Quaternary Research 58, 215–225 (2002)
doi:10.1006/qres.2002.2378

Fire and Vegetation History from the Coastal Rain Forest of the Western Oregon Coast Range

Colin J. Long[1] and Cathy Whitlock

Department of Geography, University of Oregon, Eugene OR 97403

Received February 5, 2002

High-resolution charcoal and pollen analyses were used to reconstruct a 4600-yr-long history of fire and vegetation near Taylor Lake in the wettest forests of coastal Oregon. Today, fires in these forests are rare because the season of ignition does not coincide with months of dry fuels. From ca. 4600 to 2700 cal yr B.P. fire episodes occurred at intervals of 140 ± 30 yr while forest vegetation was dominated by disturbance-adapted taxa such as *Alnus rubra*. From ca. 2700 cal yr B.P. to the present, fire episodes have become less common, occurring at intervals of 240 ± 30 yr, and fire-sensitive forest taxa, such as *Tsuga heterophylla* and *Picea sitchensis*, have become more prominent. Fire occurrence during the mid-Holocene was similar to that of the more xeric forests in the eastern Coast Range and suggests that summer drought was widespread. After ca. 2700 cal yr B.P., a decrease in fire episode frequency suggests that cooler conditions and possibly increased summer fog allowed the establishment of present-day *Picea sitchensis* forests within the watershed. These results provide evidence that fire has been an important disturbance agent in the Coast Range of Oregon, and variations in fire frequency and climate have led to the establishment of present-day forests.

Key Words: Pacific Northwest; fire history; pollen records; millennial-scale climate change; paleoecology.

INTRODUCTION

Picea sitchensis forests extend from the coastal mountains of southeastern Alaska to the southern Coast Range of Oregon (Veblen and Alaback, 1996). They are the wettest forests in North America, are characterized by dense stands of *Tsuga heterophylla* (western hemlock), *Picea sitchensis* (Sitka spruce), and *Alnus rubra* (red alder), and have some of the highest growth rates in North America (Grier, 1978). The value of these forests in timber production and as unique environments to a variety of species has brought these forests to public attention during the 20th century. However, little is known of the time of their establishment or their disturbance history. It is generally thought that wind storms, which create gaps in the forest canopy as the result of blown-down trees, are the primary disturbance. These gaps provide an opportunity for colonization by new individuals (Ruth and Harris, 1979; Harcombe, 1986; Lertzman *et al.*, 1996).

The role of fire as a disturbance agent is poorly understood in *Picea sitchensis* forests of the Pacific Northwest (PNW). One source of fire history information comes from tree-ring records but such data have been difficult to acquire because wood decay rates in the Coast Ranges of Oregon and Washington exceed all other forests in the PNW (Ruth and Harris, 1979) and fire scars on tree rings are not preserved well in the dominant trees (Grier, 1978). Another difficulty is that fires have caused high tree mortality, and evidence of older fires is confined to only a few surviving individuals (Agee, 1993; Impara, 1997). The age of a cohort of trees (stand-age class data) is often used to establish the dates of past fires. Stand-age class data from western Oregon and Washington have been used to infer a mean fire interval (MFI), or the average time between fires, of between ca. 400 yr (Andrews and Cowlin, 1940) and ca. 1100 yr (Agee and Flewelling, 1983) for *Picea sitchensis* forests in the PNW. Fires described in historical journal accounts confirm that large, infrequent, stand-replacing (i.e., high severity), fires have occurred during the last few centuries (Morris, 1934; Juday, 1976; Agee, 1993). For example, the Nestucca Fire of A.D. 1848 burned approximately 120,000 ha and the Tillamook fire of A.D. 1933 burned approximately 106,000 ha of *Picea sitchensis* and *Tsuga heterophylla* forests of the northern Oregon Coast (Munger, 1944). Whether this fire regime of widespread, infrequent, high-severity fires has persisted over time in *Picea sitchensis* forests requires information on the long-term fire history and a better understanding of the relationship between fire occurrence, climate, and forest responses in these extremely mesic ecosystems.

We report on the fire and vegetation history at Taylor Lake (Lat. 46°06′02″N, Long. 123°54′24″W, elev. 4 m.) located in the *Picea sitchensis* forest on the northern Oregon coast (Fig. 1). Taylor Lake lies on the Clatsop Plain, an area of sand aggradation south of the mouth of the Columbia River (Orr *et al.*, 1992). The lake is 4 ha in size with a maximum water depth of 4.5 m and simple bathymetry and was formed when coastal dunes advanced inland as a result of late-Holocene sea-level rise and dammed a stream in the coastal headland (Rankin, 1983). Continued sand accumulation has moved the shoreline westward ca. 2 km, and eolian activity has created an 800-m-wide deflation plain west of the lake.

[1] To whom correspondence should be addressed. Fax: 541-346-2157. E-mail: clong@oregon.uoregon.edu.

0033-5894/02 $35.00

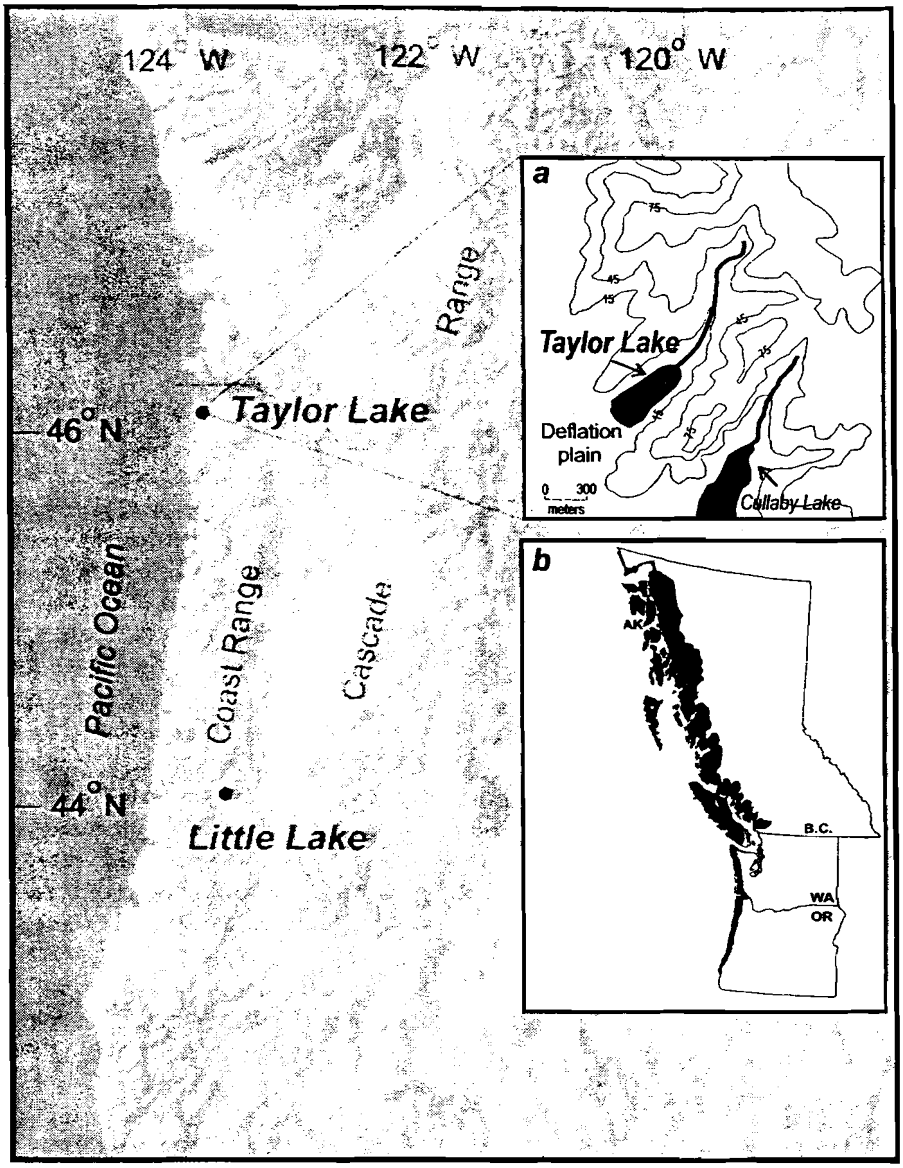

FIG. 1. Location of Taylor Lake and other physiographic features mentioned in the text. Inset *a* shows Taylor Lake and surrounding vicinity. Inset *b* shows the extent of *Picea sitchensis* forests in northwestern North America (after Alaback and Pojar, 1997).

Picea sitchensis forests are distinguished from the inland *Tsuga heterophylla* forests by the diminished presence of *Pseudotsuga menziesii* (Douglas fir) (Franklin and Dyrness, 1988; Alaback and Pojar, 1997). The major conifer taxa of *Picea sitchensis* forests, *Tsuga heterophylla, Picea sitchensis*, and *Thuja plicata* (western red cedar) have higher tolerance for salt spray, higher soil-moisture optimum, and lower tolerance for drought than *Pseudotsuga menziesii* (Minore, 1979). The present forest that surrounds Taylor Lake is typical of *Picea sitchensis* forests with *Tsuga heterophylla, Picea sitchensis*, and *Thuja plicata* as the dominant tree taxa, while *Pseudotsuga menziesii* is a minor constituent. *Alnus rubra* grows in recently disturbed sites and riparian areas. Understory taxa include *Sambucus racemosa* (elderberry), *Rubus spectabilis* (salmonberry), *Gaultheria shallon* (salal), *Polystichum munitum* (sword fern), *Blechnum spicant* (deer fern), and *Dryopteris austriaca* (mountain wood fern). *Acer circinatum* (vine maple) and *Pteridium aquilinum* (bracken) are also present on drier sites. *Salix hookeriana* (Hooker willow) and *Phalaris arundinacea* (Reed canarygrass) are abundant in wetlands on the deflation plain west of the lake. Nomenclature follows Hitchcock and Cronquist (1973).

Temperatures at Taylor Lake average 4.7°C in January and 15.1°C in August, and most of the ca. 2000 mm of precipitation falls as rain between October and May (Taylor, 1993). The A.D. 1961–1998 precipitation data from Seaside, OR (8 km south of Taylor Lake) indicate that approximately 12% (229 mm out of 1987 mm) of yearly precipitation falls between June and September (GHCN, 2000). The climate of the region is influenced by the position of the polar jet stream and the eastern Pacific subtropical high pressure system. In winter, precipitation comes from onshore storms associated with the southward movement of the polar jet stream from southeast Alaska and British Columbia to Washington and Oregon, and the contraction of the Pacific subtropical high southward. Summer conditions are typified by the expansion northward of the Pacific subtropical high as a result of increased summer insolation in the northern hemisphere. Warm conditions at midlatitudes force the polar jet northward from Washington and Oregon to southeast Alaska and British Columbia. The large-scale subsidence associated with the Pacific subtropical high inhibits the formation of precipitation (Redmond and Taylor, 1997). The Taylor Lake watershed experiences drier summers than comparable settings farther north (Veblen and Alaback, 1996). However, coastal summer fog during an otherwise dry season helps maintain moist conditions necessary for *Picea sitchensis* forest assemblages (Ruth and Harris, 1979; Alaback and Pojar, 1997). Fog occurred at Astoria, Oregon, 10 km northeast of Taylor Lake, on 65% (73 of 112 days) of days during the fire season (from July 1 to Oct 20) from A.D. 1995–2000 (NOAA, 2001).

As in other forest types, conditions necessary for fires in *Picea sitchensis* forests are low fuel moisture and an ignition source. Fires are most likely to occur from June through mid-October in conjunction with increased temperature and decreased precipitation. Fuel moisture in coastal forests is typically high throughout the year but may be lowered to flammable levels, 100% of dry weight in live fuels or ca. 20% of dry weight in dead fuels, by the end of the summer drought (Huff and Agee, 1980; Chandler *et al.*, 1983). Strong adiabatically warmed winds from the east can emanate from high-pressure centers that become established east of the Coast Range in summer. These winds dry fuels by lowering local humidity (Schroeder and Buck, 1970; Huff and Agee, 1980). Lightning, while uncommon during the summer, occurs frequently enough during the fire season (July 1–October 20) to provide an occasional ignition source (Agee, 1993; Impara, 1997). Between A.D. 1985 and 1995 an average of 6.18 ± 6.43 positively charged lightning strikes per year occurred in the 1° latitude by 1° longitude area surrounding Taylor Lake (WRCC, 1996). Positively charged lighting strokes are of longer duration and provide more energy to ignite fuels (Fuquay, 1980). In the absence of lightning, human-caused ignitions during dry, east wind conditions have been responsible for the major fires in the Coast Range during the last 100 yr (Morris, 1934; Impara, 1997).

The historical and dendrochronological evidence of fires within the Taylor Lake watershed is scant. Government surveys of the Taylor Lake watershed in A.D. 1856 did not report dead or burned timber (Trutch, 1856), and few if any fires have occurred there in the 20th century. The area was extensively logged in the early A.D. 1900s, and portions of the watershed and adjacent areas were logged in A.D. 1953, 1963, and 1972. Small slash fires accompanied the logging in A.D. 1972, and probably also in 1963 and 1953. Elsewhere, Impara (1997) used fire-scar and stand-age data to identify 27 fires over the last ca. 500 yr in a 137,500 ha transect of the central Oregon Coast Range. The fires ranged from large high-severity fires to small mixed-severity events and suggest that past fires were variable in spatial extent and severity. Teensma *et al.* (1991) used government vegetation surveys and forest stand-age maps to reconstruct preliminary maps of broad-scale patterns of fire for the entire Oregon Coast Range from A.D. 1850 to 1940. Their evidence along with reconstruction of fires using historical documentation by Morris (1934) and Juday (1976) identified large burned areas throughout the Coast Range from A.D. 1849 to 1910.

Conceptual Model for Charcoal Data Analysis

The approach used in this study follows that of Clark and Royall (1996), Long *et al.* (1998), and Millspaugh *et al.* (2000) in that it divides the charcoal record into two components: background and peaks. The background component consists of the low-frequency accumulation of charcoal particles, which varies as a result of changes in (1) fuel composition, (2) inputs of secondary charcoal introduced from the watershed or littoral zone of the lake, or (3) charcoal introduced from fires outside the watershed. The peak component is a higher frequency signal composed primarily of the charcoal produced by a single fire or series of closely spaced fires (hereafter termed a fire episode) in the watershed or sometimes in adjacent watersheds (Millspaugh and Whitlock, 1995; Whitlock and Millspaugh, 1996). A locally weighted mean function (Cleveland, 1979) is used to determine the background component. The peak component is identified as positive deviations from background that exceed a "threshold-ratio value." The selection of the locally weighted window width and the threshold-ratio value is based on two factors. The first is the estimated fire-return intervals for the area. The second is the correlation of the uppermost peaks in charcoal to times of known or suspected fires within the catchment. The fire episode frequency is calculated by averaging peak occurrence over time.

METHODS

High-resolution charcoal analysis was used to reconstruct the fire history, measurements of sediment magnetic-susceptibility provided information on the sedimentation history, and pollen analysis was the basis for reconstructing the vegetation history at Taylor Lake.

Field Collection

A 0.45-m-long short core and a 3.43-m-long core were collected within 5 m of each other from the deepest part of the lake. The short core was obtained with an 8-cm-diameter gravity sampler that preserved the mud–water interface intact. The core was extruded in the field at 1-cm intervals and stored in plastic bags. The long core was obtained with a 5-cm-diameter modified Livingstone sampler. Core segments were extruded in the field wrapped in cellophane and aluminum foil and transported to the laboratory where they were refrigerated along with the short-core samples.

TABLE 1

^{210}Pb Dates[a] for the Short Core and Calibrated and Uncalibrated ^{14}C Radiocarbon Ages and Age Model[b] for Long Core from Taylor Lake

Depth (cm)	Age (A.D.)	Error of age ±1 SD	Depth (cm)	Age (A.D.)	Error of age ±1 SD
1–2	1995	1.4	14–15	1953	12.9
3–4	1990	6.1	16–17	1945	20.7
4–5	1988	6.4	19–20	1931	23.9
5–6	1985	7.2	20–21	1921	29.6
7–8	1979	7.8	22–23	1916	34.1
9–10	1972	7.9	23–24	1911	37.8
10–11	1969	8.5	25–26	1900	50.7
11–12	1965	9.2	27–28	1889	79.8
13–14	1957	10.6			

Depth (m)	Calibrated age (cal yr B.P. ±2 SD)	Uncalibrated ^{14}C age	Material	Lab no.
0.23[c]	90		Sediment	
0.69–0.72	1060 (951–1161)	1130 ± 40	Charcoal	AA-30945
1.59–1.60	2470 (2348–2741)	2435 ± 60	Charcoal	AA-30946
2.40–2.42	3370 (3264–3465)	3145 ± 45	Charcoal	AA-30947
3.33–3.43	4590 (4574–4802)	4160 ± 70	Sediment	Beta-120001

[a] ^{210}Pb ages provided by D. Edgington, Great Lakes Water Institute, University of Wisconsin—Milwaukee.

[b] Age model: Age (cal yr B.P.) = −238.318 + 1901(depth) − 139.610(depth)^2.

[c] Age based on correlation with ^{210}Pb dated charcoal peak in short core.

Chronology

Seventeen ^{210}Pb-age determinations were used to establish a chronology for the short core (Table 1). Ages were plotted against core depth using linear interpolation to construct an age-versus-depth curve. This analysis indicated that the 45-cm-long short core spanned the last ca. 250 yr. It should be noted that this age model extends beyond the range of ^{210}Pb dating and the age-versus-depth extrapolation was based on the assumption of a constant sedimentation rate. Four accelerator mass spectrometry (AMS) radiocarbon dates were obtained from the long core, calibrated to calendar years, and rounded to the nearest decade (CALIB 4.2, Stuiver *et al.*, 1998) (Table 1). Similar charcoal stratigraphy, including a prominent peak at 0.23-m depth provided a basis for correlation between the cores. The ^{210}Pb chronology from the short core assigned this peak an age of 90 cal yr B.P (A.D. 1911). An age model based on this event and the four calibrated radiocarbon dates indicated that the long-core record spanned ca. 4600 cal yr. Sedimentation rates ranged from 9.05 yr/cm at the top of the core to 18.52 yr/cm at the base, with an average time of 13.4 yr represented in 1 cm of sediment.

Charcoal Analysis

Subsamples of 2.5 cm^3 were taken from contiguous 1-cm intervals for the length of the short and long cores and dissagregated in a 5% solution of sodium hexametaphosphate for 24 h. Samples were then gently washed through nested sedimentological screens of 125- and 250-μm mesh size. All charcoal particles > 125 μm in minimum diameter were tallied, because studies of charcoal deposition after recent fires have shown that this particles size is not transported long distances and thus provides a record of watershed fires (Millspaugh and Whitlock, 1996; Clark and Patterson, 1997; Gardner and Whitlock, 2001). Charcoal counts were converted to concentration data (particles/cm) and then divided by the sample deposition time (yr/cm) to calculate charcoal accumulation rates (CHAR) (particles cm^{-2} yr^{-1}). The CHAR record was then converted to pseudo-annual values and averaged at 10-yr intervals to avoid biases caused by changing sedimentation rates over the length of the core.

In order to assign background window widths and threshold-ratio values, three obvious peaks were identified in the untransformed and log-transformed CHAR data for the last 1000 yr (marked by arrows in Fig. 2) and considered to be fire episodes. The three CHAR peaks were easily distinguished from background by visual inspection, and three fire episodes/1000 yr falls within the range of MFI estimates based on dendrochronological data for modern *Picea sitchensis* forests (Agee and Flewelling, 1983; Agee, 1993). A series of weighted-running means were applied to determine the most appropriate value to use in determining the average or background CHAR for each decade. Window widths of between 300 and 900 yr showed generally similar trends and an intermediate window width 600-yr was selected. A range of threshold-ratio values from 1.00 to 1.30 was evaluated to determine an optimum value for identifying peaks using a 600-yr-background window. Threshold-ratio values between 1.00 and 1.20 identified more than three peaks over that last 1000 yr. A threshold-ratio value of 1.25 was chosen to identify fire episodes above background, because it successfully identified the three peaks of the last 1000 yr but no additional peaks. Peak occurrence was then averaged using a 2000-yr moving window and fire episode frequency was plotted as fire episodes/1000 yr.

Lithologic Analysis

Weight-loss after ignition was used to determine the amount of organic and carbonate material in the sediment at 1-cm intervals in the short core and at 10-cm intervals in the long core (Dean, 1974). Subsamples of 1 cm^3 were placed in crucibles and heated for 2 h at 550°C and for 2 h at 900°C. The weight loss after each burning was used to calculate the percent of organic matter and carbonate content respectively.

Magnetic-susceptibility analysis was used to assess the input of clastic material from the watershed (Thompson and Oldfield, 1986). Fires can defoliate and destabilize slopes (Swanson, 1981; Benda, 1994) and heat soils (Longworth *et al.*, 1979), which can increase the amount of paramagnetic minerals washed into lakes (Gedye *et al.*, 2000). These processes can increase the magnetic susceptibility of lake sediments (Millspaugh and Whitlock, 1995). Other events unrelated to fires, such as variations in

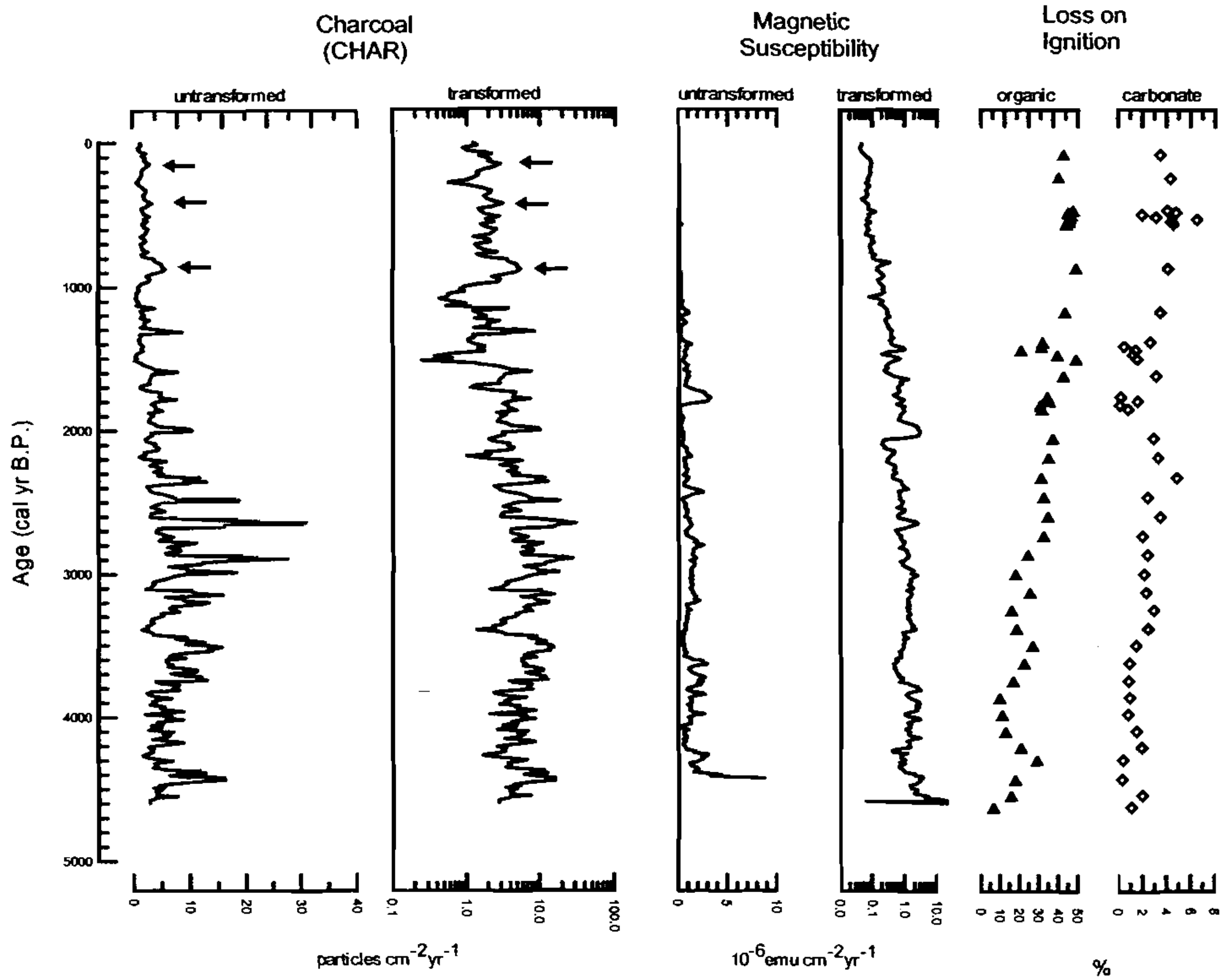

FIG. 2. Untransformed and log-transformed charcoal (CHAR), magnetic susceptibility accumulation rates, and data from loss on ignition analysis plotted against age of the long core. CHAR peaks assumed to be watershed fires in the last 1000 yr are marked with arrows (←).

surface runoff, stream flow, and mass movements within the catchment can also produce changes in clastic input that are reflected in the magnetic susceptibility of lake sediment (Dearing and Flower, 1982).

Magnetic susceptibility readings were taken on an 8-cm^3 subsample from each 1-cm interval prior to charcoal analysis. Subsamples were put into a plastic cup and placed in a cup-coil magnetic susceptibility instrument by Sapphire Instruments. Sediment magnetic-susceptibility values were converted to concentration values, expressed as electromagnetic units per cubic centimeter (emu/cm^3), and divided by the sample deposition time (yr/cm) to calculate magnetic-susceptibility accumulation rates (emu cm^{-2} yr^{-1}) (Fig. 2). These data were then analyzed in the same way as the CHAR data, with peak intervals interpreted as pulses of allochthonous input into the lake, hereafter referred to as a sedimentation episode. The conversion of bulk magnetic susceptibility values to accumulation rates was done to facilitate comparison with the CHAR record.

Pollen Analysis

Pollen analysis was undertaken at 10-cm intervals to reconstruct the vegetation at ca. 135-yr intervals. One-cm^3 samples were processed for analysis with standard methods (Faegri *et al.*, 1989). Pollen was identified to the lowest taxonomic level possible based on modern pollen collections at the University of Oregon and published atlases (Erdtman, 1969; Moore *et al.*, 1991). A minimum of 400 terrestrial grains were identified. Haploxylon-type and Diploxylon-type *Pinus* grains were assigned to *P. monticola* (western white pine) and *P. contorta* (lodgepole pine), respectively, based on their present coastal distribution. *Pseudotsuga*-type pollen was attributed to *P. menziesii*

(Douglas fir), and *Picea* pollen was assumed to represent *P. sitchensis*. *Abies* pollen was attributed to *A. grandis* (grand fir) and *A. amabilis* (Pacific silver fir), both of which grow above 1000 m elevation in the northern Oregon Coast Range as well as in the Cascade Range. Cupressaceae pollen was attributed to *Thuja plicata* (western red cedar) and *Chamaecyparis nootkatensis* (yellow cedar), which occurs rarely in the Oregon Coast Range, and is also a possible contributor. Pollen grains that could not be identified were labeled "Unknown." Terrestrial percentages were based on the sum of arboreal and nonarboreal pollen and spores, excluding *Polystichum*-type spores which overwhelmed the pollen percentages. Pollen zones were constructed based on results of a constrained cluster analysis of arboreal pollen percentages (CONISS, Grimm, 1988). Assemblages were then compared to modern pollen rain from different vegetation types to infer past vegetation and climate (Pellatt *et al.*, 1997; Minckley and Whitlock, 2000).

RESULTS

Core Description

The long core consisted of dark brown (10YR 2/2) fine-to-medium detritus gyttja with dark gray (5Y 3/1) clay lenses 1–2 cm in thickness. The bottom 2 cm consisted of a sand layer with small pebbles. Organic content averaged 18% between 3.41 and 2.00 m depth then increased to 38% from 2.00 m depth to the top of the core. Carbonate content averaged 2% (Fig. 2).

Historical CHAR and Magnetic Susceptibility Records

Two CHAR peaks, dated at A.D. 1973 and 1911, were identified by visual inspection of the short core CHAR record (Fig. 3). The peak centered on A.D. 1973 is likely associated with a slash burn following the logging of a 200-ha area on the eastern side of the watershed in the early A.D. 1970s. The peak at ca. A.D. 1911 may be the result of a local undocumented fire, on the grounds that many fires took place in A.D. 1910 in western Oregon (Morris, 1934; Juday, 1976). The associated CHAR peaks indicate that significant peak values are 2 to 3 times larger than the background levels. The magnetic susceptibility record shows low values from A.D. 1770 to 1900 (Fig. 3). Values rise at ca. A.D. 1900, reach highest levels at ca. A.D. 1906 and 1925, and then decline to present-day values. Increases in sediment magnetism after A.D. 1900 may be associated with increased erosion from logging and slash fires. Magnetic susceptibility values from the early A.D. 1900s to the present are higher than those prior to A.D. 1900, suggesting that logging or other anthropogenic activities within the catchment has provided continuous clastic input.

Prehistoric CHAR and Magnetic Susceptibility Records

The CHAR record from the long core was divided into two zones based on visual inspection of trends in peak frequency and background CHAR. The first zone TL1 (ca. 4600–2700 cal yr B.P.) is characterized by relatively high fire episode frequency and background CHAR, and the second zone TL2 (ca. 2700 cal yr B.P. to present) features a decrease in fire episode frequency and background CHAR levels (Fig. 4). Six to seven fire episodes/1000 yr occurred during Zone TL1 and then declined to three episodes/1000 yr in Zone TL2. The MFI, expressed as the average time between peaks, indicate that fire episodes in Zone TL1 occurred every 140 ± 30 yr and increased to 230 ± 30 yr in Zone TL2. Although the difference in the MFI between Zone TL1 and Zone TL2 is not statistically significant at $p = 0.05$, the trend in the data is clear. The mid-Holocene period (Zone TL1) experienced more frequent fires than the late Holocene period (Zone TL2), with the last 1000 yr recording the fewest fires in the record.

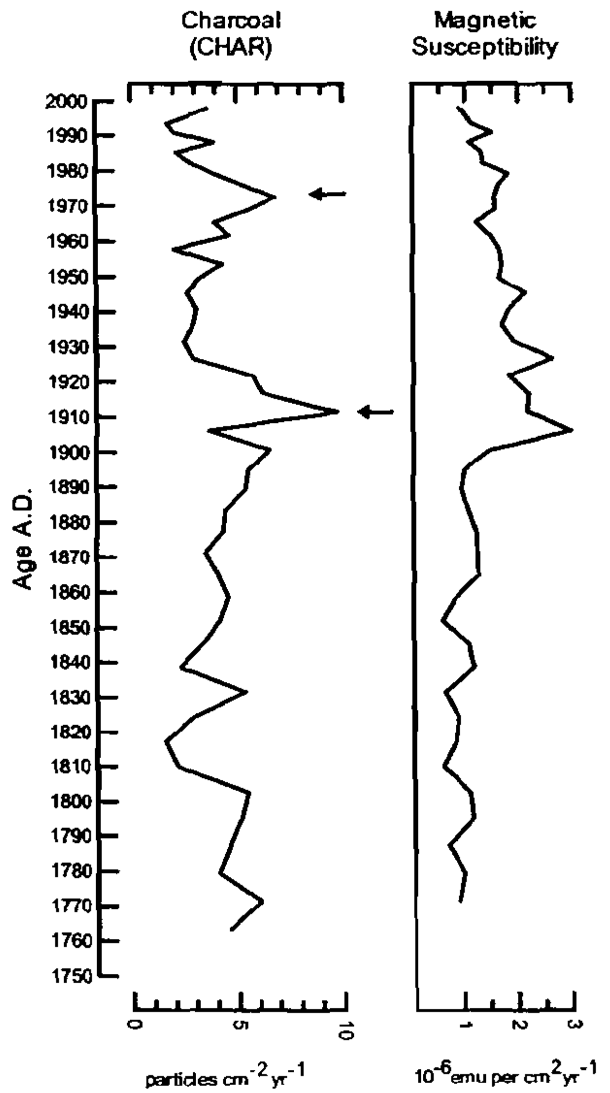

FIG. 3. Log-transformed charcoal (CHAR) and magnetic susceptibility accumulation plotted against age for the short core. Core intervals marked (←) are associated with known or suspected watershed fires in A.D. 1972 and 1911.

The change in background CHAR values between Zones TL1 and TL2 were not statistically significant but did display trends over time. CHAR values were low (5 particles cm^{-2} yr^{-1}) at 4600 cal yr B.P., then increased to moderate levels (9 particles cm^{-2} yr^{-1}) at ca. 2700 cal yr B.P., and declined to present-day values of 1–2 particles cm^{-2} yr^{-1} by ca. 1500 cal yr B.P. The increase in background CHAR between ca. 4600 to 2700 cal yr B.P. suggests that charcoal production and delivery to the lake was greater than at present. This increase could also result from redeposition of sequestered charcoal in the watershed or littoral zone, but the higher fire occurrence during this period suggests that the background CHAR values reflect greater charcoal production. The decline in background CHAR values after ca. 2700 cal yr B.P. indicates less charcoal produced during a time of few fires.

Peaks in the magnetic susceptibility record are more frequent early in the record and could be the result of local mass movement associated with severe storm or winter-related precipitation (Reneau and Dietrich, 1990) and also greater slope destabilization as a result of high fire frequency (Swanson, 1981; Benda, 1994) (Fig. 4). Six to seven sedimentation episodes/1000 yr occurred from ca. 4600 to 3600 cal yr B.P. The frequency of sedimentation episodes declined to between three and four episodes/1000 yr by ca. 2700 cal yr B.P., and three to four episodes/1000 yr were recorded in the last ca. 2700 yr. Background magnetic susceptibility records showed moderate values (5.0×10^{-6} emu cm^{-2} yr^{-1}) from 4600 to 3700 cal yr B.P., and low values from 3700 cal yr B.P. to the present (between 0.075 and 1.25×10^{-6} emu cm^{-2} yr^{-1}).

A Pearson cross-correlation between peaks in CHAR and magnetic susceptibility records showed little correlation ($r = -0.026$; $p = 0.575$). This lack of correlation suggests that charcoal peaks are not associated with fire-related erosion episodes. Benda (1994) proposed that several years may lapse between a fire and slope destabilization in the Oregon Coast Range. To examine the possibility of a lagged response in the Taylor Lake record, magnetic susceptibility peaks and CHAR peaks were compared with lags of 10, 20, 30, 40, and 50 yr. The correlation did not improve, the highest being a 20-yr time lag ($r = 0.085$; $p = 0.069$), suggesting that sedimentation events were not associated with fires in a systematic way. The weak correlation between fire occurrence (peaks in CHAR) and erosional events (peaks in magnetic susceptibility) is similar to that found at another Coast Range site (Long *et al.*, 1998). Other processes,

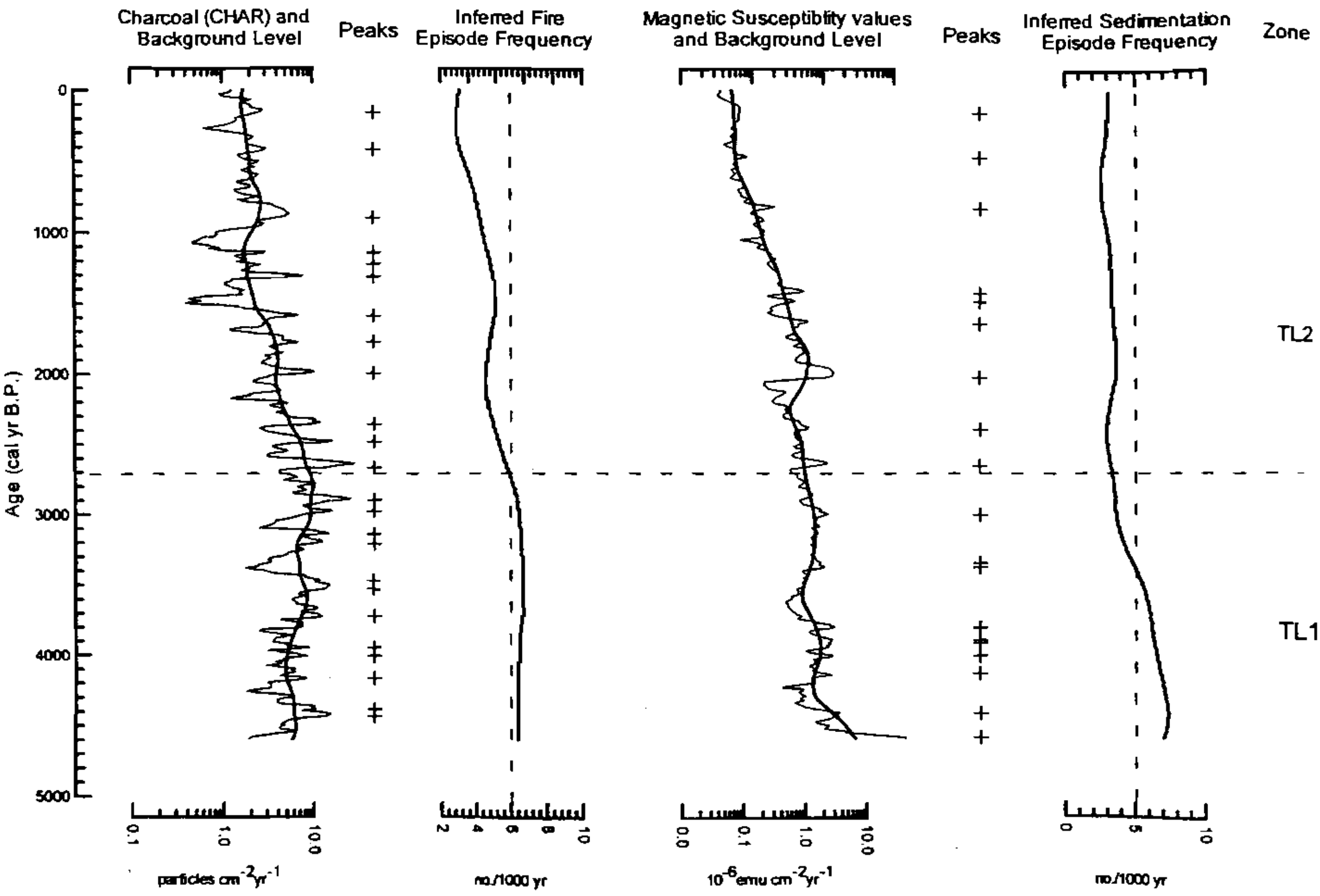

FIG. 4. Log-transformed charcoal (CHAR) and magnetic susceptibility accumulation values, background levels, peaks, and inferred fire frequency and sedimentation event frequency for the Taylor Lake long core. A 600-yr background window width and a 1.25 threshold-ratio value were used to determine background and peak series. The horizontal line marks the boundary between pollen zones TL1 (ca. 4600–2700 cal yr B.P.) and TL2 (ca. 2700 cal yr B.P.–present).

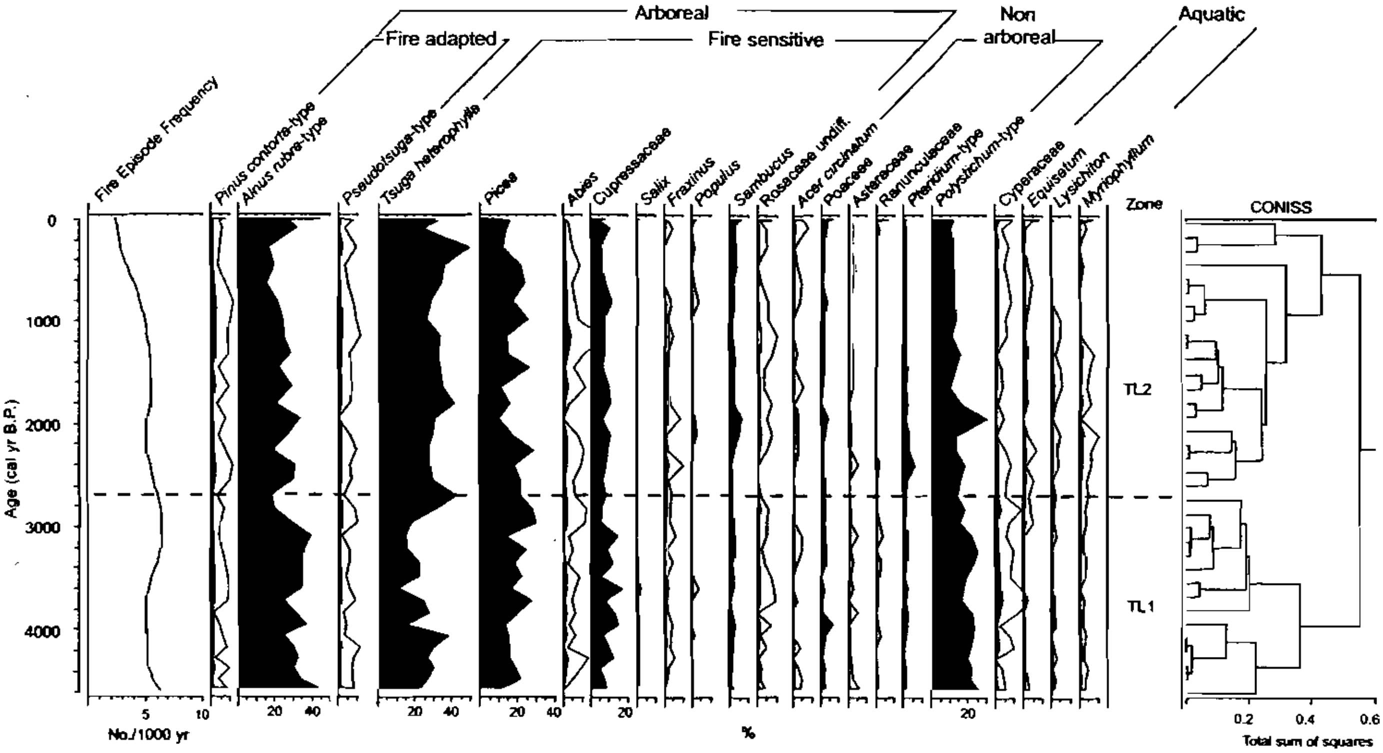

FIG. 5. Inferred fire epsiode frequency, pollen percentages, and constrained cluster analysis (CONISS, Grimm, 1988) from Taylor Lake long core. Open curves represent a 5 times exaggeration of black curves. Arboreal taxa are grouped as fire adapted or fire sensitive (after Minore, 1979, and Agee, 1993).

such as precipitation events, stream erosion, or earthquakes, may serve as primary triggers of sedimentation events.

Vegetation Reconstruction

The period from ca. 4600 to 2700 cal yr B.P. (Zone TL1) was characterized by relatively high percentages of *Alnus rubra*–type (25–44%) *Tsuga heterophylla* (10–38%), *Picea* (14–30%), and Cupressaceae (5–17%) pollen. *Pseudotsuga*-type, *Pinus contorta*-type, *Abies, Sambucus*, and *Salix* pollen occurred in smaller amounts (<3% each) (Fig. 5). *Polystichum*-type spores were the most common nonarboreal species, with lesser amounts of Poaceae and *Pteridium*-type spores. The high amounts of *Alnus* pollen and *Polystichum*-type spores for this period suggest a more open forest than at present. Based on comparisons with modern pollen samples from the *Picea sitchensis* forests (Pellatt *et al.*, 1997; Minckley and Whitlock, 2000), the vegetation was composed of stands of *Alnus rubra* on disturbed sites and riparian areas and upland forests of *Tsuga heterophylla, Picea sitchensis*, and *Thuja plicata*. The understory consisted of *Polystichum munitum* (the likely contributor of *Polystichum*-type spores) with *Pteridium aquilinum* on drier sites or open areas. *Salix hookeriana* (a probable contributor of *Salix* pollen) and grasses covered the deflation plain to the west. Interpretation of the pollen data from ca. 4600 to 2700 cal yr B.P. (TL1) suggests a cool moist climate which is similar with regional climate reconstructions for the mid-Holocene (Whitlock 1992, Worona and Whitlock 1995).

The period from ca. 2700 cal yr B.P. to present (Zone TL2) showed an overall increase in *Tsuga heterophylla* (17–49%) and a decrease in *Alnus rubra*–type (16–44%) with *Picea* (6–28%) and Cupressaceae (5–11%) percentages remaining about the same as Zone TL1 (Fig. 5). *Pseudotsuga*-type, *Pinus contorta*-type, *Abies*, and *Salix* contributed a minor component (<3% each) to the pollen rain. Nonarboreal taxa, such as Poaceae, *Pteridium*-type, and *Polystichum*-type, were most common. This zone marks the establishment of closed *Picea sitchensis* and *Tsuga heterophylla* forest and suggests increasing year-round effective moisture in the last three millennia. This is similar with regional climate reconstructions for the late-Holocene (Whitlock 1992); however pollen evidence from an inland site, Little Lake, located 240 km southeast in drier *Tsuga heterophylla* forests suggest a slight warming over the last ca. 2700 cal yr B.P. (Worona and Whitlock, 1995; Long *et al.*, 1998). The rise in *Alnus rubra*-type and decrease in *Tsuga heterophylla* and *Picea* percentages at the top of the zone are a reflection of disturbance from logging activities over the last century.

DISCUSSION

Late Holocene Fire History, Vegetation, Climate Reconstruction

Changes in Holocene climate in PNW are inferred from a network of vegetation reconstructions based on fossil pollen and plant macrofossil data. The major forcing mechanism of climate change in this region is the long-term variations in the seasonal cycle of insolation generated by changes in the timing of perihelion and tilt of the Earth's axis (Berger and Loutre, 1991; Kutzbach *et al.*, 1993). Greater-than-present summer insolation in western North America and the PNW between ca. 14000 and 8000 cal yr B.P. increased summer temperatures as well as strengthened the influence of the eastern Pacific subtropical high in summer. The result of these shifts was a warmer drier climate in the PNW than today (Whitlock and Grigg, 1999). Decreased summer insolation since ca. 8000 cal yr B.P. has weakened the eastern Pacific subtropical high and resulted in the onset of cooler wetter conditions in the PNW (Whitlock, 1992; Thompson *et al.*, 1993). Variations in climate on submillennial timescales that affect the PNW, such as El Niño/Southern Oscillation (ENSO) (Diaz and Markgraf, 1998) and the Pacific Decadal Oscillation (Biondi *et al.*, 2001), may be responsible for centennial or decadal variations in precipitation. Decadal droughts affect fuel moisture within forests, and thus fire occurrence. The Taylor Lake record matches well with the general pattern of regional climate changes. The shift from open to closed forest over the last 4600 yr indicates an increase in effective moisture (Mathewes, 1985; Thompson *et al.*, 1993). Frequent fires between ca. 4600 and 2700 cal yr B.P. at Taylor Lake suggest greater seasonal drought than today, and forest stands were dominated by disturbance-adapted *Alnus rubra* and an understory of abundant *Polystichum munitum* and *Sambucus racemosa* (Fig. 5). A MFI of 140 ± 30 yr during this period is similar to that at Little Lake located 240 km southeast in more xeric *Tsuga heterophylla* forest (Long *et al.*, 1998). After 2700 cal yr B.P., interannual drought may have been less regular, allowing fire intervals to increase to 230 ± 30 yr. Longer periods between fires allowed shade-tolerant and fire-sensitive species, such as *Tsuga heterophylla*, to become more prominent within the forest. The decrease in seasonal drought at Taylor Lake may be related to increased summer fog after ca. 2700 cal yr B.P. If climate conditions became slightly warmer and drier inland, as suggested by paleoecological data from Little Lake (Worona and Whitlock, 1995; Long *et al.*, 1998), summer fog would have become a regular feature of the coastal climate and kept local conditions cooler and wetter. The increase in *Tsuga heterophylla* and *Picea sitchensis* is consistent with more humid conditions.

Increased anthropogenic fires may also explain the higher-than-present fire frequencies from ca. 4600 to 2700 cal yr B.P. People have inhabited the Oregon Coast for at least the last 5000 yr, and as many as 30,000 people occupied the coastal strip of Oregon and Washington at the time of European contact (Suttles and Ames, 1997). Disease introduced in the early A.D. 1700s led to a rapid decline in Native American populations (Ubelaker, 1988), and the Lewis and Clark expedition estimated that only 1200 Natives lived in the vicinity of Fort Clatsop, 8 km northeast of Taylor Lake, in A.D. 1804 (Clark, 1990). Archeological evidence suggests that coastal subsistence activities did not require the deliberate use of fire inasmuch as fishing and marine invertebrates were the primary food sources (Boyd, 1999). It is possible that an occasional brush fire turned into a larger forest fire, as has occurred frequently in the last 100 yr with slash burns, but if anthropogenic activity were the sole explanation for fires during this period, it is not clear why fires became less frequent after ca. 2700 cal yr B.P. when Native populations were still high. Increased summer moisture after ca. 2700 cal yr B.P. seems to be the primary explanation for decreased fire activity regardless of the ignition source.

CONCLUSIONS

The Taylor Lake record shows that fire is an important agent of disturbance in the southern *Picea sitchensis* forests and that fire occurrence has changed with climate change over the last 4600 yr. From ca. 4600 to 2700 cal yr B.P. frequent fires helped maintain *Alnus rubra* and other disturbance-adapted taxa in the watershed, and the decrease in fire occurrence over the last 2700 yr has led to a closing of the forest, perhaps caused by protracted summer fog and shorter summer drought.

An increase in regional summer temperatures in the future is projected as a result of higher greenhouse gas concentrations in the atmosphere (IPCC, 2001). In the PNW, model simulations of future climate conditions suggest that summer temperatures will increase (Leung and Ghan, 1999). Warmer summer conditions could increase the advective fog formation along the coastal margin and keep summer conditions moist and retard fire activity. Contrary to projections of increased fire activity in the future in the interior U.S. (Franklin *et al.*, 1992; Price and Rind, 1994; Bartlein *et al.*, 1997) along the coast the present-day regime of infrequent fires may persist despite a warmer climate.

ACKNOWLEDGMENTS

We thank R. Strickland, C. A. Pearl, D. Pickering, A. Brunelle, and M. Power and for assistance in the field. R. S. Anderson, R. Mathewes, and D. Hallett provided helpful comments on the manuscript. Research was supported by National Science Foundation Grants (EAR-9906100 and ATM-0117160) to C. Whitlock.

REFERENCES

Agee, J. K. (1993). "Fire Ecology of Pacific Northwest Forests." Island Press, Washington, DC.

Agee, J. K., and Flewelling, R. (1983). A fire cycle model based on climate for the Olympic Mountains, Washington. *In* "Seventh Conference Fire and Forest Meteorology, April 25–28, 1983, Ft. Collins Colorado" (J. Means, Ed.), pp. 32–37. Am. Meteorol. Soc., Boston.

Alaback, P., and Pojar, J. (1997). Vegetation from ridgetop to seashore. *In* "The Rain Forests of Home: Profile of a North American Bioregion" (P. K. Schoonmaker, B. von Hagen, and E. C. Wolf, Ed.), pp. 69–88. Island Press, Washington, DC.

Andrews, H. J., and Cowlin, R. W. (1940). "Forest Resources of the Douglas-Fir Region." United States Department of Agriculture. miscellaneous Publication No. 389. Portland.

Bartlein, P. J., Whitlock, C., and Shafer, S. L. (1997). Future climate in the Yellowstone National Park region and its potential impact on vegetation. *Conservation Biology* **11**, 782–792.

Benda, L. E. (1994). "Stochastic Geomorphology in a Humid Mountain Landscape." Unpublished Ph.D. dissertation, University of Washington, Seattle.

Berger, A., and Loutre, M. F. (1991). Insolation values for the last 10 million years. *Quaternary Science Reviews* **10**, 297–317.

Biondi, F., Greshunov, A., and Cayan, D. R. (2001). North Pacific decadal climate variability since 1661. *Journal of Climate* **14**, 5–10.

Boyd, R. (1999). Introduction. *In* "Indians, Fire and the Land in the Pacific Northwest" (R. Boyd, Ed.), pp. 1–30. Oregon State Univ. Press, Corvallis.

Chandler, C., Cheney, P., Thomas, P., Trabaud, L., and Williams, D. (1983). "Fire in Forestry, Volume I: Forest Fire Behavior and Effects." Wiley, New York.

Clark, J. S., and Patterson, W. A., III (1997). Background and local charcoal in sediments: scales of fire evidence in the paleorecord. *In* "Sediment Records of Biomass Burning and Global Change." (J. S. Clark, H. Cachier, J. G. Goldammer, and B. Stocks, Eds.), pp. 23-48. NATO Advanced Science Institutes Series 1: Global Environmental Change, Vol. 51. Springer, New York.

Clark, J. S., and Royall, P. D. (1996). Local and regional sediment charcoal evi-

Grimm, E. C. (1988). Data analysis and display. *In* "Vegetation History" (B. Huntley and T. Webb, III, Eds.), pp. 43–76. Kluwer Academics, Dordrecht.

Harcombe, P. A. (1986). Stand development in a 130-year-old spruce-hemlock forest based on age structure and 50 years of mortality data. *Forest Ecology and Management* **14**, 41–58.

Hitchcock, C. L., and Cronquist, A. (1973). "Flora of the Pacific Northwest." Univ. of Washington Press, Seattle.

Huff, M. H., and Agee, J. K. (1980). Characteristics of large lightning fires in the Olympic Mountains, Washington. *In* "Proceedings Sixth Conference on Fire and Forest Meteorology" (R. E. Martin, R. L. Edmonds, D. A. Faulkner, J. B. Harrington, D. M. Fuquay, B. J. Stocks, and S. Barr, Eds.), pp. 117–123. Soc. Am. Foresters, Bethesda, MD.

Impara, P. C. (1997). "Spatial and Temporal Patterns of Fire in the Forests of the Central Oregon Coast Range." Unpublished Ph.D. dissertation, Oregon State University, Corvallis.

Munger, T. T. (1944). Out of the ashes of Nestucca: Two sequels to Oregon's Great Nestucca fire of a century ago. *American Forests* **50**, 342–345, 366, 368.

NOAA (2001). Regional Past Weather and Climate Data, Astoria Weather Data 1995–2000. URL: htpl://www.wrh.noaa.gov/portland/climate.

Orr, E. L., Orr, W. N., and Baldwin, E. M. (1992). "Geology of Oregon." Kendall-Hunt, Dubuque, IA.

Pellatt, M. G., Mathewes, R. W., and Walker, I. R. (1997). Pollen analysis and ordination of lake sediment-surface samples from coastal British Columbia. *Canadian Journal of Botany* **75**, 799–814.

Price, C., and Rind, D, (1994). The impact of a 2 × CO_2 climate on lightning-caused fires. *Journal of Climate* 7, 1484–1494.

Taylor, G. H. (1993). Normal annual precipitation, state of Oregon. Oregon State Climate Service, Corvallis, OR.

Teensma, P. D. A., Rienstra, J. T., and Yeiter, M. A. (1991). "Preliminary Reconstruction and Analysis of Change in Forest Stand Age Classes of the Oregon Coast Range from 1850 to 1940." Technical Note T/N OR-9, United States Department of the Interior, Bureau of Land Management, Oregon State Office, Portland.

Thompson, R., and Oldfield, F. (1986). "Environmental Magnetism." Allen and Unwin, London.

Thompson, R. S., Whitlock, C., Bartlein, P. J., Harrison, S. P., and Spaulding, W. G. (1993). Climatic changes in the western United States since 18,000 yr B.P. *In* "Global Climates Since the Last Glacial Maximum" (H. E. Wright,