QUATERNARY RESEARCH **49,** 287–298 (1998)
ARTICLE NO. QR981966

Late-Glacial Vegetation and Climate Change in Western Oregon

Laurie D. Grigg and Cathy Whitlock

Department of Geography, University of Oregon, Eugene, Oregon 97403

Received July 28, 1997

Pollen records from two sites in western Oregon provide information on late-glacial variations in vegetation and climate and on the extent and character of Younger Dryas cooling in the Pacific Northwest. A subalpine forest was present at Little Lake, central Coast Range, between 15,700 and 14,850 cal yr B.P. A warm period between 14,850 and 14,500 cal yr B.P. is suggested by an increase in *Pseudotsuga* pollen and charcoal. The recurrence of subalpine forest at 14,500 cal yr B.P. implies a return to cool conditions. Another warming trend is evidenced by the reestablishment of *Pseudotsuga* forest at 14,250 cal yr B.P. Increased haploxylon *Pinus* pollen between 12,400 and 11,000 cal yr B.P. indicates cooler winters than before. After 11,000 cal yr B.P. warm dry conditions are implied by the expansion of *Pseudotsuga.* A subalpine parkland occupied Gordon Lake, western Cascade Range, until 14,500 cal yr B.P., when it was replaced during a warming trend by a montane forest. A rise in *Pinus* pollen from 12,800 to 11,000 cal yr B.P. suggests increased summer aridity. *Pseudotsuga* dominated the vegetation after 11,000 cal yr B.P. Other records from the Pacific Northwest show an expansion of *Pinus* from ca. 13,000 to 11,000 cal yr B.P. This expansion may be a response either to submillennial climate changes of Younger Dryas age or to millennial-scale climatic variations.

INTRODUCTION

The last deglaciation is characterized in many parts of the world by a series of short-term climatic fluctuations between 16,000 and 10,000 cal yr B.P. These fluctuations are well documented in the North Atlantic region and are thought to be related to variations in the wastage of the Laurentide and Scandinavian ice sheets (Broecker, 1994; Manabe and Stouffer, 1995). The most recent late-glacial oscillation in the North Atlantic was a cooling event between 12,900 and 11,600 cal yr B.P., called the Younger Dryas event (Alley *et al.*, 1993). Recent studies identify a Younger Dryas event not only in the North Atlantic but also in the Pacific, South America, New Zealand, and western North America (Mathewes, 1993; Denton and Hendy, 1994; Kennett and Ingram, 1995; Lowell *et al.*, 1995). The evidence of a Younger Dryas event far from its purported source area raises questions about its global registration, as well as the mechanisms by which late-glacial oscillations were transmitted.

Paleoecological records from the Pacific Northwest have previously lacked the temporal resolution necessary to detect submillennial climatic variations, and evidence of a climatic oscillation during Younger Dryas time has been unclear (Heusser, 1977; Barnosky *et al.*, 1987; Mathewes, 1993). In this study, we describe two detailed pollen records of the late-glacial period from western Oregon and compare them with published data from the Pacific Northwest and northeast Pacific. The western Oregon sites include a low-elevation unglaciated site near the coast and an inland high-elevation glaciated site. A synchronous response at both sites would suggest a regional climatic change. Asynchrony would reflect variations in local climate conditions and environmental factors.

Western Oregon is presently characterized by mild, wet winters and warm, dry summers. Most of the annual precipitation (75–85%) occurs in winter from mid-latitude cyclones originating from the Gulf of Alaska, which are associated with a southward displaced polar jet stream (Mock, 1996). Winter precipitation falls predominantly as rain in the Coast Range and Willamette Valley and as snow in the Cascade Range. In summer, the jet stream shifts northward and the eastern Pacific Subtropical High strengthens, producing warm, dry summers.

A steep precipitation gradient exists from west to east as a result of the orographic effect of the Cascade and Coast Ranges, and vegetation patterns are distributed accordingly (Franklin and Dyrness, 1988). The Coast Range and foothills of the western Cascade Range lie within the *Tsuga heterophylla* Zone (250–1000 m elevation), which is dominated by *Pseudotsuga menziesii* (Douglas fir), *T. heterophylla* (western hemlock), *Thuja plicata* (western red cedar), and *Alnus rubra* (red alder), with minor amounts of *Abies grandis* (grand fir) and *Pinus monticola* (western white pine). The Willamette Valley (120–240 m elevation) features *Quercus garryana* (Oregon white oak) with lesser amounts of *Pseudotsuga, Acer macrophyllum* (bigleaf maple), *A. rubra, Corylus cornuta* (western hazel), and *Fraxinus latifolia* (Oregon ash). Higher in the Cascade Range is the *Abies amabilis* Zone (1000–1500 m elevation), which supports *A. amabilis* (Pacific silver fir), *T. heterophylla, Abies procera* (noble fir), *Pseudotsuga, T. plicata,* and *P. monticola.* The highest-

0033-5894/98 $25.00

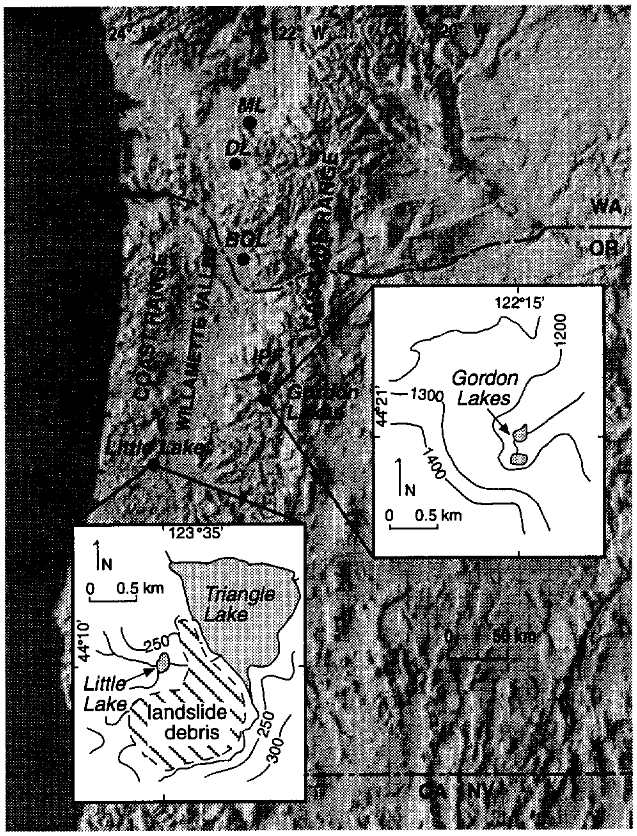

FIG. 1. Map showing Oregon study sites and other sites in Washington (WA) and Oregon (OR) used for regional comparisons (ML, Mineral Lake; DL, Davis Lake; BGL, Battle Ground Lake; IPF, Indian Prairie Fen; CA, California; NV, Nevada). Insets show topographic maps of Little Lake and Gordon Lake, with elevations in meters.

elevation forests comprise the *Tsuga mertensiana* Zone (1500-2000 m elevation) and are composed of *T. mertensiana* (mountain hemlock), *A. amabilis,* and minor amounts of *Picea engelmannii* (Engelmann spruce), *A. procera, P. monticola, Pseudotsuga,* and *Chamaecyparis nootkatensis* (Alaska-cedar). In dry settings, *Abies lasiocarpa* (subalpine fir) and *Pinus contorta* (lodgepole pine) are also components of high-elevation forests. Open subalpine parkland of *Pinus albicaulis* (whitebark pine), *A. lasiocarpa, T. mertensiana, P. contorta,* and alpine herb species are found between 1900 and 2200 m elevation.

Little Lake (Lat. 44°10′N, Long. 123°35′W, 217 m elevation), which has a surface area of 1.5 ha (Fig. 1), is located in the central Coast Range 45 km east of the Pacific Ocean. It was formed ca. 42,000 cal yr B.P. by a landslide that dammed Little Creek (Fig. 1). January and July averaged temperatures from Noti, Oregon, 14 km to the southeast, are 4.4° and 18.4°C and mean annual precipitation is 1500 mm (NOAA Climatological Data Annual Summary 1964–1991). Little Lake lies within the *Tsuga heterophylla* Zone, but the local forest is composed primarily of second-growth *Pseudotsuga.* The lake sediments have provided a pollen record of the past 42,000 cal yr B.P. (Worona and Whitlock, 1995).

Gordon Lake (Lat. 44°21′N, Long. 122°15′W, 1177 m elevation) is located in the western Cascade Range (Fig. 1) in a small cirque of presumed late Pleistocene age. The surface area of Gordon Lake is 2.6 ha. January and July averaged temperatures from a nearby automated climate station (Lat. 44°23′N, Long. 122°10′W, 1067 m elevation) are 0.83° and 15.06°C and mean annual precipitation is 2134 mm (K. T. Redmond, pers. commun., 1997). Gordon Lake lies within the *Abies amabilis* Zone. The local forest contains *A. amabilis, T. heterophylla, Pseudotsuga,* and *A. procera* and minor amounts of *T. plicata* and *P. monticola.*

METHODS

Cores from Little and Gordon lakes were obtained with a 5-cm-diameter piston sampler (Wright *et al.,* 1983). Samples of 1 cm^3 were taken every 4 to 8 cm (average of 177-yr intervals) from Little Lake and every 4 to 8 cm (average of 154-yr intervals) from Gordon Lake. Samples were processed for pollen analysis with standard methods (Cwynar *et al.,* 1979; Faegri *et al.,* 1989). Known amounts of *Lycopodium* spores were added to each sample to calculate pollen concentration (grains/cm^3). Pollen was examined at 400 and 1000× magnification and identified to the lowest taxonomic level possible based on modern pollen collections and published atlases (McAndrews *et al.,* 1973; Moore and Webb, 1978). Pollen grains not assigned to a taxon were called "Unknown" or "Indeterminate" if corroded, degraded, torn, or hidden. One thousand to 2000 terrestrial pollen grains were counted for each sample. Terrestrial pollen percentages were based on a sum of terrestrial pollen and spores. Percentages of aquatic taxa were calculated with the total sum of pollen and spores. Pollen accumulation rates (PAR: grains/cm^2/yr) were determined by dividing pollen concentrations by deposition time (yr/cm) (Davis *et al.,* 1973). Pollen zones were constructed based on the results of a constrained cluster analysis program (CONISS; Grimm, 1988).

Pinus pollen grains were separated into haploxylon and diploxylon types if the distal membrane was preserved. The haploxylon-type *Pinus* species in Pacific Northwest forests are *P. monticola* and *P. albicaulis.* Their identification was based on modern ecological associations, on phytogeography, and, where possible, by the occurrence of plant macrofossils. Diploxylon-type pollen was attributed to *P. contorta,* but *Pinus ponderosa* is a possible, although unlikely contributor. *Pinus* pollen lacking an intact distal membrane was identified as "Undifferentiated *Pinus.*" *Pseudotsuga*-type pollen was assumed to represent *P. menziesii* rather than *Larix,* which is rare west of the Cascade Range. *Abies* and

TABLE 1
Uncalibrated Radiocarbon Dates and Calibrated Ages for Little Lake and Gordon Lake

Depth (m)	Material dated	Uncalibrated age (^{14}C yr B.P.)	Lab No.	Calibrated age (cal yr B.P.) used in age model and 2σ range
		Little Lake		
3.45–3.55	Lake sediment	2840 ± 70	Beta-48600	2940 (2780–3150)
5.45–5.55	Lake sediment	4260 ± 70	Beta-48601	4840 (4570–4980)
7.74–7.75	Mazama Ash	6845 ± 50	(Bacon, 1983)	7630 (7710–7550)
8.75–8.85	Lake sediment	8270 ± 80	Beta-48602	9250 (9440–8990)
10.35–10.45	Lake sediment	10,790 ± 80	Beta-48603	12,720 (12,900–12,520)
10.78–10.86	Charcoal, wood	10,400 ± 70	AA-20630	12,300 (12,510–11,990)
11.01–11.09	Charcoal, wood	11,320 ± 190	AA-20631	13,230 (13,680–12,850)
11.49–11.55	Charcoal, wood	10,940 ± 180	AA-20632	12,860 (13,230–12,490)
11.65–11.75	Lake sediment	12,720 ± 140	Beta-55124	14,980 (15,490–14,490)
11.94–12.00	Charcoal, wood	11,990 ± 120	AA-20634	13,980 (14,370–13,640)
12.35–12.45	Lake sediment	13,640 ± 390	Beta-48604	16,350 (17,300–15,180)
13.25–13.35	Lake sediment	15,920 ± 230	Beta-55125	18,800 (19,340–18,340)
14.25–14.35	Lake sediment	18,750 ± 280	Beta-55126	22,380
15.35–15.45	Lake sediment	22,280 ± 640	Beta-48605	26,340
16.40–16.50	Lake sediment	28,090 ± 470	Beta-58162	31,300
17.42–17.52	Lake sediment	36,300 ± 750	Beta-55127	38,310
17.61–17.65	Wood	>39,670	Beta-48607	>41,180
18.08–18.18	Lake sediment	27,480 ± 640	Beta-48606	30,780
		Gordon Lake		
4.99–5.09	Lake sediment	2840 ± 60	Beta-81666	2940 (3150–2780)
7.01–7.11	Lake sediment	4710 ± 60	Beta-81667	5460, 5360[a], 5330 (5580–5310)
9.75–9.85	Lake sediment	6080 ± 70	Beta-81668	6900 (7160–6780)
11.09–11.10	Mazama Ash	6845 ± 50	(Bacon, 1983)	7630 (7710–7550)
16.94–16.98	Charcoal, needles	10,010 ± 70	AA-20627	11,210, 11,350, 11,110[a] (11,840–11,000)
17.33–17.35	Charcoal, needles	10,400 ± 75	AA-20628	12,300 (12,530–11,960)
17.86–17.90	Wood, needles	11,520 ± 80	AA-20629	13,430 (13,700–13,230)
18.15–18.20	Lake sediment	12,640 ± 150	Beta-72839	14,860 (15,390–14,360)

[a] Calibrated age used in age model.

Picea pollen could not be identified to species level without the presence of macrofossils.

The late-glacial charcoal record from Little Lake was reconstructed by C. Long (unpublished data). Charcoal analysis followed the methods outlined in Millspaugh and Whitlock (1995) and is based on counts of macroscopic charcoal (pieces >125 μm in diameter) at 1-cm intervals. Charcoal counts were converted to charcoal concentration (pieces/cm^3), which was then divided by deposition time to calculate charcoal accummulation rates (pieces/cm^2/yr).

LITHOLOGY AND CHRONOLOGY

We examined the Little Lake core from 15,700 to 10,200 cal yr B.P. (12.34–9.68 m depth). The core consists of inorganic clay, silt, and fine sand (~9% organic matter) from 12.34 to 10.68 m depth and silty, fine-detritus gyttja from 10.68 to 9.68 m. Loss-on-ignition data for Little Lake during the late-glacial period shows little variation in organic content (8–11%) (Worona and Whitlock, 1995). The Gordon Lake core was analyzed between 15,500 and 8500 cal yr B.P. (18.48–14.64 m depth). The base of the Gordon Lake core (18.48–18.20 m depth) is inorganic clay, fine sand, and silt. The core contains faintly laminated gyttja from 18.48–18.62 m depth. The lithologic boundary at 18.48 m probably marks the transition from an unproductive lake fed by glacial meltwater to a more productive system and greater vegetation cover.

Age-vs-depth relations were based on bulk-sediment radiocarbon dates, AMS radiocarbon dates on plant macrofossils and charcoal, and Bacon's (1983) age for the Mazama Ash (Table 1). Radiocarbon ages <19,262 ^{14}C yr B.P. were converted to calibrated years (cal yr B.P.) using the CALIB 3.0 program (Stuiver and Reimer, 1993). Dates younger than 10,100 cal yr B.P. were calibrated with the bidecadal tree-ring data set and dates older than 10,100 cal yr B.P. with the U/Th and ^{14}C coral data set (Stuiver and Becker, 1993; Stuiver and Pearson, 1993; Pearson and Stuiver, 1993; Kromer and Becker, 1993; Bard *et al.*, 1993). Ages >19,262 ^{14}C yr B.P. were calibrated using methods described by Bard *et al.* (1990) and Mazaud *et al.* (1991).

The chronology at Little Lake is based on 13 bulk radiocarbon dates, the age of Mazama Ash, and four AMS dates (Table 1). The bulk-sediment dates were consistently 500 to 1000 cal yr older than the AMS dates. Similar age differences noted in other studies have been attributed to ancient carbon from reworked organic material or submerged aquatics in carbonate systems (Törnqvist *et al.*, 1992). The bedrock surrounding Little Lake consists of noncarbonate volcanic sediments. However, a steep catchment and relatively inorganic sediments (8–11% organics) suggest considerable allochthonous input during the late-glacial period; erosion of ancient soils may have been a source of old carbon.

A weighted, third-order polynomial equation was used to construct an age-vs-depth curve for the Little Lake core (Fig. 2). The gradual changes in deposition rate generated with this regression are consistent with those displayed by loss-on-ignition data and lithology (Worona and Whitlock, 1995). Four models using different combinations of weightings for each date were considered in selecting the final age-vs-depth model. The model chosen gave heavier weighting to AMS dates and agreed well with regional pollen stratigraphic correlations (Grigg, 1996). The four models differed in late-glacial age assignments by <150 yr.

Four bulk radiocarbon dates, the age of Mazama Ash, and three AMS dates were used to construct a chronology for Gordon Lake (Table 1). A weighted, fourth-order polynomial was used to fit a sedimentation curve for Gordon Lake (Fig. 2). Two age models, with chronologies that differed by <150 yr were considered; we use the model that gives heavier weighting to the AMS dates.

LATE-GLACIAL POLLEN RECORDS

Little Lake

The upper part of Zone LL-2 through the base of Zone LL-4 of Worona and Whitlock (1995) spans the late-glacial period (15,700–10,200 cal yr B.P.). Based on our more de-

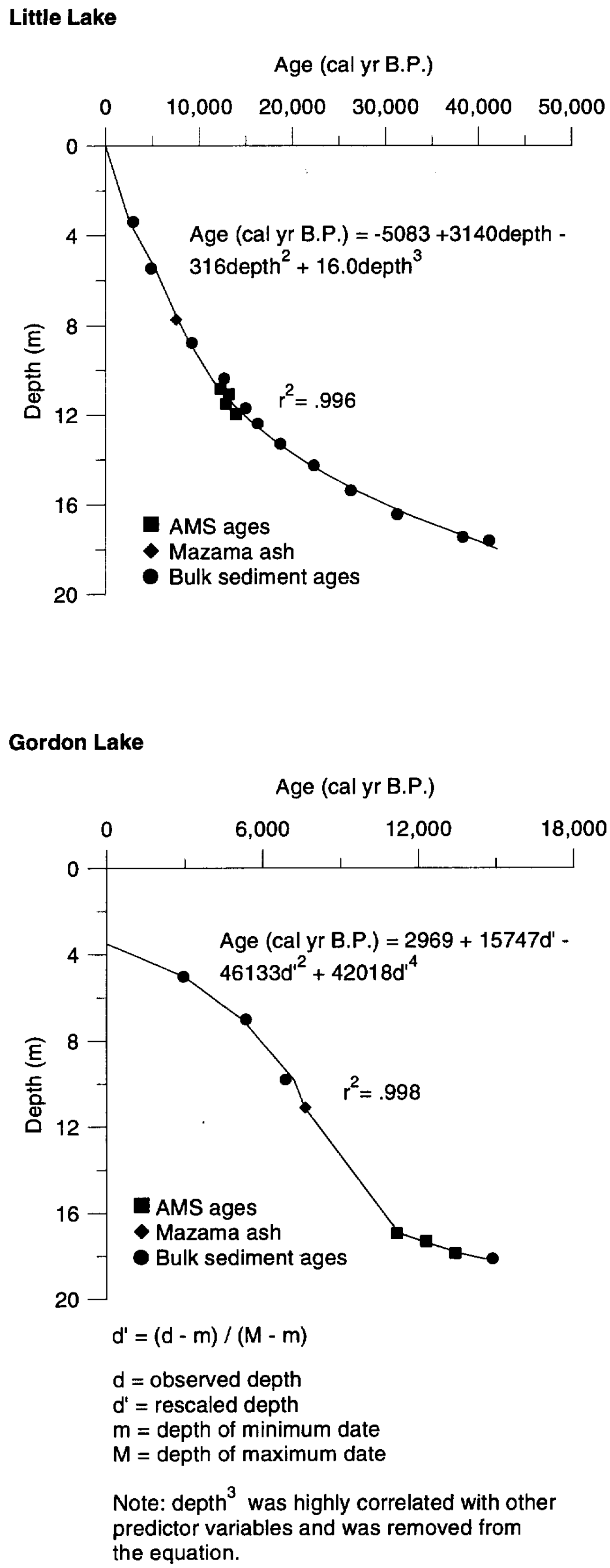

FIG. 2. Age-vs-depth curves and regression equations for Little Lake and Gordon Lake. Age calculations were derived using calibrated ages (see Table 1). Standard deviations for radiocarbon and calibrated ages are listed in Table 1.

TABLE 2
Description of Little Lake Pollen Percentage Diagram

Pollen zone	Depth (m)	Age: Cal yr B.P.	Age: ^{14}C yr B.P.[a]	Description
LL-4a	9.68–10.20	10,200–11,000	9230–9830	High percentages of *Alnus rubra*-type (40%), *Pseudotsuga*-type (20%), and *Pteridium*-type (10%); *Corylus* (15%) and *Quercus* (<5%) abundant at the top. Total accumulation rates of terrestrial pollen and spores (PAR): 7600–15,200 g/cm^2/yr.
LL-3d	10.20–10.88	11,000–12,400	9830–10,480	High percentages of *Pinus* (30%), mostly haploxylon-type (10%); *A. rubra*-type (30–40%) and *Pseudotsuga*-type (10%) less abundant than in Zone LL-3b; *Abies* and *Tsuga heterophylla* (<5%). PAR: 7800–11,800 g/cm^2/yr.
LL-3c	10.88–11.78	12,400–14,250	10,480–12,200	High percentages of *Pseudotsuga*-type (5–15%), *A. rubra*-type (40%), *Dryopteris*-type (10–20%), and *Pteridium*-type (5%). PAR: 4400–10,700 g/cm^2/yr.
LL-3b	11.78–11.85	14,250–14,500	12,200–12,390	High percentages of *Picea* (>20%), *Pinus*, mostly diploxylon-type (10–25%), *Abies*, *T. heterophyilla*, *Alnus sinuata*-type (>10%), *Tsuga mertensiana* (>5%), and minor amounts (3–5%) of Poaceae, Cyperaceae, and *Artemisia*. PAR: 4600–6800 g/cm^2/yr.
LL-3a	11.85–12.10	14,500–14,850	12,390–12,630	High percentages of *A. rubra*-type (40%), *Pseudotsuga*-type (>5%), *Drypoteris*-type (15–20%), and *Pteridium*-type (>5%). Trace amounts of Cupressaceae and *Spiraea*-type (<5% each). PAR: 4300–5900 g/cm^2/yr.
LL-2a	12.10–12.34	14,850–15,700	12,630–13,170	High percentages of *Picea* (15–25%), and *T. mertensiana* (10–15%) at the base. Abundant *Abies* (5–20%), *Pinus*, mostly diploxylon-type (10–25%), *A. sinuata*-type (5–15%), Poaceae (>5%), Cyperaceae (>5%), and *Dryopteris*-type (5–10%). Trace amounts of *Artemisia*-type and Other Asteraceae (<5% each). PAR: 3800–6200 g/cm^2/yr.

[a] ^{14}C ages derived from CALIB 3.0 calibration curves (Stuiver and Reimer, 1993). Errors associated with the transformation of calendar ages to the equivalent ^{14}C ages were not included.

tailed examination, we divided these zones into subzones, in order to reconstruct the late-glacial vegetation history (Table 2; Fig. 3). Pollen accumulation rates support the percentage trends throughout the late-glacial period. The base of Subzone LL-2 (15,700-15,250 cal yr B.P.) indicates a period of open subalpine forests similar to that of the *T. mertensiana* Zone at present (Heusser, 1978). Needles of *Picea engelmannii, P. contorta,* and *T. mertensiana* suggest that these trees grew near the lake (Worona and Whitlock, 1995). The *Abies* pollen is most likely from *A. lasiocarpa*. Pollen spectra from the upper part of Subzone LL-2 (15,250–14,850 cal yr B.P.; 13,170–12,630 ^{14}C yr B.P.) resemble those from present-day closed montane forests in the *A. amabilis* Zone (Heusser, 1978; Sea and Whitlock, 1995). The forest was probably composed of *T. heterophylla, A. amabilis, A. procera,* and lesser amounts of *P. engelmannii, P. contorta,* and *T. mertensiana. Alnus sinuata* likely grew in disturbed and wet areas, and *A. rubra* and *Salix* were present in riparian zones.

Subzone LL-3a (14,850–14,500 cal yr B.P.; 12,630–12,390 ^{14}C yr B.P.) resembles modern pollen percentages from *Pseudotsuga*-dominated forests in the *T. heterophylla* Zone (Heusser, 1978; Worona and Whitlock, 1995). *Polystichum minutum* (the likely contributor of *Dryopteris*-type spores) was probably an important component of the closed forest understory and *Alnus rubra, Sambucus,* and *Pteridium* likely grew in disturbed areas. Subzone LL-3b (14,500–14,250 cal yr B.P.; 12,390–12,200 ^{14}C yr B.P.) represents the return to a subalpine forest of *Picea, T. mertensiana, Abies,* and *T. heterophylla.* The transition from Subzone LL-3b to Subzone LL-3c (14,250–12,400 cal yr B.P.; 12,200–10,480 ^{14}C yr B.P.) includes a sequence of pollen assemblages that shift from *Picea, Pinus* (mostly diploxylon-type), and *T. mertensiana* to *T. heterophylla* and *Abies* and finally to *Pseudotsuga*-type and *A. rubra*-type. This progression suggests a second transition from subalpine forest to a *Pseudotsuga*-dominated forest. *Abies* pollen at this time could have been from *A. amabilis* and *A. procera* and/or from low-elevation *A. grandis.*

Subzone LL-3d (12,400–11,000 cal yr B.P.; 10,480–9830 ^{14}C yr B.P.) lacks a clear modern analog. Haploxylon-type *Pinus* pollen is assumed to be *P. monticola* on the basis of its association with *Pseudotsuga*-type pollen. *P. monticola* is a minor component of both the present-day *A. amabilis* Zone and the *T. mertensiana* Zone, as a result of the introduction of white pine blister rust (Fowells, 1965). A better modern analogue may be the montane forests of northeastern Washington and northern Idaho, where *P. monticola* is relatively abundant. Modern pollen data from these regions, however, lack the high

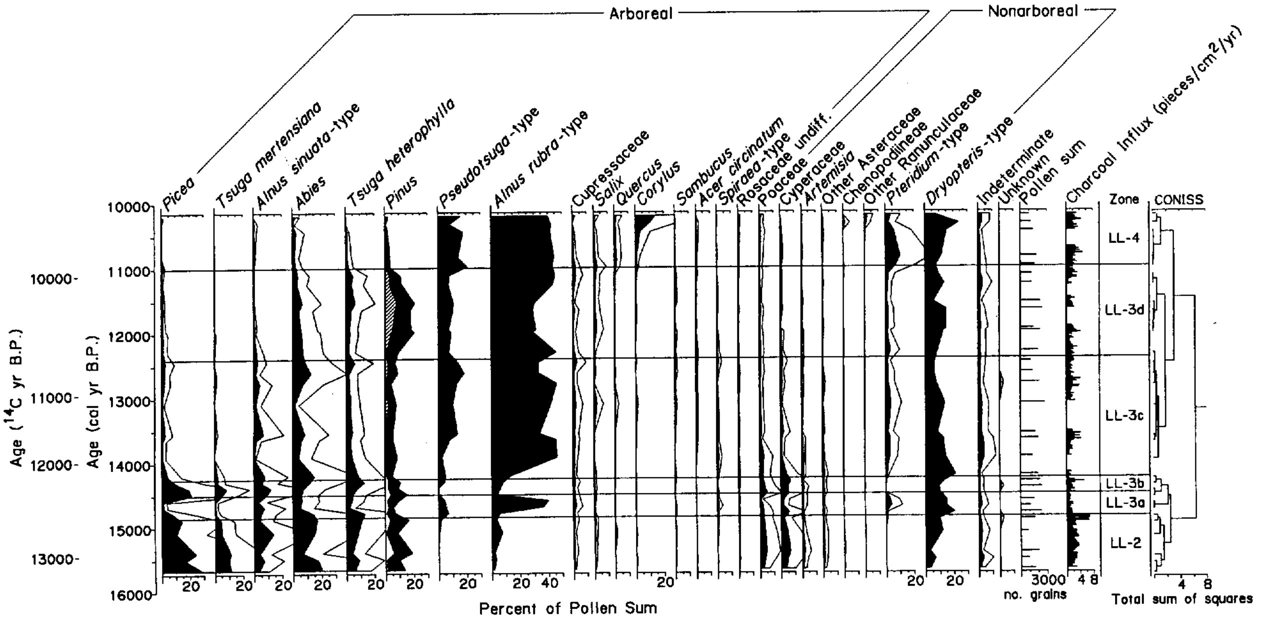

FIG. 3. Pollen percentages for selected taxa and charcoal accumulation rates during the late-glacial period at Little Lake. Open curves represent fivefold exaggeration of black curves. *Pinus* includes diploxylon-type (*P. contorta* or *P. ponderosa*) represented by the open curve, haploxylon-type (*P. monticola* or *P. albicaulis*) shown with a hatched curve, and *Pinus* undifferentiated shown with a black curve. Pollen zones were calculated using a constrained cluster analysis (CONISS; Grimm, 1988). Radiocarbon age scale was derived using the CALIB 3.0 calibration curve (Stuiver and Reimer, 1993).

percentages of *Alnus* seen in Subzone LL-3d (Heusser, 1978; Mack *et al.,* 1978). The subzone apparently represents a mixed conifer forest composed of elements from mid- to low-elevation forests west and east of the Cascade Range. *P. monticola* probably occupied cool, moist slopes, while *Pseudotsuga* was abundant on dry, warm slopes.

Subzone LL-4 (11,000–10,200 cal yr B.P.; 9830–9230 ^{14}C yr B.P.) is similar to modern pollen percentages from the Willamette Valley and low-elevation forests of the Cascade Range (Heusser, 1978). *Pseudotsuga* was the dominant conifer, and *Quercus garryana* grew in dry areas subject to frequent fire. *Corylus cornuta, Acer circinatum,* and *Polysticum* occupied the understory, *A. rubra* and *Salix* were present in riparian areas, and *Pteridium* was seen in disturbed areas.

Gordon Lake

The late-glacial record at Gordon Lake is divided into three pollen zones (Table 3; Fig. 4). Changes in pollen accumulation rates are similar to those described for the percentage data. Zone GOR-1 (15,500–14,500 cal yr B.P.; 13,040–12,390 ^{14}C yr B.P.) is similar to modern pollen spectra from subalpine parkland in the *T. mertensiana* Zone (Heusser, 1978), with high percentages of *P. engelmannii, P. contorta,* Poaceae, Cyperaceae, and *Artemisia. T. mertensiana, T. heterophylla,* and *Abies* were probably present in smaller amounts. High *Picea* percentages at the base of the zone suggests that it was an important early colonizer of deglaciated areas in the Cascade Range.

The presence of *Abies, Pinus, Alnus sinuata*-type, *T. mertensiana,* and *Dryopteris*-type in Subzone GOR-2a (14,500–12,800 cal yr B.P.; 12,390–10,870 ^{14}C yr B.P.) indicates closed forest similar to the upper *A. amabilis* and lower *T. mertensiana* zones at present (Heusser, 1978; Sea and Whitlock, 1995). *Alnus rubra, Salix,* and *Spiraea* were probably present in mesic and riparian areas. High percentages of *A. sinuata*-type pollen imply areas of heavy snow accumulation (Franklin and Dyrness, 1988). Subzone GOR-2b (12,800–11,000 cal yr B.P.; 10,870–9830 ^{14}C yr B.P.) resembles modern spectra from the *T. mertensiana* Zone, except for the abundance of *Pinus* and *Abies* pollen, which is similar to assemblages from high-elevation forests in the eastern Cascade Range (Heusser, 1978). The forest was dominated by *Pinus* (either *P. monticola* or *P. albicaulis* and *P. contorta*), *Abies* (*A. lasiocarpa, A. amabilis,* and *A. procera*), and *T. mertensiana.* The local presence of *Abies lasiocarpa* and *T. mertensiana* is confirmed by the occurrence of their needles within Subzone GOR-2b.

High values of *Pseudotsuga*-type in Zone GOR-3 (11,000–8500 cal yr B.P.; 9830–7760 ^{14}C yr B.P.) mark the beginning of the Holocene. *Pseudotsuga*-type, *Pinus,* and *Abies* percentages resemble those from modern forests at

TABLE 3
Description of Gordon Lake Pollen Percentage Diagram

Pollen zone	Depth (m)	Age: Cal yr B.P.	Age: ^{14}C yr B.P.[a]	Description
GOR-3	14.64–16.90	8500–11,000	7760–9830	High percentages of *Pseudotsuga*-type (15–20%), *Alnus sinuata* (30–40%), *Pinus* (20%), *Dryopteris*-type (10–15%). Increases in *Pteridium*-type (5%), *Quercus* (5%), and *Corylus* (2%). Total accumulation rate of terrestrial pollen and spores (PAR): 6200–32,400 g/cm^2/yr.
GOR-2b	16.90–17.60	11,000–12,800	9830–10,870	High percentages of *Pinus* (30%), both haploxylon and diploxylon-types, *Abies* (10–20%), *A. sinuata* (20–30%), and *Tsuga mertensiana* (5–10%). Lesser amounts of *Tsuga heterophylla* (2–5%), *Pseudotsuga*-type (2–5%), *Pteridium*-type (2%), and *Dryopteris*-type (5%). PAR: 2900–7800 g/cm^2/yr.
GOR-2a	17.60–18.25	12,800–14,500	10,870–12,390	High percentages of *Abies* (10–20%), *Pinus* (15–20%), *A. sinuata* (20–30%), *T. mertensiana* (5–10%), and *Dryopteris*-type (10–20%). Lesser amounts of *T. heterophylla* (2–5%), *Pseudotsuga*-type (2–5%), *Picea* (2%), *Pteridium*-type (2–5%). PAR: 2800–5300 g/cm^2/yr.
GOR-1	18.25–18.48	14,500–15,500	12,390–13,040	High percentages of *Picea* (20–40%), *Pinus* (20–35%), mostly diploxylon-type, Poaceae (10%), Cyperaceae (5–10%), and *Artemisia* (10–15%). Lesser amounts of *T. mertensiana* (5–10%), *Abies* (5–10%), *T. heterophylla* (2–5%), *A. sinuata* (5%), and *Dryopteris*-type (2–10%). PAR: 1400–2500 g/cm^2/yr.

[a] ^{14}C ages derived from CALIB 3.0 calibration curves (Stuiver and Reimer, 1993). Errors associated with the transformation of calendar ages to the equivalent ^{14}C ages were not included.

mid- and low-elevations in the western Cascade Range (Heusser, 1978; Sea and Whitlock, 1995). *Abies, P. monticola,* and *P. contorta* may have occupied north-facing slopes while *Pseudotsuga* grew on south-facing slopes. *Quercus* and *Corylus* pollen probably came from the Willamette Valley.

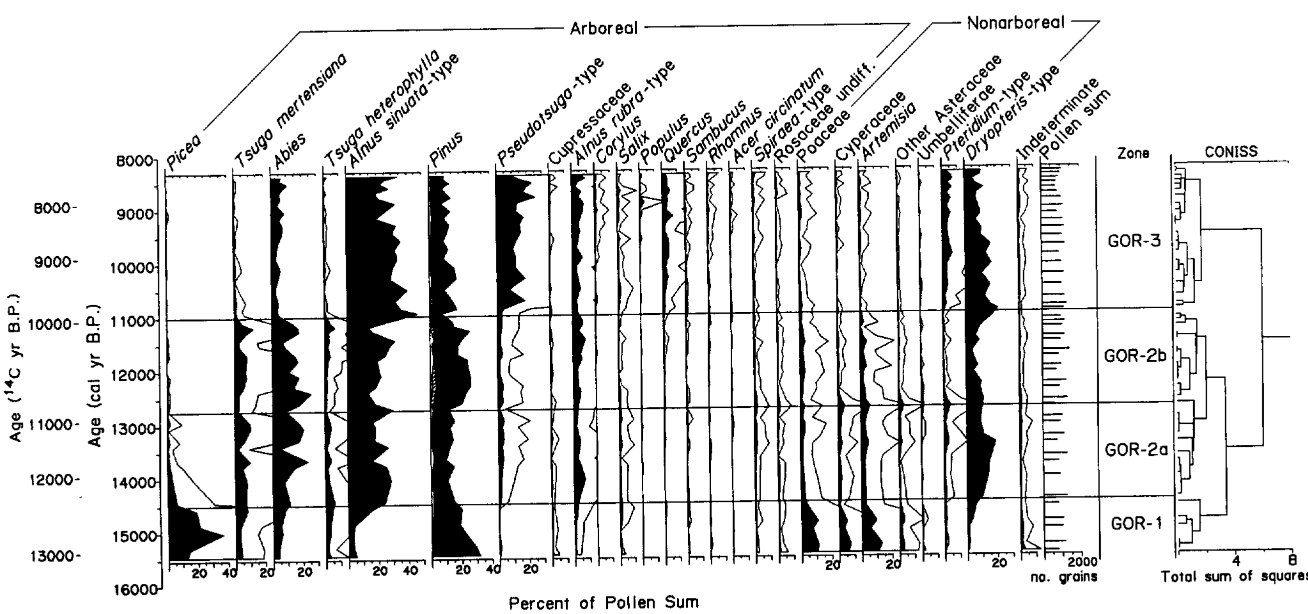

FIG. 4. Pollen percentages for selected taxa during the late-glacial period at Gordon Lake. Open curves represent fivefold exaggeration of black curves. *Pinus* includes diploxylon-type (*P. contorta* or *P. ponderosa*) represented by the open curve, haploxylon-type (*P. monticola* or *P. albicaulis*) shown with a hatched curve, and *Pinus* undifferentiated shown with a black curve. Pollen zones were calculated using a constrained cluster analysis (CONISS; Grimm, 1988). Radiocarbon age scale was derived using the CALIB 3.0 calibration curve (Stuiver and Reimer, 1993).

DISCUSSION

15,700–12,400 cal yr B.P. (LL-2a–LL-3c): Full-Glacial to Late-Glacial Transition

The shift from open subalpine to closed montane forest at Little Lake between 15,700 and 14,850 cal yr B.P. implies a climatic warming equivalent to a 500- to 1000-m decrease in elevation. Present conditions of subalpine and montane forests in the Pacific Northwest suggest that annual temperatures increased by 2–3°C, but were still ca. 4°C below present. Continued warming after 14,850 cal yr B.P. (LL-3a) is evidenced by the sudden appearance (within a century) of a forest with *Pseudotsuga, A. rubra,* and *Pteridium.* These taxa imply an increase in annual temperature of 3–5°C and a decrease in annual precipitation of 100–300 mm, compared to previous conditions. A prominent charcoal peak in sediments dating from 14,900 to 14,850 cal yr B.P. precedes the vegetation change and suggests a brief episode of increased fire activity (Fig. 3) (C. Long, unpublished data). *Pseudotsuga, A. rubra,* and *Pteridium* are all disturbance-adapted species; thus a shift toward warmer drier conditions, coupled with one or a series of fires, would have enabled the rapid vegetational change (Minore, 1979; Agee, 1993). This vegetation and fire regime was apparently short-lived, lasting <350 yr.

At 14,500 cal yr B.P. (LL-3b) the recurrence of a subalpine forest implies a return to cool humid conditions. The sudden appearance of subalpine taxa suggests that they were growing in the central Coast Range and could spread into the Little Lake region with little delay. Warming is again registered between 14,400 and 14,250 cal yr B.P. during the transition to Subzone LL-3c by the reappearance of montane and temperate taxa. A second charcoal peak, represented by one sample, precedes high percentages of *Pseudotsuga* and further implicates fire as a catalyst of vegetational change. The dominance of *Pseudotsuga* and the presence of *Abies, P. monticola,* and *T. heterophylla* from 14,250 to 12,400 cal yr B.P. indicate that conditions were warmer and wetter than before, although cooler than today (Fig. 5).

The vegetation at Gordon Lake between 15,700 and 14,500 cal yr B.P. (GOR-1) implies that conditions were ca. 3°C colder than present. Gordon Lake does not register a warm period comparable to the *Pseudotsuga* interval at Little Lake (LL-3a) (Fig. 5). The record does suggests that subalpine parkland was replaced at 14,500 cal yr B.P. by a closed montane forest (GOR-2a) as a result of a 1–3°C warming and a slight increase in precipitation.

Indian Prairie Fen, just north of Gordon Lake, also shows a subalpine parkland between 16,400 and 14,500 cal yr B.P. (Sea and Whitlock, 1995) (Figs. 1 and 6). A minor peak (<2%) of *Pseudotsuga* pollen at Indian Prairie Fen between 15,750 and 14,750 cal yr B.P. may correspond with LL-3a.

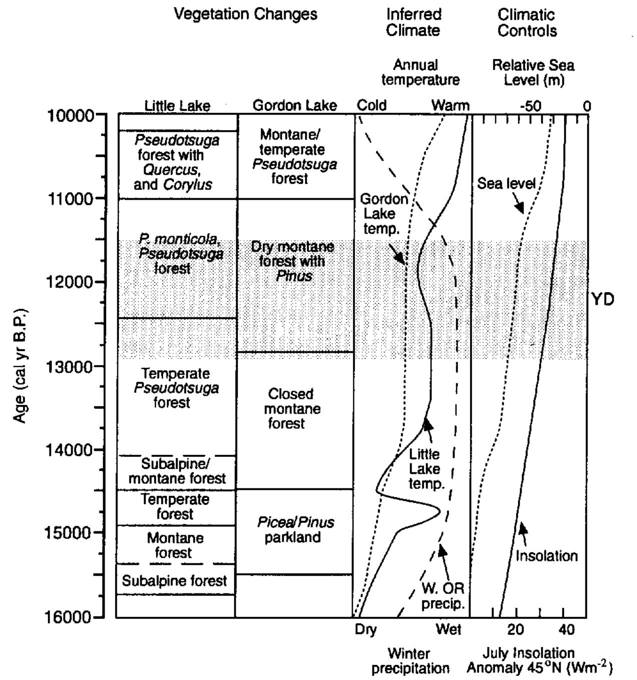

FIG. 5. Submillennial changes in vegetation and the inferred climate during the late-glacial period at Little Lake and Gordon Lake. Variations in millennial-scale climatic controls in the Pacific Northwest during the late-glacial period are shown in the far right panel. July insolation anomaly (difference from present) from Berger (1978). Relative sea level curve is an estimate of changes in ice sheet volume (Fairbanks, 1989). Shaded area indicates the Younger Dryas period in the North Atlantic.

A closed montane forest developed at Indian Prairie Fen after 14,500 cal yr B.P. In the southern Puget Lowland, Battle Ground Lake records an open *Picea–Pinus* parkland and cool humid conditions from 18,000 to 13,000 cal yr B.P., with little modification (Barnosky, 1985) (Figs. 1 and 6). The establishment of closed forest occurred there after 13,000 cal yr B.P.

Several points may explain why a brief warming at 14,850 to 14,500 cal yr B.P. is thus far recorded only at Little Lake. Evidence for the warming may have been overlooked because of the coarse sampling interval of other records, though not at Gordon Lake. In addition, the vegetation at Little Lake is probably more sensitive to submillennial variations in climate and fire frequency, because of its location, vegetation, or climate at that time. For example, the abundance of fuels and relatively warm climate at 15,300 cal yr B.P. may have made Little Lake more susceptible to fire than sites near glacial margins. Fire, in turn, likely triggered a response to short-term climate change at Little Lake.

12,400–11,000 cal yr B.P. (LL-3d): Pinus Period

At 12,400 cal yr B.P. an increase in *P. monticola* and decrease in *Pseudotsuga* and *A. rubra* at Little Lake suggest

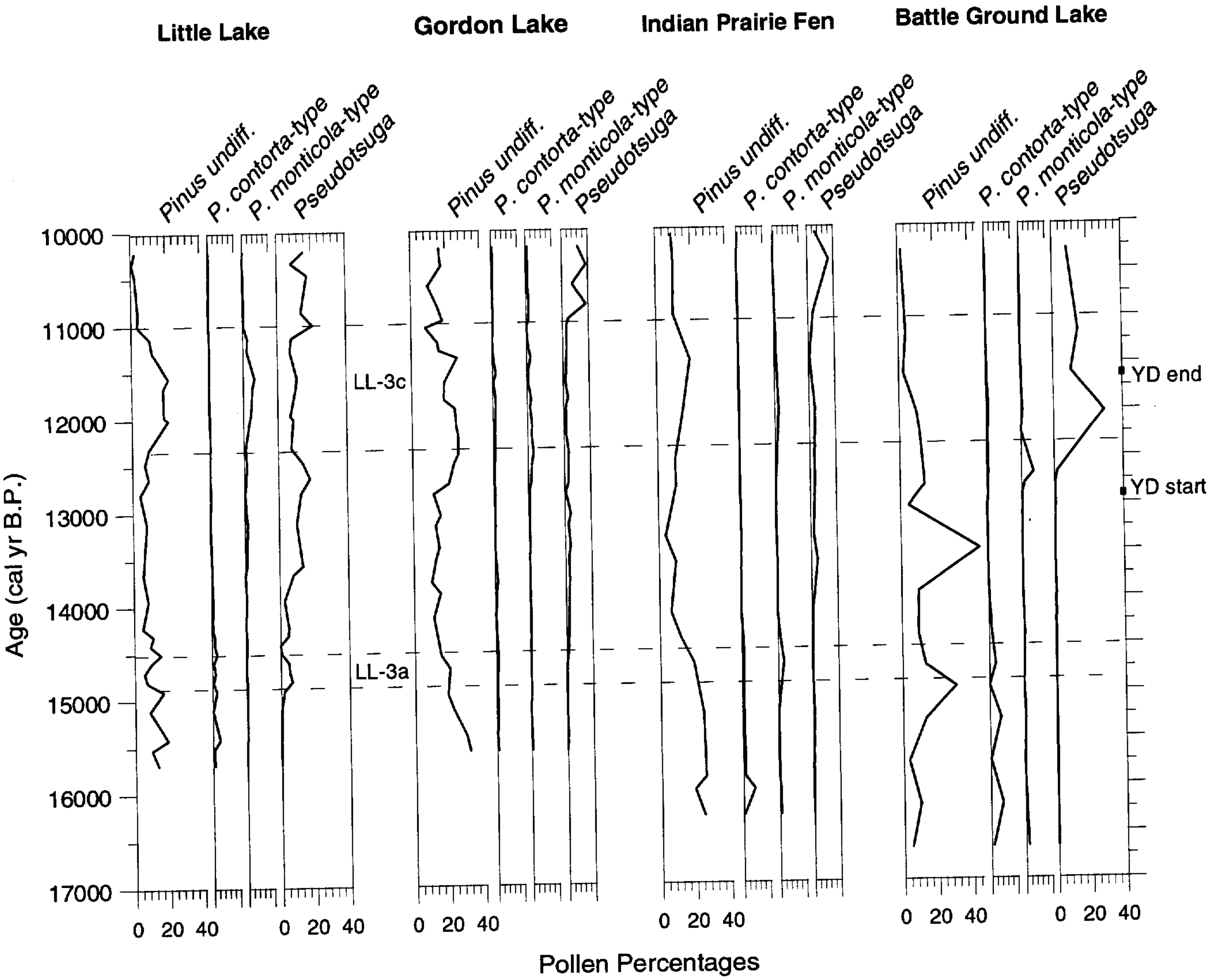

FIG. 6. Late-glacial pollen percentages of *Pinus* and *Pseudotsuga*-type from sites in western Oregon and southern Washington. Dashed lines show the position of Zone LL-3a and Zone LL-3c. The start and end of the Younger Dryas period in the North Atlantic are also shown.

cooler winters and drier summers than before. Areas that support abundant *P. monticola* today are characterized by warm dry summers and cold winters with heavy snow accumulation (Fowells, 1965; Franklin and Dyrness, 1988). *P. monticola* is limited by drought at low elevations and cold temperatures at high elevations. *Pseudotsuga* is abundant at low elevations as a result of its relative intolerance to frost and snow and its ability to resist drought (Minore, 1979). Thus, expansion of *P. monticola* and decline of *Pseudotsuga* were likely a result of cooler winters and more snowfall. The decline in *A. rubra,* which favors mild climates, also supports this interpretation. Absence of montane and subalpine species, such as *Picea, T. mertensiana,* and *A. sinuata,* however, indicates that cooling was not great enough to result in a large shift in vegetation.

An increase in *Pinus* at Gordon Lake between 12,800 and 11,000 cal yr B.P. (GOR-2b) marks the transition from mesophytic to xerophytic montane forest and coincides with the expansion of *P. monticola* at Little Lake. This transition implies greater seasonality than before with warmer drier summers. Additional evidence of a climatic change comes from the slight decline in *Tsuga heterophylla,* which is severely limited by summer drought (Minore, 1979). *T. mertensiana* and *Abies* percentages do not change in this interval, suggesting the persistence of cool winter conditions with abundant precipitation (Fig. 5).

A regional climatic change is implied by expansion of *Pinus* in the Pacific Northwest between 13,000 and 11,000 cal yr B.P. At Indian Prairie Fen, an increase in *Pinus* (haploxylon-type) between 12,500 and 11,000 cal yr B.P.

and increases in *T. mertensiana* and *Abies* between 11,800 and 11,000 cal yr B.P., overlap with the *Pinus* period at Little Lake and Gordon Lake (ca. 12,600–11,000 cal yr B.P.) (Fig. 6). Battle Ground Lake has several peaks of *Pinus* during the late-glacial period: an increase in diploxylon-type (referred to *P. contorta*) from 15,750 to 14,750 cal yr B.P., followed by an increase in *Pinus* undifferentiated from 14,800 to 12,750 cal yr B.P. and finally by an increase in haploxylon-type from 12,750 to 11,500 cal yr B.P. (Fig. 6). The initial expansion of *P. contorta* has been ascribed to favorable local edaphic conditions (Barnosky, 1985), but the later haploxylon-type *Pinus* expansion (12,750–11,500 cal yr B.P.) coincides with that in western Oregon, so suggests a regional event. Davis and Mineral lakes in the southern Puget Trough also show a peak in *Pinus* (mostly diploxylon-type) that generally coincides with the *Pinus* interval in western Oregon (Barnosky, 1981; Tsukada *et al.,* 1981).

One explanation for the *Pinus* period may be that summers became progressively warmer and drier in the late-glacial period (although conditions were not as xerothermic as in the early Holocene), while winters remained cooler than at present. Increased seasonality in the Pacific Northwest between ca. 13,000 and 11,000 cal yr B.P. is explained by the combined effects of changes in the amplitude of the seasonal cycle of insolation and the position and strength of the winter jet stream. Variations in the seasonal cycle of insolation and in the size of the Laurentide ice sheet have been shown to be important controls of late Quaternary climate variations on millennial time scales in the Pacific Northwest (Whitlock and Bartlein, 1997). Summer insolation was greatest between 11,000 and 10,000 cal yr B.P. (Thompson *et al.,* 1993), which resulted in higher summer temperatures and less precipitation. The corresponding decrease in winter insolation would have been effective at keeping winter temperatures cooler than at present. Paleoclimate model simulations for 14,000 cal yr B.P. indicate that the jet stream had moved to its present position and that the Laurentide ice sheet still produced a steep temperature gradient at its southern margin, strengthening the jet stream. The combination of these effects has been attributed to the increase in winter storms in the Pacific Northwest during late-glacial time (Thompson *et al.,* 1993). Ice volumes were likely large enough (Fairbanks, 1989; Peltier, 1994) to influence the North American jet throughout the late-glacial period (Fig. 6).

An alternative explanation for the *Pinus* period may be submillennial-scale climate changes associated with the Younger Dryas event. If so, the evidence is most convincing at Little Lake, where a cooling is registered between 12,400 and 11,000 cal yr B.P. The interval, however, lags Younger Dryas cooling in the North Atlantic (12,900–11,600 cal yr B.P.) by several centuries (Fig. 6). The already cool conditions at Gordon Lake, Indian Prairie Fen, and Battle Ground Lake may have obscured evidence of the beginning of a Younger Dryas event; however, as in the Little Lake record, cool conditions at these sites ended abruptly and lag Younger Dryas cooling in the North Atlantic.

11,000–10,200 cal yr B.P. (LL-4a): Early Holocene Transition

After 11,000 cal yr B.P., the vegetation at Little Lake was comparable to modern closed forests growing at low elevations in the western Cascade Range. High percentages of *Pseudotsuga, Quercus,* and *Corylus* indicate decreased precipitation and warmer drier summers than before. After 11,000 cal yr B.P. *Pseudotsuga* forest was present at Gordon Lake and Indian Prairie Fen, and *Pseudotsuga* and *Quercus* grew near Battle Ground Lake (Fig. 6). The climate conditions implied by these assemblages are attributed to the effects of increased summer insolation, including a stronger-than-present eastern Pacific Subtropical High which is presently associated with drought in the Pacific Northwest (Fig. 5) (Thompson *et al.,* 1993).

Climatic Change in the Northeast Pacific Region during Younger Dryas Time

Cooling during Younger Dryas time has been reported in some, but not all, marine and terrestrial records in the northeast Pacific region. Increases in *T. mertensiana* or herb taxa at 11 coastal sites in British Columbia between ca. 12,600 and 11,000 cal yr B.P. are interpreted as a shift toward cooler and wetter conditions (Mathewes, 1993). An increase in herbaceous taxa between 12,700 and 10,900 cal yr B.P. from southeastern Alaska also suggests Younger Dryas cooling (Engstrom *et al.,* 1990). The coastal location of all these sites suggests a transmittal mechanism that involves changes in oceanic circulation (Mathewes, 1993; Hansen and Engstrom, 1996). An oceanic mechanism is also implied by evidence of increased bioturbation and cool-water planktonic foraminifers between 12,970 and 11,200 cal yr B.P. from the Santa Barbara Basin (Kennett and Ingram, 1995) and by similar changes of Younger Dryas age in the Guaymas Basin in the Gulf of California (Keigwin and Jones, 1990).

Most records from the northeast Pacific, including those in western Oregon, show that conditions throughout the late-glacial period were not as cool as during the full-glacial period. Another feature in common is that the late-glacial cooling episode in the northeast Pacific lasted until 11,000 cal yr B.P., 600 yr after the end of the Younger Dryas event in the North Atlantic. This discrepancy may represent lags in the transmission of the Younger Dryas event to the northeast Pacific region or, in the case of pollen records, lags in the vegetational response to regional climate changes. The lack of correspondence could also reflect age control problems caused by the plateau in the radiocarbon calibration curve between 10,900 and 12,300 cal yr B.P.; a broad range of calendar ages can correspond to a single ^{14}C age (Stuiver and Reimer, 1993; Bartlein *et al.*, 1995). As a result, calendar and radiocarbon ages from the interval that includes the end of the Younger Dryas event are poorly constrained.

The generally weak registration of a Younger Dryas event at Little and Gordon lakes in comparison to other sites in the northeast Pacific may be explained by their relatively inland location. Inland sites would have been more sensitive to the effects of increasing summer radiation, particularly to steepening temperature and precipitation gradients, and less sensitive to subtle changes in ocean conditions. The absence of evidence for glacial advances during Younger Dryas time at inland sites in the Washington Cascade Range is noteworthy and supports this hypothesis (Heine, in press; Porter, 1976, 1978).

The Younger Dryas signal is strongest in marine and coastal records in the northeast Pacific. The records implicate changes in atmospheric and/or oceanic circulation that originated over the North Pacific Ocean as the proximal cause of late-glacial submillennial variations in the region. Recent paleoclimatic model simulations suggest that atmospheric cooling of the North Pacific during Younger Dryas time led to decreased sea-surface temperatures and the expansion of sea-ice, which caused further changes to atmospheric circulation in the North Pacific (Mikolajewicz *et al.*, 1997). The model also indicates that changes in oceanic circulation contributed to cooling in the North Pacific and that the greatest decrease in sea-surface temperatures occurred north of 50°N. Variability of sea surface temperatures along the northeast Pacific coast, as a result of changes in upwelling, may explain stronger terrestrial Younger Dryas evidence from coastal British Columbia and Alaska in comparison to western Oregon and Washington.

CONCLUSIONS

(1) The establishment of *Pseudotsuga* at Little Lake between 14,850 and 14,500 cal yr B.P. indicates a brief period of warm dry conditions. The rapid response of vegetation to this submillennial-scale climatic change was triggered by an increase in fire frequency. The return of subalpine forests suggests cooling after 14,500 cal yr B.P. The absence of evidence for this climatic oscillation at other sites in the Pacific Northwest may be related to either their wide sampling interval, the insensitivity of their vegetation at that time to climate change, or the absence of fire to catalyze the change.

(2) A period of *Pinus* dominance at several sites in the Pacific Northwest suggests a regional shift in climate between ca. 13,000 and 11,000 cal yr B.P. toward greater seasonality. The *Pinus* period can be explained either by the combined effects of changes in the seasonal cycle of insolation and atmospheric circulation that occurred during the late-glacial period or by a short-term cooling event. At Little Lake, cooling during the *Pinus* period occurred between 12,400 and 11,000 cal yr B.P. and overlaps in time with the Younger Dryas event in the North Atlantic. Persistence of cool winter conditions at other sites until 11,000 cal yr B.P. may or may not be related to Younger Dryas cooling.

(3) Comparisons with other sites from the northeast Pacific imply that the proximity of a site to the ocean may have determined its sensitivity to short-term climate fluctuations. While millennial-scale controls of climate are clearly registered at all sites, the geographic pattern of Younger Dryas evidence suggests that changes in atmospheric circulation over the North Pacific and fluctuations in sea-surface temperatures may account for the better registration of submillennial climate variations along the coast.

ACKNOWLEDGMENTS

We thank P. J. Bartlein, C. J. Mock, and C. Long for discussion and comments on the manuscript. Research was supported by National Science Foundation Grant ATM 9615822 to C. Whitlock.

REFERENCES

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

Alley, R. B., Meese, D. A., Shuman, C. A., Gow, A. J., Taylor, K. C., Grootes, P. M., White, J. W. C., Ram, M., Waddington, E. D., Mayewski, P. A., and Zielinski, G. A. (1993). Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. *Nature* **362,** 527–529.

Bacon, C. R. (1983). Eruptive history of Mount Mazama and Crater Lake caldera, Cascade Range, U.S.A. *Journal of Volcanology and Geothermal Research* **18,** 57–115.

Bard, E., Hamelin, B., Fairbanks, R. G., and Zindler, A. (1990). Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals. *Nature* **345,** 405–410.

Bard, E., Arnold, M., Fairbanks, R. G., and Hamelin, B. (1993). ^{230}Th-^{234}U and ^{14}C ages obtained by mass spectrometry on corals. *Radiocarbon* **35,** 191–199.

Barnosky, C. W. (1981). A record of late Quaternary vegetation from Davis Lake, southern Puget Lowland, Washington. *Quaternary Research* **16,** 221–239.

Barnosky, C. W. (1985). Late Quaternary vegetation near Battle Ground Lake, southern Puget Trough, Washington. *Geological Society of America Bulletin* **96,** 263–271.

Barnosky, C. W., Anderson, P. M., Bartlein, P. J. (1987). The northwestern U.S. during deglaciation: Vegetational history and paleoclimatic implications. *In* "North America and Adjacent Oceans during the Last Glaciation" (W. F. Ruddiman and H. E. Wright, Jr., Eds.), The Geology of North America, Vol. K-3, pp. 289-232. Geological Society of America, Boulder, CO.

Bartlein, P. J., Edwards, M. E., Shafer, S. L., and Barker, E. D., Jr. (1995). Calibration of radiocarbon ages and the interpretation of paleoenvironmental records. *Quaternary Research* **44,** 417–424.

Berger, A. (1978). Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quaternary Research* **9,** 139–167.

Broecker, W. S. (1994). Massive iceberg discharges as triggers for global change. *Nature* **372,** 421–424.

Cwynar, L. C., Burden, E., and McAndrews, J. H. (1979). An inexpensive method for concentrating pollen and spores from fine grained sediments. *Canadian Journal of Earth Sciences* **16,** 1115–1120.

Davis, M. B., Brubaker, L. B., and Webb, T., III. (1973). Calibration of absolute influx. *In* "Quaternary Plant Ecology" (H. J. B. Birks and R. G. West, Eds.), pp. 109–116. Wiley, New York.

Denton, G. H., and Hendy, C. H. (1994). Younger Dryas age advance of Franz Josef Glacier in the Southern Alps of New Zealand. *Science* **264,** 1434–1437.

Engstrom, D. R., Hansen, B. C. S., and Wright, H. E., Jr. (1990). A possible Younger Dryas record in southwestern Alaska. *Science* **250,** 1383–1385.

Faegri, K., Kaland, P. E., and Krzywinski, K. (1989). "Textbook of Pollen Analysis." Wiley, London.

Fairbanks, R. G. (1989). A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* **342,** 637–642.

Fowells, H. A. (1965). "Silvics of Forest Trees of the United States." U.S. Department of Agriculture Handbook 271.

Franklin, J. F., and Dryness, C. T. (1988). "Natural Vegetation of Oregon and Washington." Oregon State Univ. Press, Corvallis.

Grigg, L. D. (1996). "Late-Glacial Vegetation and Climate Change in Western Oregon." Unpublished MA thesis, Univ. of Oregon.

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

Hansen, B. C. S., and Engstrom, D. R. (1996). Vegetation history of Pleasant Island, southeastern Alaska since 13,000 yr B.P. *Quaternary Research* **46,** 161–175.

Heine, J. T. Extent, timing, and climatic implications of glacier advances near Mount Rainier, Washington, USA, at the Pleistocene/Holocene transition. *Quaternary Science Reviews* **17,** in press.

Heusser, C. J. (1977). Quaternary palynology of the Pacific Slope of Washington. *Quaternary Research* **8,** 282–306.

Heusser, C. J. (1978). Modern pollen spectra from western Oregon. *Bulletin of the Torrey Botanical Club* **105,** 14–17.

Keigwin, L. D., and Jones, G. A. (1990). Deglacial climatic oscillations in the Gulf of California. *Paleoceanography* **5,** 1009–1023.

Kennett, J. P., and Ingram, B. L. (1995). A 20,000-year record of ocean circulation and climate change from the Santa Barbara basin. *Nature* **377,** 305–308.

Kromer, B., and Becker, B. (1993). German oak and pine ^{14}C calibration, 7200-9400 BC. *Radiocarbon* **35,** 125–135.

Lowell, T. V., Heusser, C. J., Anderson, B. G., Moreno, P. I., Hauser, A., Heusser, L. E., Schlüchter, C., Marchant, D. R., and Denton, G. H. (1995). Interhemispheric correlation of late Pleistocene glacial events. *Science* **269,** 1541–1549.

Mack, R. N., Bryant, V. M., and Pell, W. (1978). Modern forest pollen spectra from eastern Washington and northern Idaho. *Botanical Gazette* **138,** 249–255.

Manabe, S., and Stouffer, R. J. (1995). Simulation of abrupt climate change induced by freshwater input to the North Atlantic Ocean. *Nature* **378,** 165–167.

Mathewes, R. W. (1993). Evidence for Younger Dryas-age cooling on the North Pacific coast of America. *Quaternary Science Reviews* **12,** 321–331.

Mazaud, A., Laj, C., Bard, E., Arnold, M., and Tric, E. (1991). Geomagnetic field control of ^{14}C production over the last 80 ky: Implications for the radiocarbon time-scale. *Geophysical Research Letters* **18,** 1885–1888.

McAndrews, J. H., Berti, A. A., and Norris, G. (1973). "Key to the Quaternary Pollen and Spores of the Great Lakes Region." Life Sciences Miscellaneous Publications, Royal Ontario Museum Toronto, Canada.

Mikolajewicz, U., Crowley, T. J., Schiller, A., and Voss, R. (1997). Modelling teleconnections between the North Atlantic and Pacific during the Younger Dryas. *Nature* **387,** 384–387.

Millspaugh, S. H., and Whitlock, C. (1995). A 750-year fire history based on lake sediment records in central Yellowstone National Park. *Holocene* **5,** 283–292.

Minore, D. (1979). "Comparative Autecological Characteristics of Northwestern Tree Species—A Literature Review." United States Department of Agriculture Forest Service General Technical Report PNW-87.

Mock, C. J. (1996). Climatic controls and spatial variations of precipitation in the western United States. *Journal of Climate* **9,** 1111–1125.

Moore, P. D., and Webb, J. A. (1978). "An Illustrated Guide to Pollen Analysis." Wiley, New York.

National Oceanic and Atmospheric Administration (1964–1991). Multiple summaries of climatological data, Oregon section. Asheville, NC.

Pearson, G. W., and Stuiver, M. (1993). High-precision bidecadal calibration of the radiocarbon time scale, 500–2500 BC. *Radiocarbon* **35,** 25–33.

Peltier, W. R. (1994). Ice age paleotopography. *Science* **265,** 195–201.

Porter, S. C. (1976). Pleistocene glaciation in the southern part of the North Cascade Range, Washington. *Geological Society of America Bulletin* **87,** 61–75.

Porter, S. C. (1978). Glacier Peak tephra in the North Cascade Range, Washington: Stratigraphy, distribution, and relationship to late-glacial events. *Quaternary Research* **10,** 30–41.

Sea, D. S., and Whitlock, C. (1995). Postglacial vegetation and climate of the Cascade Range, central Oregon. *Quaternary Research* **43,** 370–381.

Stuiver, M., and Becker, B. (1993). High-precision decadal calibration of the radiocarbon time scale, AD 1950–6000 BC. *Radiocarbon* **35,** 35–65.

Stuiver, M., and Pearson, G. W. (1993). High-precision bidecadal calibration of the radiocarbon time scale, AD 1950–500 BC. and 2500–6000 BC. *Radiocarbon* **35,** 1–23.

Stuiver, M., and Reimer, P. J. (1993). Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program. *Radiocarbon* **35,** 215–230.

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, Jr., J. E. Kutzbach, W. F. Ruddiman, F. A. Street-Perrott, T. Webb III, and P. J. Bartlein, Eds.), pp. 468–513. Univ. of Minnesota Press, Minneapolis, MN.

Törnqvist, T. E., De Jong, A. F. M., Oosterbaan, W. A., and Van Der Borg, K. (1992). Accurate dating of organic deposits by AMS ^{14}C measurement of macrofossils. *Radiocarbon* **34,** 566–577.

Tsukada, M., Sugita, S., and Hibbert, D. M. (1981). Paleoecology of the Pacific Northwest I. Late Quaternary vegetation and climate. *Verhandlungen der Internationalen Vereingung für theoretische und angewandte Limnologie* **21,** 730–737.

Whitlock, C., and Bartlein, P. J. (1997). Vegetation and climate change in northwest America during the past 125 kyr. *Nature* **388,** 57–61.

Worona, M. A., and Whitlock, C. (1995). Late-Quaternary vegetation and climate history near Little Lake, central Coast Range, Oregon. *Geological Society of America Bulletin* **107,** 867–876.

Wright, H. E., Jr., Mann, D. H., and Glaser, P. H. (1983). Piston cores for peat and lake sediments. *Ecology* **65,** 657–659.