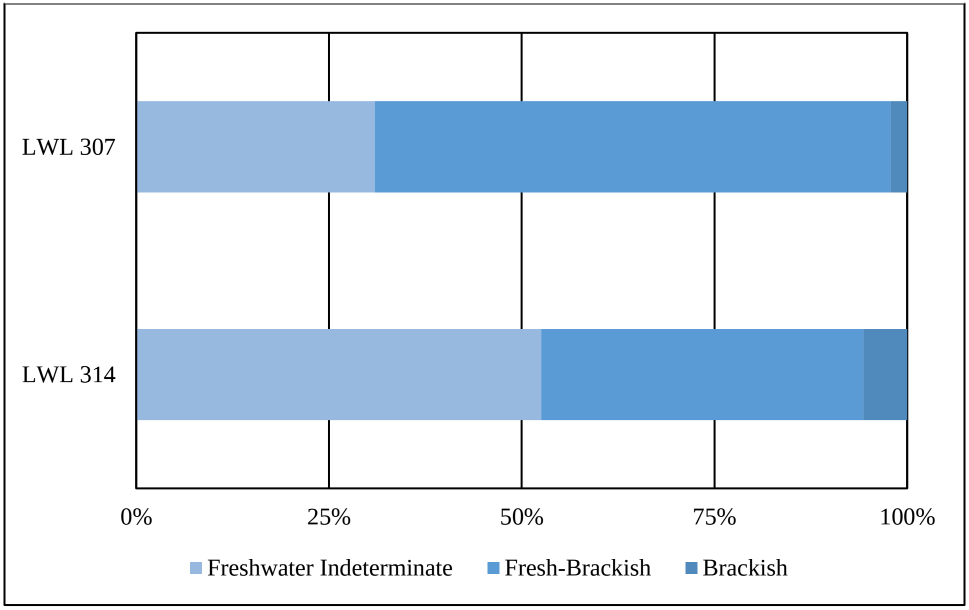
***Figure 19.*** *Percentages of selected pollen and spore taxa at Little Woss Lake. Total pollen accumulation rate (PAR) is included at the right in grains/cm$^2$/cal yr (x1000). A 10x exaggeration has been applied to selected infrequent taxa. Pollen and spore zones were determined by CONISS analysis (Grimm 1987).*



*Diatom Analysis*

As an additional proxy for both relative sea level change and for the composition of the lake water, diatom species distribution was documented across halobian classes (after Hustedt 1953). From this, I inferred water salinity affiliation in the lower portion of the sequence from Little Woss Lake (see Fedje et al. 2018 for a detailed explanation of this method). Diatoms were identified from 307 cm LD (13,877 cal BP) and 314 cm LD (14,137 cal BP) (see Figure 20).

***Figure 20.*** *Percentages of each halobian class as represented in diatom assemblages at Little Woss Lake. The contribution of each class to the assemblage can be used as an indicator of relative salinity in the basin.*



The sample at 307 cm LD was collected from the upper portion of the transition from fibrous gyttja above to the silty gyttja at the base of the core. This sample contained no diatom species that are exclusively found in freshwater (0% halobian class 1), a moderate amount of diatoms that are generally freshwater species (31% halobian class 2), a majority of fresh-brackish diatom species (67% halobian class 3), a small amount of diatoms that are found in brackish water (2% halobian class 4), and no diatom species that are found in marine contexts (0% halobian class 5).

The deepest diatom sample from the Little Woss Lake core was collected at 314 cm LD within fine silty gyttja. This sample contained no diatom species that are exclusively found in freshwater (0% halobian class 1), a majority of generally freshwater diatoms (52% halobian class 2), a significant proportion of fresh-brackish diatoms (42% halobian class 3), a small percentage of diatom species that are found in brackish water (6% halobian class 4), and no diatoms that are found in marine contexts (0% halobian class 5).

### *eDNA Analysis*

Ancient eDNA analysis of the sediments from Little Woss Lake corroborates the presence of species from the pollen record and indicates the presence of animals and other plant taxa not provided by other proxies. The inclusion of ancient eDNA analysis in this study is experimental, and thus the results are limited but useful for palaeoenvironmental reconstruction when combined with other proxies.

Out of the five samples from Little Woss Lake that underwent extraction and amplification to isolate ancient DNA fragments, two samples yielded enough material for successful sequencing. The first of these, collected at 312 cm LD (14,077 cal BP), indicates the presence of numerous plant and animal species in and around Little Woss Lake. The second sequenced sample from Little Woss Lake, collected at 314 cm LD (14,137 cal BP), did not yield plant or animal sequences following bioinformatic quality control procedures.

### *LWL eDNA Sample 1: 312 cm LD (14,077 cal BP)*

Plant taxa detected in this sample include shore pine, black cottonwood (*Populus balsamifera* ssp. *trichocarpa*), yellow pond-lily, pondweed, sedges, and grasses. Animal sequences matched to the sample include Chinook salmon (*Oncorhynchus tshawytscha*), brown bear, Pacific black-legged tick (*Ixodes pacificus*) and mountain pine beetle (*Dendroctonus ponderosae*). Other taxa including possible or confirmed contaminants were detected in the sample but have been omitted from this analysis. A table depicting all Viridiplantae and Metazoa eDNA identifications including contaminants is presented in Appendix C.

### *Other Indicators*

In this study, most of the sampling and rigorous analysis focused on describing late Pleistocene pollen and diatom assemblages and experimenting with the viability of using eDNA in palaeoecological research in the region. However, other micro- and macrofossil indicators were opportunistically recorded in the lower core section from Little Woss Lake.

In addition to the appearance of *Nuphar* pollen grains at low proportions through most of the samples from Little Woss Lake, microfossil trichomes from *Nuphar* appear throughout most of the sequence to a depth of at least 292 cm LD (13,355 cal BP). Combining the pollen and other fossil evidence demonstrates that yellow pond-lily consistently grew in Little Woss Lake through at least the late Pleistocene and early Holocene. Freshwater colonial green algae (*Pediastrum*) was also noted in the early period at 312 cm LD (14,077 cal BP).

Resting eggs belonging to *Daphnia* water fleas were recorded in two samples from the basal sequence of a parallel core from Little Woss Lake (Hiruta and Tochinai 2014). Radiocarbon samples from the same depths on arthropod chitin, terrestrial monocot leaves, and a needle of *Pinus* yielded dates of approximately 13,670 cal BP (11,826 ± 169 $^{14}$C BP, UOC-8924) and 14,050 cal BP (12,170 ± 62 $^{14}$C BP, UOC-8925). These remains indicate that water fleas inhabited the lake in the early period following deglaciation and that pine was growing nearby in the catchment.

### 4.7.2 Topknot Lake

*Site Description*

Topknot Lake (50.563752°, -128.189976°) is located within Quatsino First Nation territory on the outer west coast of northern Vancouver Island, British Columbia, Canada. At 13 m above modern sea level on the rolling Nahwitti Plateau, the lake rests among low hills to the east and south and a small ridge to the west, less than 2 km beyond which lies the open Pacific Ocean. Topknot Lake is approximately 12 km west-northwest of the community of Winter Harbour near the mouth of Quatsino Sound and is 3 km southeast of Raft Cove. The catchment of the lake is relatively small, with runoff from hillsides to the east, west, and south flowing into the sides of the lake and collecting in a wetland which flows into the lake's southern end. Topknot Lake drains to the north, feeding into the Macjack River, and from there into the eastern side of Raft Cove and the Pacific Ocean beyond. The lake measures approximately 1.1 km north-south and 600 m east-west, with an area of 38.5 hectares (GeoBC 2019).

The underlying bedrock geology around Topknot Lake consists largely of Bonanza Group volcanics running north-south along the outer west coast between Cape Scott and the mouth of Quatsino Sound, mostly dark grey-green basaltic to andesitic flows (Howes 1983; Nixon et al. 2011a). Other areas farther inland include some Queen Charlotte Group sandstone and conglomerate formations, and the eastern side of northern Vancouver Island consists largely of dark grey-green Karmutsen Formation basalt flows (Nixon et al. 2011a).

Topknot Lake falls within the Southern Very Wet Hypermaritime subzone of the Coastal Western Hemlock biogeoclimatic ecosystem classification (CWHvh1; BC Forest Analysis and Inventory Branch 2018). This subzone covers most of Nahwitti Plateau, wrapping around the northern tip of Vancouver Island from Winter Harbour to Cape Scott, around Cape Sutil and past Shushartie nearly to Hardy Bay. The low relief of the northern end of the island means that topography has relatively little impact on precipitation and climatic regime in the region compared with elsewhere in coastal BC, and thus vegetation communities are relatively consistent across much of the plateau. The vegetation of the CWHvh1 subzone is characterized by trees including western hemlock, western redcedar, and amabilis fir, with lesser amounts of yellow-cedar and mountain hemlock. Diverse shrubs and herbs cover the understory in this zone including salal, deer fern, red huckleberry, Alaskan blueberry, false azalea, bunchberry, and twinflower, among other species (Green and Klinka 1994). Species observed around the margins of Topknot Lake in May 2018 include western redcedar, Sitka spruce, mountain hemlock, red alder, salmonberry, Labrador tea, sedges, false azalea, salal, and Nootka rose.

### *Chronology and Stratigraphy*

#### Radiocarbon Dates

Twelve radiocarbon samples were successfully dated from the Livingstone core sections collected from Topknot Lake. These samples range in depth from 920 to 509 cm dbp with median calibrated ages from 17,796 to 9,811 cal BP. Despite thorough sampling control, identification, and documentation of $^{14}$C remains, the dates below 701 cm dbp are not all in stratigraphic order. However, all samples collected from this lower part of the core demonstrate ages of at least 16,000 cal BP, and many are much older. The characteristics of these samples and their resulting ages are described in Table 3 below. The dates in the table below were calibrated using CALIB 7.1.0 (Stuiver et al. 2019) with IntCal13, the terrestrial northern hemisphere calibration curve (Reimer et al. 2013).

***Table 3.*** *Radiocarbon assay results from Topknot Lake including sample collection depth, type of material dated, lab ID from UOC (A.E. Lalonde AMS Laboratory) as well as conventional $^{14}$C ages and calibrated age ranges derived from CALIB 7.1.0 using IntCal13 (Stuiver et al. 2019).*

| **Depth (cm dbp)** | **Material** | **Lab ID** | **Radiocarbon age ($^{14}$C BP)** | **Calibrated age range (cal BP, ±2σ)** | **Median calibrated age (cal BP)** |
|---|---|---|---|---|---|
| 509 | matted leaves in fibrous organics and diatomaceous gyttja | UOC-7369 | 8792 ± 31 | 9,922-9,679 (98.0%)<br>10,114-10,087 (2.0%) | 9,811 |
| 526 | *Nuphar* leaf in fibrous organics and diatomaceous gyttja | UOC-8914 | 11252 ± 31 | 13,168-13,055 (100.0%) | 13,108 |
| 538 | *Nuphar* seed and leaf in fibrous organics and diatomaceous gyttja | UOC-8915 | 11565 ± 37 | 13,472-13,305 (100.0%) | 13,401 |
| 553 | monocot leaf fragment in fibrous organics and diatomaceous gyttja | UOC-8916 | 12890 ± 45 | 15,606-15,201 (100.0%) | 15,385 |
| 591 | matted leaves in diatomaceous gyttja | UOC-7368 | 13049 ± 39 | 15,828-15,383 (100.0%) | 15,641 |
| 622 | monocot leaf fragment in diatomaceous gyttja | UOC-7367 | 13232 ± 47 | 16,086-15,727 (100.0%) | 15,906 |
| 701 | monocot leaf fragment in diatomaceous gyttja | UOC-7366 | 14033 ± 43 | 17,264-16,805 (100.0%) | 17,053 |
| 799 | monocot leaf and stem (Cyperaceae) in diatomaceous silts | UOC-8917 | 14536 ± 43 | 17,905-17,549 (100.0%) | 17,717 |
| 799 | monocot leaf and stem (Cyperaceae) in diatomaceous silts | UOC-7365 | 14611 ± 40 | 17,961-17,632 (100.0%) | 17,796 |
| 837.5 | monocot leaf fragment in diatomaceous silts | UOC-7364 | 13350 ± 49 | 16,246-15,866 (100.0%) | 16,064 |
| 865 | monocot leaf (Cyperaceae) in massive fine silts | UOC-8918 | 14154 ± 66 | 17,472-17,003 (100.0%) | 17,232 |
| 920 | monocot leaf fragments and arthropod chitin in massive fine silts | UOC-8920 | 14109 ± 61 | 17,422-16,939 (100.0%) | 17,167 |

### Stratigraphy

In the field, we used a Livingstone corer to sample from 475 to 1125 cm dbp. The platform was suspended approximately 10 cm above the surface of the lake, and the water depth at the coring site was approximately 1 m. All references to depths in the cores from Topknot Lake (cm dbp) are made to the surface of the coring platform. The upper portion of the sediment sequence, 0-360 cm dbp, was collected using a percussion coring method, and 360-475 cm dbp was not collected by either method. These upper sediments were not analyzed as part of this study. See Figure 21.

Stratigraphy for the core analyzed from Topknot Lake consists of the following units in relation to their measured dbp, beginning at the base of the sequence:

(1) **TKL Unit 1, >1125 cm dbp**: The base of the record consists of laminated clayey silts with unknown deposits beneath. Due to limitations of coring tools, sediments below 1125 cm dbp were not recovered or analyzed as part of this study. However, it is likely that deeper sediments exist at the site that could be recovered with additional coring work.

(2) **TKL Unit 2, 1125-1057 cm dbp (68 cm thick)**: This unit consists of laminated light grey fine silts composed of progressively more clay with depth. Dark grey bands in laminations consist of coarser silts at the bottom with a sharp contact and fine upward into light grey silts. Very little organic material was observed in this unit—only a few fragments of moss, aquatic plants, and arthropod chitin were collected using pipette and dissecting scope. Not enough material was recovered for a successful radiocarbon date from this section. Scanning electron microscopy (SEM) imagery reveals that the sediment in this unit consists largely of angular silt and clay particles (Figure 22i).

(3) **TKL Unit 3, 1057-845 cm dbp (212 cm thick)**: This unit consists of massive medium grey fine silts with a lens of laminated fine light grey silt and clay between 1030-1020 cm dbp. Little organic material was found in the lower section of the unit with fragmentary moss, aquatic plants, and arthropod chitin collected from 1057-935 cm dbp. The upper section of the unit contains more organic material, especially from 935-905 cm dbp. Fragments of terrestrial monocot leaves (Cyperaceae), arthropod chitin, mosses, and aquatic plants were collected from 930-920 cm dbp. The deepest successful $^{14}$C sample on fragmentary monocot leaves and arthropod chitin was from 920 cm dbp (UOC-8920, 14109 ± 61 $^{14}$C BP). From 916-905 cm dbp, organics include terrestrial monocot leaves

and stems (sedge), arthropod chitin, and aquatic plants as documented in sediments below, as well as significant assemblage of water flea resting eggs and green algae/stonewort (*Chara*) reproductive oospores (see Figure 22a,b). SEM imagery of sample at 870 cm dbp indicates that the upper component is partially composed of diatom remains (~15%), but no diatoms were visible at 920 cm dbp or below (see Figure 22h,i). A terrestrial monocot leaf (Cyperaceae) macrofossil $^{14}$C sample was collected using forceps at 865 cm dbp (UOC-8918, 14154 ± 66 $^{14}$C BP).

(4) **TKL Unit 4, 845-755 cm dbp (90 cm thick)**: This unit consists of organic light brownish-grey diatomaceous silts. SEM imagery provides evidence for a major diatom fraction in the sediment at 836 cm dbp (up to ~40% of total) and at 797 cm dbp (up to ~75% of total) (see Figure 22c). Three $^{14}$C samples were dated from this component at 837.5 cm dbp (monocot leaf fragment, UOC-7364, 13350 ± 49 $^{14}$C BP) and at 799 cm dbp (two separate samples of Cyperaceae leaf and stem macrofossils, UOC-7365, 14611 ± 40 $^{14}$C BP; and UOC-8917, 14536 ± 43 $^{14}$C BP, see Figure 22d).

(5) **TKL Unit 5, 755-591 cm dbp (164 cm thick)**: This unit consists of silty medium brownish-grey diatomaceous gyttja. SEM imagery and test slides of this unit also provide evidence for major diatom fraction in sediment at 701 cm dbp (up to ~75% or more of total) (see Figure 22e). Three $^{14}$C samples were dated from this component at 701 cm dbp (monocot leaf fragment, UOC-7366, 14033 ± 43 $^{14}$C BP), 622 cm dbp (monocot leaf fragment, UOC-7367, 13232 ± 47 $^{14}$C BP), and 591 cm dbp (matted organics including leaves, UOC-7368, 13049 ± 39 $^{14}$C BP).

(6) **TKL Unit 6, 591-475 cm dbp (116 cm thick)**: This unit consists of matted brown fibrous organic layers interbedded with dark brownish-grey diatomaceous gyttja. Mats of leaves, sedge, yellow pond-lily, and organic debris up to 6 cm thick are contained in a silty matrix. Four $^{14}$C samples were dated from this component at 553 cm dbp (monocot leaf fragment, UOC-8916, 12890 ± 45 $^{14}$C BP), 538 cm dbp (*Nuphar* seed and leaf, UOC-

8915, 11565 ± 37 $^{14}$C BP), 526 cm dbp (*Nuphar* leaf, UOC-8914, 11252 ± 31 $^{14}$C BP), and 509 cm dbp (matted leaves and fibrous organics, UOC-7369, 8792 ± 31 $^{14}$C BP). The radiocarbon samples in this unit cover over 5,000 years of the terminal Pleistocene and early Holocene at Topknot Lake (ca. 15,641-9,811 cal BP) but represent a comparatively small portion of the total recovered sediment (116 cm). The lower units described previously cover 2,000 years (ca. 17,796-15,641 cal BP) but are much more vertically extensive (329 cm).

(7) **TKL Unit 7, <475 cm dbp**: The upper lake sediments were not collected by Livingstone corer. Material from 360-0 cm dbp was collected using a percussion corer but was not analyzed as part of this study. There is a gap between 475-360 cm dbp that was not collected using either method.

Many of the lower samples from Topknot Lake have extremely low concentrations of pollen and diatoms. SEM images taken of these sediments support this, as they show little evidence of organic remains at these depths (Figure 22h,i). Pollen slides from these basal deposits also contain numerous examples of iron-stained and degraded palynomorphs which may have been redeposited by fluvial or aeolian processes. The mineral fraction for this lowest stratigraphic unit is largely composed of silts (3.9-62.5 μm; Wentworth 1922). The clast shape is consistently sub-angular to angular from the lower diatomaceous deposits to the basal minerogenic sediments in Topknot Lake (875 to 1118 cm dbp).

***Figure 21.*** *Diagram depicting stratigraphy of analyzed sediments from Topknot Lake. Calibrated radiocarbon ages are plotted on the left and are based on the results of the age-depth model shown in Figure 23 (Model B). Uncalibrated* $^{14}$*C ages are plotted on the right.*

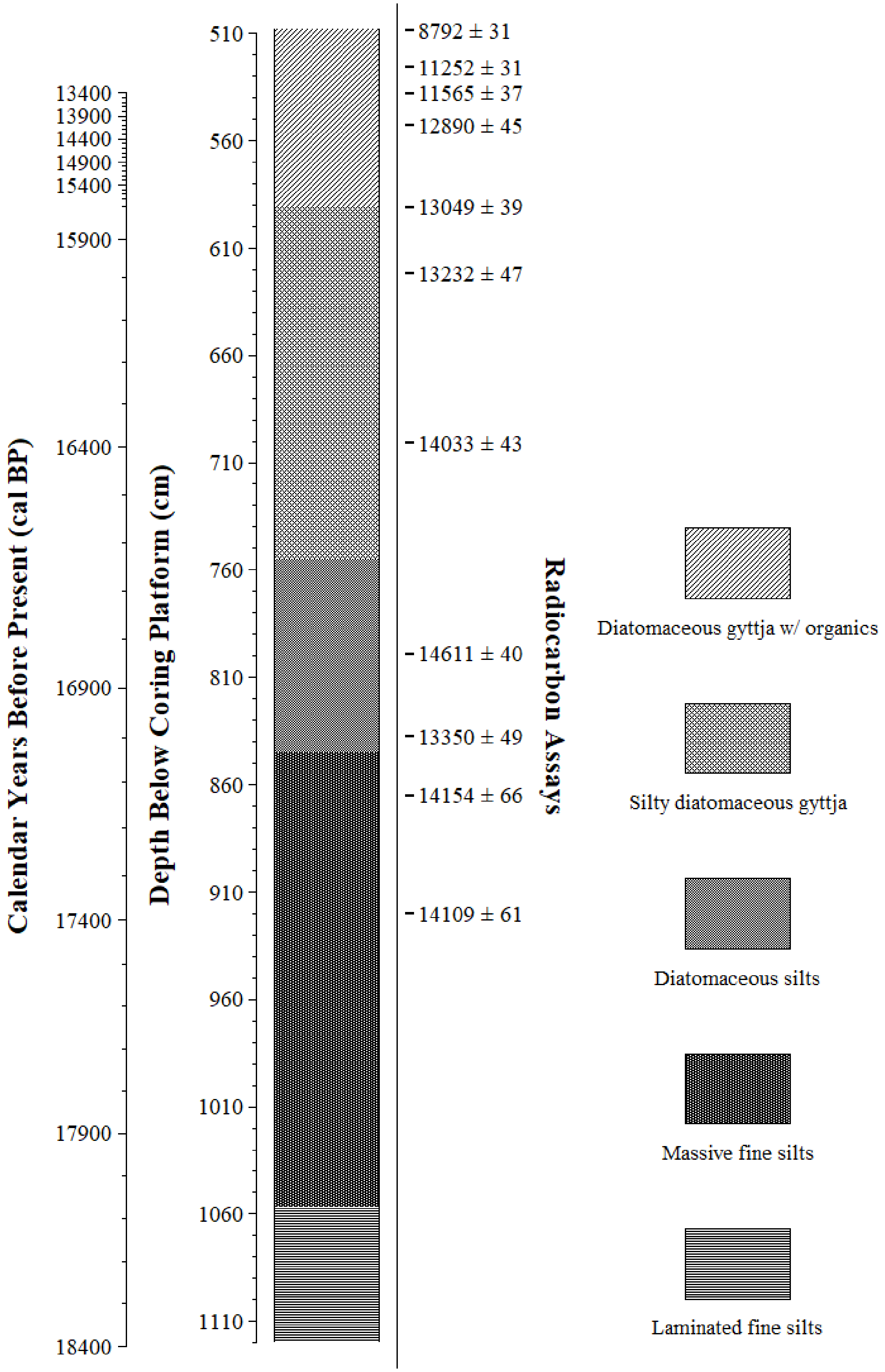
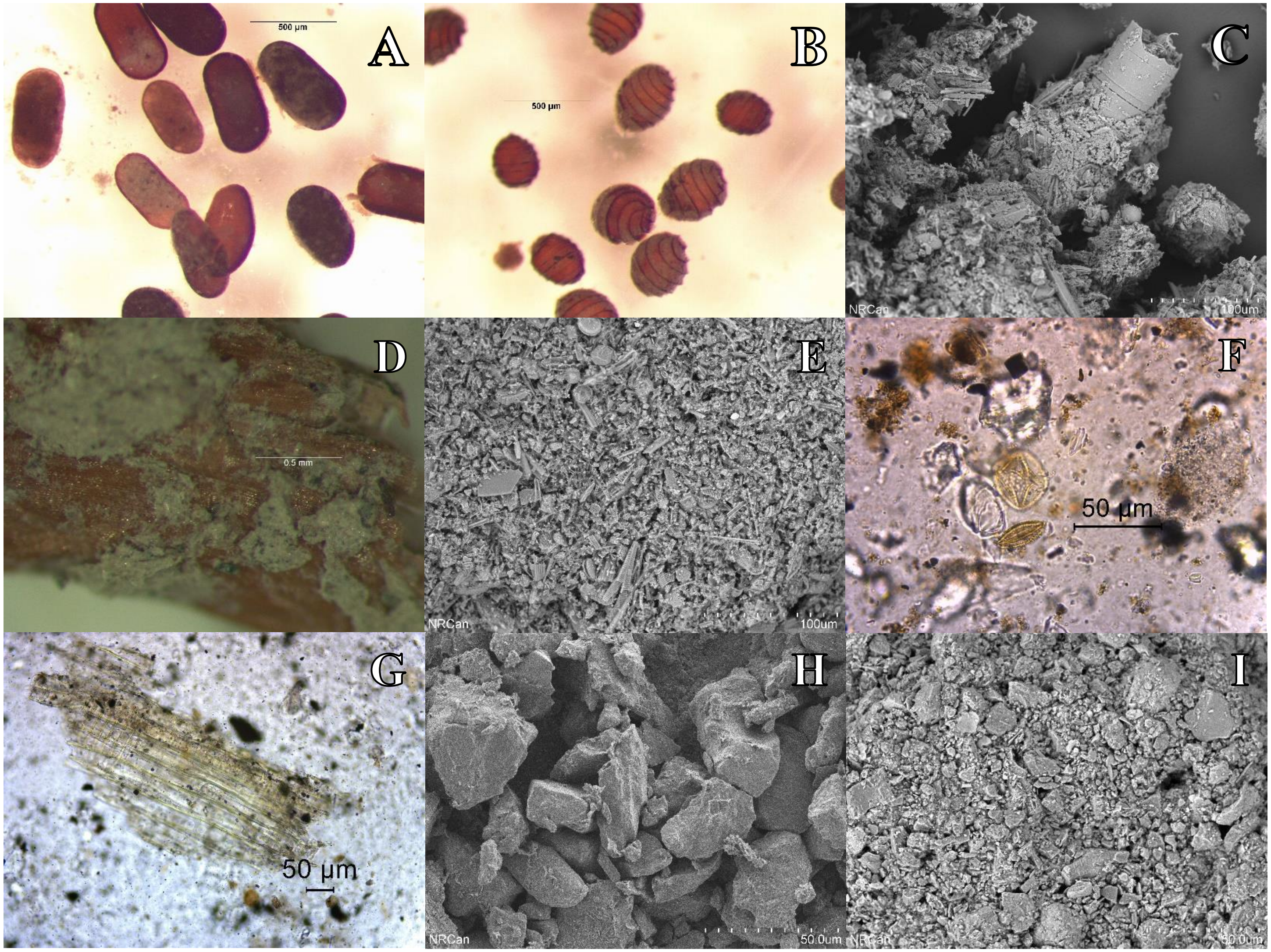


***Figure 22.*** *Microscope images of macro- and microfossil remains from Topknot Lake. Varying scales—see each image.* ***A)*** *Eggs of* Daphnia *(water flea) at 910 cm dbp (17,327 cal BP).* ***B)*** *Reproductive oospores belonging to* Chara *(green algae/stonewort) at 910 cm dbp.* ***C)*** *Example of diatomaceous component of sediment at 836 cm dbp (17,144 cal BP) in silts.* ***D)*** *Monocot leaf fragment (Cyperaceae) collected for* $^{14}$*C dating at 799 cm dbp (16,817 cal BP).* ***E)*** *Example of diatomaceous component of sediment at 701 cm dbp (16,387 cal BP) in gyttja.* ***F)*** *Examples of multiple co-occurring* Salix *(willow) pollen morphotypes at 910 cm dbp, immediately to the left of the scale bar.* ***G)*** *Radial section of* Picea *wood (spruce) preserved on pollen slide at 935 cm dbp (17,448 cal BP).* ***H)*** *Example of inorganic component of sediment at 920 cm dbp (17,373 cal BP) in fine silts.* ***I)*** *Example of inorganic component of sediment near base of core at 1118 cm dbp in laminated fine silts.*



### *Age-Depth Models*

The uncalibrated $^{14}$C dates listed in Table 3 were used to create two continuous age-depth models (Figure 23) for the lower deposits at Topknot Lake using the Bacon software package for R x64 3.6.1 (Blaauw and Christen 2011; R Core Team 2019). Bacon uses the IntCal13 terrestrial northern hemisphere calibration curve in its analysis (Reimer et al. 2013). In order to accommodate the differential accumulation rates apparent in the upper and lower sections of the record, I introduced a boundary in the age-depth model at 591 cm dbp (the base of TKL Unit 6 where it contacts TKL Unit 5 below). This boundary allows Bacon to model separate accumulation rates for the sections above and below 591 cm dbp to more accurately reflect the age of these units. Additionally, the duplicate sample from 799 cm dbp (UOC-8917) was removed from modelling. This sample yielded a confirmatory date for the initial sample from that depth (UOC-7365) but including it as a duplicate would influence the model to output older ages throughout the lower portion of the sequence.

The first model (A) is limited to the portion of the record from Topknot Lake for which radiocarbon dates were run (509-920 cm dbp). The second model (B) extrapolates to the bottom of the record (920-1125 cm dbp) using the depositional priors established for the sediment above. The continuous dates from the second model (B) were used to create the chronology associated with the pollen diagram from Topknot Lake. The age-depth model for this extrapolated basal section of the record in model B is only speculative, but the upper section is equally as robust as model A.

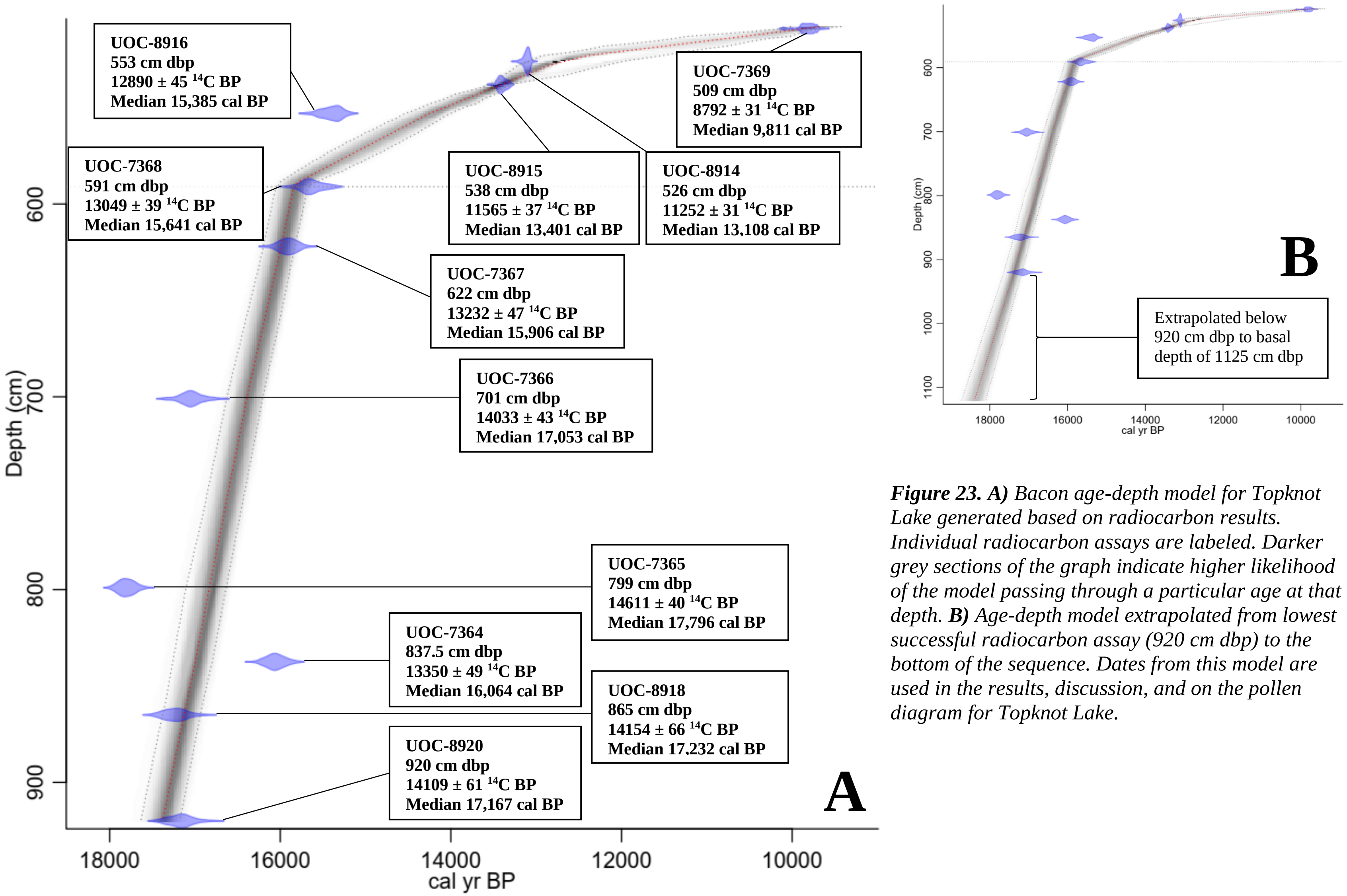


***Figure 23. A)*** *Bacon age-depth model for Topknot Lake generated based on radiocarbon results. Individual radiocarbon assays are labeled. Darker grey sections of the graph indicate higher likelihood of the model passing through a particular age at that depth.* ***B)*** *Age-depth model extrapolated from lowest successful radiocarbon assay (920 cm dbp) to the bottom of the sequence. Dates from this model are used in the results, discussion, and on the pollen diagram for Topknot Lake.*

*Pollen Analysis*

Thirteen pollen samples from Topknot Lake have been divided into three major pollen zones (TKL-1 at the base of the analyzed sequence up to TKL-3 at the top) using CONISS cluster analysis (Grimm 1987). One zone, TKL-2, has been divided into two subzones, TKL-2a and TKL-2b based on having comparable but distinct assemblages. The pollen and spore zones identified in the lake sediments are highly divergent from one another, with >2.0 total sum of squares distance.

Arboreal assemblages (trees and shrubs) dominate two of the zones (TKL-1 and TKL-3), but are each characterized by different major species. The deepest zone, TKL-1 (18,383-17,448 cal BP), is characterized by *Picea* and *Pinus* as well as notable components of *Abies* and Poaceae, with lesser incidence of other herbaceous and pteridophyte taxa. Pollen accumulation rates in zone TKL-1 are extremely low and reflect the inorganic sediment of this unit (see Figure 22h,i), ranging from ~40 to 170 grains/cm$^2$/cal yr. Additionally, iron-stained and highly degraded pollen were noted throughout the samples from TKL-1. These grains may represent redeposited material and were not included in the pollen counts presented in this thesis. In the second-deepest zone, TKL-2a (17,448-17,166 cal BP), coniferous species almost disappear from the record coincident with a marked increase in *Salix* pollen. Herbaceous pollen, especially Poaceae and Cyperaceae, are more abundant in this zone and a greater number of pollen and spore taxa were identified overall. Few aquatics or pteridophytes are apparent in this part of the record in either TKL-2a or the next sub-zone, TKL-2b. Pollen accumulation rates in TKL-2a range from ~425 to 540 grains/cm$^2$/cal yr. Despite low pollen influx the accumulation rate in this zone is nevertheless an order of magnitude greater than the lowest accumulation rates in TKL-1, driven by a notable increase in *Salix* pollen. In TKL-2b (17,166-15,946 cal BP) I observed low values of conifer pollen but noted diverse shrub and herb types at higher incidence than any other part of the record. Poaceae and Cyperaceae predominate but the assemblage includes varied taxa: Ericaceae, Rosaceae, Asteraceae, *Artemisia*, and *Sanguisorba*, among others. Pollen influx increases notably in TKL-2b, ranging from ~1800 grains/cm$^2$/cal yr at the base of the zone to an early peak of ~5100 grains/cm$^2$/cal yr before slowly decreasing upward through the rest of the zone to ~3300 grains/cm$^2$/cal yr. This steady PAR is the result of the continual influx of herbaceous pollen—especially Cyperaceae and Poaceae—throughout TKL-2b. The uppermost zone, TKL-3 (15,946-13,412 cal BP), is characterized by the return of arboreal conditions dominated by *Alnus* and *Pinus*. Little herbaceous pollen is apparent in this zone, but notable increases in pollen and spores from both aquatic vascular plants and pteridophytes were observed. The pollen accumulation rate at the base of TKL-3 (~25,800 grains/cm$^2$/cal yr) is dramatically higher than any of the other samples from Topknot Lake, driven by a marked spike in *Pinus* pollen in this sample as well as a notable increase in monolete-type fern spores. Influx drops again through the rest of this zone, ranging from ~1400 to 2800 grains/cm$^2$/cal yr. These accumulation rates are much lower than at the base of TKL-3 due to much less frequent *Pinus* pollen in the rest of the zone, though the accumulation rate is propped up by increased influx of *Alnus* and *Potamogeton* pollen at the top of the zone. More detailed descriptions of these zones are provided below, and relative pollen and spore contributions are graphed in Figure 24.

### *TKL-1: 1120-935 cm dbp; 18,383-17,448 cal BP*[1]

The basal pollen zone of Topknot Lake is mostly composed of tree and shrub species, ranging from 57.7% to 64.6% of the total. Herbaceous species constitute much of the remainder, ranging from 25.6% to 38.5% of the total pollen counted, with a small proportion of pteridophytes (3.8% to 10.0%). The most common arboreal and shrubby taxa in this zone include *Picea* (15.4% to 29.9%), *Pinus* (6.7% to 15.4%), *Abies* (6.7% to 11.5%), and *Alnus* (3.7% to

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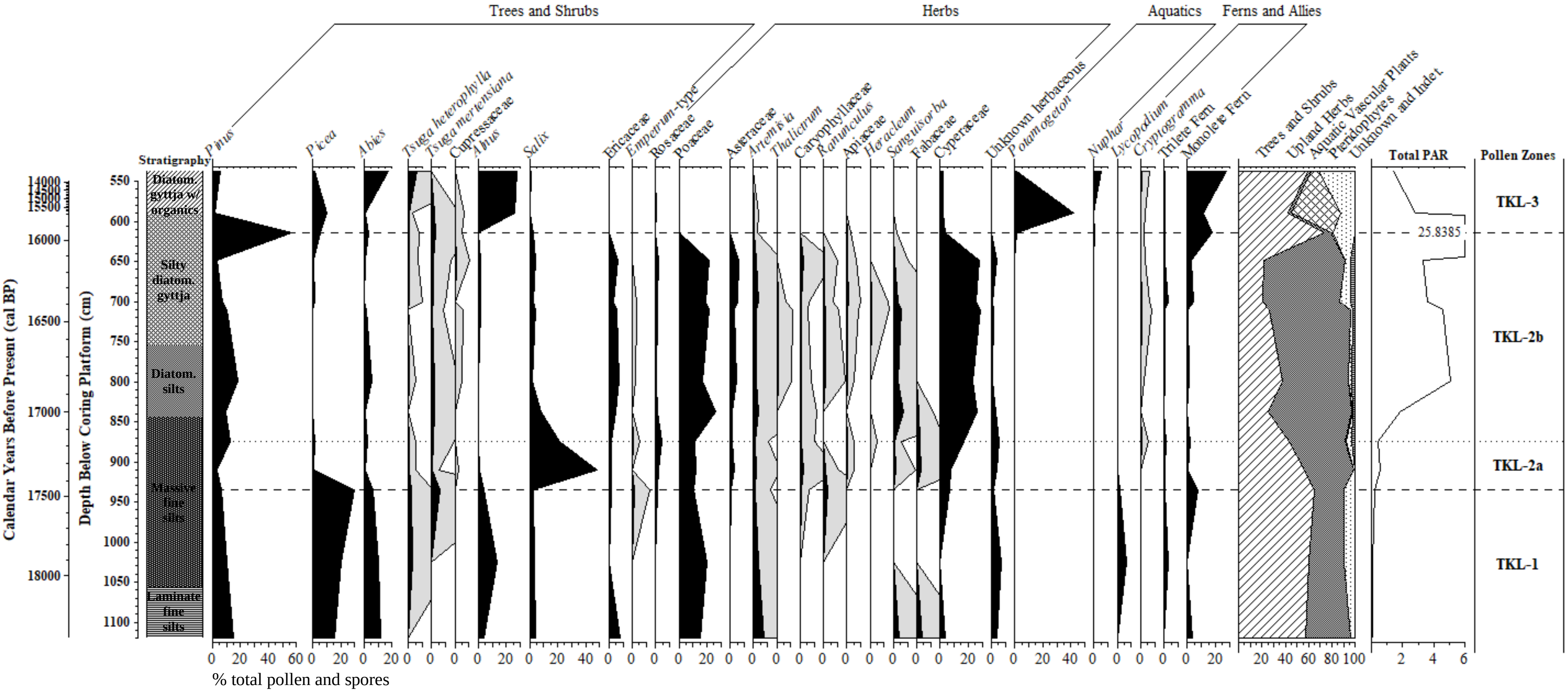
[1] These samples were collected from below the deepest successful $^{14}$C date. Ages in this zone have been extrapolated using the age-depth model described previously based on the deposition rate of sediment above with comparable stratigraphy.

13.3%). Lesser components of *Tsuga mertensiana* (0% to 6.1%) and *T. heterophylla* (0% to 3.3%) were also noted in this zone. Shrubs present in this part of the record include *Salix* (3.0% to 3.8%), Ericaceae (0% to 7.7%), and *Empetrum*-type pollen (0% to 1.2%).

The sizable herbaceous component of this zone is dominated by Poaceae (10.4% to 20.0%) along with *Artemisia* (1.2% to 7.7%) and Cyperaceae (0% to 6.7%) and lesser amounts of forbs such as Fabaceae (0% to 3.8%), *Sanguisorba* (0% to 3.8%), *Ranunculus* (0% to 3.0%), and Rosaceae (0% to 1.2%). Trace amounts (<1%) of Asteraceae and Caryophyllaceae were also observed in this zone. Some pollen grains were identifiable as belonging to herbaceous types (1.8% to 6.7%), but poor preservation precluded more precise classification.

Pteridophytes constitute a relatively small proportion of the pollen identified in TKL-1, but are represented by several groups including monolete-type ferns (0% to 7.3%), *Lycopodium* (0% to 6.7%, not to be confused with the exotic *Lycopodium* spike added to the samples to determine pollen concentrations), and unspecified trilete-type ferns (0% to 3.3%).

***Figure 24.*** *Percentages of selected pollen and spore taxa at Topknot Lake. Total pollen accumulation rate (PAR) is graphed at the right in grains/cm$^2$/cal yr (x1000) for each sample except for the sample at 614 cm dbp, which has been truncated due to scale variability with the deepest, pollen-poor samples. A 10x exaggeration has been applied to selected infrequent taxa. Pollen and spore zones were determined by CONISS analysis (Grimm 1987).*



### *TKL-2a: 935-875 cm dbp; 17,448-17,166 cal BP*

The second-deepest pollen and spore assemblage identified at Topknot Lake is characterized by a precipitous drop in tree pollen, which ranges from 7.3% to 19.6% of the total in this zone. Shrub taxa, largely *Salix*, are much better represented in this zone than any other, ranging from 23.3% to 49.2%. Herbaceous taxa also contribute a higher proportion of pollen than in TKL-1, from 42.7% to 48.1%. Both pteridophytes (0.5% to 4.8%) and aquatic vascular plants (0% to 1.6%) were observed at low amounts in this zone. The comparatively low arboreal contribution to this zone consists mostly of *Pinus* (3.0% to 12.7%) with minor proportions of *Abies* (1.1% to 2.6%), *Picea* (0.8% to 1.6%), *Tsuga mertensiana* (0.5% to 1.6%), and *Alnus* (0% to 1.1%), with trace amounts (<1%) of *T. heterophylla*, *Betula*, and Cupressaceae pollen. The extensive shrub assemblage in this zone is represented by a narrow spread of taxa, including *Salix* (21.2% to 47.8%), Ericaceae (1.4% to 1.6%) and trace amounts (<1%) of *Empetrum*-type pollen. The *Salix* grains observed in this zone belong to numerous morphotypes (see Figure 22f).

Diverse herbaceous pollen types constitute a higher proportion of the assemblage than in zone TKL-1. Cyperaceae (7.3% to 16.9%) and Poaceae (11.6% to 11.8%) form the bulk of the herb component, but many other taxa are also present including Rosaceae (2.5% to 4.8%), Fabaceae (2.1% to 3.2%), Asteraceae (1.1% to 3.2%), Caryophyllaceae (1.1% to 2.7%), Brassicaceae (0.3% to 2.1%), *Artemisia* (1.1% to 1.9%), *Sanguisorba* (0.5% to 1.6%), Liliaceae (0% to 1.3%), and *Ranunculus* (0% to 1.1%), with trace amounts (<1%) of *Polemonium*, Apiaceae, *Heracleum maximum*, *Ambrosia*, *Pedicularis*, *Gentiana*, and Lamiaceae. There was also a high proportion of poorly preserved pollen grains belonging to herbaceous species (3.5% to 5.3%).

I observed fewer pteridophytes in this zone than in TKL-1, though these are also more diverse than below. Monolete-type ferns (0.3% to 2.1%) and unspecified trilete-type ferns (0.3%

to 1.1%) are most frequent, though other ferns and allies are also present in trace amounts (<1%), including *Lycopodium*, *Cryptogramma*, and *Equisetum*. Aquatic vascular plants are only represented in this zone by *Isoëtes* (0% to 1.6%).

### *TKL-2b: 875-614 cm dbp; 17,166-15,946 cal BP*

As TKL-2a below, I observed comparatively low proportions of tree pollen in TKL-2b. However, the spike in shrub pollen driven by *Salix* in TKL-2a is also reduced. Trees and shrubs contribute 20.3% to 37.5% of the total in this zone. Herbaceous taxa are abundant and varied, representing 56.7% to 71.3%. The remainder consists of relatively few pteridophytes (0% to 9.8%) and aquatic vascular plants (0% to 1.1%). The most abundant tree taxon in this zone is *Pinus* (3.4% to 18.2%), with small contributions of *Abies* (0% to 5.5%), *Tsuga mertensiana* (0.8% to 2.3%), *Alnus* (0% to 2.0%), *Picea* (0% to 1.6%), Cupressaceae (0% to 1.0%), and *T. heterophylla* (0% to 1.0%) and trace amounts (<1%) of *Pseudotsuga menziesii* and *Populus*. Compared with TKL-2a, *Salix* contributes less but is still notable in the shrub assemblage (2.1% to 8.0%) and Ericaceae pollen is relatively abundant, occurring in several morphotypes (3.3% to 7.1%). Other shrubs observed in this zone include trace amounts (<1%) of *Shepherdia canadensis*, Caprifoliaceae, *Cornus*, and *Empetrum*-type pollen.

The herbaceous component of zone TKL-2b is exceptionally abundant, reaching 80% of total pollen and spores. Cyperaceae (23.5% to 29.1%) and Poaceae (17.2% to 26.4%) form the bulk of the herbaceous element, but many other taxa are also represented including *Sanguisorba* (1.0% to 6.9%), Asteraceae (1.1% to 5.7%), *Artemisia* (1.9% to 3.6%), Caryophyllaceae (0.5% to 2.4%), Rosaceae (0.5% to 2.4%), *Ranunculus* (0% to 1.6%), *Heracleum maximum* (0% to 1.4%), Fabaceae (0% to 1.1%), *Thalictrum* (0% to 1.1%), Saxifragaceae (0% to 1.0%), *Polemonium* (0% to 1.0%), and Apiaceae (0% to 1.0%), with trace amounts (<1%) of Onagraceae, Brassicaceae, Polygonaceae, and Liliaceae.

Pteridophytes occur infrequently in TKL-2b and are mostly represented by monolete-type spores (0% to 4.6%) and unspecified trilete-type spores (0% to 2.6%) with a smaller proportion of *Equisetum* (0% to 1.6%) and trace amounts (<1%) of *Cryptogramma*, *Pteridium aquilinum*, and *Selaginella* spores. Aquatic vascular plants were rarely observed, including *Typha* (0% to 1.1%) and trace amounts (<1%) of *Triglochin*. The only aquatic bryophyte recorded in this zone was a trace amount (<1%) of *Sphagnum* spores.

### *TKL-3: 614-538 cm dbp; 15,946-13,412 cal BP*

TKL-3 demonstrates a marked shift in assemblage composition from the preceding zones. The arboreal component in TKL-3 increases sharply, with a modest increase in pteridophytes and a marked decrease in herbaceous pollen. Trees and shrubs contribute 42.0% to 73.2% of the total in this zone, an increase of up to 50% from the zones below. Conversely, herbaceous species form only 2.8% to 5.7% of the total pollen, a respective decrease of over 50%. Pteridophyte spores increase to between 12.7% and 31.3% of the total. Aquatic vascular plants see a small increase in most of the samples from TKL-3 (2.3% to 6.5% of the total) except for a spike in the middle of this zone (to 42.0% of the total) driven by the overwhelming presence of *Potamogeton* pollen in the sample at 589 cm dbp. The tree assemblage at the base of this zone is characterized by the dominance of *Pinus* pollen (56.5%), dropping to between 1.9% and 5.9% in the upper portion as it is replaced by *Alnus* (0% at the base of the zone and increasing to between 26.1% and 27.6% in the upper portion) and other taxa including *Abies* (1.3% to 17.3%), *Picea* (1.2% to 9.9%), *Tsuga heterophylla* (0.3% to 6.2%), and *T. mertensiana* (0% to 2.6%), with trace amounts (<1%) of Cupressaceae, *Populus*, and *Pseudotsuga menziesii* pollen. Shrub species contribute little pollen in this zone and are only represented by *Salix* (0% to 2.3%).

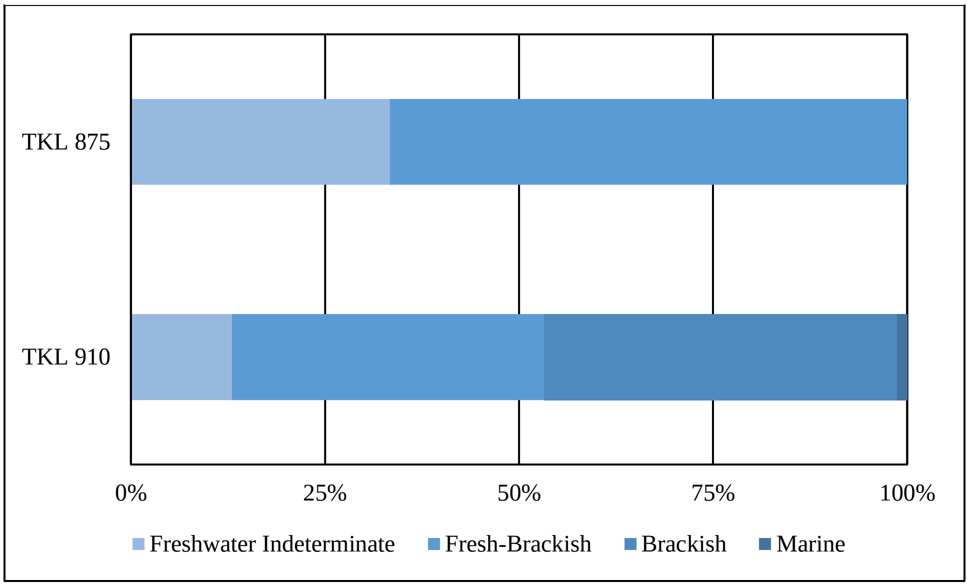
The marked increase in arboreal pollen in this zone is paired with an even greater decrease in herbaceous pollen. Cyperaceae is still present in low proportion (1.9% to 3.9%), but other herbs only occur in trace amounts (<1%), including Poaceae, Rosaceae, *Artemisia*, Liliaceae, Apiaceae, and *Sanguisorba*.

Spores derived from pteridophytes are much more plentiful in this zone, including monolete-type spores (11.5% to 27.5%), *Pteridium aquilinum* (0% to 1.2%), and trace amounts (<1%) of *Cryptogramma*, *Equisetum*, and unspecified trilete-type spores. Aquatic vascular plants were observed at somewhat higher proportions through most of this zone, including *Nuphar* (0% to 5.3%), *Potamogeton* (1.2% to 1.3%) and trace amounts of *Typha*. However, as mentioned earlier, I recorded an exceptional spike in *Potamogeton* up to 42.0% of the total pollen in the sample at 589 cm dbp (15,783 cal BP). This peak is not coincident with a marked increase in the pollen or spores of other aquatic plants in the sample.

### *Diatom Analysis*

As with Little Woss Lake, I used the proportions of diatom species by halobian class as an additional proxy for relative sea level change and for the composition/inferred salinity of the water in Topknot Lake (see Fedje et al. 2018 for a detailed explanation of this method). Diatoms were examined at 875 cm dbp (17,166 cal BP) and 910 cm dbp (17,327 cal BP) (see Figure 25). Slides were also prepared for analysis at 1005 cm dbp and 1115 cm dbp, but neither yielded any diatom remains.

***Figure 25.*** *Percentages of each halobian class as represented in diatom assemblages at Topknot Lake. The contribution of each class to the assemblage can be used as an indicator of relative salinity in the basin.*



The sample at 875 cm dbp was taken the upper portion of the large unit of massive fine silts and is coincident with both a pollen sample and the successful eDNA sample from Topknot Lake. This sample contained no diatom species that are exclusively found in freshwater (0% halobian class 1), a moderate amount of diatoms that are generally freshwater species (33% halobian class 2), a majority of fresh-brackish diatom species (67% halobian class 3), and no diatom species that are found exclusively in brackish or marine contexts (0% halobian classes 4 or 5).

The sample at 910 cm dbp was taken from within the same massive fine silts, but lower in the unit. This sample contained no diatom species that are exclusively found in freshwater (0% halobian class 1), a small number of generally freshwater diatoms (13% halobian class 2), a significant proportion of fresh-brackish diatoms (40% halobian class 3), an even larger percentage of diatom species that are found in brackish water (45% halobian class 4), and exactly 1 broken diatom specimen usually found in marine contexts (1% halobian class 5).

### *eDNA Analysis*

Ancient eDNA analysis of the sediments from Topknot Lake corroborates the presence of species from the pollen record and indicates the presence of animals and other plant taxa not provided by other proxies. The inclusion of ancient eDNA analysis in this study is experimental, and thus the results are limited but useful.

Out of the nine samples from Topknot Lake that underwent extraction and amplification to isolate ancient DNA fragments, only one sample yielded enough material for successful sequencing. This sample, collected from 875 cm dbp (17,166 cal BP), indicates the presence of numerous plant taxa and limited animal species in and around Topknot Lake.

### *TKL eDNA Sample 1: 875 cm dbp (17,166 cal BP)*

Plants in this sample include sequences matched with true fir, spruce, and general pine family species (Pinaceae—possibly pine, spruce, or fir), black cottonwood, willow, northern Jacob's ladder (*Polemonium boreale*), bistort (*Bistorta*), asters, and unspecific berries (*Rubus*). A single non-contaminant animal species matched with this sample—mountain pine beetle. Other taxa including possible or confirmed contaminants were detected in the sample but have been omitted from this analysis. A table depicting all Viridiplantae and Metazoa eDNA identifications including contaminants is presented in Appendix C.

### *Other Indicators*

My analysis of the material from Topknot Lake focused on producing extensive pollen and diatom assemblages as well as testing the viability of ancient eDNA analysis for future palaeoecological studies in the region. Other micro- and macrofossil remains were also recorded throughout the core from Topknot Lake.

I observed *Nuphar* trichomes in three of the uppermost pollen samples, including at 538, 589, and 614 cm dbp. This supports the pollen evidence for *Nuphar* in two of these samples (538 and 614 cm dbp) and extends its presence to the remaining sample in zone TKL-2b. *Nuphar* trichomes were also observed in the sample from 910 cm dbp (17,327 cal BP) near the base of zone TKL-2a, but not in any of the intervening samples. Remains of freshwater colonial green algae including *Pediastrum* were observed throughout most of the sequence at Topknot Lake, from the uppermost sample at 538 cm dbp down to 837.5 cm dbp (13,412 to 16,983 cal BP).

Arthropod remains observed in pollen samples were not identified, but mouth parts, body segments, and eggs were noted throughout the lower samples from Topknot Lake, from 710 to 1120 cm dbp (16,427 to 18,383[1] cal BP). These remains are corroborated by observations of these lower sediments under dissecting microscope during $^{14}C$ sampling. Mouth parts, body sections, and other chitinous fragments were observed in material ranging from 910 to 1105 cm dbp (17,327 to 18,309 cal BP, extrapolated). At 910 cm dbp, sediment examined under the dissecting microscope contained a major fraction of *Daphnia* (water flea) resting eggs (Hiruta and Tochinai 2014) as well as *Chara* oospore reproductive bodies (green algae/stonewort) (see Figure 22a,b).

Deeper still, on a pollen slide from 935 cm dbp, I observed a fragment of wood in radial section (Figure 22g). Based on the presence of small pit apertures, resin canals, transversal tracheids, and absent spiral thickenings in the tracheid walls, I believe the fragment to be derived from *in situ Picea* in the catchment of Topknot Lake at 17,450 cal BP.

## 4.8 Discussion

### 4.8.1 Late Pleistocene Glacial History of Northern Vancouver Island

The radiocarbon dates and stratigraphy observed at Little Woss Lake and Topknot Lake demonstrate different glacial histories for these two regions, likely driven by factors including proximity to ice accumulation centres, local physiography, and the temperature-regulating effect of the Pacific Ocean, among others.

The record described here for Little Woss Lake demonstrates that deglaciation of Woss Lake and the Nimpkish River Valley occurred sometime prior to ca. 14,300 cal BP. This chronology is consistent with the basal post-glacial dates of most other areas on northern Vancouver Island including Misty Lake (14,900 cal BP), Brooks Peninsula (14,500 cal BP), Pellucidar Cave (14,500 cal BP), and Port Eliza Cave (14,350 cal BP) (Hebda 1997; Lacourse 2005; Al-Suwaidi et al. 2006; Steffen and Fulton 2018).

The oldest previously documented deglacial age on northern Vancouver Island is from the basal deposits at Bear Cove Bog, dating to ca. 16,450 cal BP (Hebda 1983). The record from Topknot Lake, however, extends back much further, to at least 18,000-17,500 cal BP—some 3,000 years older than most sites on northern Vancouver Island and at least 1,000 years older than the basal deposits at Bear Cove Bog. The age of the sediments at both Topknot Lake and Bear Cove Bog may be explained by differential deglacial processes, ice flow, and local topography in coastal environments when compared with those in the Vancouver Island Ranges and Coast Mountains.

### 1. *Differential Deglacial Processes*

The earliest areas from which ice retreated at the end of the Fraser Glaciation were distant from the sources of Cordilleran ice or were located on refugial plains. These include parts of Haida Gwaii (Cape Ball, 18,600 cal BP) and the continental shelf off the central coast of BC (Explorer Ridge, 18,850 cal BP) (Warner et al. 1982; Blaise et al. 1990). Many other records up and down the coast indicate deglacial processes beginning somewhat later, generally after ca. 17,000 cal BP among the islands, bays, and major waterways of exposed coastal environments.

Glacial retreat is documented after this time in southern Alaska (Sanak Island, 17,000 cal BP; Copper River Delta, 16,700 cal BP), southeastern Alaska (Alexander Archipelago, 17,000 cal BP), Haida Gwaii (Dogfish Bank, 17,365 cal BP; West Side Pond, 16,890 cal BP; western Dixon Entrance, 16,700 cal BP; eastern Dixon Entrance, 15,800 cal BP), and the central coast (eastern Queen Charlotte Sound, 16,400 cal BP; Calvert Island, 16,600 cal BP) (Luternauer et al. 1989a; Barrie and Conway 1999; Davies et al. 2011; Lacourse et al. 2005; Addison et al. 2012; Misarti et al. 2012; Darvill et al. 2018; Lesnek et al. 2018). Though deglaciated relatively early, the climate and ecology of these areas would still have been influenced after ca. 17,000 cal BP by the cold air of major ice source areas including the Alaska Range, St. Elias Mountains, and Coast Mountains (Hicock et al. 1982; Hebda and Whitlock 1997).

These early offshore dates contrast with much later dates that represent the retreat of glaciers onto land and into mountainous regions across the Pacific Coast Ranges, including in the Gulf of Alaska (14,790 cal BP), interior portions of the Alexander Archipelago (15,300-13,600 cal BP), and parts of Haida Gwaii (SG̱ang Gwaay, 14,350 cal BP) and the central coast (Seymour Inlet, 13,800 cal BP) (Hansen and Engstrom 1996; Hebda et al. 2005; Galloway et al. 2008; Davies et al. 2011; Briner et al. 2017; Lesnek et al. 2018; Ager 2019). The later interior dates correspond closely to the basal postglacial dates from northern Vancouver Island noted above. This deglacial window from ca. 15,300-14,000 cal BP is probably the result of global climatic warming during the Bølling-Allerød coupled with thicker glacial cover up to 2000 m over central parts of the ice sheet as well as top-down decay of mountain glaciers and subsequent lingering dead ice masses in valley bottoms (Clague and James 2002). These factors would have affected areas closer to the interior to a greater degree than thinner calving sea ice or coastal glaciers at the margin of the Cordilleran ice, which may have melted earlier.

On Vancouver Island, areas farthest from the glaciers emerging from mainland inlets would have been covered by less ice volume, if they were covered at all. Outer coastal sites like Topknot Lake would have been glaciated for a shorter period and would have undergone ice mass decay first, followed by inner coastal sites like Bear Cove Bog and Misty Lake, and finally by sites in the fjords and valleys themselves. Additionally, coastal sites would not have been as susceptible to the long-term postglacial presence of dead ice masses as those in interior mountain valleys.

### *2. Ice Flow*

In addition to differential deglacial chronologies based on thickness and proximity to interior ice systems, the direction and character of ice flow would also have affected the temporal distribution of glacial maxima on the coast. On land, glacier flow is generally confined to existing topography including mountain icefields and river valleys (Clague and James 2002). Eventually as ice pushes to the sea it grinds upon but remains constrained by coastal fjords and channels. However, as it spreads into open waters the advancing ice is no longer limited by valley slopes and forms broad piedmont lobes that vary in thickness and flow, shaped by island and sea floor topography. The occurrence of these processes has been demonstrated for portions of Haida Gwaii during the LLGM (Barrie and Conway 1999; Shaw et al. 2019). As mainland glaciers grew during the early part of the Fraser Glaciation, encroaching tongues of Cordilleran ice emerging from the Nass and Skeena rivers met and coalesced with local ice from Graham Island before being deflected into two directions. One part of the lobe flowed northwest into deeper waters of Dixon Entrance and along the northern edge of Haida Gwaii, and a second part of the lobe flowed south through Hecate Strait along the eastern side of the then-exposed Hecate Lowlands (Barrie and Conway 1999; Mathewes and Clague 2017; Shaw et al. 2019). These flow patterns meant that thinner local ice rather than thick Cordilleran ice covered much of Haida Gwaii and parts of Hecate Strait, potentially allowing for the maintenance of refugia in areas such as the Hecate

Lowlands and early deglaciation elsewhere nearby such as at Cape Ball and Dogfish Bank even while other parts of the archipelago (e.g. SG̱ang Gwaay) remained covered (Warner et al. 1982; Heusser 1989; Hebda et al. 2005; Lacourse et al. 2005; Mathewes and Clague 2017; Shaw et al. 2019). The second tongue of Cordilleran ice, the Hecate Outlet Glacier, continued south before turning to the west into Moresby Trough, parallel with other ice tongues flowing across Queen Charlotte Sound into Mitchell's Trough and Goose Island Trough (Shaw et al. 2019). These troughs have raised banks between them that would have been subaerial as a result of isostatic forebulge effects during and following the local glacial maximum (Clague 1983; Shugar et al. 2014). Retreat along the flow lines from these maxima would have rapidly exposed the inter-trough banks as well as the broad Hecate Lowlands in the postglacial period.

On Vancouver Island, ice emerging from the fjords of Knight Inlet, Kingcome Inlet, and other smaller channels eventually flowed west into Queen Charlotte Strait, possibly meeting with local ice from the Vancouver Island Ranges at the mouth of Nimpkish Lake or on the eastern side of Quatsino Sound. The path of least resistance for these growing glaciers would be to flow to the northwest into Queen Charlotte Sound and to the open Pacific Ocean, as well as to begin to flow west across the northern end of Vancouver Island (Howes 1983). As climatic conditions ameliorated and ice began to retreat, it would have retreated first from higher elevations with thin ice as a result of downwasting and then along dominant flow lines up the channels and fjords (Clague and James 2002). This pattern would lead to sites such as Topknot Lake being exposed early (ca. 18,000-17,500 cal BP) by the retreating flow, if it was covered at all, followed by Bear Cove Bog (16,450 cal BP), then Misty Lake and Little Woss Lake (ca. 15,000-14,000 cal BP) as the ice continued to regress toward local centres of accumulation and the mainland fjords (Hebda 1983; Lacourse 2005).

### 3. *Local Topography*

Differences in physiography may also have led to differential glacial cover and retreat on northern Vancouver Island during the Fraser Glaciation. We might expect outer coastal sites that were distant from the main sources of Cordilleran ice to have undergone deglaciation sooner than interior sites. Given this, west coast areas like Brooks Peninsula (basal date of 14,500 cal BP) and Port Eliza Cave (postglacial basal date of 14,350 cal BP) should have significantly earlier basal dates than more glacio-proximal sites like Misty Lake (basal date of 14,900 cal BP), but they are relatively close in age (Hebda 1997; Lacourse 2005; Al-Suwaidi et al. 2006). Both Brooks Peninsula and Port Eliza Cave are located at the edge of the northern Vancouver Island Ranges and are heavily influenced by the effects of this mountainous physiography, which likely produced local ice near these sites. Brooks Peninsula is covered with high mountain peaks and ridges, several of which include north-facing cirques that may have incubated small glaciers separate from the main Cordilleran ice sheet during the LLGM (Howes 1997). The area around Port Eliza Cave may have been similarly influenced by local ice emerging from the west coast fjords of the Vancouver Island Ranges (Ward et al. 2003; Al-Suwaidi et al. 2006). As climate ameliorated during the global Bølling-Allerød warming, Cordilleran ice may have retreated from areas influenced by mainland fjords (Misty Lake) at the same time as areas that had been more influenced by local ice and mountainous topography (Little Woss Lake, Pellucidar Cave), even those nearer to the coast (Brooks Peninsula, Port Eliza Cave) (Hebda 1997; Lacourse 2005; Al-Suwaidi et al. 2006; Steffen and Fulton 2018).

Though not far to the north, sites like Topknot Lake and Bear Cove Bog may have conversely escaped such lengthy glaciation due to the subdued topography of the Nahwitti Plateau (Holland 1976). Without a source of local ice and being more distant from valley axis glaciers and the Cordilleran advance, both sites may have been covered more thinly and for less time during the coastal ice maximum.

### 4.8.2 Late Pleistocene Palaeoecology of Northern Vancouver Island

The earliest pollen records following the LLGM of the Fraser Glaciation across the northwest coast of North America consist of non-arboreal plant communities generally dominated by dry- and cold-adapted species including sedges, grasses, dwarf willow, sagewort, and diverse herbaceous taxa. Such assemblages appear to indicate dry periglacial environments on parts of Haida Gwaii (Cape Ball, Dogfish Bank, Langara Island, and West Side Pond ca. 17,000-15,900 cal BP; later Hippa Island, ca. 14,000 cal BP) and southern Alaska (Sanak Island, 16,300 cal BP) (Warner 1984; Barrie et al. 1993; Lacourse et al. 2003; Lacourse et al. 2005; Lacourse et al. 2012; Misarti et al. 2012). These early cold herb-shrub tundra components are comparable to numerous GLGM and deglacial Beringian records to the north and west in what are now Alaska and far eastern Russia (Colinvaux 1964; Rampton 1971; Ager 1975; West 1981; Anderson and Brubaker 1996; Colinvaux 1996; Savvinova et al. 1996) as well as pre-glacial (>19,500 cal BP) records at Port Eliza Cave on northern Vancouver Island (Al-Suwaidi et al. 2006) and at several sites on southeastern Vancouver Island (Miskelly 2012).

However, most late Pleistocene pollen records on the coast demonstrate pioneering communities of shore pine often accompanied by alder and mixed fern assemblages. These are inferred to represent open woodland environments during a relatively cool and dry climate (Hebda 1995). Such assemblages have been documented in southeastern Alaska (Cwynar 1990; Hansen and Engstrom 1996; Ager 2019), on parts of southern Haida Gwaii (Hebda et al. 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Mathewes et al. 2019), the north and central coast of mainland BC (Lacourse et al. 2003; Galloway et al. 2008; McLaren 2008; Eamer 2015), and northern and southern Vancouver Island (Hebda 1983; Hebda 1997; Brown and Hebda 2002; Lacourse 2005). Most of these pine woodland assemblages correspond to the ca. 15,300-14,000 cal BP deglaciation window described previously. The earliest assemblages from Topknot Lake (herb-shrub tundra or parkland) and Little Woss Lake (open conifer woodland) correspond generally to these two classes of pioneering vegetation communities.

### *Interior and Leeward Coastal Sites – Little Woss Lake*

Broadly speaking, the basal pollen zone at Little Woss Lake (LWL-1, ca. 14,150 cal BP) is comparable to other pollen records across the Pacific northwest coast that represent dry and cool conifer woodlands (Hebda 1983; Hebda and Whitlock 1997; Brown and Hebda 2002; Lacourse 2005; Eamer 2015; Ager 2019). At Little Woss Lake, the basal zone is dominated by fir, shore pine, and alder, with few herb and shrub species including <3% monolete-type ferns, <2% sedges, and <1% willow and heathers. The early predominance of firs (peak 39.4% of total pollen) rather than by pine is unique and may be explained by the physiography of the area around Woss Lake and the Nimpkish River Valley. The north-south orientation of Nimpkish Lake and Woss Lake may have created a funnel for cold and dry anticyclonic continental winds blowing from the Cordilleran and Laurentide ice to the north and east and from local ice sourced in the Vancouver Island Ranges (Hebda and Whitlock 1997). Even today, Little Woss Lake experiences a comparatively drier interior-like climate than the wet, cool climate often associated with coastal British Columbia (CWHxm2—the Western Very Dry Maritime Subzone of the Coastal Western Hemlock biogeoclimatic ecosystem classification; BC Forest Analysis and Inventory Branch 2018). This local interior orientation is represented in the modern environment by the presence of Douglas-fir, a species whose range on Vancouver Island is now generally limited to drier, warmer areas to the south (Pojar and MacKinnon 2004). Under such conditions, *Abies*-type pollen in the late Pleistocene assemblage may have been derived from dry-adapted subalpine fir (*Abies lasiocarpa*) or possibly grand fir (*A. grandis*) rather than amabilis fir (*A. amabilis*), which generally favours wetter environments (Heinrichs et al. 2002; Mazzucchi 2010). Today, subalpine fir is known to grow near sea level in the drainages of major rivers within colder airflows and grand fir prefers dry to moist coniferous forest in rainshadow areas (Pojar and

MacKinnon 2004). These conditions may be analogous to the early postglacial climate at Little Woss Lake. As other pine zone assemblages on northern Vancouver Island are derived from near-coastal sites where temperatures are regulated by the nearby Pacific Ocean (Bear Cove Bog, Misty Lake, Brooks Peninsula), this brief *Abies* zone at Little Woss Lake could represent a late Pleistocene interior-coastal assemblage not yet documented in British Columbia. Additional studies focusing on interior lakes on Vancouver Island may reveal similar cold- and dry-adapted assemblages.

Numerous water flea (*Daphnia*) resting eggs were observed coincident with the early dry assemblage documented in zone LWL-1 and extending into LWL-2 (Hiruta and Tochinai 2014). High incidence of these species is often indicative of eutrophic basins during the early stages of development, suggesting ample postglacial nutrient load in the lake during this early period (Fredskild 1983).

Between LWL-1 and LWL-2 (14,077 cal BP), taxa identified using ancient eDNA also indicate partially open environments around Little Woss Lake at this time. These analyses demonstrate the presence of both woodland taxa such as shore pine (with associated mountain pine beetle) and grasses as well as lake-marginal species including black cottonwood, yellow pond-lily, pondweed, and sedges. Furthermore, this record indicates animal species within or proximal to the basin, including Chinook salmon and brown bear (with possible associated Pacific black-legged tick).

Following this brief basal zone, the next period at Little Woss Lake (ca. 14,000-12,000 cal BP) is composed of typical late Pleistocene open woodlands of shore pine (peak 65.4% of total pollen) and alder (peak 39.0% of total pollen) on Vancouver Island (Hebda 1983; Hebda 1997; Brown and Hebda 2002; Lacourse 2005). Fir pollen values drop precipitously in this period along with the appearance of small amounts of spruce, mountain hemlock, and some western hemlock, all of which are adapted to warmer and moister conditions (Hebda 1983; Hebda and Whitlock 1997). Non-arboreal taxa also increase, filling the meadows and lakesides with herbaceous species including grasses, sedges, asters, rose-family, and sagewort as well as diverse species of ferns and their allies including bracken fern (*Pteridium aquilinum*), parsley ferns (*Cryptogramma*), and horsetail (*Equisetum*), along with other polypodiate fern taxa. Diverse ferns as well as dry-adapted species like sagewort in this component suggest that while Woss Lake and the Nimpkish River Valley were wetter through this period than in the earlier interval, the climate was still relatively dry.

Through the cooler and moister Younger Dryas chronozone and into the early Holocene, thick mixed-conifer forests expand in the record at Little Woss Lake (ca. 12,000-11,100 cal BP). Shore pine and alder were still abundant, though with increased proportions of shade-tolerant taxa like western hemlock (peak 22.0% of total pollen), mountain hemlock (peak 8.6% of total pollen), and spruce (peak 9.8% of total pollen). Similar mixed-conifer vegetation reconstructions for this period have been published from other sites on northern Vancouver Island, including at Bear Cove Bog and Misty Lake (Hebda 1983; Lacourse 2005). The associated cool and moist climate appears to have prevailed into the early Holocene. Mathewes (1993) notes that increases in moisture-dependent taxa, especially mountain hemlock, are characteristic of the Younger Dryas on the north Pacific Coast. However, Mathewes (1993) also notes that Younger Dryas assemblages are generally accompanied by thinning forests and increases in non-arboreal pollen, conditions which were not observed in the record at Little Woss Lake. Instead, herb and shrub species that depend on drier conditions and gaps in the canopy were less common through this period. However, the lakeshore at this time was still covered in sedges and grasses as well as some other herbs (aster family and rose family), horsetail, bracken fern, and parsley fern. The uppermost subzone, LWL-3b, reveals the disappearance of mountain hemlock and marked increase in alder, conditions which are typically associated with climate warming during the early Holocene (Hebda and Whitlock 1997). Additional samples from the upper sediments at Little Woss Lake would further refine the reconstruction of Holocene vegetation and chronology.

In addition to the pollen and macrofossil record, both diatom samples analyzed from the base of the core from Little Woss Lake (307 cm LD and 314 cm LD) contain high proportions of freshwater or fresh-brackish diatom species, and no marine species were observed at either depth. As all the sediment above these components consists exclusively of fibrous gyttja, these samples indicate that sea level has not transgressed Little Woss Lake (153 m above modern) or likely the rest of Woss Lake Valley since at least ca. 14,137 cal BP.

### *Exposed Outer Coastal Sites – Topknot Lake*

While post-LLGM vegetation succession in the vicinity of Little Woss Lake is largely comparable with other late Pleistocene records described on Vancouver Island, the record from Topknot Lake is unique in the region. The early environmental record of Topknot Lake is contained within inorganic fine grey silts older than ca. 17,400 cal BP. In order to date the lowest segment of the core, I extrapolated the age-depth model from the deepest radiocarbon assay at 920 cm dbp using the same priors for sedimentation rate as in the section above.

With respect to the vegetation reconstruction, pollen concentrations at these depths are very low (between 41 and 174 grains/cm$^2$/cal yr), and so interpretation of this part of the record is tentative. Additionally, pollen counts from the samples in this zone are low with several samples yielding fewer than 300 pollen grains (minimum 26 grains at 1120 cm dbp). Several other sites in coastal BC with early non-arboreal assemblages have basal sections with low pollen counts, including Dogfish Bank and Kilgii Gwaay (Lacourse et al. 2005; Mathewes et al. 2019). Acknowledging the limitations above, the record from this early period at Topknot Lake (ca.

18,380-17,450 cal BP, extrapolated) is characterized by a mixed-conifer assemblage dominated by spruce (peak 29.9% of total pollen) with pine, fir, and alder at moderate proportions and low values of mountain hemlock and western hemlock. The remainder of the assemblage consists of two general groups: dry-adapted herb/shrub species and meadow/wetland plants in the riparian zones. The dry-adapted species in this zone, including grasses (peak 20.0% of total pollen), sagewort, heathers, and dwarf willow are possibly representative of pollen rain from adjacent slopes and uplands in the broader area. The meadow/wetland plants observed in this component include sedge, buttercup, burnet, and other rose family taxa. These may have been growing around the lake margin and in the wetlands that feed the lake's southern end. The Fabaceae-type pollen recorded in this zone may represent dry meadow lupines or vetches on nearby slopes (Underhill 1986; Pojar and MacKinnon 2004). The relatively low proportions of tree pollen in this part of the record likely indicates an open woodland or parkland environment with stands of conifers in a sea of grasses and dry-adapted herb and shrub species.

Overall, the composition of this earliest vegetation zone is difficult to interpret, especially considering the low pollen counts. The relatively large spruce component of TKL-1 may not represent pure Sitka spruce stands because the rest of the assemblage—including pine and the herbaceous taxa—does not indicate the kind of moist maritime environment it favours (Pojar and MacKinnon 2004). Other potential spruce species including Engelmann spruce (*Picea engelmannii*), white spruce (*P. glauca*), and black spruce (*P. mariana*) do not occur on Vancouver Island and are uncommon in coastal British Columbia today (Klinkenberg 2019a), though hybrids of these species with Sitka spruce can be found in low elevation valleys and along major river corridors of the Pacific Coast Ranges (Pojar and MacKinnon 2004). The fragment of spruce wood found in the transitional pollen sample between zone TKL-1 and TKL-2a (935 cm dbp, see Figure 22g) indicates that the spruce pollen in zone TKL-1 is local rather than the result of long-distance transport. However, some of the pollen and the wood fragment may be the result of redeposited nearby slopewash or meltwater during periods of high sedimentation in the early period rather than the result of vegetation directly around the lake margin (McAndrews 1984; Mathewes et al. 2019). Similar assemblages are known from older pre-LLGM deposits on southeastern Vancouver Island where they are interpreted to represent subalpine parkland (Miskelly 2012).

Following this mixed-conifer-grass-herb stage, there was a brief period dominated by willows (peak 47.8% of total pollen). The pollen may have been derived from a band of wetland species at the lake margin or possibly dry-adapted prostrate species like arctic willow (*Salix arctica*). Several species must have been nearby, as numerous willow pollen morphotypes were observed in these samples. A similar peak in willow pollen have been observed in the early record (ca. 16,500 cal BP) at West Side Pond in Haida Gwaii (Lacourse et al. 2005). Pine pollen still occurs at proportions up to 12.7% of the total, though this may be accounted for by long-distance transport, which is well-documented for Pinaceae taxa (McAndrews 1984; Hebda and Allen 1993; Rousseau et al. 2008; Szczepanek et al. 2017; Takeuchi et al. 2019). Herbs and shrubs expanded in proportion and diversity through this period, indicating an open landscape. Grasses and especially sedges were common, with the slopes and lake flats covered in wildflowers of the aster family and rose family, as well as heathers, crowberry, buttercup (*Ranunculus* sp.), burnet (*Sanguisorba* sp.), Jacob’s-ladder, cow-parsnip, lilies (Liliaceae), louseworts (*Pedicularis* sp.), and possibly chickweeds and other members of the pink family (Caryophyllaceae), bitter-cress (Brassicaceae, mustard family), lupines and vetches (Fabaceae, pea family), and others. Club-mosses (*Lycopodium*), parsley ferns, and horsetail also grew nearby. Sagewort continued to be present through this period, though its pollen is less abundant. Many of these wildflowers and herbs are adapted to dry or rocky slopes (Pojar and MacKinnon 2004). Such diverse non-arboreal assemblages are not commonly recorded following the LLGM on the coast (Hebda 1995; Hebda and Whitlock 1997). However, comparable assemblages have been documented during older cold and dry glacial climates elsewhere on Vancouver Island (Al-Suwaidi et al. 2006; Miskelly 2012) as well as in areas that remained unglaciated during local glacial maxima or were deglaciated early, including on parts of Haida Gwaii, the islands off southern Alaska, and on the western side of the Olympic Mountains of Washington state beyond the reach of Cordilleran ice (Warner et al. 1982; Heusser 1983; Warner 1984; Heusser 1989; Barrie et al. 1993; Lacourse et al. 2003; Lacourse et al. 2005; Misarti et al. 2012).

The ancient environmental DNA sample from Topknot Lake at 875 cm dbp (ca. 17,166 cal BP) further supports the local presence of non-arboreal species such as willows, Jacob’s ladder, bistort, and asters in the lake catchment. The eDNA at this depth also indicates that conifers including fir and spruce as well as black cottonwood were present near the lake in this period, despite low percentages of arboreal pollen in this zone. During this interval, the climate at Topknot Lake became colder and drier than in TKL-1, possibly as a result of the approach of Cordilleran ice (ca. 17,450-17,150 cal BP) combined with changes in north Pacific airflow and the southward push of the jet stream (Hebda and Whitlock 1997).

The following thousand years at Topknot Lake (ca. 17,150-16,000 cal BP) continued to be dominated by non-arboreal dry-adapted taxa on the uplands and slopes with some moisture-dependent species populating the lower slopes around the lake itself. Low proportions of pine (peak 18.2% of total pollen) and other conifers continue through this period but most of the assemblage to consists of herb and shrub taxa, many in greater abundance than in preceding zones. Willows are present but do not dominate the assemblage as in the previous period; however, other shrubs including soapberry, bunchberry/dwarf dogwood, valerian (Caprifoliaceae), and several pollen morphotypes of heathers occur. The hinterland of the lake was populated by all the herb species noted in the previous period as well as others including meadowrue (*Thalictrum* sp.), saxifrages (Saxifragaceae), fireweed (*Epilobium angustifolium*) and other members of the evening primrose family (Onagraceae), and species in the buckwheat family (Polygonaceae). As with zone TKL-2a, this assemblage suggests an open environment with herbs and low shrubby vegetation, comparable with records from earlier cold and dry full-glacial climates elsewhere on the coast (Al-Suwaidi et al. 2006; Miskelly 2012; Mathewes and Clague 2017) and in Beringia (Colinvaux 1964; Rampton 1971; Ager 1975; West 1981; Anderson and Brubaker 1996; Colinvaux 1996; Savvinova et al. 1996).

After ca. 16,000 cal BP, the record at Topknot Lake indicates a climatic amelioration and the resurgence of arboreal species, especially shore pine (up to 56.5% of total pollen) and alder (up to 27.6% of total pollen), but also fir, spruce, western hemlock, and mountain hemlock. Shrubs and shade-intolerant herbaceous species nearly disappear from the lake area, with little evidence of willow and infrequent grasses, sagewort, lilies, rose-family herbs, and burnet occupying the gaps in the woodland canopy. Fern and ally species appear in high proportions (peak 31.3% of total), including bracken, parsley ferns, and horsetail in the open forest and around the lake margins.

Few aquatic species were observed through the early record from Topknot Lake, only appearing in pollen assemblages after ca. 16,000 cal BP. This may be a result of oligotrophic lake conditions and high sediment accumulation in the early part of the record (indicated by inorganic silts at the bottom of the sequence, see Figure 22h,i), followed by eutrophication and build-up of diatoms other algae (e.g. *Pediastrum* and *Chara*) (see Figure 22a,b). The abundance of both water flea resting eggs and green algae/stonewort oospores in the diatomaceous gyttja at 910 cm dbp (17,327 cal BP) is indicative of significant nutrient deposition in the lake along with freshwater conditions with some alkalinity. Similar assemblages have been documented at low-latitude arctic lakes in Greenland (Fredskild 1983) as well as full glacial and early postglacial lakes on Haida Gwaii (Warner et al. 1982; Warner 1984; Mathewes and Clague 2017). Comparing the fossil evidence from Topknot Lake with these parallel assemblages indicates that the basin became cold and eutrophic after ca. 17,300 cal BP.

Diatom assemblages from the same depositional unit also support this interpretation. Both samples analyzed from this component, at 875 cm dbp (ca. 17,166 cal BP) and 910 cm dbp (ca. 17,327 cal BP), indicate largely freshwater diatom populations in the lake, though the assemblage at 910 cm dbp consists of more species associated with brackish environments. The other proxies at this depth (*Daphnia* and *Chara*) indicate that the brackish assemblage is more likely associated with cold, somewhat alkaline freshwater conditions. The lack of glaciomarine clays or marine diatom species in the record (save for a single specimen noted at 910 cm dbp) demonstrates that relative sea level did not reach or surpass 13 m above modern sea level in this region at least back to ca. 17,300 cal BP. This diatomaceous gyttja continues into the upper portion of the record I examined from Topknot Lake. However, above 591 cm dbp it becomes interbedded with mats of fibrous organics, coincident with a marked spike in pondweed (up to 42.0% total pollen) and the appearance of yellow pond-lily.

The composition of the uppermost assemblage at Topknot Lake after ca. 16,000 cal BP is coeval with the pine-dominated basal zone at Bear Cove Bog on the eastern side of Vancouver Island (Hebda 1983) and mirrors many early post-LLGM sites on the coast. Other sites from Alaska to southern Vancouver Island—including Little Woss Lake—demonstrate comparable pine woodland environments, but generally not until after ca. 15,000 cal BP (Cwynar 1990; Hansen and Engstrom 1996; Hebda 1997; Brown and Hebda 2002; Lacourse et al. 2003; Hebda et al. 2005; Lacourse 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005; Galloway et al. 2008; McLaren 2008; Ager 2019). Though not as moist as later parts of the Holocene, this assemblage likely represents a warmer and wetter time between ca. 16,000 and 13,400 cal BP than the cold and dry non-arboreal tundra conditions of the previous two thousand years (Hebda and Whitlock 1997). I did not analyze Younger Dryas or Holocene pollen assemblages at Topknot Lake.

The record from Topknot Lake demonstrate a continuous dry/cold but vegetated landscape on the outer coast of northern Vancouver Island for at least 2,000 years (ca. 18,000-16,000 cal BP) during the late Pleistocene. This sequence extends the palaeoenvironmental record well into the hypothesized maximum of the Fraser Glaciation in coastal British Columbia at ca. 18,200 to 17,600 cal BP (Clague et al. 1980; Hicock and Armstrong 1985; Clague and James 2002; Dyke 2004). The open herb/shrub-tundra environments at Topknot Lake at this time are similar to the pre-19,500 cal BP record at Port Eliza Cave to the south (Al-Suwaidi et al. 2006) and may represent ecological continuity throughout the LLGM on the outer coast of northern Vancouver Island.

### 4.8.3 Regional Stability and Glacial Refugia

The broad strokes of climate, glacial history, and biotic succession in northwestern North America during the late Pleistocene and early Holocene have been known for at least half a century (Hansen 1947; Heusser 1960; Armstrong et al. 1965; and many others). Rooted in these early efforts, the story of the Fraser Glaciation across this coast has been one of extrapolating the specific to the general. The sites and areas that have been most extensively studied (e.g. the Fraser Lowland and Haida Gwaii) became the moulds to which the intervening landscape has conformed. This approach is fundamental to empirical historical science and allows us to compare and discuss and build upon the work of ourselves and others. However, it can also lead to smoothing over the specific circumstances that differentiate regions, local areas, and sites from one another in the interest of a cohesive narrative. It is exactly these specifics—the blemishes in the story—that drive new questions and research.

For example, researchers have spent decades conducting archaeological excavations, speaking with elders, coring, sampling sediment exposures, and scanning the coasts of Haida Gwaii to reconstruct the past. In so doing they have unearthed many of these blemishes—the early deglaciation of Cape Ball; the effects of isostatic forebulge to create markedly lower sea levels in Hecate Strait; the impact of local ice on Cordilleran glacial flow and extent; the existence of the Hecate Refugium; the antiquity of settlement by Indigenous peoples and preservation of ethnobotanical remains at sites like Kilgii Gwaay, Gaadu Din, and elsewhere; to name a few (Warner et al. 1982; Clague 1983; Warner 1984; Clague 1989a; Heusser 1989; Josenhans et al. 1997; Barrie and Conway 1999; Fedje and Mathewes 2005; Lacourse et al. 2005; Fedje et al. 2011; Cohen 2014; Mathewes and Clague 2017; Shaw et al. 2019; and others). This research has exposed the unique history of Haida Gwaii during and since the Fraser Glaciation and its relationship to the people, plants, animals, and landscape of the archipelago.

Few regions have seen as extensive consideration as Haida Gwaii, but with more scholars investigating the ‘empty space’ between older research sites we begin to unearth snapshots of the coastal past that are better informed, more local, and perhaps more interesting than the previously smoothed surface (McLaren 2008; Mackie et al. 2011; McLaren et al. 2014; McLaren et al. 2015; Letham et al. 2016; Mathewes and Clague 2017; Darvill et al. 2018; Fedje et al. 2018; Lesnek et al. 2018; Mackie et al. 2018; Shaw et al. 2019; and many others). There is no doubt that large-scale hemispheric cooling, shifts in the jet stream, cycling in seasonal perihelion, continental climate changes, land exposure by eustatic sea level lowering, and countless other factors have facilitated change in late Pleistocene environments on the northwest coast of North America (Hays et al. 1976; Hicock et al. 1982; Barnosky et al. 1987; Thompson et al. 1993; Hebda and Whitlock 1997; Dyke 2004; Dyke 2005; Peltier and Fairbanks 2006; Chang et al. 2008; Hendy and Cosma 2008; Clark et al. 2009; Nisancioglu 2009; Seguinot et al. 2016; and others). However, the most complete—and often surprising—picture can be found at the intersection of these global ‘smoothing’ effects with regional and local processes. It is this intersection that demonstrates the distinctive history of Haida Gwaii, and it is a similar local expression of large-scale factors that multidisciplinary research seeks today, including on northern Vancouver Island.

Having described the early environmental histories of Little Woss Lake and Topknot Lake and noted the effects of differential deglacial processes, ice flow patterns, and local topography across the north Pacific coast of North America, I will now discuss these sites in regional context. The basal dates from Topknot Lake (ca. 18,000-17,500 cal BP) on the outer west coast of northern Vancouver Island fall between the early dates on Haida Gwaii and the continental shelf off the central coast (ca. 19,400-18,500 cal BP; Warner et al. 1982; Blaise et al. 1990) and those derived from islands, bays, and major outer waterways proximal to the Pacific Coast Ranges from southern Alaska to northern Vancouver Island (ca. 17,000-16,000 cal BP; Luternauer et al. 1989a; Barrie and Conway 1999; Lacourse et al. 2005; Davies et al. 2011; Addison et al. 2012; Misarti et al. 2012; Darvill et al. 2018; Lesnek et al. 2018).

Cordilleran ice originating from the mainland fjords during the glacial advance would likely have occupied Queen Charlotte Strait and pushed into Queen Charlotte Sound before flowing across the Nahwitti Plateau, covering some outer coastal areas of northern Vancouver Island last and retreating from them first (Howes 1983; Clague and James 2002). Additionally, as there is no direct evidence of glaciomarine deposits in Topknot Lake, relative sea level apparently did not transgress beyond 13 m above modern sea level in this area during the late Pleistocene. Deglaciation began near Topknot Lake no later than ca. 18,000 cal BP, and the effects of an isostatic forebulge would have lowered relative sea level significantly after this time (Clague 1983). As ice continued to retreat, the migration of this forebulge eventually exposed portions of the continental shelf off Vancouver Island and nearby Cook Bank, lowering relative sea level to at least 135 m below modern by ca. 15,400-14,100 cal BP and 95 m below modern by ca. 12,600-

12,100 cal BP (Luternauer et al. 1989b; Hetherington et al. 2004; Shugar et al. 2014). Consistently low relative sea levels also suggest limited isostatic depression of the crust by local glacial ice in the area of Topknot Lake, especially when compared with much higher relative sea levels (nearly 200 m asl) on the eastern side of Vancouver Island, closer to the sources of Cordilleran ice in the Coast Mountains (Fedje et al. 2018). Together, limited glacial cover, thousands of years of lower relative sea levels in adjacent coastal regions, and established vegetation communities on the outer western coast of northern Vancouver Island during the late Pleistocene indicate a stable refugial landscape during the LLGM.

Following maximum coastal ice cover on the coast ca. 18,200-17,600 cal BP, Cordilleran glaciers continued to retreat southward and eastward across northern Vancouver Island. These decaying ice masses first exposed Bear Cove Bog (ca. 16,450 cal BP) followed by Misty Lake (ca. 14,900 cal BP) due to the patterns of ice flow discussed previously as well as the temperature-moderating effect of the ocean on coastal sites, which may have accelerated already rapid deglaciation (Hebda 1983; Lacourse 2005). Little Woss Lake, nestled in the Vancouver Island Ranges, was deglaciated after these coastal sites (ca. 14,300 cal BP) as Cordilleran and local Vancouver Island ice retreated (Howes 1983). Retreat of Cordilleran ice from all these sites as well as others up and down the coast is largely coincident with the arrival of warmer and drier pine-dominated vegetation assemblages described above (Hebda 1995; Lacourse et al. 2003). This warming climate is first represented by the Bølling-Allerød (ca. 14,700-12,800 cal BP), but is then interrupted by the colder Younger Dryas (ca. 12,800-11,500 cal BP) before the onset of rapid warming through the early Holocene (Johnsen et al. 1992; Alley 2000). Among other factors, this warming was aided by a gradual orbital procession to peak northern summertime perihelion—the time of the year when Earth is closest to the sun—between ca. 14,000 and 7,000 cal BP (Hays et al. 1976; Hebda and Whitlock 1997; Nisancioglu 2009). Younger Dryas-aged cooling and moistening is indicated at numerous sites on northern Vancouver Island by increased variety in conifer species, especially increases in mountain hemlock (Mathewes 1993). Records from Little Woss Lake, Bear Cove Bog, Misty Lake, and several sites on Brooks Peninsula demonstrate this climatic shift (Hebda 1983; Hebda 1997; Lacourse 2005). At these same sites, the cool and moist Younger Dryas consistently transitioned to warm and dry early Holocene conditions between ca. 11,500 and 11,100 cal BP, as indicated by the northward migration of Douglas-fir into the region (Lacourse 2005). This species is uncommon on northern Vancouver Island today, only found at dry interior-like sites such as Little Woss Lake (BC Forest Analysis and Inventory Branch 2018). By the mid-Holocene—after ca. 7,000 cal BP—climate in the region became moist and temperate enough for the dominance of species like western redcedar that continue to be important to the monumental woodworking and canoe-building technologies of coastal Indigenous peoples (Hebda and Mathewes 1984).

### 4.8.4 Implications for the Coastal Migration Route and the Peopling of the Americas

The late Pleistocene history of northern Vancouver Island has long been a missing piece in the palaeoenvironmental context of northwestern North America. Because of this, the region is of considerable interest in understanding glacial and postglacial environmental dynamics on the coast. However, it is also a poorly known portion of the hypothesized migration route for the first peopling of the Americas along the Pacific coast. With evidence mounting against the interior ice-free corridor as an initial migration route (Heintzman et al. 2016; Pedersen et al. 2016; Froese et al. 2019; Waters 2019), the Pacific coast has recently seen new scrutiny as a viable alternative (Braje et al. 2017; Potter et al. 2017; Potter et al. 2018; Froese et al. 2019; Braje et al. 2019a). By demonstrating ice-free environments on the outer west coast of northern Vancouver Island ca. 18,000-17,500 cal BP and by adding to the late glacial and postglacial environmental history of the region, we have begun to close a key gap along the coastal corridor. Questions remain, however: how do we evaluate the viability of living on the coast through the late Pleistocene?

How might people have accessed these refugial landscapes and how did they eat, survive, and thrive here?

Recent genetic evidence suggests that early peoples entering the Americas separated from their Beringian forebears between approximately 22,000 and 18,100 cal BP, and further diverged as they spread south of the Cordilleran and Laurentide ice sheets between 17,500 and 14,600 cal BP (Moreno-Mayar et al. 2018a; 2018b; Raff 2019). Combining this genetic perspective with archaeological evidence in the Americas dating back to at least ca. 15,300 cal BP (Davis et al. 2019) and possibly earlier (Dillehay et al. 2015), I will consider the coastal migration route between ca. 19,000 to 15,000 cal BP.

The network of islands, fjords, and ice tongues snaking their way across the late glacial Pacific coast would undoubtedly have required that early peoples used watercraft. Indirect evidence suggests that humans have conducted such voyages for tens of thousands of years, crossing the palaeo-islands and straits of Sunda to Sahul and appearing on remote archipelagos around the world (O'Connell et al. 2010; O'Connor 2010; O'Connor 2015; Clarkson et al. 2017; Bird et al. 2019). Medium-to-long distance open-ocean boat voyages (up to ~60 km) are indicated on the Pacific Rim at multiple sites in the islands of Japanese archipelago by ca. 36,000 cal BP (Kaifu et al. 2015). Collection and open-ocean transport of obsidian across colder north Pacific waters off Honshu is demonstrated in this early period by the stone's presence in archaeological contexts (Ikeya 2015). In North America, boat travel to Triquet Island off BC's central coast is implied by ca. 14,000-13,000 cal BP and to the Channel Islands off the coast of California by ca. 13,000-12,000 cal BP, with coastal adaptations documented archaeologically elsewhere along the Pacific coast of the Americas extending back to ca. 15,000 cal BP (Keefer et al. 1998; Sandweiss et al. 1998; Erlandson et al. 2007; Erlandson et al. 2008; Erlandson et al. 2011; Reitz et al. 2016; Dillehay et al. 2017; Gauvreau and McLaren 2017). Unfortunately, boats made of organic materials are unlikely to preserve in the archaeological record. Instead, we must rely on evidence for humans in places that are difficult to access by other means, or on resources that cannot be acquired except using robust maritime toolkits. Researchers have considered the feasibility of simple wood or skin boats in early postglacial environments in the Americas and Europe with little consensus (Bjerck 2009; Mackie et al. 2011; Bjerck and Zangrando 2013; Breivik 2014; Fletcher 2015; Bjerck et al. 2016; Breivik et al. 2016; Bjerck 2017). However, the challenges often posed to these reviews and studies—namely, that few food resources were available to early peoples in late glacial environments, and that maintaining boats would not be possible in environments assumed to be treeless—need not necessarily be an obstacle. Along the coastal migration route, palaeoecological records derived from pollen, macrofossils, eDNA, and other proxies now demonstrate the presence of terrestrial environments offering both food resources and toolmaking/boatbuilding supplies for early peoples:

In the north, Mann and Peteet (1994) suggest deglaciation of Kodiak Island off the Alaska coast beginning by ca. 17,900 cal BP. The basal record from Deep Lake on Sanak Island off southwestern Alaska indicates that the landscape was deglaciated by ca. 17,000 cal BP and vegetated after ca. 16,300 cal BP with tundra species, especially dwarf willows, grasses, and sedges (Misarti et al. 2012). Few terrestrial pollen records are available further to the east around the Gulf of Alaska, though deglacial estimates for the region suggest that ice was in retreat by ca. 16,700 cal BP and began opening the waterways, exposing the modern fjordland coast by no later than ca. 14,800 cal BP (Davies et al. 2011). Additionally, the remains of marine invertebrates and cold-adapted pinnipeds have been recovered from the Bering Glacier foreland in southern Alaska that date to as early as 16,000 cal BP, indicating an environment that could have provided substantial food resources for early peoples (Yesner et al. 2019).

In southeastern Alaska, the late Pleistocene palaeoenvironment between the channels, islands, and fjords is complex, reflecting shifting ice and vegetation cover through time. The timing of deglaciation on this part of the coast was variable, and lower relative sea level in the region due to isostatic forebulge effects suggests that lowland offshore refugia may have been present through the LLGM (Carrara et al. 2007). These low-lying coasts as well as the outer islands of the Alexander Archipelago were uncovered by retreating Cordilleran ice no later than ca. 17,000 cal BP (Lesnek et al. 2018). Even at this time, some cold-adapted animal species likely populated southeastern Alaska, including ringed seals, northern sea lions, and arctic fox as well as some waterbirds and small mammals (Heaton and Grady 2003; Wigen 2005). At Hummingbird Lake on Baranof Island in the outer Alexander Archipelago, Ager (2019) demonstrates a well-established conifer parkland of pine, willow, heathers, sedges, and grasses by ca. 15,240 cal BP. Palaeontological evidence demonstrates the presence of many animal species in the terminal Pleistocene as well, including bear, caribou, river otter, and wolverine (Heaton and Grady 2003; Wigen 2005).

Haida Gwaii, unique in its detached position from the rest of coastal British Columbia, also provides evidence for well-established terrestrial and marine palaeoenvironments. Though local ice remained in the central mountains of Haida Gwaii (Clague 1989a), the periphery of the archipelago was deglaciated earlier than other regions of the coast, especially those closer to the mainland sources of Cordilleran ice. Parts of Haida Gwaii were ice-free at least to ca. 18,600 cal BP at Cape Ball on the eastern side of Graham Island (Warner et al. 1982). Nearby, Dogfish Bank in Hecate Strait was free of ice by at least ca. 17,365 cal BP (Lacourse et al. 2005). However, the outer portions of Dixon Entrance at the northern end of the islands, under the influence of continental ice flowing along sea-floor troughs, were uncovered after ca. 16,700 cal BP (Barrie and Conway 1999; Shaw et al. 2019). Though few large land mammals were likely present on the islands at this time, palaeontological and genetic evidence for brown bear, black bear, ermine, Steller’s jay, Haida Gwaii slug, and chum, sockeye, and coho salmon suggests that populations of these species survived through the depths of glaciation (Kondzela et al. 1994; Byun et al. 1997; Smith et al. 2001; Fleming and Cook 2002; Burg et al. 2005; Reimchen and Byun 2005; Beacham et al. 2006; Ovaska and Sopuck 2013).

Pollen records are scant for this earliest time, but the records from Cape Ball and Dogfish Bank demonstrate herb-shrub tundra species including dwarf willow, grasses, sedges, and shrub species like crowberry and bunchberries by ca. 17,080 cal BP, likely part of the Hecate Refugium off the eastern side of Graham Island (Heusser 1989; Barrie et al. 1993; Lacourse and Mathewes 2005; Lacourse et al. 2005; Mathewes and Clague 2017). Sediments from southwestern and northwestern Haida Gwaii indicate that a similar herb-shrub tundra was established by ca. 16,890 cal BP at West Side Pond and 16,250 cal BP at Langara Island, respectively (Heusser 1995; Lacourse et al. 2005). After ca. 14,300 cal BP, mixed-conifer forests of pine, hemlock, and alder with an open fern understory spread throughout the islands (Hebda et al. 2005; Lacourse and Mathewes 2005; Lacourse et al. 2005), though some sites such as Hippa Island off the western side of Graham Island remained cold and dry through this time with herbaceous assemblages (Lacourse et al. 2012).

South of Haida Gwaii, ice tongues that had filled the troughs of Queen Charlotte Sound began retreating by ca. 18,100 cal BP, and deglaciation of the outer islands of the central coast was underway after ca. 17,600 cal BP (Darvill et al. 2018; Shaw et al. 2019). As further north, this retreat and subsequent crustal forebulge effect exposed low-lying islands to a depth of 135 m below modern sea level on the outer coast over what is now Goose Bank and Cook Bank by at least 15,400 cal BP (Shugar et al. 2014). Vegetation records for these early submerged landscapes do not yet extend as far as back as the earliest archaeological sites south of the ice sheets (>15,500 cal BP, Waters et al. 2019). However, these outer islands remained stable for thousands of years, with pine-dominated forests present at 95 m below modern sea level on Cook Bank by 12,400 cal BP (Lacourse et al. 2003).

The new record from Topknot Lake connects northern Vancouver Island with the early herb-shrub tundra assemblages in Haida Gwaii and further north. Cordilleran ice retreated from the waters off the northern tip of Vancouver Island by ca. 18,850 cal BP (Blaise et al. 1990), and if Topknot Lake was glaciated at all, it was ice-free by ca. 18,000-17,500 cal BP. Diverse herbs and shrubs are represented early at Topknot Lake (at least ca. 17,500-16,000 cal BP), including willows, grasses, sedges, sagewort, crowberry, cow-parsnip, soapberry, and several heather-family species—possibly representing species like salal, bog cranberry, blueberry, bog laurel, or Labrador tea. Microfossil spruce wood was found *in situ* on pollen slides at ca. 17,500 cal BP, and the presence of fir, spruce, willow, and other herb/shrub tundra taxa are also confirmed in this period by ancient environmental DNA evidence. On the glacio-proximal eastern side of northern Vancouver Island and Queen Charlotte Strait, sites such as at Bear Cove Bog demonstrate deglaciation earlier than ca. 16,450 cal BP along with pine parkland vegetation (Hebda 1983). Other sites across the north end of the island including Misty Lake, Brooks Peninsula, Port Eliza Cave, and Little Woss Lake indicate pine parkland environments as well as large and small mammal species after ca. 15,000 cal BP (Hebda 1997; Lacourse 2005; Al-Suwaidi et al. 2006; Steffen and Fulton 2018). Palaeoecological gaps remain through the early period on the central parts of western Vancouver Island, but the same pine, alder, and fern-dominated parkland is evident on the western side of southern Vancouver Island by at least 15,550 cal BP (Brown and Hebda 2002).

These palaeoenvironmental indicators suggest that on the Pacific northwest coast of North America, the earliest ecological niches that could have been tapped by humans may have been pelagic and near-shore environments where early people hunted pinnipeds or cetaceans and supplemented their diet with shellfish as well as land-based hunting and gathering in peripheral lowlands or refugia (Hetherington and Reid 2003; Kenady et al. 2011; Waters et al. 2011; Bjerck et al. 2016; Yesner et al. 2019). Sea mammal hunting is demonstrated in several early archaeological assemblages from Haida Gwaii, the north and central coast, and northern Vancouver Island, including at Triquet Island (Fargo 2013; Gauvreau and McLaren 2017), Far West Point (McLaren 2008), Kilgii Gwaay (Fedje et al. 2011; Cohen 2014), and the O'Connor Site (Chapman 1982). The late glacial lowland plains of coastal BC also provided habitats for megafauna that could have been hunted by early peoples, including ancient bison, American mastodon, and Jefferson's ground sloth (*Megalonyx jeffersonii*) as well as other smaller species including deer, caribou, mountain goat, and bears (Heaton and Grady 2003; Wigen 2005; Al-Suwaidi et al. 2006; Kenady et al. 2011; Waters et al. 2011; Steffen and Fulton 2018). While these species may not all have been present in the depths of the Fraser Glaciation, they may have arrived soon after climatic conditions began to ameliorate in the terminal Pleistocene.

Collectively, these glacial and biotic histories along the Pacific Rim of northwestern North America demonstrate ecological and environmental continuity extending into the hypothesized maximum of the Fraser Glaciation on the coast ca. 18,000 cal BP (Clague et al. 1980). Perhaps more importantly, these palaeoecological records show that marine and terrestrial species including sea mammals, megafauna, birds, and shellfish as well as many varieties of berries, trees, roots, and other plant-based resources would have been available to early peoples along the Pacific coastal route by ca. 16,000 cal BP or earlier. In some places including southeastern Alaska, Hecate Lowlands, and the outer coast of northern Vancouver Island, larger animal species and robust plant communities may have been available throughout the glacial period. Resources for building and maintaining skin boats and a suite of organic tool technologies were available throughout the coastal route at this time, including versatile branches of willow and other shrubs. In some places, as at Topknot Lake, early peoples would have had access to strong conifer wood at least as early as ca. 17,500 cal BP.

## 4.9 Conclusions

The data from Topknot Lake demonstrate that the outer west coast of northern Vancouver Island had a very different history during the Fraser Glaciation than other sites closer to the sources of Cordilleran ice. Between 18,000-16,000 cal BP, the herb-shrub assemblages from Topknot Lake are more comparable with early postglacial and refugial environments in Haida Gwaii (Warner 1984; Barrie et al. 1993; Heusser 1995; Lacourse et al. 2005) or pre-glacial records on Vancouver Island (Al-Suwaidi et al. 2006; Miskelly 2012) than with post-LLGM records elsewhere in the region (Hebda 1983; Hebda 1997; Lacourse 2005; Al-Suwaidi et al. 2006). After 16,000 cal BP, the ecosystems on the outer west coast during the Bølling-Allerød warming begin to resemble the pine woodlands from elsewhere in the region. Further inland, Little Woss Lake demonstrates continuity with other early pine-dominated northern Vancouver Island sites down to the base of the record. However, both sites also demonstrate the important influence of local physiography on the extent and timing of glaciation across the region. Topknot Lake's early radiocarbon dates and vegetation assemblage align with both a later and less-extensive glaciation over the low topography of the outer coast when compared with the lengthy glaciation of inland sites, especially those like Little Woss Lake in the interior glacio-proximal valleys of the Vancouver Island Ranges. Additionally, for thousands of years, relative sea level off nearby Cook Bank and other now-inundated coastal plains was significantly lower than modern (Luternauer et al. 1989b; Hetherington et al. 2004; Shugar et al. 2014). Together with the early ice-free conditions at Topknot Lake, these observations suggest both less extensive glaciation and more exposed land in the region. The differing glacial and ecological histories of these two sites, especially when considered in the broader context of north Pacific coast of North America, demonstrate that the outer coast of northern Vancouver Island and the adjacent coastal plains were either unglaciated refugia throughout the Fraser Glaciation or had well-established floral and faunal communities before the first peoples lived on these lands and seas.

Methodologically, this study investigated the potential for ancient environmental DNA analysis to broaden our ability to answer questions about the makeup of coastal ecosystems. Though both sites yielded numerous plant taxa, matches with animal sequences were limited. Also, the resolution of these samples (one for each site) is too poor for reliable interpretation without analyzing additional proxies. Though 74 samples were taken across both study sites, DNA was successfully extracted from 14 and only 2 provided any meaningful sequences. In both successfully sequenced samples, numerous confirmed and possible contaminant species were noted in the data following the quality control measures introduced as part of the bioinformatic analysis. These had to be manually identified and removed and were not included in the discussion.

Currently, the cost vs. return of ancient eDNA in terms of both human effort and research funding is disadvantageous to the method. However, the descriptive power of eDNA is substantial as it clearly demonstrates the local signal of well-established biotic communities at Topknot Lake ca. 17,000 cal BP and at Little Woss Lake ca. 14,000 cal BP, something that pollen can only imply. Macrofossil and other non-pollen microfossils can also demonstrate local presence, but not at the scale of eDNA where thousands of species matches are derived from a single sample. Because of this, the ancient eDNA records from this study are valuable and complementary to the other methods I used.

Both descriptively and methodologically, this study demonstrates the importance of continued, widespread multi-disciplinary research on late Pleistocene palaeoenvironments in western North America. The records from Topknot Lake and Little Woss Lake are only two of many that will shed light on the dynamics of geological and ecological processes in this region. Future research would benefit from a deeper understanding of the late Pleistocene glacial development of Canada’s west coast, and Vancouver Island in particular. Haida Gwaii has seen extensive research from many palaeoecological and archaeological avenues, as have parts of the central coast of BC. Previous research on the late Pleistocene along the outer west coast of Vancouver Island has been locally intensive (Hebda and Haggarty 1997; Ward et al. 2003; Al-Suwaidi et al. 2006) but not geographically extensive enough to document the apparent complexity of the Fraser Glaciation. Additional studies on the outer coast of northern Vancouver Island—and especially research focusing on now-submerged coastal lowlands off the island’s tip—will be crucial in the task of understanding these environments and the relationships of early peoples with them. Furthermore, research on early postglacial environments further south along the coastal fringe of Vancouver Island may yet constrain a potential coastal migration route and should be a priority. I hope that continued experimentation with ancient environmental DNA will factor into this research, as it will undoubtedly continue to improve in accuracy, ease, and cost.

# 5. Conclusions and Future Directions

At the outset of this thesis, I outlined several questions that informed my approach to this research. The first broad question, focused on the palaeoenvironmental history of northern Vancouver Island, was theoretical:

1. At what time during the late Pleistocene did northwestern and north-central Vancouver Island become viable environments for human habitation?
    a) What are the long-term trends in the makeup of biotic communities through the late Pleistocene and early Holocene on northwestern and north-central Vancouver Island, as recorded in the sediments of Topknot Lake and Little Woss Lake?
    b) Using the presence or absence of plant and animal species at these sites as proxies for glacial proximity, what can these records tell us about late glacial Cordilleran ice advance and retreat in these regions?
    c) Can palaeoecological records like those recovered from the two study sites reveal the presence of refugia in these regions during the LGM along the coast of British Columbia?

The second broad question, focused on expanding the range of methods for palaeoenvironmental reconstruction available to researchers on the Pacific northwest coast of North America, was methodological:

2. Is the use of ancient environmental DNA (eDNA) analysis valuable for reconstructing the palaeoenvironments of coastal British Columbia?
    a) Can fragmentary ancient eDNA be successfully extracted and analyzed from samples collected in coastal BC?

b) Is eDNA analysis complementary to established approaches (including pollen and macrofossils) when analysing the presence or absence of species on the coast?

Throughout this thesis, each question above has been considered. The greatest effort has focused on the first broad question—the palaeoecology of northern Vancouver Island and its place in the late Pleistocene coastal migration route into the Americas. The answer to the second question—the utility and challenges of using ancient environmental DNA in palaeoecological and archaeological analysis—has been demonstrated through the results of the first question and will be summarized below.

By addressing the first broad question, I placed northern Vancouver Island in the context of the late Pleistocene peopling of the Americas along the Pacific coast. Prior to this study, the environmental history of the outer west coast of northern Vancouver Island was poorly known. Certain adjacent areas have extensive published records, including Bear Cove Bog (Hebda 1983), Misty Lake (Lacourse 2005), Brooks Peninsula (Hebda 1997), Port Eliza Cave (Al-Suwaidi et al. 2006), and Cook Bank (Luternauer et al. 1989b; Lacourse et al. 2003). However, the outer coast of the Nahwitti Plateau was a notable gap in the data.

As a result of this study, the data from Topknot Lake demonstrate that the outer west coast had a much different late glacial environmental history than other sites in the region. Most importantly, the record at Topknot Lake was shown to extend back in time to the LLGM. The sparse tundra-woodland and herb-shrub tundra environment in the early record from Topknot Lake (18,000-16,000 cal BP) are comparable with the early postglacial and refugial environments of Haida Gwaii (Warner 1984; Barrie et al. 1993; Heusser 1995; Lacourse et al. 2005) rather than the other sites on northern Vancouver Island. After this time the vegetation communities on the outer west coast resembled the mixed-conifer pine forests from elsewhere in the region. The record from Little Woss Lake demonstrates ecological continuity with other northern Vancouver Island sites of similar age, with pine woodland in the earliest period (14,300 cal BP). Both sites illustrate the importance of local physiography to the extent and timing of glaciation across the region. Topknot Lake's early radiocarbon dates and vegetation assemblage align with both a shorter and less-extensive glaciation over the low topography of the outer coast when compared with the lengthy glaciation of inner sites, especially those like Little Woss Lake in the interior glacio-proximal valleys of the Vancouver Island Ranges. Additionally, for thousands of years in the late Pleistocene relative sea level off nearby Cook Bank and other now-inundated coastal plains was much lower than modern (Luternauer et al. 1989b; Hetherington et al. 2004; Shugar et al. 2014). The different glacial and ecological histories of Topknot Lake and Little Woss Lake, especially when considered in the broader context of Pacific northwest coast of North America, demonstrate that the outer coast of northern Vancouver Island and the adjacent coastal plains were either unglaciated refugia throughout the Fraser Glaciation, or were exposed and vegetated well before the time that the first peoples lived on these lands and seas.

The second question I posed was methodological, asking whether ancient environmental DNA analysis has the potential to broaden our ability to answer questions about the makeup of late Pleistocene ecosystems on the Pacific northwest coast of North America. The utility of analyzing ancient pollen assemblages to understand past vegetation communities in northwestern North America is indisputable. But as with any specific ecological proxy—sedimentology, diatoms, macrofossils, charcoal, trace element analysis, and others—pollen relies on specific assumptions and biases that can influence the interpretation of the record (Hebda 1981; Moore et al. 1995; Bennett and Willis 2001). The ability of eDNA to detect animals, bacteria, and other biota in addition to plants is inherently valuable, though the method has been successful even when used exclusively for identifying the presence of plants remains (Jørgensen et al. 2012b; Pedersen et al. 2013).

Though both sites yielded numerous plant taxa, matches with animal sequences were limited. Also, the resolution of these samples (one for each site) is too poor for reliable interpretation without analyzing additional proxies. With 74 samples taken across both study sites, only 14 successfully extracted, and only 2 yielding any meaningful sequences, the cost vs. return of ancient eDNA in terms of both human effort and research funding is disadvantageous to the method. Additionally, in both successfully sequenced samples, numerous confirmed and possible contaminant species were noted in the data following the quality control measures introduced as part of the bioinformatic analysis. These had to be manually identified and removed and were not included in the discussion.

However, the descriptive power of eDNA is substantial as it clearly demonstrates the local signal of well-established biotic communities at Topknot Lake ca. 17,000 cal BP and at Little Woss Lake ca. 14,000 cal BP, something that pollen can only imply. Macrofossil and other non-pollen microfossils can also demonstrate local presence, but not at the scale of eDNA where thousands of species matches are derived from a single sample. Because of this, the ancient eDNA records from this study are valuable and complementary to the other methods I used.

Perhaps the greatest strength of eDNA analysis is the detection of ancient animal species that inhabited the lake catchment. Whether as small as voles and marmots or as large as short-faced bears and mastodons, the animals living in an ecosystem are key indicators for the reliability of the late Pleistocene coastal environment for early peoples. This study has demonstrated the limited presence of animal species near the basin at Little Woss Lake, including brown bear, Chinook salmon, and arthropods; however, the results from Topknot Lake were extremely limited. This study was unable to replicate the success of others in demonstrating more extensive metazoan records (Hofreiter et al. 2003; Willerslev et al. 2003; Matisoo-Smith et al. 2008; Haile et al. 2009; Pedersen et al. 2016).

The limitations of eDNA encountered during this study are likely those of a method in its infancy. As with any of the other palaeoenvironmental proxies listed above, the challenges of new approaches are often methodological and technological. Recognition of these drawbacks is important when designing this kind of research. However, ancient geneticists have demonstrated some remarkable feats over the past two decades, and the computing power available for DNA sequencing and bioinformatics continues to grow with each passing year (Pedersen et al. 2015). The expansion of technique and analytical power in the field will undoubtedly continue to increase the reliability and accessibility of this method while decreasing the cost. Even with the difficulties of its use as a proxy in this study, it would be a mistake to discard ancient eDNA analysis when researching the palaeoecology of the Pacific northwest coast of North America.

Both descriptively and methodologically, this study demonstrates the importance of continued, widespread multi-disciplinary research on late Pleistocene palaeoenvironments in western North America. The records from Topknot Lake and Little Woss Lake are only two of many that will shed light on the dynamics of geological and ecological processes in this region. Future research would benefit from a deeper understanding of the late Pleistocene glacial development of Canada’s west coast, and Vancouver Island in particular. Haida Gwaii has seen extensive research from many palaeoecological and archaeological avenues, as have parts of the central coast of BC. Previous research on the late Pleistocene along the outer west coast of Vancouver Island has been locally intensive (Hebda and Haggarty 1997; Ward et al. 2003; Al-Suwaidi et al. 2006) but not geographically extensive enough to document the apparent complexity of the Fraser Glaciation. Additional studies on the outer coast of northern Vancouver Island—and especially research focusing on now-submerged coastal lowlands off the island’s tip—will be crucial in the task of understanding these environments and the relationships of early peoples with them. Furthermore, research on early postglacial environments to the south along the outer fringe of Vancouver Island should be a priority to better understand the coastal corridor. I hope that continued experimentation with ancient environmental DNA will factor into this research, as it will undoubtedly continue to improve in accuracy, ease, and cost.

Most importantly, it is vital that late Pleistocene archaeological research such as that being conducted as part of the Northern Vancouver Island Archaeology and Palaeoecology Project continue as more is revealed about the glacial coast. Sea level histories, biotic communities, and the flow of ancient glaciers paint a compelling picture, yet the human voice embedded within these dynamic landscapes and seascapes is elusive. Only by continuing to listen to the elders, the trees, and the stones will we come closer to these answers.

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# Appendix A: Lists of Plant and Animal Classifications

Plant taxa described in the text, including scientific and common names

| Scientific name | Common name |
|---|---|
| *Abies* | true firs |
| *Abies amabilis* | amabilis fir |
| *Abies grandis* | grand fir |
| *Abies lasiocarpa* | subalpine fir |
| *Achlys triphylla* | vanilla-leaf |
| *Alnus* | alder |
| *Alnus rubra* | red alder |
| *Ambrosia* | ragweed |
| *Amelanchier alnifolia* | saskatoon |
| Apiaceae | carrot family |
| *Artemisia* | sagewort |
| Asteraceae | aster family |
| *Betula* | birch |
| *Bistorta* | bistort |
| *Blechnum spicant* | deer fern |
| Brassicaceae | mustard family |
| *Campanula* | harebell |
| Caprifoliaceae | valerian family |
| *Carex sitchensis* | Sitka sedge |
| Caryophyllaceae | pink family |
| *Chara* | green algae/stonewort |
| *Cornus* | dogwood/bunchberry |
| *Cornus canadensis* | bunchberry |
| *Cryptogramma* | parsley fern |
| Cupressaceae | cedar family |
| Cyperaceae | sedge family |
| *Empetrum nigrum* | crowberry |
| *Epilobium angustifolium* | fireweed |
| *Equisetum* | horsetail |
| Ericaceae | heather family |
| Fabaceae | pea family |
| *Gaultheria shallon* | salal |
| *Gentiana* | gentian |
| *Heracleum maximum* | cow-parsnip |
| *Isoëtes* | quillwort |
| *Juniperus* | juniper |
| *Kalmia microphylla* | western bog-laurel |
| Lamiaceae | mint family |
| Liliaceae | lily family |
| *Linnaea borealis* | twinflower |

| **Scientific name** | **Common name** |
|---|---|
| *Lonicera involucrata* | black twinberry |
| *Lycopodium* | club-moss |
| *Lysichiton americanus* | skunk cabbage |
| *Mahonia nervosa* | dull Oregon-grape |
| Malvaceae | mallow family |
| *Menyanthes trifoliata* | buckbean |
| *Menziesia ferruginea* | false azalea |
| *Myrica gale* | sweet gale |
| *Nuphar polysepala* | yellow pond-lily |
| Onagraceae | evening primrose family |
| *Pediastrum* | colonial green algae |
| *Pedicularis* | lousewort |
| *Pentagramma triangularis* | goldenback fern |
| *Picea* | spruce |
| *Picea engelmannii* | Engelmann spruce |
| *Picea glauca* | white spruce |
| *Picea mariana* | black spruce |
| *Picea sitchensis* | Sitka spruce |
| *Pinus* | pine |
| *Pinus contorta* | shore pine |
| *Plantago* | plantain |
| Poaceae | grass family |
| *Polemonium* | Jacob’s ladder |
| *Polemonium boreale* | northern Jacob's ladder |
| Polygonaceae | buckwheat family |
| Polypodiaceae | polypody fern family |
| *Polystichum munitum* | sword fern |
| *Polytrichum* | haircap moss |
| *Populus* | cottonwood |
| *Populus balsamifera* ssp. *trichocarpa* | black cottonwood |
| *Potamogeton* | pondweed |
| *Prunus emarginata* | bitter cherry |
| *Prunus virginiana* | choke cherry |
| *Pseudotsuga menziesii* | Douglas-fir |
| *Pteridium aquilinum* | bracken fern |
| *Ranunculus* | buttercup |
| *Rhododendron groenlandicum* | Labrador tea |
| Rosaceae | rose family |
| *Rosa nutkana* | Nootka rose |
| *Rubus* | berries/bramble fruit |
| *Rubus spectabilis* | salmonberry |
| *Salix* | willow |
| *Salix arctica* | arctic willow |

| **Scientific name** | **Common name** |
|---|---|
| *Sambucus* | elderberry |
| *Sanguisorba* | burnet |
| Saxifragaceae | saxifrage family |
| *Selaginella* | spikemoss |
| *Shepherdia canadensis* | soapberry |
| *Sphagnum* | peat-moss |
| *Taxus brevifolia* | western yew |
| *Thalictrum* | meadowrue |
| *Thuja plicata* | western redcedar |
| *Triglochin* | arrow-grass |
| *Tsuga* | hemlock |
| *Tsuga heterophylla* | western hemlock |
| *Tsuga mertensiana* | mountain hemlock |
| *Typha* | cattail |
| *Vaccinium alaskaense* | Alaskan blueberry |
| *Vaccinium oxycoccos* | bog cranberry |
| *Vaccinium parviflorum* | red huckleberry |
| *Xanthocyparis nootkatensis* | yellow-cedar |

Animal taxa described in the text, including scientific and common names

| **Scientific name** | **Common name** |
|---|---|
| Alcidae | auklets |
| Anatidae | ducks and geese |
| *Arctodus simus* | short-faced bear |
| *Balanus* | barnacle |
| *Bison antiquus* | ancient bison |
| *Callorhinus ursinus* | northern fur seal |
| *Castor canadensis* | Canadian beaver |
| *Cerorhinca monocerata* | rhinoceros auklet |
| Cervidae | deer family |
| *Chen caerulescens* | snow goose |
| *Clangula hyemalis* | long-tailed duck |
| *Corvus corax* | common raven |
| *Clupea pallasi* | Pacific herring |
| *Cyanocitta stelleri* | Steller's jay |
| *Daphnia* | water flea |
| *Dendroctonus ponderosae* | mountain pine beetle |
| *Enhydra lutris* | sea otter |
| *Erethizon dorsatum* | porcupine |
| *Eumetopias jubatus* | Steller sea lion |
| *Fratercula* | puffins |
| Gadidae | cod |
| Gaviidae | loons |
| *Gulo gulo* | wolverine |
| *Haliaeetus leucocephalus* | bald eagle |
| Hexagrammidae | greenling |
| *Hippoglossus stenolepis* | Pacific halibut |
| *Ixodes pacificus* | Pacific black-legged tick |
| *Lagenorhynchus obliquidens* | Pacific white-sided dolphin |
| Laridae | gulls |
| *Lemmus trimucronatus* | brown lemming |
| *Lontra canadensis* | river otter |
| *Mammut americanum* | American mastodon |
| *Mammuthus primigenius* | woolly mammoth |
| *Marmota caligata* | hoary marmot |
| *Martes americana* | American marten |
| *Megalonyx jeffersonii* | Jefferson's ground sloth |
| *Mergus merganser* | common merganser |
| *Microtus longicaudus* | long-tailed vole |
| *Microtus townsendii* | Townsend's vole |
| *Mustela erminea* | ermine |
| *Mytilus* | mussel |
| *Neovison vison* | American mink |

| **Scientific name** | **Common name** |
|---|---|
| *Odocoileus hemionus* | mule deer |
| *Oncorhynchus* | salmon |
| *Oncorhynchus gorbuscha* | pink salmon |
| *Oncorhynchus keta* | chum salmon |
| *Oncorhynchus kisutch* | coho salmon |
| *Oncorhynchus mykiss* | rainbow trout |
| *Oncorhynchus nerka* | sockeye salmon |
| *Oncorhynchus tshawytscha* | Chinook salmon |
| *Ophiodon elongatus* | lingcod |
| *Oreamnos americanus* | mountain goat |
| *Passerculus sandwichensis* | savannah sparrow |
| *Peromyscus maniculatus* | deer mouse |
| *Phalacrocorax* | cormorant |
| *Phalacrocorax auritus* | double-crested cormorant |
| *Phenacomys intermedius* | heather vole |
| *Phoca vitulina* | harbour seal |
| Phocoenidae | porpoise |
| *Phoebastria* | albatross |
| *Phoebastria albatrus* | short-tailed albatross |
| Pleuronectidae | flatfish |
| *Pusa hispida* | ringed seal |
| *Rangifer tarandus* | caribou |
| *Salvelinus malma* | Dolly Varden |
| *Sebastes* | rockfish |
| *Staala gwaii* | Haida Gwaii slug |
| *Squalus suckleyi* | Pacific spiny dogfish |
| *Thaleichthys pacificus* | eulachon |
| *Uria aalge* | common murre |
| *Ursus americanus* | black bear |
| *Ursus arctos* | brown bear |
| *Vulpes lagopus* | arctic fox |
| *Vulpes vulpes* | red fox |

## Appendix B: Postglacial Archaeology on the Northwest Coast

This appendix expands on the late Pleistocene/early Holocene archaeological sites noted on Figure 14 and in Table 1. The order of this appendix corresponds to the numbers on the map and in the table and is generally chronological from oldest to most recent based on currently published radiocarbon dates.

### *Triquet Island (EkTb-9) – Site 1, 14,086-13,613 cal BP*

Triquet Island, one of the outer islands of the central coast of British Columbia, is home to a diverse array of archaeological remains, including fish traps, culturally modified trees (CMTs), shell middens, and lithic scatters (Gauvreau and McLaren 2017). The most well-documented of these sites is EkTb-9, an extensive area curving around a north-facing bay that contains shell midden, petroforms, and intertidal lithic scatters. The oldest radiocarbon date derived from this site extends back to 14,086-13,613 cal BP from a sample of charcoal (Gauvreau and McLaren 2017). Artifacts recovered from these early deposits at EkTb-9 include a cache of lithics, including 10 flakes, three cores, and shatter. (Gauvreau and McLaren 2017).

### *Ayer Pond (45SJ454) – Site 2, ca. 13,950-13,750 cal BP*

Ayer Pond is located on the eastern side of Orcas Island, among the San Juan Islands of Washington state. Three separate wetlands on the island have yielded late Pleistocene remains of ancient bison (*Bison antiquus*), but the earliest date has been documented at Ayer Pond at ca. 13,950-13,750 cal BP (11,990 ± 25 $^{14}$C BP; Wilson et al. 2009; Kenady et al. 2011). The remains seem to show evidence of butchery, including spiral fracture, element selection, and a few cutmarks, for which modern sources and scavenging have been largely ruled out. Though apparently unrelated to the bison remains, evidence for a hearth feature and dacite lithic debitage were also discovered as part of reconnaissance around the edge of the pond (Kenady et al. 2011). These materials were not dated.

### *Manis Mastodon – Site 3, 13,860-13,763 cal BP*

The Manis Mastodon site is located near Sequim, Washington, on the southern side of the Strait of Juan de Fuca. Materials at the site consist of the remains of a disarticulated American mastodon with a bone/antler projectile point embedded in the dorsal surface of the proximal end of a rib, near where it joins the vertebra (Gustafson et al. 1979; Waters et al. 2011). Some of the bones from the site also show evidence of spiral fracture, flaking, and cutmarks that indicate intentional butchery by humans. Organic remains associated with the mastodon were initially dated to ca. 14,000 cal BP, and later, tusk ivory and the rib with the embedded projectile point yielded AMS dates averaging 13,860-13,763 cal BP (Waters et al. 2011).

### *Kildidt Narrows (ElTa-18) – Site 4, 13,673-10,692 cal BP*

ElTa-18, the Kildidt Narrows site, is located on the central coast of British Columbia, on the eastern side of Kildidt Inlet on Hunter Island. Among several sites near the ancient village at Namu that were cored and dated by Cannon (2000), charcoal from the Kildidt Narrows site is the oldest, dating to 11,400 cal BP. Later excavations undertaken by McLaren et al. (2015) at the same location as Cannon's (2000) previous work produced additional radiocarbon ages dating between 13,673-10,692 cal BP. Three small chipped stone chert artifacts were recovered from this layer, including the tip of a biface, a retouched flake fragment, and a small flake (McLaren et al. 2015). A 50cm by 50 cm unit placed in the nearby upper intertidal also yielded a wood chip dating to 10,640-10,517 cal BP, indicating the potential for preserved organic material from the early Holocene at this and similar sites (McLaren et al. 2015; McLaren et al. 2019a).

### *Meay Channel I (EjTa-4) – Site 5, 13,317-13,241 cal BP*

The Meay Channel I site, EjTa-4, is located on the central coast of British Columbia, in a small bay on Calvert Island, just north of Kwakshua Channel. Some of the oldest remains at the site include preserved footprints in the intertidal mud that represent the trails of at least three human individuals, including one juvenile (McLaren et al. 2018). The oldest component at the site (Stratum X), dating to 13,317-13,241 cal BP based on preserved twigs of shore pine, also yielded fine-grained quartzite cores, retouched flake tools, and spall tools in addition to the footprints (McLaren et al. 2018). Archaeological remains from the more recent upper component of the site include objects made of bone, antler, shell, and chipped stone (McLaren et al. 2015).

### *K1 Cave (FgUc-6) – Site 6, 13,150-12,150 cal BP*

K1 Cave is a karst cave system located on Moresby Island along the western coast of Haida Gwaii in British Columbia. K1 has yielded many faunal remains, including common merganser (*Mergus merganser*), deer mouse (*Peromyscus maniculatus*), black bear, and mule deer, as well as trace amounts of fish and amphibians (Fedje et al. 2004; Ramsey et al. 2004). Archaeological remains from the site include the bases of two foliate spearpoints made of chert that are comparable to the Pebble Tool Tradition of early chipped stone artifacts on the northwest coast of North America (Fedje et al. 2004; Fedje et al. 2011). The age of these artifacts is bounded by two radiocarbon dates on bone, one above and one below, which suggest a date range between 13,150 and 12,150 cal BP (10,960 ± 35 and 10,510 ± 35 $^{14}$C BP, Fedje et al. 2011). Because of the location of the artifacts in the dark zone of the cave with no other anthropogenic features or artifacts nearby, it is likely the spearpoints were brought into the cave by wounded bears after people attempted to hunt them near the cave mouth (Fedje et al. 2011).

### *Puget Sound Clovis – Up to ca. 13,000 cal BP*

Aside from a small number of other fluted points found within the southern Salish Sea, none of which are from datable contexts, Clovis points represent some of the earliest documented human occupation in the region (Croes et al. 2008). Clovis points (or pieces thereof) have been collected from sites throughout central and southern Puget Sound, as well as on Whidbey Island

(Croes et al. 2008). These points, often brought to light by collectors rather than professional archaeologists, usually do not have well-dated contexts, and so are dated based on the Clovis typology as up to ca. 13,000 cal BP (Croes et al. 2008). Nevertheless, they represent a key marker of late Pleistocene assemblages developing on the coast.

### *Bear Creek Site (45KI839) – Site 7, 12,770-11,274 cal BP*

Located along Bear Creek on the east side of the Sammamish River in Redmond, Washington, the site consists of several discrete tool-use and processing locations in well-stratified deposits (Kopperl et al. 2015). The assemblage at Bear Creek consists mostly of expedient flake tools, but also contains some bifaces and retouched tools, including apparent stemmed and concave-based points. Materials collected from the site include over 1,000 flakes and shatter, more than 50 cores, several hammerstones, unifaces, bifaces, scrapers, gravers, and projectile points (Kopperl et al. 2015). The oldest radiocarbon date on carbonized wood from the same layer as the projectile point bases at this site yielded an age of 12,770-12,596 cal BP, and is bounded by dates from the layer above, the oldest of which yielded an age of ca. 11,817-11,274 cal BP (Kopperl et al. 2015).

### *Gaadu Din 1 (1693T) – Site 8, 12,600-11,500 cal BP*

Gaadu Din 1 is a cave site located on Huxley Island in southeastern Haida Gwaii in British Columbia. Two flaked stone spear points and two flake tools associated with charcoal and bear bone on the cave floor were dated to between 12,600-11,500 cal BP, and a bone point tip was also directly dated to ca. 11,800 cal BP (Fedje et al. 2011). The spear points may have been brought into the cave by wounded bears as a result of terminal Pleistocene bear hunting by early peoples on the coast. Faunal remains recovered from the cave include salmon (mostly sockeye), Dolly Varden, rainbow trout (*Oncorhynchus mykiss*), several bird species, brown bear, black bear, and deer (Fedje et al. 2011). The majority of the faunal assemblage may be the result of natural depositional processes rather than human agency.

### *Gaadu Din 2 (1906T) – Site 8, 12,500-10,700 cal BP*

Gaadu Din 2 is a cave site located on Huxley Island in southeastern Haida Gwaii in British Columbia and may represent an intermittently used late Pleistocene hunting camp or shelter (Fedje et al. 2011). Several stone tools, including two small bifacial point tips, 11 small resharpening flakes, and several hammerstones were recovered from cave floor sediments near a stratified hearth feature. Elsewhere in the cave, a large bifacial projectile point and a bifacial knife were found on the surface. The hearth feature yielded dates between 12,500-10,700 cal BP (Fedje et al. 2011). Few faunal remains were recovered from the site, though the assemblage did include a small number of fish bones and a poorly preserved piece of black bear bone (Fedje et al. 2011; Mackie et al. 2011).

### *Stave Watershed Sites – Site 9, ca. 12,404-12,051 cal BP*

Several late Pleistocene archaeological sites (DhRn-16, DhRn-21, DhRn-29, DhRo-11, DhRo-53) have been recorded during low-water surveys in the Stave Lake watershed on the northern margin of the Fraser Valley on the Lower Mainland of British Columbia (McLaren 2017). Many lithics identified as part of surface scatters throughout the watershed demonstrate attributes in common with late Pleistocene/early Holocene stemmed and foliate points found in nearby regions of western North America. Subsurface testing was also conducted at several of the sites in the area. The oldest of these is the Cardinalis Creek Mouth Site (DhRn-29), which yielded a chert biface preform and several flakes, and was dated to ca. 12,404-12,051 cal BP (10,370 ± 40 $^{14}$C BP) based on charcoal samples (McLaren 2017). A large flake tool was also uncovered at the nearby Devil’s Point Site (DhRn-16) and was dated based on associated charcoal to a similar period of ca. 12,381-11,829 cal BP (10,290 ± 50 $^{14}$C BP) (McLaren 2017).

### *Werner Bay – Site 10, ca. 12,100-11,700 cal BP*

Located in the waters of Juan Perez Sound off the eastern side of southern Haida Gwaii in British Columbia, a single utilized stone flake was recovered from the sea floor of Werner Bay using a bucket dredge at 53m below modern sea level (Fedje and Josenhans 2000; Fedje et al. 2005a). Based on sea level reconstructions of the region, this landscape was last above water at ca. 12,100-11,700 cal BP (10,200 $^{14}C$ BP, Fedje et al. 2005a; Mackie et al. 2011).

### *Shuká Káa (On Your Knees Cave; 49-PET-408) – Site 11, ca. 12,000 cal BP*

Shuká Káa is a karst cave located at the northern end Prince of Wales Island in southeastern Alaska (Dixon et al. 1997). Several periods of human use are documented at this site, with the oldest beginning ca. 12,000 cal BP (10,300 ± 50 $^{14}C$ BP) with the presence of a worked bone tool (Dixon 2013). The more extensive use of the cave, however, dates to ca. 10,350 cal BP (ca. 9,200 $^{14}C$ BP), and is associated with the remains of a man in his mid-twenties, the oldest human remains known on the Pacific northwest coast (Dixon et al. 1997; Kemp et al. 2007). Carbon isotope analysis of the remains suggests that the man ate a diet emphasizing marine foods. This, along with the presence of obsidian tools fingerprinted with XRF to Mount Edziza in northern British Columbia and Suemez Island in southeast Alaska, suggest that the people using the site at this time had the ability to travel the coast and subsisted on marine resources (Lee 2001; Dixon 2013).

### *Namu (ElSx-1) – Site 12, ca. 11,600-11,550 cal BP*

Namu is located on the central coast of British Columbia on the mainland along the eastern shore of Fitz Hugh Sound. The site at Namu contains extensive shell midden deposits post-dating ca. 6,800 cal BP (6,000 $^{14}C$ BP) and a consistent occupation through the early Holocene, with the deepest deposits from ca. 11,600-11,550 cal BP (9,720 ± 140 $^{14}C$ BP, R.

Carlson 1996). The earliest artifacts recovered from Namu consist of stone tools, including foliate bifaces and cobble tools. (R. Carlson 1996; McLaren et al. 2015). Recent excavations on a road-cut terrace above the village site in 2014 yielded 38 subsurface artifacts in early Holocene contexts (most ca. 11,000-10,000 cal BP), including cores, core tools, cobble tools, and flakes (Porter 2015). Microblade technology appears slightly later (ca. 10,000-9,000 cal BP), and bone and antler artifacts as well as faunal remains are not documented from the site until ca. 7,000 cal BP (McLaren et al. 2015; Porter 2015).

### *Far West Point (GcTr-6) – Site 13, ca. 11,200-11,050 cal BP*

Located on the north coast of British Columbia to the northwest of Prince Rupert and near the border with the Alaska Panhandle, Far West Point (GcTr-6) is located on the western side of Dunira Island, part of the Dundas Islands (McLaren 2008; Martindale et al. 2009). The site consists of shell midden deposits as well as underlying cultural strata consisting of charcoal-rich black silt, for which the earliest radiocarbon date is ca. 11,200-11,050 cal BP (9,690 $^{14}$C BP, McLaren 2008). A small number of lithics were recovered from the lowest strata, including one cobble core, one spall tool, and a flake, all coarse-grained basalt. Faunal remains recovered from this site date to later components (post 7,000 cal BP) and include sea mammal and fish bones, as well as sea urchin, mussel, and clam remains (McLaren 2008). Other nearby shell midden sites on the Dundas Islands have returned dates of ca. 7,800 to 6,800 cal BP (7,000-6,000 $^{14}$C BP) (Martindale et al. 2009).

### *Kilgii Gwaay (1325T) – Site 14, 10,700 cal BP*

The site at Kilgii Gwaay is located on Ellen Island at the southern end of Haida Gwaii in British Columbia and includes a large faunal component as well as water-saturated artifacts (Fedje et al. 2005a; Cohen 2014; Cohen 2019). The stone tool assemblage at the site is large and is almost exclusively unifacial, including scrapers, gravers, burins, choppers, and discoidal cores.

Bone artifacts include barbed points, perforators, and pieces of bone exhibiting cutmarks and spiral fracture as a result of anthropogenic butchery (Fedje et al. 2005a). The organic artifacts recovered from the water-saturated component of the site include twine, withes wrapped with split spruce root, and wooden wedges, as well as numerous wood chips, stakes, and other pieces (Fedje et al. 2005a; Cohen 2014; Cohen 2019). Most of these wooden artifacts were manufactured from Sitka spruce and western hemlock, with some evidence of the use of alder and bitter cherry (*Prunus emarginata*) (Cohen 2014; Cohen 2019). The cultural component at the site is securely dated to 10,700 cal BP (Fedje et al. 2011). Faunal remains recovered from the site are extensive and demonstrate an emphasis on marine resources, including rockfish, Pacific spiny dogfish, lingcod, greenling, Pacific halibut (*Hippoglossus stenolepis*), salmon, Pacific herring, mussel, auklets (Alcidae), albatross (*Phoebastria*), cormorant (*Phalacrocorax*), goose, loon, harbour seal, northern sea lion, sea otter, and black bear (Fedje et al. 2005a). Collectively, the artifact and faunal assemblage suggest summer occupancy at the site (Fedje et al. 2011; Cohen 2014; Cohen 2019).

### *Ground Hog Bay 2 – Site 15, 10,683-10,303 cal BP*

Ground Hog Bay 2 is located on mainland Alaska to the west of Juneau and east of Glacier Bay. The site rests upon an elevated beach terrace and has yielded a basal lithic assemblage consisting of bifaces, scrapers, choppers, flake cores, gravers, and utilized flakes (Ackerman 1996). This lowest component is overlain by an assemblage containing microblades and microblade cores but lacking in bifaces. Many of the microblade cores from this site are frontally fluted and wedge-shaped and are similar in form to those found in central and southwestern Alaska from ca. 12,800-11,500 cal BP (11,000-10,000 $^{14}$C BP) (Ackerman 1996). X-ray fluorescence performed on obsidian from Ground Hog Bay 2 shows that the material is sourced to Mount Edziza in northern British Columba, indicating the presence of extensive trade or high mobility throughout the region at this time (Ackerman 1996). No preserved organic materials were recovered from this site. Basal sediments from the site date to 10,683-10,303 cal BP (Chatters et al. 2012).

### *Richardson Island (1127T) – Site 16, ca. 10,600-10,400 cal BP*

The Richardson Island site is located on the western side of Richardson Island, east of Moresby Island in southern Haida Gwaii, British Columbia (Storey 2008; Mackie et al. 2011; Waber 2011). Several features, including 11 post-moulds and numerous layered hearths (clustered within a 200-year use period) were uncovered at the site. The oldest of these hearth features was radiocarbon dated to ca. 10,600-10,400 cal BP (9,290 $^{14}$C BP, Storey 2008). The site yielded lithics of both the Kinggi Complex and the Early Moresby Tradition, including gravers/burins, spokeshaves, bifacial knives, projectile points, scrapers, choppers, spall tools, microblades, wedges, utilized flakes, abraders, cores, and hammerstones (McLaren and Smith 2008; Storey 2008; Waber 2011). Three bone point tips were also found during excavation. The floral and faunal assemblage recovered from the Richardson Island site is extensive, including rockfish, salmon, Pacific spiny dogfish, lingcod, flatfish, greenling, Pacific herring, and several species of mammals and birds as well as samples of western hemlock, pine, sedge, grass, blueberry/huckleberry (*Vaccinium* sp.), and saskatoon (*Amelanchier alnifolia*) (Storey 2008).

### *Hidden Falls – Site 17, 10,553-9,785 cal BP*

The Hidden Falls site is located on the eastern side of Baranof Island in the Alexander Archipelago of southeastern Alaska (Ackerman 1996). The site was found on an elevated beach terrace and has yielded a lithic assemblage comparable to that at Ground Hog Bay 2. Artifacts recovered from the Hidden Falls site include microblades and microblade cores, flake cores, bifaces, scrapers, choppers, gravers, and utilized flakes. Many of the microblade cores from this site are frontally fluted and wedge-shaped and are similar in form to those found in central and southwestern Alaska from ca. 12,800-11,500 cal BP (11,000-10,000 $^{14}$C BP) (Ackerman 1996).

No organic materials were recovered from this site. The lowest layers of this site date to 10,553-9,785 cal BP (Chatters et al. 2012).

### *Arrow Creek 1 (766T) and 2 (925T) – Site 18, ca. 10,500-10,200 cal BP*

Arrow Creek sites 1 and 2 are both situated in the Arrow Creek estuary in Matheson Inlet on the east coast of Moresby Island in southern Haida Gwaii, British Columbia (Fedje and Christensen 1999). Arrow Creek 1 is located on a raised alluvial fan above the modern tidal limit, and yielded a few scattered stone tools dating to ca. 9,830 cal BP (8,800 $^{14}$C BP), while the rest of the assemblage, including numerous microblades, scraper planes, and cores, date to somewhat more recently, between ca. 9,200 and 6,300 cal BP (between 8,200 and 5,500 $^{14}$C BP) (Fedje and Christensen 1999). Arrow Creek 2, located at the modern tidal limit nearby, also yielded stone tools, the barnacles on several of which were dated to between ca. 10,500-10,200 cal BP (9,500-9,200 $^{14}$C BP, Fedje and Christensen 1999). Faunal remains were not recovered from either of these sites, with organic remains limited to preserved charcoal (Fedje and Christensen 1999; Mackie et al. 2011).

### *Milliken (DjRi-3) – Site 19, ca. 10,500-9,650 cal BP*

Located on the eastern side of the lower Fraser River Canyon to the northeast of Yale, the Milliken site consists of two components, the earlier Milliken component ca. 10,500-9,650 cal BP (9,000 ± 150 $^{14}$C BP) and the later Mazama component (Mitchell and Pokotylo 1996). The Milliken component of the site consists of 450 artifacts, including foliate points and foliate point bases, bifacial and unifacial tools, bifacially and unifacially retouched flakes, pebble core tools, and several other tools types including a pebble graver, flaked slate and schist, ground and sawn stone objects, and several hammerstones (Mitchell and Pokotylo 1996). The lower layers of the site also contain several hearth features, nearly 100 stake and/or post holes, and a rock alignment or wall feature. Collectively, these have been proposed as infrastructure for a summer salmon fishing camp (Mitchell and Pokotylo 1996). Floral and faunal remains recovered from the Milliken component include at least thirty carbonized choke cherry (*Prunus virginiana*) pits and other unidentified seeds as well as one fragment of calcined mammal bone (Mitchell and Pokotylo 1996).

### *Rice Creek (CRG-234, 235, 238) – Site 20, ca. 10,250-10,100 cal BP*

The Rice Creek site is located on Heceta Island in southeastern Alaska and includes 11 flakes recovered from surface exposures in tree throws or along stream banks (Ackerman 1996). No diagnostic artifacts were recovered at this site, but the earliest sampled date for the assemblage is ca. 10,250-10,100 cal BP (9,000 $^{14}$C BP, Ackerman 1996).

### *Pruth Bay (EjTa-15) – Site 21, 10,151-9,924 cal BP*

Pruth Bay (EjTa-15) is located on the central coast of British Columbia, at the far western end of Kwakshua Channel on Calvert Island. Archaeological remains at the site include canoe runs and a scatter of chipped stone tools in the intertidal zone as well as thin shell midden deposits underlain by other non-shell cultural strata on the upper beach (McLaren et al. 2015). The lower strata contain a significant volume of charcoal, as well as 69 obsidian, basalt, chert, and andesite tools and flakes. The basal occupation layer at this site has yielded dates of 10,151-9,924 cal BP from a hearth feature (McLaren et al. 2015).

### *Lyell Bay South (1354T) and East (1355T) – Site 22, ca. 10,000-9,450 cal BP*

Lyell Bay South and Lyell Bay East are two raised beach sites in Lyell Bay on the western side of Lyell Island in southern Haida Gwaii, British Columbia (Fedje and Christensen 1999). Excavations at Lyell Bay South resulted in the collection of 610 lithic artifacts, including microblades, multidirectional and unidirectional cores, scrapers, and a cobble chopper, with the lowest levels of the excavation represented by a single biface thinning flake. The basal deposits at

Lyell Bay West date to ca. 9,550-9,450 cal BP (8,450 $^{14}$C BP, Fedje and Christensen 1999). Lyell Bay East yielded comparatively fewer lithics (274), including microblades, spokeshaves, cobble choppers, multidirectional cores, and flaked tools as well as a single biface fragment near the base of the excavations. The basal deposits of Lyell Bay East date to ca. 10,000-9,600 cal BP (8,800 $^{14}$C BP, Fedje and Christensen 1999).

### *Millard Creek (DkSf-2) – Site 23, ca. 9,650-8,650 cal BP*

The Millard Creek site is located on the southwestern side of Comox Harbour, at the mouth of Millard Creek to the south of Courtenay/Comox on central Vancouver Island. Cultural remains excavated at the site include extensive shell midden deposits as well as human remains, which have since been reburied there (BC Archaeology Branch 2019). The oldest component at the site, within Locality A, dates to ca. 9,650-8,650 cal BP (8,300 ± 200 $^{14}$C BP, BC Archaeology Branch 2019). Artifacts recovered from the site during initial excavations at the site include a stemmed point, choppers, possible scrapers, and several pieces of worked bone (Capes 1970). Later excavations at the site yielded sandstone abraders, hammerstones, bone awls, and antler wedges (Grant 2004).

### *Glenrose Cannery (DgRr-6) – Site 24, ca. 9,550-8,450 cal BP*

The Glenrose Cannery site is located on the Lower Mainland of British Columbia, on the southern side of the Fraser River across from Annacis Island. The palaeogeography of the site indicates that the Panorama Ridge area adjacent to the archaeological deposits was probably a peninsula extending into the Strait of Georgia adjacent to the mouth of the Fraser River at the time of occupation (Matson 1996). This suggests that at ca. 9,550-8,450 cal BP (8,150 ± 250 $^{14}$C BP), the site was not inland as it now appears (due to the growth of the Fraser River delta) but was instead coastal (Matson 1996). Generally, the organic materials from the early component at this site are poorly preserved, so the assemblage mostly consists of lithics with limited antler and bone artifacts (Matson 1996). In the early Old Cordilleran component of the site 614 artifacts were recovered, including 269 cobble tools (mostly unifacial choppers and cortex spalls with some bifacial tools, scraper planes, and hammerstones), four complete foliate projectile points, six leaf-shaped knives, and one stemmed point, along with large crude bifaces, many retouched unifacial flakes, and several abrader stones and miscellaneous ground stone pieces (Matson 1996). The bone and antler assemblages consist of 11 antler wedges, one antler punch, and one barbed antler point, as well as three needle point bone awls, one ulna tool, one bone blade, and six worked bone fragments. The faunal assemblage from the oldest component at the Glenrose Cannery site consists of elk, deer, *Canis sp*., beaver, seal, salmon, starry flounder, eulachon, stickleback, peamouth, sturgeon, bay mussel, and clam (Matson 1996).

### *Chuck Lake (CRG-237) – Site 20, ca. 9,500-9,000 cal BP*

The Chuck Lake site is located on Heceta Island in southeastern Alaska and consists of several isolated scatters of lithic debitage as well as several shell deposits (Ackerman 1996). The oldest date from the site is ca. 9,500-9,000 cal BP from midden deposits (8,220 ± 125 $^{14}$C BP, Ackerman 1996). The artifact assemblage from this site includes microblades and microblade cores, cobble flake cores, scrapers, hammerstones, whetstones, cobble spall tools, flakes, and a barbed bone point. Faunal material identified in the midden deposits mostly consists of shellfish including clams, cockles, mussels, barnacles, and marine gastropods, but also includes cod, greenling, sculpin, and rockfish as well as beaver, sea lion, deer, cormorant, grebe, and loon (Ackerman 1996).

### *Two-Flake Terrace (GbTo-185/P011-1) – Site 25, 9,304-9,028 cal BP*

Two-Flake Terrace (GbTo-185/P011-1) is a site located on a raised terrace at the northern end of Digby Island to the west of Prince Rupert on the north coast of British Columbia (Letham et al. 2016). The site area, located 10-12 m above modern sea level, was delineated through shovel testing and probing, and contains evidence of repeated use of hearth features and stone tool-making dating to 9,304-9,028 cal BP (Letham et al. 2016).

### *Bear Cove (EeSu-8) – Site 26, 8,900 cal BP*

Bear Cove is located on the eastern shore of Port Hardy on Northern Vancouver Island, with radiocarbon dates extending back to 8,900 cal BP (8,020 ± 110 $^{14}$C BP, C. Carlson 2003). Excavations conducted in the late 1970s at the site uncovered intact shell midden deposits underlain by non-shell cultural strata, including brown and black silts and gravels (C. Carlson 1979; C. Carlson 2003). The earliest dates at the site are derived from charcoal in these lower deposits, from which numerous chipped stone artifacts were recovered, assigned to the Pebble Tool Tradition by C. Carlson (2003). In addition, the early component of the site (8,900-4,900 cal BP) contains well-preserved faunal remains, including rockfish, ratfish, greenling, herring, sculpin, dogfish, flatfish, cod, salmon, deer, beaver, otter, dolphin, seal, sea lion, duck, gull, loon, cormorant, heron, murre, *Corvus sp.*, and bald eagle (C. Carlson 2003).

### *Lawn Point (FiTx-3) – Site 27, ca. 8,450-7,950 cal BP*

The site at Lawn Point (FiTx-3) is a raised beach site on the eastern side of Graham Island in central Haida Gwaii, British Columbia. Most of the lithic assemblage from the site, including 551 artifacts and 2,225 flakes, is characterized by pebble and flake tools, pebble cores, and microblades and microblade cores (Fladmark 1986; Fladmark 1989). The deepest radiocarbon date available from the site places it at ca. 8,450-7,950 cal BP (7,200 $^{14}$C BP), though sea level history of the area suggests that the site may have been in use 1,000-2,000 years earlier (Fedje et al. 2005c).

### *Thorne River (CRG-177) – Site 28, ca. 8,400-8,200 cal BP*

The Thorne River site is located on the eastern side of Prince of Wales Island in southeastern Alaska. Artifacts recovered from the site include microblades and microblade cores, flakes, flake cores, pebble tools, burins, and a few bifaces, with basal dates of the deposits at ca. 8,400-8,200 cal BP (7,520 ± 59 $^{14}$C BP, Ackerman 1996). The microblade cores, burins, and bifaces from this site are generally manufactured from obsidian (Ackerman 1996).

### *Argonaut Hill Forks (GaTw-9) – Site 29, ca. 8,000-7,900 cal BP*

The Argonaut Hill Forks site (GaTw-9) is a subsurface lithic scatter site located on the northeastern side of Graham Island in northern Haida Gwaii, British Columbia. Evaluative testing of three terraces around this site yielded 28 stone artifacts, including choppers, spall tools, and flakes manufactured from locally available raw materials (Sanders 2009). Four radiocarbon dates indicate that the earliest occupation of the site dates to at least ca. 8,000-7,900 cal BP (7,140 ± 20 $^{14}$C BP, Sanders 2009).

# Appendix C: Proxy Data from Little Woss Lake and Topknot Lake

## Pollen Count Data

All the data that follow are raw counts from each sample, listed by core depth in cm LD (Little Woss Lake) and cm dbp (Topknot Lake). The number of exotic *Lycopodium* spores counted for a particular sample is listed at the bottom of each table along with the pollen count totals. These totals do not include the number of exotics counted.

Little Woss Lake

| **Taxa** | **172** | **187** | **202** | **217** | **232** | **247** | **262** | **277** | **292** | **302** | **307** | **310** | **312** | **314** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *Pinus* sp. | 156 | 91 | 112 | 168 | 150 | 132 | 170 | 126 | 96 | 171 | 234 | 250 | 107 | 100 |
| *Picea* sp. |  | 20 | 33 | 15 | 8 | 14 | 8 | 13 | 16 | 3 | 2 |  | 1 | 4 |
| *Abies* sp. | 2 | 29 | 8 | 17 | 14 | 10 | 29 | 14 | 54 | 36 | 11 | 8 | 93 | 127 |
| *Tsuga heterophylla* | 10 | 74 | 74 | 79 | 69 | 24 | 17 | 9 | 7 | 5 | 2 | 8 | 1 | 2 |
| *Tsuga mertensiana* | 2 | 4 | 29 | 22 | 27 | 20 | 15 | 11 | 26 | 24 | 14 | 10 |  | 4 |
| *Pseudotsuga* | 7 | 1 |  |  | 1 | 1 | 1 |  |  |  |  |  |  |  |
| Unknown conifer |  |  |  |  |  |  |  |  | 4 | 4 | 3 |  |  | 9 |
| Cupressaceae | 2 |  |  |  | 2 |  |  |  |  |  |  |  | 1 |  |
| *Populus* | 3 | 6 | 2 | 3 | 5 | 2 |  | 3 | 1 | 3 | 3 |  |  | 1 |
| *Betula* | 1 | 1 |  |  |  |  |  |  |  |  |  |  |  |  |
| *Alnus* sp. | 162 | 142 | 64 | 93 | 67 | 108 | 111 | 131 | 135 | 104 | 103 | 89 | 86 | 66 |
| *Salix* | 13 | 6 | 1 | 1 | 1 | 5 | 2 | 2 | 1 | 6 | 1 | 3 | 2 | 1 |
| *Myrica* |  |  |  |  |  | 3 |  |  | 1 | 1 |  |  |  |  |
| Ericaceae Type 1 (>36 µm) | 4 | 2 |  |  | 1 |  |  |  |  |  |  |  |  |  |
| Ericaceae Type 2 (30-36 µm) |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| Ericaceae Type 3 (<30 µm) |  | 1 |  |  |  |  |  |  | 2 |  |  |  |  | 1 |
| Rosaceae small |  |  |  |  | 1 |  |  |  | 1 |  |  |  |  |  |
| Rosaceae large |  |  | 1 |  | 1 |  |  |  |  |  |  |  |  |  |
| *Sambucus* | 1 |  |  |  |  |  |  |  |  |  |  |  |  |  |

| **Taxa** | **172** | **187** | **202** | **217** | **232** | **247** | **262** | **277** | **292** | **302** | **307** | **310** | **312** | **314** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *Cornus* | | | | | | 1 | | | | | | | | |
| Poaceae | 2 | | | 1 | | 6 | 3 | 1 | 1 | | | | | |
| Asteraceae Tub. | | | 1 | | | 2 | | | | | | | | |
| *Ambrosia* | 1 | | | | | | | | | | | | | |
| *Artemisia* | | | | | | | | | | | 1 | 2 | | |
| *Plantago* | | | 1 | | | | | | | | | | | |
| *Lysichiton* | 1 | | | | | | | | | | | | | |
| Unknown herbaceous | | 1 | | 1 | | | | | | | | | 1 | |
| Cyperaceae | 4 | 3 | 4 | 1 | 1 | 22 | 10 | 2 | 2 | | 2 | 1 | 5 | 3 |
| *Typha* | | 2 | | | | | | | | 1 | | | | 1 |
| *Menyanthes* | | 2 | | | | | | | | | | | | |
| *Potamogeton* | 2 | 3 | | | | | 4 | 6 | | | 1 | | | |
| *Nuphar* | | 2 | | 1 | 1 | 2 | 2 | 1 | 2 | 1 | 6 | 1 | 1 | 1 |
| *Isoetes* | | | | | | | | 1 | | | | | | |
| *Pteridium* | 7 | 5 | | | | 4 | | | 2 | | | | | |
| Cryptogram | | | 1 | 1 | | 1 | 3 | 1 | | | | | 1 | 1 |
| Trilete smooth | | 1 | | | 1 | 5 | 6 | 2 | | 1 | 3 | 1 | 2 | |
| Monolete fern - verrucate | | | 2 | | | 1 | | | | 1 | | | 3 | |
| Monolete fern - rugulate | | | 1 | | | | | | | | | | | |
| Monolete fern - undifferentiated | | 6 | 3 | 3 | 2 | 6 | 13 | 8 | 17 | 10 | 9 | 9 | 6 | |
| *Equisetum* | 12 | 5 | | 3 | 4 | 3 | 6 | 5 | 1 | | 4 | | | 1 |
| *Pityrograma* | | | | | | | | | | | | | 1 | |
| **Exotic Marker (*Lycopodium*)** | **45** | **142** | **49** | **50** | **36** | **35** | **51** | **34** | **102** | **47** | **91** | **95** | **40** | **40** |
| **Total (-Exotics)** | **392** | **407** | **337** | **409** | **356** | **372** | **400** | **336** | **369** | **371** | **399** | **382** | **311** | **322** |

### Topknot Lake

| **Taxa** | **538** | **589** | **614** | **649** | **700** | **710** | **799** | **837.5** | **875** | **910** | **935** | **1025** | **1120** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *Pinus* sp. | 19 | 6 | 217 | 10 | 22 | 37 | 69 | 8 | 24 | 11 | 11 | 3 | 4 |
| *Picea* sp. | 4 | 31 | 20 | 1 | 5 | 1 | 1 |  | 3 | 3 | 49 | 6 | 4 |
| *Abies* sp. | 56 | 4 | 14 | 2 |  | 7 | 21 | 1 | 5 | 4 | 11 | 3 | 3 |
| *Tsuga heterophylla* | 20 | 1 | 3 | 2 | 3 |  | 2 |  | 1 | 2 | 3 | 1 |  |
| *Tsuga mertensiana* |  | 6 | 10 | 4 | 3 | 3 | 7 | 2 | 3 | 2 | 10 |  |  |
| *Pseudotsuga* | 1 |  |  | 1 |  |  |  |  |  |  |  |  |  |
| Unknown conifer |  |  | 3 | 2 | 3 | 2 |  |  |  |  | 6 |  |  |
| Cupressaceae |  | 2 | 2 | 3 |  | 2 | 2 |  |  | 1 |  |  |  |
| *Populus* |  |  | 3 | 1 |  |  | 2 |  |  |  |  |  |  |
| *Betula* |  |  |  |  |  |  |  |  | 1 |  |  |  |  |
| *Alnus* sp. | 89 | 82 |  | 4 | 6 | 6 | 1 |  |  | 4 | 6 | 4 | 1 |
| *Salix* | 3 |  | 9 | 13 | 8 | 14 | 8 | 7 | 40 | 178 | 5 | 1 | 1 |
| Ericaceae Type 1 (>36 µm) |  |  |  | 6 | 1 | 4 | 8 | 1 |  | 1 | 2 |  |  |
| Ericaceae Type 2 (30-36 µm) |  |  |  | 4 | 2 | 10 | 13 | 1 | 1 | 1 |  |  |  |
| Ericaceae Type 3 (<30 µm) |  |  |  | 9 | 7 | 7 | 6 | 2 | 2 | 3 | 1 |  | 2 |
| *Empetrum* |  |  |  |  | 1 | 1 | 1 |  | 1 |  | 2 |  |  |
| Rosaceae small |  | 1 |  | 5 | 1 | 3 |  |  | 2 | 1 | 2 |  |  |
| Rosaceae large |  | 2 |  | 2 | 3 | 3 | 2 | 1 | 7 | 8 |  |  |  |
| *Sambucus* |  |  |  |  |  |  |  |  |  |  |  |  |  |
| *Shepherdia* |  |  |  | 1 | 1 |  |  |  |  |  |  |  |  |
| Caprifoliaceae |  |  |  |  |  | 1 | 1 |  |  |  |  |  |  |
| *Cornus* |  |  |  | 1 |  |  |  |  |  |  |  |  |  |
| Poaceae |  |  | 1 | 63 | 59 | 79 | 65 | 23 | 22 | 44 | 17 | 6 | 4 |
| Asteraceae Tub. |  |  |  | 17 | 17 | 11 | 15 | 1 | 2 | 12 | 1 |  |  |
| Asteraceae Lig. |  |  |  |  |  | 1 | 2 |  |  |  |  |  |  |

| **Taxa** | **538** | **589** | **614** | **649** | **700** | **710** | **799** | **837.5** | **875** | **910** | **935** | **1025** | **1120** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *Ambrosia* | | | | | | | | | | 1 | | | |
| *Artemisia* | | 1 | 1 | 6 | 11 | 7 | 8 | 3 | 2 | 7 | 2 | 1 | 2 |
| Saxifragaceae | | | | 3 | | | | | | | | | |
| *Thalictrum* | | | | | 2 | 4 | 4 | | | | | | |
| Caryophyllaceae | | | | 7 | 2 | 2 | 3 | 1 | 2 | 10 | 1 | | |
| Onagraceae | | | | | 2 | | 2 | | | | | | |
| *Ranunculus* | | | | 3 | 2 | 4 | 6 | | | 4 | 5 | | |
| *Pedicularis* | | | | | | | | | | 1 | | | |
| Malvaceae | | | 1 | | | 1 | 1 | | | | | | |
| Brassicaceae | | | | 1 | | | 1 | | 4 | 1 | | | |
| Polygonaceae | | | | | 1 | | | | | | | | |
| Liliaceae | | | 1 | | 1 | | | | | 5 | | | |
| Apiaceae | | | 1 | 2 | 3 | 3 | 2 | | 1 | 2 | | | |
| *Heracleum* | | | | | 4 | 5 | | | 1 | | | | |
| *Gentiana* | | | | | | | | | | 1 | | | |
| *Sanguisorba* | | | 1 | 3 | 10 | 20 | 12 | 6 | 1 | 6 | | | 1 |
| *Polemonium* | | | | | 3 | | | | 1 | 3 | | | |
| Fabaceae | | | | | | | | 1 | 4 | 12 | | | 1 |
| Lamaiceae/Gal. | | | | | | | | | | 1 | | | |
| *Triglochin* | | | | | | | 1 | | | | | | |
| Unknown herbaceous | 1 | | 1 | 11 | 2 | 6 | 3 | 3 | 10 | 13 | 3 | 2 | 1 |
| Cyperaceae | 8 | 6 | 15 | 82 | 78 | 106 | 89 | 23 | 32 | 27 | 11 | | 1 |
| *Typha* | | | 2 | | | | | 1 | | | | | |
| *Potamogeton* | 4 | 132 | 5 | | | | | | | | | | |
| *Nuphar* | 17 | | 2 | | | | | | | | | | |
| *Selaginella* | | | | | 1 | | | | | | | | |
| *Lycopodium* | | | | | | | | | 1 | | 3 | 2 | |

| **Taxa** | **538** | **589** | **614** | **649** | **700** | **710** | **799** | **837.5** | **875** | **910** | **935** | **1025** | **1120** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *Isoetes* | | | | | | | | | 3 | | | | |
| *Pteridium* | 4 | | | 2 | | | | | | | | | |
| Cryptogram | 2 | 1 | 1 | 1 | 2 | 3 | 1 | | 1 | | | | |
| Trilete smooth | 3 | 2 | 2 | 3 | 8 | 2 | 1 | | 2 | 1 | 1 | 1 | |
| Trilete rugulate | | | | | | | 1 | | | | | | |
| Monolete fern - verrucate | 15 | 5 | 7 | 1 | 1 | | 1 | | | 1 | 2 | | |
| Monolete fern - rugulate | 11 | 5 | 2 | 3 | 2 | | | | | | | | |
| Monolete fern - undifferentiated | 63 | 26 | 58 | 4 | 11 | 2 | 4 | | 4 | | 10 | | 1 |
| *Equisetum* | 3 | 1 | 2 | 1 | 5 | | | | 1 | | | | |
| *Sphagnum* | | | | | 1 | | | | | | | | |
| Unknown | | | | 12 | 11 | 7 | 13 | 2 | 5 | 1 | | | |
| **Exotic Marker (*Lycopodium*)** | **105** | **70** | **55** | **371** | **355** | **369** | **351** | **221** | **1982** | **3054** | **3925** | **2509** | **2616** |
| **Total (-Exotics)** | **323** | **314** | **384** | **296** | **305** | **364** | **379** | **87** | **189** | **372** | **164** | **30** | **26** |

## eDNA Taxa Identification

The data below are grouped by clade, either Viridiplantae (plants) or Metazoa (animals). Scientific names are listed based on aligned reads to genomic databases including NCBI Nucleotide, RefSeq, and the Genome Taxonomy Database (GTDB). ‘Hits’ lists the number of times that a sequence from the sample matched a known sequence in one of the genomic databases with no deviation (edit distance 0). The column on the right lists whether I considered the hit a contaminant, either based on lab/sample contamination or a poor genomic reference sequence. Only species that were not considered contaminants were included in the results.

Little Woss Lake, 312cm LD

| Group | Scientific name | Common name/type | Hits | Contaminant? |
|---|---|---|---|---|
| Viridiplantae | *Nephroselmis olivacea* | green algae | 1 | Possible |
| Viridiplantae | *Choricystis* | green algae | 1 | Possible |
| Viridiplantae | *Blidingia* sp. | green algae | 1 | Possible |
| Viridiplantae | *Pinus contorta* | shore pine | 1 | No |
| Viridiplantae | *Momordica charantia* | bitter melon | 1 | Yes |
| Viridiplantae | *Ricinus communis* | castor bean | 3 | Yes |
| Viridiplantae | *Populus* | cottonwood | 1 | No |
| Viridiplantae | *Populus trichocarpa* | black cottonwood | 1 | No |
| Viridiplantae | *Brassica* | mustards | 1 | Possible |
| Viridiplantae | *Gossypium hirsutum* | Mexican cotton | 1 | Yes |
| Viridiplantae | Potamogetonaceae | pondweed family | 2 | No |
| Viridiplantae | *Potamogeton* | pondweeds | 3 | No |
| Viridiplantae | *Potamogeton gramineus* | grassy pondweed | 4 | No |
| Viridiplantae | *Potamogeton praelongus* | whitestem pondweed | 1 | No |
| Viridiplantae | *Stuckenia pectinata* | sago pondweed | 1 | No |
| Viridiplantae | *Phormium tenax* | New Zealand flax | 1 | Yes |
| Viridiplantae | Cyperaceae | sedges | 1 | No |
| Viridiplantae | Poaceae | grasses | 1 | No |

| **Group** | **Scientific name** | **Common name/type** | **Hits** | **Contaminant?** |
|---|---|---|---|---|
| Viridiplantae | Nymphaceae | water lilies | 1 | No |
| Viridiplantae | *Nuphar* | yellow pond-lily | 1 | No |
| Metazoa | *Dicentrarchus labrax* | European bass | 1 | Yes |
| Metazoa | *Oncorhynchus tshawytscha* | Chinook salmon | 42 | No |
| Metazoa | Catarrhini | old world monkeys | 1 | Yes |
| Metazoa | *Ursus arctos horribilis* | grizzly bear | 1 | No |
| Metazoa | *Lipotes vexillifer* | baiji | 171 | Yes |
| Metazoa | *Meleagris gallopavo* | wild turkey | 2 | Yes |
| Metazoa | *Parastrongyloides trichosuri* | possum-specific nematode worm | 1 | Possible |
| Metazoa | *Necator americanus* | American hookworm | 1 | Possible |
| Metazoa | *Ixodes scapularis* | deer tick | 1 | No |
| Metazoa | *Spodoptera litura* | tobacco cutworm | 15 | Yes |
| Metazoa | *Plutella xylostella* | diamondback moth | 1 | Yes |
| Metazoa | *Dendroctonus ponderosae* | mountain pine beetle | 5 | No |
| Metazoa | *Culex quinquefasciatus* | southern house mosquito | 1 | Yes |
| Metazoa | *Trichogramma pretiosum* | parasite wasps | 1 | Yes |

### Topknot Lake, 875 cm dbp

| Group | Scientific name | Common name/type | Hits | Contaminant? |
|---|---|---|---:|---|
| Viridiplantae | Pinaceae | pine family | 2 | No |
| Viridiplantae | *Abies* | true firs | 2 | No |
| Viridiplantae | *Abies balsamea* | balsam fir | 1 | No |
| Viridiplantae | *Picea glauca* | white spruce | 2 | No |
| Viridiplantae | *Actinidia* | kiwi fruit | 1 | Yes |
| Viridiplantae | *Polemonium boreale* | northern Jacob's-ladder | 1 | No |
| Viridiplantae | Lamiids | sub-aster family | 1 | No |
| Viridiplantae | *Bistorta officinalis* | common bistort | 1 | No |
| Viridiplantae | *Fallopia multiflora* | tuber fleeceflower | 1 | Yes |
| Viridiplantae | *Cucumis melo* | muskmelon | 1 | Yes |
| Viridiplantae | *Populus* | cottonwood | 2 | No |
| Viridiplantae | *Populus trichocarpa* | black cottonwood | 2 | No |
| Viridiplantae | *Rubus* | various berries | 1 | No |
| Viridiplantae | *Salix* | willow | 1 | No |
| Viridiplantae | *Elaeis guineensis* | African oil palm | 1 | Yes |
| Viridiplantae | Triticeae | wheats | 1 | Yes |
| Viridiplantae | Triticinae | wheats | 2 | Yes |
| Viridiplantae | *Triticum aestivum* | common wheat | 1 | Yes |
| Metazoa | *Pygocentrus nattereri* | red-belled piranha | 1 | Yes |
| Metazoa | Haplorrhini | dry-nose primates | 1 | Yes |
| Metazoa | Simiiformes | monkeys/apes | 11 | Yes |
| Metazoa | Catarrhini | old world monkeys | 20 | Yes |
| Metazoa | Hominoidea | apes | 29 | Yes |
| Metazoa | *Pan troglodytes* | chimpanzee | 1 | Yes |
| Metazoa | *Lipotes vexillifer* | baiji | 2 | Yes |

| **Group** | **Scientific name** | **Common name/type** | **Hits** | **Contaminant?** |
|---|---|---|---|---|
| Metazoa | *Sus scrofa* | wild boar | 6 | Yes |
| Metazoa | *Meleagris gallopavo* | wild turkey | 1 | Yes |
| Metazoa | *Gallus gallus* | red junglefowl (chicken) | 2 | Yes |
| Metazoa | *Dendroctonus ponderosae* | mountain pine beetle | 5 | No |

## Diatom Count Data

All the data that follow are raw counts from each sample, listed by core depth in cm LD (Little Woss Lake) and cm dbp (Topknot Lake). Overall total as well as the totals for each halobian class are summed at the base of each table.

Little Woss Lake

| Taxa | Halobian Class | 307 | 314 |
|---|---|---|---|
| *Brachysira vitrea* | 2 | 4 | |
| *Brachysira brebissonii* | 2 | 4 | |
| *Cocconeis pseudothumensis* | 2 | | 8 |
| *Encyonema minuta* | 2 | 4 | 1 |
| *Encyonema hebridicum* | 2 | 1 | 3 |
| *Encyonema silesiacum* | 2 | 2 | |
| *Fragilaria construens* | 2 | 5 | 8 |
| *Gomphonema subtile* | 2 | | 2 |
| *Gomphonema truncatum* | 2 | 1 | |
| *Navicula cryptotenella* | 2 | | 5 |
| *Navicula rhynchocephala* | 2 | | 1 |
| *Navicula staffordiae* | 2 | | 11 |
| *Navicula aurora* | 2 | 5 | 13 |
| *Neidium iridis* | 2 | 1 | |
| *Pinnularia brebissonii* | 2 | 3 | 5 |
| *Pinnularia maior* | 2 | | 2 |
| *Pinnularia turfosiphila* | 2 | | 3 |
| *Rhopalodia gibba* | 2 | 2 | 2 |
| *Sellaphora laevissima* | 2 | 5 | 10 |
| *Sellaphora pupula* | 2 | 3 | 6 |
| *Stauroneis anceps* | 2 | | 2 |
| *Stauroneis phoenicenteron* | 2 | 3 | 1 |
| *Encyonopsis descriptiformis* | 3 | 35 | |
| *Amphorilia* | 3 | 3 | |
| *Nitzschia semirobusta* | 3 | 25 | |
| *Epithemia adnata* | 3 | 1 | |
| *Epithemia sorex* | 3 | 1 | |
| *Martyana martyi* | 3 | 1 | 5 |
| *Navicula schweigeri* | 3 | | 3 |
| *Pseudostaurosira brevistrata* | 3 | 7 | 15 |
| *Pseudostaurosira parasitica* | 3 | 2 | 23 |
| *Rhoicosphenia abbreviata* | 3 | 10 | |
| *Staurosirella lapponica* | 3 | 1 | |
| *Staurosirella leptostauron* | 3 | 5 | 20 |
| *Staurosirella leptostauron dubia* | 3 | 2 | |

| **Taxa** | **Halobian Class** | **307** | **314** |
|---|---|---|---|
| *Denticula kuetzingii* | 4 | | 5 |
| *Denticula valida* | 4 | 3 | 1 |
| *Rhopalodia musculus* | 4 | | 3 |
| **Total Halobian 1** | | **0** | **0** |
| **Total Halobian 2** | | **43** | **83** |
| **Total Halobian 3** | | **93** | **66** |
| **Total Halobian 4** | | **3** | **9** |
| **Total Halobian 5** | | **0** | **0** |
| **Total** | | **139** | **158** |

Topknot Lake

| **Taxa** | **Halobian Class** | **875** | **910** |
|---|---|---|---|
| *Cocconeis placentula* | 2 | | 3 |
| *Cymbella muelleri* | 2 | | 1 |
| *Eunotia lapponica* | 2 | | 1 |
| *Fragilaria capucina* | 2 | | 1 |
| *Gyrosigma acuminatum* | 2 | 2 | |
| *Rhopalodia gibba* | 2 | 1 | 1 |
| *Stephanodiscus oregonicus* | 2 | | 1 |
| *Synedra ulna* | 2 | | 1 |
| *Tabellaria floculosa,fenestrata* | 2 | | 1 |
| *Amphora libyca* | 3 | 2 | |
| *Epithemia adnata* | 3 | 1 | 7 |
| *Epithemia sorex* | 3 | 1 | 7 |
| *Fragilariforma virescens* | 3 | | 8 |
| *Fragilaria mesolepta* | 3 | | 2 |
| *Fragilaria tenera* | 3 | | 1 |
| *Gyrosigma attenuatum* | 3 | | 4 |
| *Pseudostaurosira brevistrata* | 3 | 2 | 2 |
| *Campylodiscus hibernicus* | 4 | | 34 |
| *Tabularia tabulata* | 4 | | 1 |
| *Coscinodiscus radiatus* | 5 | | 1 |
| **Total Halobian 1** | | **0** | **0** |
| **Total Halobian 2** | | **3** | **10** |
| **Total Halobian 3** | | **6** | **31** |
| **Total Halobian 4** | | **0** | **35** |
| **Total Halobian 5** | | **0** | **1** |
| **Total** | | **9** | **77** |