Paleoenvironmental Change on the Olympic Peninsula, Washington:

Forests and Climate from the Last Glaciation to the Present

Daniel G. Gavin[1], David M. Fisher[1], Erin M. Herring[1],
Ariana White[1], and Linda B. Brubaker[2]

1. Department of Geography, University of Oregon, Eugene, Oregon.

2. School of Environment and Forest Sciences, University of Washington, Seattle, Washington.

FINAL REPORT TO OLYMPIC NATIONAL PARK

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Table of Contents

List of Figures

List of Tables

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I would be remiss not to mention the excellent recent revision of *Olympic National Park: A Natural History* by Tim McNulty (2009). I am grateful for Tim McNulty's astute observation, excellent research, and detailed factual presentation. McNulty's book provided the ideal springboard for several sections of this report that are afield from my own expertise.

1. Introduction

> "Understanding present-day ecosystems must include the concept of time. The fullest understanding of this world, its origin, current status, trends and future will come from the seamless integration of neobiology with paleobiology, and a knowledge of the environmental dynamics that drive the process of change."
>
> – Alan Graham, *A Natural History of the New World*, 2011

The landscape of the Olympic Peninsula is impressive from every angle. A distinct geographic area bordered by the Pacific Ocean, the Strait of Juan de Fuca, and the Puget Lowlands, it contains high-relief mountains, a gradient of rainfall unmatched anywhere in the coterminous United States, forests capable of producing some of the largest trees in the world, and a distinct biota marked by endemic species. The peninsula is island-like in some respects, with a lowland connection in the south, but in other respects it functions as the southern end of a temperate rainforest ecosystem that extends along the archipelago of coastal islands of British Columbia and southeast Alaska. It is this geography that has resulted in the accumulation, and exclusion, of the diversity of life arrayed across the peninsula and on full display within the Olympic National Park.

The complex arrangement of ecosystems within the Olympic Peninsula is normally described as a product of the regional climate and geology. However, at a finer resolution, say, of a tributary of a major river, one may see that the exact patterns of forest cover is a result of a history of fires on certain years and windthrow on other years (D. L. Peterson, E. G. Schreiner, et al. 1997). Similarly, if one focuses on a single species, the pattern of its presence and absence over space may be as much a factor of environmental gradients as of its recent history of dispersal and interactions with other species. Thus, making sense of the biotic patterns on the peninsula, or any geographic area, requires a historical perspective. What events, especially those related to past climate changes and disturbances, led to the occurrence of the particular set of species and their distributions today? Obtaining answers to this question is never more important than it is today. The rapid climate changes currently underway are unlike any during the last 11,000 years (Marcott et al. 2013), but the potential impact of these climate changes on the biota is difficult to gauge (McMahon et al. 2011).

Paleoecology is the study of past environments using fossil evidence. Rather than only reconstructing past ecosystems (i.e., species composition, habitat types, etc.), paleoecologists address ecological questions on timescales longer than observable in human lifespans (Schoonmaker and Foster 1991). The use of fossil data creates a set of unique challenges with respect to interpreting fossil representation of past organisms and ecosystems (the field of taphonomy). However, the payback of such studies is large. The changes in the climate and vegetation over millennial time scales are fundamental background information on the assembly of the natural communities of a region. As retrospective studies are very labor intensive, knowledge is accumulated in steps, with detailed studies of information-rich sites providing the largest gains. The Olympic Peninsula has no shortage of important paleoecology studies (especially paleovegetation from pollen records) and the larger region has several robust paleoclimate reconstructions.

Several factors contribute to the difficulty of understanding biotic response to climate change and other agents of change in ecosystems, but paleoecology can effectively address some of these challenges. For example, species responses to climate change are mediated by changing disturbance regimes. Paleoecological studies in the western US have been addressing this issue by using detailed charcoal stratigraphy of fire events in conjunction with the pollen record (Whitlock et al. 2008). Paleoecological studies also provide the context for the persistence of biotic hotspots. Mountain regions may maintain biodiversity through significant climate change in 'refugia': locations where components of diversity retreat to and expand from during periods of unfavorable climate (Keppel et al. 2012). The modern biogeography of the Olympic Peninsula suggests these mountains were an important refugium south of the ice sheet during the glaciation. Such studies can also inform on the resilience of ecosystems to disturbances in the past, showing how many ecosystems recover quickly while others may not (Willis et al. 2010).

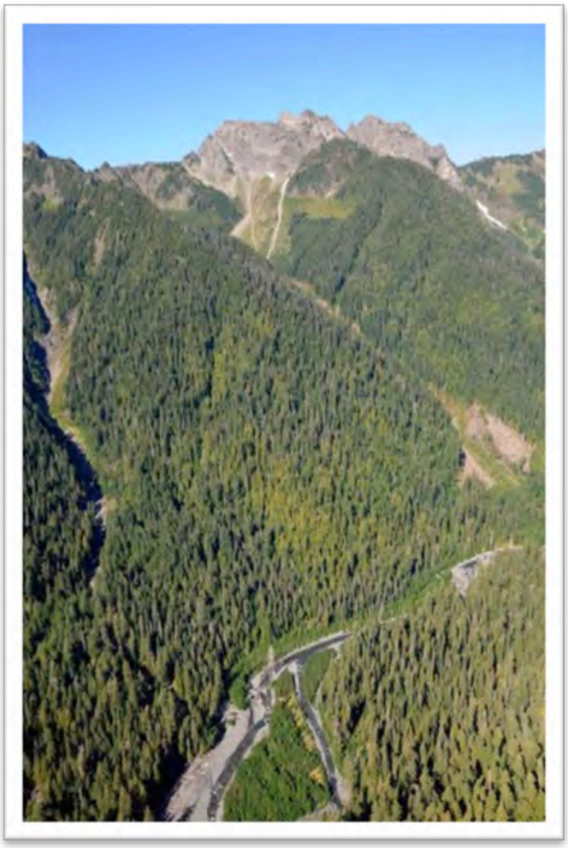

Figure 1. Typical forests on a south-facing aspect in the western portion of Olympic National Park, showing an elevation range from 300 m (Queets River at river-mile 43) to the summits of Thor and Woden peaks (1750 m elevation). Forest types include riparian early-successional red alder, avalanche-track Sitka alder, and old growth forests of western hemlock, western redcedar, Pacific silver fir and mountain hemlock. Photo by Larry Workman. (https://ssl.panoramio.com/photo/14517121).

1.1 Aims of this report

This report aims to bring together decades of research on the modern natural environment of the region, then review past research on paleoenvironmental change since the Late Pleistocene, and finally present five sediment records of changing forest composition and fire over the last 14,000 years. The last 14,000 years begins with the first major warming coming out of the Last Glacial Maximum and the retreat of the continental ice sheet, shortly followed by the establishment of continuous forests by 13,000 years ago. This period is easily studied due to the abundance of geologic records, from lake sediment, ocean sediments, and soils. During this 14,000-year period, there have been abrupt climate changes, forests changing composition rapidly, and large changes in fire regimes. Several tree species did not make their first appearance until several millennia after the ice sheet retreated, while others may have been present in small refugia close to the ice-sheet margin. Tree species that are associated with specific elevational zones today have moved elevationally due to changing climate, but sometimes in unexpected ways. The occurrence of fire, currently a minor component of natural disturbances on the Olympic Peninsula, was at times much greater than it is today. All of these dynamics provide many opportunities for understanding the changing climate and its potential future impacts. While the report uses a variety of data sources, a major focus is on pollen records. Maps and other data graphics are used extensively to illustrate points.

The setting of the Olympic Peninsula is presented in Chapter 2. The modern climate gradients on the Peninsula and recent trends in climate over the past 100 years demonstrate how climate change is beginning to accelerate on the peninsula. The vegetation of the peninsula is described with respect to five major forest zones and their association with climatic gradients. Last, the natural disturbance regime is reviewed, including fire, wind, insects, and geomorphic disturbances.

Chapter three presents the geological history and the historical development of biodiversity on the Olympic Peninsula. The long-term geologic history, beginning in the middle Cenozoic, sets the stage for the origin of the Olympic Mountains. What climate changes accompanied the development of the Olympic Mountains during the past 30 million years? How did this region become dominated by some of the largest conifer species in the world? Patterns of endemism and species disjunction are also presented.

Chapter four present the history of post-glacial paleoclimate change on the Olympic Peninsula. Various proxies of past climate are synthesized, including the record of glaciation from the Last Glacial Maximum (21,000 years ago) to the present.

Chapter five presents the postglacial vegetation history of the biota as determined from a set of pollen records, including the five most recently developed records over the peninsula. Various techniques are used to demonstrate changing vegetation patterns and occurrence of fire. A regional synthesis of mapped pollen data from western Washington is also presented.

Chapter six presents a brief review of important archeological sites and places their occurrence in time and space in the context of the environmental changes occurring on the peninsula.

Chapter seven presents several areas of knowledge gaps and additional needed research.

The report does not explicitly address impacts of ongoing and projected climate changes on the peninsula. There are several recent assessments related to this topic (Jenkins et al. 2002, Halofsky et al. 2011, Devine et al. 2012). However, the data presented here are relevant towards understanding the mechanisms governing ecological change and historical precedents of the modern changing environment on the Olympic Peninsula.

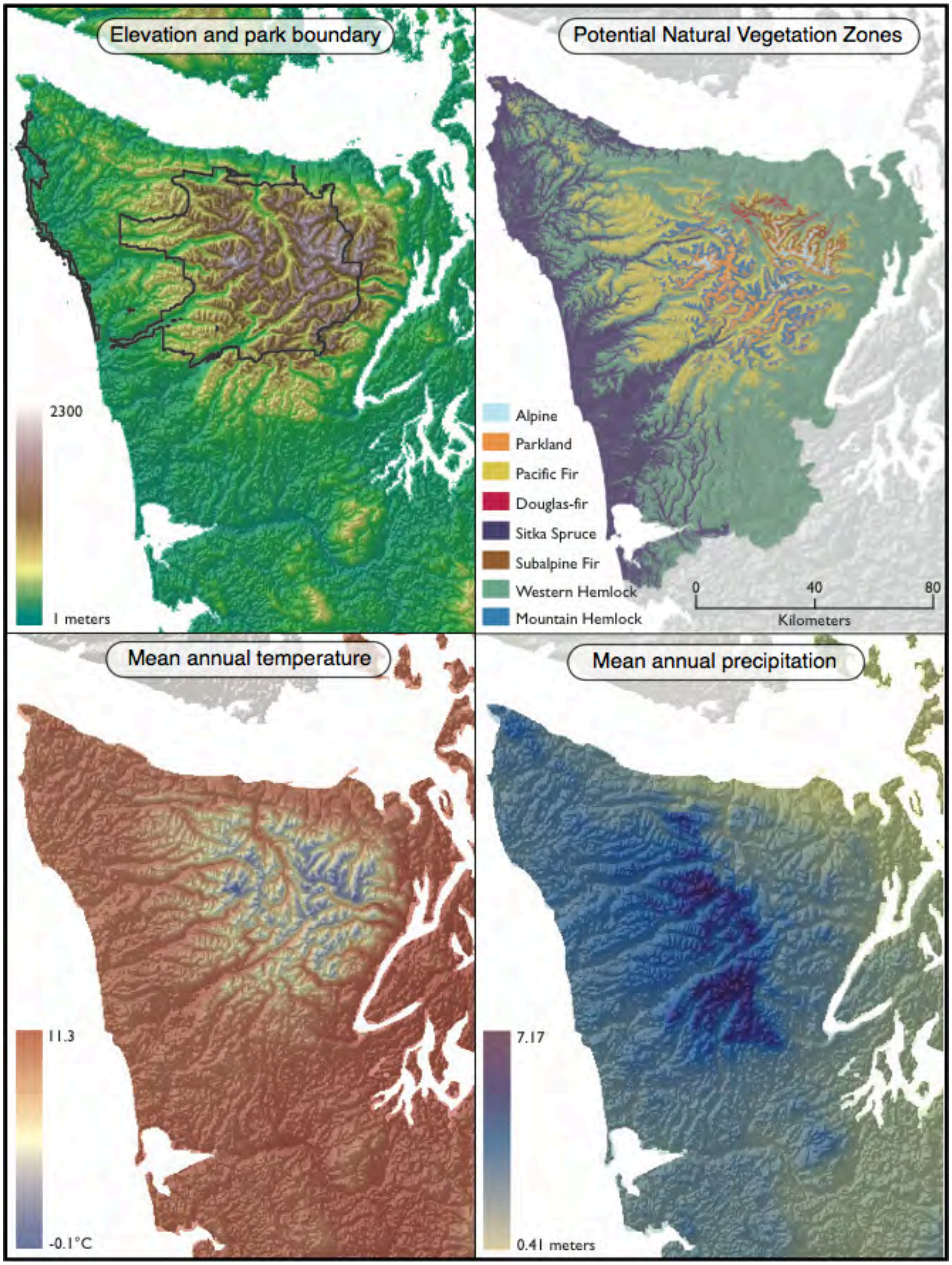

Figure 2. Elevation, vegetation and climate on the Olympic Peninsula, Washington. Potential Natural Vegetation Zones are from Henderson et al. (2011) and the mean temperature and precipitation is from the PRISM model (Daly et al. 2002). Olympic National Park is shown in the elevation map.

2. The modern landscape of the Olympic Peninsula

> "Olympic National Park is of remarkable beauty, and is the largest protected area in the temperate region of the world that includes in one complex ecosystems from ocean edge through temperate rainforest, alpine meadows and glaciated mountain peaks. It contains one of the world's largest stands of virgin temperate rainforest, and includes many of the largest coniferous tree species on earth"
>
> – UNESCO World Heritage Centre, Statement of Significance for Olympic NP.

2.1 Geography of the Olympic Peninsula and Olympic National Park

The Olympic Mountains encompass a roughly 7000 km^2 area, occupying about 50% of the Olympic Peninsula in northwest Washington. A large coastal plain occupies the western third of the peninsula, while the mountain massif occupies the interior and extends nearly to the north and east coasts. The mountains, reaching an elevation of nearly 2400 m (8000 feet), represent a distinct range in the long string of coastal mountains from British Columbia to California. The mountains were formed by a scraping off, or accretion, of ocean-floor sediments comprised mainly of sandstone and shale onto the continental margin. A distinct feature of the Olympics is a band of basalt rocks in a horseshoe pattern around the north, northeast, and eastern portion of the mountains, called the Crescent Formation. Olympic National Park occupies 3696 km^2 on the peninsula, 95% of which is located on the Olympic Mountains and 5% located along a coastal strip on the western Olympic Peninsula. Of the land area under 350 m elevation (59% of the peninsula), only 6.5% is protected within the park. In contrast, 57% of the area above 350 m elevation is within the park. The park and surrounding wilderness areas contain one of the largest contiguous intact temperate forests in the world.

The modern topography of the Olympic Mountains is largely a product of Pleistocene glaciations, which created broad U-shaped valleys extending westward from the interior. These valleys are broadest on the southwest side of the peninsula, whereas valleys on the east and south side are predominantly V-shaped, resulting in a variety of landforms over a short distance (Montgomery 2002). Glaciers today are restricted to the highest elevation cirque basins and to the Olympic massif and are retreating rapidly.

2.2 Regional climate on the Olympic Peninsula

The Pacific Northwest climate is characterized by dry summers and wet winters with moderate temperatures year-round. This pattern is the result of the mid-latitude location of the Pacific Northwest receiving westerly wind from the Pacific Ocean and the presence of large mountain ranges. A seasonal pattern of large semi-permanent pressure cells over the Pacific Ocean is the primary control of the seasonal patterns of precipitation. During summer, a strong Pacific subtropical high-pressure system occurs over the eastern Pacific. Clockwise flow from this system produces a steady westerly and northwesterly flow with descending air resulting in low humidity and clear skies. During the winter, the subtropical high moves southward as the Aleutian low-pressure system intensifies and moves south from the northern Gulf of Alaska. This large low-pressure system picks up moisture over the relatively warm Pacific Ocean and delivers it to the Pacific Northwest. The counter-clockwise flow around the Aleutian Low brings air arriving from the south and southwest during the winter, which results in a maritime air

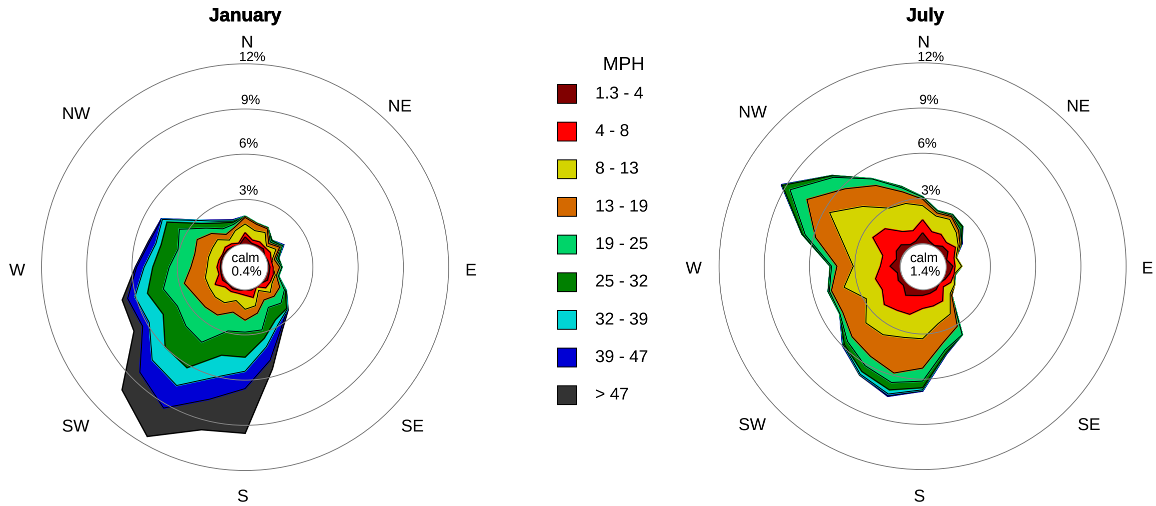

Figure 3. Wind rose diagrams for 850 mb winds at Quillayute State Airport near Forks, WA for the period 1990-2012. The diagrams show the proportion of measurements (twice per day by weather balloon) at a given wind direction and speed. The 850 mb elevation is high enough (generally 5000 feet) such that topography does not influence wind direction and speed. Note the predominately strong SSW winds in January and the gentle NW to SW winds in July. Data were accessed from http://www.esrl.noaa.gov/raobs/

mass and moderated winter temperatures (Figure 3). This circulation pattern can, at times, entrain moisture from the subtropics in an “atmospheric river” which can cause intense precipitation sustained for many days.

The Olympic Mountains produce a pronounced rainfall gradient. Moisture-laden air, as it moves inland, is lifted over the mountains and cooled to below the dew point temperature, producing clouds and precipitation. This orographic precipitation occurs on the southwest-facing facets of the major mountain ranges. Precipitation is also greatly affected by individual ridge systems: a recent study has shown that an 800 m high ridge in the western Olympic Mountains receives 50% more rainfall than the valley bottoms (Anders et al. 2007, Minder et al. 2008). Indeed, the highest precipitation in the coterminous United States, estimated by precipitation models to be over 5 m, occurs at Mount Olympus (Figure 2). As soon as air reaches the leeward side of the mountains, it descends and warms, rainfall ceases, and clouds dissipate. Because of the consistent southwesterly direction of airflow bringing moisture to the peninsula, the southwest portion of the Olympic Mountains receives the vast bulk of precipitation (Figure 2). The rainshadow climate in Sequim, which has an average annual rainfall of only 418 mm (1916-1980), is located only 55 km from the >5000 mm precipitation at Mount Olympus.

The mountains also modify temperatures on a daily basis. During the summer, the difference in temperature between the land and the ocean is the greatest, causing a sea breeze from the cool ocean to the warm land, and an upslope breeze into the mountains. At night, the breeze reverses in the mountains, bringing a cooler downslope breeze. In the fall, this land breeze can result in valley fog that may take many hours to dissipate each day.

Climate diagrams illustrate the seasonal pattern of precipitation and temperature at five sites across the peninsula (Figure 4). These diagrams show the distinct seasonal pattern of precipitation and the distinct gradient created by orographic precipitation and

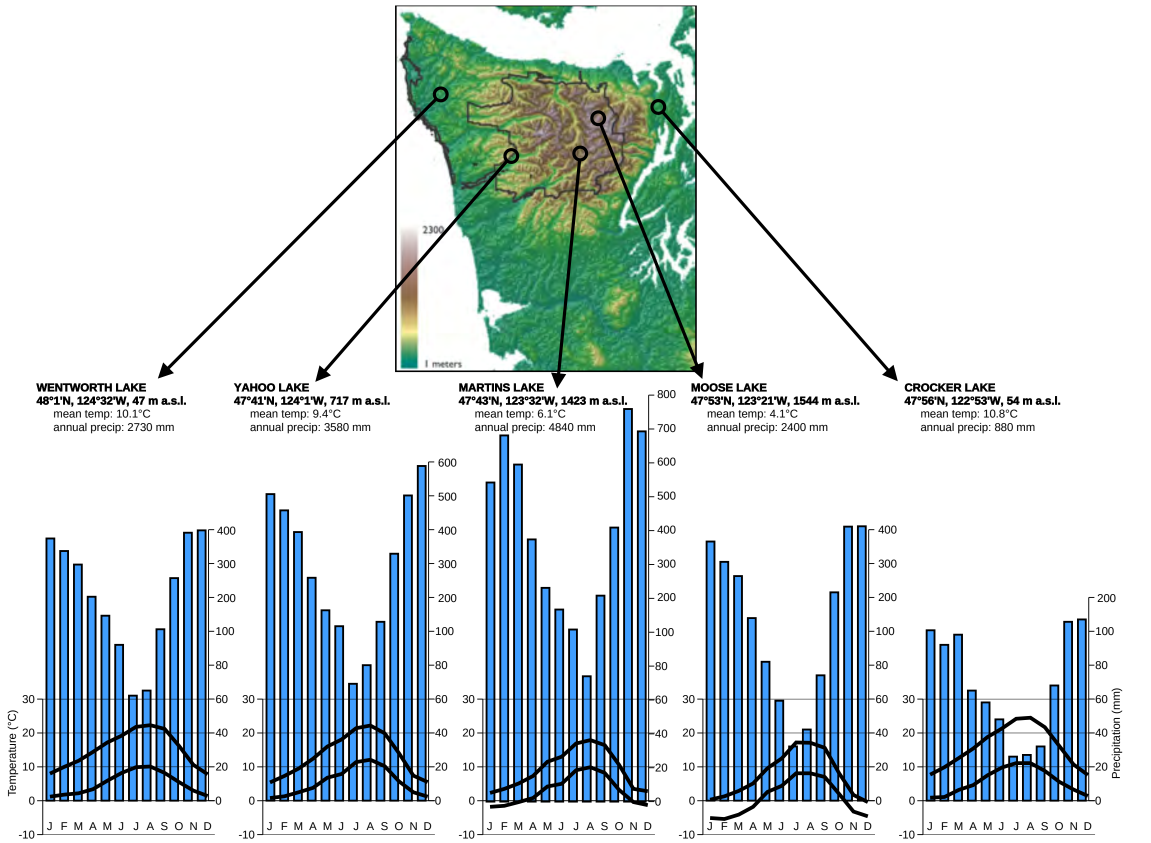

Figure 4. Climate diagrams from five locations straddling the climatic gradient on the Olympic Peninsula. Climate data are 1970-1999 mean monthly precipitation (blue bars) and mean monthly maximum and minimum temperature (black lines) from PRISM (Daly et al. 2002) and interpolated to local sites using the CLIMATEWNA software (Wang et al. 2012).

the rain shadow. Summer maximum temperatures are cooler on the west side (18 °C) than the east side (24 °C) due to the moderating sea breeze on the west side. In contrast, summer minimum temperatures are remarkably uniform over the peninsula (ca. 10 °C), likely due to cold air drainages originating in the mountains and reducing air temperature near the ground at night at the low elevations. Mean winter temperature reaches below freezing only in the upper-elevations on the east side, though temperatures straddle the freezing mark at the highest elevations on the west side, where snow accumulation is very high. The diurnal range of temperature is smaller at high elevation (8°C in August) than at low elevation (14 °C in August on the east-side) because high elevations receive wind that has not been greatly affected by ground level warming and cooling.

2.3 Indicators of recent climate change on the Olympic Peninsula

Over the last 100 years, the greatest change in seasonal climate has been an increase in monthly minimum temperatures in the western Olympics. At Forks, both the January and July minimum temperature has increased at a rate of 0.1 °C/decade since 1895, but since 1960 this rate has increased to 0.34°C/decade for January or 0.32°C/decade for July. At Port Townsend, in the rain shadow, there has been a more muted increase in minimum temperatures since 1960 (0.19 and 0.16 °C/decade for January and July, respectively). At both locations, monthly maximum temperatures do

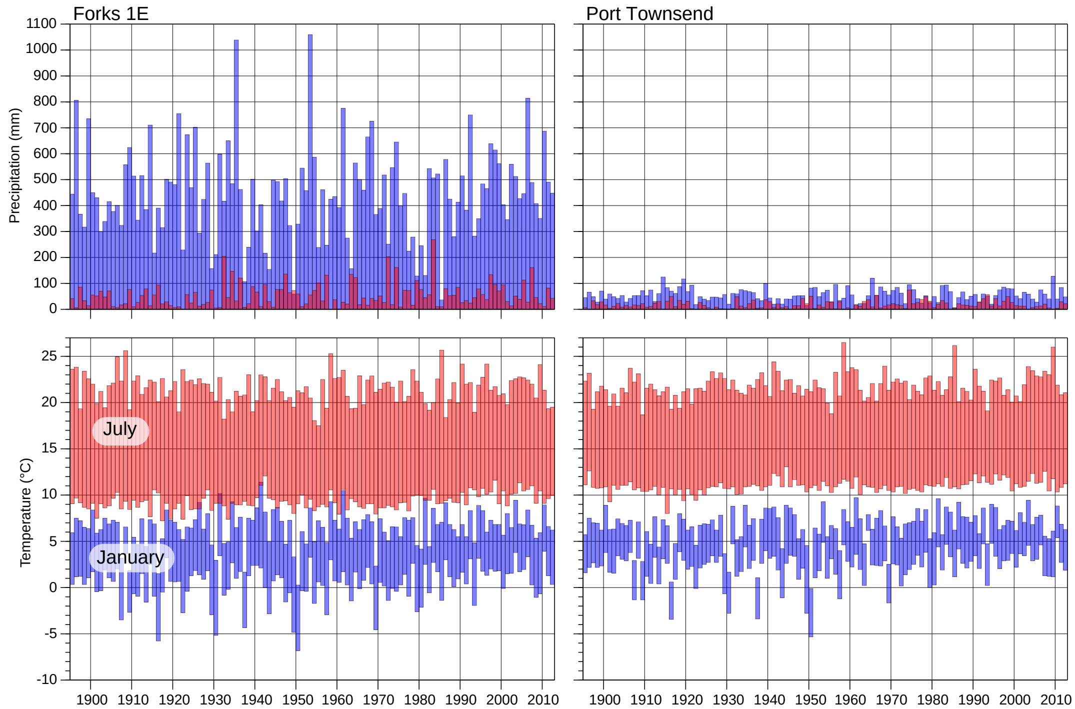

Figure 5. January and July precipitation (top) and minimum and maximum temperatures (bottom) at Forks and Port Townsend WA. Data are from the Historical Climatology Network version 2 (Menne et al. 2009).

not trend more than 0.1 °C/decade. Winter precipitation at Forks increased at a rate of 14.3 mm/decade since 1960, though no change was detected for summer precipitation at Forks and no change in seasonal precipitation was detected at Port Townsend.

The trends in mean climate mask some of the more dramatic changes occurring on the Olympic Peninsula with respect to snowpack. A small increase in winter temperature can have a large effect on snowpack because of the fairly warm temperatures in the subalpine and alpine zones. (Nolin and Daly 2006) estimate that 61% of the snowpack area of the Olympic Mountains has a snow-pack that is considered “at risk” due to the large quantity of snow that occurs at temperatures within 2 °C of the rain-snow boundary. Spring snowpack at Hurricane Ridge has declined steadily since the beginning of surveys in 1949 until the late 1990s and has remained low despite one high snow year in 1998 (Figure 6). The cause of the decline in snowpack can be attributed to a 2 °C increase in winter minimum temperature on wet days between the periods 1948-1976 and 1976 to 1999 (Conway et al. 1999). The trend in snowpack generally follows the phases of the Pacific Decadal Oscillation (PDO), a decadal-scale pattern of sea-surface temperature that is strongly correlated to spring temperature and snowfall (Mantua et al. 1997). Interannual variability in snowpack also correlates strongly with the El Niño Southern Oscillation (ENSO); when both ENSO and PDO are in a cold phase spring snowpack is greater than normal. Regardless of the vagaries of the modes of ENSO and PDO, the data clearly show a link between increasing winter temperature and declining spring snow pack.

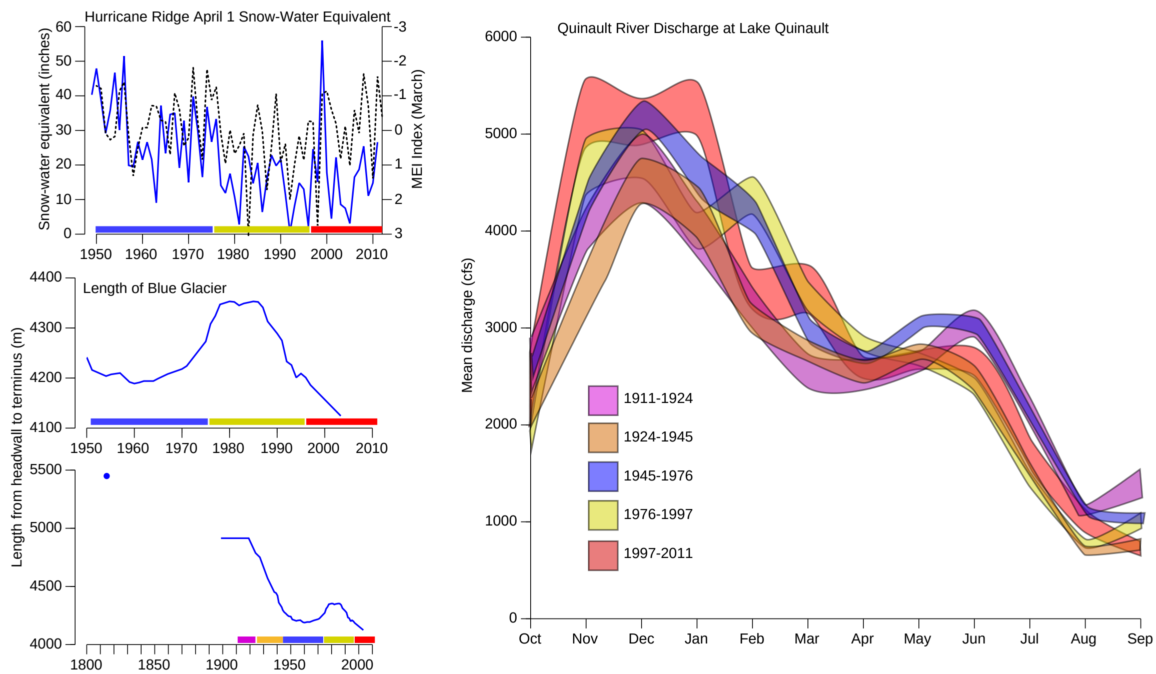

Figure 6. Top-left: Snow-water equivalent on April 1 at Hurricane Ridge snow course from 1949 to 2011 (data from the Natural Resources Conservation Service). Dashed line is the Multivariate ENSO Index for March, in which negative values indicate La Niña conditions. Middle-left: The length of Blue Glacier from 1950-2003. Lower-left: Length of Blue Glacier from 1815 to 2003 (data from Heusser (1957), Spicer (1989 and Conway et al. (1999), and online from the University of Washington). Phases of the Pacific Decadal Oscillation (PDO) are marked by colored lines (Deser et al. 2004). Right: Monthly discharge from the Quinault River at Lake Quinault. Data have been lumped into time segments corresponding to phases of the PDO and plotted as color bands that span the highest and lowest mean monthly discharge within each time period. A distinct 'warm phase' of the PDO, from 1976-1997, was marked by higher early-spring discharge and a smaller snowmelt pulse in May and June.

The history of Blue Glacier on Mount Olympus reveals a longer-term trend of declining snow (Figure 6). The glacier has retreated almost 1.5 km from its early-19th century maximum, despite a period of re-advance in the 1970s. The negative mass balance of recent decades can be attributed to the increase in temperature on days with snowfall (Conway et al. 1999). Earlier snowmelt and less snow accumulation not only cause glacial retreat, but also changing river hydrology with a decline in the snowmelt pulse in May and June (Figure 6). Further projected warming is expected to accentuate this pattern, reducing spring and early summer streamflow (Halofsky et al. 2011).

2.4 Vegetation zones of the Olympic Peninsula and climate controls on forest vegetation

Forest and subalpine meadow composition on the peninsula has been described in several monographs (Fonda and Bliss 1969, Kuramoto and Bliss 1970, Franklin and Dyrness 1988, Henderson et al. 1989, Schreiner 1994, Buckingham et al. 1995). Recently, (Henderson et al. 2011) developed a mapping algorithm to describe the extent of eight 'vegetation zones' as shown in Figure 2. These zones broadly follow zones described by previous authors, and generally correspond to the dominant late-successional species. Vegetation zones are generally defined by the dominant vegetation type on a 'zonal' site with a typical soil type and drainage, and thus do not capture fine-scale patterns created by steep slopes or wetlands, but they do reflect the effect of slope aspect on the temperature regime (Figure 7). Henderson et al. (2011) formalized the definition of the vegetation zones on the Olympic Peninsula using a classification scheme based on a hierarchy of indicator tree species. In this system, forested vegetation zones are assigned to the first indicator species of the list to reach more than 10% cover. This hierarchical list of species (and corresponding vegetation zones) is Sitka spruce, mountain hemlock, Pacific silver fir, western hemlock, subalpine fir, and Douglas-fir. Thus, a plot is in the Douglas-fir Zone only if none of the five preceding species in the list reach over 10% cover. The western hemlock zone is based on the sum of western hemlock and western redcedar cover.

This system was used to develop maps of potential Natural Vegetation Zones on the peninsula (Henderson et al. 2011). The classification scheme was applied to >1000 vegetation plots on the peninsula. These plots were then used in statistical models to develop predictive equations in which environmental variables were used to predict boundary elevations among vegetation zones. These models, based on fog, aspect, cold air drainages, temperature, and precipitation, resulted in an accuracy of ca. 70% for predicting the correct vegetation zone. Descriptions of the vegetation zones below are based on a the regional vegetation classifications of Franklin and Dyrness (1988) and Henderson et al. (1989).

The tree species on the Olympic Peninsula have very large geographic ranges, all of which extend into warmer and colder climates than occur on the Olympic Peninsula. Therefore, the controls on species distributions within the Olympic Peninsula, including the orderly layering of the vegetation zones, must reflect other aspects of the seasonal climate aside from temperature alone, as well as biotic interactions. For each vegetation zone, mean temperature and precipitation values are provided. In addition, a simple water balance model is described for each vegetation zone, including an index of productivity (actual evapotranspiration) and drought stress (deficit). Studies of climatic effects on particular life-history stages of the dominant tree species are also mentioned.

2.4.1 Sitka Spruce Zone

The Sitka Spruce Zone occupies the low-elevation coastal forests that rarely extend above 250 m elevation and that experience a hypermaritime climate. Fires are very infrequent and prior to the logging era forests were dominated by late-successional species but subjected to windthrow disturbances. Dominant tree species are Sitka spruce, western hemlock, and western redcedar. Common shrubs include salal, salmonberry, red huckleberry, and vine maple. Herbaceous species include oxalis, swordfern, ladyfern, deerfern, and foamflower. Understory regeneration of hemlock, spruce, and redcedar is normally restricted to 'nurse logs' that can supply a source of moisture, which is unavailable to the short roots of tree seedlings when in competition with dense moss mats on the forest floor (Harmon and Franklin 1989). River terraces along the Hoh River have been interpreted as a long-term chronosequence (Fonda 1974). Early successional communities on river floodplains are rapidly colonized by red alder and willows (Van Pelt et al. 2008), but Sitka spruce also establish amidst the hardwoods and eventually replace the hardwoods when they die out after 80 years (Stolnack and Naiman 2010). Red

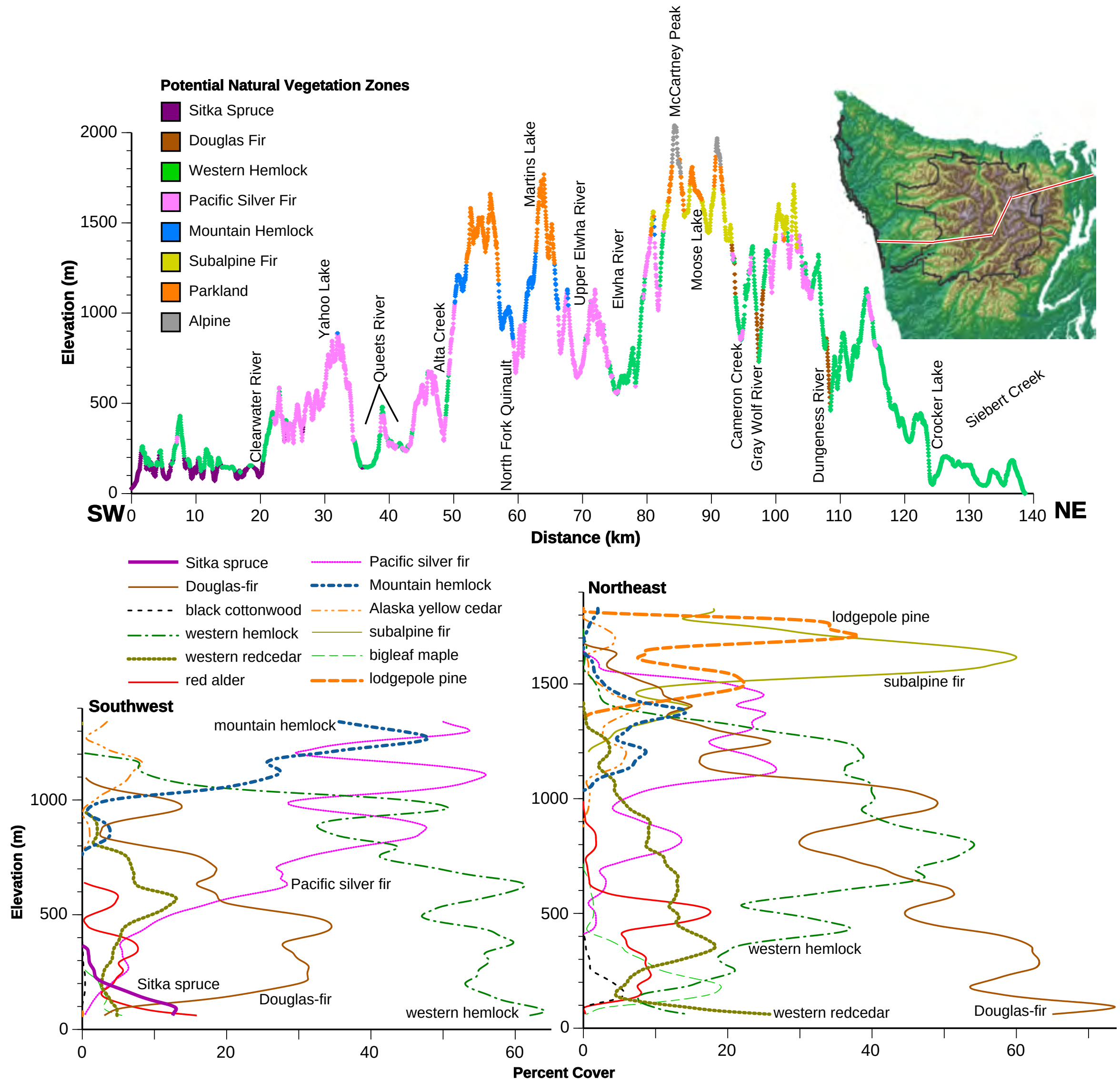

Figure 7. Top: An elevation profile across the Olympic Peninsula showing the pattern of potential natural vegetation zones (Henderson et al. 2011). Note how the elevation of zonal boundaries varies with the effects of aspect, cold-air drainages in narrow valleys, and location along transect. Bottom: Percent cover of tree species in 639 ecological plots in the Olympic National Forest (downloaded from Ecoshare.info). Plots were divided into southwest and northeast areas. Curves are a locally-weighted running median in a 300-m window.

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alder is particularly important for building soil nitrogen pools (Luken and Fonda 1983). Elsewhere, disturbances (such as from fire) may be followed by a dense shrub phase, including salmonberry and red elderberry. As Sitka spruce is less shade tolerant than western hemlock, it requires moderate-sized canopy gaps for it to reach the canopy (Taylor 1990). The oldest terraces are dominated by western hemlock and western redcedar.

There are three distinct variants within the Sitka spruce zone. The structurally complex “Olympic rainforests” of the major river valleys (Hoh, Quinault, and Queets) contain massive Sitka spruce and western hemlock with understory of epiphyte-laden bigleaf maple and open understory glades maintained by browsing Roosevelt Elk (Schreiner et al. 1996). An ungulate-exclosure study revealed a strong effect of ungulate herbivory on herb, shrub, and tree density in the understory, including selective browsing that may increase Sitka spruce recruitment relative to other conifers (Woodward et al. 1994). The response of tree-seedling abundance to the changing elk populations over time, however, requires further investigation, as an early study on western hemlock ages failed to show a relationship (Harmon and Franklin 1983). On the coastal plain, a nutrient-poor muskeg swamp forest is dominated by western redcedar and shore pine with a very dense salal understory. Last, treeless prairies dominated by bracken fern occur in several places. Maintenance of these open prairies was likely the result of Native American burning practices (Reagan 1909, Anderson 2009).

The Sitka Spruce Zone has an average temperature of 4.3°C in January, 15.6°C in July, and 2700 mm of precipitation. This area is marked by a strong fog effect where up to 1 m of extra precipitation may occur through fog drip. Sitka spruce has low drought tolerance, generally requiring over a meter of annual precipitation, and is also very tolerant of flooding disturbances (Figure 9; Peterson et al. 1997b, Niinemets and Valladares 2006). Sitka spruce is favored in hollows and toe slopes where topographic effects yield high soil moisture. At its upper elevational limits, Sitka spruce is limited to riparian areas (Figure 7). During summer a strong diurnal sea-breeze and land-breeze pattern moderates the daily high and low temperatures (Mass 1982). Mean winter temperature and actual evapotranspiration are greater than uphill in the Western Hemlock Zone, and there is very little moisture deficit through the year (Figure 8). Tree-ring studies have found tree growth in this zone to be poorly correlated with any climate variable, though there is a possibility there is a correlation over multi-year frequencies (Nakawatase and Peterson 2006). In contrast, Holman and Peterson (2006) found high inter-annual sensitivity in Sitka spruce growth rates, and a potential for very fast growth, and suggested that this forest type is the most sensitive to climate changes on the Olympic Peninsula. Considering the competitive environment of Sitka spruce forests, and the importance of canopy disturbances in its growth rate, it is likely that climate-change effects in this forest zone may occur indirectly via the disturbance regime.

2.4.2 Western Hemlock Zone

The Western Hemlock Zone, comprising roughly half of the peninsula, occurs in a narrow elevational band above the Sitka Spruce Zone in the west, but occupies a broad band from sea level to nearly 1200 m in the drier northeast portion. The Western Hemlock Zone corresponds to the areas where other indicator species (Pacific silver fir and Sitka spruce) are rare or absent; though western hemlock may be common outside of this zone (Figure 7). In the eastern and northern areas, forest fires have been common over the past several centuries, leading to dominance by even-aged seral Douglas-fir stands over broad areas. In wetter areas of the western peninsula the zone contains much more western hemlock and western redcedar, though pure climax forests are rare. Western white pine may occur sporadically, though this species has declined dramatically in recent decades (Harvey et al. 2008). Common shrubs include salal, vine maple, Oregon grape, red huckleberry, and salmonberry. Herbaceous species include swordfern, deerfern, oxalis, beargrass, twinflower, prince’s pine, evergreen violet, vanillaleaf, trillium, and foamflower. The early successional patterns within this zone are highly variable due to site differences and seed sources. On dry sites, Douglas-fir can invade

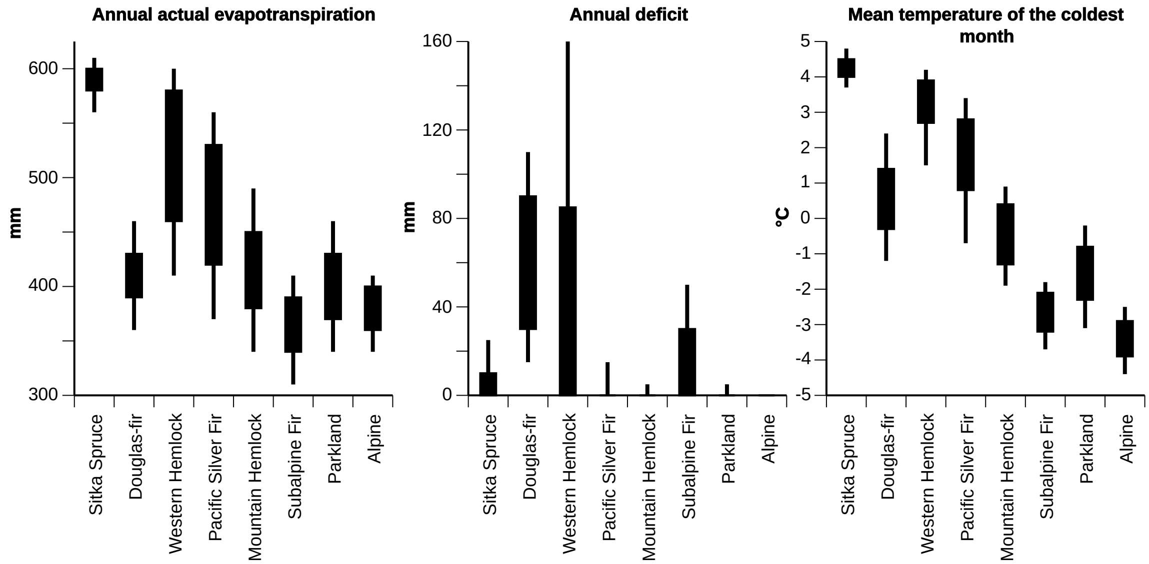

Figure 8. Distribution of three bioclimatic variables within eight potential natural vegetation zones on the Olympic Peninsula. Box plots show the 10^{th}-90^{th} percentiles (thin lines) and the 25^{th}-75^{th} percentiles (boxes) of each variable within each vegetation zone. This figure was developed by overlaying the Potential Vegetation Zones of (Henderson et al. 2011) with a simple water balance model (Willmott et al. 1985, Gavin and Hu 2006) applied to mean monthly temperature and precipitation from the PRISM spatial climate data calculated for the 1970-1999 period at a grid resolution of 400 m (Daly et al. 2002).

quickly and produce even-aged stands within 10 years of the disturbance (Winter et al. 2002), while in wet sites fireweed and shrub phases and/or a red alder phase may persist for decades. There is a broad continuum of variants of the Western Hemlock Zone related to site moisture and productivity, each with well-defined understory indicator species (see Henderson et al. 1989).

The Western Hemlock Zone has an average temperature of 3.7 °C in January, 6.2 °C in July, and 2200 mm of precipitation. On the western peninsula, western hemlock is common and is favored by its very high shade tolerance in forests with little fire disturbance (Figure 9) and its abundance in climates with high evapotranspiration (i.e., warm and wet summers; Gavin and Hu 2006). In the dry northeastern area this zone is dominated by Douglas-fir, which spans a wide range of elevations reflecting recent widespread fires. Douglas-fir growth is limited by moisture deficits except in the wettest portions of the peninsula (Littell et al. 2008). Western redcedar is commonly associated with western hemlock, and is favored on the warmest sites with some growing season precipitation or some topographic moisture (Harrington and Gould 2010). Moisture deficits span a wide range of values, though generally the soil water balance model here did not show any deficit in the wetter areas (Figure 8). Tree-ring studies have shown poor climatic control on the elevational boundaries of the dominant species in this zone, suggesting that competitive interactions may be a major determinant of species elevational boundaries (Ettinger et al. 2011).

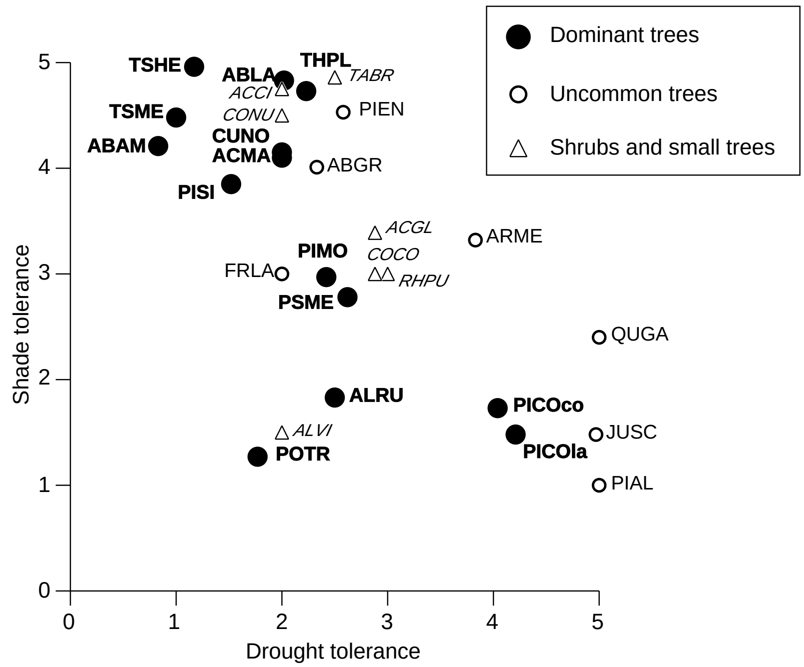

Figure 9. Relative rankings of shade tolerance and drought tolerance of trees and selected shrubs on the Olympic Peninsula as reported in Niinemets and Valladares (2006). These characteristics explain a large amount of the geographic variation in species abundances. Low shade tolerance is associated with disturbed habitats (e.g. after fire, blowdowns, or along riparian corridors) and low drought tolerance is associated with the wettest climates on the peninsula, and low drought and shade tolerance occur in species restricted to wet disturbed sites (avalanche paths and along rivers). Tolerance to cold and length of growing season, not shown here, also varies considerably among species and explains much elevational variation. Species codes listed alphabetically: ABAM: Pacific silver fir, ABLA: subalpine fir, ABGR: grand fir, ACCI: vine maple ACGL: Rocky Mountain maple, ACMA: bigleaf maple, ALRU: red alder, ALVI: slide alder, ARME: madrone, COCO: beaked hazel, CONU: Pacific dogwood, CUNO: Alaska yellow cedar, FRLA: Oregon white ash, JUSC: Rocky Mountain juniper, PIMO: western white pine, PISI: Sitka spruce, PICOco: shore pine, PICOla: lodgepole pine, PIAL: whitebark pine, PIEN: Engelmann spruce, POTR: black cottonwood, PSME: Douglas-fir, QUGA: garry oak, RHPU: cascara, TABR: Pacific yew, THPL: western redcedar, TSHE: western hemlock, TSME: mountain hemlock.

2.4.3 Pacific Silver Fir Zone

The Pacific Silver Fir Zone occurs at mid elevations, above the Western Hemlock Zone and below the Mountain Hemlock Zone. It may reach down to 300 m elevation on north-facing slopes on the western peninsula, but is restricted to areas >1000 m elevation in the drier eastern peninsula. The upper elevation limit of the zone is also much higher in the east (1400 m) than the west (1000 m). The dominant tree species are Pacific silver fir, western redcedar, and western hemlock, though Douglas-fir may be present up to 1000 m elevation as relicts from fires occurring during warmer periods 500 to 800 years ago. Common shrubs include several huckleberry species, white rhododendron, mountain ash, false huckleberry, salal and Oregon grape. Herbaceous species include bunchberry, queen’s cup, twisted stalk, vanillaleaf, false lily-of-the-valley, deerfern, swordfern, five-leaved bramble, foamflower, and trillium. Succession after disturbance may begin with a mixture of western hemlock and Douglas-fir, which eventually are replaced by the very shade-tolerant and snow-load tolerant Pacific silver fir.

There are few distinct variants within the Pacific Silver Fir Zone. Sitka alder communities form where snow avalanches recur frequently, or where creeping snow and high water tables inhibit seedling establishment. Alaska yellow cedar may be the only conifer that survives the conditions in these sites. Over broader areas, wet treeless meadows may be dominated by a thick cover of bracken fern and thimbleberry. These meadows appear to be compositionally stable and not slowly converting to forest.

The Pacific Silver Fir Zone occupies the lower-montane elevations and is more extensive in wet areas. This zone has an average temperature of 2.0 °C in January, 15.1 °C in July, and 3500 mm of precipitation. Pacific silver fir is one of the least drought tolerant trees on the peninsula (Figure 9). Henderson et al. (2011) associated a lower elevational limit of Pacific silver fir with areas of high precipitation, cold air drainages, north-facing aspects, and high topographic moisture. While actual evapotranspiration in the Pacific Silver Fir Zone may be broadly similar with the Western Hemlock Zone, moisture deficits are much lower and winter temperatures are notably lower in the Pacific Silver Fir Zone than at lower elevations (Figure 8) such that snowpack may accumulate for a portion of the year. Pacific silver fir has greater mechanical strength and flexibility under snow loads compared to western hemlock.

2.4.4 Mountain Hemlock Zone

The Mountain Hemlock Zone is the highest forested zone in the wettest portion of the peninsula, where it occurs from 1200 to 1500 m elevation. It is also present in patches on north-facing slopes in the northeast. In heavy-snow areas, subalpine parklands occur within the upper portions of the Mountain Hemlock Zone, while in low-snow areas moisture limits the occurrence of mountain hemlock. The zone is dominated by mountain hemlock and Pacific silver fir, with minor amounts of Alaska yellow cedar. Common shrubs include several huckleberry species, false huckleberry, mountain ash, white rhododendron, and red heather. Herbaceous species include avalanche lily, five-leaved bramble, deerfern, queen’s cup, beargrass, and one-sided pyrola. (Agee and Smith 1984) found that after fires tree establishment generally mirrors that of the pre-existing forest, though climate strongly modulated recruitment patterns (see below). Fire may promote huckleberry species in meadows. There are few variants of the Mountain Hemlock Zone. Non-forested vegetation is discussed below.

The Mountain Hemlock Zone is the highest forest zone in wet areas and is absent in dry areas. This zone has an average temperature of -0.2 °C in January, 13.1 °C in July, and 3500 mm of precipitation. Its upper elevational limit is controlled by topographic moisture and is extended upwards under lower precipitation and on south-facing slopes. These patterns are readily explained by an observed relationship between mountain hemlock and spring snowpack (Means 1990). Heavy snow pack, however, limits the growth of mountain hemlock by delaying the start of the growing season, while high growth is associated with warm summer temperature (Gedalof and Smith 2001, Nakawatase and Peterson 2006). Climate strongly influences the upper-elevational limit of trees in this zone as revealed by a strong growth-climate relationship at treeline, while no such relationship existed at the lower elevational limit suggesting climatic control of other life-history stages (e.g., germination and establishment) and/or non-climatic factors

(Ettinger et al. 2011). Moisture deficits are absent in this zone, and actual evapotranspiration and winter temperatures are notably lower than in the Pacific Silver Fir Zone (Figure 8).

2.4.5 Subalpine Fir Zone

The Subalpine Fir Zone occurs in the dry northeastern peninsula at elevations generally between 1300 m and 1800 m elevation, and is the highest forest zone in the rainshadow climate. Depending on aspect and location, its lower boundary may abut the Western Hemlock, Mountain Hemlock, Douglas-Fir or Pacific Silver Fir Zones. Its upper boundary grades into Subalpine Parkland. The zone is dominated by subalpine fir and/or lodgepole pine, but may contain some Douglas-fir. Shrubs include big huckleberry, common juniper, white rhododendron, and pachistima. Herbaceous plants include subalpine lupine, Sitka valerian, one-sided pyrola, Martindale's lomatium, and white hawkweed. Fire is the major disturbance in this region, and tree establishment requires over five decades and is dependent on climate, substrate, and seed source (Agee and Smith 1984). Seedling survivorship is low in most meadows in the parkland zone, though heath-shrub communities tend to have safe sites for seedling survival (Soll 1994).

This zone has an average temperature of -2.6 °C in January, 12.2 °C in July, and 2300 mm of precipitation. The zone is most prevalent on dry south-facing slopes. It has a distinct moisture deficit approaching 50 mm (based on the simple water balance model used here) and after fire drought-adapted lodgepole pine may be abundant. The zone experiences distinctly lower winter temperatures than in the Mountain Hemlock Zone (Figure 8). Subalpine fir, like mountain hemlock, has high growth rates on years with low snowpack and high temperatures, and this relationship is especially strong at high elevation high-snowpack sites (Ettl and Peterson 1995, Peterson 1998). Growth is negatively correlated with temperature of the previous year, possibly via the impact of cone production and respiration on carbohydrate reserves. Furthermore, growth is correlated with summer precipitation at the middle and lower elevation ranges of subalpine fir, indicating sensitivity to drought varies elevationally (Ettl and Peterson 1995).

2.4.6 Douglas-fir Zone

The Douglas-fir Zone is the most spatially limited forested zone on the peninsula, located near the driest portions of the Western Hemlock Zone on south and west facing slopes. It is observed in the Dungeness, Maiden Creek, and Elwha drainages. Douglas-fir is the dominant tree species, and grand fir, lodgepole pine, Rocky Mountain juniper, and madrone may be present. Conditions are too dry for western hemlock to be the late-successional dominant species (Fonda and Bliss 1969). Shrubs include kinniinnick, Oregon grape, serviceberry, oceanspray, baldhip rose, snowberry, and salal. Herbaceous species present include western fescue, vanillaleaf, white hawksweed, prince's pine, Scouler's harebell, bigleaf sandwort, and starflower. Succcession after disturbance is normally dominated by Douglas-fir during all stages. One distinct variant is the very dry kinnikinnick series, on shallow rocky soils with low fertility and frequent fire.

The Douglas-fir Zone has an average temperature of 0.5 °C in January, 15.0 °C in July, and 1900 mm of precipitation. Much of the area is composed of south-facing aspects and locations with low topographic moisture, making it the most consistently dry zone, with deficits between 20 and 100 mm. The combination of lower rainfall and cooler temperatures also limits the actual evapotranspiration to an average of 410 mm.

2.4.7 Subalpine Parkland Zone

The Subalpine Parkland Zone occurs throughout the high elevations on the peninsula. This zone is comprised of scattered trees, subalpine meadows, and barrens. Its lower boundary is as low as 1100 m on north-facing slopes in wet areas and 1700 m in the dry northeast. The broad range of temperature and moisture in this zone results in several well-defined meadow community types. The first large survey by Kuramoto and Bliss (1970) revealed eight community types. Within meadows, consistent snowmelt gradients and topographic moisture also produce pronounced gradients in community types over short distances (Belsky and Del Moral 1982). Further sampling by Schreiner (1994) expanded the eight communities of Kuramoto and Bliss (1970) to a system of 16 community types that cluster into four groups: scree and talus communities, *Phlox diffusa* communities, *Carex specatabilis* communities, and *Luetkea-Saxifraga* communities. Description of these communities is beyond the scope of this report.

The Subalpine Parkland Zone has an average temperature of -1.5 °C in January, 12.1 °C in July, and 2300 mm of precipitation. The boundary between the forested and parkland zones is likely highly dynamic and influenced by fire and subsequent slow tree regeneration. Invasion of dry meadows by subalpine fir requires heath-shrub microsites and normal or above-average growing-season moisture, such as when the snowpack provides moisture to the soil, which occurred during a high-snowpack period from 1956-1980 (Agee and Smith 1984, Woodward et al. 1995). In contrast, colonization in wet meadows occurs during a dry climate when snowpack does not limit the growing season, as occurred during 1921-1945 (Agee and Smith 1984, Woodward et al. 1995). Repeat photography of the Parkland Zone has also revealed meadow invasion by conifers as well as impacts from mountain goats (Schreiner 1994).

2.5 Natural disturbance regimes of the Olympic Peninsula

2.5.1 Fire

> "But the most magnificent thing I found, and to me it was an amazing discovery, was that every part of the Reserve I saw appeared to have been cleared by fire within the last few centuries. The mineral soil under the humus, wherever it was exposed about the roots of windfalls, was overlayed by a layer of charcoal and ashes. Continuous stretches of miles without a break were covered with a uniform growth of Douglas fir from two to five feet in diameter, entirely unscarred by fire. Among them numerous rotting stumps of much larger trees did bear the marks of burning. I did not see a single young seedling of Douglas fir under the forest cover, not a single opening made by fire which did not contain them. Fires conditioned and controlled the forest of the Olympics."
>
> – Gifford Pinchot, remarking on his 1897 visit to the Bogachiel area. *Breaking New Ground* (Pinchot 1947).

Forest fire, even when recurring only every few-hundred years, has a major effect on the composition and structure of forests. Agee (1993) presented a rigorous synthesis of the fire ecology on the peninsula and Henderson (Henderson and Peter 1981, Henderson et al. 1989) has synthesized evidence of fire history from various sources (Figure 10). The following brief review of the above-cited studies also includes references to more recent studies where available.

On the western Olympic Peninsula, the recent historical record suggests fire is sufficiently rare and of limited extent, though old-growth forests bear evidence of extensive fire in past centuries. Douglas-fir is mostly limited to areas that burned within the last 500 years because fire is required for its establishment in dense forested settings. Douglas-fir also is intolerant of waterlogging that occurs on level terrain in the coastal lowlands. Mapped fire occurrences show few major fires during the past 100 years on the coastal plain. Where Douglas-fir does exist, its age confirms the rarity of fire. Huff (1995) reported 515 year-old Douglas-fir stands in the Hoh drainage. Fahnestock and Agee (1983) reported a fire cycle (a fire-interval estimate based on the age distribution of

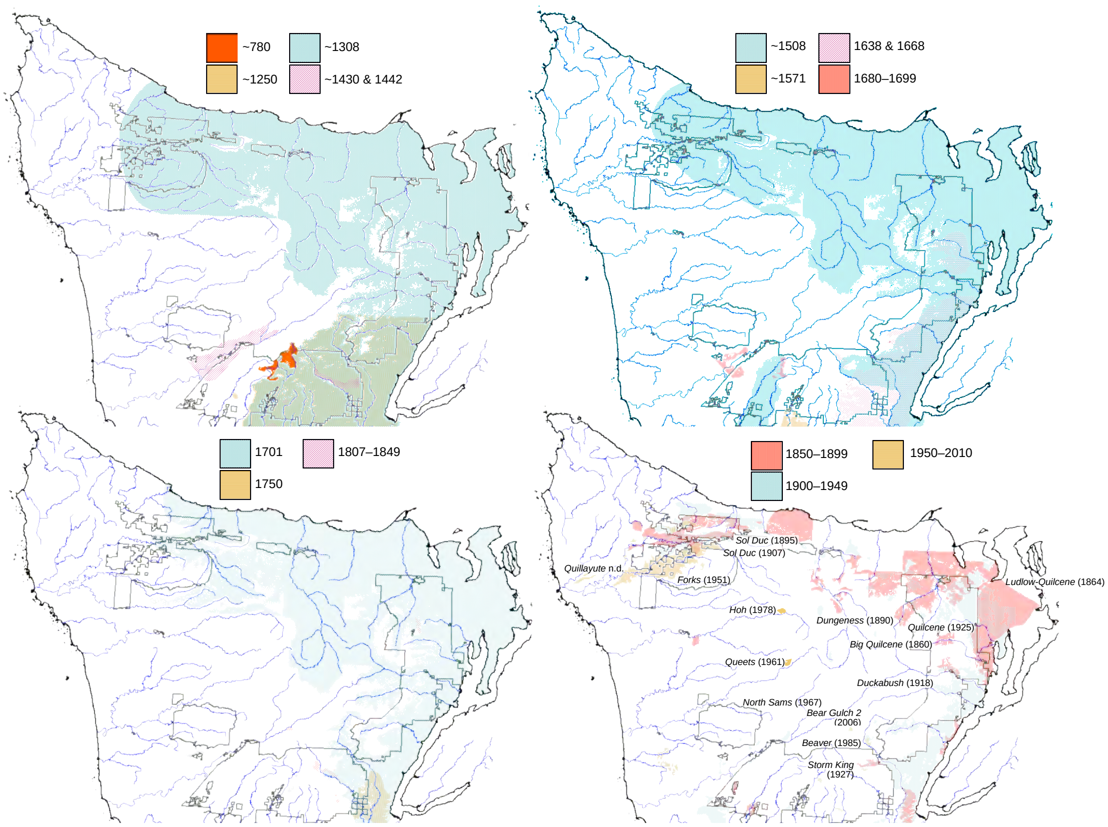

Figure 10. Mapped fire events on the Olympic Peninsula compiled from historical records and Douglas-fir stand ages using data from Henderson and Peter (1981) and Henderson et al. (1989). See these sources for more detailed narratives of more severe fire years. This fire history focused on fires occurring within the Olympic National Forest (gray outline on maps) with an emphasis in the southeastern region. The extents of some fires are crude extrapolations and in some cases are truncated because fires extended beyond the focal area for particular studies (e.g., the 1250 AD fire). Most fires within the Park are not shown (extent of the Queets and Hoh fires are approximations). Of the pre-1860 fires, most fire dates are estimates and fire extents are rough extrapolations, though the 1701 fire is accurately dated in several locations (Jan Henderson, *personal communication*).

trees over a large area) of 1100 years in Sitka spruce forests, though the actual fire cycle may be longer because many tree ages dated to wind disturbance rather than fire (Agee 1993). Agee and Flewelling (1983) used a probabilistic fire occurrence model based on modern climate and lightning occurrence and calculated a fire rotation of over 4000 years for the western Olympic National Park, though this method could only approximate the effects of historical drought. In contrast to the fire cycle statistic, the fire rotation is the amount of time required to burn an area equal to the size of one's study area, which corresponds to the average fire interval at a point if fires were spatially random (Heinselman 1973). For the western hemlock zone, such a long fire rotation is not in agreement with evidence of fires hundreds of years ago (Huff 1995), though the extent of such ancient fires are difficult to reconstruct. An intensive study of fire in a similar forest type on Vancouver Island found that 20% of the lower slopes and terraces of a watershed had not burned in over 6000 years (Gavin et al. 2003a). Thus it is clear that wind disturbance is much more prevalent than fire.

At mid elevations, above river terraces, fire is more common because south-facing aspects are dry during the late summer months. A recent fire on the western Olympic Mountains, the 96 ha Queets Fire in 1961, appears to be typical of fires inferred from stand ages (Figure 12). This fire was of high severity as a result of few fire adaptations of the dominant tree species. However, the fire was restricted to steep south-facing slopes and did not spread into terrace and riparian forests, into upper-elevation forests with late snowmelt date, nor did it extend more than a few kilometers cross-slope. This pattern agrees with a model of fire occurrence described for western Vancouver Island (Schmidt 1960). In contrast, the 425-ha Hoh Fire, ignited by lightning after nearly three months of no rainfall in 1978, overcame some of these topographic barriers and extended into both riparian and subalpine forests on a south-facing slope of the upper Hoh River.

Very unlike the west-side forests, the east-side forest in the Western Hemlock Zone has been shaped by fire over the past few hundred years. Indeed, evidence from the

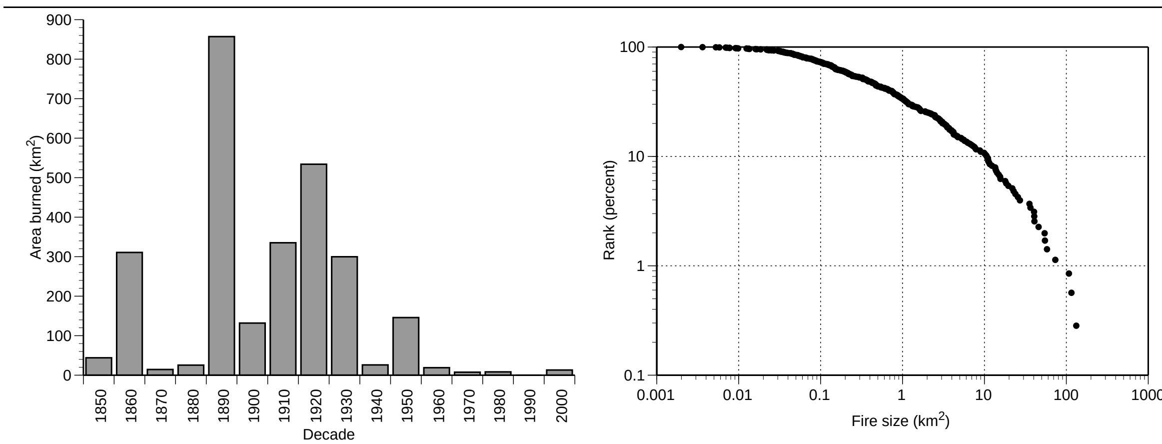

Figure 11. Left: Area burned in decadal time steps from 1850 to present from the mapped fires presented in Figure 10 and from the Monitoring Trends in Burn Severity project (Eidenshink et al. 2007). Right: Inverse cumulative size distribution of recorded fires on the Olympic Peninsula from 1850 to present. Most fires were small: 90% of the 353 reported fires were smaller than 10 km^2, though this percentage is likely to be greater because many small fires during the 19^{th} century were not recorded.

ages of Douglas-fir trees, which establish almost exclusively after fire, suggests massively extensive fires in ca. 1308, ca. 1508 and 1701. Such fires were unlike anything that has occurred in recent history (Figure 10). Extensive fires occurred during settlement times (1860-1910), especially in the northeastern portion of the peninsula and along the Hood Canal. These fires are almost all attributable to land-clearance activities and escaped fires from logging operations. Several fires in the Lake Constance area (Sol Duc fires of 1895 and 1907 and the Forks Fire of 1951) had explosive crown-fire runs as the result of strong east winds (Agee 1993). Such winds occur in the Strait of Juan de Fuca when a thermal trough develops over the Pacific Ocean (Brewer et al. 2012). Most major fires on the peninsula occur during these east wind events (Agee 1993). Fire activity of the last 50 years is a tiny fraction of the area burned during the early 20th century (Figure 11) likely due to increased suppression activity and a reduction in human-caused ignitions.

Figure 12. Aerial photograph of the 1961 Queets Fire scar photographed three years after the fire. Note the high severity of the burn but also several small patches of surviving trees.

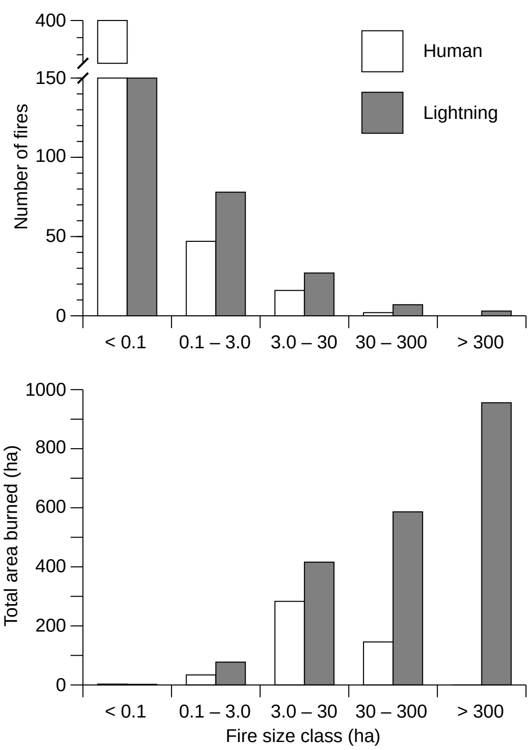

Figure 13. Number of fires and area burned by fire in the Olympic National Park from 1916 to 1975 as reported in (Pickford et al. 1980). Human vs. lightning caused fires are plotted separately.

A compilation of 747 fire events within the Olympic National Park from 1916 to 1975 revealed an extremely episodic pattern of natural fire (Pickford et al. 1980). The proportion of human-caused fires decreased dramatically since 1946 (burning a total of 10 ha). Only 274 fires were lightning-caused, which were strongly biased to only six years. The majority of the area burned by lightning fire (76%) was due to the 10 fires that were greater than 30 ha in size. In the driest portion of the Park, for example, very steep rocky terrain results in discontinuous fuel and a heterogeneous pattern of burn severity, as occurred in the 2009 Heat Wave complex. Thus, the natural pattern of fire is not only limited by fuel moisture and ignition, but also by fuel continuity interrupted by topography.

Tree-ring fire history studies corroborate this episodic pattern of fire over longer time periods. In the Morse

Creek drainage south of Port Angeles, Douglas-fir trees reveal growth releases corresponding to two major fire periods at ca. 1700 and 1895, in addition to many other fires of smaller extent (Wetzel and Fonda 2000). At a scale of 200 ha, fire recurs at an average of 21 years. This relatively short interval suggests that fires must have been mostly moderate or low in severity and were punctuated by the episodic large high-severity fires at 200-year intervals (though criteria for identifying fires from tree-rings were less rigorous than for most other such studies). In the Elwha River south of Lake Mills, another tree-ring fire scar study mapped fire events for the last 440 years (Wendel and Zabowski 2010). The natural fire rotation was 146 years prior to 1775, when large fires burned in 1568, 1661, 1676, and 1729. No fires were detected from 1775 to 1850 due to a combination of Little Ice Age climate and dramatic decline in Native American population. During the period of settler influx the fire rotation dropped to 53 years. Wendel and Zabowski (2010) suggested this period of high fire was partly due to a feedback involving fire-induced loading of fine fuels causing reburns (Agee and Huff 1987), with large fires detected in 1868, 1890 and 1898. Few fires burned after 1926. The conservative criteria for identifying fires used in this study (fires defined by three scars) resulted in longer fire interval estimates than found by Wetzel and Fonda (2000) in a similar forest type.

The effective exclusion of fire over the past 60 years has resulted in increased density of understory vegetation and a simplification of forest structure relative to what has occurred during earlier centuries of more frequent fire (Weisberg 2004). As in other forests in the Pacific Northwest, this condition has prompted the application of prescribed fire. Initial application of a low-intensity prescribed fire in the Maiden Creek watershed was successful in reducing understory biomass while incurring little canopy-tree mortality (Fonda and Binney 2011).

2.5.2 Wind

Severe winter cyclonic storms can produce winds that approach or surpass Hurricane forces (74 mph) leading to extensive blowdown. Tree-ring studies along the Oregon coast have found that wind disturbances have been the primary control on tree growth rates, and the growth-release evidence of high-wind events was most common in the first half of the 20th century (Knapp and Hadley 2012).

The three largest wind events on the Olympic Peninsula occurred in 1921 (The 21 Blow), 1962 (The Columbus Day Storm) and 1997. The 1921 storm was reported to have reached wind speeds over 100 mph along the coast resulting in 40% of forest blown down on the southwestern flanks of the Olympic Mountains (Mass and Dotson 2010). Windthrow from the 21 Blow was reported to have been strongly biased to western hemlock and Pacific silver fir, as those species have weaker boles compared to western redcedar and Douglas-fir. The blowdown also resulted in a release of the advance regeneration in the forests, which was predominately the shade-tolerant western hemlock (Dale et al. 1986). The Columbus Day Storm originated as tropical cyclone in the western Pacific that was entrained into the westerlies and moved quickly to the Pacific Northwest (Lynott and Cramer 1966). The majority of damage was focused in the southern peninsula. Another storm in 1997 had winds focused along the coast with most blowdown occurring south and north of Grays Harbor. 11,600 hectares of blowdown was mapped within a 30-mile wide coastal strip, much of it occurring on tribal land (Forest Health Program 2008).

2.5.3 Other disturbance agents

Forest insects comprise a minor role in the disturbance regime as no widespread outbreaks have been reported and disease surveys are not regularly conducted within the Park (Henderson et al. 1989). Few species have been ranked as a severe threat (Devine et al. 2012). Balsam wooly adelgid (*Adelges piceae*), a non-native defoliating insect that attacks subalpine fir, caused a peak in defoliation in 2009 followed by mortality (Devine et al. 2012). There is concern regarding future potential for this species to cause additional mortality, including mortality of other true fir species (Devine et al. 2012). Western balsam bark beetles may also pose a threat for high-elevation subalpine fir, and weakened trees are more susceptible to the beetles. Hemlock looper (*Lambdina fiscellaria*), a defoliating lepidopteran insect, may have outbreaks on western hemlock causing high levels of defoliation within restricted patches. Outbreaks are initiated following periods of higher temperature and lower precipitation, which may become more common in the future (McCloskey et al. 2009). Other forest insects may be similarly favored by such climate changes (Lysak et al. 2006).

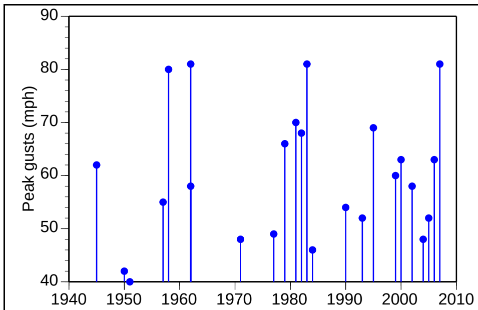

Figure 14. Peak gusts at Hoquiam WA of 24 storms between 1945 and 2010, as reported on the Washington State Climatologist web page (www.climate.washington.edu/stormking). Wind gusts of ca. 80 mph recur at approximately 20-year intervals.

The dramatic reduction of western white pine constitutes another new unique disturbance within the Park. This species may have been common throughout the peninsula but has suffered mortality from introduced white pine blister rust (*Cronartium ribicola*). Western white pine was detected in only 5% of the 697 forest ecology plots on the peninsula (Figure 7), which likely reflects a reduction from an unknown level prior to the blister rust. Whitebark pine, the other five-needle pine (subgenus Strobus) in the region that is susceptible to the blister rust, is widely scattered in small clumps at high elevation in the eastern and northeastern peninsula. This highly discontinuous pattern may be slowing the spread of blister rust.

Herbivory by ungulates, especially elk, constitutes another form of disturbance. It is unclear, over long periods (e.g., hundreds of years) what level of grazing occurred in coastal forests. Elk grazing creates glades in the Sitka spruce zone by limiting tree recruitment, especially maple and cottonwood in riparian areas (see above). The extirpation of wolves from the peninsula in the early 1900s has resulted in a rapid increase in elk and reduced recruitment of riparian trees (Beschta and Ripple 2008). Thus, recent high levels of herbivory may be viewed as a punctuated modern disturbance for plant populations. This decline of tree regeneration may have reduced bank stability and resulted in a widening of the active channel of the Queets River within the park versus outside the park where elk populations are smaller (Beschta and Ripple 2008).

This relatively new expanse of bare alluvium, in turn, reduces the flows of biota and nutrients between riparian forests and the active river channel.

The Olympic marmot creates disturbances in meadows from selective herbivory and burrowing activities that create large mounds of soil. Moderate levels of marmot activity may reduce abundance of the most common species and therefore increase diversity, but where marmot populations are high palatable species are greatly reduced and ruderal (weedy) species increase (Del Moral 1984). Black bear damage to young forests is very widespread on the Olympic Peninsula. One bear may girdle 60-70 trees per day while feeding on phloem, with preference for trees 20-40 cm in diameter (Ziegltrum 2004). Management outside the park has created extensive stands of trees suitable for such feeding and bear damage has become an economic and wildlife management issue.

Geomorphic disturbances are frequent due to the steep terrain within the Park, the sedimentary bedrock, and the high rainfall and snowfall. An inventory of landslides by Gerstel (1999) based on aerial photographs over approximately half of the western peninsula outside of the Park, identified 465 deep-seated (below rooting depth) landslides. Another database of mapped landslides shows that 3.8% of the mountainous landscape of the south-central portion of the Park has the signature of a landslide disturbance (Figure 15). The greatly enhanced precipitation on ridge-tops is a major factor affecting the geographic distribution of landsliding (Minder et al. 2009). Indeed, the orientation of landsliding is biased to the south and southwest aspects (Figure 15), which receive more rainfall than leeward aspects. The same orientation of landslides was noted in a similar setting on Vancouver Island (Jakob 2000).

Very large deep-seated paleo-landslides may have been triggered by major earthquakes. A series of rock avalanches in the southeastern Olympics all date to the period of 1000 to 1300 years ago, based on radiocarbon dating of drowned trees in landslide-dammed lakes (Schuster et al. 1992). Another series of very large landslides from Storm King Mountain dammed the ancient Lake Crescent and changed its outlet to the Lyre River. While some of these landslides may have occurred within the last 500 years, the age of the largest landslides remain unknown (Logan and Schuster 1991). Both areas of landslides mentioned here occurred on basalt bedrock, which may require seismic activity to produce major landslides (Schuster et al. 1992).

The coast of the Olympic Peninsula has experience repeated seismic events due to its proximity to the Cascadia subduction zone, which lies offshore. However, the geology of the peninsula itself seems to be particularly resistant to seismic activity. Only 14 earthquakes of magnitude 5.5 or greater have occurred near the peninsula since 1965 (http://earthquake.usgs.gov/earthquakes/eqarchives/). Another source indicates only four other notable earthquakes earlier in the century (Yeats 2004). Three other events far from Washington produces tsunamis with minor impacts on the peninsula (Yeats 2004). It is clear from the recent seismic record that the peninsula experiences significant hazard risk from activity along the Cascadia Subduction Zone, as half of the recent earthquakes were identified as “offshore”.

Paleoseismic evidence for large subduction-zone earthquakes is increasing. Atwater (1987) suggests that methods involving judicious stratigraphy can pinpoint locations of subduction-zone earthquake related subsidence in coastal low-lying areas, and that the 1964 Seattle earthquake caused such subsidence, mostly in estuaries. Willapa Bay sediments recorded seven such events in the last 7000 years (Atwater 1987). In 1700 AD, a major earthquake of magnitude 9 affected much of the Pacific Northwest southward from the southern peninsula (Atwater and Yamaguchi 1991, Atwater et al. 2005). The impact of these events on coastal villages is discussed in chapter 6.

Large landslides provide opportunities for primary succession in a landscape where other large disturbances are a rare occurrence (excepting recent logging). High light levels and exposed mineral soils on fresh landslides are ideal for the establishment of Douglas-fir and pine species that otherwise require high-severity fire for establishment (Geertsema and Pojar 2007). Considering the aerial extent and frequency of natural landsliding, these events are likely crucial in maintaining a wide diversity of disturbance-adapted species within the park (Swanson et al. 1988, 2010).

Today, forest harvest constitutes the main form of disturbance on the peninsula. An inventory of late-successional old-growth forest from Landsat imagery finds that old growth forests are concentrated in the western half of the park, as historical fire disturbances reduce late-successional forest area in the Elwha watershed and further east, but also that few patches of old growth larger than 2 ha exist outside of the park or wilderness areas (Moeur et al. 2011). Currently, the natural disturbance regimes outside of the park are likely very different from that within the park due to extensive young-aged stands and changed species composition in forests replanted after harvest.

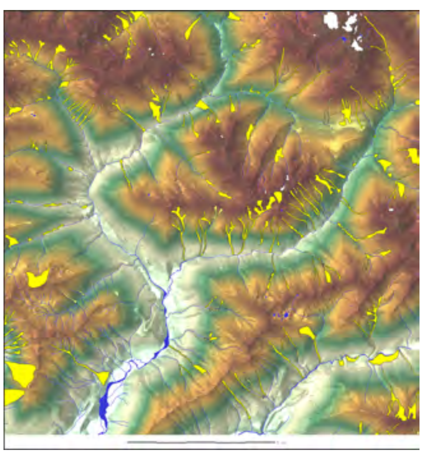


Figure 15. Left: Mapped landslides (yellow) over the upper Quinault River in the Olympic National Park. Recent landslides and older revegetated landslides are not differentiated. From the Washington State Landslide Database (http://www.dnr.wa.gov/ResearchScience/Pages/PubData.aspx). Right: The Grouse Creek Landslide of 1997, which was triggered by a rain-on-snow event. The runout of the landslide is 2.4 km. Photo by Karl Wegmann (http://www.panoramio.com/photo/6558592).

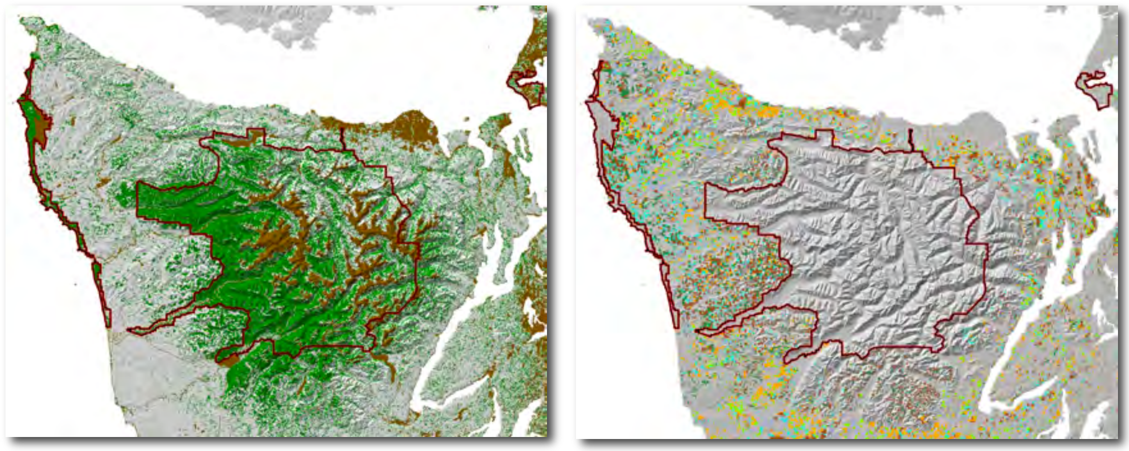

Figure 16. Left: Late successional old growth forest (green) and non-forested areas (brown). Right: Harvested forest stands from 1984 to 2002. Colors indicate the year harvested. National Park land shown in red outline. See (Moeur et al. 2011) for remote sensing methods.

3. Geology and historical biogeography of the Olympic Peninsula

> "The gap that separates the Olympic Mountains from the Cascade Range, …is only about a hundred miles wide and is filled by the dense forest already mentioned, affording continuity of range…, but the higher parts of the Olympics… have been disconnected from corresponding zones to the north and east for at least several million years, a period long enough to admit a considerable amount of differentiation in the species stranded here..."
>
> – C. Hart Merriam, undated manuscript, quoted in Schultz (1994)

3.1 Late Cenozoic geologic and climatic history

During the Eocene, about 50 million years ago, a deep marine rift formed parallel to the continent as the oceanic plate moved obliquely to the continental plate (Figure 17). This rift may have caused immense amounts of lava to accumulate, forming a large region of basalts under the ocean. Sediments accumulated on both the continental and oceanic side of the basalts. As the oceanic plate pushed against the continental plate, both the oceanic sediments and the basalt were compressed and turned upward. This process of subduction under the continental plate resulted in the accretion of a large wedge of sediments against the continent, which were pushed vertically and forced into an eastward-plunging arch (Tabor 1975). Younger sedimentary rocks were the last to be pushed against the continent. Because the basalt mass, and the continent behind it, were much more rigid than the oceanic sediments, the marine sandstones, siltstones and conglomerates making up the core of the Olympic Mountains were greatly disrupted and folded. The horseshoe shape of the basalt Crescent Formation (Figure 18) may have resulted from the ocean plate forcing the basalt mass into the "inside corner" between Vancouver Island and the continent. Beginning around 20 million years ago, the lighter sedimentary rocks rose relative to the dense oceanic crust leading to the uplift of the Olympic Mountains. Ages on the central core rocks of the peninsula, dated by the zircon fission track method, suggests the uplift of the mountains has been occurring for 17 million years and that the erosion of the terrain is in rough equilibrium with the uplift (Pazzaglia and Brandon 2001, Wegmann and Pazzaglia 2002). However, there remains much uncertainty about the age of the marine sedimentary rocks, which are often mapped within broad time periods (e.g., Miocene to Eocene).

During the uplift of the Olympic Mountains, from the middle Miocene to present, the Olympic Mountains experienced large climatic changes. A warming in the Middle Miocene, the Miocene Climatic Optimum from 18 to 16 Ma, interrupted a general global cooling trend during the Cenozoic. Vegetation at this time in the Pacific Northwest indicates summer moisture and humidity sufficient to maintain a diversity of broadleaf deciduous trees. The climatic changes from the Miocene into the Pliocene involved a steepening of the equator-to-pole temperature gradient, which would have strengthened the Pacific subtropical high and increased summer aridity in western North America (Wolfe 1985). This climate change resulted in great losses of the deciduous element that required warm and moist summers while promoting expansion of conifer forests. In addition, as uplift of the Cascade and Sierra Nevada Ranges proceeded, the interior rainshadow climate developed and resulted in the development of xeric steppe vegetation. Expansion of conifers and woodland vegetation continued into the Pliocene. Despite this major change in vegetation formation, 90 genera of the Pacific Northwest Miocene flora occur today on the Olympic Peninsula (Buckingham et al. 1995). Beginning 2.6 million years ago (the Pleistocene), global climate completed its transition from a "greenhouse" world that lacks ice sheets to an "ice-house" world marked by cyclic glaciations in the northern hemisphere. Glacial periods lasting ca. 90,000 years were marked by rapid climate changes and ice sheets extending over most of Canada. In contrast, interglacial periods lasting ca. 10,000 years resulted in species expanding their ranges into higher latitudes. The Late Pleistocene and the Holocene (current interglacial) are the focus of the remainder of this report.

3.2 Endemism, disjunction, and the insular nature of the Olympic biota

The insular (island-like) nature of the plant and animal life of the Olympic Peninsula has been widely noted (Houston et al. 1994, McNulty 2009). Several features of the biodiversity on the peninsula are consistent with long-term isolation and a functioning of the mountains is ice-age refugium during the Pleistocene. First, there are 27 plant and animal taxa known to be endemic to the Olympic Peninsula. There are also a large number of species on the peninsula that are disjunct from their closest conspecifics by hundreds of kilometers. Second, there are also 13 species common in the Cascade Range that are absent from the peninsula, suggesting the operation of a long-term barrier to dispersal.

The endemic taxa of the Olympic Peninsula are found in a small set of habitat types on the peninsula, suggesting that these habitats are a long-term feature of the peninsula though not necessarily in their current extent and spatial distribution. Of the 27 endemic taxa, 13 have ranges restricted to subalpine or higher elevations in the north or northeast. In particular, five species are associated with scree or rocky slopes. These dry meadow and woodland habitat is more extensive in the modern Rocky Mountains and may have been more common in the past on the Olympic Peninsula. Another eight taxa are associated with small river settings and shaded sites on the southwestern Olympics. One species, the Pacific coast tiger beetle, is a specialist of sand dunes. Other endemic species may have become extinct in the Pleistocene, such as a trechine beetle discovered in the beach cliff at Kalaloch (Cong 1997). Species continue to be discovered (e.g., Barr 2011) and therefore these patterns are only likely to be strengthened with further taxonomic work.

Another set of species on the peninsula is disjunct by long distances (100s of km) from the remainder of their population, providing additional evidence for the existence of Pleistocene refugia. This list of species is almost too long and difficult to compile, but generally fall into two groups. One group of such disjunctions is a pattern of distribution on the peninsula and in the boreal and tundra biomes of the Yukon and Alaska. These species likely had ranges split during the Pleistocene glaciations, with the Olympic Peninsula constituting an important southern refugium. An example of this pattern is the long-stalked whitlow grass. The other pattern has a distribution on the peninsula disjunct from an interior Rocky Mountain distribution but absent from the Cascade Range. Examples of this pattern are the least-bladdery milkvetch and western hedysarum (Houston et al. 1994).

Isolation of the Olympic Peninsula has resulted in several species that have not naturally colonized the peninsula from the Cascade Range. Notable mammals and birds in the group include grizzly bear, wolverine, red fox (introduced in the 20th century), coyote, lynx, water vole, golden-mantled ground squirrel, northern bog lemming, porcupine, pika, mountain sheep, mountain goat (introduced in the 20th century), and the white-tailed ptarmigan. Absent trees include noble fir, ponderosa pine, subalpine larch, western larch, and western juniper (Houston et al. 1994). The mountain goat was introduced to the peninsula in the 1920's. Recognition that this species, which impacts meadows and endemic plant species, was historically absent prompted efforts by the Park to control populations. Lyman (1995) has questioned the strength of evidence supporting its non-native status, as the fossil record within the mountainous part of the peninsula is very limited. For other invasive species on islands, there is some precedent for paleoecological studies to challenge the assumed invasive status of certain species (Lyman 2006, Willis et al. 2007).

In contrast the unique biogeography resulting from Olympic Peninsula's insular nature, another set of biogeographic patterns result from its limited connectivity to the archipelago of landbridge islands to the north (connected to mainland during lower sea levels) and to drier mountain ranges to the east. Houston et al. (1994) highlight several plant species that are endemic to the landbridge islands with their southern limit on the Olympic Peninsula, suggesting a northward expansion from an Olympic refugium. Another set of species are restricted to dry sites on the Olympic Peninsula, but are common in the eastern Cascades and/or Rocky Mountains (Englemann spruce, whitebark pine, and quaking aspen). Kuramoto and Bliss (1970) suggest that the phlox-bunchgrass community of the northeastern subalpine meadows is a remnant of the Early Holocene. However, considering the number of endemic taxa this vegetation type supports, it might have a longer history on the peninsula.

In summary, the distributions of several plant species and communities suggest the existence of glacial refugia in the north and northeastern Olympic Mountains (Houston et al. 1994). We note that both the alpine northeastern mountains and southwestern peninsula contained substantial unglaciated areas during the glacial periods. However, many endemic taxa are subspecies or otherwise highly related to taxa off of the peninsula and thus are likely 'neoendemics' that differentiated since deglaciation. Other endemics are far from their sister taxa both geographically and evolutionarily, likely date to deeper evolutionary splits, and may be considered 'paleoendemics'. The potential that such endemics remained in situ through the major climate changes of the Late Quaternary is intriguing but strong support for this would require both phylogeographic studies and a better understanding of past climate at fine (microclimatic) resolution (Hampe and Jump 2011).

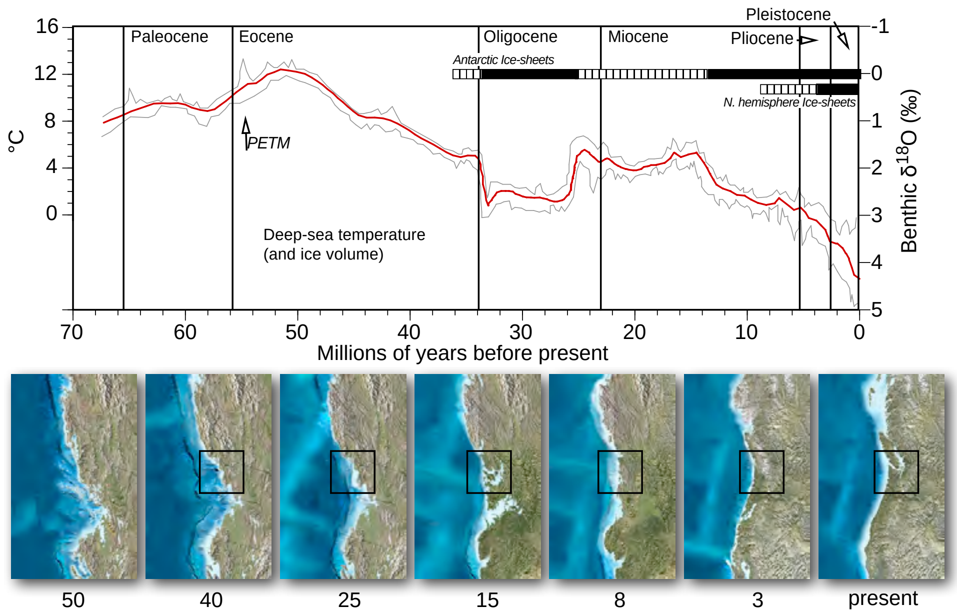

Figure 17. Top: Global climate trajectory during the Cenozoic inferred from oxygen isotope ratios of benthic ocean sediment (Figure modified from (Zachos et al. 2001). Bottom: Paleogeography of western North America. Colors correspond to vegetation cover. Maps from Ron Blakey, Colorado Plateau Geosystems, Inc (http://www.cpgeosystems.com/paleomaps.html).

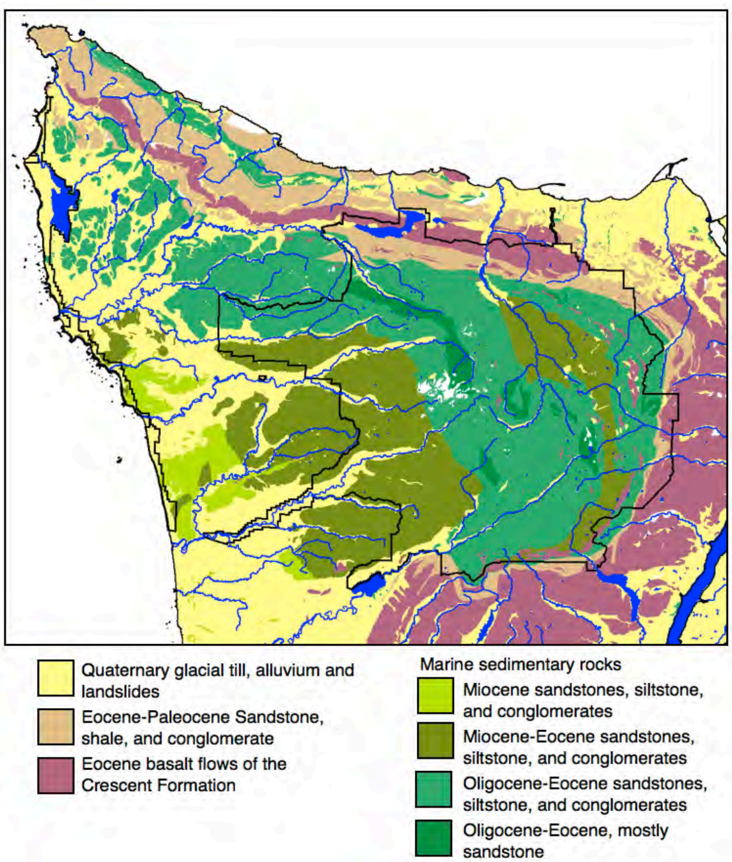

Figure 18. Geological map of the Olympic Peninsula showing major units differentiated by age class. Data from (Washington Division of Geology and Earth Resources staff 2008).

Table 1. Endemic taxa of the Olympic Peninsula. Based on lists from Park Service web pages (http://www.nps.gov/olym/naturescience/endemic-animals.htm), Houston et al. (1994), Buckingham et al. (1995), and McNulty (2009). Note some taxa have ranges extending partly outside the peninsula, and published lists differ depending on geographic extent off of the peninsula. Note this list does not include at least 10 species occurring on Vancouver Island and the Olympic Peninsula.

Scientific Name	Common Name	Habitat association (and quadrant on Peninsula)	Source (for habitat association)
Vascular plants			
Astragaus australis var. *olympicus*	Olympic Mountain milkvetch	Subalpine, open sites (northeastern mountains)	Buckingham et al. 1995
Campanula piperi	Piper's bellflower	Montane to alpine, rocky sites (Northeast to central)	Buckingham et al. 1995
Corallorhiza maculate var. *ozettensis*	Spotted coral root	Lowland, partial shade (Southwest)	Tisch 2001
Erigeron flettii	Flett's fleabane	Subalpine to alpine, open sites (Northeast to south)	Buckingham et al. 1995
Erigeron peregrinus ssp. *peregrinus* var. *thompsonii*	Thompson's wandering fleabane	Lowland, bog sites (Southwest)	Buckingham et al. 1995
Erythronium quinaultense	Quinault fawn lily	Lowland, open or partially open (Southwest)	Allen 2001
Petrophytum hendersonii	Olympic rock mat	Montane to alpine, rocky sites (South, north and east)	Buckingham et al. 1995
Senecio neowebsteri	Olympic Mountain groundsel	Subalpine to alpine, scree sites (North, northeast, and central)	Buckingham et al. 1995
Synthyris pinnatifida var. *lanuginosa*	Olympic Mountain synthyris	Subalpine to alpine, scree sites (North, northeast, and central)	Buckingham et al. 1995
Taraxacum olympicum	Olympic Mountain dandelion	Subalpine to alpine, open sites (North)	Buckingham et al. 1995
Viola flettii	Flett's violet	Subalpine to alpine, rocky sites (North, northeast, and central)	Buckingham et al. 1995
Amphibians			
Rhyacotriton olympicus	Olympic torrent salamander	Lowlands to montane, steep gradient streams (Southwest)	Adams and Bury 2002
Mammals			
Marmota olympus	Olympic marmot	Subalpine, open meadows (Throughout)	Edelman 2003
Tamias amoenus caurinus	Olympic yellow-pine chipmunk	Subalpine, forest and parkland (North and northeast)	Sutton 1992, Demboski and Sullivan 2003
Scapanus townsendii olympicus	Olympic snow mole	Subalpine (Northeast)	Johnson and Yates 1980, Carraway et al. 1993

Scientific Name	Common Name	Habitat association (and quadrant on Peninsula)	Source (for habitat association)
Thomomys mazama melanops	Olympic Mazama pocket gopher	Meadows, young forest (North and northeast)	Verts and Carraway 2000
Mustela erminea olympica	Olympic ermine	Throughout (North, central and southeast)	Hall 1945
Fish			
Novumbra hubbsi	Olympic mudminnow	Low gradient rivers, muddy sediment (South)	McPhail 1967
Orthoptera (grasshoppers)			
Nisquallia olympica	Olympic grasshopper	Subalpine to alpine, scree sites (Throughout?)	Rehn 1952
Lepidoptera (butterflies and moths)			
Hesperia comma hulbirti	Hulbirt's skipper	Subalpine to alpine	Lindsey 1939
Oeneis chryxus valerata	Olympic arctic	Subalpine to alpine (Northeast)	Yake 2005
Coleoptera (beetles)			
Nebria danmanni	Mann's gazzelle beetle	Montane to subalpine (Northeast)	Kavanaugh 1981
Nebria acuta quileute	Quileute gazelle beetle	River banks at mid elevation (North)	Kavanaugh 1979
Cicindela bellissima frechini	Pacific coast tiger beetle	Sand dunes and deflation plains (West)	Leffler 1979
Bryelmis rivularis	Riffle beetle	Streams 3-6 m wide, woody debris (West, and NW Oregon)	Barr 2011
Diplopoda (millipedes)			
Tubaphe levii	Olympic peninsula millipede	Lowlands (West)	Causey 1954
Mollusks			
Hemphillia burringtoni	Arionid jumping slug	Dense riparian forest	Burke 2005

4. Postglacial climate on the Olympic Peninsula

4.1 Forcing factors in the climate system

The three major controls of Earth's climate - incoming solar radiation (insolation), ice extent, and greenhouse gases - have changed dramatically from the glacial maximum to the present (Figure 19). Earth's orbit varies slowly on cycles lasting from 20,000 to 100,000 years. These Milankovitch cycles involve the shape of the orbit (eccentricity), the tilt of Earth's axis (obliquity) and the seasonal precession of the perihelion, or closest earth-sun distance, which all combine to affect the latitudinal distribution and seasonal cycle of incoming solar radiation (insolation). These cycles are the pacemaker of glaciations, as the onset of ice sheet growth occurs when northern hemisphere summer insolation is low. The extent of glacial ice sheets provides a positive feedback in the climate system. Extensive ice sheets during the Last Glacial Maximum (LGM, 24-19 ka or thousands of years ago) reflected short-wave solar radiation back to space, affecting the global balance of incoming and outgoing radiation, as well as modifying the locations of high and low pressure systems, the location of jet streams, and routes of moisture to continental interiors. Greenhouse gases, including carbon dioxide and methane, were lower during the LGM, thus increasing the amount of outgoing longwave radiation escaping from Earth's surface and atmosphere to space. The large-scale controls on Earth's climate were successfully simulated in the Cooperative Holocene Mapping (COHMAP) project which provided the first strong link between data and climate models on the climates since the LGM (COHMAP Members 1988, Kutzbach et al. 1998).

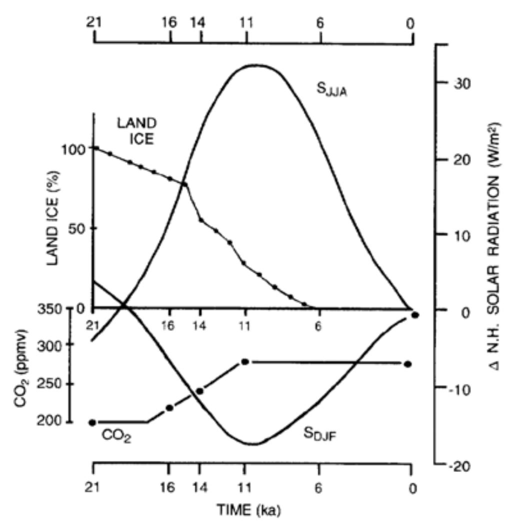

Figure 19. Major forcing factors of climate change for the past 21,000 years. S: Summer (JJA) and winter (DJF) insolation anomalies for the northern hemisphere; CO_2 concentrations and land ice (percent of maximum cover) are also shown. Reprinted from (Kutzbach et al. 1998).

In addition to these well-known forcing agents of climate change, other components of the climate system add shorter-term variability, resulting in a complex history marked by variability occurring over a wide range of timescales. Solar forcing due to variation in solar intensity, although of low magnitude, may become amplified by affecting upper-atmospheric ozone concentrations. Volcanic activity may shield insolation for several years. Changes in large-scale features of ocean and atmospheric circulation may experience sudden shifts, especially during deglaciation, producing abrupt climate changes, some of which may persist for decades to millennia (Gavin et al. 2011). During the glacial period, these fluctuations were pronounced and are referred to as stadial (cold) and interstadial (warm) events.

Paleoclimatologists use a variety of methods to reconstruct these fluctuations and to attribute the fluctuations to factors that are known to affect the climate. The discussion below summarizes a selection of well-dated paleoclimate reconstructions and evidence of glacial fluctuations occurring on and nearby the Olympic Peninsula. Changes in vegetation and fire occurrence are purposely avoided in this section because they are responses to the climate change that may lag the actual climate changes by more than 100 years. In contrast, more direct proxies of climate should respond with much shorter lag times and are normally more straightforward in their climatic interpretation compared to vegetation proxies (Shuman et al. 2004).

4.2 The Late Pleistocene and the Last Glacial Maximum: >60–19 ka

The late Pleistocene, preceding the Last Glacial Maximum (LGM, 24-19 ka), was a period of alternating cold glacial advances and warmer periods (Figure 20). Large valley glaciers on the western peninsula repeatedly advanced and formed extensive moraines, one of which created Lake Quinault (Figure 21). Radiocarbon dates on these moraines indicate maximum glacier extents older than 50 ka (Lyman Rapids advances), ca. 38 ka (Hoh Oxbow I advance), ca. 32 ka (Hoh Oxbow 2 advance, which was the greatest ice extent of the western Olympics during the late Pleistocene) and ca. 22 ka

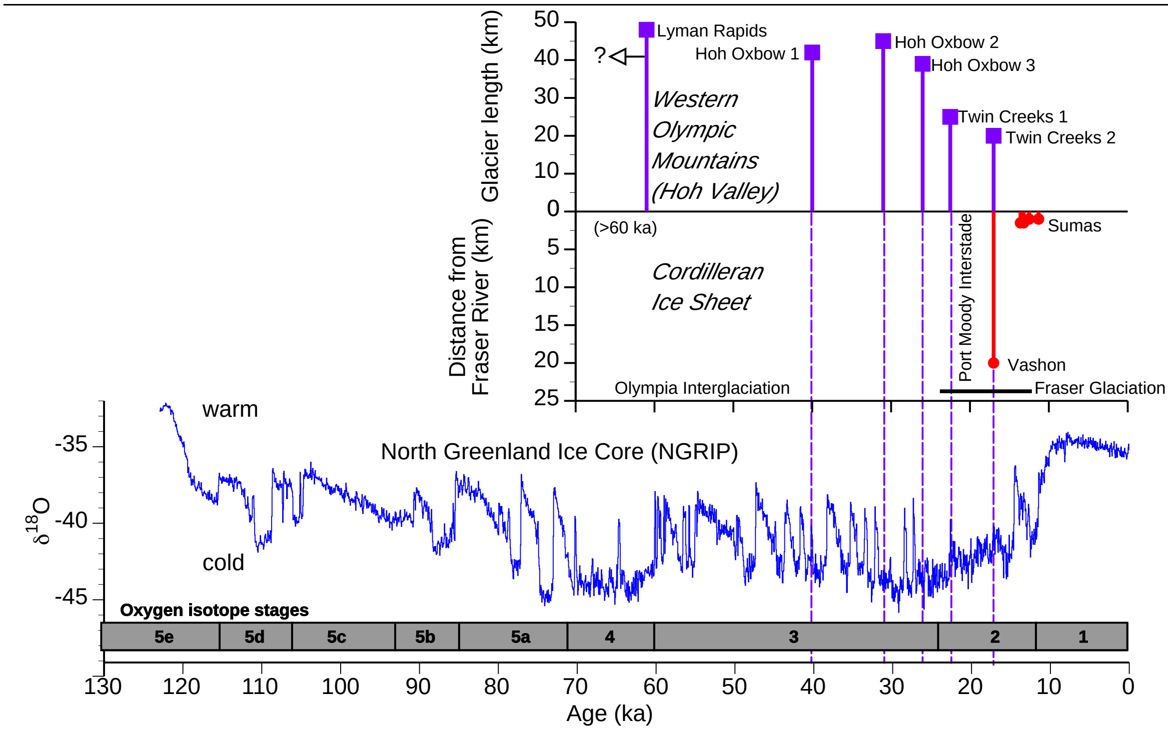

Figure 20. Top: Extents of late Pleistocene glacial advances in the Hoh Valley of the western Olympic Mountains (purple) and of the Cordilleran Ice Sheet in the Puget Sound (red). Dates are from Booth et al. (2003) and Thackray (2008). Bottom: The Greendland oxygen isotope record (North Greenland Ice Core Project members 2004) indicating temperature in Greenland at 50-year intervals to the last interglacial (marine oxygen isotope stage 5e). Note that the dates of the Hoh glacial advances generally date to stadial (cold) periods during stage 3 (dashed lines) and that the very extensive Lyman Rapids glacial advance falls during the cold stage 4.

(Twin Creeks 1 advance; Thackray 2001, 2008). These events correlate broadly with alpine glaciers in the Skagit Valley (Riedel et al. 2010) and on Mount Rainer (i.e., the Evans Creek Stade; Booth et al. 2003). The climate of the period between glaciations was not much colder than present. Fossil beetles can be used to reconstruct past temperatures based on the tolerances of the species in an assemblage. A fossil beetle assemblage from a beach cliff at Kalaloch showed that the latter part of the Olympia Interglacation, 48–40 ka, was only about 1°C colder during July compared to present (Cong and Ashworth 1996).

Glacial activity increased the supply of sediment to the rivers, which created broad floodplains and carved into hillslopes of the western Olympic Mountains. Rivers on the western Olympics have now incised through these floodplains and into bedrock leaving them as strath terraces (Pazzaglia and Brandon 2001). Despite these glacial advances on the western Olympics, there is little evidence of advances of Cordilleran ice

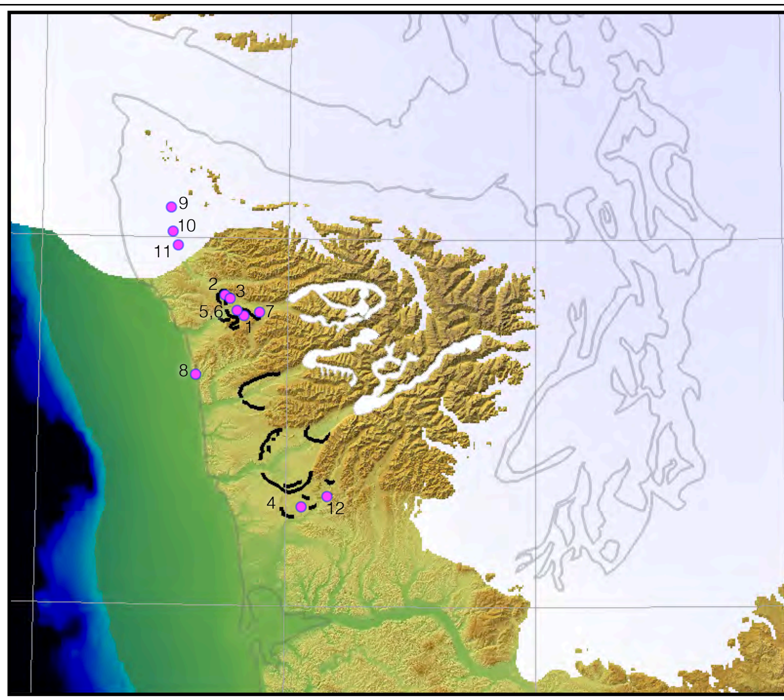

Figure 21. The maximum extent of the Vashon Stade advance into Puget Sound and the Strait of Juan de Fuca, paleogeography based on a 120-m decrease in sea level, and modern coastlines shown by a gray line (though this is an overestimate as it does not consider isostatic depression, see James et al. (2009). Alpine glaciers show extent of the Twin Creeks 2 Stade (Thackray 2001, which was likely coeval with the Vashon Stade of the Cordilleran Ice Sheet. (17 ka). Black lines show moraines dating to the maximum late Pleistocene extent of the valley glaciers in the Hoh, Queets, Quinault, and Humptulips rivers, dating to ca. 23 ka. Glacier extents are from Porter and Swanson (1998) and Thackray (2001); numerous alpine glaciers are not mapped. Numbered sites are lakes and bogs studied by Cal Heusser and discussed in Chapter 5: 1–3: Hoh River Valley sites 1-3 (Heusser 1964) ; site 2, Bogachiel Bog, was also studied by (Hansen 1941, Heusser 1978). 4: Humptulips Bog (Heusser 1964, Heusser et al. 1999) 5–7: Hoh bogs 1–3 (Heusser 1974). 8: Kalaloch sea cliff (Heusser 1972). 9–11: Wessler Bog, Wentworth Lake, and Soleduck Bog, respectively (Heusser 1973).

into Puget Sound, though such evidence may have been obliterated by the subsequent Vashon Advance (Booth et al. 2003).

During the LGM, the seasonal pattern of insolation was similar to the present day although extensive ice sheets were located to the north and atmospheric CO_2 was as low as 180 ppm (65% of preindustrial levels). The high albedo and high elevation of the Cordilleran and Laurentide ice sheets produced very cold air. This cold air resulted in a strong high-pressure system that produced a clockwise (anticyclonic) air flow, which was especially strong during winter. The anticyclonic winds spun off the ice sheets producing an east-to-west wind south of the glacial limits in Washington State; the loess hills of eastern Washington date to the Pleistocene when these winds were strong. The westerly flow of the winter jet stream was thus diverted to the south, resulting in less annual precipitation than occurs today (Kutzbach et al. 1998).

The climate of the LGM was colder than present, though few quantitative climate reconstructions date to this time period. Beetle assemblages from the Kalaloch sea cliff that date to the period of the Hoh Oxbow 2 advance up to 19 ka indicate a treeless environment and temperatures at least 3 °C colder than present (Cong and Ashworth 1996). Regional estimates derived from pollen data suggest that the mean annual temperature was 5–8 °C colder than today (Thompson et al. 2003, Bartlein et al. 2010). A quantitative reconstruction from pollen data from a bog near the Humptulips River indicate a ca. 5 °C decrease in summer temperature during the LGM (Heusser et al. 1999). In agreement with these estimates, a model of glacier growth for the LGM achieved the best fit to the observed ice extent under a summer temperature of 7.2 °C (from current sea-level temperature of 14 °C) and precipitation 40% of modern (Hellwig 2010).

After the LGM, a short interstadial event in southwest British Columbia dubbed the Port Moody interstadial, was warm and moist enough to support a subalpine forest vegetation (Lian et al. 2001). The authors speculated that the general view of a LGM climate that was cold and dry, as predicted from the COHMAP models, was not generally correct through the entire period, but rather significantly warm interstadials occurred. Quantitative temperature reconstruction from fossil beetles near Seattle dating to 18 ka indicate summer temperatures to be only slightly (ca. 2 °C) colder than present but significantly colder during the winter (Miller et al. 1985, Ashworth et al. 2000, Ashworth 2003).

4.3 The Late Glacial: 19 to 11.6 ka

The period of deglaciation at the end of the Pleistocene, between 19 and 11.6 ka, involved fluctuations on timescales of centuries, decades, and year-to-year, as well as abrupt changes. During this period summer insolation was increasing and atmospheric CO_2 was increasing (Figure 19). The period is marked by a complex series of climate changes as the location of the jet stream was adjusting as the ice sheet was waning, which also resulted in large outburst floods that affected ocean circulation. As the ice sheets retreated north, the glacial anticyclone weakened and the westerly jet stream moved northward resulting in increased onshore flow and increased precipitation in the Pacific Northwest. A large increase in moisture at 17 ka caused glacial advances throughout the Pacific Northwest. This resulted in the maximum extent of Cordilleran ice during the

Vashon advance into Puget Sound (Figure 21). This correlated in time to the last major advance of valley glaciers on the Olympic Peninsula (the Twin Creeks II advance of Thackray 2001). The Vashon lobe had completely retreated by 16 ka. This rapid retreat was facilitated by calving into a proglacial lake dammed south of the ice front (Porter and Swanson 1998). Most subalpine cirque lakes in the Pacific Northwest date to 14.5 ka, indicating that alpine glaciers did not retreat to their near-modern extent for another 1500 years after the Vashon retreat. The loss of the alpine glaciers occurred very close in time to the onset of the Bølling-Allerød warm interstadial event (14.7-12.9 ka) that is well known in Europe.

Many paleoclimate proxies are available from the beginning of the Bølling-Allerød since that is the time in western Washington when most lake basins were deglaciated. The next major event to occur was the Younger Dryas stadial between 12.9 and 11.6 ka. This event, which is a distinct event in the northern Atlantic, has been identified to various degrees in the Pacific Northwest. Some records show that the Younger Dryas-like climate event in the Northwest slightly lagged events in the North Atlantic (Mathewes et al. 1993). However, more recent well-dated climate records from a speleothem (stalagmite) in Oregon Caves (Vacco et al. 2005), and the organic content of lake sediments on the Olympic Peninsula (Gavin et al. in review), are synchronous with the Younger Dryas. A reconstruction of sea-surface temperatures indicates a nearly 4 °C cooling during the Younger Dryas relative to today, while a July temperature reconstruction from British Columbia based on fossil midges (chironomids) reveals a more muted cooling of ca. 2 °C (Figure 22).

The re-advance of Cordilleran ice in southwest British Columbia during the Late Glacial, termed the Sumas Stade, has generated much discussion. A series of moraines in the Fraser River Lowland date to 13.7 to 13.2 ka (Clague et al. 1997, Kovanen and Easterbrook 2002), which agrees with a moraine age of 13.5 ka in a nearby fjord (Menounos et al. 2009). Another moraine at Howe Sound north of Vancouver dates to the earliest part of the Younger Dryas at 12.8 ka (Friele and Clague 2002, Menounos et al. 2009) Kovanen and Easterbrook (2002) also report two minor re-advances that date to within the Younger Dryas Chronozone. However, a detailed pollen record from near the Fraser Lowland indicates that the Younger Dryas climate record is one of gradual warming followed by a cold interval, occurring too late to be consistent with the latter Sumas advances (Pellatt et al. 2002). A further discussion provides some doubt that some of the Younger-Dryas glacial re-advances are correct (Easterbrook 2004, Pellatt et al. 2004). Several moraines in the southern Coast Mountains of British Columbia and Mount Baker have moraines with poorly constrained ages that may date to the Younger Dryas (Osborn et al. 2012). Regardless of the exact chronology, several glacial advances occurred in the southern Coast Mountains between the 13.7 and 12.8 ka.

4.4 The Early Holocene: 11.6 to 6 ka

The increased summer insolation during the Holocene led to an intensified Pacific Subtropical High pressure system, which created warm, stable, dry air to its east (i.e., the Pacific Northwest). At the same time, decreased winter insolation may have resulted in winter temperatures colder than today and thus an overall more continental climate with greater seasonal cycle of temperature. This period has been broadly termed the

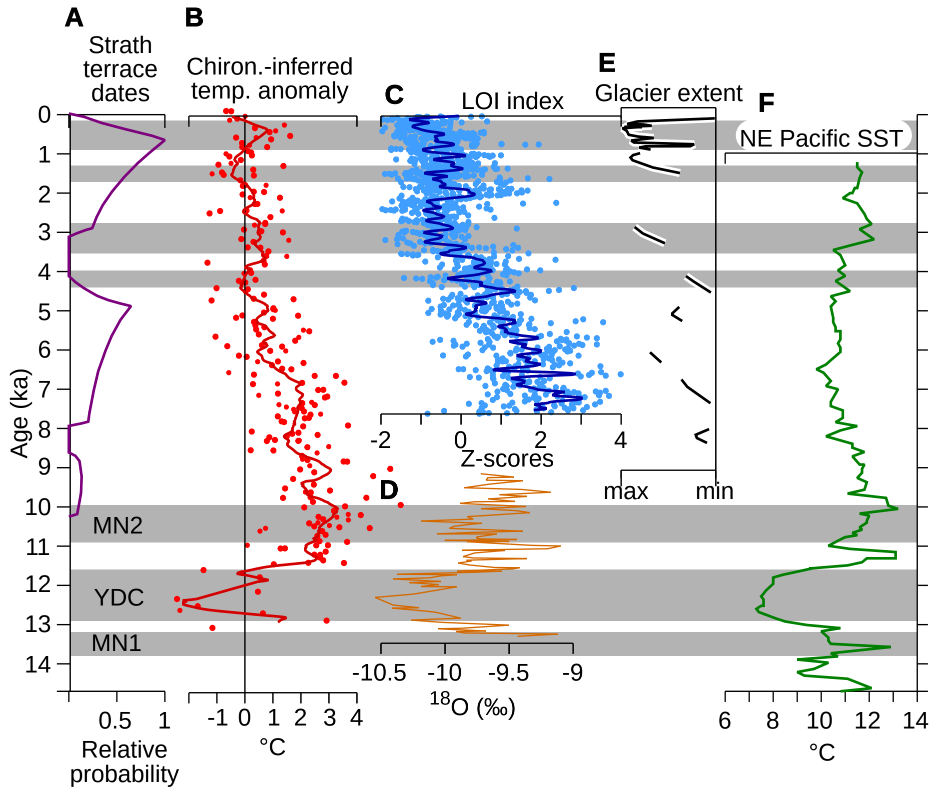

Figure 22. Paleoclimate indicators from the Pacific Northwest. A) The strath terrace dates are a summary of a set of 32 radiocarbon dates directly overlaying strath terraces on the Clearwater River west of Yahoo Lake (Wegmann and Pazzaglia 2002). B) The chironomid-inferred July temperature anomaly is based on four lakes in southern British Columbia. Points indicate the temperature anomaly from a 0-2 ka base period, and the smooth line shows a locally weighted (loess) smoothing in a 600-yr window and its standard error (gray lines). C) The LOI (loss-on-ignition) index is from glacial-fed lakes in southwest British Columbia that are sensitive to times of glacier advances (Menounos et al. 2008). D) The speleothem $\delta^{18}O$ record is from Oregon Caves National Monument (Vacco et al. 2005). E) Generalized alpine glacier extent in western Canada (Clague et al. 2009). F) The Northeast Pacific SST (sea-surface temperature) is inferred from alkenone records from a core near northern California (Barron et al. 2003) , which broadly agrees with a lower-resolution record closer to the Olympic Peninsula (not shown; Kienast and McKay, 2001). Horizontal shading indicates times of glacial advances described by Heine (1998) for the McNeeley (MN) 1 and 2 and by Menounos et al. (2009) for the advances from 4.2 ka to present. The Younger Dryas Chronozone (YDC), notably not a period of glacier advance in the region, is also shown.

'hypsithermal', the 'xerothermic' or the 'Holocene climatic optimum'. Fossil chironomid remains recovered from lake sediments in southern British Columbia suggest Early Holocene summer temperatures at 4°C warmer than present, with the warmest temperature broadly between 11.5 and 10 ka (Figure 22). In contrast, sea-surface temperatures off of northern California were only 1-2° warmer than present, which may be the result of a moderating influence of upwelling during the Early Holocene due to a moderately strong California Current (Barron et al. 2003).

The Early Holocene was likely not uniformly warm and dry through time, but rather marked by distinct century-scale periods of increased moisture. The remnants of the waning ice sheet to the north may have still been influencing the jet stream across western North America (Gavin et al. 2011). For example, dated lake sediments from Mount Rainier suggest a glacial advance between 11 and 10 ka (Heine 1998) which is synchronous with a cooling of sea-surface temperatures and cooler temperatures indicated in southern Oregon (Figure 22). Gavin et al. (in review) have found forest compositional changes on the Olympic Peninsula that support increased moisture between 11 and 10 ka. However, this interpretation does not entirely agree with the chironomid-based temperature reconstruction, though dating errors may prevent detection of short-term minor temperature change. Because similar glacial advances have not been found elsewhere in western Washington, the validity of the McNeeley 2 advance has been questioned (Menounos et al. 2009).

4.5 The Middle-to-Late Holocene: 6 ka to present

The millennial-scale decreasing trend in summer insolation through the middle Holocene resulted in a weakening of the Pacific Subtropical High and the onset of cooler summer temperatures. This latter part of the Holocene is termed the "neoglacial" because many alpine glaciers began advancement downslope about 4000 years ago. Many of these glacial advances were synchronous across the West, including an event at 3.3–2.9 ka (Figure 22). Of all the glacial advances during the Holocene, almost without exception the largest was a series of Little Ice Age glacier advances from 1350 to 1850 AD. During this cooling phase, the decadal-scale pattern called El Niño Southern Oscillation (ENSO) increased in frequency and intensity, with distinct millennial-scale variability (Moy et al. 2002). The ENSO pattern in the Pacific Northwest is associated with warm/dry winters alternating with cool/wet winters. A longer decadal-scale pattern called the Pacific Decadal Oscillation is superimposed on the ENSO pattern, but there is no paleoclimate record that has been able to document of PDO-like variability over millennial time periods.

5. Late Quaternary vegetation and fire history of the Olympic Peninsula

> "Climate changes during the late-Quaternary have been complex, and the details of the vegetational response to climate change are also complex."
>
> – Thompson Webb, *Is vegetation in equilibrium with climate?* (1986).
>
> "For every complex problem there is a simple solution. And it is always wrong."
>
> – H.L. Mencken.

5.1 Pioneering studies

The Olympic Peninsula has experienced a complex climate history and a complex vegetation history over the late Quaternary. Reconstructing both histories and understanding how they are linked mechanistically is a major goal in the field of Earth system science. Advances are made when there is intensive development of paleo-records and models in certain locations. The Olympic Peninsula has the potential to be such a location, as the area received attention from the pioneering years of pollen analysis up to the present.

The vegetation history of the Olympic Peninsula was first described by pioneering paleoecologists who were working when pollen analysis from lake sediments was a new approach. Henry P. Hansen developed dozens of pollen profiles in the Pacific Northwest while a PhD student at the University of Washington in the 1930s and as faculty at Oregon State University in the 1940s. His work preceded radioistopic methods of dating sediment (i.e., radiocarbon dating). For example, he described a coarse pollen profile from a low-elevation bog near the Bogachiel River (Hansen 1941) that was dominated by mountain hemlock and pine. Radiocarbon dating would have likely indicated he had reached Late Glacial sediments, but since there was no basis at the time to make this interpretation he speculated that pollen had been transported downhill in streams to his bog site.

In Hansen's (1947) monograph "Postglacial forest succession, climate, and chronology in the Pacific Northwest", he summarized over 70 pollen profiles and suggested that the "postglacial climatic trends have been much the same over the entire region, although varying in degree." Edward Deevey, in a review of Hansen's book, is more pessimistic about the claim that such a coherent trend exists in Hansen's data, and instead asserts that the data indicate complex successional patterns strongly influenced by fire and other disturbances (Deevey 1948). Nevertheless, the sequence described by Hansen remains largely valid: cool and moist Late Glacial, followed by increased warming and drying until 8000 years ago, followed by maximum warmth, and cooling during the last 4000 years.

Hansen (1947) provides a comprehensive review of the geography and vegetation and of the origin of lake and bog basins in the Pacific Northwest, pollen morphology, coring methods, and the pollen stratigraphy at several dozen sites. In the Puget Lowland, he describes the replacement of lodgepole pine by Douglas-fir as climate 'ameliorated' in the Late Glacial, and then increased western hemlock after the deposition of the Mazama tephra (Crater Lake eruption that is now aged to 7627±100 years ago). Hansen acknowledged that this sequence could be due to natural succession on the deglaciated landscape, as this sequence of species is ordered by increasing shade tolerance and with increasing requirements for organic and moist soils. However, Hansen also noted some synchronous fluctuations in western hemlock that are not expected due to succession and therefore favored a climate-change explanation. For example, he equated the Puget Sound Early Holocene vegetation to that of the Willamette Valley today. In contrast, along the coastal strip, Hansen found very little evidence for vegetation change during the Holocene and suggested that the hyper-maritime climate along the coast is buffered from regional climate change. These sites were from the Oregon coast where the relative stasis of Holocene vegetation change has been confirmed by subsequent studies (Long and Whitlock 2002). Hansen conducted no further studies from the western Olympic Peninsula.

5.2 The Late Pleistocene

Calvin J. Heusser conducted two decades of research on the vegetation and glacial history of the western Olympics. In 1960 he published the monograph "Late Pleistocene Environments of North Pacific North America" which summarized the state of knowledge of tree biogeography and climate history (Heusser 1960). Heusser included dozens of mostly undated pollen diagrams with a focus on southeastern Alaska but also included preliminary results from Washington and Oregon.

Heusser's first major study was the pollen analysis of three peat bog sections near the Hoh River and one longer section near the Humptulips River (Heusser 1964). The three pollen records from the Hoh, two of which had basal dates at ca. 15,000 ^{14}C yr BP (before present), showed the now well-known progression of pollen types of grass and pine and mountain hemlock transitioning to alder and eventually to western hemlock and Sitka spruce. The longer Humptulips record extended back to before 50 ka in under 6.3 m of fibrous and sphagnum peat and lake sediment. The record was marked by fluctuating levels of pine (likely shore pine) before the Holocene, with varying levels of Sitka spruce, grass, and sedges. Based on a data set of modern pollen assemblages up the Pacific coast, Heusser (1964) assigned pollen analogs to fossil samples to reconstruct the history of temperature and precipitation. These curves indicated a 10°F decrease in mean July temperature, a decrease in precipitation in the late Pleistocene, and fluctuations that crudely match the hypothesized sequence of glacial advances.

Heusser (1964) remarked that the Olympic Peninsula vegetation record was very dynamic relative to other locations on the Pacific coast. He also noted that the Humptulips pollen diagram indicated clear presence of trees through the peak of the LGM (the 'Late Wisconsin'), though sampling was coarse and dating control was marginal by today's standards. In particular, the LGM vegetation was interpreted as open parkland of mountain hemlock with some subalpine fir, and locations with Sitka spruce and shore pine along with a rich herbaceous community. The earlier 'Middle Wisconsin' had fewer tree species, when a valley glacier was only within 2 km of the site. Heusser's (1964) results support the assertion that southwestern Olympic Peninsula functioned as a refugium for tree species through the glacial periods, and suggested that such refugia did not likely extend northward into the purported small unglaciated areas north of Washington.

Heusser later developed a longer and more detailed late Pleistocene pollen record from the beach cliff exposure almost 30 m in height near Kalaloch, Washington (Heusser 1972). The pollen record is marked by large fluctuations between tree and herb pollen types, suggesting frequent changes in forest density related to temperature change and many submillennial-scale fluctuations not yet inferred from the glacial chronology. The upper part of the beach cliff reveals the transition from a mountain hemlock parkland prior to the LGM to a mostly treeless environment during the LGM. Earlier in the Pleistocene, during the Olympia Interglaciation, western hemlock and Sitka spruce pollen were abundant suggesting a nonglacial interval with an environment similar to modern montane forests. The Kalaloch chronology was re-interpreted by Thackray (2001) to extend back to the last interglacial period or between 105 and 125 ka, rather than the 70 ka estimated by Heusser. Regardless of the ages of the stadial (glacial) events in this section, the pollen data indicate that during these cold periods some trees, albeit at low density, persisted in the area and a high level of plant diversity was able to persist in this coastal refugium within tens of kilometers of extensive glaciation (Heusser 1972).

Additional pollen records obtained near the Hoh River, located south of the Juan de Fuca lobe but within mid-Wisconsin recessional moraines, provide a view of the Pleistocene refugium much closer to the glacial limits than seen at Kalaloch (Heusser 1974). One of the three bogs contained a sequence dated to ca. 21 ka, which revealed a tundra vegetation assemblage, with only ca. 10% lodgepole pine pollen, dominating until the Holocene warmth. Heusser (1974) noted that all of the pollen types in the diverse tundra-period in the Hoh bog have modern representatives on the peninsula, with the exception of *Selaginella selaginoides*, which is now found on northern Vancouver Island.

A combination of the Hoh peat core with the Kalaloch section provides a continuous pollen record spanning to before 50 ka (Heusser 1977). The interpretation from this sequence was confirmed by a peat profile near the Bogachiel River (Heusser 1978). The peat in this section is bracketed by radiocarbon dates between 30 ka and 20 ka, and the pollen record indicates mountain hemlock parkland transitioning into a tundra-like vegetation during the LGM. However, the rapid transition from the LGM to one typical of the Early Holocene suggests that much of the Late Glacial (16-12 ka) is missing from this record.

The combined Hoh and Kalaloch pollen record provided an opportunity to use pollen data with multivariate statistics to reconstruct the late Quaternary temperature. Heusser et al. (1980) compared the long pollen record with a network of modern pollen composition from northwest California to south-central Alaska to reconstruct mean July temperature and annual precipitation. This method assumes that modern vegetation is in equilibrium with climate (i.e., there are no migration lags or other factors affecting tree distribution other than climate), which is still debated. By extracting four principal components from the pollen data, which are most related to pine, hemlock, and grass pollen types, transfer functions were constructed relating these new variables to climate, though no measures of uncertainty were presented. Applied to the fossil record, the results show that the middle Holocene was slightly (1.5 °C) cooler than present, the Early Holocene was similar to present, and the LGM was 4 °C cooler than present. These

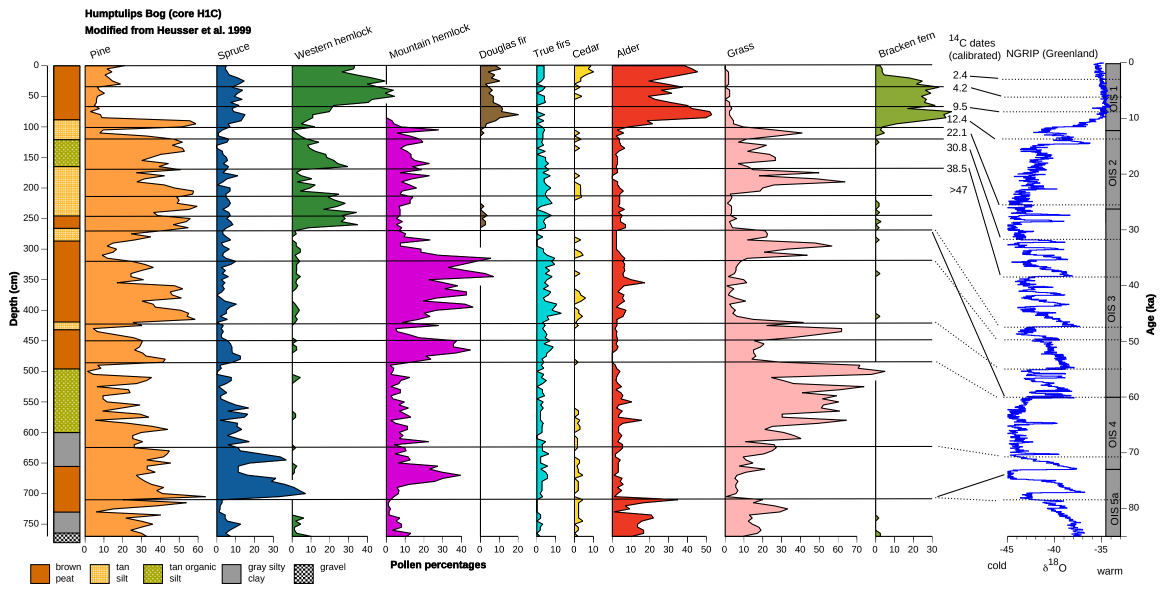

Figure 23. Pollen diagram from Humptulips Bog, redrawn from Heusser et al. (1999) and abbreviated to a few key pollen types. Horizontal lines are pollen zones defined by Heusser et al. The oxygen isotope record from northern Greenland (NGRIP) and the marine-based oxygen isotope stages (OIS) are shown for correlation with the pollen record. Solid sloping lines on the right indicate age assignments by Heusser et al. (1999) while dashed sloping lines are alternate age possibilities based on the timing of interstadials in the NGRIP core (North Greenland Ice Core Project members 2004).

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results are a considerable underestimation of the actual variability currently reconstructed from pollen data (Bartlein et al. 2010). The precipitation reconstruction shows a dry Early Holocene consistent with evidence of fire and Douglas-fir, and a dry LGM (ca. 70% of modern precipitation). Both temperature and precipitation are much more variable during the period prior to the LGM.

Late Pleistocene vegetation was described in greatest detail by a revision of the Humptulips Bog core originally published in 1964 (Heusser et al. 1999). A 7.7 m bog core was analyzed for pollen at 5-cm intervals, and the upper 170 cm could be dated by radiocarbon before ages were too old for radiometric dating. An abbreviated version of this pollen diagram, showing ten important pollen types, is shown in Figure 23. This pollen record reveals the Holocene as distinct from the rest of the core, suggesting to Heusser et al. (1999) that the base of the core did not reach the previous interglacial. The Holocene vegetation is marked by high values of alder and western hemlock, with consistently high values of Douglas-fir and bracken, suggesting regional fire disturbances. A distinct fluctuation in grass and pine at 90 cm depth is consistent with the Younger Dryas, though below that period radiocarbon ages suggest a hiatus exists in the core at the time of the LGM. The lower 6.7 m, however, has generally lower alder and greatly fluctuating levels of grass and pine pollen. Heusser et al. (1999) correlated the entire period with low western hemlock and alder pollen but high in grass and mountain hemlock to oxygen isotope stage (OIS) 4, which compresses the majority of the core to a 12,000-year period of time. An alternate interpretation is that the units of silt and silty clay correlate to OIS 4 and other fluctuations in grass pollen date to the numerous interstadial (warm) events that occurred in OIS 3 (see dashed lines in Figure 23). Resolving this will likely require obtaining new core material and using advanced means of absolute dating such as optically-stimulated luminescence on quartz grains.

In addition to the Humptulips bog, Heusser et al. (1999) analyzed a peat section on the West Fork Humptulips River, occurring under several meters of laminated silty lake sediment. This peat section was marked by western hemlock, Sitka spruce and alder pollen. Based on its infinite radiocarbon age (>50 ka) and the similarity to modern forests, Heusser suggests that it dates to the previous interglacial period (ca. 120 ka, or OIS 5e).

Regardless of the details of the Humptulips chronology, this record clearly reveals that for the majority of the Pleistocene the regional vegetation was significantly different from today and fluctuated greatly through time. Heusser et al. (1999) estimated a ≥5°C decline during the LGM and several times when tree line occurred close to the site, which is at 100 m above sea level. The biota that is believed to have persisted on the peninsula though the Pleistocene, therefore, did so associated within a regional vegetation that was characterized by mountain-hemlock parkland and a tree-line tundra environment, which returned several times during the Pleistocene and persisted for several millennia each time.

5.3 The Late Glacial and Holocene

Heusser (1973) obtained more detail on the Late Glacial and Holocene vegetation changes of the western Olympic Peninsula from three additional bog and lake pollen records located just north of the southern limit of the Juan de Fuca Lobe (Heusser 1973). Wessler Bog, located 6.2 km east of the north tip of Ozette Lake, has a radiocarbon date at its base of its 10-m core at 14,460 ^{14}C yr BP, which calibrates to an age of 17.5 ka. This suggests an initial uneven retreat of the ice at the same time or slightly earlier than the retreat of the Vashon Lobe in the Puget Trough (Heusser 1973). Interestingly, many small diameter trees found embedded in glacial till in the same area dated to roughly 13,100 ^{14}C years BP (calibrated to ca. 15.8 ka). Heusser (1973) interpreted this as lodgepole pine trees that quickly colonized on ablation moraines as the ice sheet was stagnating for many centuries. Eventually the final loss of the ice after ca. 15.8 ka resulted in collapse of moraines and burial of the colonizing forest. While this sequence of events is paralleled in the Puget Trough, the exceptionally long period of stagnating ice inferred from these dates needs to be confirmed with better constraint on the timing of deglaciation.

The pollen records from Wessler Bog, Wentworth Lake, and Soleduc Bog indicate parallel environmental sequences (Heusser 1973). The sites were colonized by lodgepole pine followed by the warm Early Holocene period with alder and small quantities of Douglas-fir (presently very rare in the region) with a gradual transition to cool-moist late-successional cedar-hemlock forests. Few radiocarbon dates constrain these pollen records. A more detailed record from Wentworth Lake is presented later in this chapter.

Other pollen-based temperature reconstructions were conducted at sites in southwest British Columbia. The statistical method of Heusser et al. (1980) was applied to a higher-resolution 14,000-year pollen record from Marion Lake near Vancouver, British Columbia (Mathewes and Heusser 1981). These results generally confirmed the previous study, but showed a more distinct 'xerothermic' period in the Early Holocene that began to become cooler and drier a few centuries before the Mazama ash (7.6 ka).

The extensive work by Hansen and Heusser remains the best view into the LGM and earlier environments on the peninsula, but unfortunately there remain two limitations to many of these studies. First, the radiocarbon dating methods used may have been susceptible to contamination of older carbon from groundwater sources, causing some dates to be older than that of the associated pollen record. The overall insufficiency of dates at all sites limits their utility for regional correlation. Second, the pollen of Cupressaceae, the family that contains western redcedar and Alaska yellow cedar, was inconsistently identified. Heusser did indicate these taxa on his diagrams but pollen percentages almost never exceeded 2%; Bogachiel Bog and the Humptulips cores are the exceptions (Heusser 1978, Heusser et al. 1999). Although this pollen type is sometimes difficult to distinguish from algal cysts or other pollen types, palynologists have been able to reliably identify it using modern chemical processing techniques to obtain a cleaner sample for microscopic identification.

Other studies have found that the Cupressaceae record is very dynamic. Canadian palynologists Rolf Mathewes and Richard Hebda showed that western redcedar increased to modern levels of abundance in a south-to-north pattern, from 8 ka or earlier in southern Washington to 4 ka on northern Vancouver Island. As Native American cultures utilize large-diameter western redcedar, the availability of this important resource was likely quite recent in coastal British Columbia (Hebda and Mathewes 1984).

Other researchers developed detailed pollen records in the Puget Trough in the 1980s (Barnosky 1981, Tsukada et al. 1981, Leopold et al. 1982, Sugita and Tsukada 1982). These records, constrained by multiple radiocarbon dates, provided a more detailed ecological interpretation of the Late Glacial and Holocene vegetation and climate histories. A synthesis of pollen records indicated a 1000-m lowering of alpine treeline during the LGM, with humid conditions occurring only on the western Olympic Peninsula (Whitlock 1992). At the time of maximum ice extent of the Puget Lobe, mesic subalpine parkland characterized many of the low-elevation sites both east and west of the Olympic Mountains. Following the retreat of the Vashon Lobe, the arrival of temperate taxa resulted in rapid vegetation changes between 15 and 10 ka (Whitlock 1992). These vegetation changes will be the focus of the next section of this chapter with a focus on five postglacial pollen records from the Olympic Peninsula.

5.4 Postglacial vegetation and fire at five sites on the Olympic Peninsula.

Beginning in 1993, a paleoecological study of the Olympic Peninsula was initiated by Linda Brubaker from the University of Washington. Sediment cores were collected from small lakes located at sites spanning across both the precipitation and the elevational gradient. These records have been published with the exception of Wentworth Lake (a site originally studied by Heusser 1973), but these sites have not been analyzed simultaneously to demonstrate the concurrent changes in vegetation across the range of vegetation zones on the peninsula. In the remainder of this chapter these pollen records are presented in new diagrams. They are described briefly and the significant findings are highlighted. Then, the sites are summarized using a modern analog method in which the pollen data are compared with modern pollen data to link past vegetation to modern forest zones. In the next section, these sites are synthesized with the existing paleoecological records from the region, especially southern Vancouver Island and the Puget Lowland.

5.4.1 Study sites

Wentworth Lake is a 6 m deep, 14 ha kettle lake located at 47 m asl, 7 km SE of the southern tip of Ozette Lake in the northwest Olympic Peninsula. The terrain around most of the lake is level and therefore the size of the watershed is indeterminate, and there are no inflowing streams. Aerial photographs indicate fluctuating water levels, with a 3 m drop in water from early to late summer, exposing the NW portion of the lake. The forests around the lake are composed of western redcedar, western hemlock and Sitka spruce. This site was first studied by (Heusser 1973).

Yahoo Lake a 17.8 m deep, 3.7 ha cirque lake located at 717 m asl on a ridge above Stequaleho Creek and 5.5 km north of the Queets River. The lake has a watershed of only 9.6 ha and no inflowing streams. The forests are composed on western hemlock, western redcedar, Pacific silver fir, and minor amounts of Douglas-fir and Sitka spruce.

Martins Lake is a 9.6 m deep, 0.8 ha small moraine-dammed lake. It is the larger of two lakes located at 1423 m asl on a ridge in the center of the park on the north side of Mount Christie and south of Low Divide. The watershed is very restricted (ca. 3 ha) and there are no streams inflowing or outflowing. The lake is in subalpine parkland, with small patches of mountain hemlock and Pacific silver fir. Common meadow vegetation is red heather, white mountain heather, American bistort, Cascade blueberry, glacier lily, and sedge species. Sitka alder occurs in avalanche chutes. The site has been published in Gavin et al. (2001).

Moose Lake is a 7.8 m deep, 3.7 ha cirque lake located at 1544 m asl in Grand Valley 7 km SE of Hurricane Ridge. Its watershed, ca. 400 ha in size, includes steep slopes extending to above 2000 m asl which feeds three intermittent inflowing streams. The watershed is composed of talus, scree, and parkland of subalpine fir. Less common trees include mountain hemlock, Alaska yellow cedar, and lodgepole pine. Meadows host a high diversity of herbaceous species, with distributions primarily controlled by moisture availability. The site has been published in Gavin et al. (2001) and Brubaker and McLachlan (1996).

Crocker Lake is a 5.1 m deep 26 ha kettle lake located at 54 m asl in the Leland Valley in the northwestern peninsula, 6.5 km south of Discovery Bay. The watershed of Crocker Lake includes the area drained by Andrews Creek, estimated to be over 1000 ha. Forests in the area are heavily managed Douglas-fir stands; red alder and western redcedar were likely important historically and especially along the streams. Despite being in the Western Hemlock Zone, this late-successional species is rare because of dominance of Douglas-fir after disturbance in this drier climate. The site has been published in McLachlan and Brubaker (1995) and Brubaker and McLachlan (1996).

Coordinates for all study sites are in Table 2 and climographs of mean monthly temperature and precipitation are in Figure 4.

5.4.2 Field and laboratory methods

Sediment cores were taken using a Livingstone piston corer operated from a plywood platform on inflatable rafts that were anchored over the deepest part of each lake (Wright et al. 1984). Cores were wrapped in aluminum foil and transported to the University of Washington. Coring was performed during the summers of 1993 and 1994.

Radiocarbon dates were obtained on bulk sediment using conventional dating or on plant macrofossils using accelerator mass spectrometry (AMS). Two to seven dates were obtained from each site, or 24 dates in total. Chronologies were developed using a smoothing spline to the median age of each calibrated radiocarbon date (Reimer et al. 2009). Bulk radiocarbon dates were assessed for potential contamination by carbonate sources by comparison with independent information on glacial retreat dates or the ages of the Mazama tephra. Chronologies were revised at the three published sites (Crocker, Moose, and Martins lakes).

Sediment core organic matter was quantified using loss-on-ignition at 550°C. Pollen analysis was conducted on 1-cm^3 samples following standard methods (Faegri and Iversen 2000) with the addition of a 7 µm sieve step for basal samples with high clay content (Cwynar et al. 1979). Sitka alder and red alder pollen were separated based on pore morphology (Sugita 1990). Pollen percentages were based on a sum of at least 350 terrestrial pollen grains.

Plant macrofossils and charcoal were analyzed in sediment subsamples taken at intervals between 1 and 10 cm. For Wentworth Lake, 54 samples were taken contiguously at interval of approximately 250 years, corresponding to 7-20 cm intervals in the core. For Yahoo Lake, 415 samples, each 5 cm^3, were taken contiguously at 1 cm intervals. For Martins Lake, 47 samples were taken contiguously at 5-cm intervals. For Moose Lake, 41 subsamples were taken contiguously at 7 to 25-cm intervals. For Crocker Lake, 26 subsamples each spanning 10 to 25 cm of core, were taken irregularly down core. For all lakes, sediment subsamples were wet sieved at 0.5 mm for charcoal and 1.18 mm for macrofossil identification. For all sites except Yahoo Lake, sediments were soaked in sodium hexametaphosphate solution prior to sieving. For Yahoo Lake, sediment subsamples were wet sieved at 0.15 and 1.18 mm. The >1180 µm fraction was saved for macrofossil identification. The 150–1180 µm fraction was treated with a 5% KOH solution at 40°C for 20 minutes and then washed through 500 and 150 µm sieves with a gentle flow of water. The 150–500 µm fraction was saved for charcoal analysis. Plant macrofossils and charcoal particles were tallied under 20X–70X magnification. Macrofossils were identified using published keys for conifer foliage (Dunwiddie 1985) and a reference collection at the University of Washington. Needle fragments were combined and expressed as needle equivalents and *Thuja* branchlets were tallied individually following Dunwiddie (1987). Charcoal was expressed as an accumulation rate (CHAR; pieces cm^{-2} yr^{-1}).

Pollen data from the fossil cores were compared to the modern pollen assemblages on the Olympic Peninsula using the modern analog technique. This approach calculates a dissimilarity coefficient between each fossil pollen assemblage and

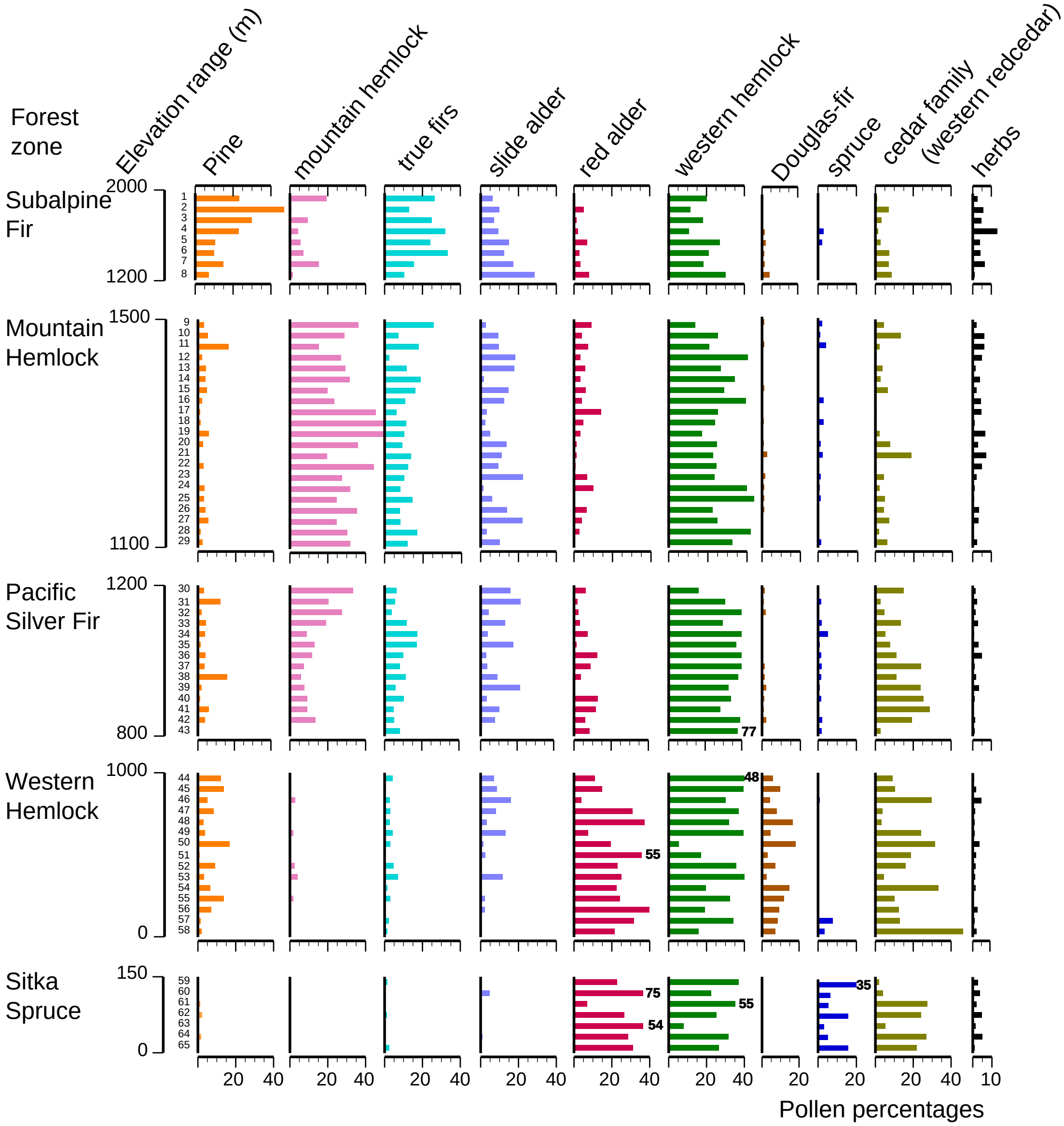

Figure 24. Modern pollen assemblages from 65 small lakes on the Olympic Peninsula. These data were compared to fossil pollen assemblages using the modern analog technique. Figure modified from Gavin et al. (2005).

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each modern pollen assemblage and assesses whether the coefficient falls below a threshold indicating a pollen analog. The modern assemblages come from surface-sediments from 65 small lakes located in areas where vegetation is little disturbed; heavily logged areas were avoided (Figure 24, Gavin et al. 2005). Heusser (1969) conducted a similar study on the Olympic Peninsula but used subcanopy (moss polster) pollen assemblages that showed a strong pollen/vegetation correlation. However, those sites are not adequate analogs for lacustrine pollen assemblages because pollen assemblages from subcanopy locations are characteristically different than those from lakes, especially near the treeline (Minckley and Whitlock 2000).

We used the square chord distance dissimilarity coefficient, calculated as $D_{m,f} = \sum_{k=1}^{n} \left(\sqrt{p_{m,k}} - \sqrt{p_{f,k}}\right)^2$ where $D_{m,f}$ is the squared chord distance between a modern and fossil assemblage, and $p_{m,k}$ and $p_{f,k}$ are pollen proportions of pollen taxa k in modern and fossil assemblages, respectively, and there are a total of n pollen taxa. Gavin et al. (2005) found that distances of 0.15 to 0.30 occurred among modern assemblages within vegetation zones on the Olympic Peninsula. Those threshold distances could be used as conservative and liberal criteria, respectively, for assessing analogs.

5.4.3 Sediment stratigraphy and chronology

The 937 cm sediment core from Wentworth Lake is composed of silty clay with occasional sandy layers of 1-3 cm in length below 530 cm (LOI 2–5%), transitional to brown uniform gyttja from 530-490 cm, and organic gyttja above 470 cm (LOI 30–45%). The five AMS radiocarbon dates fell on a smooth line in the zone above the silty gray clay and in line with a very faint but discernable Mazama tephra (Figure 25). The silty clay layer contained no pollen and we believe it was rapidly deposited. No basal radiocarbon date was obtained.

The 430-cm core from Yahoo Lake is composed of uniform light gray silt in the lower 82 cm (LOI 4–8%), with the exception of an organic transitional layer within the silt unit (LOI up to 40%). The upper 348 cm is finely laminated organic sediment (LOI 45–70%) with the exception of a 1-cm Mazama tephra. The five AMS radiocarbon dates fall on a smooth curve and in line with the Mazama tephra.

The 277-cm core from Martins Lake is composed of uniform silty clay (LOI 2–8%) in the lower 136 cm and laminated organic gyttja (LOI 13–25%) in the upper 137 cm; these units are separated by a 4 cm Mazama ash. Two AMS radiocarbon dates fall on a smooth curve in line with the Mazama tephra. Lack of macrofossils for dating the lower portion of the core necessitated correlating the pine pollen decline at Martins with the same feature at Yahoo Lake.

The 708-cm core from Moose Lake is composed of silty clay with sand layers (LOI 4–6%) in the basal 158 cm, and uniform organic gyttja (LOI 10–25%) in the upper 550 cm with a 1-cm narrow Mazama tephra. The age model used in Gavin et al. (2001) was revised after considering a likely error in one of the conventional (bulk) radiocarbon dates. The conventional date at 565-575 cm places the distinct decline in pine pollen over 1000 years earlier than it occurs regionally. A very old age of 30 ka at 695 cm is good indication that old carbon is contributing to the radiocarbon ages at the base of the core. Assigning an age of 11.6 ka to the pine decline, in contrast, results in a pollen stratigraphy in line with the established regional chronology. We found that the decline of pine pollen to very low levels is complete by 11.6 ka consistently at three sites in the region (Tsukada et al. 1981, Leopold et al. 1982, Brown and Hebda 2002).

. Table 2. Radiocarbon dates on five sediment cores from the Olympic Peninsula.

Depth (cm)	Lab Code	^{14}C age ± 1 SD	Material dated or age basis[1]	Calibrated age
Wentworth Lake				
90-92	AA-20319	3285 ± 65	Conifer needles	3520 (3380-3640)
183-188	AA-20320	6210 ± 120	Conifer seeds and cone	7100 (6830-7330)
206-206.5			Mazama tephra[2]	7627 (7477-7777)
281-285	AA-20321	8540 ± 120	Conifer seeds and alder seed	9530 (9260-9910)
444-445	CAMS-25018	10,980 ± 160		12,880 (12,590-13,180)
510-520	AA-20322	11,985 ± 140	Seeds	13,840 (13,460-14,160)
Yahoo Lake				
109-111	CAMS-35107	3620 ± 60	Conifer needles	3940 (3820-4090)
198-199	AA-20315	5845 ± 60	Conifer needles	6660 (6500-6790)
230-231			Mazama tephra[2]	7627 (7477-7777)
293-295	AA-20316	8175 ± 110	Conifer needles	9140 (8850-9440)
356-357	AA-20317	9980 ± 115	Conifer twig and wood	11,510 (11,210-11,830)
407-409	AA-20318	11,915 ± 95	Conifer needles	13,770 (13,480-13,980)
Martins Lake				
71-73	AA-20323	1995 ± 100	Conifer needles	1960 (1710-2160)
122-123	CAMS-32689	5390 ± 80	*Tsuga mertensiana* cone	6180 (5990-6310)
137-141			Mazama tephra[2]	7627 (7477-7777)
232		Correlation with regional *Pinus* decline		11,200
Moose Lake				
150-160	Beta-63814	3280 ± 70	Bulk sediment	3510 (3370-3640)
350-360	Beta-63815	6130 ± 90	Bulk sediment	7020 (6790-7250)
413-414			Mazama tephra[2]	7627 (7477-7777)
480-488	AA-20324	6705 ± 70[4]	woody material	7570 (7460-7680)
565-575	Beta-63816	11,040 ± 70[3]	Bulk sediment	12,930 (12,710-13,110)
590		Correlation with regional *Pinus* decline		11,600
695-705	Beta-63817	30,709 ± 1490[3]	Bulk sediment	35,280 (31,900-38,580)
Crocker Lake				
200-210	Beta-60299	1780 ± 70	Bulk sediment	1700 (1550-1870)
450-460	Beta-60300	4760 ± 80	Bulk sediment	5490 (5320-5610)
601-611	Beta-60301	7530 ± 130[3,4]	Bulk sediment	8330 (8150-8560)
613-614			Mazama tephra[2]	7627 (7477-7777
800-810	Beta-60302	9830 ± 140[3]	Bulk sediment	11,280 (10,760-11,770)
900		Correlation with regional *Pinus* decline		11,600
950-960	Beta-60303	11,540 ± 240[3,5]	Bulk sediment	13,410 (12,900-13,900)
1174	Beta-70546	10,420 ± 60[6]	Wood	12,300 (12,090-12,530)
1174	Beta-70549	11,340 ± 60[7]	Leaves	13,200 (13,110-13,340)

1. Bulk sediment ages are conventional radiocarbon dates, while all other dates on plant material are accelerator mass spectrometry radiocarbon dates.
2. Mazama tephra date is from Zdanowicz et al. (1999)
3. Date not used due to likely old carbon in bulk sediment
4. Date not used because not in stratigraphic order with Mazama tephra
5. Small sample (0.4 gram) with extended running time
6. Date not used because post-dates known glacial retreat by > 1000 years.
7. Date above glacial sediments from Cedar Swamp (McLachlan and Brubaker 1995), here assigned to base of Crocker Lake core.

The 1181 cm core from Crocker Lake is composed of silty clay with sand layers (LOI 2–13%) in the lower 251 cm and organic gyttja (LOI 20–40%) for the upper-most 930-cm with the exception of a 1-cm Mazama tephra. The age model used in McLachlan and Brubaker (1995) was revised after considering two likely errors in the current set of dates. First, a conventional radiocarbon date 13 cm above the Mazama tephra was 800 radiocarbon years older than the known age of the tephra, suggesting that other conventional dates in the core are also too old due to old carbon fixed by algal productivity in the lake (see Brown 1994 for precision dating of the Mazama tephra at

Crocker Lake using AMS dates on a pollen extraction). As the conventional dates are considered unreliable, we used a correlation of the decline in pine pollen as used at Moose Lake. Second, an AMS radiocarbon date near the base of the core is 900 years younger than the base of Cedar Swamp, a site only a few km away (McLachlan and Brubaker 1995). We therefore use the Cedar Swamp date under the assumption that deglaciation was synchronous. We also note, however, that the Leland Creek spillway, in which Crocker Lake is located, was actively draining a proglacial lake behind the Vashon lobe 2000 years earlier (Porter and Swanson 1998). The landscape may have supported dead-ice and been unstable for many centuries before the lake basins formed and organic matter began contributing to the core sites, such as at the other low-elevation site (Wentworth Lake).

Pollen diagrams are plotted with macrofossil occurrences overlaid on the pollen percentages (Figures 26-30) and the modern analog analysis results are plotted for all sites together (Figure 31).

5.4.4 Late Glacial: 14–11.6 ka

A very dynamic environment existed during the several millennia following deglaciation. Climatic fluctuations were pronounced as the waning Cordilleran and Laurentide Ice Sheets were shrinking in size. A landscape of recently deposited till was eroding rapidly before the last stagnant ice melted and vegetation could stabilize soils. This process of landscape stabilization, though often considered to be very rapid in the Pacific Northwest (e.g., Whitlock 1992), likely was variable over space depending on dates of deglaciation, topography, and the nature of the till deposits. Lodgepole pine was the dominant species colonizing the landscape. Today, lodgepole pine is restricted on the peninsula, either in early successional forests in the Subalpine Fir Zone or as a coastal variety restricted to muskegs.

Forest cover first appeared on the western Olympic Mountains shortly after deglaciation of the lake sites. Prior to 14.2 ka, the treeline was below 700 m on western Olympics, as indicated by high grass, sedge, and sagebrush pollen (Figure 27). No other pollen record exists at the same time at lower elevations. Stagnating ice during the retreat of the Juan de Fuca lobe may have precluded a pollen record at Wentworth Lake, while Heusser's sites (Heusser 1973, 1974, Heusser et al. 1999) extend further back but resolve this time period poorly. After a decline in pine at 13.6 ka at both Wentworth and Yahoo lakes, a diverse subalpine forest established at Yahoo (mountain hemlock, Pacific silver fir, western hemlock and occasional Sitka spruce) and Sitka spruce–western hemlock established at Wentworth Lake. At Yahoo Lake, distinct charcoal peaks indicate that intense fires occurred at least six times before 11.6 ka. The coarser Wentworth Lake charcoal record indicates that the highest CHAR values occurred at 13.0–13.6 ka. The combination of species that existed at this time is a poor analog to any modern forest on the Olympic Peninsula (Figure 26).

Unlike the wet west coast sites, the subalpine sites and the rainshadow site showed very low forest density during the Late Glacial that were similar to the Subalpine Fir Zone on the peninsula today, though very sparse tree cover, or only tundra, existed at high elevations. Martins Lake was high in pine and alder (both regionally dispersed pollen taxa) and had very low pollen influx, low organic matter, no charcoal, and no macrofossils. Moose Lake also had high pine percentages, and the lack of any macrofossil or charcoal in the core also indicates sparse or no tree cover.

The low elevation site in the rainshadow (Crocker Lake), supported a diverse macrofossil assemblage in the deepest sediments, containing Sitka spruce, subalpine fir, Douglas-fir, and lodgepole pine needles[1]. The age of these sediments are poorly constrained, but estimated to be 13 ka based on correlation with nearby Cedar Swamp. The macrofossil sample comes from a 50-cm sediment subsample that may integrate much time. These needles and the high charcoal value at the base of the core may represent soil developed on the stagnating ice landscape that later became submerged in the lake basin. Low pollen accumulation rate and occurrence of herbaceous and fern pollen (up to 40%) suggests very open vegetation, which included slide alder in avalanche tracks. Overall, the surprising combination of species (i.e., subalpine fir and Sitka spruce) suggest a no-analog vegetation type.

Additional pollen records from the region support this picture. Nearby, the Manis Mastodon site located in the center of the rainshadow at Sequim indicates open vegetation, including cactus, from 13.8 to 12.9 ka (Petersen et al. 1983). On Vancouver Island, (Brown and Hebda 2003) found a herbaceous-dominated vegetation prior to 14.6 ka followed by a lodgepole pine woodland, as found at Yahoo Lake. The latter period of lodgepole pine woodland or parkland vegetation coincides with the warming associated with the Bolling Allerod period and the widespread retreat of cirque glaciers throughout the Pacific Northwest.

The vegetation responses to the climate changes of the Younger Dryas Chronozone (YDC, 12.9–11.6 ka) are variable among the sites. Paleoclimate proxies from both southern Oregon and British Columbia reveal the YDC as a distinct cold event, and thus the Olympic Peninsula likely experienced the same event (Figure 22). Yahoo Lake reveals this event most clearly as a period of low organic matter, decreased pollen influx, and fewer charcoal peaks. Wentworth Lake had few macrofossils and low charcoal, and Moose Lake had a low pollen accumulation rate, during the YDC. The balance of evidence suggests a colder climate that was perhaps also drier, that caused reduced biomass and fewer fires (Gavin et al. in review).

The major vegetation change at the end of the YDC, marked by a decline in pine, may have been locally controlled as soils at each site stabilized and began to support mesic forests at different times. However, dating control at this time was poor at Moose, Martins and Crocker, such that the pine decline itself was used to assign ages. Improved radiocarbon chronologies are needed to resolve how this vegetation transition varied over space.

In summary, the pollen records indicate that a very strong rainfall gradient, as occurs today, also occurred during the Late Glacial. On the western peninsula, the occurrence of mountain hemlock forest at mid elevations, but not higher, suggests a depressed treeline while at Crocker Lake, on the east side, was higher pine pollen and

[1] The presence of the Douglas-fir and lodgepole pine needles, not published in McLachlan and Brubaker (1995), was identified by Gavin from archived material.

indicators of open parkland vegetation. All sites except Wentworth Lake showed a similarity, for at least part of the Late Glacial, to the Subalpine Fir Zone (Figure 31). However, the high elevation sites were likely in a very sparse parkland or tundra; even the forest at Yahoo Lake only showed similarity to the Subalpine Fir Zone during the YDC.

5.4.5 Early Holocene: 11.6 – 6 ka

The start of the Early Holocene is marked by a great increase in warmth and dramatic changes in forest composition across the Olympic Peninsula. Yahoo Lake exemplifies the rapid turnover in forest composition. At 12 ka, the site supported a diverse mix of subalpine and low-elevation species: Sitka spruce, western hemlock, mountain hemlock, and Pacific silver fir, similar to what occurs in southeast Alaska today (except for presence of Pacific silver fir). Douglas-fir pollen appears first shortly after 12 ka, and a fire (revealed by a sharp charcoal peak) at 11.6 ka resulted in the initial loss of mountain hemlock and pine, and a decline in spruce. However, it was not until a second fire at 11.2 ka that Pacific silver fir was lost and replaced by Douglas-fir, as shown by the macrofossil record. Red alder dramatically increases, and after 11.2 ka the landscape is one of frequent fire with disturbance-adapted Douglas-fir and red alder, with a minor component of western hemlock. One fire at 10.8 ka resulted in a return of Sitka spruce to the site, a successional pattern not expected for this site, but Douglas-fir returned during a period of frequent fire from 10 to 9 ka. The forest at this time was similar to low-elevation forests on the east side of the peninsula. Occurrence of bracken fern and mistletoe pollen (likely on western hemlock) is consistent with open forests that are conducive to the spread and growth of these species. The return to mesic forest species began at 9.5 ka with a distinct increase in western hemlock, followed by the return of abundant Pacific silver fir at 8.6 ka and the decline of Douglas-fir after 8 ka. At this time, the "analog" forest type also moved from Western Hemlock Zone sites to the vegetation currently at Yahoo Lake (Figure 27).

The two low elevation sites show a similar pattern to that at Yahoo Lake. At Crocker Lake the dominance by Douglas-fir and red alder is more pronounced than on the west side. The important mesic-site species, western redcedar, increases in abundance at 8 ka. Charcoal was surprisingly not more abundant at Crocker Lake during the Early Holocene compared to the Late Holocene. The large size of Crocker Lake might affect transport of macroscopic charcoal to the site. At Wentworth Lake, red alder and Douglas-fir also increase but Douglas-fir is a minor component and Sitka spruce was able to remain abundant. Charcoal was more abundant than later in the Holocene. When fires occurred, red alder and Sitka spruce may have been the most successful at establishment. This resulted in a vegetation type with no analog to modern forest types on the peninsula (Figure 32). Underscoring this no-analog condition is the surprising finding of a subalpine fir needle at 10.5 ka during the period of high red alder. To the authors' knowledge, the nearest location of subalpine fir to Wentworth Lake is in subalpine forest on a dry south-facing slope near Lake Crescent 57 m to the east. Although fir pollen was never abundant at this site, we must acknowledge the possibility that subalpine fir established during the Late Glacial on very well-drained glacial moraines near the lake and survived for several thousand years into the warm and dry Early Holocene. While a combination of Sitka spruce and subalpine fir is nonsensical from the perspective of modern vegetation gradients on the Olympic Peninsula, these species do co-mingle in fjordlands of southeast Alaska today and they were also detected together in the Late Glacial at Crocker Lake (see above). The alternative explanation, that subalpine fir needles were dispersed westward, is very improbable, though Pisaric (2002) found that convection from a forest fire in Montana transported abundant subalpine fir needles up to 20 km.

In the subalpine zone, the vegetation type during the Early Holocene is unexpected based regional vegetation responses at that time. The warm summer temperature at this time, up to 4°C warmer, would suggest that tree line should have been higher than present and forest density should be greater in the parkland vegetation zone, as has been shown elsewhere in the Pacific Northwest (Rochefort et al. 1994). At Martins Lake, the sediment remained inorganic, the charcoal levels remained very low, the pollen was dominated by regionally dispersed Sitka alder, and only three small macrofossils of subalpine fir needles were found from 10.5 to 7.6 ka. The lack of a higher treeline at Martins Lake during the Early Holocene presents a conundrum. Increased temperature is expected to alleviate constraints on tree growth in mountain hemlock forests (Zolbrod and Peterson 1999, Halofsky et al. 2011). (Gavin et al. 2001) speculated that snow avalanching during the Early Holocene was more common than today, due to the colder winter temperatures that may have occurred at the time. Frequent disturbances from snow avalanches would have favored slide alder thickets over conifer trees. At Moose Lake there is a similar dominance by slide alder pollen, but the first tree macrofossil (Engelmann spruce) occurs at the start of the Holocene and subalpine fir occurs locally throughout the Early Holocene. In contrast, trees did not become common at Martins Lake until directly after the Mazama tephra deposition, when subalpine fir (currently absent from the area), mountain hemlock and Alaska yellow cedar established. One possibility is that the Mazama ash, though only 2 cm in thickness, was sufficient to fill voids in talus slopes and thus provide a mulch for conifer seedlings to survive the summer drought (Gavin et al. 2001). In line with this hypothesis is that the most drought-adapted conifer, subalpine fir, was the only conifer present before the Mazama ash and was the one species that dominated for the period directly after the ash deposition but eventually was replaced by mesic-adapted Pacific silver fir. At Moose Lake, the ash deposition did not have as great an effect perhaps because Moose Lake is located on a valley floor that may receive fine-grained material suitable for seedling germination from upslope areas, while Martins Lake is on a ridge. Alaska yellow cedar increases in abundance at Moose Lake at the time of the Mazama tephra; the climatic cause of this increase is not clear.

In summary, the distinct Early Holocene forests at low elevation, dominated by western hemlock and red alder, are similar to those described by earlier studies in the Puget Lowland (Whitlock 1992). The one site where a high-resolution macrofossil record was possible (Yahoo Lake) showed that turnover of long-lived trees was often lagged behind climate change and required major disturbances (Gavin et al. in review), a prediction emerging from theories and models of forest dynamics (Davis 1986, Kitzberger et al. 2012). The vegetation changes at Yahoo Lake in the Early Holocene, such as upslope movement of Sitka spruce, are supported by another detailed macrofossil record near Vancouver BC (Wainman and Mathewes 1987). The vegetation changes in the subalpine, however, were not predicted by expectations of increased treeline under warmer summer conditions. The controls on treeline in the Olympic Mountains today are complex and vary greatly over space (Woodward et al. 1995, Peterson 1998). Without more paleoecological records, the pattern of subalpine forest development, and the role of soils and climate on forest colonization of the subalpine landscape, remains unresolved.

5.4.6 Middle and Late Holocene: 6–0 ka

During this time period, cooler and moister summers resulted in lower fire occurrence and the establishment of the dense, deep-shade tolerant vegetation currently typical in western Washington. At all sites, modern vegetation types were achieved by 6 ka, and in general forest composition changed only small amounts thereafter (Figure 32). The period around 6 ka marks the time to when most forests began to resemble today's old growth Douglas-fir forests (Brubaker and McLachlan 1996). This change is most clearly seen at Yahoo Lake where there were very long fire intervals (>500 yr) and increased late-successional species (hemlock, Pacific silver fir, and redcedar). Paleoclimate records indicate climatic fluctuations mainly in the form of neoglacial advances (Figure 22). While not extensively studied, neoglacial advances on the Olympic Peninsula likely correlate well with other sites in the Pacific Northwest (Long 1975, Hellwig 2010). However, these climate fluctuations were not sufficiently large enough to affect broad-scale vegetation patterns. The long lifespans of the late-successional forest trees buffered forest compositional change to the decadal (or longer) climate fluctuations of the Late Holocene. One exception is a late-Holocene increase in fire at 2.5 ka at Yahoo and Wentworth lakes, which is seen in the Cascades as well (Gavin et al. 2007, Prichard et al. 2009). Whether this fire-increase is the result of increased climatic variability (favoring intermittent fire-susceptible conditions) or increased burning by humans (for example, for berries and hunting habitat) has been widely speculated and will be further discussed below (Hallett et al. 2003, Lepofsky et al. 2005).

Other vegetation changes occurring during the Late Holocene may represent progressive autogenic ecosystem development rather than direct response to climate fluctuations. The clearest expression of this change is at Wentworth Lake where western redcedar increases progressively through the Late Holocene (Figure 26), similar to that found at Whyac Lake on western Vancouver Island (Brown and Hebda 2002) and near Vancouver (Wainman and Mathewes 1987). This likely represents development of redcedar-dominated muskeg forests on the coastal strip due to accumulation of organic matter and increased soil acidity that favors redcedar along with increasing moisture. A similar pattern of retrogressive ecosystem development has been found in coastal forests in southeast Alaska (Hansen and Engstrom 1996). Another pollen record near Crocker Lake, taken from a wetland site, revealed a sequence of hydrosere succession controlled by infilling of the basin with sediment (McLachlan and Brubaker 1995).

While regional pollen records from lakes do not reveal major vegetation change during the Late Holocene, pollen records from smaller basins and soils reveal more dynamic vegetation at smaller spatial scales. Pollen records from "small hollows" (small bedrock or gravel depressions only ca. 5 m in diameter) record vegetation at the stand scale, or within about 50 m. Similarly, meadow soils will record pollen from primarily insect-pollinated species in much greater proportion than they occur in small lakes. The following three studies of small hollows and soils were conducted within or near the park.

Three small hollow sites were studied on terraces of the Queets River (Greenwald and Brubaker 2001). The two sites within 800 m of the river date back only 500 yr when side channels were abandoned, or a major flood affected the site. These sites show in general a succession from red alder to Sitka spruce and western hemlock. The third site, adjacent to the valley wall recorded consistently high abundance of red alder over the past 5000 years suggesting a long-term presence of riparian forests with frequent flooding disturbance. The charcoal record indicated only two fires over the past 5000 years. An increase in hemlock over the last 1000 years may reflect cooling during the Little Ice Age (Greenwald and Brubaker 2001).

A study of small hollows on Orcas Island, in the rainshadow of the Olympic Mountains, revealed histories from 7 ka to present (Sugimura et al. 2008). They showed that the rocky outcrop of Mount Constitution supported lodgepole pine forests for most of the Late Holocene, likely due to the poor soil conditions supporting a positive feedback between vegetation and fire. However, a period from 5.3 to 2 ka supported reduced fire and less pine at sites that had slightly more soil and thus ability to hold moisture during wetter periods (Sugimura et al. 2008).

Subalpine meadows contain high plant diversity within small regions, but these meadows are “invisible” to any pollen records from lake sites. Gavin and Brubaker (1999) studied a 6000-year record from three soil profiles from a ridge top at 1770 m elevation in Royal Basin in the northeastern portion of the park. While the soil sites were more difficult to study due to poor pollen preservation and low temporal resolution, the study did reveal that late-snowmelt meadow communities dominated by *Carex nigricans* was the most sensitive to climate change. That site type apparently switched between the monodominant sedge community and more diverse mesic meadow types at certain times in the past. Other sites revealed a medieval warm period dominated by American bistort (Gavin and Brubaker 1999).

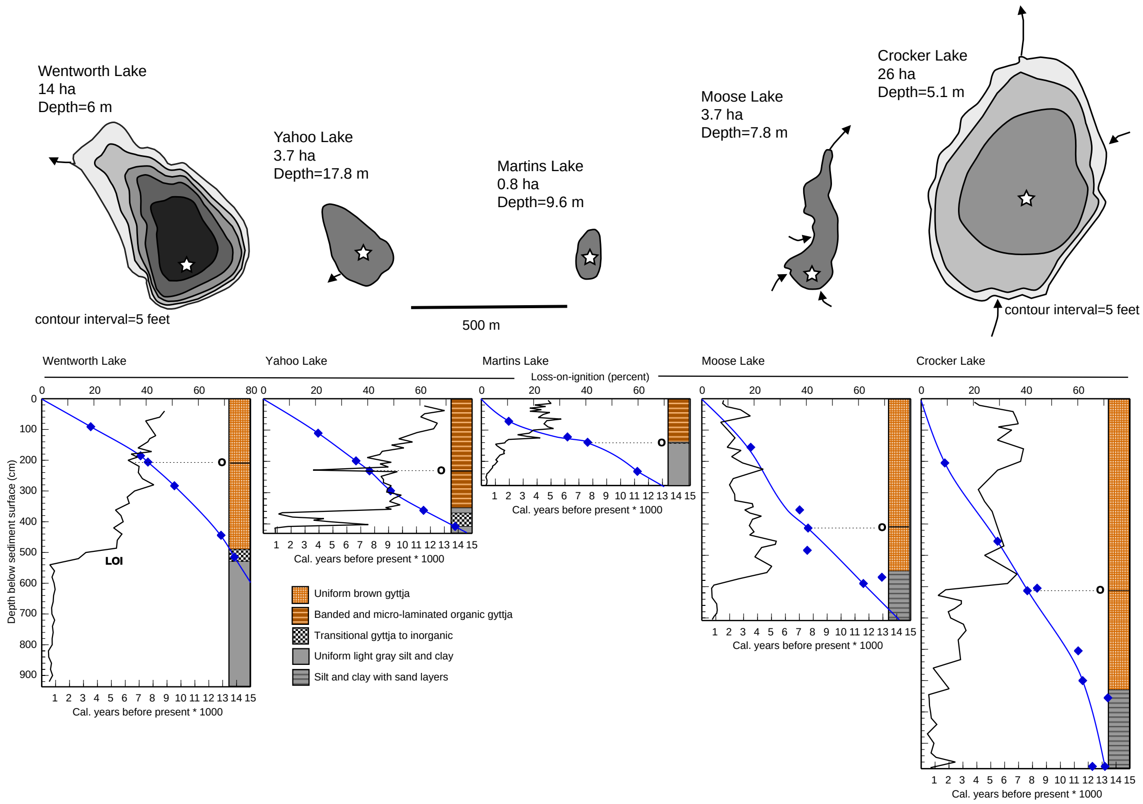


Figure 25. Maps, age-depth relationships, and stratigraphy of sediment cores from five lakes on the Olympic Peninsula. Bathymetric maps shown for two lakes where bathymetric data were available. Blue diamonds indicate radiocarbon dates and the black line is loss-on-ignition at 550°C, a measure of organic matter in lake sediment.

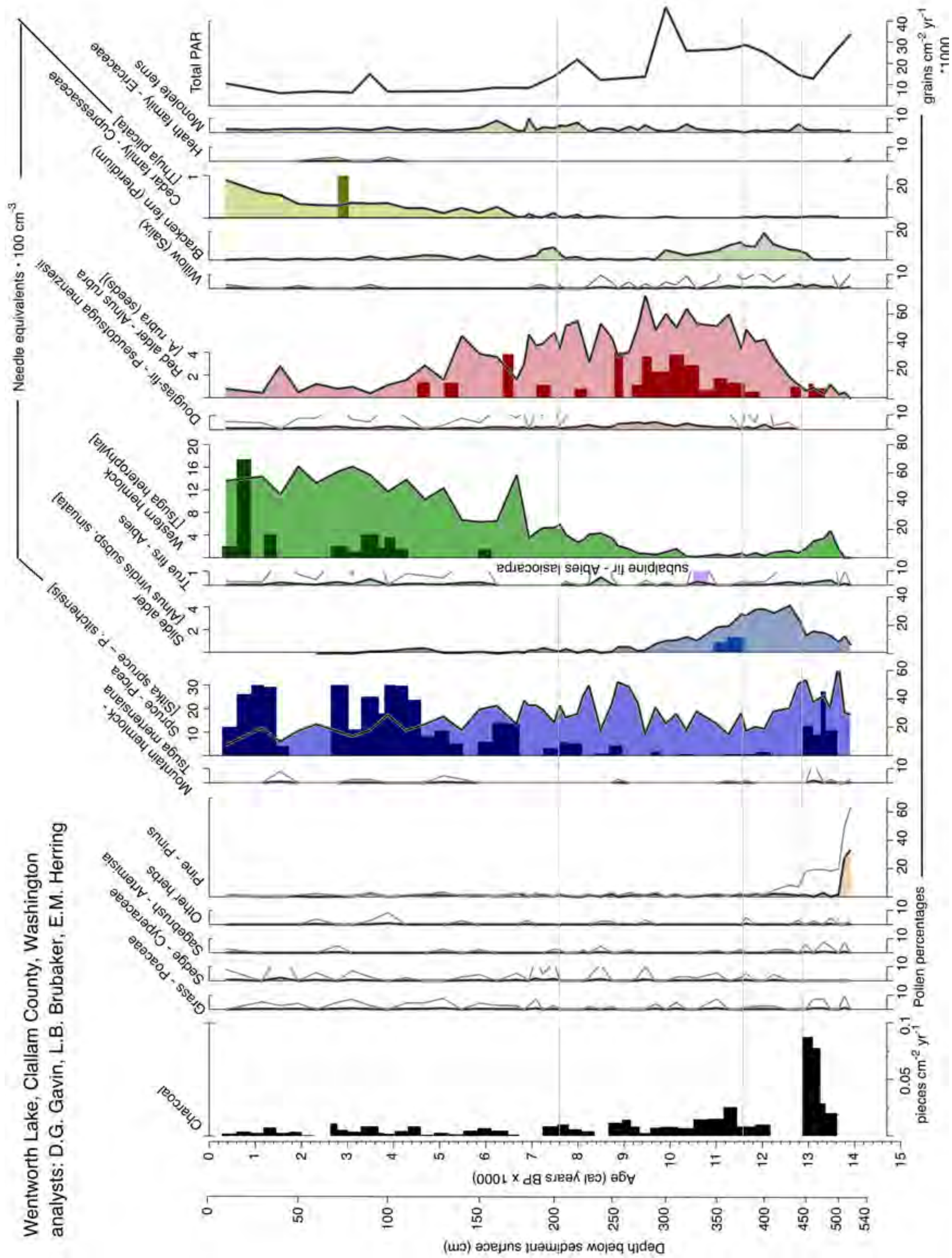


Figure 26. Pollen and macrofossil diagram for Wentworth Lake. Each pollen taxon is labeled above each graph in both common and scientific names. Taxon names in brackets refer to the macrofossil data, graphed as bar charts superimposed over the filled line plot for pollen percentages. Horizontal lines indicate the ages of the Mazama tephra (7.6 ka), and the start and end of the Younger Dryas (12.9–11.6 ka). Gray lines on the pollen graphs indicate 10X exaggeration for rare pollen taxa. PAR=pollen accumulation rate.

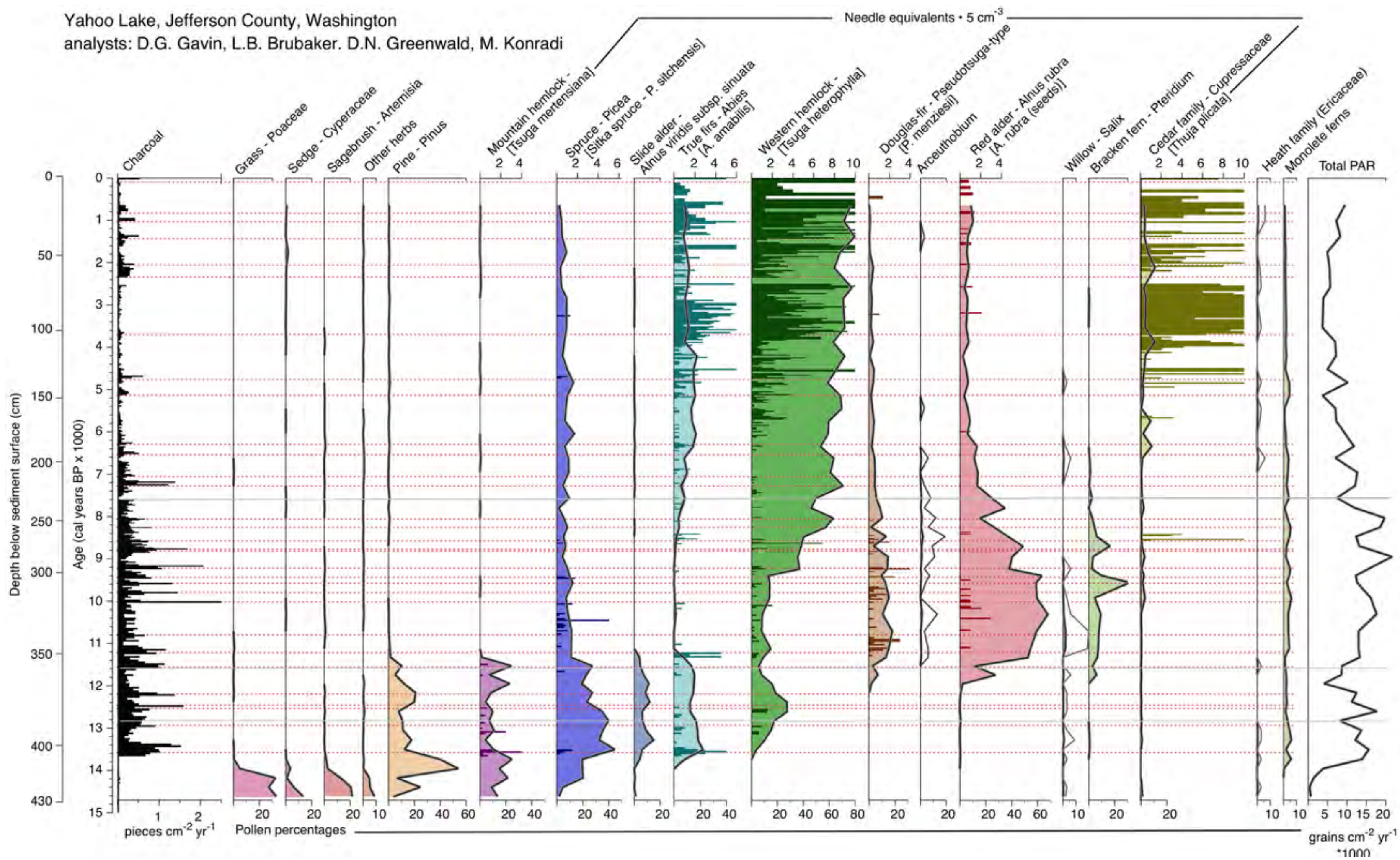


Figure 27. Pollen and macrofossil diagram for Yahoo Lake (Gavin et al. in review). Diagram is constructed as for Wentworth Lake. Identified fire events from the 1-cm charcoal stratigraphy are shown by dashed red lines. Methods for identifying fire events are described in (Higuera et al. 2010).

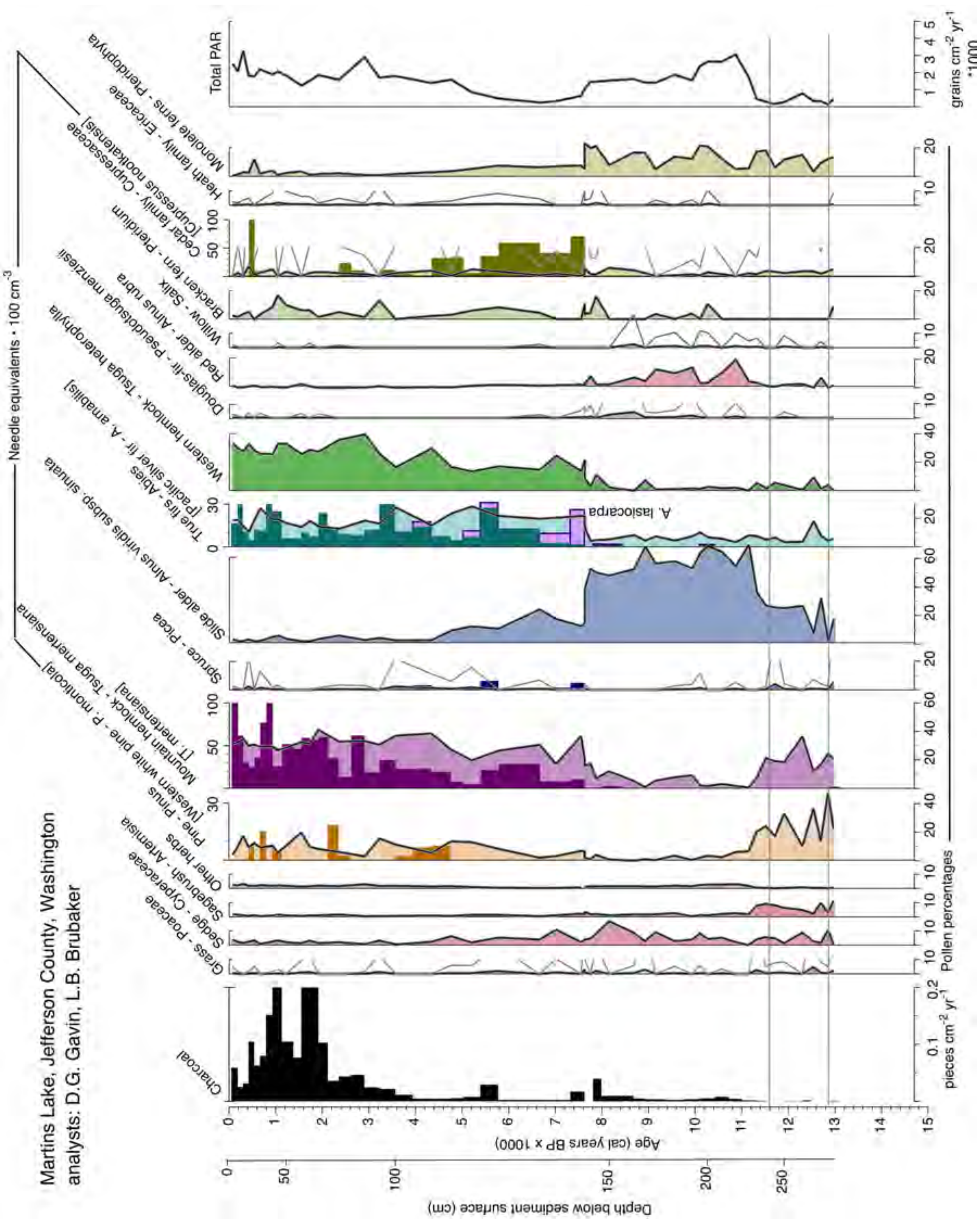


Figure 28. Pollen and macrofossil diagram for Martins Lake (Gavin et al. 2001). Diagram is constructed as for Wentworth Lake.

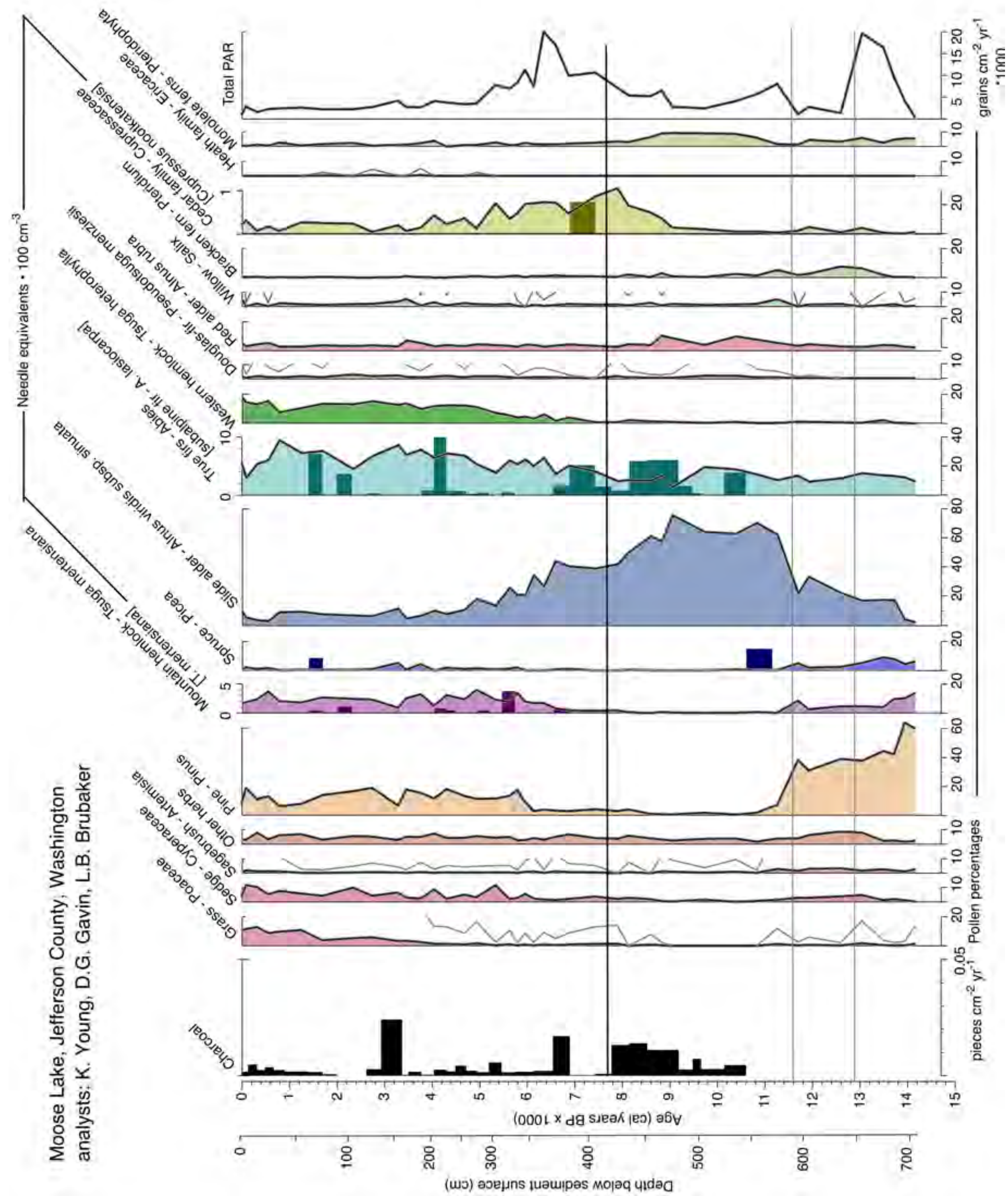


Figure 29. Pollen and macrofossil diagram for Moose Lake (Gavin et al. 2001). Diagram is constructed as for Wentworth Lake.

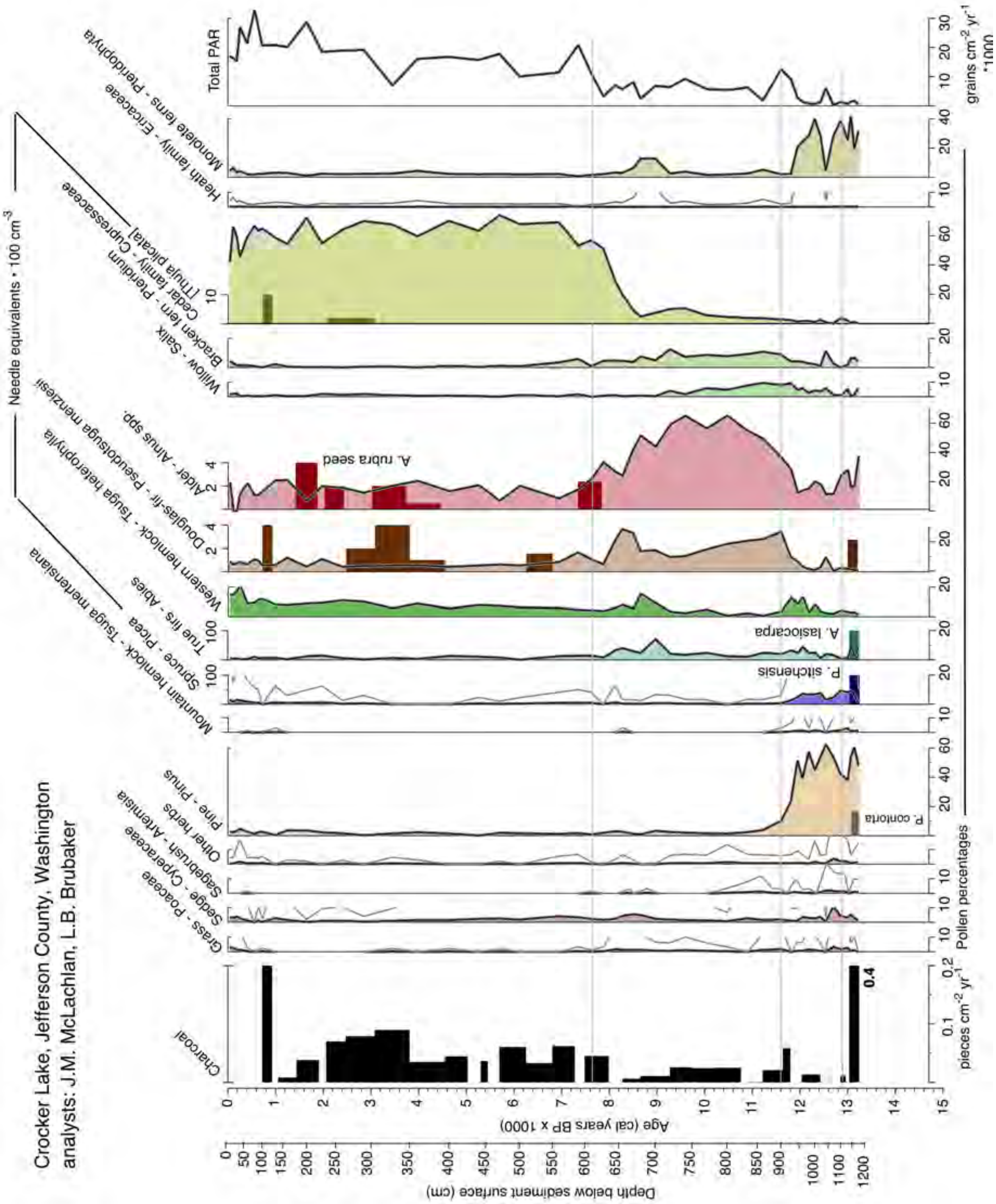


Figure 30. Pollen and macrofossil diagram for Crocker Lake (McLachlan and Brubaker 1995). Diagram is constructed as for Wentworth Lake.

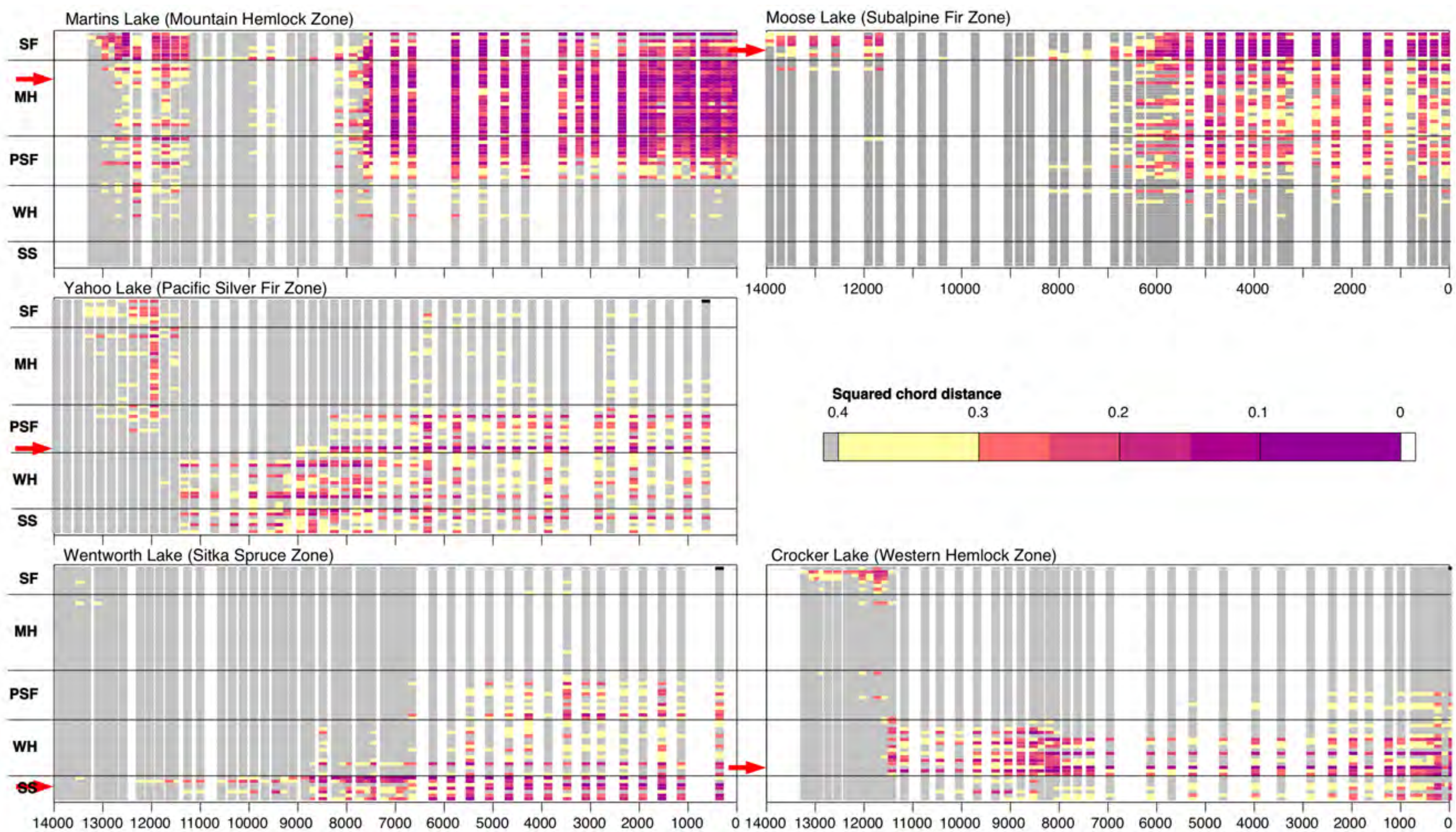


Figure 31. Modern analog analysis for five pollen records on the Olympic Peninsula. For each lake, each column indicates the value of a dissimilarity coefficient (SCD) between the 65 modern pollen assemblages (arrayed by vegetation zone) with a fossil pollen assemblage.

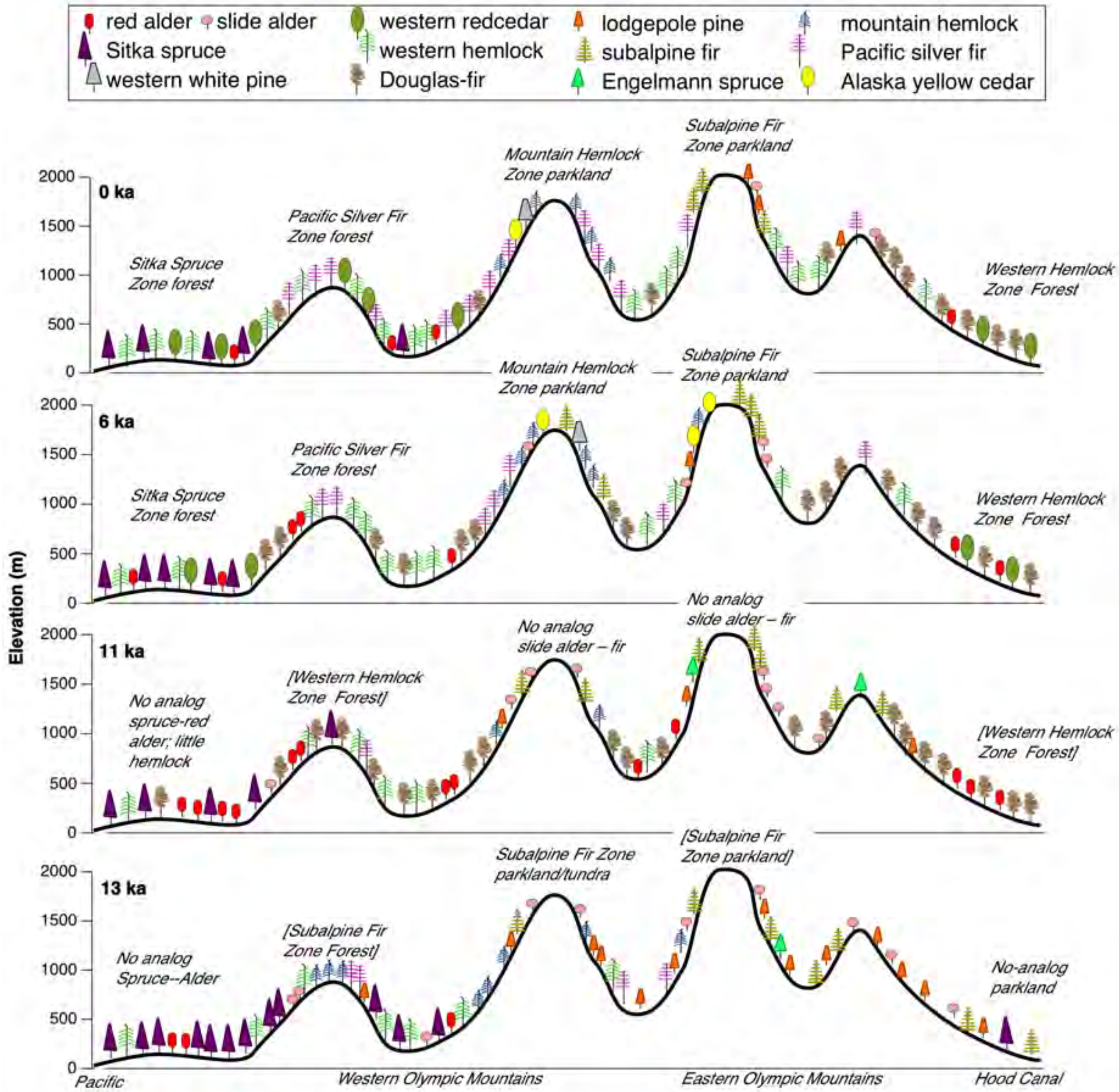


Figure 32. Schematic diagram showing a profile of dominant tree species on a longitudinal transect across the Olympic Peninsula for four time periods. The analog assignments are given above the location of each pollen record along the transect (West to east: Wentworth, Yahoo, Martins, Moose and Crocker lakes). Brackets indicate only marginal analogs were found.

Table 3. Late Quaternary pollen records and other sites mentioned in the text. Site numbers refer to Figure 33. Asterisks indicate sites used for mapping pollen abundances.

Site	Name	Latitude	Longitude	Elev. (m)	Lake size (ha)	References
1	Frozen Lake	49° 36'N	121° 28'W	1180	3	(Hallett et al. 2003)
2	Pinecrest Lake*	49° 30'N	121° 26'W	320	1.8	(Mathewes et al. 1972, Mathewes and Rouse 1975)
3	Squeah Lake	49° 29'N	121° 24'W	250	2.2	(Mathewes and Rouse 1975)
4	Marion Lake*	49° 19'N	122° 33'W	305	16	(Mathewes 1973, Mathewes and Heusser 1981)
5	Mike Lake*	49° 16'N	122° 32'W	225	4.8	(Pellatt et al. 2002)
6	Mt. Barr Cirque	49° 16'N	121° 31'W	1376	2	(Hallett et al. 2003)
7	Clayoquot Lake	49° 12'N	125° 30'W	17	47	Gavin et al. 2003
8	Porphyry Lake*	48° 54'N	123° 50'W	1100	0.3	(Brown and Hebda 2003)
9	Mosquito Lake Bog	48° 46'N	122° 07'W	198	–	(Hansen and Easterbrook 1974)
10	Whyac Lake*	48° 40'N	124° 51'W	10	2.5	(Brown and Hebda 2002, 2003)
11	Panther Potholes*	48° 40'N	121° 02'W	1100	0.3	(Prichard et al. 2009)
12	Mt Constitution	48° 39'N	122° 50'W	660	–	(Sugimura et al. 2008)
13	Ayer Pond	48° 37'N	122° 49'W	50	–	(Kenady et al. 2011)
14	Pixie Lake*	48° 36'N	124° 12'W	70	5.3	(Brown and Hebda 2002, 2003)
15	Walker Lake*	48° 32'N	124° 00'W	950	0.4	(Brown and Hebda 2003)
16	East Sooke Fen*	48° 21'N	123° 41'W	155	0.1	(Brown and Hebda 2002)
17	Kirk Lake*	48° 15'N	121° 37'W	190	0.6	(Cwynar 1987)
18	Ebey's Prairie	48° 12'N	122° 42'W	25	–	(Weiser and Lepofsky 2009)
19	Ahlstrom's Prairie	48° 10'N	124° 43'W	60	–	(Anderson 2009)
20	Manis Mastadon Site	48° 03'N	123° 07'W	170	–	(Petersen et al. 1983, Waters et al. 2011)
21	Wentworth Lake*	48° 01'N	124° 32'W	47	14	This study
22	Crocker Lake*	47° 56'N	122° 53'W	54	26	(McLachlan and Brubaker 1995)
23	Cedar Swamp*	47° 54'N	122° 53'W	57	–	McLachlan and Brubaker 1995
24	Moose Lake*	47° 53'N	123° 21'W	1544	3.7	(Brubaker and McLachlan 1996, Gavin et al. 2001)
25	Royal Basin	47° 49'N	123° 13'W	1770	–	(Gavin and Brubaker 1999)
26	Hall Lake*	47° 48'N	122° 19'W	104	3.1	(Sugita and Tsukada 1982)
27	Martins Lake*	47° 43'N	123° 32'W	1423	0.8	(Gavin et al. 2001)
28	Yahoo Lake*	47° 41'N	124° 01'W	717	3.7	Gavin et al. in review
29	Queets Hollows	47° 38'N	123° 58'W	120	–	(Greenwald and Brubaker 2001)
30	Humptulips Bog	47° 17'N	123° 55'W	120	–	(Heusser et al. 1999)
31	Nisqually Lake*	47° 02'N	122° 38'W	65	8	(Hibbert 1979)
32	Reflection Pond 1	46° 46'N	121° 43'W	1482	0.6	(Dunwiddie 1986)
33	Mineral Lake*	46° 44'N	122° 10'W	436	103	(Sugita and Tsukada 1982)
34	Davis Lake*	46° 32'N	122° 15'W	290	–	(Barnosky 1981)

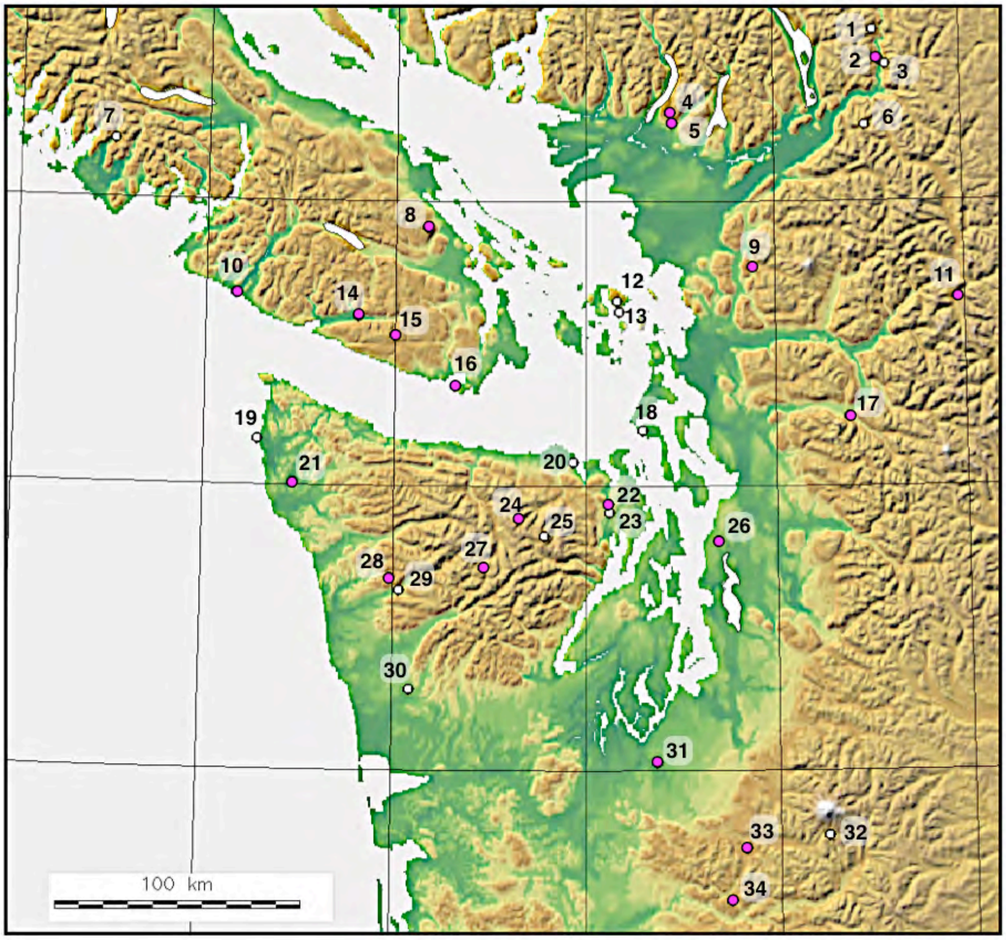


Figure 33. Map of fossil record localities discussed in the text. Red sites are radiocarbon-dates pollen records that are mapped in the regional vegetation syntheses.

5.5 Regional synthesis of vegetation changes in the Pacific Northwest

Single pollen diagrams, as described above, provide a wealth of information regarding the progression of vegetation and climate changes at a site. However, the spatial patterns of change require synthesizing the information from many pollen diagrams. To do this, we compiled data from 20 pollen records from western Washington and southwest British Columbia (Figure 33), updating an earlier effort (Tattersall 1999). We re-analyzed the data from these sites using a multistep method to ensure consistency in the interpretation among sites (see Appendix A for detailed methods). We then produced maps of pollen abundance for the major tree species at roughly 1000-year time slices from 17,000 years ago (the time of the maximum Cordilleran ice extent) to present. Limitations in the accuracy of the radiocarbon chronology at many sites preclude a detailed interpretation of between-site differences. Furthermore, because sites occur across a range of elevations and microsite types, sites nearby each other may have very different pollen percentages. However, the maps do reveal regional patterns of vegetation change, with no two species (or species groups) showing the same pattern. Below we provide both species-level interpretation and regional vegetation interpretations.

5.5.1 Species dominant during the Late-Glacial: pine, spruce, mountain hemlock, and poplar

Pine pollen is largely comprised of lodgepole pine during the Late Glacial. The very high abundance of pine at sites directly adjacent to the receding ice sheet suggests that is was able to rapidly colonize on glacial outwash and/or till soils. It declines, however, at 11.5 ka when the climate warms and it must compete against a broad diversity of species that are able to grow to larger sizes than lodgepole. It likely remained abundant throught the Early Holocene on rocky outcrops and south facing slopes, as seen at Panther Potholes in the North Cascades (Prichard et al. 2009). The generally low abundance of pine pollen through the remainder of the Holocene throughout the region is likely a mix of both lodgepole pollen on dry sites, shore pine from poor soils on the coastal plain, or western white pine from montane sites.

Spruce pollen is also abundant during the Late Glacial, but this pollen type mainly represents Sitka spruce with some unknown level of Engelmann spruce (the pollen is indistinguishable at the species level). On the coast, Sitka spruce dominated through the Late Glacial, as evidenced by macrofossils at Yahoo Lake, and remains abundant until present. As Sitka spruce occurs today close to periglacial environments in southeast and south-central Alaska, it is not surprising that it could be common shortly following deglaciation along the coast. Further inland, a high spruce period at Mineral Lake from 16 to 14 ka was assumed to be Engelmann spruce, as this species is common in subalpine parkland settings as assumed to be the case during the Late Glacial (Tsukada and Sugita 1982). However, further north at Crocker Lake and Kirk Lake, at 13 ka macrofossils indicate Sitka spruce was present in a parkland setting (Cwynar 1987, McLachlan and Brubaker 1995). A combination of genetic and fossil evidence indicates Sitka spruce migrated northward to Haida Gwaii as early as 13.4 ka (Lacourse et al. 2005, Holliday et al. 2010). We suspect that Sitka spruce generally had a broader distribution during the Late Glacial.

Mountain hemlock is also an early colonizer in the deglacial landscape. It increases during the increase in temperature between 15 and 14 ka on the Olympic Peninsula and on Vancouver Island. At 11.5 ka it moves upslope to the its modern suablpine elevational zone, with the exception of the case of Martins Lake (see above) where soil development may have slowed its upslope progression. A general increase of mountain hemlock during the Late Holocene is consistent with neoglacial cooling.

Poplar pollen was common during the Late Glacial in the Puget Trough, especially in the foothills of the western Cascades; on the Olympic Peninsula it was present at lower levels (<5 % pollen). It is likely that this pollen represents black cottonwood in riparian forests where glacial outwash sediments provided extensive habitat. Its abundance declined into the Holocene as this habitat succeeded to less disturbed forests with organic soils.

Environmental interpretation of the Late Glacial conditions remains controversial because of the difficulty in interpreting the pollen data. Most studies have interpretted the Puget Lowland to support open parkland that was similar to northern Rocky Mountain subalpine fir-Engelmann spruce forests (Barnosky 1984, Whitlock 1992). However, we note that the Engelmann spruce macrofossils from the Puget Lowland generally date to older than 16 ka (e.g., at Mineral Lake and Port Moody BC; Tsukada et al. 1981, Hicock et al. 1982). After 15 ka the climate became distinctly more maritime with Sitka spruce sometimes co-occuring with subalpine fir such as at Crocker Lake at 13 ka. W e suggest that the best analog for the postglacial sequence begins with continental conditions that favored subalpine parkland. This transitioned at 15 to 14 ka to a maritime forest similar to that in coastal southeast Alaska where Sitka spruce co-occurs with mountain hemlock and in a few places with subalpine fir (Figure 34). The transition from continental to maritime climate remains poorly constrained. An intriguing possibility is that the two spruce species hybridized, which is possible with modern populations (Kiss 1989). Sitka spruce also hybridizes with white spruce, highly related to Engelmann spruce, in northeastern British Columbia. Such hybridization would contribute to the difficulty of climatic interpretation and to the difficulty in distinguishing the species in the fossil record.

Figure 34. Two perspectives on the Late Glacial environment of the Puget Lowlands. Top: Cool maritime forests of Sitka spruce, cottonwood, mountain hemlock, and subalpine fir that may be similar to the Late Glacial from 15 to 12 ka. (location: Salmon River near Hyder, AK). Bottom: Continental tree-line forests of the interior. Engelmann spruce and subalpine fir with lodgepole pine dominate that may be typical of LGM and possibly Vashon Stade periods from 21 to 17 ka. (location: Misty Peak, Salmo-Priest Wilderness of northeast Washington). Top photo: http://www.panoramio.com/photo/25281602; bottom photo: http://www.panoramio.com/photo/54378334.

5.5.2 Species dominant during the Early Holocene: Douglas-fir, oak, alder, and maple

Douglas-fir is one of the few species that was likely absent from western Washington until the last stages of the Late Glacial. The possibility that this pollen represents western larch is remote as that species is found only east of the Cascade Range. However, it is not until the Early Holocene (11.5 ka) when it dramatically increases in abundance and macrofossils are found. Its pollen is not identified at a well-dated location prior to 13.5 ka. At 13 ka it is present at low levels (<5%) in the Puget Lowland and southern Vancouver Island, though the radiocarbon chronology is poor at the sites that indicate its presence at this time. As currently mapped, the evidence supports a rapid northward migration at low abundance followed by a rapid inrease in abundance upon warming. However, with better radiocarbon chronologies it may indicate a more uniform migration occuring closer to 11.5 ka, perhaps with a migration lag to Vancouver Island. Its abundance declines in wetter areas by 8 ka, but it remains fairly abundant in the Puget Lowland and eastern Vancouver Island into the Late Holocene.

Oak pollen, representing garry oak, first appears at 13 ka in the southern Puget Lowland but becomes more common from 10 to 6 ka. As for Douglas-fir, the warm and dry conditions of the Early Holocene, and the frequent fires, allowed expansion of the oak prairies in the Puget Lowland (Leopold and Boyd 1999).

Alder pollen represents two common species: red alder and Sitka alder. Both produce abundant pollen. Because too few studies distinguished the pollen, the mapped values show total alder percentages. The period of regionally highest abundance, from 12 to 8 ka, coincides with the period of high Douglas-fir.

Maple is poorly represented by pollen because maple species are primarily insect pollinated. Bigleaf maple is likely the largest contributor of pollen. Although the pollen almost never exceeds 2%, it is more common in the middle Holocene. There is little spatial patterning of maple pollen.

The interpretation of the Early Holocene from these mapped data is similar to that discussed above for the Olympic Peninsula sites. High levels of red alder and Douglas-fir pollen suggest widespread fire and a competitive post-fire environment between these two species. A possible analog is the Oregon Coast Range where alder is aggressive on hillslopes, while it is more relegated to riparian zones today on the western Olympic Peninsula. The pollen assemblages at this time are, in fact, somewhat similar to the pollen assemblages on the modern landscapes of the western peninsula dominated by clear-cut logging (Heusser 1969).

The spatial gradient of Douglas-fir pollen indicates that fire and lower precipitation allowed for a great expansion of Douglas-fir forests in the Puget Lowland, while a lack of this gradient for alder indicates that disturbances may have been occurring as frequently on the Pacific Coast but that alder may have been a more aggressive in the wetter climate. This interpretation of rainfall effects on the Douglas-fir distribution was used to reconstruct of the spatial gradient of precipitation on southern Vancouver Island using a ratio of Douglas-fir and western hemlock pollen (Brown et al. 2006).

5.5.3 Species dominant during the Late Holocene: true firs, western hemlock, and western redcedar

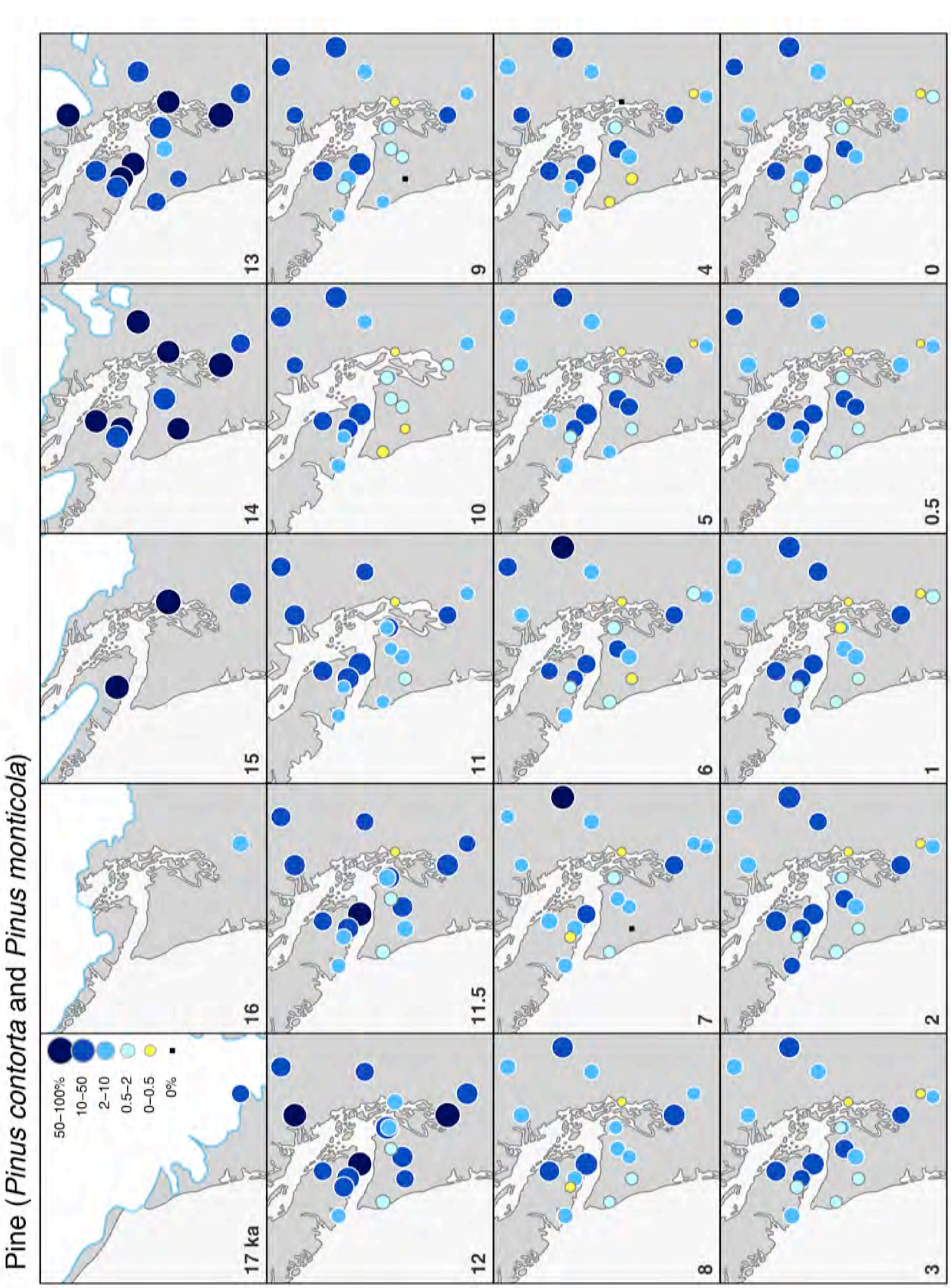
The true firs are represented by two common species in the region: subalpine fir and Pacific silver fir. Because these two species occur in quite different climatic zones, there is no strong spatial patterns in the true-fir pollen data. However, the pollen type is large and poorly dispersed, and so low abundances are more meaningful for this species than for others. The results show that Late Glacial presence in the Puget Lowland was likely subalpine fir but on the west coast it was Pacific silver fir, as shown by macrofossils (Gavin et al. in review, Cwynar 1987). The Early Holocene provided poor climate for Pacific silver fir, while subalpine fir retreated to rainshadowed high-elevation sites. The Late Holocene increase in true-fir pollen represents an expansion of Pacific silver fir in late-successional forests.

Western hemlock, the common late-successional tree species below the subalpine elevations, follows a similar history as Pacific silver fir, but because the pollen is so widespread its spatial pattern is less distinct. It is present by 14 ka mainly in the Puget Lowland and Vancouver Island. Before 13 ka it increases rapidly on the western Olympics but then declines to low levels (<5%) at 11.5 ka. It remains most abundant on the Pacific slope through the Early Holocene, but increases to over 10% in the Puget Lowland by 7 ka.

The cedar family pollen represents two tree species (western redcedar at low elevations and Alaska yellow cedar in the subalpine) and possibly a small tree Rocky Mountain juniper that is today restricted to the Sequim area. These species are, at most sites, poorly represented by pollen. This makes it particularly difficult to determine which species was important on the landscape during the long period of low pollen abundance from 15 to 11 ka. Macrofossils indicate that increases at 9 ka at Moose Lake is likely entirely Alaska yellow cedar. In contrast, the majority of sites at low elevation show a clear northward progression from 10 ka at Mineral Lake to 6 or 4 ka on Vancouver Island. This northward progression was noted by Hebda and Mathewes (1984) as important context for Native American pre-history, as large cedar logs were likely not available until the large increase in cedar pollen. Genetic studies of western redcedar suggest it occurred in a southern refugium south of Washington during the LGM (O'Connell et al. 2008).

The interpretation of the Late Holocene mapped pollen data is consistent with regionally uniform increase in available moisture and a decrease in fire, allowing extensive old growth forests to develop. While climatic fluctuations still occurred, they may have been of lower magnitude than the Early Holocene. Thus, with forests dominated by long-lived trees species, and little large-scale migration, changes in forest composition were relatively minor. However, higher resolution pollen records that are well dated should reveal regional patterns in changing forest composition during the Late Holocene. For example, elevational shifts in species establishment are easily inferred from forest structure over elevational transects, which are likely climate driven on century time-scales. Thus, "vegetational inertia" to climate change during the Late Holocene may be an important, but not overwhelming, factor slowing the response of forests to climate change.

Figure 35. (See following 11 pages). Pollen abundance mapped at 20 time slices for 11 common pollen types. The maps for present-day pollen abundances (0 ka) show in green the vegetation zone for the taxa corresponding to each pollen type, only if the pollen type corresponds to a species that corresponds with a distinct vegetation zone. Alder (*Alnus*), for example, is not mapped because red alder and Sitka alder together occur over the entire region. Ice extent and paleogeography are from (Dyke et al. 2003) Note that ice extent in these maps is conservative, and does not show advances of ice in southwest British Columbia between 13.7 and 13.5 ka.



Pine (Pinus contorta and Pinus monticola)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35A

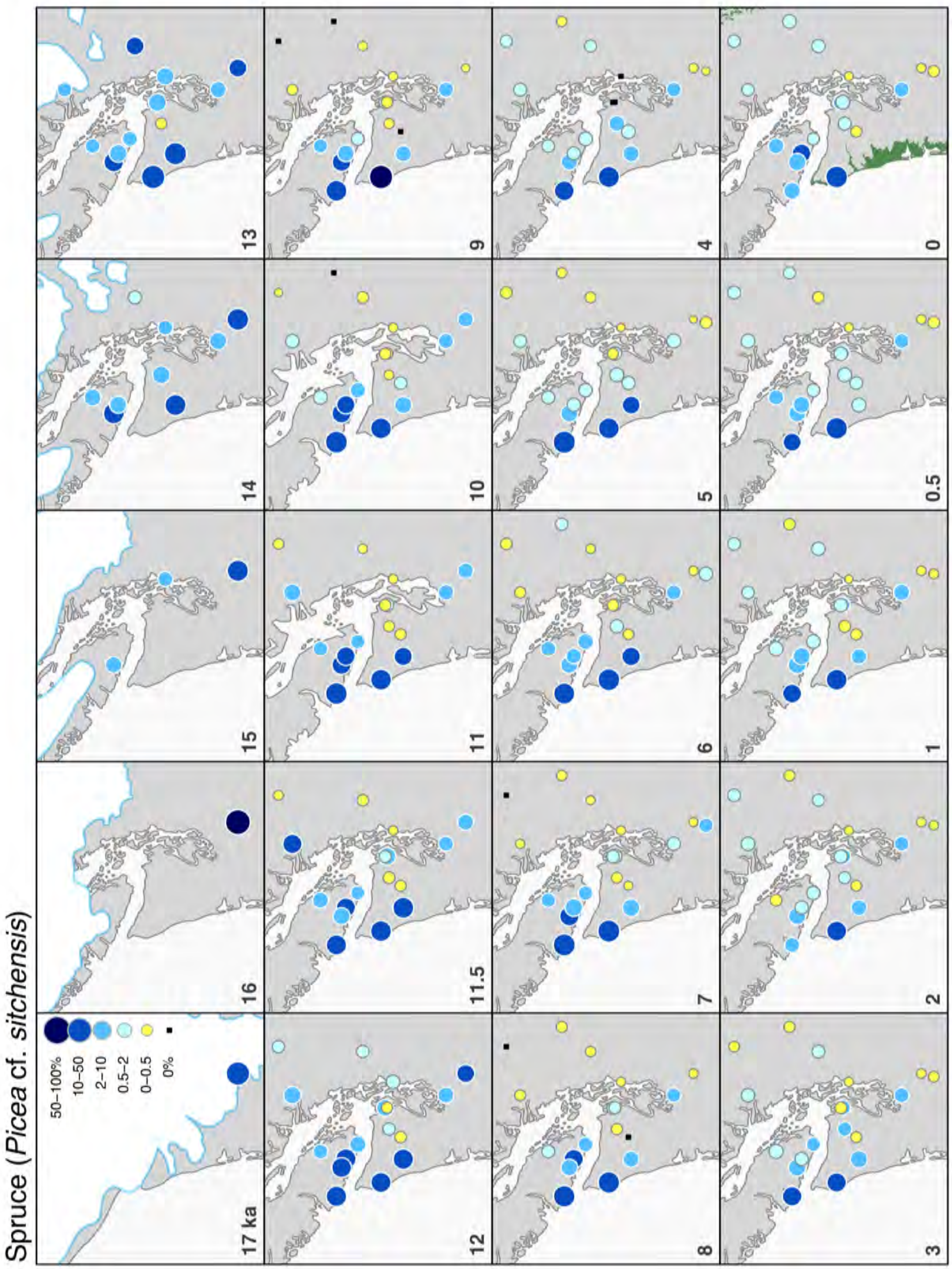
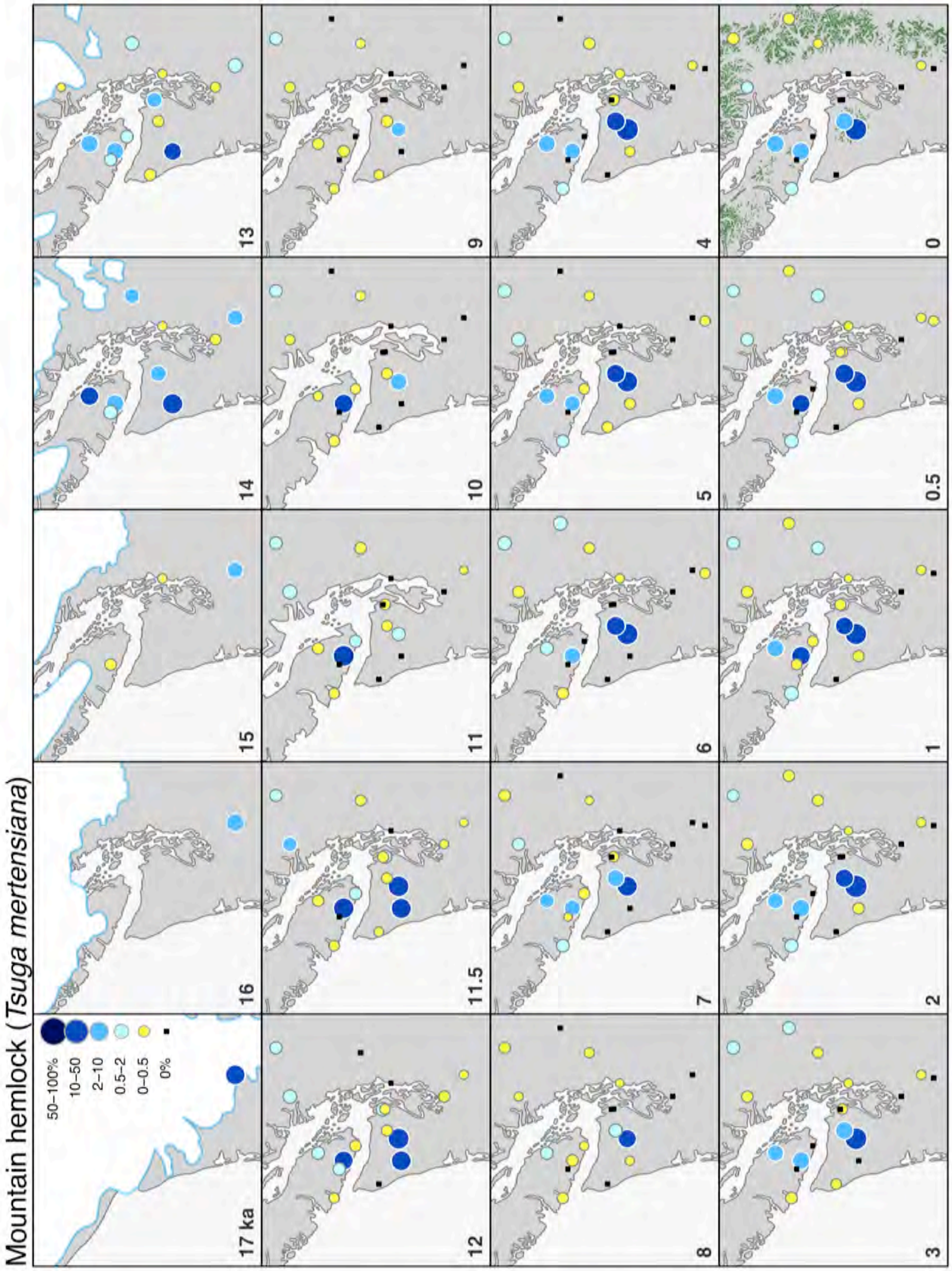

Spruce (Picea cf. sitchensis)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35B



Mountain hemlock (Tsuga mertensiana)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35C

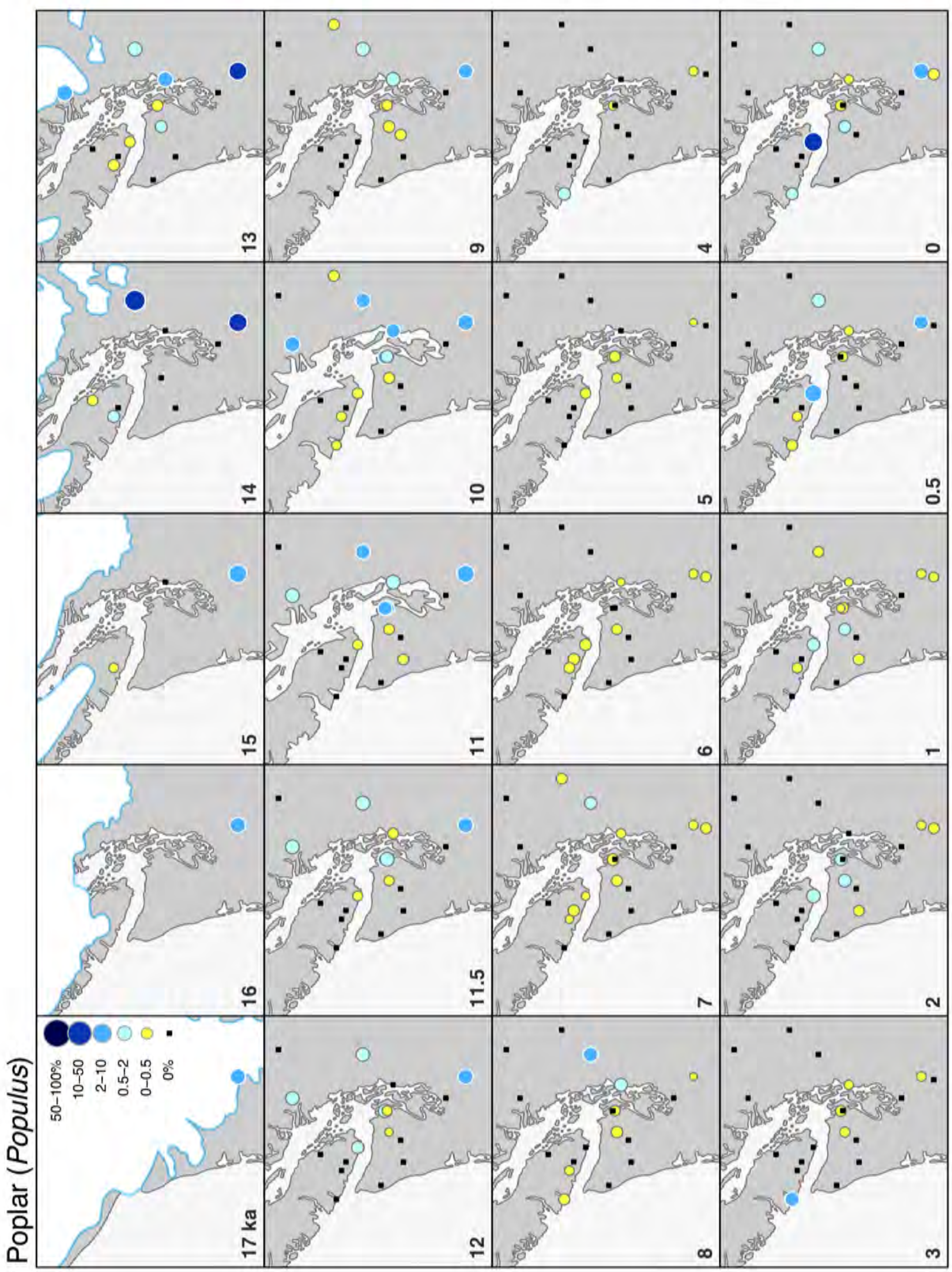

Poplar (Populus)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35D

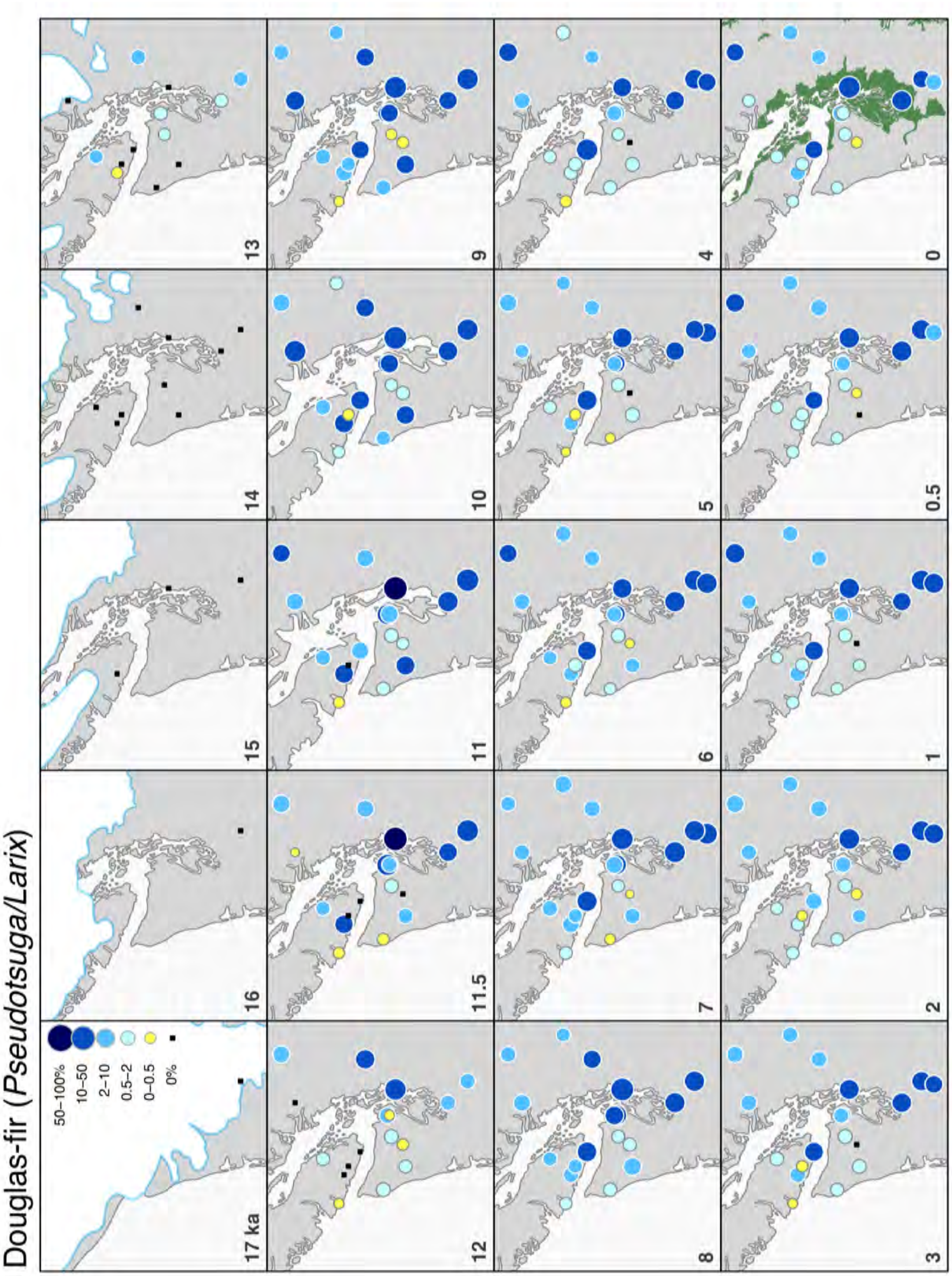

Douglas-fir (*Pseudotsuga/Larix*)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35E

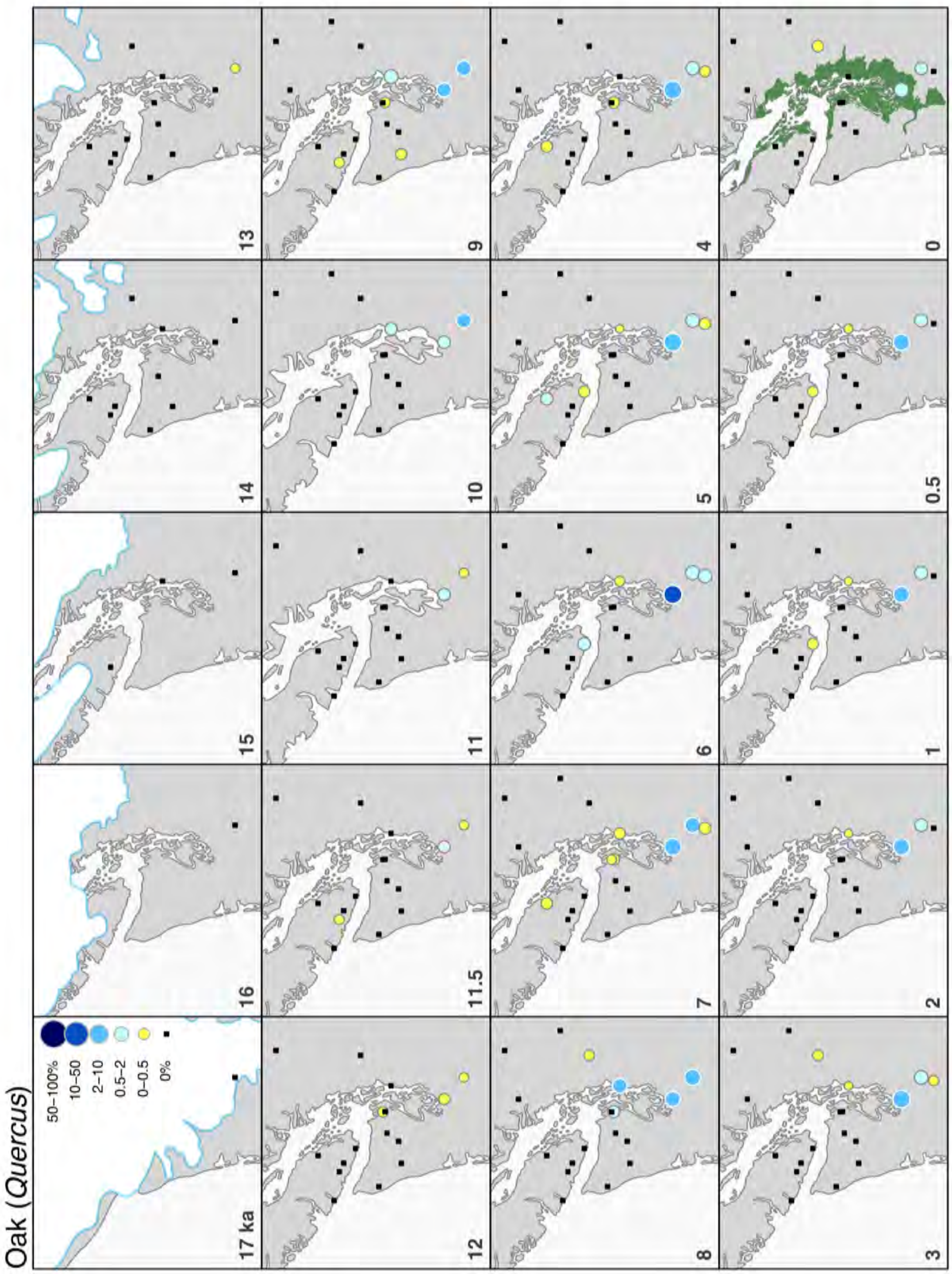

Oak (Quercus)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35F

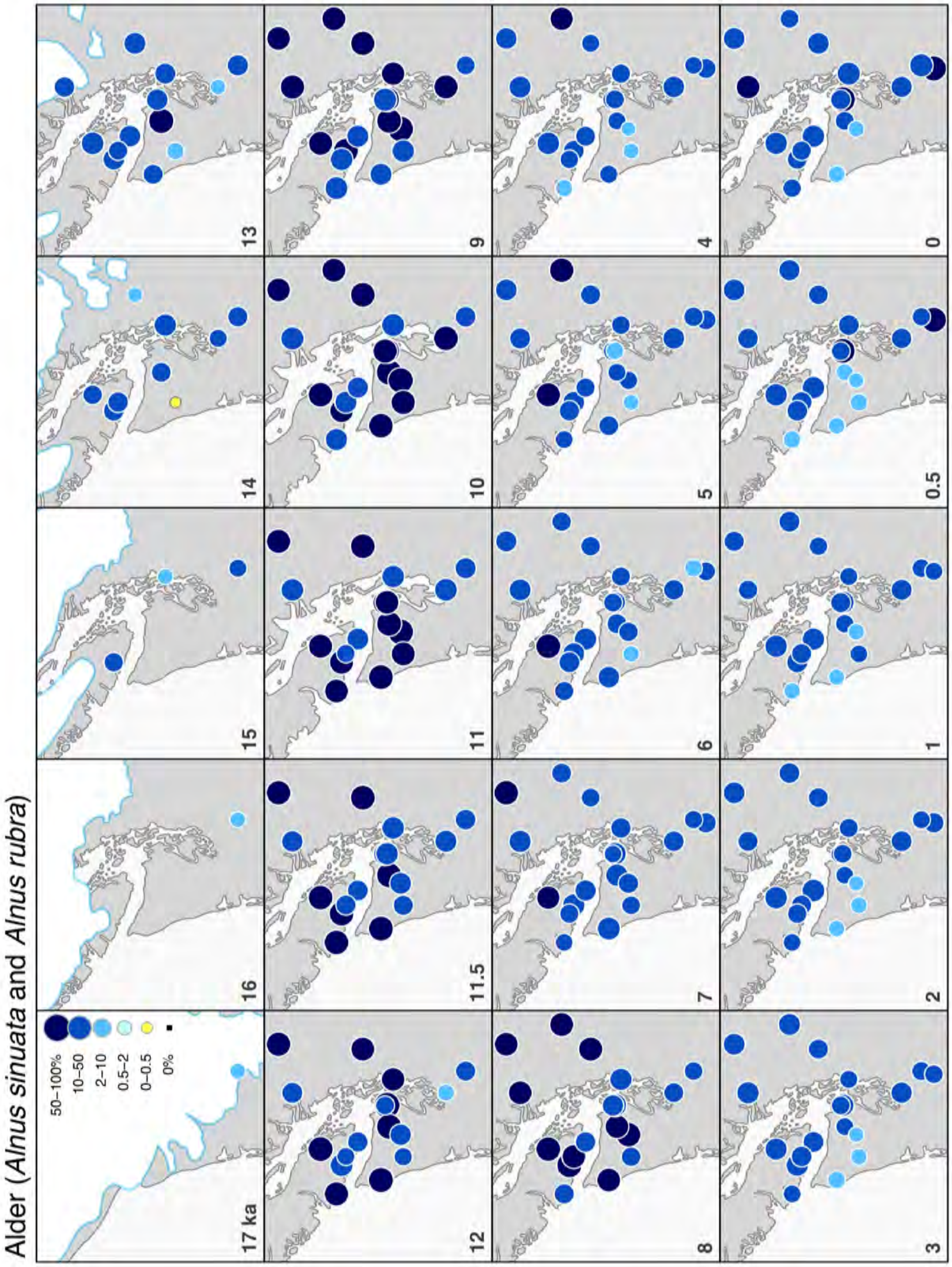

Alder (Alnus sinuata and Alnus rubra)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35G

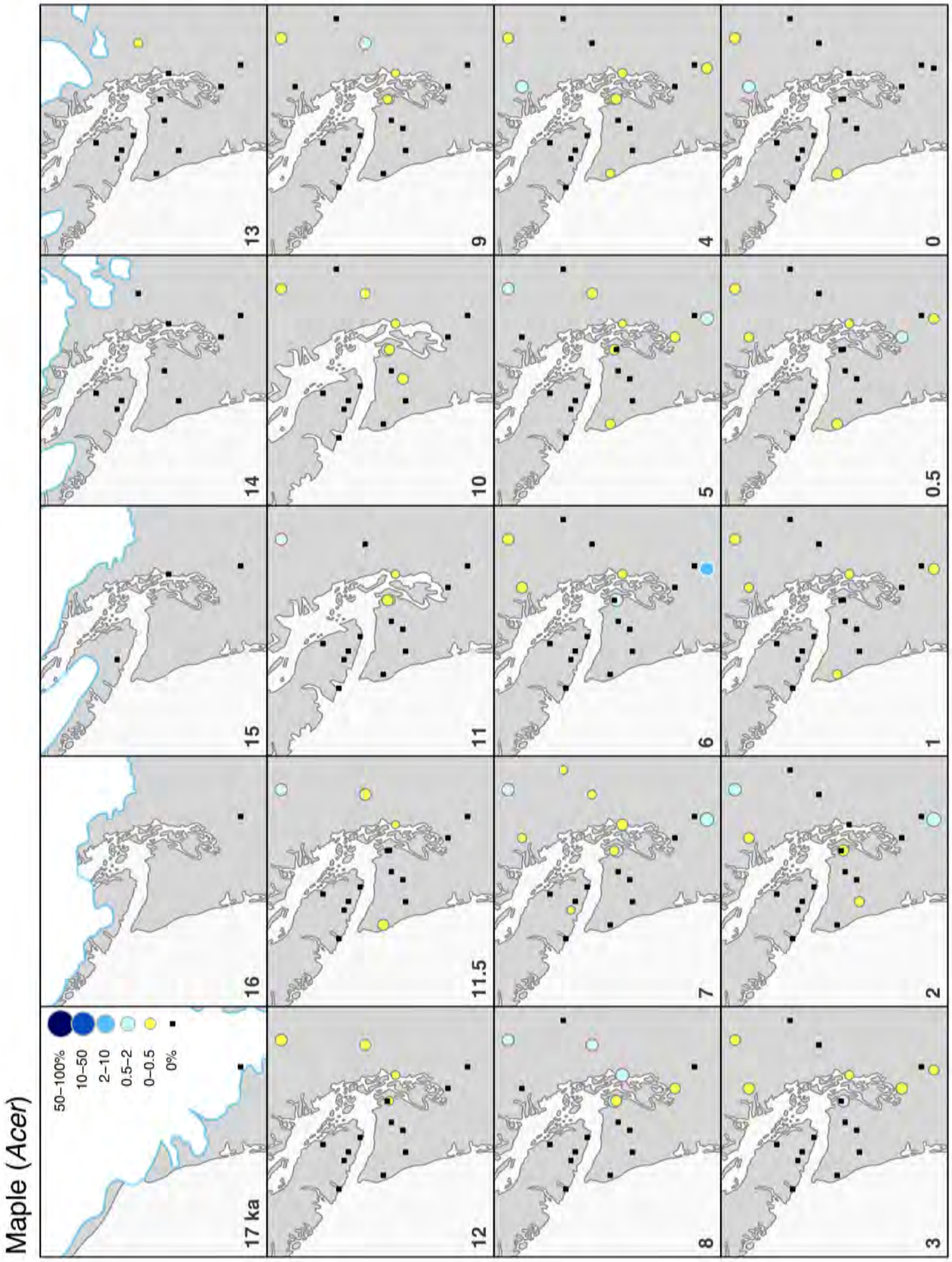

Maple (Acer)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35H

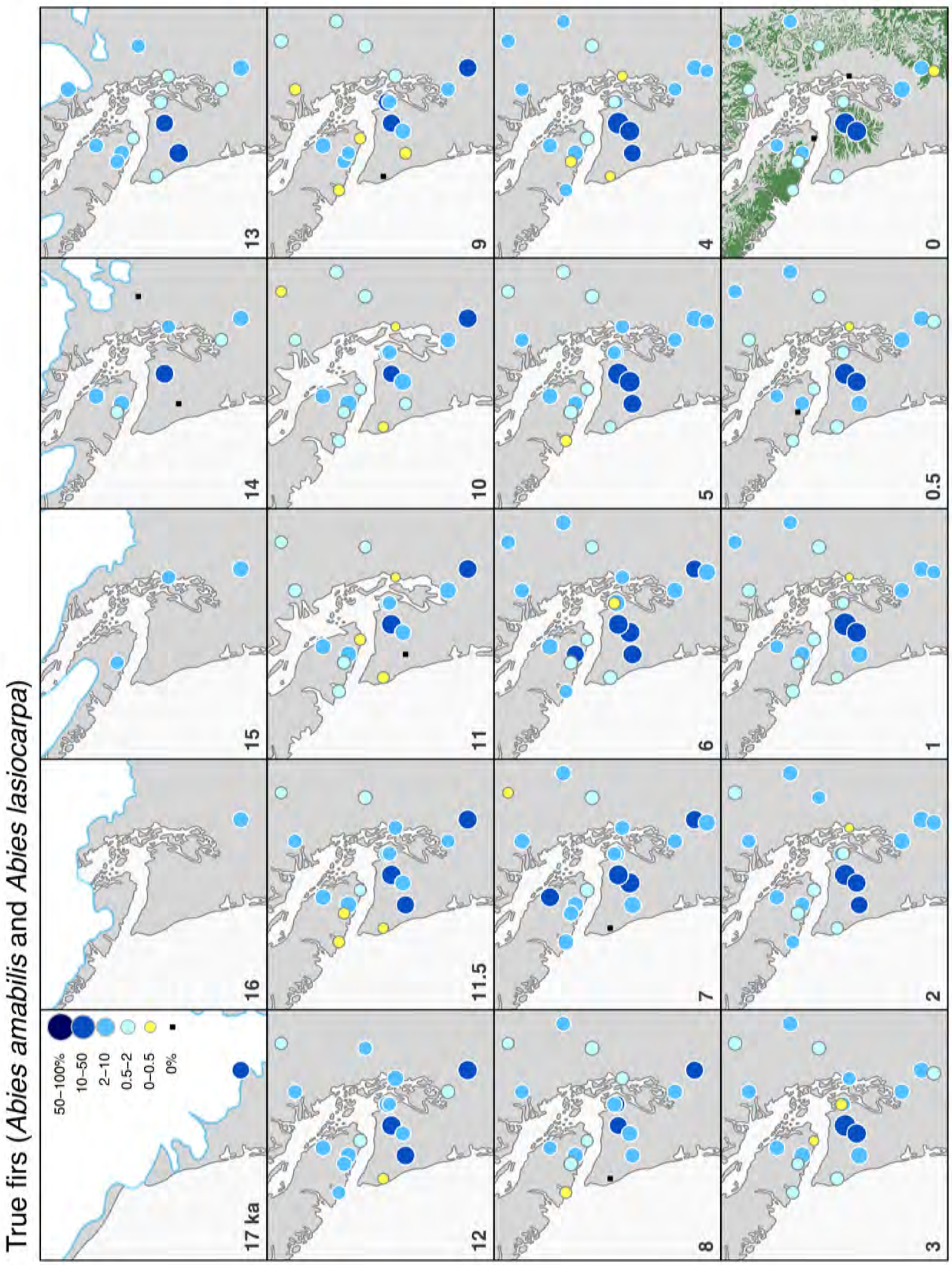

True firs (*Abies amabilis* and *Abies lasiocarpa*)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35I

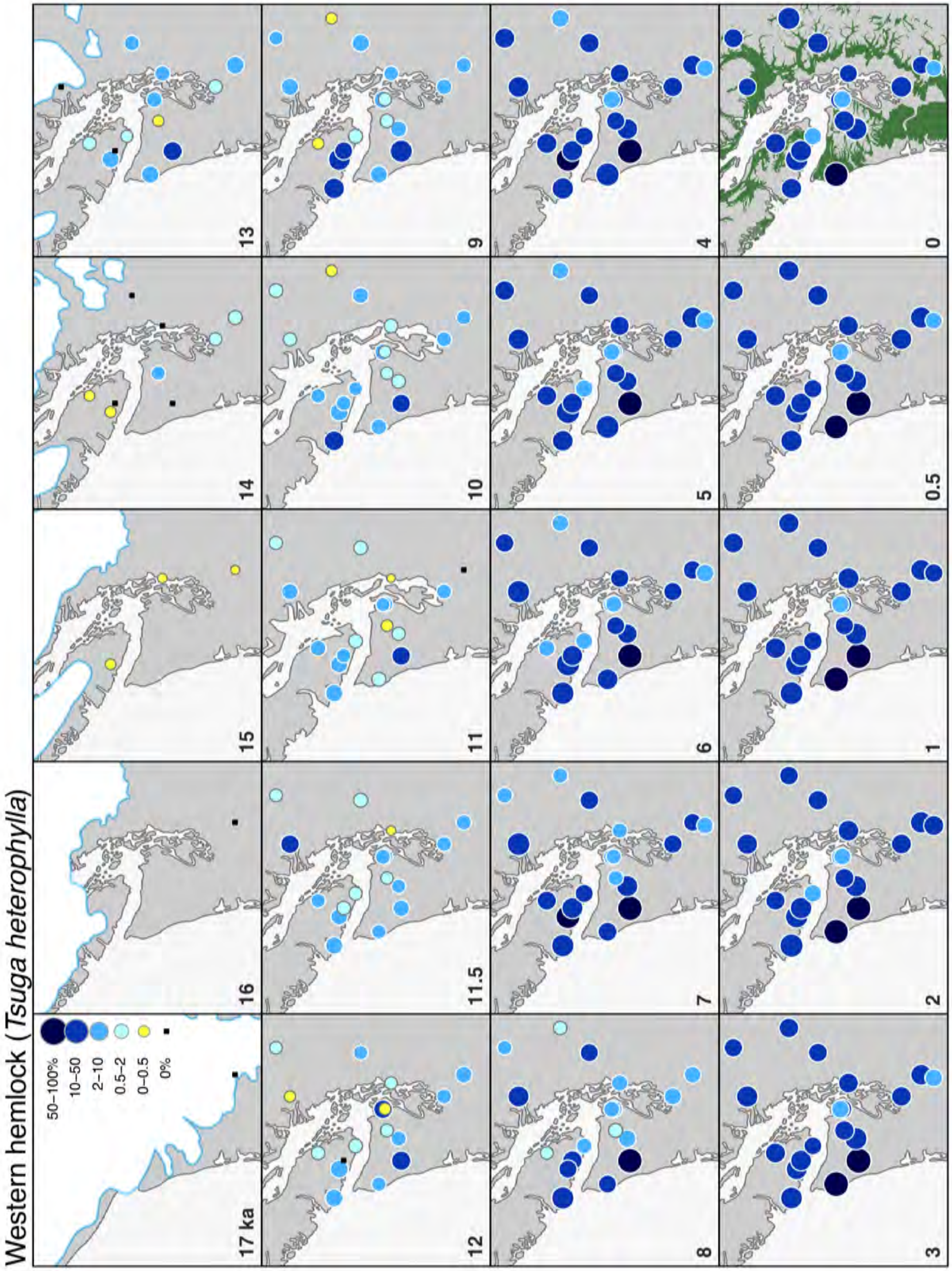
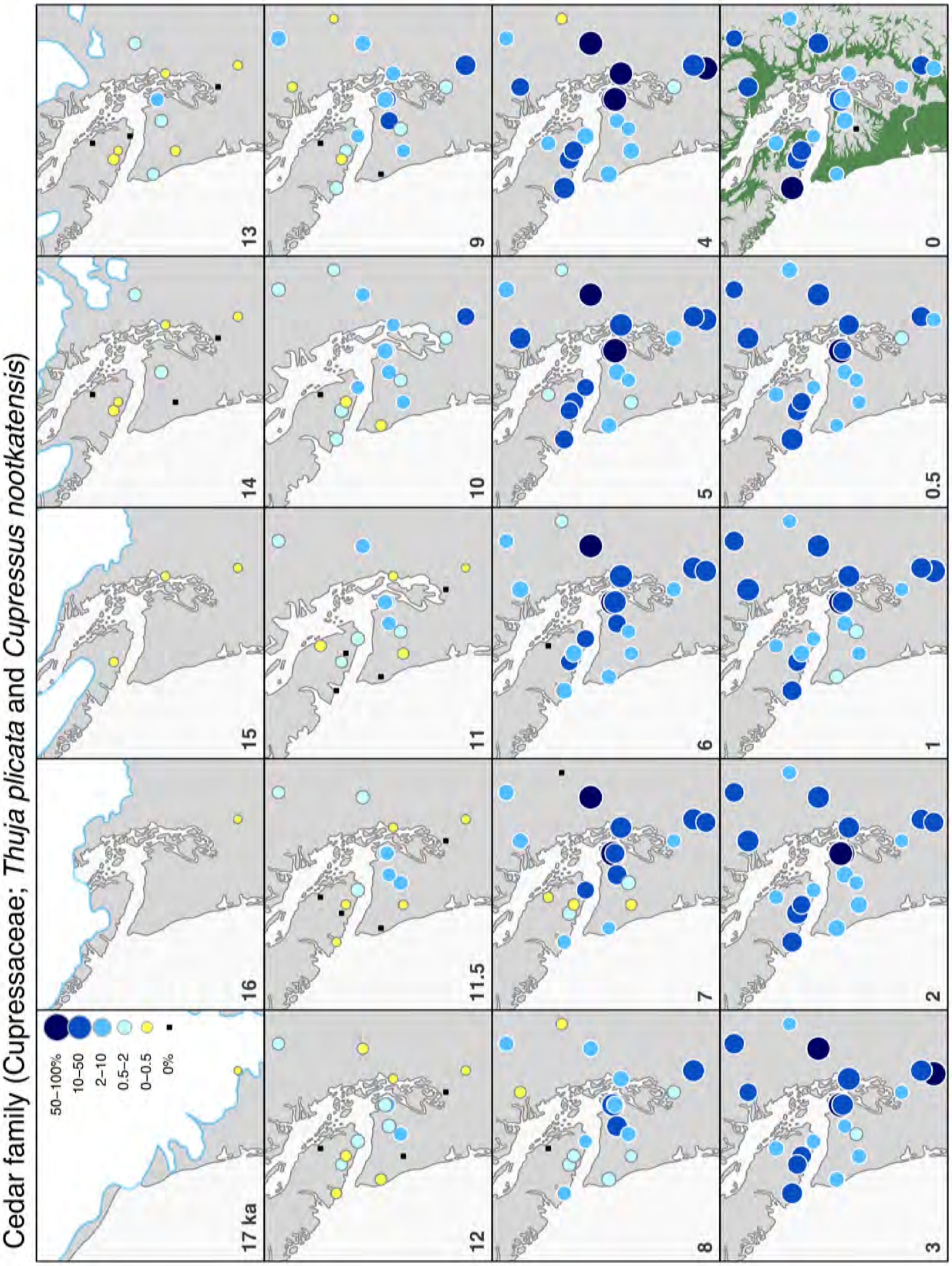

Western hemlock (Tsuga heterophylla)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35J



Cedar family (Cupressaceae; Thuja plicata and Cupressus nootkatensis)
50–100%
10–50
2–10
0.5–2
0–0.5
0%
17 ka
16
15
14
13
12
11.5
11
10
9
8
7
6
5
4
3
2
1
0.5
0

Figure 35K

6. Brief review of the archeological record in a context of environmental change

Humans have inhabited the Olympic Peninsula from ca. 13 ka to present and experienced a dynamic changing environment. While the general climate presents a challenge for preserving archeological sites, several robust and well-dated discoveries on the peninsula provide an opportunity to correlate evidence of human prehistory with the paleoecological record. Here we present a brief review of some well-studied sites and discuss their paleoenvironmental context. A detailed review of Olympic Peninsula prehistory, and the difficulty of obtaining well-dated sites, is in Schalk (1988).

6.1 The Manis Mastodon Site

This site south of the town of Sequim was discovered during excavation for a pond in a pasture in 1977. The discovery of this mastodon, and associated bison, caribou, and muskrat, is recounted in several places (Gustafson et al. 1979, Kirk and Daugherty 2007). The significance of the find was immediately apparent upon discovery of a bone projectile point embedded in a rib. The age of the mastodon was determined by radiocarbon dates of wood associated with the bones, with the most precise date yielding a calibrated age of 13.6–13.8 ka (Gustafson 1979). This event pre-dates the Clovis period that began at 13 ka. More recently, Waters et al. (2011) obtained additional evidence on the embedded bone. A CT scan indicated that the embedded bone was a sharpened mastodon bone and bone growth indicated that the mastodon was not killed directly by the injury. AMS radiocarbon dates on bone collagen place the age at 13.76-13.86 ka. This age is remarkably similar to butchered bison bones found in Ayers Pond on Orcas Island, which had a bone collagen radiocarbon age that calibrated to 13.78-13.89 ka (Kenady et al. 2011). The statistically identical age of these two sites, both discovered accidentally, suggests this was an important time in human presence in the region. Both sites, in shallow water, are consistent with a practice of storing carcasses in bogs that would freeze over (Kirk and Daugherty 2007).

The date of these discoveries, 13.8 ka, is within a period of great climatic variability. The pollen record from the peat soils at the mastodon site revealed open meadow conditions persisted from 13.8 to 12.9 ka (Petersen et al. 1983). Presence of prickly-pear cactus pollen and hornwort seeds, both species that have northern limits close to the Olympic Peninsula today, suggest summer temperatures were similar to today. The site was also drier than present, based on the presence of Douglas-fir forests before recent land clearance. However, this may have been due simply to the fact that Douglas-fir had not arrived to the area despite suitable climate (i.e., it was absent in western Washington at 13.8 ka because it was dispersal limited). In contrast to the evidence of open vegetation at Manis, our two best-dated pollen records on the peninsula, (Yahoo and Wentworth lakes on the western peninsula) and Pixie Lake on Vancouver Island (Brown and Hebda 2002) show that 13.8 ka dates to within 100 years of a rapid transition to dense forests. At these sites, pine pollen decreased and the pollen of mesic conifer species increased, including Sitka spruce, Pacific silver fir, mountain hemlock, as well as the initial increase of western hemlock. That these vegetation changes were synchronous at three well-dated sites and that multiple species increased in abundance suggests a large increase in moisture and possibly an increase in winter temperature (which may have precluded establishment of cold-sensitive species such as western hemlock). At Manis, increased tree pollen occurred closer to 10 ka, at a time when western hemlock at Wentworth Lake makes a second increase and a diverse mammal fauna occurred on Whidbey Island (Mustoe et al. 2005).

The large increase in moisture at 13.8 ka would have increased the precipitation gradient on the peninsula and thus would have increased the climatic uniqueness (and relative habitability) of the rainshadow environment. The increase in moisture at this time also fueled glacial advances. The well-studied Sumas moraines on the lower mainland of British Columbia date to 13.7 to 13.2 ka, indicating valley glaciers descending the Fraser River close to the modern coastline (see section 4.3). The northeastern Olympic Peninsula would have been an oasis between a periglacial environment on the mainland and cold and wet climate on the west coast. After this period of increased forest density, the colder and likely drier climate that returned during the Younger Dryas (12.9–11.6 ka) is marked by very little archeological and faunal evidence in the region (Fedje et al. 2011).

6.2 Late Holocene cultural phases

The scarcity of archeological sites dating to the Early Holocene, and the few sites dating to the Late Holocene, precludes making strong correlations between environmental changes and cultures on the Olympic Peninsula. Most dates of Early Holocene artifacts are spear tips and tools on mountain ridge sites indicating hunting and gathering (McNulty 2009). In contrast, coastal sites are not evident until the Late Holocene. A recently published sea-level curve for southeastern Vancouver Island (James et al. 2009) suggests that the lack of earlier coastal sites is due to sea-level rise, as all coastal sites date to the period after modern sea levels stabilized at 4 ka (Figure 36). For the Late Holocene, however, there is sufficient evidence of changing culture and resource use that can be placed into the context of the new paleoenvironmental reconstructions presented here.

The Locarno Beach Cultural Phase occurred from 3.5 to 2.4 ka (Croes 1989). The archeological record indicates that this is a period of increased storage economy, but little cultural differentiation across the region (Croes 1989). Climate and vegetation records indicate that Locarno Beach was a period of increasing cool and wet climate. The Tiedemann Advance in the southern Coast Mountains occurred 3.3 to 2.9 ka (Figure 22; Menounos et al. 2009). The detailed record at Yahoo Lake shows an absence of fire and an increase in late-successional Pacific silver fir and western redcedar. Interestingly, sea-surface temperatures increase at 3.3 ka which could have resulted in increased moisture advecting onshore (Barron et al. 2003). The only well dated site on the Olympic Peninsula to this period is the Hoko River site on the northern Olympic coast, which spans the period from 3.2 to 2.7 ka (calibrated ages presented in (Ames 2005). The site was interpreted as a fishing camp, with hook and net artifacts, and provided evidence of the development of both marine and riverine fishing (Croes and Hackenberger 1988).

The Marpole Cultural Phase occurred from 2.4 to 1.2 ka (Lepofsky et al. 2005). It is marked by increased numbers of large houses and settlements of multiple houses, regional art forms, and symbols of prestige (Grier 2003). This is a time of a lack of

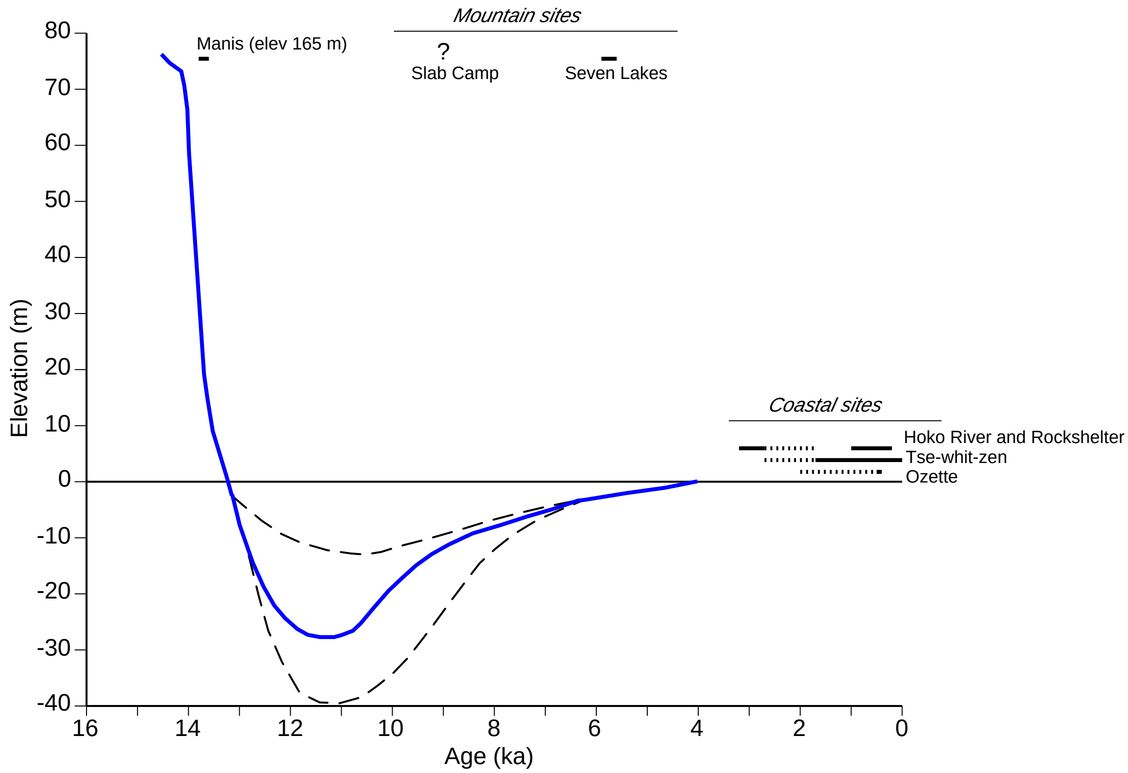


Figure 36. Sea-level curve for southeastern Vancouver Island showing the combined effects of isostasy and sea-level rise on relative sea level (James et al. 2009). The dashed lines represent the range of possible values constrained by radiocarbon dates, and the blue line is the preferred curve by James. Note a sea-level rise of almost 10 m between 8 ka and 4 ka. Dates of important archeological sites are plotted above. Solid lines indicate age-ranges of occupancy of the sites and dashed lines connect outlier dates to the remainder of dates at the site. Note that the coastal site elevations are approximate. See Schalk (1988) for a full list of dates.

glacial advances (Figure 22) and increase fire occurrence throughout the Pacific Northwest (Gavin et al. 2007). Lepofsky et al. (2005) explored the environmental changes at the time possibly stemming from warm and dry autumn climate that promoted natural fire occurrence as well as anthropogenic fire. Increased fire in mountain meadows may be human-caused. The climate at the time may have resulted in greater ease of traveling to subalpine elevations, with flammable conditions persisting into fall allowing for burning to promote blueberries. Indeed, on the Olympic Peninsula, fire increases during Marpole time at Yahoo and Martins lakes, the latter occurs in subalpine meadows. While there are relatively fewer sites on the Olympic Peninsula dating to Marpole time compared to the Georgia Strait and Fraser River, the recent excavations of the large village of Tse-whit-zen in Port Angeles suggests some regional coherence between archeological sites on the peninsula and elsewhere. This site was occupied from 1.7 ka to historic times, with the oldest radiocarbon age of 2.7 ka.

The Ozette village site, dating to ca. 1500 AD, deserves special mention because of the excellent preservation and large number of artifacts. This site is a set of at least four cedar plank houses that were buried in a mudflow. The quick burial and water saturation resulted in excellent presentation of over 55,000 wood and fiber artifacts

(Ames 2005). The oldest midden age at Ozette indicates the site dates back to (at least) the early Marpole period.

A study by Hutchinson and McMillan (1997) advanced the theory that stories of village abandonment following seismic and tsunami events are common among coastal Washington tribes, and that these intergenerational stories should be taken as cues to search for archaeological evidence of events that may have convinced tribal peoples that their village was unsafe for habitation. The most famous of these villages is Ozette, but its connection to a known seismic event is not grounded by accurate dating. The initial estimated age of ca. 1500 AD has been re-evaluated to align the age with the major earthquake in 1700 AD (Kirk and Daugherty 2007). However, some paleosiesmic studies suggest that subduction zone earthquakes recur on a roughly 500-year interval, with only the most recent event in 1700 AD (Atwater 1987, Atwater et al. 2005). Recent paleoseismology studies from marine turbidites have shown that an earthquake dating to 1500 AD occurred along the Olympic Peninsula (the “T2” event in Goldfinger et al. 2012). This event aligns more closely with the reported radiocarbon dates from Ozette (Schalk 1988).

7. Needed studies and actions

While this report depicts a rich and detailed environmental history on the Olympic Peninsula based on sediment-based studies, there remains potential for improving our understanding in several areas. At the millennial scale, our results confirm previous studies regarding the general sequence of vegetation change at low elevations (Heusser 1977, Whitlock 1992). Examination of mapped vegetation patterns and sub-millennial variability in the records raises many questions for future research. We provide these recommendations as a list, and their order is not meant to convey relative importance.

- The chronologies of many sites, based on radiocarbon dates, are too poor to allow constructing regional patterns of forest composition in single time slices. For example, determining the synchroneity of Late Glacial vegetation changes is limited by the difficulty to age organic-poor sediments. However, dating by AMS methods, which requires as little as 50 mg of plant material, has become standard only within the last 15 years. Better-dated records are required to verify existing records and to accurately map the spatial spread of populations. In some cases, archived material exists from sediment cores that may be used to provide new chronologies on existing pollen records.
- Several specific aspects of vegetation history were unexpected and further investigation may reveal important aspects of vegetation-fire-climate interactions.
    - The outer west coast revealed less sensitivity to changing regional climate, perhaps due to the moderating effect of onshore winds (as initially proposed by Henry Hansen). However, upslope of the coastal plain in the Pacific Silver Fir zone, the forest experience a very dynamic history. This increased sensitivity may be partly explained as an artifact of different sampling resolutions among sites. However, replicating sites with similar basin characteristics as at Yahoo Lake may result in records that are more comparable across the vegetation zones on the peninsula.
    - The Manis Mastodon site is world-famous as a rare site demonstrating human hunting of megafauna. The relative contribution of hunting versus climate change towards megafauna extinction remains debated. Studies that reconstruct the ecological changes during megafauna decline, and reconstruct the decline itself, have the potential to help distinguish among the mechanisms causing the extinctions. Such a study in Indiana used the dung-fungus *Sporormiella* to reconstruct declining megafauna abundance (Gill et al. 2009). The Sequim area, and sites on the Kitsap Peninsula, contain numerous deep kettle lakes that may provide excellent undisturbed sediment for studying the relative sequences of vegetation change, faunal change, and forest fire.
    - All sites showed abrupt responses to climate change at the beginning of the Holocene suggesting that despite the long life spans of the dominant tree species vegetation generally responded to climate change with minimal lag. However, were there significant periods of disequilibrium resulting in no-analog communities? For example, the association of Sitka spruce and subalpine fir would be considered impossible based on modern plant geography on the peninsula, though it occurred twice at low elevations in the Late Glacial and Early

Holocene. Similarly, the mapping of pollen data provided some evidence that Douglas-fir arrival to the region lagged behind climate change.

    - Slow colonization of Mountain Hemlock Zone forest presents a conundrum that requires more study. Was this site limited by soil texture, and how common was this in the Olympics, and what is the implication for development of meadows and subalpine forests during the Early Holocene?
    - The Early Holocene landscape supported frequent fire and levels of Douglas-fir and red alder that approach what occurs on industrially managed forest-land today. What was the spatial mosaic of this vegetation type? Did alder colonize hillslopes similar to what occurs today in the Oregon Coast Range? Has red alder greatly improved the nitrogen status of Early Holocene soils, boosting forest productivity for subsequent millennia?
- The detailed paleoecological records spanning a climatic gradient provides an opportunity to use paleoecological records to calibrate vegetation models. Dynamic vegetation models, such as LPJ-GUESS, can use climate inputs to predict the likely set of taxa on the landscape (e.g., Miller et al. 2008). Such an approach should be feasible considering the strong abiotic controls of modern vegetation on the peninsula and the strong climatic control of vegetation in the past. Our initial application of LPJ-GUESS, requiring some modifications to the model, is providing promising results showing a match between the pollen records and the simulated vegetation. Such a model could then be used to predict vegetation change under a trajectory of changing climate into the future. For example, two ocean-atmosphere general circulation models (CGCM3 and HADGEM1) run under the greenhouse gas scenario of > 600 ppm CO_2 by the year 2080 are projecting a 3 to 5 °C increase in December-February mean temperature, a ca. 4 °C increase in June-August temperature, and a 15 to 40% decrease in summer precipitation in an area including Yahoo Lake (Wang et al. 2012). Such a summer climate appears to be close to the warm and dry periods of the Early Holocene at 10 ka, as inferred from the chironomid temperature reconstruction. A rapid reversion of the forest to the Early Holocene state would be dramatic. However, such a compositional shift will require the occurrence of large fires to overcome strong neighborhood effects on species composition. The 10 ka analog for future climates is crude because of differences in the seasonal climatic patterns between the two periods. A dynamic vegetation model could address such differences.
- There is potential to infer more dimensions of the paleoenvironment using interdiscisplinary biogeographic methods of modeling distributions, paleoecology, and genetics.
    - Can phylogeographic studies determine which of the endemic and disjunct taxa are likely to have been long-term (>20,000 years) residents of the peninsula? A common approach is to use molecular genetic approaches to show whether a disjunct population diverged from other populations (or sister taxa) before the Last Glacial Maximum (LGM), thus suggesting persistence in a glacial refugium (Hu et al. 2009). Very few such studies to our knowledge exist on the Olympic Peninsula.
    - Can we improve upon the reconstructions of vegetation and climate during the glacial period? Most work has focused on lakes that formed during the initial

glacial reteat around 14 ka. Cal Heusser's long pollen records from bogs, however, extend further back and provide an important view of the LGM, but hiatuses in the sediment and poor chronological control makes interpretation difficult in many cases. As discussed above, there is potential to improve these long records with better radiocarbon dating and possibly dating to before 40 ka using optical-spin luminescence.
  - What is the climatic niche of the 'paleoendemic' species? Does our understanding of past climate agree with their long-term persistence on the landscape? Can this approach provide new information regarding plasticity of the niche?
- Paleoecological studies may be designed to address knowledge gaps in specific settings.
  - The prairies in Sitka spruce forests near Ozette, Forks, and Quillayute are widely regarded as anthropogenic, with significant ecological and ethnographic support for their origin and maintenance (Reagan 1909, Anderson 2009). However, their long-term presence and their biotic significance is not known. Radiocarbon dating soil charcoal and examining any stratigraphic record near these sites could provide a history of these sites that has the potential to be older than coastal sites that are lost from sea level rise. Much work has begun on this topic (D. Conca, personal communication).
  - Similarly, intentional burning by people for berries in subalpine meadows has not been studied on the Olympic Peninsula.
  - The elk glades in riparian forests appear to be a recent phenomenon due to high elk populations following wolf extirpation. What is the longer history of forest density in terrace forests of the Hoh, Bogachiel, Queets and Quinault Rivers? Paleoecological studies, carefully done, could address this question though use of "small hollow" sediments that record local forest density. One such study on the Queets River showed the potential for this approach, but the study was not located in areas of high elk density (Greenwald and Brubaker 2001).
  - Fire occurrence the Olympic Peninsula is highly episodic. Fire periods in the past were unlike any in recent history. The dendrochronological record, much of it already collected, has much potential to describe the extent of fire on particular years or decades (e.g., Henderson and Peter 1981, Wendel and Zabowski 2010). Such studies could provide precedents for high-fire-risk years during severe drought.

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Appendix A: Treatment of pollen data for production of regional maps

We obtained pollen data from the Neotoma Paleoecology Database (*http://www.neotomadb.org*), from published diagrams, and from unpublished data of the authors. We included all sites with a radiocarbon-based chronology from west of the Cascades and from the southern Puget Lowlands to southern Vancouver Island. If data were not available in digital format, we digitized the values from published pollen diagrams using GraphClick software.

Creating pollen maps requires quality control on each site contributing data. Without such assessment, outlier values may greatly affect interpretation at each site. For each of the 20 sites, we developed new relationships between depth in the sediment and age of the sediment (age models) by fitting a spline curve through all age-control points (calibrated radiocarbon dates, tephras, and core top dates). These relationships have already been described for the five sites on the Olympic Peninsula. We avoided using samples that required extrapolation below the lowest radiocarbon date. Top ages were based on the publication year of the paper if the coring date was not available.

Pollen percentages were expressed as a percentage of all arboreal pollen taxa (*Abies, Acer, Alnus, Betula, Chrysolepis/Lithocarpus, Thuja plicata, Fraxinus, Picea, Pinus, Populus, Pseudotsuga menziesii, Quercus, Tsuga heterophylla, Tsuga mertensiana*). Some pollen records differentiate *Alnus rubra* and *Alnus sinuata* while others do not. For consistency across sites, we map the pollen type *Alnus*, combining the percentages of the two species when necessary. While traditional methods express pollen as a percentage of all terrestrial pollen types, our intent was to limit interpretation to the dominant tree species and was the simplest means to avoid issues with between-site variability of non-arboreal pollen taxa that could affect pollen abundance.

For each pollen taxon, pollen percentages were estimated by linear interpolation between the samples straddling each of 20 time slices. Percentages were square-root transformed and symbolized to emphasize differences in low abundances. Basemaps of ice extent and coastlines are from Dyke et al. (2003). Data were interpolated and mapped with the R computing environment, which allowed simple updating of data.

Table A1. Late Quaternary pollen records used for mapping pollen abundance. Site numbers refer to Figure 33. Notes indicate special consideration of certain chronologies. See Table 2 for site locations.

Site	Name	Source[1]	Chronology[2]	Other notes	References
23	Cedar Swamp	Author	B:3 A:3 T:1 D:1 (0.3–12.1)	Lower part of core not counted. Removed one bulk date due to AMS date at same level.	(McLachlan and Brubaker 1995)
22	Crocker Lake	Author	B:4 A:1 T:1 D:4 (0.03–13.2)	Tied pine decline to 11.6 ka and basal age tied to Cedar Swamp AMS age	(McLachlan and Brubaker 1995)
34	Davis Lake	Neotoma	B:16 T:1 (-0.01–7.1)	Did not use data below Mazama because 1) suspect hard-water effect on bulk dates in clay-rich sections, and 2) pollen differed greatly from nearby Mineral Lake	(Barnosky 1981)
16	East Sooke Fen	Neotoma	B:4 T:1 (0.2–13.5)		(Brown and Hebda 2002)
26	Hall Lake	Digitized	B:6 T:1 D:1 (0.03–15.2)	Radiocarbon times scale translated to depths, then to calibrated ages.	(Sugita and Tsukada 1982)
17	Kirk Lake	Neotoma	B:4 T:1 (-0.03–14.1)		(Cwynar 1987)
4	Marion Lake	Neotoma	B:7 T:1 (0.04–9.3)	Used Mike Lake for the period before 9.3 ka due to its stronger chronology	(Mathewes 1973, Mathewes and Heusser 1981)
27	Martins Lake	Author	A:2 T:1	Pine decline tied to 11.2 ka	(Gavin et al. 2001)
5	Mike Lake	Digitized	A:5[3] (9.9–13.8)		(Pellatt et al. 2002)
33	Mineral Lake	Digitized	B:7 T:4[4] D:1 (-0.03–20.8)	Radiocarbon times scale on diagram translated to depths before translating to calibrated ages.	(Sugita and Tsukada 1982)
24	Moose Lake	Author	B:4 A:1 T:1 D:3 (-0.01–14.3)	Pine decline tied to 11.6 ka	(Brubaker and McLachlan 1996, Gavin et al. 2001)
31	Nisqually Lake	Digitized	B:1 T:1 (0.04–14.9)		(Hibbert 1979)
11	Panther Potholes	Author	A:6 T:3 (-0.04–10.6)		(Prichard et al. 2009)
2	Pinecrest Lake	Neotoma	B:1 T:1 (0.04–12.9)		(Mathewes et al. 1972, Mathewes and Rouse 1975)
14	Pixie Lake	Neotoma	B:7 T:1 (0.02–15.6)	Did not extrapolate below lowest ^{14}C age	(Brown and Hebda 2002, 2003)
8	Porphyry Lake	Neotoma	B:4 (0.30–14.7)	Did not extrapolate below lowest ^{14}C age	(Brown and Hebda 2003)
15	Walker Lake	Neotoma	B:4 T:1 (0.03–14.2)	Did not extrapolate below lowest ^{14}C age	(Brown and Hebda 2003)
21	Wentworth Lake	Author	A:5 T:1 (0.3–13.9)		This study

10	Whyac Lake	Neotoma	B:4 (0.3–12.7)	Did not extrapolate below lowest ^{14}C age	(Brown and Hebda 2002, 2003)
28	Yahoo Lake	Author	A:5 T:1 (0.6–14.6)		Gavin et al. in review

1. "Neotoma" is the Neotoma Paleoecology Database. "Digitized" indicates values were hand-digitized from pollen diagram graphics using GraphClick software. "Author" indicates data were provided directly by researchers.

2. B: number of bulk (conventional) radiocarbon dates, A: number of AMS (accelerator mass spectrometry) radiocarbon dates, T: number of tephras with a known age, D: number dates not used because deemed unreliable. Values in parentheses indicate total age range used in maps (in ka, thousands of calibrated years before present).

3. Four dates on pollen concentrates, one on a needle.

4. Three tephra ages not used (insufficient information to date).