

COLUMBIA **RIVER** BASIN ECOSYSTEMS: LATE QUATERNARY ENVIRONMENTS

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## INTRODUCTION

### The Past in Perspective

After more than a century of exploiting the Northwest's natural resources, mounting population and societal concerns as well as maturing ecological perspectives are causing us to reconsider management practices. For example, in their assessment of management effects on ecosystems, Johnson and others (1994) recommended reintroducing the natural disturbances that direct regional vegetation dynamics. Knowledge of past vegetational associations and ecological processes that illustrate the variety of pathways to current landscape mosaics strengthen understandings of these dynamics (Arno and others 1995; Foster and others 1992; Hann 1991; Schoonmaker and Foster 1991).

In the span of a human life-time we might explain vegetation history through observing plant succession and the apparent enduring harmony of climax communities. The scales of our observations, however, restrain notions of nature's abiding balance (Tausch and others 1993) and short term observations furnish few analogs for the magnitude and pace of past and potential vegetation changes. Traditional, idealized stable cycles, always returning to the "natural" state, are just a few frames in a continuous movie (Graham 1988). They are illusions of forests and steppe primeval, the pristine vegetation of the imagination, or Thoreau's "bog in our brain and bowels". (Torrey and Allen 1985, p. 43).

Effects of cows, plows, and alien weeds have accelerated the movie, but still leave room for short-term predictability with no necessary appreciation for the longer view. The specter of sudden global warming and the fates of ecosystems (Brubaker 1988; Franklin and others 1991) are another matter. Appreciating the changeable character of biotic communities is crucial to future skillful management (Johnson and Mayeux 1992). Knowledge about how vegetation varies resides in fossil records--nature's archives--and deciphering these chronicles is the ecological challenge for the future.

Two principles emerge from Quaternary vegetation studies world-wide: change is **continual** and change is unpredictable. First, though perhaps imperceptible in the geographic center of mixed montane forest or steppe, sensitive communities on the fringes often show that plants, even **long-lived** trees, respond rapidly to environmental perturbations. For example, buried needles, cones and whole logs of spruce and subalpine **fir** above **treeline** in the Canadian Rockies show **upslope** advances to **>10** m above present **treeline** between 9000 and 5000 B.P. and again near 1000 B.P. (Luckman 1990). These warm ( $+0.5-1.5^{\circ}\text{C}$ ), temperature-controlled responses at **treeline** affect only a small area of vegetation. In the terrain of the Columbia River Drainage, however, even slight changes at the lower forest border have a great influence on total areas of vegetation types.

The Northwest's steppe-conifer ecotone has been in flux over the past 12,000 years. In eastern Oregon, historic and past successes of western juniper are impressive because a small increase in effective moisture brings broad expanses of juniper woodlands (Miller and Rose 1995). Likewise, the Holocene has seen significant vegetational changes along the fluctuating **steppe-ponderosa pine** border in eastern Washington. Change is perpetuated not only by plant responses to climate, but by disturbances that accompany climatic change; fire and disease often follow drought **If** vegetation changes lag climate changes (Davis 1989; Franklin and others 1991) then, in the long view, vegetation often is out of equilibrium with climate and is subject to rapid variation through natural disturbances. Understanding the history and probable consequences of these disturbances gives land managers potent tools for selecting ecologically and socially acceptable alternatives for influencing the course of changing ecosystems (Arno and others 1995).

Unpredictability, the second principle, is best illustrated by unusual late glacial and Holocene assemblages of both plants and animals. These nonanalog assemblages suggest that individuals, rather than communities or vegetation zones, react to climatic change and that fossil assemblages and modern communities are loosely organized collections of individually distributed species. With disturbances--both natural and human made--species may be redistributed along different environmental gradients, in different directions and at different rates, and may reassemble in unpredictable ways (Graham and Grimm 1990). So, the **recency** of some familiar plant associations is evident from the fossil record; present communities are not necessarily good guides to the past nor to the future (Botkin 1990, p. 60-62); and, in Quaternary vegetation studies surprises are the rule (Franklin and others 1991). Yet, with fossil plant records and the vision of Janus we may come to understand the past and thereby achieve a more sagacious view of future ecosystems.

## **Late Pleistocene Environments**

The remarkable vegetational shifts described for the Columbia River Basin during the last 20,000, 10,000, or 1000 years are not unusual. In fact, no vegetation on earth escaped the repeated challenges of glacial climates and interglacial adjustments, or the short, sharp climatic shifts during each episode. With every glacial-interglacial cycle species displaced by climate, ice, water, and competition faced local extinction; they responded through growth form and physiology (Van de Water and others 1994; Wigand 1995), migration; or adaptation. At varying paces in different places species repeatedly abandoned then reclaimed the same terrain, but not always with the same associates.

On a global scale, during the last 20,000 years the grip of glacial climate squeezed North Africa dry, allowing cold winds to heap-up Saharan sands and sweep dust clouds far out to sea. It brought polar desert and steppe to Europe, and coniferous forests to the eastern U.S. North American deserts sported snow-capped mountains, pluvial lakes and pygmy woodlands while savanna spread in the New World tropics.

Early Holocene monsoons nourished deflation basins, **shrubby** steppe and Neolithic pastoralists, then retreated to the Sahara's southern fringe. With postglacial warming, temperate deciduous forests regained dominance in Europe and the eastern U.S. Western America's pluvial lakes rapidly shrank to ephemeral **playas** and woodlands abandoned our deserts. Savanna gave way to shrubs and trees in tropical South America

The last glaciation brought Laurentide or **Cordilleran** Ice sheets to the Northwest and ice caps grew in the larger ranges (**Waitt** and **Thorson** 1983). The greatest area of ice south of the continental ice sheets covered the Yellowstone Plateau; large ice caps also formed over the Bitterroot and Sawtooth ranges, and along the spine of the Cascades. Other ranges supported smaller ice caps or cirque and **valley** glaciers that receded to near present limits by 12,000 B.P. Periodic readvances of alpine glaciers and downslope retreat of upper treelines mark cold moist episodes of the Holocene (Davis 1988).

Pleistocene lakes drowned vast areas of the Northwest and Great Basin and, along with glaciers, acted as barriers to plant movements. They also may have contributed to restriction or demise of populations by their growth and by catastrophic draining. When the lakes shrank or desiccated, plants colonized fine-grained substrates that had accumulated over thousands of years. Some lake floors (e.g., Glacial Lake Missoula) probably were stabilized rapidly by grasses and sagebrushes, whereas others (e.g., **Pluvial** Lake Bonneville) remain sparsely vegetated and still return wind-born sediment to surrounding ranges and adjacent valleys. The late histories of these two lakes illustrate the rapidity and intensity with which brief climatic fluctuations altered the Pleistocene landscape on a grand scale.

Though annual averages and seasonal patterns of temperature and precipitation remain uncertain, from about 20,000 to 13,000 B.P. precipitation sometimes exceeded evaporation throughout **the** Western U.S. and many basins now dry, or nearly so, filled to overflowing (Benson and others 1990). Strandlines of Lake Bonneville, the largest of these basins, remain obvious on slopes to 300 m above the present Great Salt Lake, Utah. When this highest beach formed, the lake--with a surface area of 50,000 **km<sup>2</sup>**--spread into Nevada and Idaho. Then

overflow and downcutting at Red Rock Pass, Idaho, sent a catastrophic flood cascading northward via the Snake and Columbia rivers to the Pacific Ocean. Consequently, for a short while the northeastern Great Basin became a part of the Columbia River drainage.

Subsequently, Lake Bonneville fell about 100 m to its newly cut outlet where it remained long enough to form a distinctive strandline. By 12,000-10,000 B.P., with rising temperatures and Holocene circulation patterns, Lake Bonneville waters had evaporated to the area of present Great Salt Lake.

Along the margins of continental glaciers in Washington, Idaho and Montana, proglacial lakes grew in ice-dammed drainages. The largest of these, Glacial Lake Missoula, formed when ice extended southward in the Purcell Trench and blocked the Clark Fork Valley with a 600-meter-high ice dam at the Montana-Idaho border. Rivers and melting glaciers fed the resulting lake until it spread over 3000 km<sup>2</sup>, with a volume comparable to that of Lake Ontario. Faint wave-formed beaches still mark the hillsides to 290 m above Missoula, Montana. Although estimates of the number vary, periodic catastrophic outburst floods, or jokulhlaups, began with recurrent failures of the ice-dam. With each “Spokane” or “Missoula” flood as much as 2150 km<sup>3</sup> of water poured through the breach, Lake Pend Oreille, and the Spokane Valley. Icy torrents, hundreds of meters deep and tens of kilometers wide, then surged across central Washington, violating drainage divides, stripping **Palouse** loess, and raising gigantic basalt blocks **from** their beds. Washington’s distinctive scablands are the product of some of the greatest floods known from the geologic record; the last of these occurred about 13,000 B.P. (Baker and others 1991; Waitt 1985).

## **Late Quaternary Climates In The West**

Variable surface boundary conditions, or controls, that influenced broad climatic circulation patterns through the late Quaternary include mass of the ice sheets, extent of sea-ice, sea-surface and deep-ocean temperatures, land albedo, dust, and soil moisture (Kutzbach and others 1993). In addition, external forcing came from orbitally caused changes in seasonal distribution of insolation—the **Croll-Milankovitch** hypothesis (Gates 1993). Three aspects of the earth’s orbit around the

sun, and their periodicity, determine the timing of these changes: eccentricity (100,000 years), obliquity (41,000 years), precession (22,000 years). Concentrations of greenhouse gases, such as carbon dioxide and methane, also influenced external forcing and are now implicated in global warming. Today's atmosphere holds 55% more CO<sub>2</sub> than it did during the last glacial maximum (18,000 B.P.).

On a continental scale, generalizations **from** calculations of external forcing include (Kutzbach and Ruddiman 1993):

1. At the time of full glacial conditions (18,000 B.P.) seasonal and latitudinal distributions of solar radiation were close to today's values. Glacial boundary conditions--principally ice sheets and cold oceans--set ice age atmospheric circulation patterns that brought cold dry conditions to the Northwest and pluvial lakes to the Great Basin.
2. External forcing assumed more importance in rapid climate changes after 15,000 **B.P.** Glaciers wasted, oceans warmed, and solar insolation increased with heightened seasonal differences.
3. By 12,000 B.P. summer solar radiation had increased by 7% from 18,000 B.P. values; westerlies brought Pacific moisture to the Northwest.
4. By 9000 **B.P.** summer solar radiation achieved maximum values in the Northern Hemisphere. It averaged 8% higher than today in July and 8% lower in January.
5. After 6000 B.P. seasonal insolation extremes declined toward modern values but, with **loss** of the ice sheets, seasonal insolation continued to be a dominant factor in Northern Hemisphere climate.

In the western U. S. precipitation patterns and temperature are influenced by distance from the coast, latitude and elevation. Mountain ranges also have a profound influence on climate and biota because they produce strong elevational gradients in temperature and precipitation, and store snow that sustains streams through the summer. Most importantly, the Olympic, Cascade and Sierra Nevada ranges intercept the westerly flow of moisture and cast semi-arid rain shadows eastward.

Seasonal rainfall patterns vary geographically. The “Mexican monsoon,” from the Gulf of Mexico or the subtropical Pacific, brings summer rains to the southern deserts, whereas winter westerlies nourish the Northwest. These contrasting precipitation patterns often are in opposition and are also a feature of late Pleistocene climates (Thompson and others 1993). For example, effective moisture was apparently low in the Northwest at 9000 B.P. while the **Southwestern** deserts welcomed more summer storms. However, proxy data do not confirm climatic model extension of this moisture into northwestern Nevada as shown in Figure 1. There, low elevation juniper woodlands decrease in the late glacial and have withdrawn by 9000 B.P.

Though models and simulations may explain the pace and large-scale features of climate change, proxy data--such as plant macrofossils and pollen, tree-rings, lake levels, and oxygen and carbon isotopes--may better reveal the magnitude and rate of change and local variations. They also may delimit climatic boundaries (Neilson 1993). A late glacial, Hemisphere-wide reversal from deglacial warming to rapid cooling (Younger **Dryas**, 11,000-10,000 B.P.) was first detected in pollen and plant macrofossils **from** Scandinavia. It has been **confirmed** by dated glacial readvances, by  $^{18}\text{O}$  measurements **from** lake and ice cores, and by other evidence for significant climatic shifts at this time, including fossil pollen from the Pacific Northwest (Dansgaard and others 1989; Engstrom and others 1990; Johnson and others 1994; Mathewes 1993; Peteet and others 1990; Wright 1989; Zhisheng and others 1993). Causes of the sudden Younger **Dryas** cooling--by an astonishing  $6^{\circ}\text{C}$ , or so--remain clouded, whereas fossil and isotope studies clearly confirm important late glacial fluctuations of climate. Likewise, plant and insect fossils from coastal British Columbia and, perhaps, in the unglaciated Yukon as well (Mathewes 1991a) suggest an unexpected relatively warmer, wetter episode at the height of glacial conditions (18,000-19,000 B.P.). Yet, climatic model simulations predict very cold and dry conditions for northwestern Washington at 18,000 B.P. In the Inter-mountain West, topographic features significantly modulate the effects of large-scale circulation patterns. Modeled environmental changes must be **confirmed** by proxy data (Mock and Bartlein 1995).



A thousand years ago, during the medieval warm episode, subarctic people prospered world-wide and Europe's agriculture, population, cities, and commerce grew. Vikings expanded west to Greenland and to North America while immigrants from Alaska spread Thule Eskimo adaptations eastward across the Arctic islands to Greenland. Then in the A.D. **1300s**, the Little Ice Age gave Europe short, cool, stormy summers and harsh, frigid winters; crop failure and famine followed. Glaciers readvanced in the Alps and Rocky Mountains and sea ice spread southward. The Norse abandoned their Greenland colonies and Thule peoples withdrew to favored locales. As little as **1°C** cooling had affected the most simple and complex of cultures alike. Still, even with historic records for validation of climate change models (Grove **1988**), we do not fully comprehend the cause(s) for **theis** recent and rapid chilling.

The chronology and position of shifting climatic boundaries in the western U.S. are important in explaining the apparent latitudinal differences discerned from fossil pollen and **woodrat midden** assemblages. For example, the degree of penetration of the "Mexican monsoon" determined the limits of typical Sonoran Desert plants and prehistoric agriculturists (Petersen **1994a, b**); or the related position of the Pacific High controlled the flow of moisture into the Northwest. **Whitlock** and Bartlein (1993) employed divergent pollen records to study and explain a Holocene climatic boundary in the Yellowstone, Wyoming, region.

Fossil pollen, plant macrofossils from **woodrat middens**, and changing lake levels hold the history of past climates in the western U. S. Thompson and others (1993) combined these lines of evidence (climate proxy data) with results of climate model simulations to assess both local environments and regional patterns of specific climatic variables. They mapped inferred climatic conditions expressed as effective moisture (relative to today's) for the time of maximum glaciation at 18,000 **B.P.**, the deglacial period of transition to Holocene conditions at 12,000 **B.P.**, the period of **greatest** summer insolation around 9000 **B.P.**, and 6000 **B.P.**, the traditional apogee of the Altithermal. Their studies afford a chronology of region-wide climate change and a basis of comparison with the **Quaternary** vegetation records from the Columbia River Basin.

## STUDY AREA

Columbia River Basin vegetation ranges from the semi-desert and shrub-steppe of eastern Washington, Oregon and southern Idaho, to grasslands, to the moist maritime forest of northern Idaho and adjacent portions of Montana and Washington, to mountains carpeted with a diverse array of coniferous forests to alpine meadows. The geology of the Columbia River Basin is a legacy of Miocene lava flows, glacial topography, volcanoes, and colossal floods (U.S.G.S. 1973).

The Columbia River drains an area of about 650,000 km<sup>2</sup> (Pitzer 1994). With its origins in the Canadian Rocky Mountains west of Banff National Park, Alberta, it flows in a northwesterly direction for about 330 km before turning south through Arrow Lakes to the Canadian border. Its major tributaries include the Kootenai, Clark Fork, Spokane, Okanogan, Snake, Yakima, John Day, Deschutes, and Willamette rivers. The Columbia winds across Washington in a succession of great bends to a point near **Walla Walla** where it turns west and becomes the Washington-Oregon state line. Below The Dalles it cuts through the Cascade Mountains that divide the green timbered coastal slopes to the west from semi-arid sagebrush steppes to the east. Figure 2 shows the study area boundaries and sites that have yielded information on paleovegetation within the U.S. portion of the Columbia River Basin. Several sites beyond the boundaries are included because of their proximity to the Columbia River Basin or because their records illustrate some specific aspect of regional vegetation or climate history.

### Tephra

Volcanic ash layers (tephra) that allow precise correlation of vegetational events within the northern Rocky Mountains (Carrara 1989) and far beyond will prove the greatest asset of this region, over others, in deciphering patterns of Quaternary vegetation history. Multiple eruptions of Cascade volcanoes such as Mount **Mazama** (Crater Lake), Oregon (Bacon 1983), and Glacier Peak (Foit and others 1993; Mehringer and Foit 1990; Mehringer and others 1984), and Mount St. Helens, Washington (Crandell 1987; Mullineaux 1986; **Yamaguchi** and others 1990) cast long

shadows of distinctive tephra layers down-wind (fig. 3). Tephra of individual eruptions may be distinguished and correlated by stratigraphic position, radiocarbon or tree-ring age, and by distinctive mineral suites and glass chemistry.

Figure 4 shows how Mazama and Glacier Peak tephra appear in lake sediments' in distant Montana. These two tephra are tune stratigraphic markers important for correlating vegetation changes from the Cascades as far as Glacier National Park and, perhaps, Yeliiowstone as well (Carrara 1989; **Whitlock** 1993). Figure 5 depicts multiple tephra layers from Glacier Peak and Mount St. Helens in late-glacial lake sediments of eastern Washington.

Late-Quaternary volcanic eruptions, wild fires, and **fires** set by the prehistoric inhabitants no doubt influenced vegetation history. **In** extreme situations the ecological impacts of past volcanic activity are obvious. Yet, the immediate effects, and long-term consequences of any prehistoric catastrophe, even the eruption of Mount Mazama (Crater Lake, Oregon) are not always discernible. One cannot say exactly how most vegetation of the northwestern U.S. would be different if Mount Mazama, Mount St Helens, or Glacier Peak had been inactive in the last 20,000 years, nor in what ways the magnitude and timing of their eruptions determined or altered the course of Holocene vegetational "development." Certainly, pumice soils of east central Oregon favor lodgepole pine. For the Yellowstone National Park region, **Whitlock** (1993) also ascribes differing vegetational histories to volcanic soils.

Effects of an **ashfall** on terrestrial or aquatic ecosystems depends on thickness of primary and secondary tephra, season and duration of the **ashfall**, and time separating recurrent eruptions. Therefore, estimates of the depositional chronology of **ashfalls** is essential to evaluating their influences (**Mehring** and others 1977b; Blinman and others 1979). Sagebrush pollen **from** Mazama tephra at Lost Trail Pass, Montana (fig. 4), suggested an autumn **ashfall**. A graded bed of clean tephra with little pollen and lack of seasonal indicators implies that Glacier Peak tephra fell at Sheep Mountain Bog, Montana, when the lake was ice-free and probably in late summer (Mehring and others 1984).

## **Ecoregions**

For common comparison this review is organized around ecoregions even though their boundaries and characteristics have no doubt changed with time. The term Ecoregion was introduced by Crowley (1967) to refer to an area with relatively homogenous ecological systems. The notion of ecoregions is further developed by Omernik (1987) as “regions of relative homogeneity in ecological systems or in relationships between organisms and their **environment.**” Ecoregions are meant to provide a framework for conceptualizing “regional representativeness” and a range of ecological conditions or qualities that are “realistically attainable” (Omernik and Galant 1986). Combinations of vegetation, physiography, climate, soils, and geology delineate ecoregions from each other. Additionally, ecoregions have descriptive titles and general characteristics based on land surface form, potential natural vegetation, land use, and soils (Bailey and others 1994).

Six of Bailey’s ecoregions occur in the study area (fig. 6). These include:

1. Northern Rocky Mountains Forest-Steppe--Coniferous Forest--Alpine Meadow Province
2. Middle Rocky Mountains Steppe--Coniferous Forest--Alpine Meadow Province
3. Southern Rocky Mountains Steppe--Open Woodland--Coniferous Forest--Alpine Meadow Province
4. Inter-mountain Semi-Desert Province
5. Great Plains-Palouse Dry Steppe Province
6. Cascade Mixed Forest--Coniferous Forest--Alpine Meadow Province.

## **PALEOVEGETATION AND CLIMATE**

### **Vegetation and Climate West of the Cascades**

Because the Northwest’s climatic patterns are shaped over the Pacific Ocean, primary features of climate-induced vegetation history are related on both sides of the Cascades. With few exceptions late **Quaternary** fossil plant sequences published for the region east of the Cascades are younger than 13,000 **B.P.** West of the Cascades pollen and macrofossil records are old enough to

be useful in evaluating full-glacial and interstadial conditions. Also, they frequently show more detail in the number of tree species and in the order of their arrivals or departures. In the Puget Trough, for instance, fossil sequences reveal vegetation of the coldest episodes south of continental glaciers as well as the northward progress of developing postglacial forests. By **contrast**, sites to the east are scattered and forest is discontinuous at low to mid elevations (Franklin and Dyrness 1973, **fig. 27**). There, causes of latitudinal or elevational differences **in** species' ranges or associations must **be** evaluated by details of specific--often geographically and ecologically isolated--data and by recognizing common patterns in fossil records from vastly different present-day plant communities.

The following description highlights the main features of western Washington's and Oregon's Quaternary vegetation history to provide a broader chronologic and paleoclimatic context. The emphasis follows **Whitlock** (1992) and Sea and **Whitlock** (1995) who reviewed and explained the fossil records with reference to climate models (COHMAP Members 1988; Thompson and others 1993). Also, Worona and **Whitlock** (1995) describe a **40,000-year** pollen and plant macrofossil sequence **from** the central Coast Range, Oregon.

During the last glaciation (20,000-16,000 B.P.) continental ice sheets cooled the middle latitudes while winter storm tracks shifted south leaving the Northwest cold and dry. Even the western Olympic Peninsula lowlands sported vegetation resembling today's cool maritime subalpine parkland of the Olympic Mountains. In the lowlands scattered trees included spruce, pines, mountain hemlock (*Tsuga mertensiuna*) and **western hemlock** (*T. heterophylla*). Treeless alpine-like **communities** occupied glacial margins and exposed sites.

Dominance of grass, sedge, and sagebrush pollen with macrofossils of **Engelmann** spruce (*Picea engelmannii*) suggest that the Puget Trough lowlands sustained open alpine parklands (fig. 7). **Precipitation** was much reduced and mean annual temperatures were perhaps 5-7°C cooler than today. The most likely modern analog is found east of the Cascades in the alpine reaches of the northern Rocky Mountains. Fossil pollen, plant macrofossils and beetle remains suggest that climatic fluctuations within the period of maximum glaciation (24,000-16,000 B.P.) brought an

episode of relatively warm moist conditions to the Fraser-Puget Lowlands 19,000- 18,000 B.P. Thus, Mathewes (1991a) questions notions of continuous cold and dry glacial climates along the coast of Washington and British Columbia.

With a warming sea-surface after 15,000 B.P., an ameliorated coastal climate enhanced the success of spruce, alder and western hemlock. Between 14,000-10,000 B.P., deglaciation and appearance of temperate taxa marked the late-glacial to Holocene transition. Early in this period Cascade treelines 500-1000 m lower than present suggest mean annual temperatures at least 3°C below those of today. Precipitation increased and annual temperatures rose to near modern values by about 10,000 B.P. New species, forming transitional communities without modern analogs, filled freshly deglaciated terrain of the Puget Trough. Lodgepole pine (*Pinus contorta*), the most successful invader, was soon joined by alders, spruce, Douglas-fir (*Pseudotsuga menziesii*), and western hemlock. Similarly, just south of the ice sheet, temperate low elevation taxa first mixed with, then succeeded montane species.

Early Holocene introduction of xerothermic communities began about 10,000 B.P. when greater summer radiation brought warmer drier summers to the Northwest. Whitlock and Bartlein (1993) suggest that summer dry areas of today were even drier about 9000 B.P., while summer wet areas (e.g., the Southwest U.S. watered by the “Mexican monsoon”) were wetter than today (fig. 1). Their notions for the Northwest find support in numerous records from both sides of the Cascades. For example, between 10,000 and 5000 B.P. the Olympic Peninsula and Puget Trough fossil pollen sequences show more Douglas-fir and bracken fern than today. Also, local prairies expanded, and sediments of this period contain abundant charcoal that attests to the increased occurrence of wildfires. In the southern Puget Trough, Washington, the late-glacial forest gave way to open forest or savanna similar to the historic vegetation of the Willamette Valley, Oregon. In all, the early and mid-Holocene paleovegetation implies drier summers. Under the influence of higher temperatures Cascades montane species were displaced upslope by 250-500 m.

After about 5400 B.P. increasing precipitation and rising water tables were apparent again on both sides of the Cascades. In western Washington, mesophytic taxa such as western redcedar

(*Thuja plicata*), western white pine (*Pinus monticola*), and western hemlock assumed more importance; prairies became smaller and modern plant associations began to emerge.

These highlights of vegetation over the last 20,000 years (Whitlock 1992) portray the major climatic episodes and their influences. These generalities, however, often come from 'detailed data illustrating the continual short, sharp climatic shifts that, directly (e.g., soil moisture) or indirectly (e.g., fire and disease), produced rapid changes in the Northwest's vegetation.

Fossil records of the Northwest's late Quaternary vegetation have been the subject of sporadic study since the 1930s when Henry P. Hansen (1947) pioneered pollen analysis in Washington, **Oregon, Idaho**, and Montana. His studies proved that the Northwest's vegetation had been dynamic throughout the late Quaternary in response to climate, volcanic eruptions and fire. Over the past decade several reviewers have synthesized the late Quaternary fossil pollen and plant macrofossils of the region and have discussed the chronologies and causes of vegetation change (Barnosky 1985; Barnosky and others 1987; Chatters, in press; Hebda 1995; Heusser 1985; Johnson and others 1994; Mehringer 1985; Thompson and others 1993; Whitlock 1992).

## **Southern British Columbia**

Any review of the vegetation and climate of the U.S. portion of the Columbia River drainage would be incomplete without mention of late Quaternary conditions in adjacent British Columbia--home waters for the Columbia River since it was freed from the glacial grip. Fossil pollen and plant **macrofossils** indicate a similar three-part pattern of Holocene climatic states for the southern interior, central interior plateau, the Columbia Mountains, and the southern Canadian Rockies (Hebda 1995).

1. Following a brief treeless interlude with wasting of continental and mountain glaciers, conditions warmer and dryer than at present ("xerothermic" climate) prevailed from about 10,000 to 7,000 **B.P.**; mean annual temperatures could have risen by as much as 2-4<sup>o</sup>C above modern values. At first sagebrush and grasses formed open **xeric** communities in the southern interior. Precipitation began to increase toward the end of this period.

2. Still warm, but moist (“mesothermic”), conditions marked the next 3000 years (7000 to 4000 B.P.). Temperatures ranged from modern values to perhaps 1-2<sup>o</sup>C warmer, but precipitation increased throughout southern British Columbia--especially from 7000-6000 B.P. During this time trees descended below their early Holocene lower limits; Douglas-fir invaded sagebrush steppe. Western hemlock (followed by western redcedar) expanded, lake levels rose, and charcoal content of lake cores declined with changing fire regimes.
3. Advancing glaciers, expanding bogs, descending treelines, and increasing success of plants favoring cool moist environments signaled the shift to modern climate conditions perhaps beginning at 5400 B.P., but apparent everywhere by 4000 B.P. Modern vegetation emerged between 4000 and 2000 B.P. and, of course, continued to adjust to short, sharp climatic shifts and related disturbances.

## **Northern Rocky Mountains Ecoregion**

The Okanogan Highlands occur in the northern portion of this ecoregion and exhibit a varied topography including glacial scoured mountains with lakes and rivers plus both broad and narrow valleys. **Xeric** big sagebrush (*Artemisia tridentata*) and ponderosa pine communities occur west of the Columbia River. Subalpine fir occupies the cold, higher elevations. More **mesic** sites characterize the eastern portion of the area as evidenced by grand-fir, western hemlock western redcedar, white pine and western larch. The climate is maritime influenced; June and July are wet months.

The **Flathead** Valley, where glaciers were the major landscaping force, dominates the central portion of this ecoregion. Major conifer species include western larch, Douglas-fir and ponderosa pine forests along with western hemlock, western **redcedar** and grand fir. The deeply dissected Bitterroot Mountains occupy the southern part of the ecoregion .

**Mack** and others (1978a, b, c, d; 1979; 1983) studied several sites, from the San **Po**il to Priest River valleys of northeastern Washington, northern Idaho, and adjacent Montana; figure 8 summarizes some of their results. As a group, these sites along with Williams Lake Fen and



Goose Lake, Washington, show the local sequence of forest history and also suggest regional trends:

1. Sediments dating to at least 12,000 B.P. show an initial treeless episode dominated by pollen of sagebrush and grass. Spruce, fir, lodgepole, and other pines arrive before the fall of Glacier Peak tephra (11,250 B.P.).
2. Mixed conifers, sometimes with birch, **dominate** for the next 1000 years or so.
3. By 10,000 B.P. grass or sagebrush begin to assume dominance at lower elevations. To the east in northern Idaho and adjacent Montana, larch and Douglas-fir, along with diploxylon pine pollen (lodgepole or ponderosa), and small but persistent percentages of grass and sagebrush pollen indicate predominance of widespread steppe and dry interior forests.
4. By 4000 B.P., as sagebrush steppe retreated toward eastern Washington's arid central core, the forest fringe (ponderosa and lodgepole pine, and larch or Douglas-fir) also had begun to advance west and south, and perhaps downslope from the Cascades as well. Spruce and fir pollen increased at sites within forested areas.
5. Between 2500 and 1000 B.P. fossil counterparts of modern forest became apparent at most sites and eastern Washington's grasslands finally began to achieve their historic importance.

The Mountains of northern Idaho and adjacent Washington, Montana and British Columbia receive abundant snow fall and spring rain (Thompson and others 1993, fig. 18.3). As a result, they support low to mid-elevation moist maritime forests (fig. 9) of western hemlock, western redcedar, western white pine, grand fir, and yew which are more widely associated and distributed nearer the coast. These interior disjunct distributions, by themselves, raise many questions about plant geography; but, the apparent late Holocene arrival of western hemlock (*Tsuga heterophylla*) and, later, western redcedar (*Thuja plicata*) makes the inquiry even more significant (Hansen 1939; Mack and others 1978a).

When did hemlock, cedar and associates typifying present habitat types arrive and spread, in what order, and by what route (Cooper and others 1991)? What role did stand-replacing fires, soils, nutrient cycling, and succession (Turner and Franz 1985a, b) have in their initial success and

regeneration (Moeur 1992)? When did they become abundant? All in all, what is the historical process by which the interior moist maritime forest communities, with such long-lived trees, became so successful in apparently less than 3000 years? These are just a few questions whose answers lie in macrofossils and pollen from the lakes and bogs of this region.

Before 12,000 B.P., as glacial ice finally wasted in western Washington, western hemlock already was present in the Puget Lowlands and Olympic Peninsula. By 10,000 B.P. it had reached the northwestern flank of the Cascade Mountains at Kirk Lake (190 m), the northeastern Olympic Peninsula in the rain shadow of the Olympic Mountains at the **Manis** Mastodon Site, near Sequim, Washington, and southwestern British Columbia (fig. 10). From there it radiated rapidly northward along the coast to Alaska (Cwynar 1993; Peteet 1991). It more slowly made its way to the interior along the Fraser River, arriving near **Fishblue** Lake about 8,000 B.P. Finally, by 7000 B.P., it had reached the northeasterly (interior) extent of its southwestern British Columbia range near Horseshoe Lake. For now, the trail to the interior ends there because lake and bog deposits between the coastal and interior distributions thus far have failed to reveal western hemlock's track through southern British Columbia and northern Washington. The continual abundance of western hemlock within its coastal range is revealed throughout the Holocene by occasional pollen grains transported downwind to sites in Washington, Idaho and Montana. There are, however, no fossil records--pollen or macrofossils--suggesting a southern interior migration route **from** refugia south of the ice sheet.

As shown in Figure 10, the western hemlock trail picks up again, about 3000 (4500 B.P.) years later, at Lower Little **Slocan** Lake, southeastern British Columbia. By 2500 B.P. it is recognized at Big Meadow, **northeastern Washington**. Though the dating is not clear and hemlock pollen is sparse, it probably arrived in northwestern Montana in the vicinity of Tepee Lake and **McKillop** Creek Pond by 2000 B.P. and further south at Smeads Bench Bog about 1500 BP. Pollen profiles from eastern British Columbia show western hemlock's presence in the vicinity of Bluebird and Pemberton Hill lakes about 3000 B.P. and perhaps even earlier (4000-3000 B.P.) at Dunbar Valley. These  $^{14}\text{C}$  dates are all **from** organic matter in cores where percentages of western

hemlock pollen suggest its local or regional presence. Direct accelerator  $^{14}\text{C}$  dates on western hemlock pollen and on a needle confirm its arrival at Hager Pond, Idaho, about 2400 years ago and its greatest abundance about 800 years later (fig. 11; Mehringer and Van de Water, in prep.). Furthermore, preliminary analyses indicate that western **redcedar** (*Thuja plicata*) arrived after western hemlock had become well established near Hager Pond.

Much remains to be learned about the Holocene dispersal and relationships of the interior moist maritime forest species. It is not certain that these “arrival” ages for western hemlock in northern Idaho and southeastern British Columbia represent a chronology of initial migration. Perhaps, western hemlock had been present, but rare, in some part(s) of its eastern range and multiplied rapidly with onset of cool moist conditions of the late Holocene. Still, examples from studies of postglacial hemlock (*Tsuga canadensis*) migration to the Great Lakes region, cedar expansion north along the British Columbia coast, and interior lodgepole pine’s journey to the central Yukon Territory give hope for eventual understandings (Davis 1989; Davis and others 1986; Hebda and Mathewes 1984, 1991b; MacDonald and Cwynar 1985).

## **Middle Rocky Mountains Ecoregion**

The central portion of this ecoregion is a part of the Idaho Batholith where glaciers **sculpted** high mountain cirques and broad, deeply incised valleys. Forest **communities** vary with **elevation**--ponderosa pine to grand fir--Douglas-fir to spruce-fir forest. Glaciated landforms of the Bitterroot Mountains occupy the northcentral portion of the ecoregion. Douglas-fir and western ponderosa pine forests dominate the lower slopes, whereas higher elevations support mixed conifers including Douglas-fir, subalpine fir and, in the northern part, western larch.

The Big Belt Mountains, west-central Montana, rise above the surrounding plains and rolling hills on the eastern edge of this ecoregion. Wheatgrasses, fescues, **grama** and needlegrass dominate the foothill prairies. Douglas-fir and ponderosa pine forests cover the major mountain ranges. The Beaverhead Mountains in southwestern Montana are an area of complex geology with broad alluvial and **fluvial** valleys and high glacial alpine ridges and cirques. The vegetation is

predominately sagebrush steppe with big sagebrush, fescues, wheatgrasses and needlegrass. At middle elevations conifers are primarily Douglas-fir and lodgepole pine with some limber pine on droughty sites. The Challis Volcanic Area of central Idaho includes numerous mountain ranges of volcanic rocks. The potential natural vegetation is a mixture of forests and sagebrush steppe. Forest types include western spruce-fir, Douglas-fir and lodgepole pine. Higher elevations hold whitebark pine and subalpine fir.

The Blue Mountains, northeast Oregon, a wide, dissected uplifted plateau with mesas and buttes, occupy the southwestern portion of this ecoregion. More moist mid-elevation sites support grand fir, Douglas-fir, western larch, lodgepole pine and ponderosa pine. Drier sites sustain Great Basin sagebrush, western juniper steppe woodlands and grasslands.

In the Bitterroot Range and in smaller ranges, timing and direction of changing vegetation represented in pollen profiles suggest that forest species were reacting similarly to the climatic patterns that influenced the vegetation of northeastern Washington and southeastern British Columbia. Before the fall of Glacier Peak tephra (11,250 B.P.), subalpine conifers began to fill higher elevations of the Bitterroot Range near Lost Trail Pass Bog (2152 m) where they grow today (fig. 12). Then, before the fall of **Mazama** tephra (about 6850 B.P.), grass and sagebrush pollen percentages, along with pollen of the warmth-requiring Douglas-fir and lodgepole pine replaced the more cold-tolerant whitebark pine. Up-slope migration and persistence of Douglas-fir marks a period of undoubted mid-Holocene warming; yet, an unbroken sediment record indicates that the small pond at Lost Trail Pass did not go dry. By 5000 B.P., perhaps for the first time in 4000 years, Douglas-fir began to lose dominance to whitebark and lodgepole pines, and retreated to warmer downslope positions. Since that time and especially after 1750 B.P., the vegetation has become similar to the present.

Varying charcoal accumulation rates, pollen/charcoal ratios, and washed-in charcoal layers may reveal the history of fire in forest and steppe (Clark and Royall 1995; MacDonald and others 1991). The observation that charcoal was unusually abundant over the past 2,000 years or so near Lost Trail Pass led to questions about the late Holocene relationships between reconstructed

vegetation, charcoal layers, and charcoal and pollen abundance in lake sediments. In this case, three widely separated sites displayed the same broad chronological patterns of changing charcoal abundance over the last 7000 years or more (Mehring 1985, table 3). All sites showed the largest charcoal/pollen ratios during the last 1000 years. Perhaps this superabundance of charcoal is related to increasing populations of Native Americans who commonly used fire as a management tool. Many early observers (Barrett and Amo 1982; Gruell 1985; Leiberg 1900, p. 314-316) stress the great extent of fires before Indian populations plummeted as a result of small pox, measles, and conflicts.

**Marys Pond** (1753 m) formed behind a terminal moraine in a steep-sided canyon of the Bitterroot Range east of **Lolo Pass** on the Idaho-Montana border. It is surrounded by western larch and lodgepole pine--legacy of several fires in the last 500 years. Exploratory pollen counts indicate that conifers grew at this site when Glacier Peak erupted. The pollen record since eruption of Mount **Mazama** shares general characteristics with the Sheep Mountain Bog profile (fig. 13). A brief interval of abundant nonarborescent pollen (less pine, spruce and fir pollen) follows Mazama tephra. Otherwise diploxylon pine (lodgepole or ponderosa) and **Douglas-fir** (or larch) pollen dominate the mid-Holocene. By 4000 B.P. spruce and fir pollen together achieved dominance over Douglas-fir. By 2000 B.P. modern forest vegetation began to emerge. Microscopic charcoal counts suggest many low intensity fires throughout the last 6800 radiocarbon years with larger or more frequent fires around 3300 and 4000 B.P.

Sheep Mountain Bog (1920 m) 18 km northeast of Missoula, Montana, lies in **Douglas-fir**/subalpine fir, near the upper limit of larch (*Larix occidentalis*) and 200 m above the usual upper limit of ponderosa pine (*Pinus ponderosa*) on south-facing slopes. Mixed forests with subalpine fir (*Abies lasiocarpa*) and **Engelmann** spruce (*Picea engelmannii*) surround the bog. Haploxylon and diploxylon pine pollen are represented by macrofossils of whitebark (*Pinus albicaulis*) and lodgepole (*Pinus contorta*) pines.

According to fossil conifer needles and cones, Douglas-fir (*Pseudotsuga menziesii*) has grown near this site since at least 10,000 B.P. Its largest pollen frequencies occur between 9700

and 4000 B.P. along with pollen and macrofossils of lodgepole pine and pollen of *Physocarpus*, all of which suggests a dry, open mid-Holocene forest. Douglas-fir is common but not abundant here today. Haploxylon pine pollen (whitebark pine) is most abundant, along with macrofossils and pollen of spruce and fir, only between 11,000 and 10,000 B.P. This may reflect the short-lived cooling of the Younger Dryas episode. Before 4000 B.P. relative importance of Douglas-fir pollen diminished in relation to spruce and fir pollen which showed marked relative increases thereafter (fig. 13).

Fire frequency, as evidenced by the charcoal/pollen ratio at Sheep Mountain Bog, Montana, declined gradually between 7000 and 2000 B.P., then increased to produce a superabundance of microscopic charcoal during the past 1000 years. These peaks of charcoal in bog sediments proved comparable with tree fire-scar studies and historic records (Hemphill 1983). A remarkable record of early and middle Holocene fires comes from deposits of Sheep Mountain Bog.

The sediment sequence at Sheep Mountain Bog began nearly 13,000 radiocarbon years ago with till left by the wasting glacier that had carved the >12 m-deep depression. From that time the depression held a pond that accumulated primary and redeposited Glacier Peak and Mazama tephra, and distinct layers of charcoal. The charcoal-rich deposits were graded beds that typically formed couplets. Sand, coarse silt, and charcoal made up the lower black layer of each couplet, whereas the upper layer consisted of gray wood ash and clayey silt. In all, 167 layers (most of them couplets) were measured with a dissecting microscope (fig. 14). Between 10,600 and 4000 B.P. erosion after fire accounted for at least 23.5 percent of sediments deposited in the basin.

The charcoal, wood ash, and sediment layers resulted from wash-in to the lake after fire and, therefore, furnish a record of fires within the basin. The thickest layers most likely result from severe erosion after high intensity duff-removing fires, whereas the thinner layers represent low intensity fires. By 4000 B.P. a filtering fringe of sedge mat surrounded the gradually filling pond and slope-wash deposits no longer reached the coring site near the center of the basin. Distribution of the measured layers reveals that there was no single "natural" fire regime in the 6400 years represented by charcoal layers; rather three patterns are evident:

1. Many low intensity fires (around 6000 B.P.).
2. Many low intensity small fires with occasional high intensity fires (around 8000 B.P.).
3. Few low intensity fires followed by stand replacing **fires** (around 5400 to 4200 B.P.).

The first charcoal layer was deposited about 11,150 B.P. when lodgepole pine dominated scattered woodlands. Four narrow (0.5-2.5 mm) **laminae** form the first cluster of charcoal layers about 10,600 B.P. At this time the site supported scattered whitebark pine with spruce and fir while sagebrush occupied openings.

Between 10,200 and 9600 B.P. the number and intensity of fires suggest a different climatic regime. By the end of this period lodgepole pine and Douglas-fir largely replaced whitebark pine, spruce and fir. This first period of abundant charcoal at Sheep Mountain Bog corresponds to the time of decreasing effective moisture, of demise of conifer woodlands in east central Washington, and the largest pre-Mazama charcoal values at Williams Lake Fen. Fire frequency increased again around 8000 B.P. when Douglas-fir pollen is most abundant and pine pollen (primarily lodgepole pine) percentages are much reduced.

From 5400 to 4200 B.P. there were many low intensity fires and, judging by thickness of charcoal layers, the three most severe fires. At Sheep Mountain Bog this period saw increasing percentages of pine, spruce and fir pollen with declining Douglas-fir pollen. Perhaps there were fewer of the small or low intensity fires that marked the preceding 1000 years. Alternatively, more ground cover may have supported occasional low intensity fires, but inhibited slope wash. In any case, during this **1200-year** period the basin of Sheep Mountain Bog apparently experienced three conflagrations more intense than fires of the earlier Holocene or any of the last 1000 years or more (Hemphill 1983).

Figure 15 compares the post-Mazama importance of Douglas-fir pollen at Sheep Mountain Bog, Marys Pond and Lost Trail Pass Bog, western Montana. The diagrams illustrate **mid-** Holocene importance of Douglas-fir pollen, relative to fir and spruce, and its decline after about 5400-4000 B.P. These differences are even more substantial when one considers that the pollen profiles are dominated by pine pollen and that Douglas-fir and larch are poor pollen producers, as

compared to fir and spruce. In the broadest sense, since 5400-4000 B.P., the coniferous forests of the Bitterroot Mountains generally have been influenced by relatively higher effective moisture, whereas the previous 4000-5000 years saw more open forests suggestive of a drier or warmer climate (less effective moisture).

Comparison of the Marys Pond (1753 m) and Lost Trail Pass (2152 m) pollen curves (fig. 15) illustrates the apparent chronological differences in the late Holocene Douglas-fir decline. Lost Trail Pass, today dominated by subalpine conifers, shows continuous presence of spruce pollen with decrease (downward displacement) of Douglas-fir beginning about 5400 B.P. Marys Pond, on the other hand, indicates continuous mid-Holocene importance of Douglas-fir (or larch), especially before 4000 B.P. along with the gradual relative increase in spruce pollen--beginning before 5000 B.P. Spruce pollen increases significantly between 4000 and 2000 B.P. Thus, pollen records from both sites reflect time transgressive vegetation changes reflecting their relative positions in today's vegetation zones of the Bitterroot Mountains.

At the lower edge of coniferous forest, Blue Lake, near Lewiston, Idaho (1035 m), is now surrounded by Douglas-fir and ponderosa pine adjacent to **Palouse** grassland; varying abundance of pollen and macrofossils of these two trees suggests important variations in effective moisture over the past 4300 years. According to Smith (1983), Douglas-fir dominated during a relatively warm-moist interval **from 4300 to 4000 B.P.**; mixed Douglas-fir, lodgepole and ponderosa pine followed during a cooler period from about 4000 to 3000 B.P. Then, conifer pollen decreased from 3000 to 1700 B.P. and open ponderosa pine parkland persisted until 1000 B.P. when vegetation similar to present day emerged.

At Blue Lake, severe fires resulting in distinctive bands of washed-in charcoal, burned through Douglas-fir forests about once every hundred years between 4300 and 4000 B.P. Only two severe fires in mixed conifer forests left charcoal layers between 4000 and 3100 B.P., but such fires were nearly twice as common between 3100 and 1700 B.P. From 1700-1000 **B.P.** severe fires averaged only one in 175 years. In contrast, light surface fires, leaving no charcoal layers in the lake sediments but producing abundant microscopic charcoal, characterized the last



700 years. During this time (A.D. 1300-1900) frequent low intensity fires left multiple fire scars on surviving old growth trees (S. Arno, pers. comm.).

Late Quaternary vegetation of the Blue Mountain region of northeast Oregon (fig. 16) and adjacent Washington (including the **Wallowa**, Ochoco, Strawberry-Aldrich; Greenhorn, and **Elkhorn** ranges) remains largely unknown. A combination of related factors including fire suppression, some logging practices, and insect **infestations** have left the diverse forests (Johnson and **Clausnitzer** 1992) of the Blue Mountains teetering on the edge of an ecological disaster (**Mutch** and others 1993). The only two published pollen records from this entire **region come from** a few exploratory **samples** of Henry P. Hansen (1943) from the Anthony Lakes area. The other, from the Silves. Valley (Craddock Meadow, 1630'm) spans most of the Holocene. But, only 23 samples were analyzed and pollen was poorly preserved. Ponderosa pine, western juniper, and big sagebrush inter-finger in the **Craddock** Meadows area today. Yet, sagebrush was the dominant pollen from about 9000 B.P. until 2000 years or so after the **Mazama ashfall**. Thereafter pine pollen values rose to 60 percent suggesting presence of ponderosa pine. Two nearby **woodrat middens**, dating to 1280 (juniper twigs) and 300 **B.P.**, held species now present at the site (Reid and others 1989).

Cursory examination of two Blue Mountain sites indicate the potential for detailing forest history for northeastern Oregon. Cores from Twin Lakes in the **Wallowa** Mountains and Lost Lake near Dale, Oregon, show well preserved pollen and macrofossils in post-Mazama sediments. For instance, Lost Lake, in mixed Douglas-fir (*Pseudotsuga menziesii*) and grand fir (*Abies grandis*) forest, holds abundant charred Douglas-fir cones and needles in redeposited **Mazama** tephra. Pollen samples from **3000-year-old** sediment show mountain hemlock (*Tsuga mertensiana*) pollen values indicating a nearby source at that time, whereas a younger tephra layer blankets a **layer** of larch needles (*Larix occidentalis*). Larch grows nearby today, but the nearest mountain hemlock occur about 110 kilometers to the northeast in the **Wallowa** Mountains (Johnson and Simon 1987).

## Southern Rocky Mountain Ecoregion

Many significant paleovegetation sites in western Wyoming lie outside, but adjacent to, the Columbia River drainage; nevertheless, their proximity, numbers and dating make them important for understanding climate, **fire** and vegetation history throughout the Rocky Mountains and interior Northwest. Parts of a 140,000 year-old paleovegetation sequence from Yellowstone National Park show repeated plant responses to glacial and interstadial climates (Baker 1986). Likewise, Grays Lake (Beiswenger 1991) and Swan Lake, southeastern Idaho (Bright 1966), are key to understanding the late Pleistocene persistence or interaction between sagebrush steppe and conifer woodlands in the Northern Great Basin and Snake River Plain. The record from the Wind River Mountains, Wyoming, is particularly significant because Rapid Lake (3134 m) lies at upper **treeline** where vegetation is especially sensitive to temperature and because its sediments hold the history of climates associated with the classic glacial sequence of the Temple Lake Valley (Fall and others 1995).

Between 12,000 and 11,500 B.P., as alpine glaciers wasted in the Temple Lake Valley, Rapid Lake began accumulating herb and shrub pollen from initial alpine tundra and sagebrush steppe communities. From about 11,300 to 3000 B.P. abundant spruce, subalpine **fir**, whitebark pine, and some lodgepole pine pollen indicate a mixed subalpine forest. Throughout most of the Holocene warmer temperatures allowed **treeline** to move upward as much as 150 m. Under the influence of colder conditions after 3000 B.P. spruce and fir moved downslope to the position of the historic **treeline** and open whitebark pine parkland replaced the mixed conifer forest that had surrounded Rapid Lake for 7500 years.

Studies **from** the Yellowstone-Grand Teton area, taken as a whole, show not only the pace and extent of vegetation changes over the last 14,000 years or so, but also illustrate controls of paleovegetation. For example, **Whitlock** (1993) ascribes disparate vegetational histories to differing soils and to a climatic boundary. She concluded that andesitic and nonvolcanic geologic substrates favored success of **late-glacial** mixed forests of pine, spruce and fir, whereas these trees remained sparse on substrates producing **rhyolitic** soils where closed lodgepole pine forests have

persisted for 5000 years. Pollen profiles from southern and northern parts of the Yellowstone region revealed that areas now experiencing dry summers and wet winters were even dryer than the present in the early Holocene, whereas maximum aridity came to today's summer-wet areas in the mid-Holocene. At least in the mountainous Yellowstone region, it appears that the boundary between these systems persisted through the Holocene (Whitlock and Bartlein 1993; **Whitlock** and others 1995).

As a group, paleovegetation studies of the Yellowstone-Teton area illustrate the local sequence of **altitudinal** shifts in species ranges (fig. 12) as influenced by climatic, edaphic and biotic factors, and also show regional trends with deglaciation and Holocene warming--from alpine meadow, to subalpine parkland, to forest.

1. Before 11,500 B.P. pollen in lake sediments reflect abundance of sagebrush, grass, sedge, and alpine herb **taxa**, along with additional nonarboreal pollen from shrubs of common juniper, willow and birch; **Treeline** was at least 600 m below today's and temperatures were **4-5°C** cooler.
2. Increasing values of spruce pollen mark the next 1000 years or so, as spruce invaded shrub, steppe and meadow communities. The open spruce parkland of 11,500-10,500 B.P. may not have an analog in the modern vegetation of this area.
3. By 10,500 B.P. spruce, subalpine **fir**, whitebark pine, and sometimes lodgepole pine, formed subalpine forest communities.
4. Around 10,000 B.P. Diploxylon pine pollen of the lodgepole type (*P. contorta*)--today's dominant conifer in the region--was becoming abundant as spruce, fir and Haploxylon pine pollen (whitebark and limber pine) percentages declined.
5. From 9500-5000 **B.P.**, the time of maximum warmth, lodgepole pines were the predominate tree which formed closed forest stands. All trees moved **upslope** from their late-glacial (and present) lower and upper elevational limits and, in some areas, Douglas-fir and aspen grew with lodgepole pine. These three are fire-adapted species, that would have been favored by the greater fire frequency likely to have come with higher temperatures.

6. During the last 5000 years or so, where soils permitted, cooler wetter conditions have favored spruce, **fir** and whitebark pine over lodgepole pine and Douglas-fir. These subalpine trees returned to the downslope positions they had not held since the late-glacial.

## **Intermountain Semi-Desert Ecoregion**

Block-faulting of basalt flows has created a deeply dissected landscape in parts of the Snake River Plain and southeastern Oregon. Lava flows, cinder cones and maars, steep lava ridges, and tephra deposits attest to recent volcanism from southeastern Idaho (Craters of the Moon) to central Oregon (Jordan, Diamond and **Newberry** craters). Though the western edge of this ecoregion generally supports ponderosa forests, pumice soils of east central Oregon favor lodgepole pine. Lodgepole pine also often dominates frost prone low lying valleys and basins where ponderosa pine dominates the less **frosty** uplands. These forests blend to the east with mixed woodland of western juniper, sagebrush, bitterbrush and bunch grasses. Sagebrush and bunch grasses dominate in eastern Oregon and adjacent Idaho where western juniper is often limited to rocky outcrops.

Flat-lying basalt flows supporting sagebrush, shrub-steppe and grassland, characterize the Columbia Basin in the northwestern part of this ecoregion. Dissected loess covered plains and rolling **Palouse** hills sustain grassland of bluebunch wheatgrass and Idaho fescue. Idaho fescue with snowberry and roses commonly characterize the Palouse's shrub-steppe. Ponderosa pine forest and woodland border the **Palouse** on the east.

Although the northwestern U.S. is noted for its remarkably diverse conifer forests, east of the Cascades in Washington and Oregon, and on the Snake River Plain, Idaho, steppe covers more area than **all** forest types (**Franklin** and Dyrness 1973, fig. 27). Grassy communities extend upward from shadscale associates in the hot, dry interior basins, through sagebrush and western juniper woodlands, ponderosa pine parklands, the lush **Palouse** grasslands of eastern Washington, and mountain grasslands of southeastern Oregon. Between 20,000 and 4000 years ago steppe was even more widespread. In fact, fossil pollen or **woodrat midden** macrofossils from the Great

Basin (Thompson 1990; **Wigand** and Nowak 1992) to southeastern Idaho (Beiswenger 1991) to eastern Washington (Bamosky 1985) and Oregon (**Mehring** 1985) attest to expansion of steppe under cold continental conditions coincident with the last glaciation and to the continued importance of steppe with high temperatures of the early and mid-Holocene.

**Snake River Plain--From** 20,000 to 12,000 B.P., south of continental ice and below the mountain glaciers, cold steppe typified much of the northwestern interior of the U.S. Grays Lake, on the southeastern border of the Snake River Plain holds a long continuous pollen sequence that reflects the regional full-glacial importance of trees or sagebrush (Beiswenger 1991). There, about 30,000 **B.P.**, small increases in conifer pollen suggest a shift **from** cold, dry steppe to cold, slightly more moist full glacial conditions lasting until about 11,500 B.P. During this time conifers grew sparsely in foothill ravines and on rocky terrain. Otherwise, sagebrush step dominated low elevations on the Snake River Plain. Swan Lake, southern Idaho, sediments show abundant pine (probably from limber and lodgepole pines) and spruce pollen between 12,000 and 10,000 **B.P.**; thereafter the site supported sagebrush steppe as it does today.

Lennstrom's (1985) study of Holocene **woodrat middens** from the eastern Snake River Plain showed no significant differences in species present today and those recovered from **middens** dating to about 9960, 5800, and to times between 3200 and 1100 B.P. At about 5820, 2930 and 1150 B.P. Douglas-fir and limber pine grew where they grow today. In fact, these Holocene **middens** lacked **extralocal** species or loss of species that would indicate significant vegetation and climatic differences from today--even at 9960 B.P. Bright and Davis (1982) believed that vegetation zones of the Snake River Plain were fairly constant for 10,000 years; still, they inferred a slight **mid-Holocene** warming from pollen of Rattlesnake and Middle Butte caves. Davis and others (1986) concluded that low elevation sites on the Snake River Plain exhibited maximum aridity before deposition of Mazama tephra (the Northwest's early Holocene xerothermic), whereas high elevation sites such as Lake Cleveland and Mission Cross Bog, where moisture was less limiting, showed maximum warmth, and upward shift of vegetation zones about 6000 B.P.

**Eastern Washington--Carp** Lake, southwestern Columbia Basin, Washington, holds sediments spanning parts of the past 33,000 years (Barnosky 1985). Sagebrush and grass pollen dominate this record. Slightly more conifer pollen before 23,000 B.P. suggests temperate steppe, whereas full-glacial temperatures too cold for trees apparently produced cold sagebrush steppe. According to **Whitlock (1992)**, this sequence indicates colder and drier conditions **from** 23,000 to 10,000 B.P. with the lower **treeline** higher than today.

All other fossil pollen localities from eastern Washington are younger than Carp Lake because they lay in the paths of glaciers or **scabland** floods until after 13,000 B.P. Though the oldest localities date to **<13,000 B.P.**, without exception they show initial dominance of sagebrush and grass pollen. Sites now well within areas dominated by conifers continue to show large values of sagebrush, grass and chenopod **pollen** until some time between the fall of Mazama tephra (6856 B.P.) and about 4000 B.P. For the pre-Mazama period the most detailed analyses come from Williams Lake Fen.

At 635 m, Williams Lake Fen is located in a flood-formed plunge pool between Badger and Williams lakes 20 km south of Cheney, Washington. There, ponderosa pine meets *Artemisia tripartita-Festuca* and *Festuca-Rosa* communities that spread to the west and south (Daubenmire 1970). Initial coring and analyses established the site's potential and indicated grass and sagebrush dominance for most of the Holocene. Ponderosa pines were not important here until after 4000 B.P. (Nickmann 1979).

Recently, larger and deeper cores and additional analyses established details of late glacial tephrochronology and the sequence of vegetation that followed the last of the Lake Missoula floodwaters through the Cheney-Palouse scablands. These cores (Mehring, unpub.), dating from about 12,700 **B.P.**, bottomed in flood gravels at 10.83 m sediment depth. They hold two layers of Mazama tephra separated by 9 cm of lake deposits and five late glacial tephra--one from an unknown source (the 10,600 B.P. "Tawny Ash"), two from Glacier Peak (11,250 **B.P.**), and two (one above and one below the Glacier Peak tephra, fig. 5) from Mount St. Helens (Foit and

others 1993). A pre-Mazama pollen **profile and** selected ratios and percentages illustrate the fossil pollen sequence and suggest trends in past vegetation and climate (figs. 17,18).

Selected curves of 64 samples of pre-10,000 B.P. sediments **from** Williams Lake Fen reveal details of events, related through climate, to fluctuating lake levels, fire **frequency**, and vegetation. Fig. 19 shows the sedge and water-lily (Nymphaeaceae) pollen curves. Large values of these two categories of lake-edge plants indicate shallow water nearer the coring site. Their largest pre-Mazama values center on 9.85 m depth, corresponding to large charcoal concentrations. From figs. 19, 20 and other information, three trends are evident before 10,000 B.P.

1. Smallest pollen influx, largest percentages of spruce and birch pollen, and primary importance of sagebrush pollen suggests cold continental conditions from about 12,700 to 12,000 B.P.
2. From about 12,000 to 11,000 B.P. grass and sagebrush pollen vary whereas pine pollen generally increases.
3. The 11,000 to 10,000 B.P. period shows marked fluctuations and the largest conifer pollen values (perhaps indicating a Younger Dryas-like oscillation).

Following the second Glacier Peak **ashfall**, declining sagebrush and grass, and large pine pollen counts indicate significant and rapid expansion of conifers followed by their catastrophic decline. About 10,200 B.P., pine pollen falls from 60 to 14 percent of total terrestrial pollen and grass pollen attains values of 50-60 percent. The pine pollen peak, centered on 10,600 B.P., did not result **from** a single invader such as lodgepole pine because it includes both haploxyton (e.g., white pines such as limber and whitebark) and diploxyton types (such as lodgepole and ponderosa) as well as continued small numbers of spruce and fir pollen.

Perhaps reliable snow pack, **sufficient** to protect young conifers and shorten the period of deficient late summer-fall soil moisture, encouraged rapid expansion of mixed conifers at lower elevations in the eastern Cascade foothills and on favorable sites throughout eastern Washington. Whatever the case, climatic conditions that had brought success to conifers following the fall of Glacier Peak tephra changed suddenly and markedly.

Pollen of aquatic plants show that shrinking lakes accompanied diminishing conifers. For example, between 9.8 and 10 m (centered on 9.85 m), indicators of shallow warm water first appear in abundance. These include the only pre-Mazama concentrations of water-shield (*Brasenia*) pollen, the largest percentages of sedge and cat-tail pollen along with water-lily (*Nuphar*) pollen, sclerids and leaf hair bases. Large charcoal values indicate increased fire frequencies or intensities.

By about 10,000 **B.P.** grassland must have dominated more of eastern Washington than it does today. At Williams Lake Fen grassland apparently persisted for the next 3000 years with few differences by comparison with the preceding 3000 years. However, slight increases in pollen percentages of saltbushes, sagebrushes and other composites indicate probable up-wind expansion of more xeric steppe and newly exposed margins of shrinking lakes and ponds prior to eruption of Mount **Mazama**.

The shift from conifer to grass pollen dominance at Williams Lake Fen about 10,200 **B.P.** is perhaps the most striking change recorded in this region's history of postglacial vegetation. At Wildcat Lake (Mehringer, unpub.) and Goose Lake (Nickmann and Leopold 1985), the shift **from** conifer to grass pollen dominance was similarly dated and is as clearly revealed by close sampling. Hebda (1982) saw the same pattern in the southern interior of British Columbia where he reconstructed the post glacial vegetation sequence, emphasizing grassland, as follows:

1. 13,000-12,000 B.P., open pioneer treeless vegetation;
2. 12,000-10,000 B.P., mixed conifer forest invaded, established and declined;
3. 10,000-8000 B.P., xeric grassland maximum;
4. 8000-4500 **B.P.**, **mesic** grassland, extent somewhat reduced by Douglas-fir and ponderosa pine;
5. 4500-3000 **B.P.**, minimum grassland, **maximum** forest expansion into former grassland, and beginning of recent conditions of climate and vegetation.

From study of Wildcat Lake cores, Davis and others (1977) described effects of historic disturbance reflected in the changing pollen of aquatic, weedy, and exotic species, and response of



algae to erosion and organic enrichment. Blinman and others (1979) detailed the Mazama tephra sequence. Later, Wildcat Lake was cored again through five meters of water to a sediment depth of 27.39 m (Mehring, unpub.). Like Williams Lake Fen, the basal date is about 13,000 B.P. and Glacier Peak tephras allow exact correlation. Unlike sites further from eastern Washington's arid core, however, Wildcat Lake has remained in steppe since at least 10,000 B.P. Today it lies within the *Agropyron-Festuca* grassland bounded on the west by the warmer and drier *Artemisia-Festuca* and on the east by the more **mesic *Festuca-Rosa*** zone (Daubenmire 1970).

Ternary plots (fig. 21) of grass, sagebrush, and **saltbush (*Atriplex* and *Sarcobatus*)** separate various steppe communities as judged from plots of their fossil pollen over the last 5700 years. The same technique is helpful in distinguishing fossil pollen assemblages from Wildcat Lake (fig. 22), and together with other data leads to the following conclusions (fig. 23):

1. Pre-Mazama samples (9000 to 7000 B.P.) fall within a cluster in which both grass and sagebrush pollen are important. Sagebrush pollen values place these samples in big sagebrush (*Artemisia tridentata*) communities to the west.
2. Sagebrush communities, established locally before eruption of Mt. Mazama, expanded where soils permitted reaching their maximum eastward position >50 kilometers east of Wildcat Lake before 5400 B.P.
3. Sagebrush gradually gave way to grass between 4400 B.P. and 2400 B.P.
4. The relatively largest grass pollen values of the last 9000 years are not reached until within the last 2400 years.

**Eastern Oregon--During** the Holocene, middle elevations of Great Basin mountain ranges have supported juniper or pinyon-Juniper woodlands where subalpine conifers grew during **glacial-pluvial** period. The preliminary pollen record from Summer Lake, Oregon, spans nearly 100,000 years (**Wigand** and others 1995). It shows the response of forest, semi-arid woodlands, and steppe to Pleistocene climatic episodes and holds the promise of distinguishing the Holocene from

previous interstadial episodes, thereby broadening our perception of potential vegetational responses.

Steens Mountain, southeastern Oregon, is unusual in lacking a montane coniferous forest zone (McKenzie 1982) and is, therefore, ideal for study of changing steppe vegetation (fig. 16): There, as elsewhere at lower elevations **from** the northern Great Basin to the Columbia Basin, varying amounts of pollen produced by grasses, saltbushes, and sagebrushes are the primary clues to relative abundance of major steppe genera as influenced by climate (fig. 21).

Sites on Steens Mountain also show a long history of treeless vegetation with changing abundance of grass and sagebrush. At Fish Lake, sagebrush steppe followed retreating glaciers upward to an elevation of 2300 meters by 12,000 B.P. There, late glacial pollen spectra dominated by sagebrush and grass are distinguished **from** those of the Holocene by larger values of jumper (probably *Juniperus communis*) and pine pollen suggesting an early, but short-lived, up-wind source of conifer pollen.

Cores from Fish and Wildhorse lakes, in sagebrush and subalpine steppe, and from Diamond Pond, on the sagebrush-shadscale desert ecotone, are precisely correlated by six **volcanic** ashes deposited over the last 6850 radiocarbon years. Fossil pollen exhibits a general three-part division of Holocene vegetational change with differences in the timing of specific events at each site. These differences probably resulted from effects of temperature and precipitation at **various** elevations.

Greater abundance of sagebrush pollen in relation to grass pollen indicates relatively low effective moisture at Fish Lake (2250 m) between about 8700 and 4700 B.P. (fig. 24). The **mid-**Holocene episode of sagebrush pollen abundance began 1500 years earlier than the **temperature-**controlled upward expansion of sagebrush to Wildhorse Lake (2565 m) about 7200 B.P. **Also**, it ended at least 1000 years before grass again assumed dominance at Wildhorse Lake (about 3800 B.P.; fig. 24), marking the end of this prolonged but variable period of relatively higher temperatures and reduced snowpack (Mehring 1987).

About the same time (4000 B.P.) at Diamond Pond (Malheur Maar, 1265 meters), juniper and grass pollen percentages increase with declining values of chenopod in relation to sagebrush pollen (fig. 25). Radiocarbon-dated macrofossils of western juniper (*Juniperus occidentalis*) (fig. 26) from woodrat middens in lava tubes confirm that sagebrush and juniper grasslands replaced xeric shadscale vegetation as suggested by the pollen sequence.

Archaeological studies may expose past human settlements and use of resources which, along with associated radiocarbon dates and sediments, give clues to the details of environmental history (Chatters 1995). For example, ancient dune occupations in the semi-arid Fort Rock Valley, Oregon, tell a story of changing climate. In human terms, the Fort Rock Valley usually was marginal but for short times people came in greater numbers to gather the occasional edible windfalls from climatic changes that transformed dry basins to lakelets and sloughs. This happened sporadically throughout the Holocene; but, judging from the distribution of radiocarbon-dated occupations, it was much more common between 2000 and 4000 than it was from 5000 to 7000 B.P. (Mehringer and Cannon 1994).

In southeastern-Oregon the mid-Holocene was a time of short, sharp climatic shifts. Increase in grass and decrease in sagebrush and chenopod pollen indicate periods of greater effective moisture. These occur just above Mazama tephra in cores from Fish and Wildhorse lakes, Steens Mountain (fig. 24). Soon after the eruption of Mount Mazama people returned to a marshy Fort Rock Valley to collect fish and waterfowl (mehringer and Cannon 1994). Human responses and fossil pollen profiles both suggest that a short period of ample moisture--then rapid desiccation--followed the climactic Mazama eruption by a century or two.

In summary, according to the studies reviewed here, cold steppe dominated eastern Washington during the last glacial period. After 12,000 B.P. wasting glaciers in the north and shrinking pluvial lakes in the south witnessed a brief conifer expansion, but by 10,000 B.P. all sites now in steppe or ponderosa pine, and some in Douglas-fir mixed forest supported grass and sagebrush. In eastern Washington the forest fringe apparently had retreated 50 to 100 km in the north and east; it did not approach its present lower elevational limits again until after 4400 B.P. In

the Steens Mountain region of southeastern Oregon, steppe vegetation persisted at all elevations but underwent punctuated differences in importance of grass, sagebrush and juniper, as well as in the production of charcoal that reflects importance of fire. In southeastern Oregon it seems that western juniper woodlands became regionally important after 5000-4000 B.P.

**Juniper Woodlands**--Radiocarbon-dated plant macrofossils from ancient **woodrat** (*Neotoma* spp.) **middens** are revealing the responses of desert shrubs and forest trees to late Quaternary climatic variation in detail never achieved solely through study of fossil pollen. In combination the two complimentary methods give greater resolution than either alone (Mehringer and **Wigand** 1990).

Unfortunately, except in the most **xeric** sites, the Northwest's climate is not conducive to preservation of **woodrat middens** because the hardened, dehydrated and crystallized **woodrat** urine (amber-at), that inhibits decay and cements the mass of plant remains is water soluble. Even in protected locations it may dissolve with seasonally high humidity. At lower elevations in dry caves and rockshelters, however, **woodrat middens** may persist for at least a few hundred years as far north as the dry interior of south central British Columbia. There, Hebda and others (1990) recovered twigs of Rocky Mountain juniper (*Juniperus scopulorum*), needles of ponderosa pine and Douglas-fir, and pollen and macrofossils of associated shrubs and herbs from **middens** dating to 700 and 1150 **B.P.** The fossils represent species growing locally today. To the south, in the northern Great Basin, **woodrat middens** 10,000 to 20,000 years old are not unusual; some have been dated to >30,000 **B.P.** (Thompson 1990; **Wigand** and Nowak 1992).

At Diamond Craters (BLM Outstanding Natural Area), southeastern Oregon, and Lava Beds National Monument, northeastern California, plant remains from **woodrat middens** in lava tubes and rock shelters give a **5000-year** perspective on the spectacular historic success of western juniper (*Juniperus occidentalis*) despite modern **programs** for its eradication (Mehringer and **Wigand** 1987, 1990). Notions that today's western juniper woodlands are unprecedented--artificially induced by cattle grazing and fire suppression--must be evaluated in light of recent information. Pollen and macrofossils reveal that late Holocene (4000-2000 **B.P.**) expanses of

western juniper probably exceeded current coverage (Miller and **Wigand** 1994). As figs. 25 and 26 illustrate, the range and abundance of juniper has changed through the late Holocene. At Diamond Craters, the combination of closely sampled cores from Diamond Pond (Malheur **Maar**) and analysis of **woodrat middens** in nearby **rimrock** and lava tubes reveal the reasons for the changing importance of juniper and associated phenomena.

Both macrofossils (**Wigand** 1987) and pollen of pond weeds (*Potamogeton*, *Ruppia*, *Ceratophyllum*), pond edge species such as cat-tails and sedges, and **differing** abundances and types of algae resistant to acid extraction reveal periods of varying water depths, water quality and pond size. Deeper, **fresher** water indicators are associated with relatively more grass in relation to sagebrush pollen and more sagebrush in relation to saltbushes (*Atriplex* and *Sarcobatus*) pollen. Juniper pollen is most abundant in these same samples. Because the water table is controlled by recharge and evaporation, episodes of deeper water suggest periods of more effective moisture. It was during these periods that juniper expanded its range and that sagebrush steppe held more grass.

Confirmation of striking changes in the late Holocene history of juniper woodlands comes from macrofossils in **woodrat middens** whose ages correspond to large juniper pollen values between about 3800 and 2200 B.P., with several lesser peaks between 2000 and 700 years ago, and again in the late A.D. 1700s and mid- 1900s (fig. 25). These **middens** hold fossil plant assemblages that record the fluctuating lower elevational border of western juniper onto present barren basalt flows and into communities now dominated by sagebrush or shadscale desert species.

Additionally, at a site in the nearby **Catlow** Valley, where western juniper is common today, a 6100 year B.P. **woodrat** midden-dating **from** a **xeric** interval--lacks juniper macrofossils (**Mehring** and **Wigand** 1990). On Steens Mountain the upper elevational limit of western juniper near Fish Lake has remained unimpeded by competition **from** other coniferous trees throughout the Holocene. Yet, the Fish Lake pollen record indicates that western juniper did not expand near its

upper elevational **limit**, not even when it withdrew **upslope** from its lower elevational limit during **xeric** intervals.

According to studies of climate response functions **from** tree rings, mild winters with ample precipitation and cool springs favor wide annual rings in western juniper (**Earle** and Fritts 1986; Fritts and Xiangding 1986); the same factors would favor grass that fuels wildfires. The fossil record reveals a correspondence in the largest juniper pollen percentages, and charcoal/pollen and grass/sagebrush ratios (figs. 25, 26), and in indicators of deeper water. The **woodrat midden** fossils and the fossil pollen sequence from Diamond Craters show that western juniper's abundance and lower elevational limits changed often over the last 5000 years. On a scale of centuries, juniper increased at **times** when water tables were higher, when grass was most abundant, and when fires that favored grass discouraged sagebrush. In the short term, expansion of juniper woodland seems to have been set back by drought and sometimes ended by catastrophic fires.

Plant macrofossils from **woodrat middens** from the Pine Creek Drainage of the Clarno Basin, north central Oregon, also revealed the history of western juniper in sagebrush steppe. Croft (1989) concluded that the historic expansion of juniper woodlands is not unique and that western juniper along with bitter-brush (*Purshia tridentata*) probably was more common than now between about 4700 and 4200 B.P. Otherwise, over the last 5000 **years**, **shrubs** such as big sagebrush (*Artemisia tridentata*) and shadscale (*Atriplex confertifolia*), and grasses varied **primarily** in relative abundances.

In eastern Oregon the macrofossil record of western juniper woodland is short either because it was not important before 5000-4000 B.P. or because the fossil record is incomplete. Western juniper twigs and seeds are present at Lava Beds National Monument, northeastern California, by 5200 B.P.; at Diamond Craters, Oregon, a single seed was recovered from pond sediments dated to about 4800 B.P. (Mehring and **Wigand** 1987). **Wigand** and others (1995) believe that the juniper pollen (species unknown) curve from Bicycle Pond indicates semi-arid juniper woodland on the lower slopes of Hart Mountain, Warner Valley, southern Oregon, about

8500 B.P. However, in places where western juniper abounds in the **Catlow Valley**, Oregon, today, the oldest (6100 B.P.) Holocene **woodrat midden** does not contain juniper. Perhaps, the Warner Valley juniper pollen represents a late persistence of the early Holocene range of Utah juniper (*Juniperus osteosperma*). Just to the south, in northwestern Nevada, many more and older **woodrat middens** show a long history of shifting distributions of desert scrub, steppe and woodland, including presence of Utah juniper at **times** over the last 30,000 years (Nowak and others 1994a; **Wigand** and others 1995).

Study of 96 dated fossil plant assemblages from 24 **woodrat midden** localities in northwestern Nevada, gave evidence for the **30,000-year** presence of Utah juniper (*Juniperus osteosperma*) and arrival of pinyon pine (*Pinus monophylla*) less than 2000 years ago. **Wigand** and Nowak (1992) suggest the following sequence of changing vegetation and climate for the region along the western arm of Pluvial Lake Lahontan:

1. About 30,000 years ago, during a cold and dry period, Utah juniper (*Juniperus osteosperma*) woodland existed over a sagebrush and shadscale understory.
2. After 24,500 B.P. increased moisture encouraged spread of whitebark pine (*Pinus albicaulis*) to as low as 1380 m (1300 m below and 10 kilometers from its present nearest locality in the Sierra Nevada) and sagebrush dominated moist open slopes that now support shadscale desert.
3. By 21,500 B.P., with onset of a cool, dry glacial maximum climate, whitebark pine and Utah juniper gave way to sagebrush dominated steppe and desert scrub communities.
4. Sagebrush steppe expanded with warmer and wetter conditions of the late glacial.
5. After 10,000 B.P. much warmer and drier conditions favored shadscale communities at the expense of sagebrush steppe and juniper woodlands.
6. After 4000 B.P. sagebrush steppe and juniper woodlands regained importance under cooler and more **mesic** conditions.
7. Yellow pine (*Pinus ponderosa*) arrived in the region only about 2000 years ago.
8. Macrofossils of pinyon pine (*Pinus monophylla*) finally appeared in **woodrat middens** between 1700 and 1000 B.P. and at pinyon's northwestern range limit only **400** years ago (fig. 27).

Even with the excellent macrofossils from **woodrat middens** of northern Nevada, and adjacent California and Oregon, variable associations of fossil shrub steppe and woodland **taxa** still require explanation in terms of individual species responses, dispersal and adaptations to climate (Davis and others 1986). For example, **Nowak** and others (1994b) noted that some woodland species apparently had “remained nearly in place” or “experienced only **small** changes in elevation.” They contrasted the late Pleistocene history of Utah juniper and single-needle pinyon pine and concluded that Utah juniper had been present continuously in Northwestern Nevada for 30,000 years. They also suggested that its distribution was influenced little by late Pleistocene climatic events and that pinyon pine, on the other hand, slowly shifted northward in response to climatic change.

Of course, these species have different tolerances and responses. Though Utah juniper grew ‘in northwestern Nevada about 30,000 B.P. while pinyon pine’s northern limit was far to the south, the record of Utah juniper in northern Nevada **woodrat middens** is far from continuous. Thus far, no juniper remains from these **middens** are directly dated between about 28,500-24,500 and 20,000-12,000 B.P. Radiocarbon dates on juniper twigs and seeds from the northern Nevada **middens** cluster, as they do elsewhere; these clusters of  $^{14}\text{C}$  dates may be indicating periods of ideal conditions for **woodrats** (Mehring and **Wigand** 1990).

Additionally, single-needle pinyon and Utah juniper were, and continue to be, active players on the Great Basin’s Pleistocene stage. They both have migrated up and down mountains in response to varying climatic gradients. For example, about 19,000 and 10,000 **B.P.** Utah juniper grew at least 600 m below its present lower limit in the White Mountains on the **California-Nevada** border. About 8800 B.P. pinyon pine joined Utah juniper near the present day lower limit of pinyon woodland; Utah juniper now is absent from this site. This earliest date for pinyon in the **White Mountains** is about 3000 years later than the earliest date from the Eleana Range only 150 km to the east (fig. 27). In the mid-Holocene, about 5600 B.P., **midden** macrofossils reveal that pinyon and Utah juniper advanced 250 m **upslope** from their present upper elevational limits into what is now subalpine woodland of bristlecone and limber pine (Jennings and Elliott-Fisk 1991,



1993). The elevational response of Utah juniper is repeated in range after range in the Great Basin (Thompson 1990; Wigand and others 1995). With study of **midden** macrofossils and continuous pollen records **from** the same regions we can hope to comprehend woodland history.

## CONCLUDING REMARKS

Familiar plant associations in shrub steppe, woodland, montane forests, **moist maritime** forest, and alpine meadows consist of species that have responded independently to the varying intensities of numerous glacial-interglacial cycles. In each cycle, a relatively brief warm period separated long cold intervals; plants responded to the vagaries of climate, differing dispersal potentials, inter-species competition, genetic selection, topography, substrate, variable soil processes, volcanism, fire regimes, chance, and more recently to the numerous effects of ever-increasing human population. With few exceptions, details of past species distributions and associations are poorly understood in this region. However, evidence is mounting for better understanding of the chronology, magnitude and causes of late Quaternary vegetation change, and the roles of climate and disturbance.

By studying today's vegetation communities, without reference to the fossil record, we could not have known that the Columbia River Drainage's familiar broad distributions of woodland and steppe did not take shape until after 4000 B.P., that northern Idaho's hemlock and cedar forests were even younger, that Douglas-fir grew **upslope** from present positions and dominated the interior Northwest mountains during the early and mid-Holocene, or that a few centuries ago, and many times during the last 4000 years, juniper woodland's spread probably exceeded that of its modern expansion. The remarkably rapid demise of the Channeled Scabland's short lived, late glacial conifer woodlands is significant for revealing the process and pace of total **replacement**, and as an analog **for** the potential magnitude of future vegetation change with, perhaps, 4-5<sup>o</sup>C rapid global warming (Overpeck and others 1991). Except, of course, that large tracts committed to cultivation and cities may limit or preclude migration of vegetation in the face of changing climate.

This review illustrates 10,000 years of fire's considerable influence on this region's vegetation composition and structure. For millennia Indians saw fire as a resource management tool and enjoyed the benefits of natural and accidental **fires**. Antithetically, for most of the last century managers have been committed to controlling wildfires. But, fire suppression has backfired--encouraging disease, conflagrations, and unwelcome invaders. To the extent that we value natural ecosystems it seems sensible to honor fire's natural role.

Human influence on the Columbia River Drainage has increased steadily since the explorations of Lewis and Clark (1804-06). But, their coming was not the first to presage drastic environmental change. Another journey, 11,000 years earlier, brought humans armed with their tools of destruction. They came to the banks of the Columbia River and witnessed eruptions of Glacier Peak (Mehring and Foit 1990). They quickly hunted the Pleistocene megafauna (such as mammoth, horse and camel) to extinction or, at the very least, they are essential pieces to the jigsaw puzzle of extinction (Martin and Klein 1984). Humans remain a part of the ecosystem; it is our inclination to learn from our past and to plan for our future. Understanding the ephemeral and unpredictable nature of nature and taking the longer view promises enlightened ecosystem management. For, **like** the river of **Heraclitus**, we cannot step into the same forest twice.

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Table 1--Locations and references to sites of late Quaternary paleovegetation studies (figs. 2, 6 and 10).

Map No.	Site	References
<b>IDAHO</b>		
1	American Falls Lake Beds (1344 m)	Bright 1982
2	Bisonsweh Pond (2220 m)	Chatters 1982
3	Blue Lake, near <b>Lewiston</b> (1035 m)	Smith 1983
4	Clear Lake, near <b>Buhl</b> (915 m)	Davis, O.K., pers. <b>comm.</b>
5	Cub Lake (1840 m)	Baker 1983
6	Eastern Snake River Plain	Lennstrom 1985
7	Grays Lake (1950 m)	Beiswenger 1991
8	Hager Pond (860 m)	Hansen 1939; <b>Mack</b> and others 1978d; Mehringer and Van de Water, in prep.
9	Lake Cleveland (25 19 m)	Davis 1981, 1984; Davis and others 1986
10	Middle Butte Cave (1593 m)	Davis and Bright 1983; Davis and others 1986
11	Murphey's Rock Shelter (810 m)	Henry 1984
12	Rattlesnake Cave (1596 m)	Bright and Davis 1982; Davis and others 1986
13	Swan Lake (1450 m)	Bright 1966
<b>MONTANA</b>		
14	<b>Flathead</b> Lake (880 m)	<b>Onken</b> 1984
15	Forest Lake (1895 m)	Brant 1980
16	Kearns Basin (2200 m)	Foit and others 1993; Mehringer, unpubl.
17	Lost Trail Pass Bog (2152 m)	Mehringer, and others 1977a; Mehringer and others 1977b
18	Marias Pass (1550 m)	Carrara 1989
19	Marys Pond ( <b>Marys</b> Frog Pond) (1753 m)	Foit and others 1993; Karsian 1995
20	<b>McKillop</b> Creek Pond (920 m)	<b>Mack</b> and others 1983
21	Sheep Mountain Bog (1920 m)	<b>Hemphill</b> 1983; Johnson and others 1994; Mehringer, in prep.; Mehringer and others 1984
22	Smeads Bench Bog (960 m)	Chatters and <b>Leavell</b> 1994
23	Telegraph Creek Site (2130 m)	Brant 1980, 1982
24	Tepee Lake (1270 m)	<b>Mack</b> and others 1983
<b>NEVADA</b>		
25	Blue Lake (2432 m)	Carter 1995, Mehringer 1987
26	Mission Cross Bog (2424 m)	Thompson 1984
27	Virginia Mountains	Nowak and others 1994a, b; <b>Wigand</b> and <b>Nowak</b> 1992
<b>OREGON</b>		
28	Bicycle Pond (1520 m)	<b>Wigand</b> and others 1995
29	Clarno Basin	Croft 1989
30	Craddock Meadow (1630 m )	<b>Wigand</b> 1989
31	Diamond Pond ( <b>Malheur</b> Maar) (1265 m)	Mehringer 1987; Mehringer and <b>Wigand</b> 1987, 1990; <b>Wigand</b> , 1987
32	Dirty Shame Rockshelter (1430 m)	Sanford 1983
33	Fish Lake (2250 m)	Mehringer 1985, 1987; Verosub and others 1986
34	Lost Lake (1880)	Mehringer, in prep.
35	Summer Lake (1260 m)	<b>Wigand</b> and others 1995
36	Twin Lakes (1970)	Beck and Mehringer, in prep.
37	Wildhorse Lake (2565 m)	Mehringer 1985, 1987

## WASHINGTON

- 38 Battle Ground Lake (155 m)
- 39 Big Meadow (1040 m)
- 40 Bonaparte Meadows (1021 m)
- 41 Carp Lake (714 m)
- 42 Clear Lake (396 m)
- 43 **Creston** Bog (Wilbur) (710 m)
- 44 Goose Lake (373 m)
- 45 Kirk Lake (190 m)
- 46 **Manis** Mastodon Site (165 m)
- 47 Mud Lake (655 m)
- 48 Seed Cave (140 m)
- 49 Simpsons Flats (535 m)
- 50 Sulphur Lake (210 m)
- 51 **Touchet** Mammoth Site
- 52 Waits Lake (540 m)
- 53 Wildcat Lake (342 m)
  
- 54 **Williams** Lake Fen (635 m)

## WYOMING

- 55 Beaverdam Creek (2466 m)
- 56 Blacktail Pond (2018 m)
- 57 **Buckbean** Fen (2367 m)
- 58 Cub Creek Pond (2512 m)
- 59 Cygnet Lake Fen (2530 m)
- 60 Divide Lake (2628 m)
- 61 Emerald Lake (2634 m)
- 62 **Fallback** Lake (2597 m)
- 63 Gardiners Hole (2215 m)
- 54 Grassy Lake Reservoir (2200 m)
- 65 Hedrick Pond (2073 m)
- 66 Lily Fen & Lake (2469 m)
- 67 Loon Lake (1900 m)
- 68 Mariposa Lake (2730 m)
- 69 Rapid Lake (3 134 m)

## BRITISH COLUMBIA

- 70 Bluebird Lake
- 71 **Chilhil** Lake (915 m)
- 72 **Dunbar Valley** (1100 m)
- 73 Finney Lake (1300 m)
- 74 **Fishblue** Lake (Blue Lake) (332 m)
- 75 Horseshoe Lake (411 m)
- 76 Kelowna Bog (405 m)
- 77 Lower Little **Slocan** Lake (725 m)
- 78 Marion Lake (300 m)
- 79 Pemberton Hill Lake (1020 m)
- 80 Phair Lake (715 m)
- 81 **Pinecrest** Lake (320 m)
- 82 Squeah Lake (305 m)

- Barnosky** 1985
- Mack** and others 1978a
- Mack** and others 1979
- Barnosky** 1984, 1985
- Bartholomew 1982
- Mack** and others 1976
- Nickmann** and Leopold 1985
- Cwynar 1987
- Petersen and others 1983
- Mack** and others 1979
- Thompson 1985
- Mack** and others 1978b
- Mehringer, unpubl.
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- Mack** and others 1978c
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- Johnson and others 1994; Mehringer, in prep.;
- Nickmann** 1979

- Baker, 1986; Baker and Richmond 1978
- Gennett** 1977
- Baker 1970, 1976
- Waddington and Wright 1974; **Whitlock** 1993
- Whitlock** 1993
- Whitlock** 1993
- Whitlock** 1993
- Whitlock** 1993
- Baker 1983
- Baker, 1986; Baker and Richmond 1978
- Whitlock** 1993
- Whitlock** 1993
- Whitlock** and others 1995
- Whitlock** 1993
- Fall and others 1995

- Mathewes** 1985
- Mathewes** and King 1989
- Hazell 1979
- Hebda 1982
- Mathewes** and King 1989
- Mathewes** and King 1989
- Alley 1976
- Hebda 1995
- Mathewes** 1985
- Hebda 1995
- Mathewes** and King 1989
- Mathewes** 1985
- Mathewes** 1985

## FIGURE CAPTIONS

Figure 1. Times of maximum and minimum levels of effective moisture during the Holocene in the western U.S. (after Thompson and others 1993).

Figure 2. Boundary of study area and location of fossil pollen and **woodrat midden** macrofossil records; sites are numbered alphabetically by state and province (table 1).

Figure 3. Distribution of Glacier Peak (11,250 B.P.) and Mazama (6850 B.P.) tephras (Blinman and others 1979).

Figure 4. Glacier Peak and Mazama tephras from Lost Trail Pass Bog, Montana.

Figure 5. Late glacial age Mount St. Helens tephra lies above and below a couplet of Glacier Peak tephra (11,250 B.P.) in cores from Wiliams Lake Fen, Washington (Foit and others 1993).

Figure 6. Ecoregions of the Columbia River Drainage in the U. S. (Bailey and others 1994) and boundary of the area reported here. Circles mark localities of fossil pollen and **woodrat midden** macrofossil studies (table 1).

Figure 7. Schematic transect across southwestern Washington shows distribution of ice and dominant vegetation today, at 20,000 B.P. during the height of mountain ice caps (**Evans Creek Stade**), and at 14,000 B.P. (**Vashon Stade**) during the maximum advance of Cordilleran glaciers into the Puget Trough north of this transect (after **Barnosky** 1984; **Whitlock** 1992).

Figure 8. Summary of vegetation dominants and inferred climatic changes for sites in northeastern Washington and northern Idaho since recession of the last ice sheet (from **Mack** and others 1978d).

Figure 9. Vegetation distribution of the eastern Cascade Range and the mountains of northern Idaho (after **Barnosky** and others 1987).

Figure 10. Present day maximum distributional extent (Little 1971) and Holocene spread of western **hemlock** to interior localities since recession of the last ice sheet (table 1). Ages of arrival in the regions are inferred from the beginning of continuous presence of western hemlock (*Tsuga heterophylla*) pollen.

Figure 11. Influx of western hemlock pollen and AMS radiocarbon dating of hemlock fossil pollen and a needle, suggest its arrival at Hager Pond, Idaho, about 2400 B.P. and its greatest abundance about 1600 **B.P.**

Figure 12. Distribution of vegetation in the Bitterroot Range near Lost Trail Pass (Mehringer and others 1977a) and the Yellowstone National Park region (Baker 1986).

Figure 13. Diagram of major pollen types from Sheep Mountain Bog, Montana, since about 12,700 B.P.

Figure 14. Widths of black charcoal-rich sands and silt layers overlain by gray, silt and clay layers containing fine charcoal and **fire** ash. These couplets, and occasional single charcoal-rich layers resulted from wash-in after fire and, therefore, afford a **6000-year** record of the number and intensity of fires within the basin of Sheep Mountain Bog.

Figure 15. Comparison of the post-Mazama importance of fir, spruce and Douglas-fir at three sites in western Montana (**spruce+fir+Douglas-fir=100%**).

Figure 16. Distribution of vegetation in the **Wallowa/Blue** Mountains, Oregon (Johnson and **Clausnitzer** 1992) and Steens Mountain, Oregon (Mehringer 1987).

Figure 17. Diagram of the major pollen types from Williams Lake Fen, Washington, spans the time between the last **Scabland** Flood (about 12,700 B.P.) and the eruption of Mount **Mazama** (about 6850 B.P.).

Figure 18. Comparison of the chronology of late glacial and early Holocene vegetation change revealed by selected pollen types **from** Williams Lake Fen, Washington, and Sheep Mountain Bog, Montana (figs. 12, 17).

Figure 19. Percentage pollen diagram shows correspondence between shallow water indicators and charcoal abundance at **Williams** Lake Fen.

Figure 20. Williams Lake Fen percentages of water-lily and sedge pollen (shallow water indicators) compared with the grass and pine pollen curves. The “Tawny Ash” dates to 10,600 B.P.

Figure 21. Ternary plots of fossil pollen **from** four steppe communities in eastern Oregon and Washington. Relative abundance of saltbushes (including *Sarcobatus*), sagebrushes and grasses distinguish steppe associations that most often produce monotonous Holocene pollen profiles dominated by nonarboreal pollen and by pine pollen transported long distances (Mehringer 1985).

Figure 22. Ternary plots of saltbush, sagebrush and grass pollen from selected periods of the last 9000 radiocarbon years from Wildcat Lake, Washington. Note that the relatively largest grass values occur within the last 2400 years and that samples of the past 100 years (+) are distinctive (Mehringer 1985).

Figure 23. Today's steppe-woodland border and estimates of its position 10,000-9000 B.P. when steppe replaced conifer woodlands and forest, and 3500 to 2500 B.P. when conifer woodlands probably advanced into steppe all along eastern Washington's steppe-woodland ecotone (after Chatters, in press).

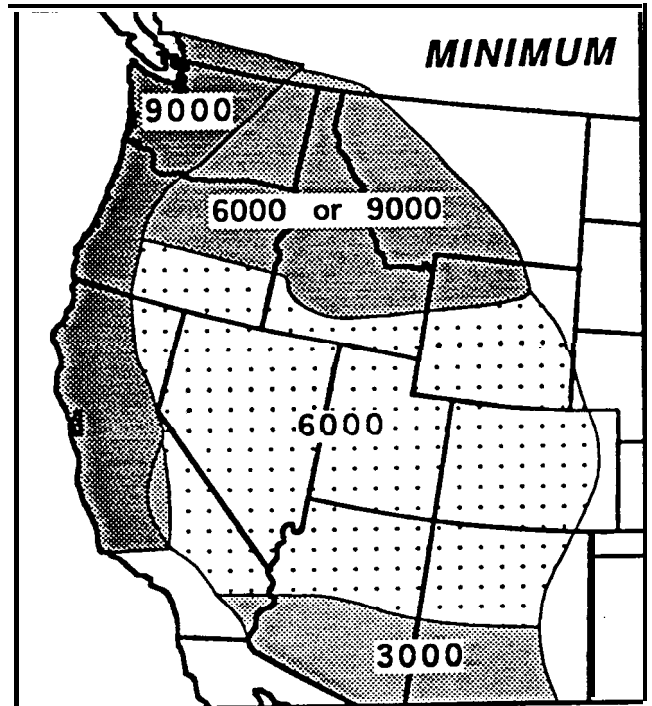
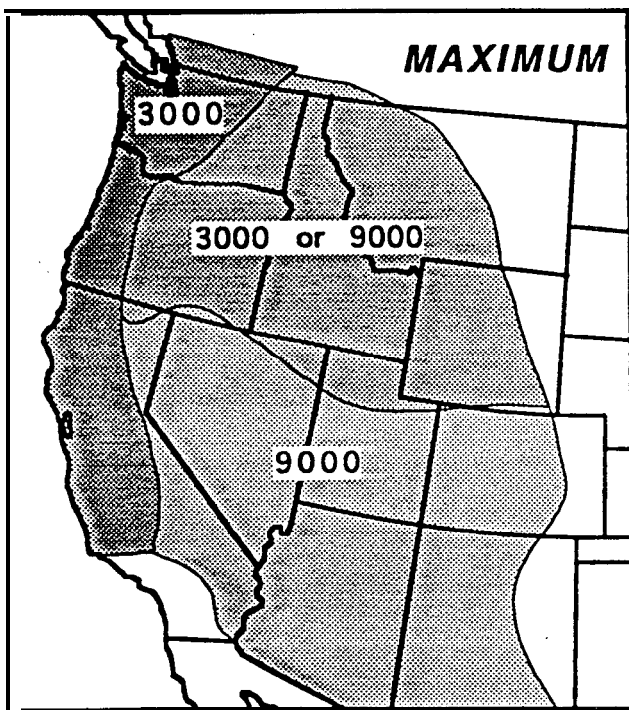
Figure 24. Ratios of sagebrush/grass pollen plotted about their means of the last 9700 and 9300 years B.P. Increase in sagebrush in relation to grass at Fish Lake (2250 m) suggests less effective moisture. The same variations at Wildhorse Lake, 315 m higher, at the current upper elevational limit of sagebrush, suggests **upslope** advance of sagebrush owing to warmer conditions with fewer snow patches lasting **into** summer (Mehringer 1985).

Figure 25. Smoothed juniper pollen percentages and ratios from the Diamond Pond core are plotted about the mean of the average values for each **500-year-interval** since 5500 B.P. Note correspondence in relative increase in juniper pollen, grass in relation to sagebrush pollen, and charcoal in relation to total terrestrial pollen beginning about 4000 B.P. Expansion of juniper woodland, increase in grass, and decline of sagebrush suggested by this diagram are confirmed by macrofossils from **woodrat middens** (Mehringer and **Wigand 1990**).

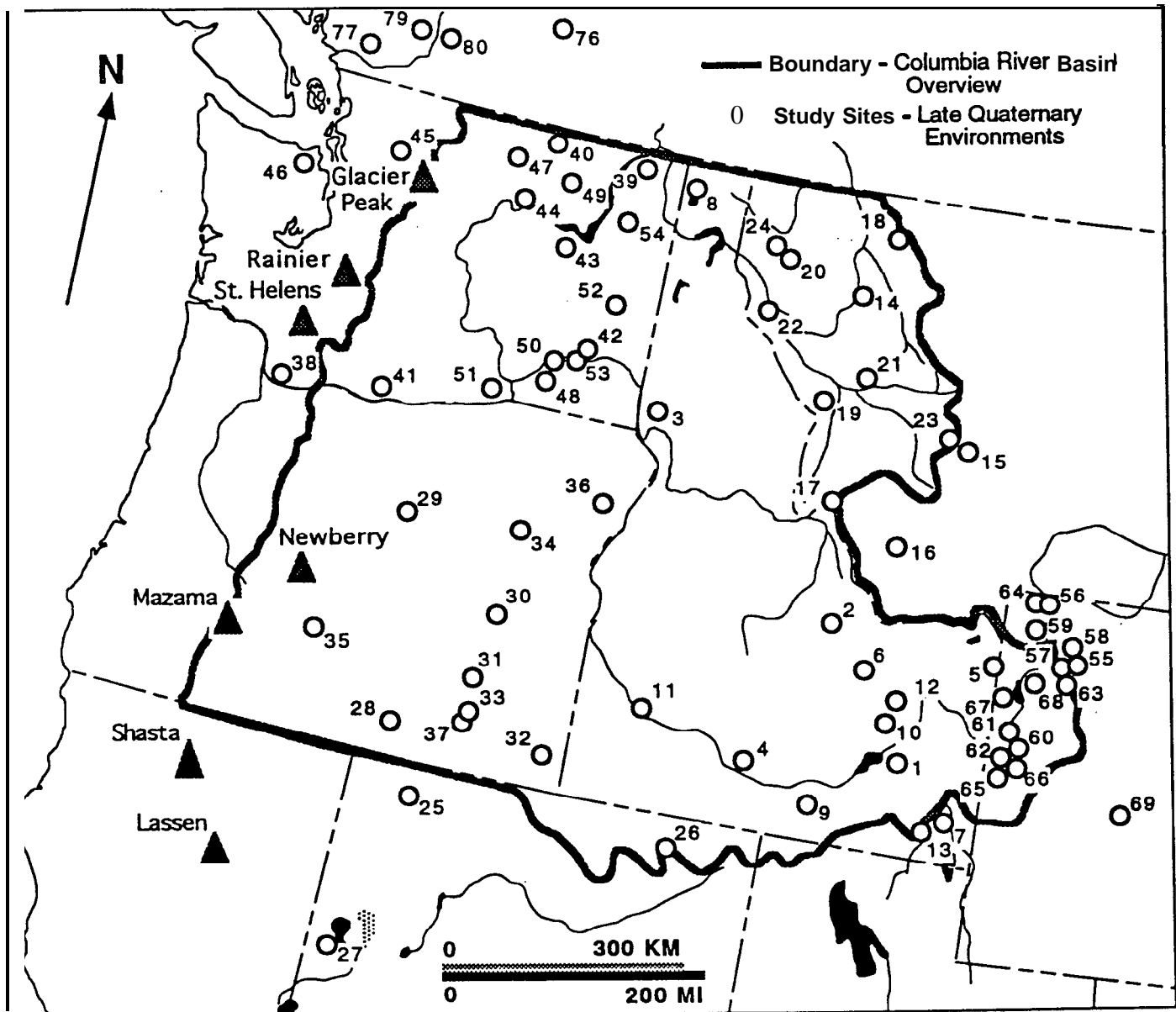
Figure 26. Smoothed Diamond Pond (**Malheur Maar**) juniper pollen percentages (fig. 25) and tree-ring corrected **woodrat midden** dates each plotted as a normal distribution with a probability of one; width at the base of each date is three standard deviations. Note that the clusters of **midden** ages correspond to the larger juniper pollen values from Diamond Pond sediments (**Mehring** and **Wigand** 1990).

Figure 27. Northward sporadic movement of pinyon pine (*Pinus monophylla*) from its late glacial positions on Clark Mountain, California (**Mehring** and **Ferguson** 1969), the Eleana Range, Nevada (**Spaulding** 1985), and other sites in southern Nevada (modified from **Nowak** and others 1994b; **Wigand**, pers. comm.). The far northwest distribution at **Hidden Valley**, Nevada, dates only from late Little Ice Age times (**Wigand** and others 1995).





**Times and Distributions of Holocene Effective Moisture**



**IDAHO**

- 1 American Falls Lake Beds
- 2 Bisonsweh Pond
- 3 Blue Lake
- 4 Clear Lake
- 5 Cub Lake
- 6 Eastern Snake River Plain
- 7 Grays Lake
- 8 Hager Pond
- 9 Lake Cleveland
- 10 Middle Butte Cave
- 11 Murphey's Rock Shelter
- 12 Rattlesnake Cave
- 13 Swan Lake

**MONTANA**

- 14 Flathead Lake
- 15 Forest Lake
- 16 Kearns Basin
- 17 Lost Trail Pass Bog
- 18 Marias Pass
- 19 Marys Pond
- 20 McKillop Creek Pond
- 21 Sheep Mountain Bog

- 22 Smeads Bench Bog
  - 23 Telegraph Creek Site
- NEVADA**

- 24 Tepee Lake
- 25 Blue Lake
- 26 Mission Cross Bog
- 27 Virginia Mountains

**OREGON**

- 28 Bicycle Pond
- 29 Clarno Basin
- 30 Craddock Meadow
- 31 Diamond Pond
- 32 Dirty Shame Rockshelter
- 33 Fish Lake
- 34 Lost Lake
- 35 Summer Lake
- 36 Twin Lakes
- 37 Wildhorse Lake

**WASHINGTON**

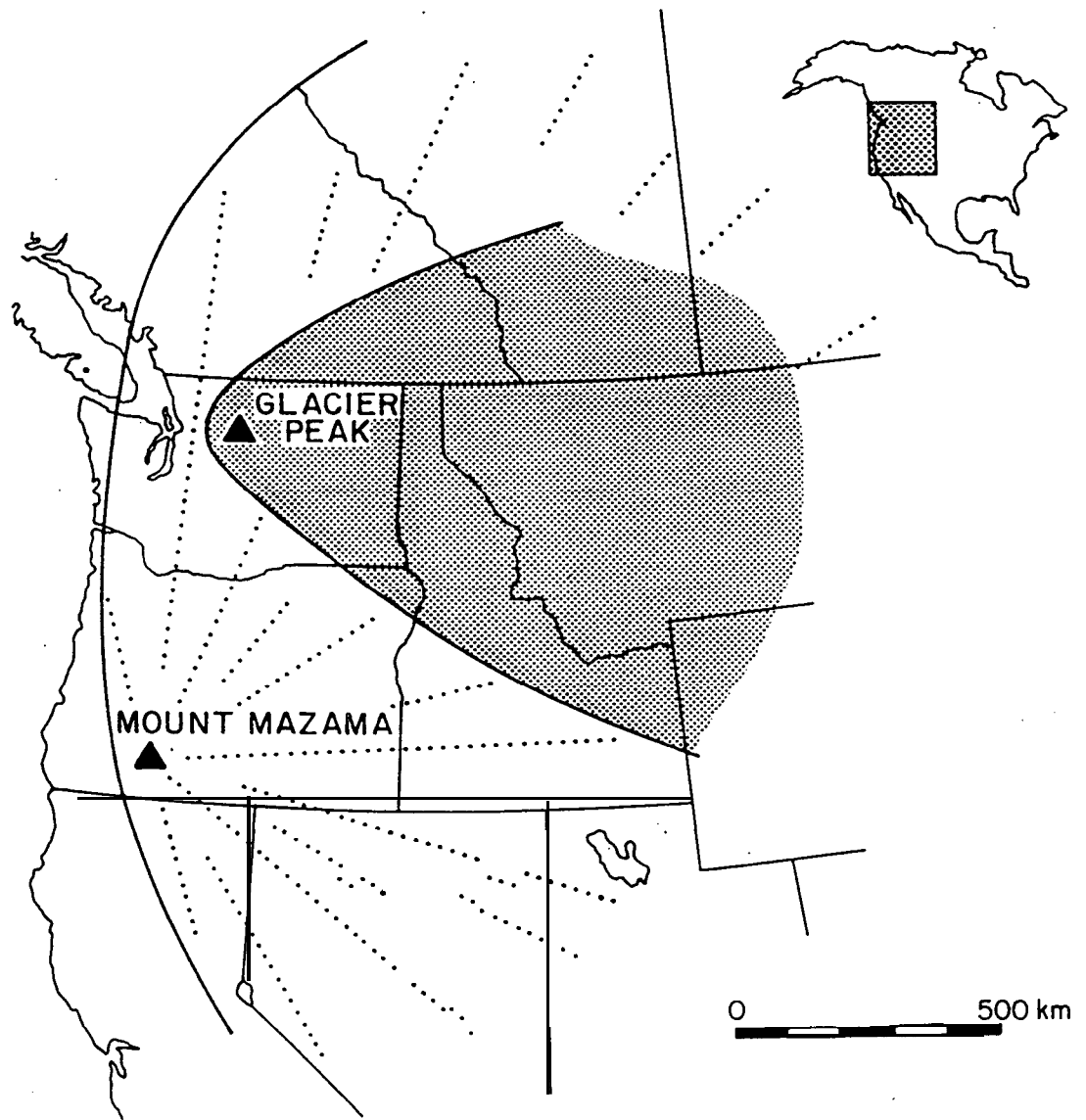
- 38 Battle Ground Lake
- 39 Big Meadow, WA
- 40 Bonaparte Meadows
- 41 Carp Lake
- 42 Clear Lake
- 43 Creston Bog
- 44 Goose Lake
- 45 Kirk Lake
- 46 Manis Mastodon Site
- 47 Mud Lake
- 48 Seed Cave
- 49 Simpsons Flat
- 50 Sulphur Lake
- 51 Touchet Mammoth Site
- 52 Waits Lake
- 53 Wildcat Lake
- 54 Williams Lake Fen

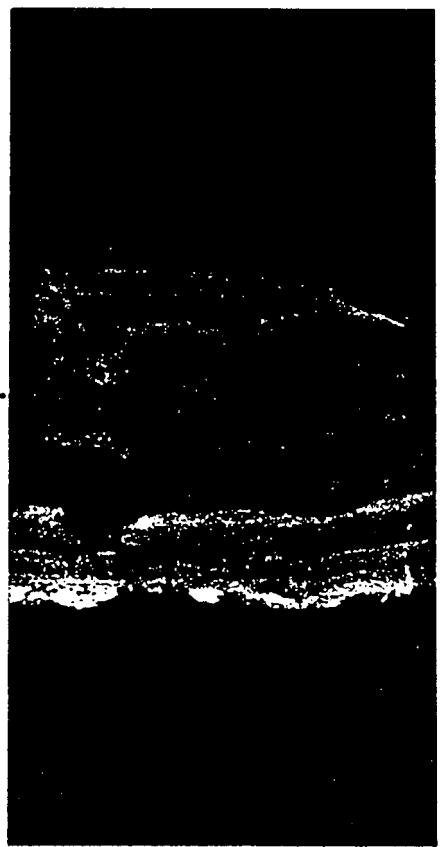
**WYOMING**

- 55 Beaverdam Creek
- 56 Blacktail Pond
- 57 Buckbean Fen
- 58 Cub Creek Pond
- 59 Cygnet Lake Fen
- 60 Divide Lake
- 61 Emerald Lake
- 62 Fallback Lake
- 63 Gardiners Hole
- 64 Grassy Lake Reservoir
- 65 Hedrick Pond
- 66 Lily Fen & Lake
- 67 Loon Lake
- 68 Mariposa Lake
- 69 Rapid Lake

**BRITISH COLUMBIA**

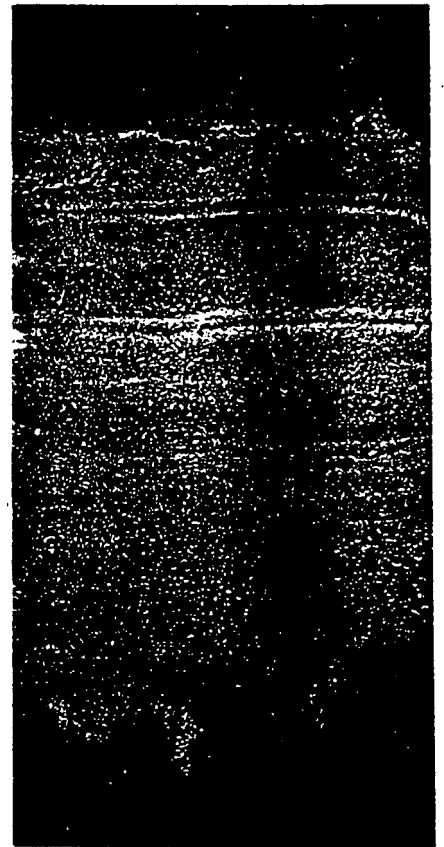
- 70 Bluebird Lake
- 71 Chilhil Lake
- 72 Dunbar Valley
- 73 Finney Lake
- 74 Fishblue Lake (Blue Lake)
- 75 Horseshoe Lake
- 76 Kelowna Bog
- 77 Lower Little Sloan Lake
- 78 Marion Lake
- 79 Pemberton Hill Lake
- 80 Phair Lake
- 81 Pincrest Lake
- 82 Squeah Lake





GLACIER PEAK ASH

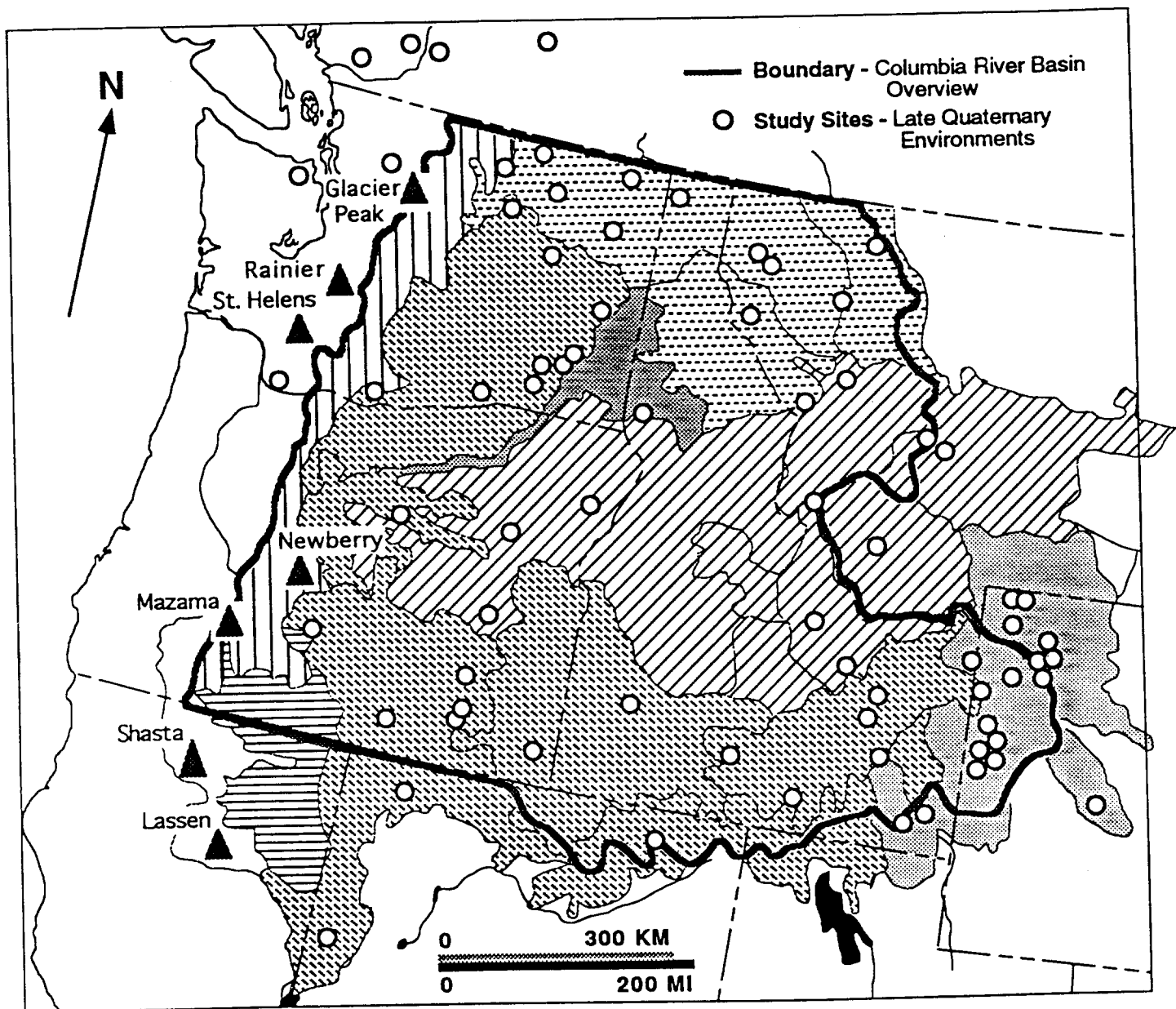
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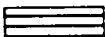

MAZAMA ASH

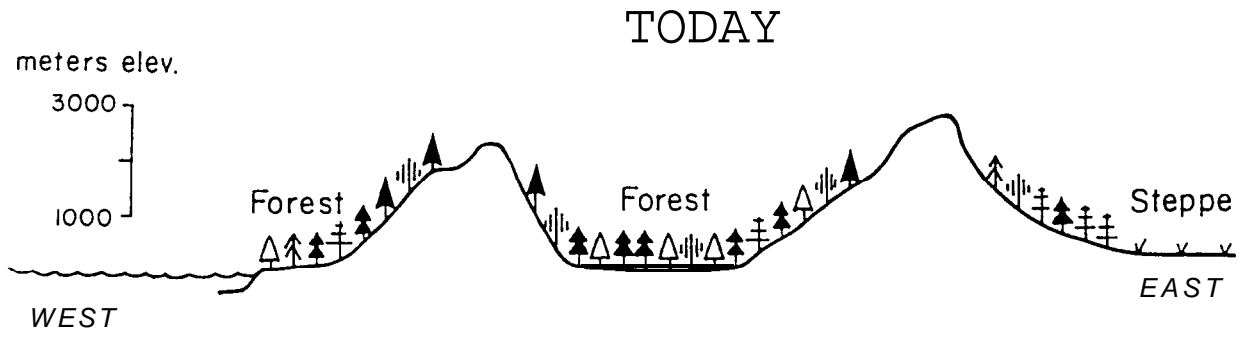


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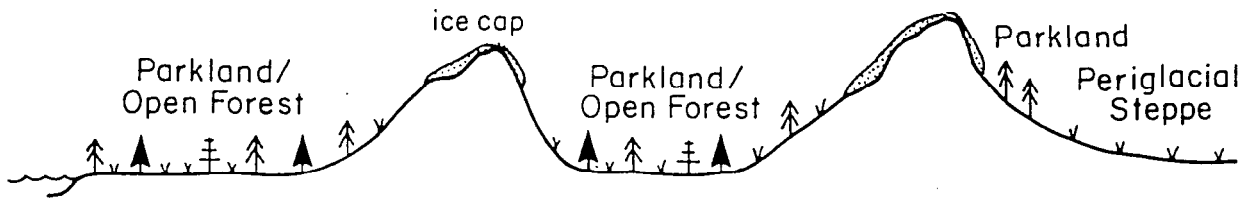


**ECOREGIONS**

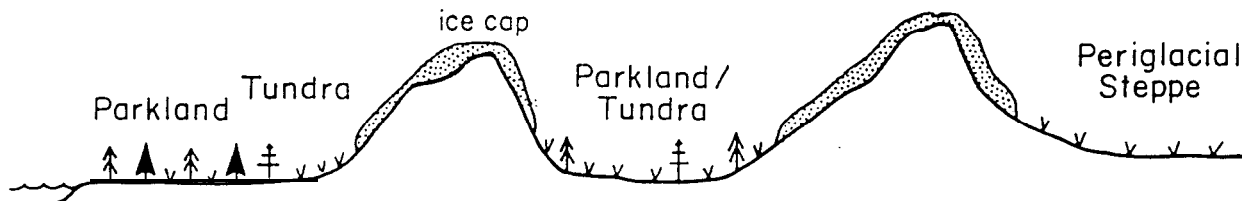
- |   |                        |   |                           |
|---|------------------------|---|---------------------------|
|  | Northern Rocky Mt.     |  | Intermountain Semi-Desert |
|  | Middle Rocky Mt.       |  | Cascade Mixed Forest      |
|  | Southern Rocky Mt.     |  | Sierra-Southern Cascade   |
|  | Palouse Prairie-Steppe |   |                           |



14,000 YEARS B.P.



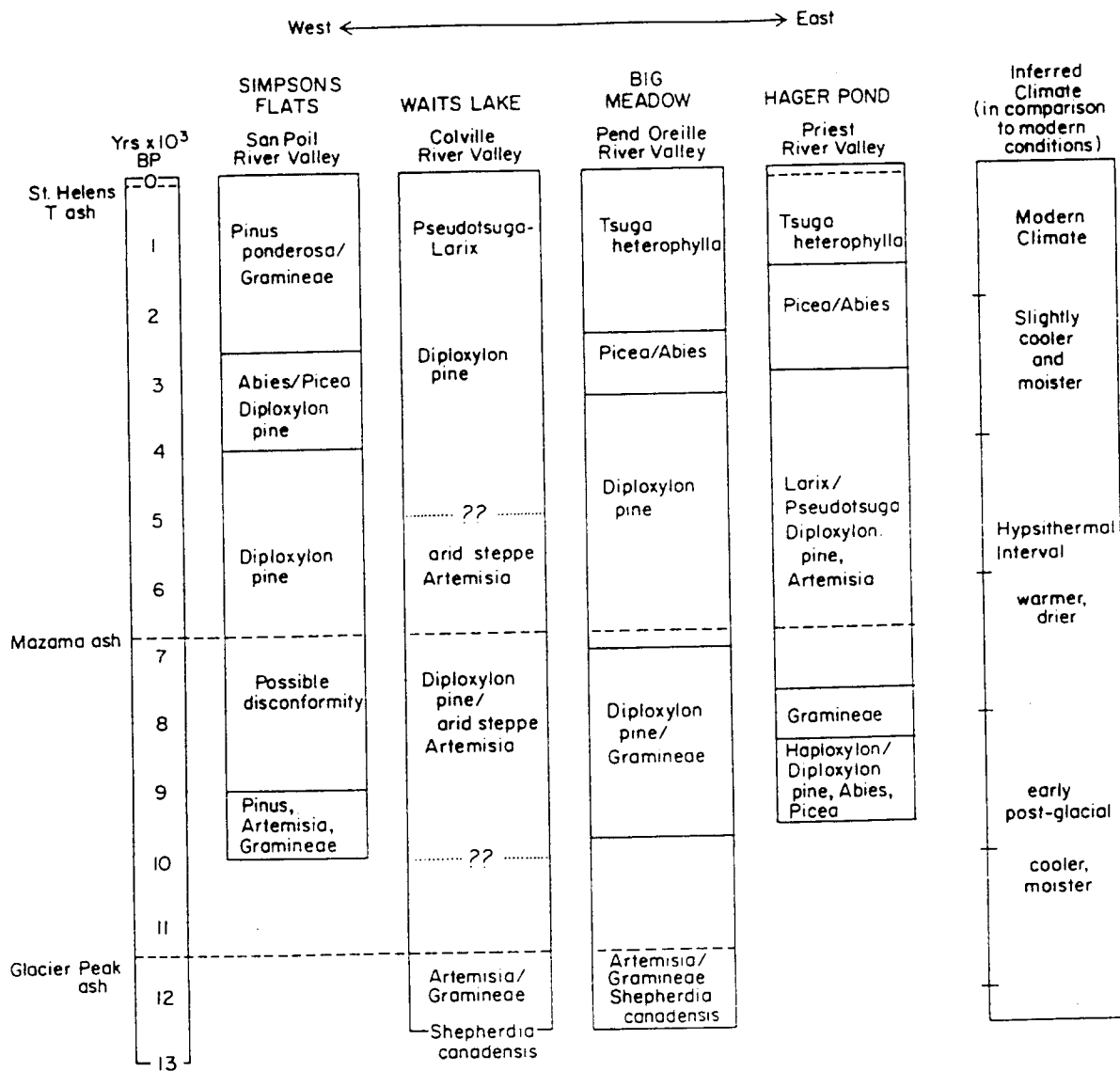
20,000 YEARS B.P.



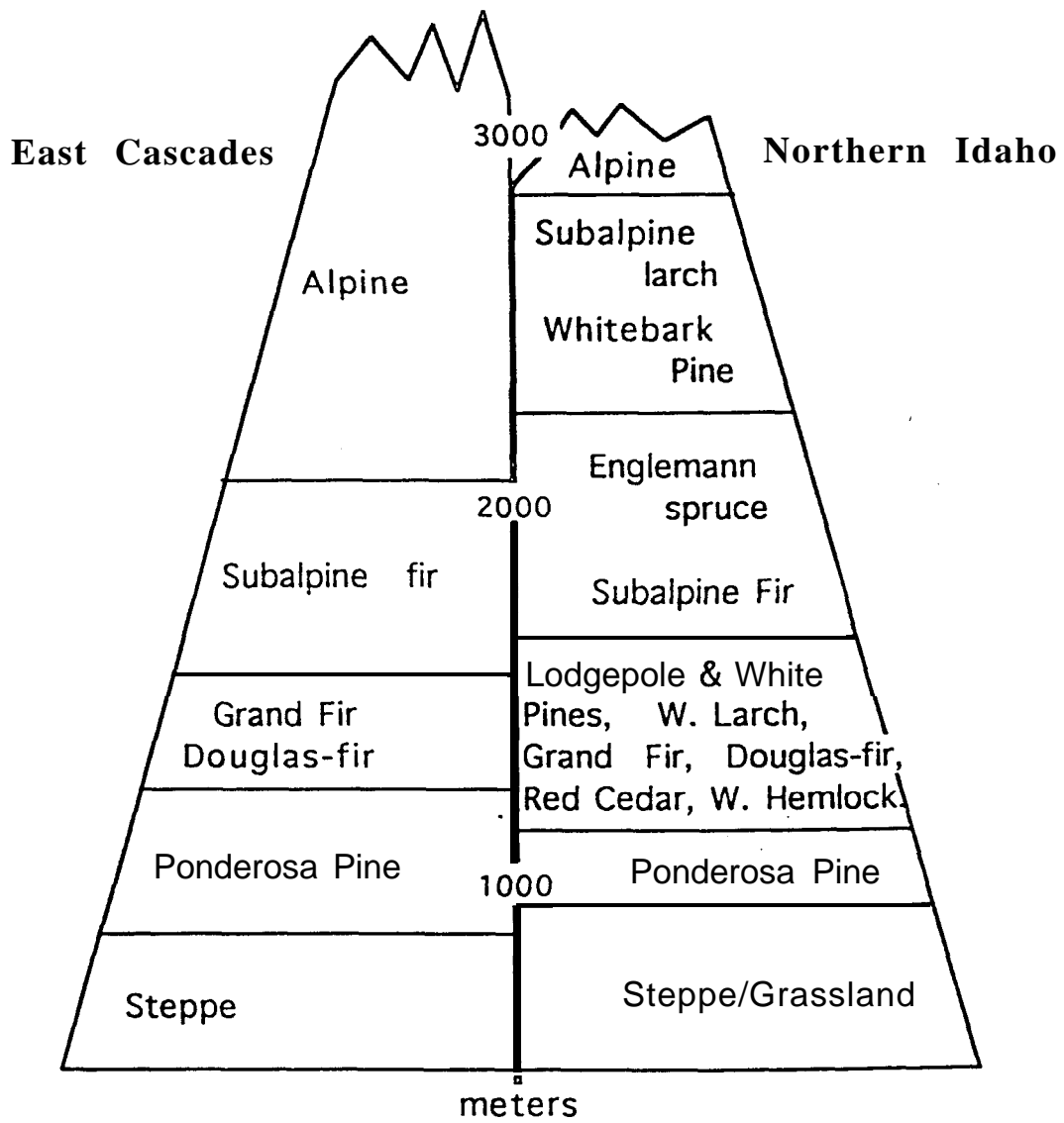
PACIFIC OCEAN    PACIFIC SLOPE    OLYMPIC MTS    PUGET TROUGH    CASCADE RANGE    COLUMBIA BASIN

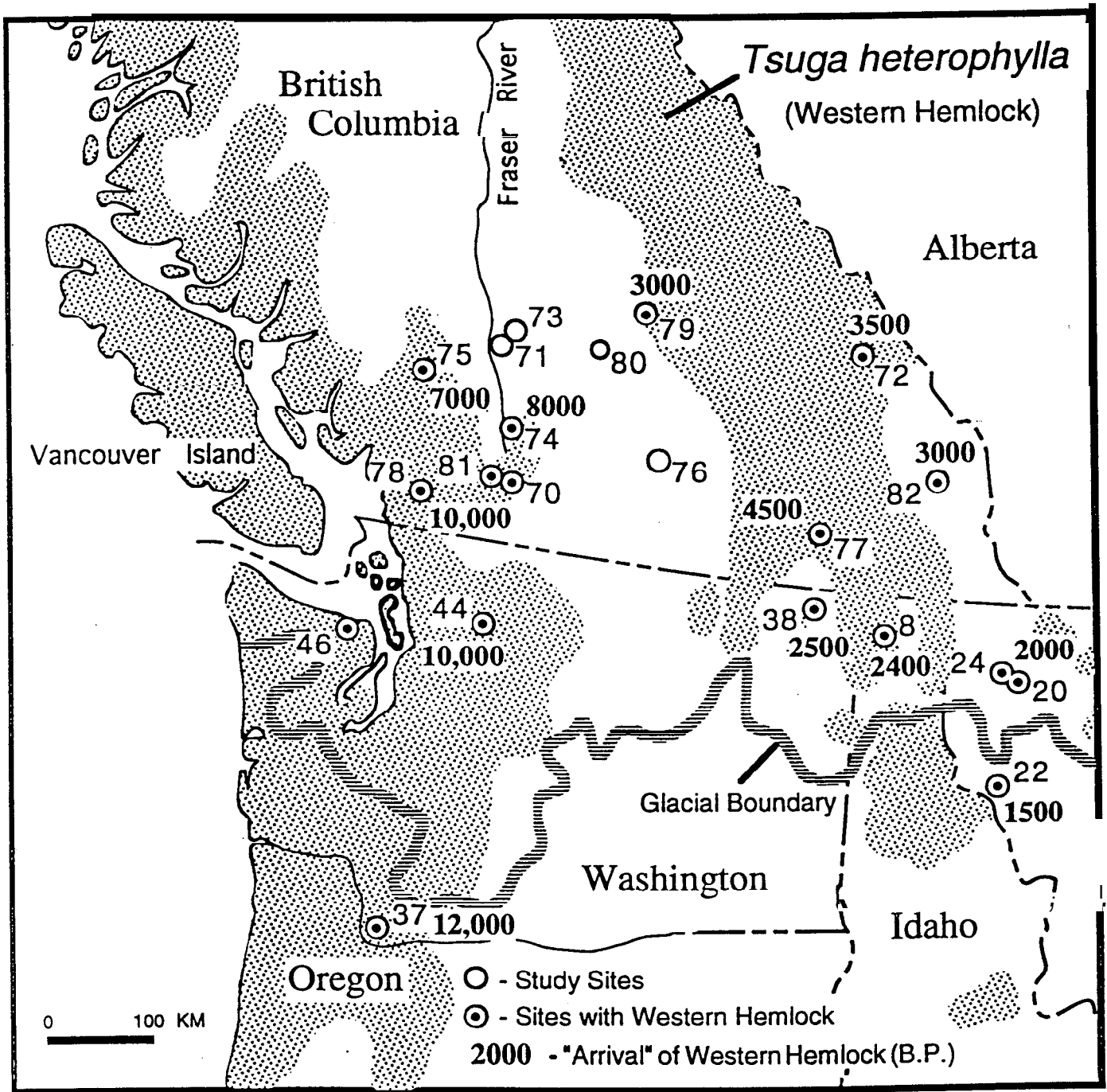
△ *Tsugo heterophyllo*      ↑ *Piceo*      𐄂 *Abies*      √ Nonarboreal  
 ▲ *Tsugo mertensiono*      ▲ *Pseudotsuga*      † *Pinus*

50 km

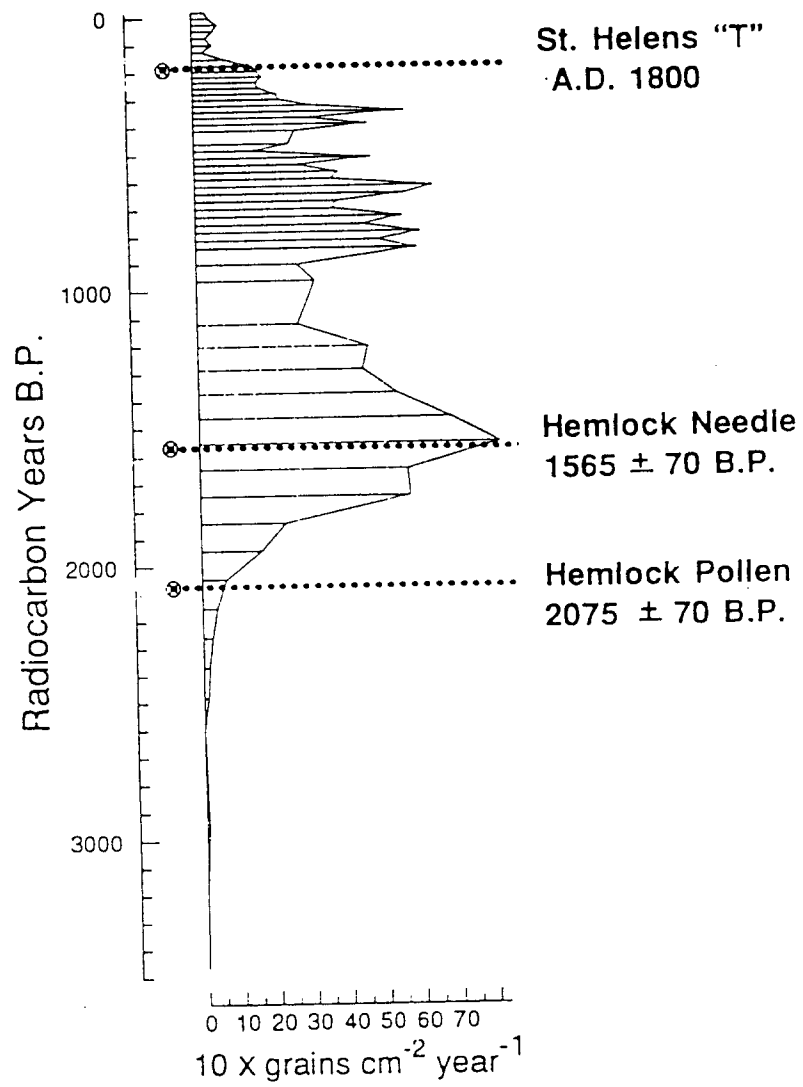


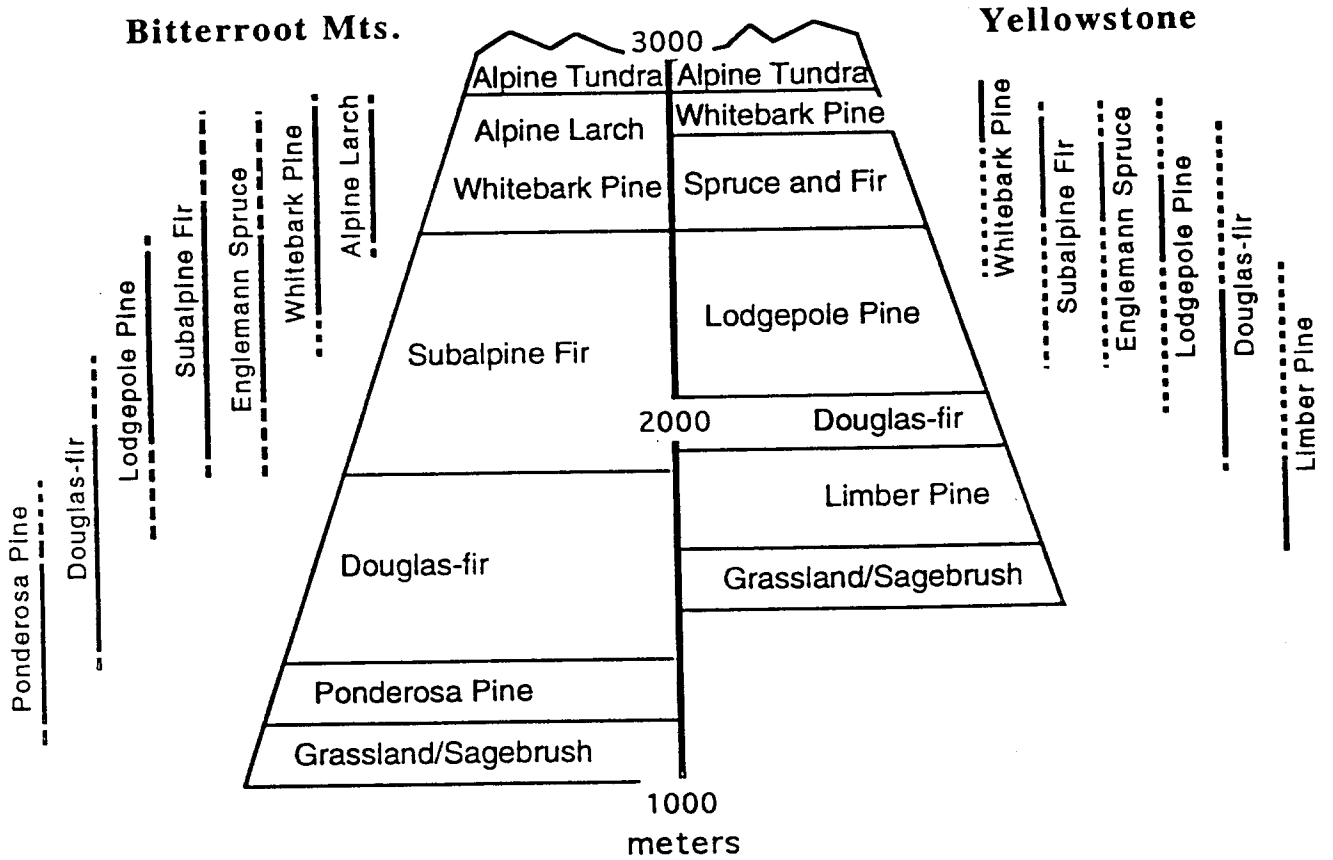




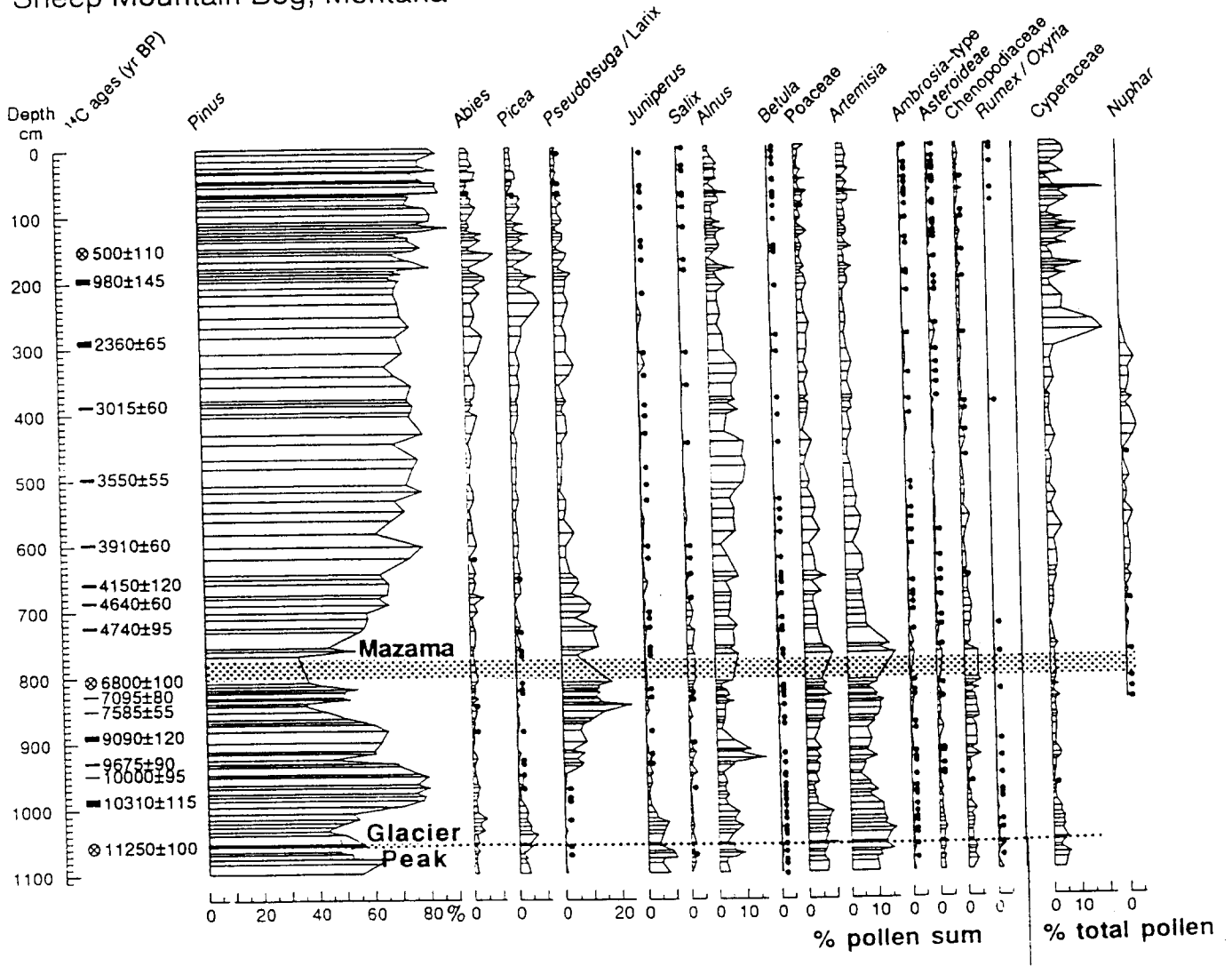


*Tsuga heterophylla* pollen  
Hager Pond, Idaho

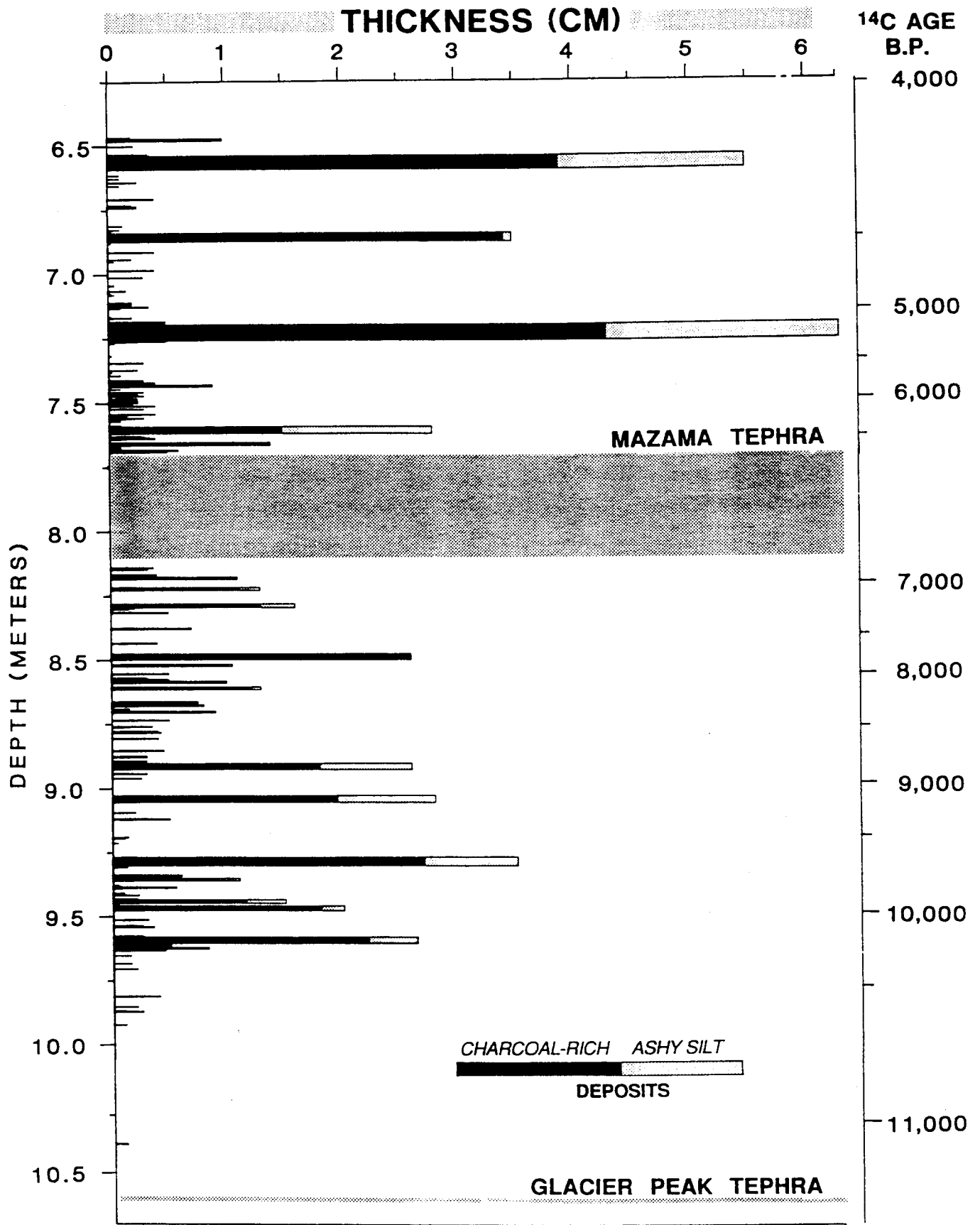


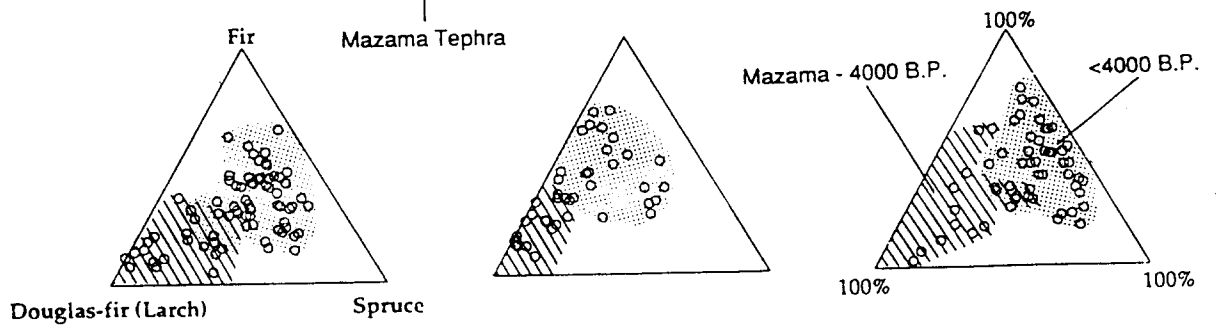
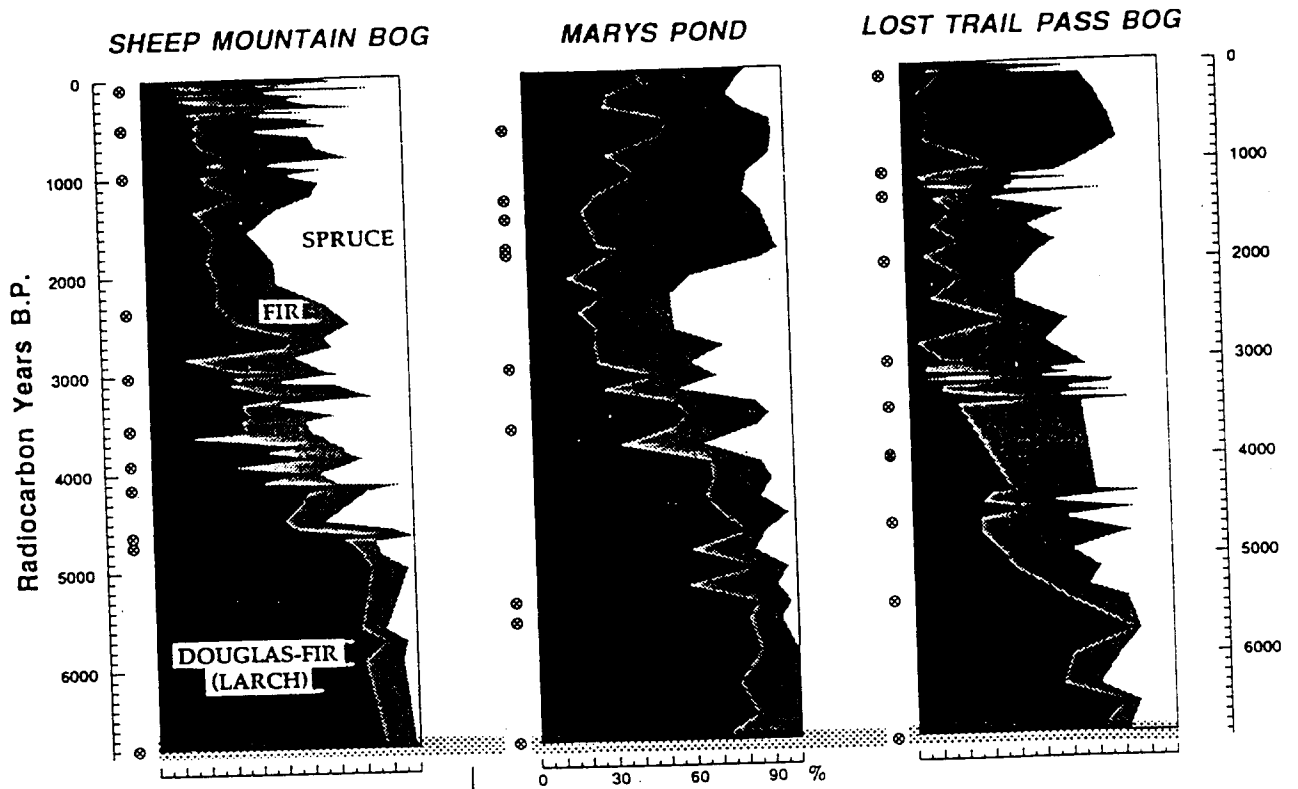


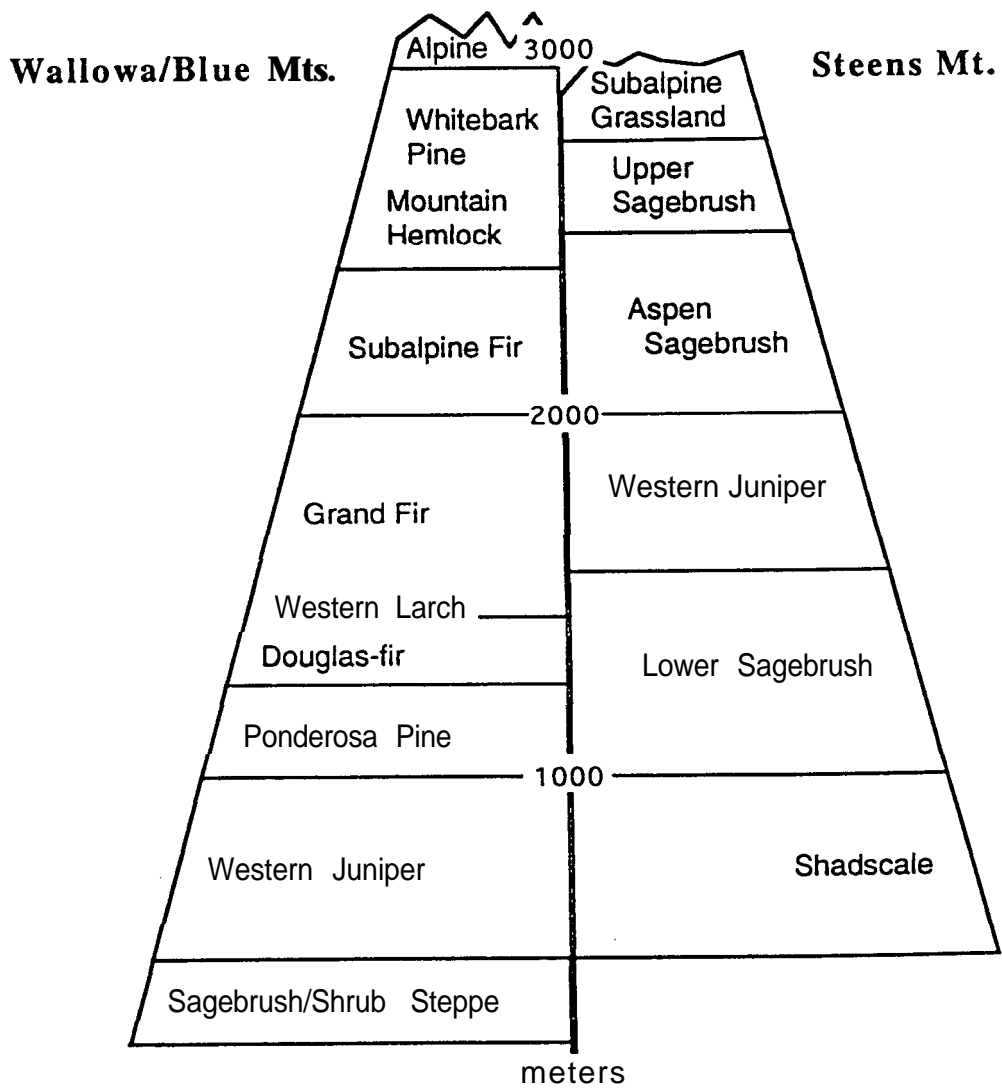
# Sheep Mountain Bog, Montana



# CHARCOAL & SEDIMENT BEDS

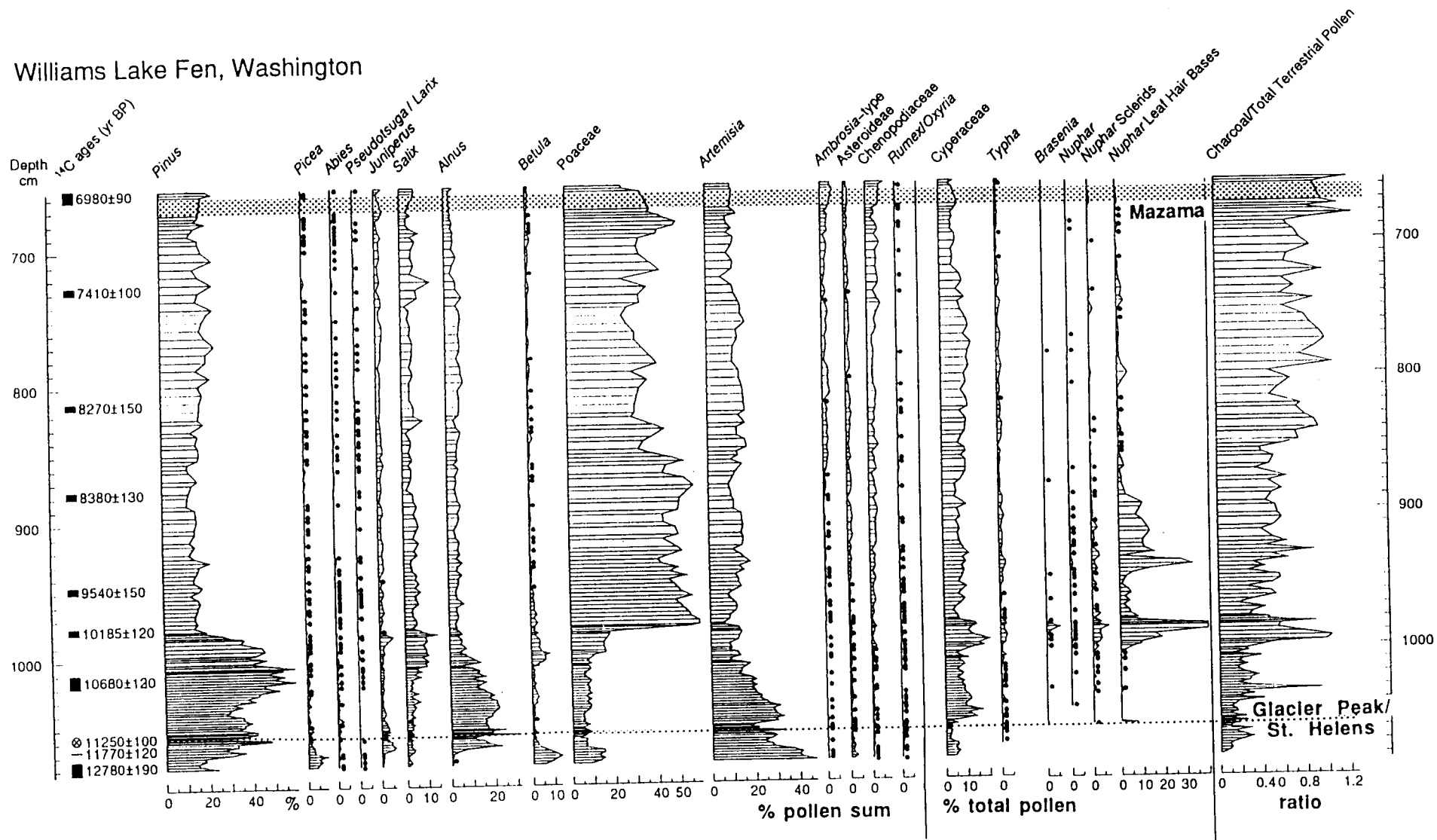






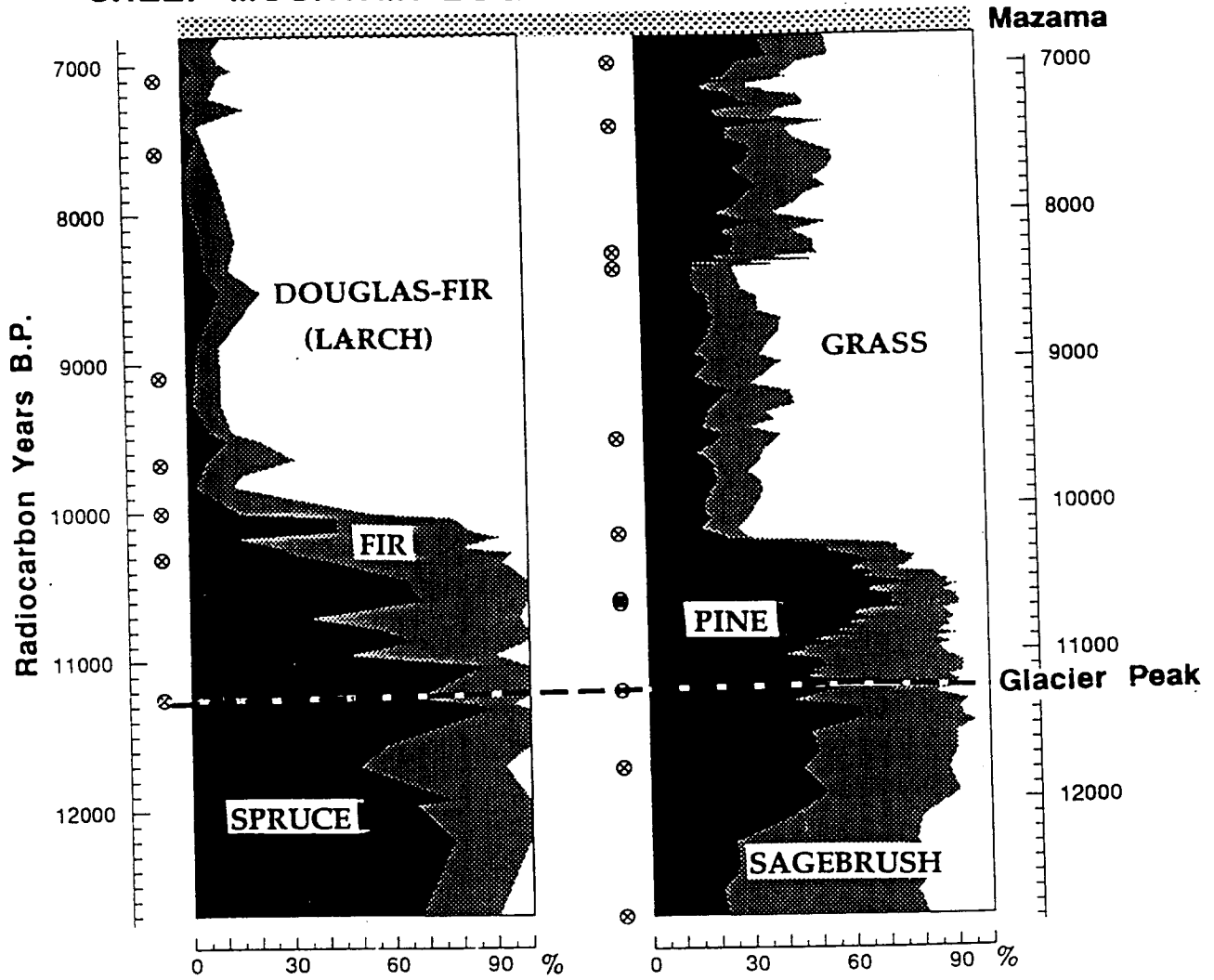


# Williams Lake Fen, Washington

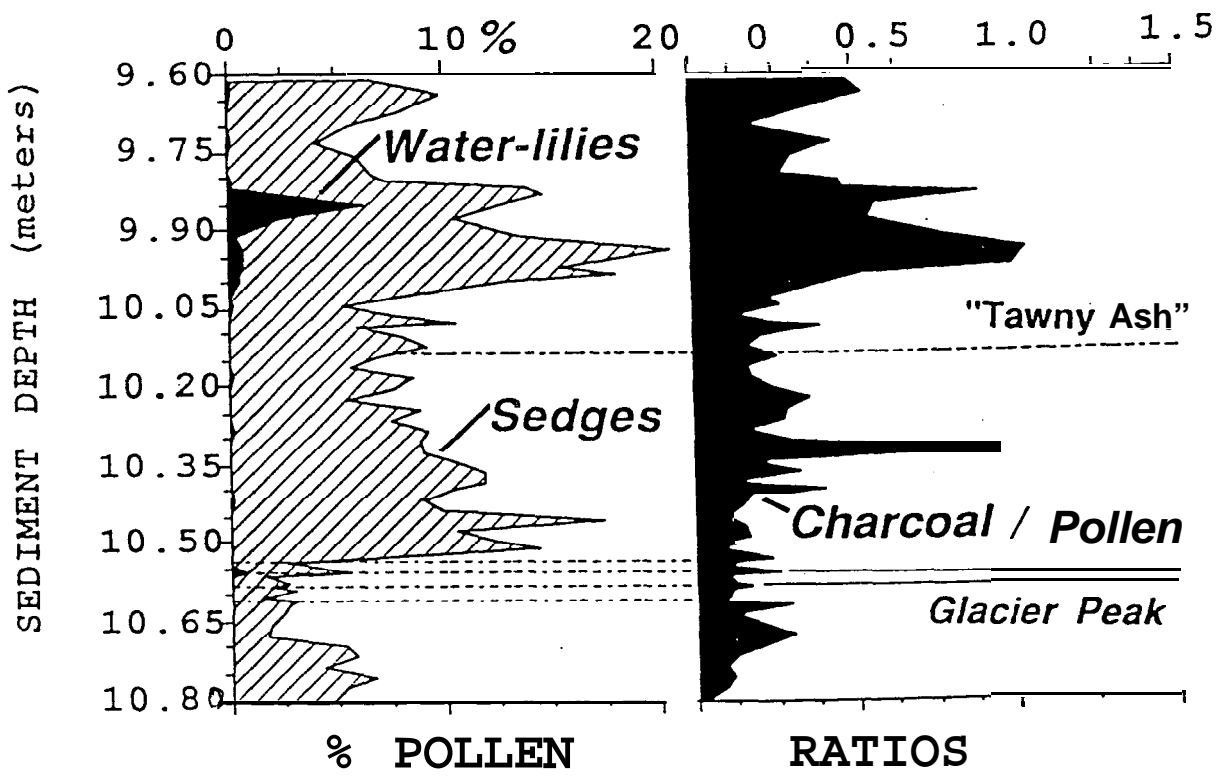


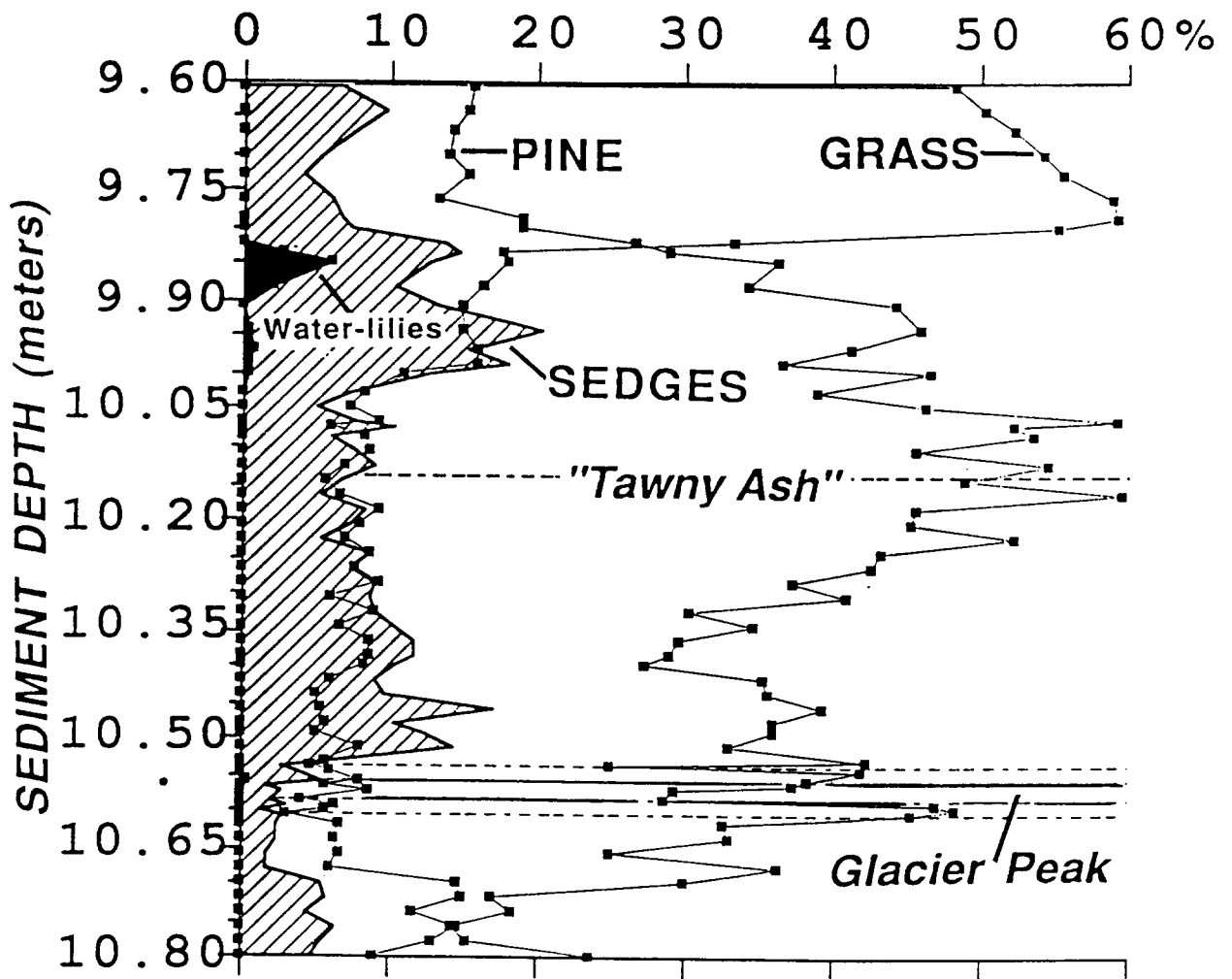
**SHEEP MOUNTAIN BOG**

**WILLIAMS LAKE FEN**

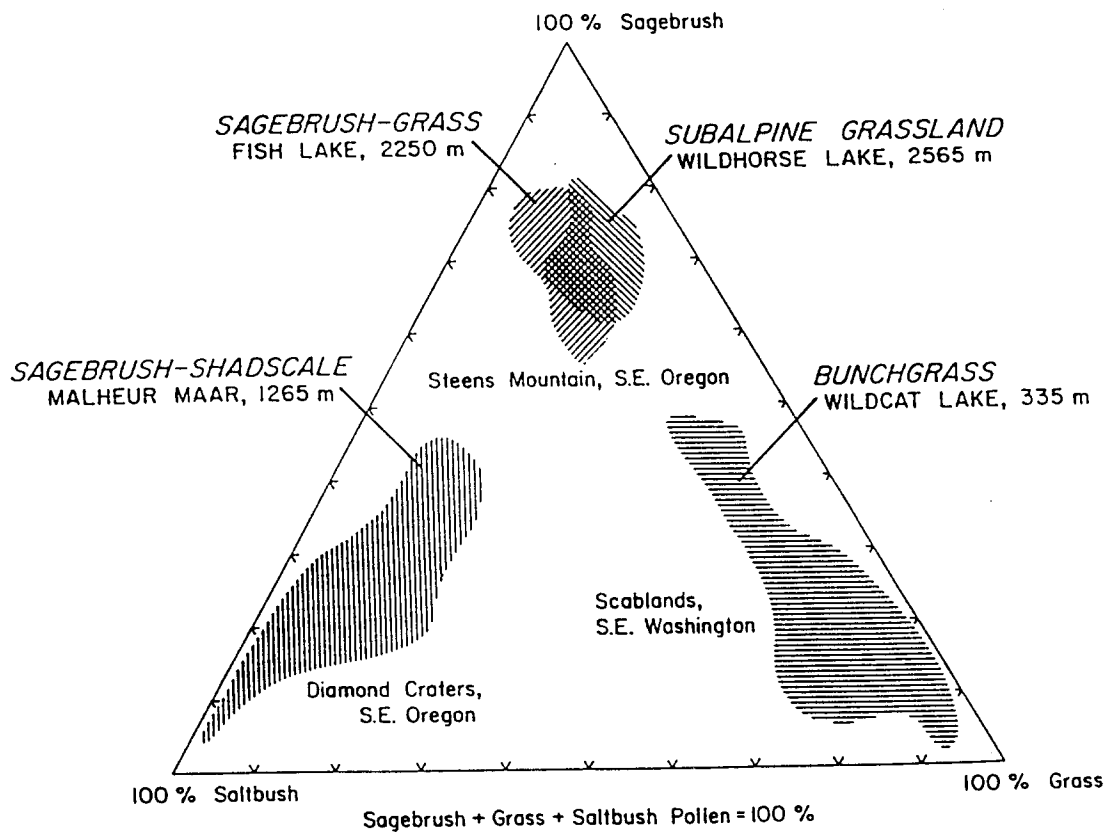


# WILLIAMS LAKE FEN

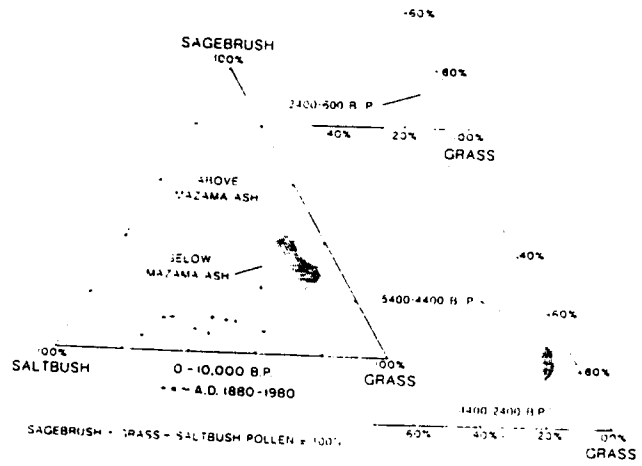




TERNARY PLOTS OF FOSSIL POLLEN (0-5700 B.P.)

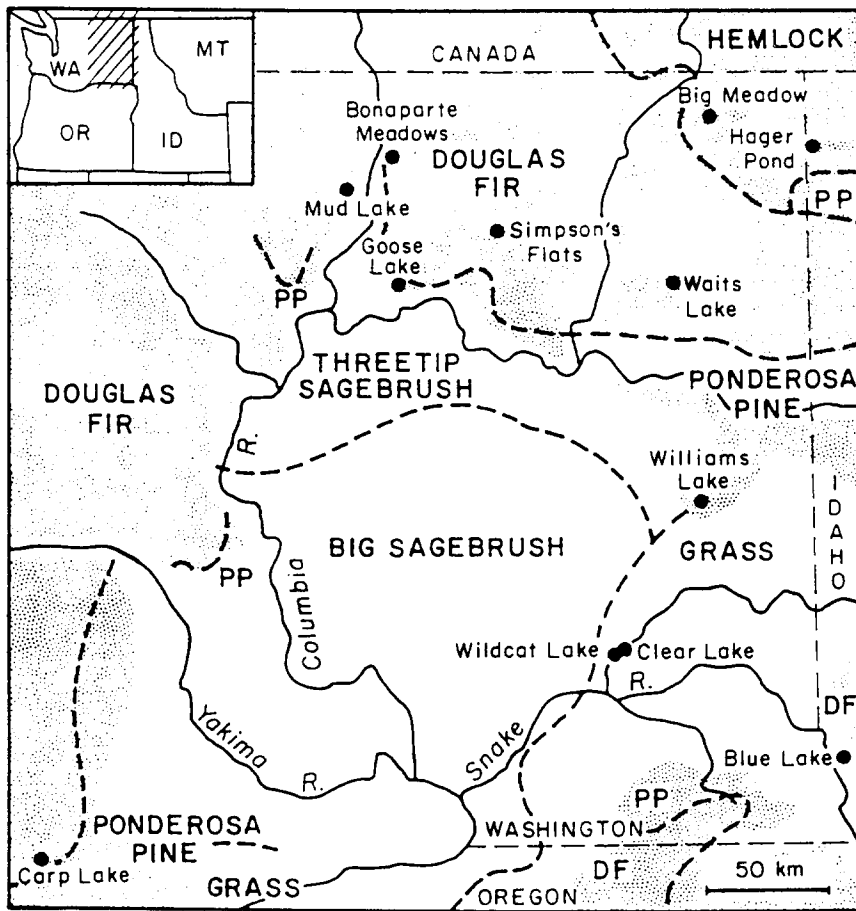


WILDCAT LAKE

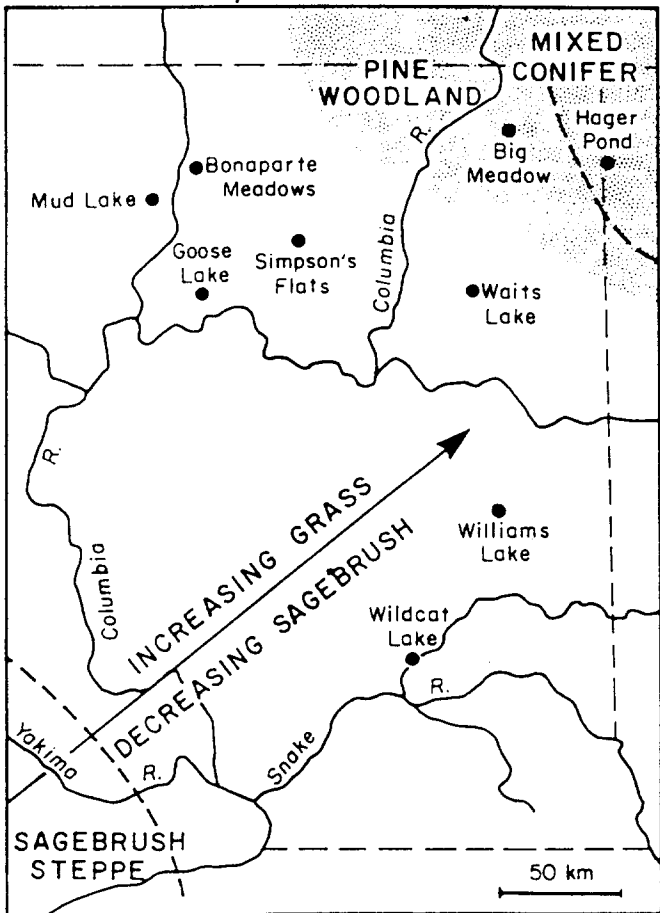


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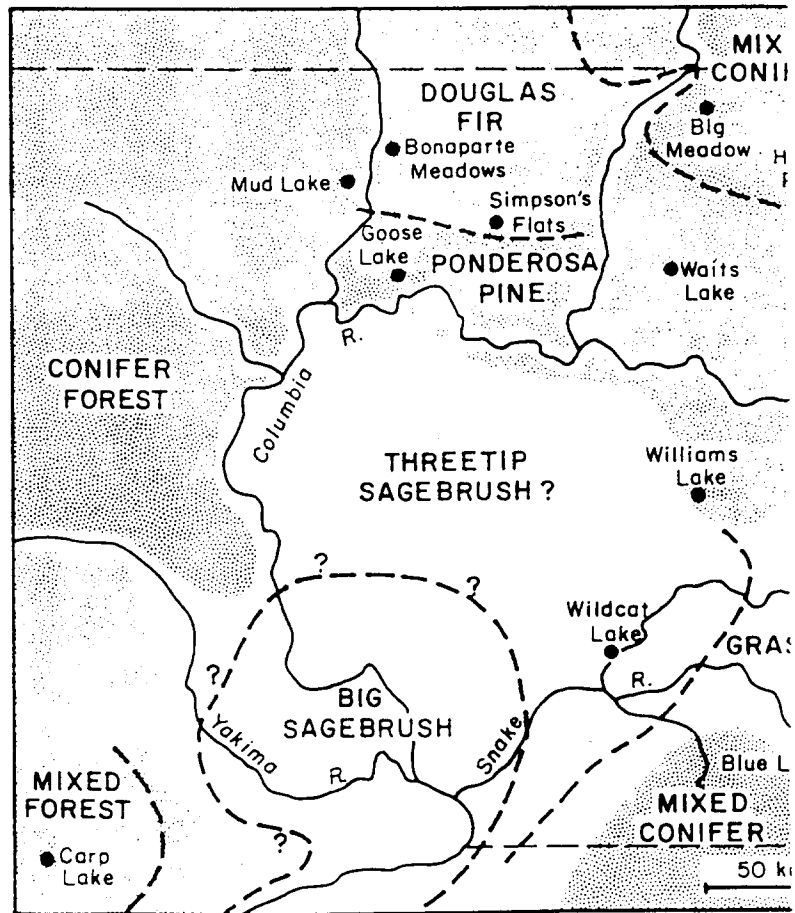
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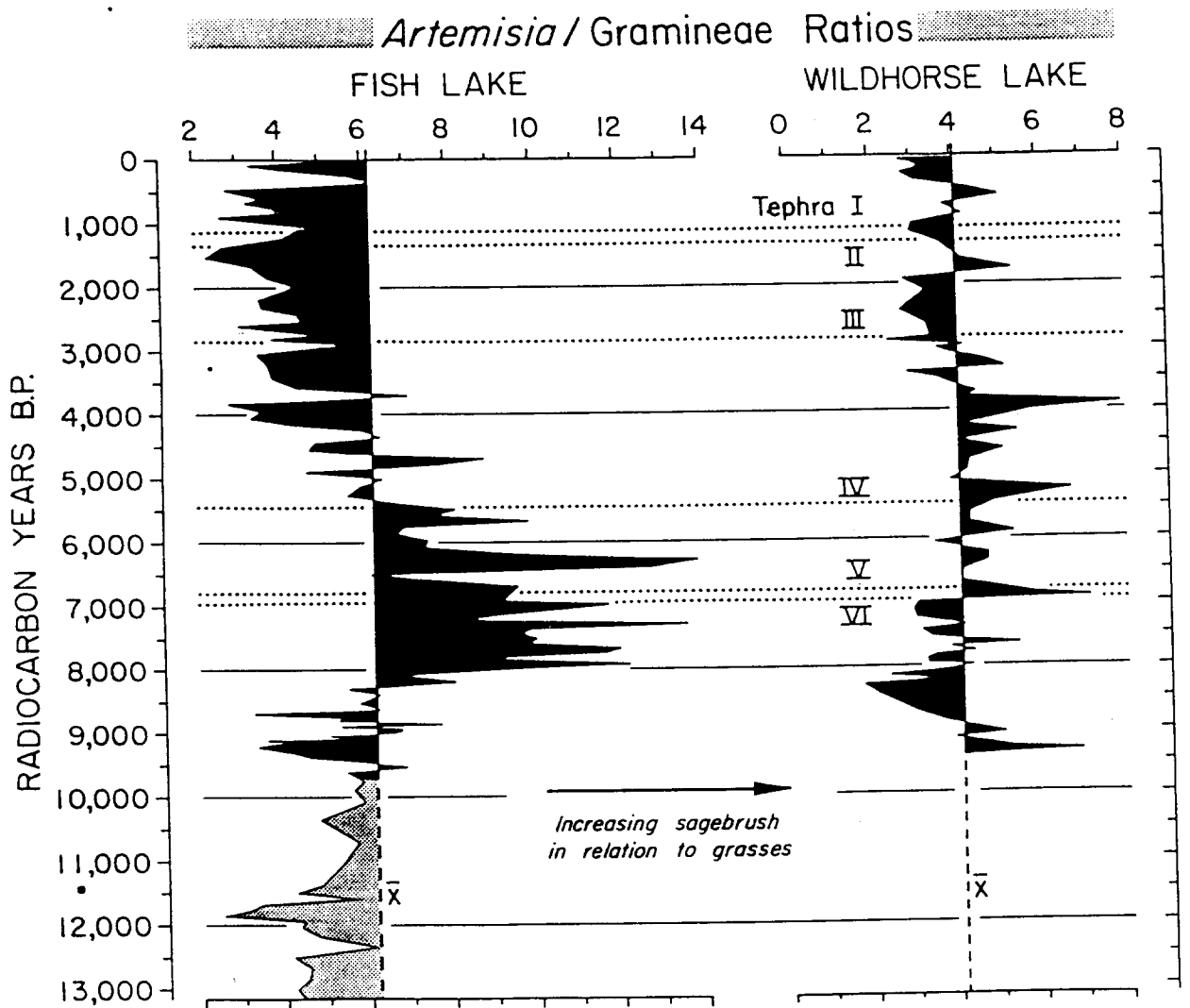


B. 10,000-9000 B.P.

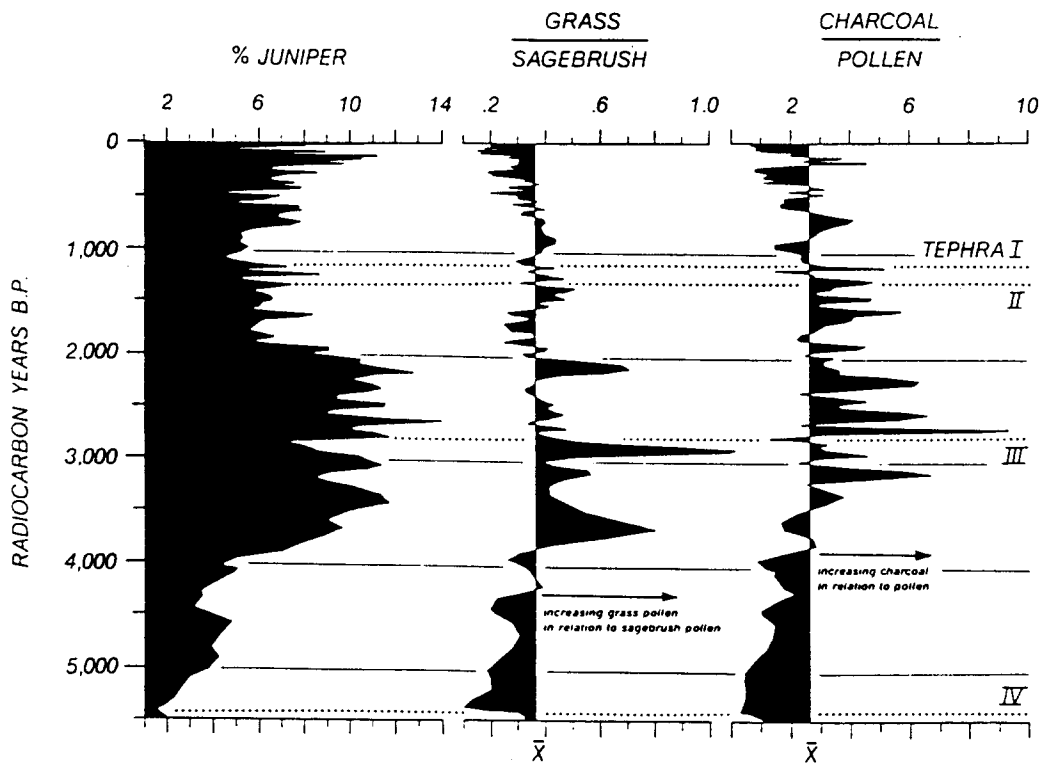


C. 3500-2500 B.P.









**Diamond Pond**

O - Juniper Seeds

