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Evidence for Millennial-Scale Climate Change During Marine Isotope Stages 2 and 3 at Little Lake, Western Oregon, U.S.A.

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Pollen and geochemical data from Little Lake, western Oregon, suggest several patterns of millennial-scale environmental change during marine isotope stage (MIS) 2 (14,100–27,600 cal yr B.P.) and the latter part of MIS 3 (27,600–42,500 cal yr B.P.). During MIS 3, a series of transitions between warm- and cold-adapted taxa indicate that temperatures oscillated by ca. 2°–4°C every 1000–3000 yr. Highs and lows in summer insolation during MIS 3 are generally associated with the warmest and coldest intervals. Warm periods at Little Lake correlate with warm sea-surface temperatures in the Santa Barbara Basin. Changes in the strength of the subtropical high and the jet stream may account for synchronous changes at the two sites. During MIS 2, shifts between mesic and xeric subalpine forests suggest changes in precipitation every 1000–3000 yr. Increases in *Tsuga heterophylla* pollen at 25,000 and 22,000 cal yr B.P. imply brief warmings. Minimum summer insolation and maximum global ice-volumes during MIS 2 correspond to cold and dry conditions. Fluctuations in precipitation at Little Lake do not correlate with changes in the Santa Barbara Basin and may be explained by variations in the strength of the glacial anticyclone and the position of the jet stream.

Key Words: **Pacific Northwest; pollen records; millennial-scale climate change; marine isotope stages 2 and 3; paleoecology.**

INTRODUCTION

Millennial-scale climate changes are evident in paleoclimate records from the late Holocene to the early Pleistocene (Bond *et al.*, 1997; Raymo *et al.*, 1998). However, much of this research has focused on the last glacial period (125,000–14,000 cal yr B.P.), which in many parts of the world is characterized by a series of abrupt and extreme climate oscillations. Recent studies suggest that some of these changes were global in extent and involved complex interactions between the atmosphere, oceans, biosphere, and cryosphere (see review by Alley, 1998). Despite these advances, questions remain about the origin and global occurrence of millennial-scale climate change.

In western North America, records from the Great Basin and the Pacific Northwest describe millennial-scale changes in pluvial lake levels, glacial extent, and vegetation during the last glacial period (Benson, 1999; Hicock *et al.*, 1999; Lin *et al.*, 1998; Whitlock and Grigg, 1999). In the Northeast Pacific, cores from the Santa Barbara Basin and the central California and southern Oregon margins show variations in sea-surface temperatures and oceanic circulation (Behl and Kennett, 1996; Gardner *et al.*, 1997; Hendy and Kennett, 1999; Lund and Mix, 1998). Millennial-scale variations in these regions have primarily been associated with shifts in the position and strength of the Eastern Pacific subtropical high, the Aleutian low, and the jet stream. However, fluctuations in the ventilation of intermediate and deep water in the North Pacific indicate that changes in North Pacific oceanic circulation may also have been involved (Lund and Mix, 1998). In addition, there is evidence that millennial-scale climate change in western North America may be partially regulated by orbital-scale controls, such as insolation and the size of the Laurentide ice sheet (Clark and Bartlein, 1995).

Vegetation changes in south-central Washington and glacial advances in southwestern British Columbia imply a series of climate oscillations that occurred ca. every 5000–8000 cal yr during the last glacial period in the Pacific Northwest (Hicock *et al.*, 1999; Whitlock and Grigg, 1999). However, records from the Santa Barbara Basin and the Great Basin show a higher frequency of climate change (ca. every 1000–3000 cal yr; Benson, 1999; Lin *et al.*, 1998). The absence of a higher frequency climate change in Pacific Northwest records may reflect a relatively coarse sampling resolution, or it may suggest different patterns of change for separate regions of western North America and the Northeast Pacific. A better understanding of millennial-scale climate change in the Pacific Northwest will

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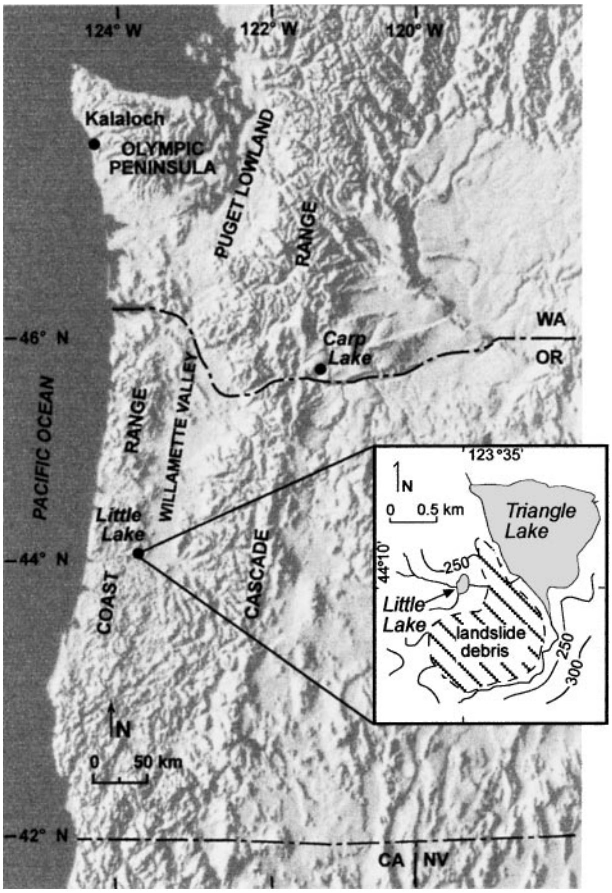

FIG. 1. Map and inset showing location and topography of Little Lake. Other sites mentioned in the text are also shown.

help identify the mechanisms responsible for these changes in western North America and the Northeast Pacific.

In this study, we present high-resolution palynological, lithologic, and geochemical data from Little Lake in the central Coast Range of Oregon (Fig. 1). This record extends from 42,500 to 27,600 cal yr B.P. and corresponds to the latter part of marine isotope stage (MIS) 3 and MIS 2. We describe the character, timings, and frequency of climate changes at Little Lake and evaluate the influence of orbital-scale climate controls. This study also aims to establish some proximal causes for these changes by comparing the Little Lake record with a high-resolution record from the Santa Barbara Basin (Behl and Kennett, 1996). Similar patterns of climate change in these two regions would imply the occurrence of large-scale shifts in atmospheric circulation over the North Pacific during the last glacial period. Differences would suggest regional-scale changes or the greater influence of specific climate controls, such as a glacial anticyclone, on one region.

Little Lake (44°10'N, 123°35'W, 217 m elevation), located 45 km east of the Pacific Ocean, was formed ca. 42,500 cal yr B.P. by a landslide (Fig. 1; Grigg and Whitlock, 1998; Long *et al.*, 1998; Worona and Whitlock, 1995). At present, Little Lake lies 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; Franklin and Dyrness, 1988). The climate of the Coast Range is characterized by mild, wet winters and warm, dry summers. Mid-latitude cyclones associated with the southward displacement of the jet stream result in 75–85% of the annual precipitation occuring in the winter (Mock, 1996). The subtropical high and the jet stream shift northward in the summer, bringing warm, dry conditions to the Pacific Northwest.

METHODS

A 17.25-m-long core was collected with a 5-cm-diameter piston sampler (Wright *et al.*, 1983) from the fen surrounding Little Lake. In the laboratory, cores were sliced longitudinally, and the lithology was described. From 10.56 to 17.22 m depth, samples of 1 cm^3 were taken for pollen analysis every 4 to 6 cm. From 11.48 to 17.22 m depth, samples of ca. 6–10 cm^3 of sediment were taken every 4 to 8 cm and analyzed for carbon content and major and trace elements. The remaining core was used for magnetic susceptibility measurements, macrofossil analysis, and radiocarbon dating.

Pollen samples were processed using standard procedures (Cwynar *et al.*, 1979; Faegri *et al.*, 1989). A known amount of *Lycopodium* spores was added to each sample to calculate the concentration of pollen (grains/cm^3). Pollen was examined at magnifications of 400× and 1000× and identified using modern pollen collections and published atlases (McAndrews *et al.*, 1973; Moore and Webb, 1978). *Pinus* (pine) pollen grains were identified as haploxylon-type and diploxylon-type when the distal membrane of the pollen grain was preserved. Those grains without an intact distal membrane were classified as "undifferentiated." *Pinus*-types, *Abies* (fir), *Picea* (spruce), and Cupressaceae (juniper family) could not be identified to the species level without the presence of macrofossils. However, modern ecological associations and phytogeography were used to determine which species were the most likely contributors. Pollen percentages were calculated based on the sum of terrestrial pollen and spores. Pollen accumulation rates (PAR; grains/cm^2/yr), pollen concentrations (grains/cm^3), and deposition times (yr/cm) were calculated using the TILIA program (Grimm, 1988). Pollen zones from Worona and Whitlock (1995) were further divided into subzones based on the results of a constrained cluster analysis (CONISS; Grimm, 1988).

Measurements of sediment magnetism indicate changes in the input or preservation of minerogenic clasts to a lake and imply variations in the allochthonous component or anoxic-vs.-oxic conditions, which can be used to infer environmental changes (Thompson and Oldfield, 1986). Magnetic susceptibility was measured every 2 cm on 6-cm^3 subsamples of sediment using a cup-coil magnetic susceptibility instrument. A software program, Magsus.pas (David Adam and Sarah Shafer, personal communication, 1993) recorded and averaged four, 4-s readings of each sample. The results were converted to concentration values of electromagnetic units (emu/cm3). Total carbon (C) was measured by combusting untreated samples of dried and crushed sediment at 960°C (Engleman *et al.*, 1985). The CO_2 released during this process was measured by titration in a coulometer cell. Samples with the highest concentration of C were analyzed by coulometry for inorganic C. These analyses and loss-on-ignition analyses (Worona and Whitlock, 1995) indicated a negligible percentage of $CaCO_3$, which sugests that most of the total C in Little Lake sediments is organic. Nitrogen concentrations were determined using a Carlo Erba NA 1500 analyzer. Samples were combusted in oxygen at 1000°C, and the resulting NO was reduced to N_2. N_2 was separated by chromatography and measured with a thermal conductivity detector. Sediments were also analyzed for 21 major and trace elements by induction-coupled, argon plasma emission spectrometry (ICP; Litchie *et al.*, 1987).

The reconstruction of vegetation and climate from pollen percentages and accumulation rates was based on recent compilations of modern pollen, vegetation, and climate data from the Pacific Northwest (Minckley and Whitlock, 2000; Pellat *et al.*, 1997). A square-chord measurement of dissimilarity (Overpeck *et al.*, 1985) was used to compare modern and fossil pollen spectra of the 12 most abundant tree types. Square-chord values of less than 0.15, of 0.15–0.20, and of 0.20–0.30 were used to identify "good analogues," "probable analogues," and "possible or weak analogues," respectively (see method described by Anderson *et al.*, 1989). Most analogues between 42,500 and 27,600 cal yr B.P., were good to probable, which allowed for the estimation of specific paleoclimate parameters (Table 1).

TABLE 1
Description, vegetation type, and inferred climate of pollen subzones within Zone LL-1

Pollen subzone age (cal yr B.P.) depth (m)	Pollen description	Vegetation type and square-chord distances	Inferred climate
LL-1h 26,490–27,810 cal yr B.P. 14.36–14.64 m	High percentages of *Pinus* (haploxylon- and diploxylon-types) (20–60%) and Abies (15–20%); increases in percentages of *Picea* (5–20%), *T. mertensiana* (5–10%), and herbs (10–25%) at end of subzone. Total accumulation of terrestrial pollen and spores (PAR): 1000–1900 g/cm^2/yr.	Montane to subalpine forest; Sq. chord = 0.15– 0.20 (probable analogue)	Cold, wet
LL-1g 27,810–28,300 cal yr B.P. 14.64–14.74 m	High percentages of *Tsuga heterophylla* (40%) and *Abies* (15%); small increases in percentages of Cupressaceae (8%) and *Pseudotsuga*-type (2–5%). PAR: 1700–2900 g/cm^2/yr.	Montane forest; Sq. chord <0.15 (good analogue)	Cool, wet
LL-1f 28,300–32,450 cal yr B.P. 14.74–15.54 m	High percentages of *Pinus* (haploxylon-type) (30–80%) and small increases in herbs (5–15%) and at the top of the subzone, *Picea* (2–5%) and *Tsuga mertensiana* (2–5%) alternate with peaks in *Abies* (5–25%), *T. heterophylla* (5–15%), *Pseudotsuga*-type (2%) and Cupressaceae (2–5%); moderate percentages of *Alnus sinuata*-type (5–15%) and *Alnus rubra*-type (2–5%) PAR: 1350–5400 g/cm^2/yr.	Alternating *Pinus*-dominated and temperate forests; Sq. chord = 0.15–0.20 (probable analogue)	Alternating cold, dry and cool, wet
LL-1e 32,450–33,330 cal yr B.P. 15.54–15.70 m	Significant increases in percentages of *Abies* (15–25%), *T. heterophylla* (5–35%), *Pseudotsuga*-type (5–7%) Cupressaceae (5–10%) and *Alnus rubra*-type (5–10%); decline in percentages of *Pinus* (haploxylon-type) (20–35%). PAR: 2500–4600 g/cm^2/yr.	Temperate forest; Sq. chord = 0.15–0.20 (probable analogue)	Warm, wet
LL-1d 33,330–35,870 cal yr B.P. 15.70–16.14 m	Generally high percentages of *Pinus* (haploxylon-type; 25–60%), herbs (10–20%), and Abies (10–15%); small increases in percentages of *T.mertensiana* (2–5%), *Picea* (2–5%) and *Alnus sinuata*-type (5–10%). PAR: 1350–3250 g/cm^2/yr.	*Pinus*-dominated subalpine forest; Sq. chord <0.15 (good analogue)	Cold, wet
LL-1c 35,870–39,500 cal yr B.P. 16.14–16.74 m	Generally high percentages of *Pinus* (haploxylon-type; 15–65%) alternate with peaks in *T. heterophylla* (10–30%) and *Abies* (5–20%); moderate percentages of herbs (15–20%), Cupressaceae and *Alnus sinuata*-type (5–10%); percentages of *Alnus rubra*-type decline and *T. mertensiana* (5%) and *Picea* (5–7%) briefly increase toward the bottom of the subzone. PAR: 900–4100 g/cm^2/yr.	Alternating *Pinus*-dominated and montane forests; Sq. chord = 0.15–0.20 (probable analogue)	Alternating cool, wet and cold, dry
LL-1b 39,500–40,310 cal yr B.P. 16.74–16.86 m	High percentages of *T. heterophylla* (35%), *Pseudotsuga*-type (5–10%), and Cupressaceae (10–15%); decline in percentages of *Pinus* (haploxylon-type; 10–15%) and *Alnus sinuata*-type (<5%). PAR: 1500–2100 g/cm^2/yr.	Temperate forest; Sq. chord = 0.15–0.20 (probable analogue)	Warm, wet
LL-1a 40,310–42,680 cal yr B.P. 16.86–17.22 m	High percentages of *Pinus* (haploxylon-type; 25–60%) alternate with high percentages of *T. heterophylla* (20–30%) and *Abies* (15–35%); moderate percentages of *Alnus sinuata*-type (5–10%). PAR: 1000–2600 g/cm^2/yr.	*Pinus*-dominated forest; Sq. chord <0.15 (good analogue)	Cool, wet

Note. PAR, pollen accumulation rate.

TABLE 2
Description, Vegetation Type, and Inferred Climate of Pollen Subzones within Zone LL-2

Pollen subzone age (cal yr B.P.) depth (m)	Pollen description	Vegetation type and square-chord distances	Inferred climate
LL-2h 14,140–15,730 cal yr B.P. 10.56–11.30 m	High percentages of *Alnus sinuata*-type (10–15%) and herbaceous taxa (20%); increasing percentages of *Tsuga heterophylla* (10%) and Pteridophytes (10%) toward top of subzone. Total accumulation of terrestrial pollen and spores. PAR: 2100–5300 g/cm^2/yr.	Developing montane forest; Sq. chord <0.15 (good analogue)	Cool, wet
LL-2g 15,730–17,250 cal yr B.P. 11.30–11.88 m	Base of zone begins with high percentages of *Picea* (20–25%) and is replaced by *Tsuga mertensiana* (20–25%), *Abies* (15–35%), and herbaceous taxa (25%); percentages in the above taxa decline after 16,540 cal yr B.P. when percentages of *T. heterophylla* (10%), *Alnus sinuata*-type (5%), and Pteridophytes (15–20%) increase. PAR: 1200–3100 g/cm^2/yr.	Wet, subalpine to montane forests; Sq. chord = 0.20–0.30 (possible analogue)	Cold, wet changing to cool, wet
LL-2f 17,250–20,220 cal yr B.P. 11.88–12.82 m	Generally high percentages of herbs (15–40%) dominate this zone; periodic increases in *Pinus* (diploxylon-type) (15–35%), *Picea* (10–20%), *T. mertensiana* (5–20%), and *Abies* (10–20%) also characterize this subzone. PAR: 740–2300 g/cm^2/yr.	Dry, subalpine forest and parkland; Sq. chord = 0.20–0.30 (possible analogue)	Coldest and driest
LL-2e 20,220–20,790 cal yr B.P. 12.82–12.98 m	High percentages of herbs (25%), *T. mertensiana* (20%), and *Picea* (20%). PAR: 1200–2400 g/cm^2/yr.	Wet, subalpine forest; sq. chord = 0.20–0.30 (possible analogue)	Cold, wet
LL-2d 20,790–23,930 cal yr B.P. 12.98–13.78 m	Generally high percentages of *Pinus* (diploxylon-type) (20–60%) alternate with peaks in *Picea* (10–20%), *T. mertensiana* (5–20%), *Abies* (5–20%), and small increases in *T. heterophylla* (5–10%); high percentages of herbs (15–25%) separate peaks in *Pinus* and associated taxa. PAR: 830–2670 g/cm^2/yr.	Alternating wet, subalpine and dry subalpine forests; Sq. chord = 0.20–0.30 (possible analogue)	Alternating cold, wet and cold, dry
LL-2c 23,930–24,960 cal yr B.P. 13.78–14.02 m	High percentages of *Pinus* (haploxylon-type; 20–60%) and herbs (5–20%). PAR: 900–2000 g/cm^2/yr.	Dry subalpine forest/parkland; Sq. chord = 0.20–0.30 (possible analogue)	Cold, dry
LL-2b 24,960–25,940 cal yr B.P. 14.02–14.24 m	High percentages of *Pinus* (diploxylon-type; 20–40%) and Abies (10–20%); moderate increase in percentages of *T. heterophylla* (5–15%) at end of subzone. PAR: 500–2500 g/cm^2/yr.	Montane forest; Sq. chord = 0.20–0.30 (possible analogue)	Cool, wet
LL-2a 25,940–26,490 cal yr B.P. 14.24–14.36 m	High percentages of *Picea* (20–30%), *T. mertensiana* (30%), and *Abies* (20–30%). PAR: 1200–2100 g/cm^2/yr.	Wet, subalpine forest; Sq. chord = 0.15–0.20 (probable analogue)	Cold, wet

Note. PAR, pollen accumulation rate.

Between 27,600 and 14,100 cal yr B.P., most analogues were possible or weak, which limited reconstructions to more qualitative estimates that were based on established relationships between present-day pollen percentages, vegetation abundance, and climate for specific taxa (Table 2; Minckley and Whitlock, 2000; Pellat *et al.*, 1997; Thompson *et al.*, 1999).

CHRONOLOGY

The chronology for the last glacial period at Little Lake is based on 11 accelerator mass spectrometry (AMS) radiocarbon dates of charcoal and plant macrofossils (Table 3). Radiocarbon ages <20,300 ^{14}C yr B.P. were calibrated using the CALIB 4.0 program (Stuiver and Reimer, 1993). Radiocarbon ages from 20,300 to 28,000 ^{14}C yr B.P. were calibrated using a lake varve data set from Lake Suigetsu, Japan (Kitagawa and van der Plicht, 1998). Calibrated age-ranges were derived using Method B from CALIB 4.0, which constructs a probability distribution of calibrated ages for any given radiocarbon age (see Stuiver and Reimer, 1993). This method includes both the sample and calibration curve error in the calculation of calibrated age-ranges.

For radiocarbon ages >28,000 ^{14}C yr B.P., a calibration curve was estimated using the Lake Suigetsu data set, a U/Th and ^{14}C dated coral sample from New Guinea, and paleomagnetic data that indicate atmospheric Δ^{14}C changes close to zero at 50,000 ^{14}C yr B.P. (Kitagawa and van der Plicht, 1988; Bard

TABLE 3
Radiocarbon Dates and Calibrated Ages from the Last Glacial Period for Little Lake

Depth (m)	Material dated	Lab No.	Radiocarbon age (^{14}C yr B.P.)	[a]Calibrated age (cal yr B.P.) and 2σ range
11.67–11.70	charcoal, wood	AA-27839	14,200 ± 170	17,030 (16590–17540)
11.89–11.93	charcoal, wood	Beta-111337	13,140 ± 160	15,650 (15,060–16,150)
11.89–11.93	lake sediment	Beta-111336	15,340 ± 130	18,260 (17,950–18,560)
12.48–12.51	charcoal, wood	AA-27840	15,220 ± 150	18,140 (17,790–18,480)
13.41–13.45	charcoal, wood	AA-27841	19,440 ± 340	23,100 (22,480–23,890)
13.99–14.01	charcoal, wood	AA-27842	22,790 ± 570	26,320 (25,130–27,730)
14.69–14.73	charcoal, wood	AA-27843	26,360 ± 540	29,741 (28,230–31,570)
15.49–15.52	charcoal, wood	AA-27844	25,540 ± 700	28,960 (27,370–30,830)
16.48–16.52	charcoal, wood	AA-31360	32,030 ± 940	36,260 (35,160–37,500)
17.26	wood	AA-27846	42,700 ± 1700	44,720 (43,800–45,780)

[a]Calibrated ages were used to derive the age model for Little Lake. See text for discussion of calibration.

et al., 1998; Mazuad *et al.*, 1991). From 28,000 to 50,000 ^{14}C yr B.P., a discrepancy of 5000 yr exists between the lake and coral data sets (Kitagawa and van der Plicht, 1998). To utilize all existing data sets, a curve was fit that approximated between the coral and lake varve data sets and assumed a Δ^{14}C of zero at 50,000 ^{14}C yr B.P. Calibrated age-ranges were derived using a probability distribution, and a second-order polynomial was used to construct an age vs. depth model for the calibrated ages (Fig. 2).

RESULTS

Lithology, Geochemical, and Magnetic Susceptibility Data

The bedrock surrounding Little Lake consists of noncarbonate volcanic sedimentary rocks. The core contains mostly inorganic clay, silt, and fine sand with discontinuous lenses of organic material (2%–3% organic carbon (OC)) from 10.56 to 14.62 m depth. The sediment from 14.62 to 16.42 m depth is similar but with more discontinuous lenses of organic material (2–4% OC), and a section of gyttja (3–5% OC) from 16.04 to 16.35 m depth. Results of ICP elemental analyses of common lithophile elements (e.g., Al, K, Mg, Ti, Li, Ni, V, and Z) did not reveal any significant patterns of variability, suggesting that the source of detrital clastic material into Little Lake remained constant.

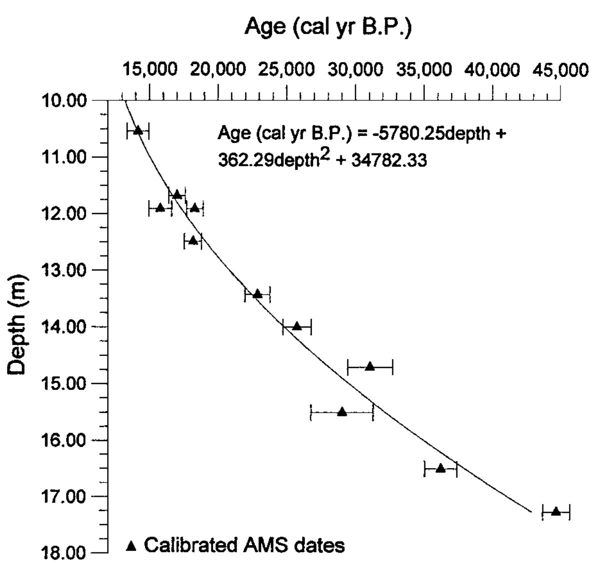

FIG. 2. Age-vs.-depth curve and regression equation for the last glacial period at Little Lake. Age model was developed using calibrated radiocarbon ages, which are shown with two sigma error bars. See text for discussion of calibration.

The concentration of OC in sediments deposited during Zone LL-1 fluctuates between 1.75 and 5% (Fig. 3). Peaks in total N generally correspond to peaks (>3%) in OC, with an overall correlation coefficient of 0.85. Values of C/N in sediments deposited during Zone LL-1 are mostly >10 (up to 16), suggesting that much of the organic matter was likely from terrestrial allochthonous sources (Meyers and Ishiwatari, 1993). Total S concentration averages >0.2% in sediments deposited during Zone LL-1, with peaks of up to 0.5% corresponding with peaks in C concentration ($r = 0.80$, $n = 71$). The concentration of Fe is low and invariant in sediments deposited during Zone LL-1, which indicates that the S is likely from organic sources, rather than from Fe sulfide mineral. High concentrations of OC and S are concurrent with low values of magnetic susceptibility. These data suggest that the preservation of organic matter periodically increased as a result of increased anoxia in the bottom waters.

During Zone LL-2, OC and S concentrations are relatively constant with an average of ca. 2% and 0.1%, respectively. Values of N average ca. 0.2, and C/N ratios average 9.5, suggesting that the organic matter is mostly autochthonous (Meyers and Ishiwatari, 1993). Generally high values of magnetic susceptibility in sediments deposited during Zone LL-2, together with low OC and S concentrations, indicate decreased preservation of organic matter and the occurrence of oxic conditions. These data are consistent with high and variable concentrations of Fe and Mn, which imply that the hypolimnion and sediment pore waters were well oxygenated, resulting in the preservation of oxidized Fe and Mn minerals.

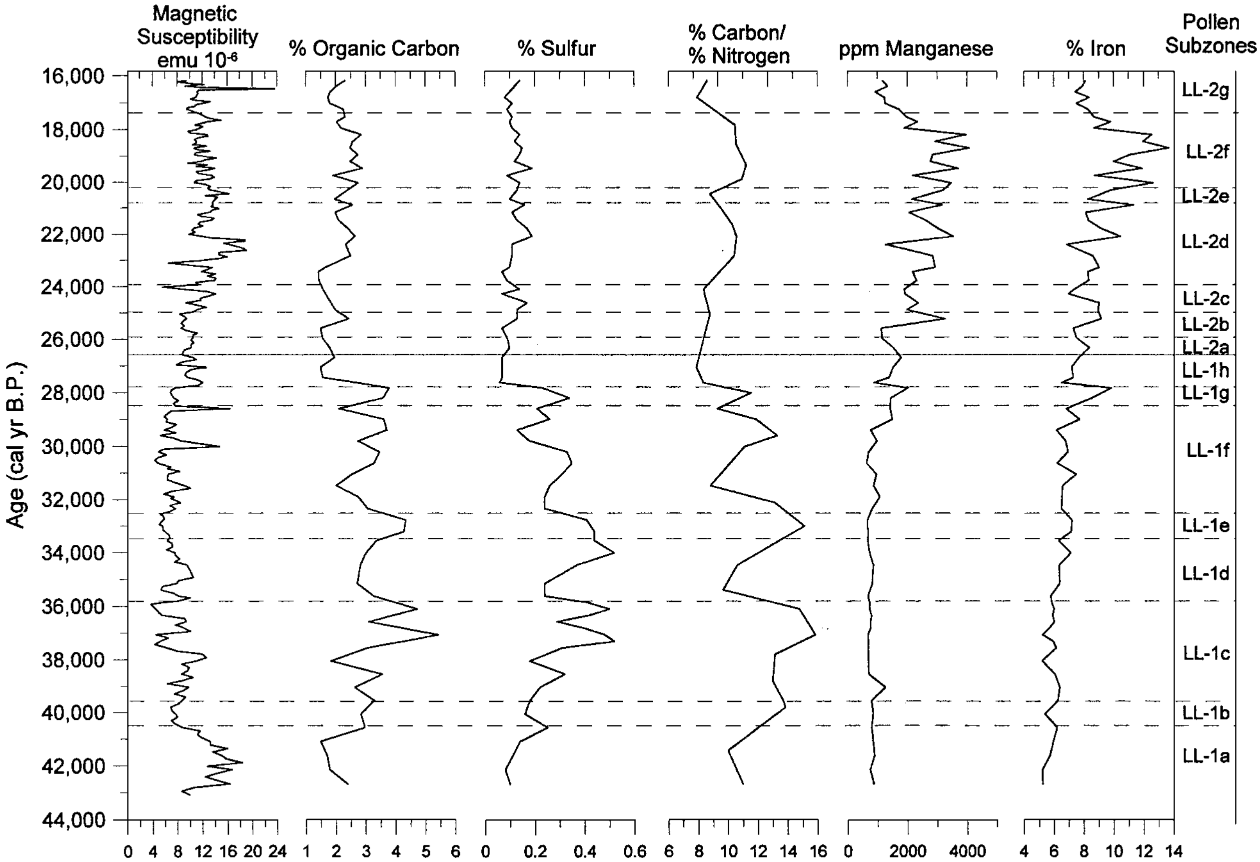

FIG. 3. Magnetic susceptibility and geochemical data from Little Lake.

Vegetation and Climate Reconstructions

The last glacial period at Little Lake is represented by pollen zones LL-1 (42,500–27,000 cal yr B.P.) and LL-2 (27,000–15,000 cal yr B.P.; Worona and Whitlock, 1995), corresponding to the latter part of MIS 3 and to MIS 2, respectively. Zone LL-1 is generally characterized by high percentages and accumulation rates of *Pinus* (mostly haploxylon-type), *Abies*, and *T. heterophylla* pollen (Table 1; Figs. 4 and 5). Zone LL-2 is dominated by generally high pollen percentages and accumulation rates of *Picea*, *T. mertensiana*, *Abies*, *Pinus* (mostly diploxylon-type), and herbaceous taxa (Table 2; Figs. 6 and 5).

Picea pollen in Zone LL-1 is most likely from *Picea engelmannii* (Engelmann spruce) based on its association with *T. mertensiana* and *Abies* in Oregon today. However, the presence of *P. sitchensis* (Sitka spruce) cannot be ruled out, because it currently grows with *T. mertensiana* in coastal British Columbia (Meidinger and Pojar, 1991). Haploxylon-type *Pinus* pollen is likely from *P. monticola* based on its occurrence with *T. heterophylla* and *Abies*. *Pseudotsuga*-type pollen is probably from *Pseudotsuga menziesii* rather than *Larix* (larch), which is rare west of the Cascade Range. The association of Cupressaceae with *T. heterophylla* and *Pseudotsuga*-type pollen suggests that it is probably from *Thuja plicata* or *Libocedrus decurrens* (incense cedar). *Abies* pollen is not consistently associated with any one taxa, which implies that it could be from a number of common species, including *A. amabilis* (Pacific silver fir), *A. grandis*, *A. procera* (noble fir), and *A. lasiocarpa* (subalpine fir). Most of the fossil pollen assemblages from Zone LL-1 are similar to modern pollen spectra from montane (ca. 900–1500 m elevation) and subalpine (ca. 1300–2000 m elevation) forests from the Cascade Range in Oregon and Washington (Franklin and Dyrness, 1988; Table 1; Figs. 4 and 6). The presence of such forests at Little Lake indicates conditions were ca. 4°–8°C cooler than today and about as wet as present. Wet conditions are consistent with C/N values, which suggest an allochthonous OC source and increased erosion. Generally high but variable values of OC and S and low magnetic susceptibility imply anoxic conditions that were probably related to relatively warm conditions and long periods of thermal stratification.

During Subzone LL-1a, high percentages of *T. heterophylla* and Cupressaceae pollen are replaced by those of *Abies* and *Pinus* (mostly haploxylon-type) pollen at ca. 42,250 cal yr B.P. This shift in dominance reflects a transition from montane to subalpine forest and from cool to cold conditions. High magnetic susceptibility and low percentages of OC suggest oxic conditions that were likely related to a relatively short period of thermal stratification (Fig. 3). The vegetation during Subzone

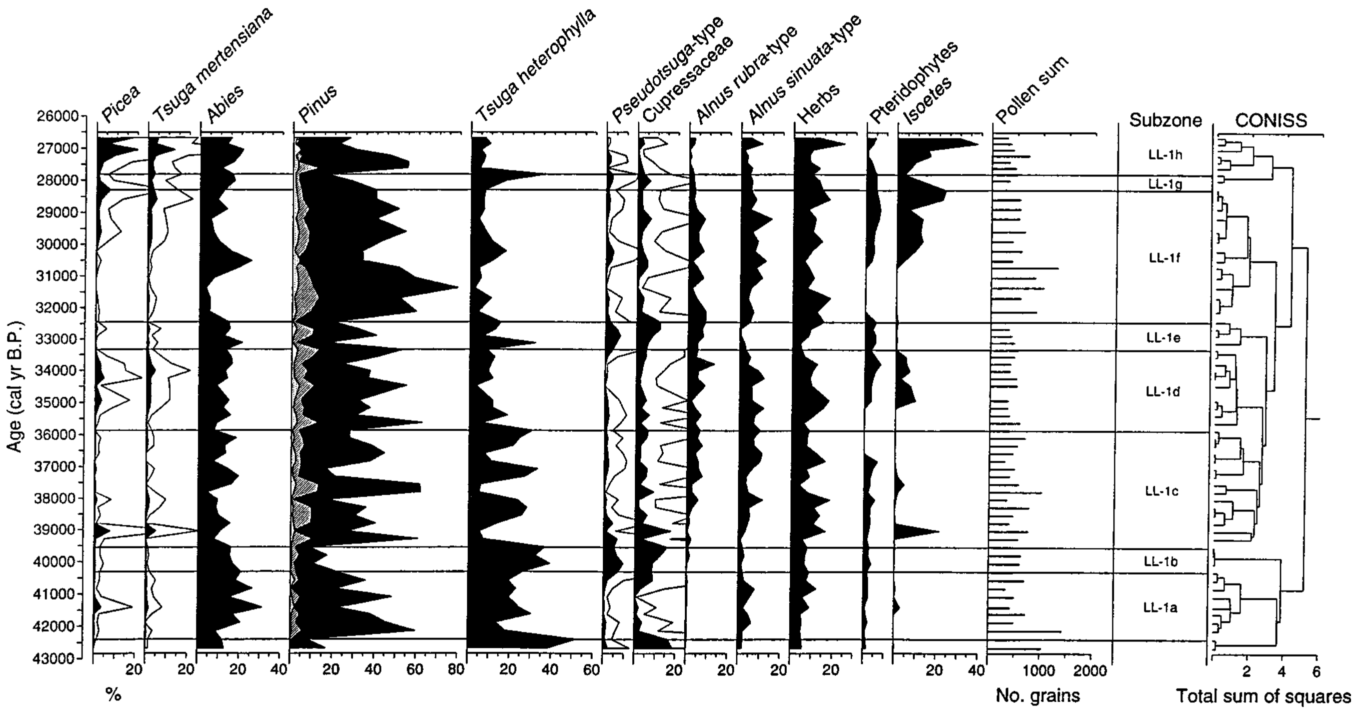

FIG. 4. Pollen percentage diagram for selected Little Lake taxa during the latter part of marine isotope stage 3. Open curves represent fivefold exaggeration of black curves. *Pinus* includes haploxylon-*type* (*P. monticola*) and is shown with a stripped curve, diploxylon-type (*P. contorta* or *P. ponderosa*) is represented by an open curve, and total *Pinus* is shown with a black curve. The results of a constrained cluster analysis are illustrated.

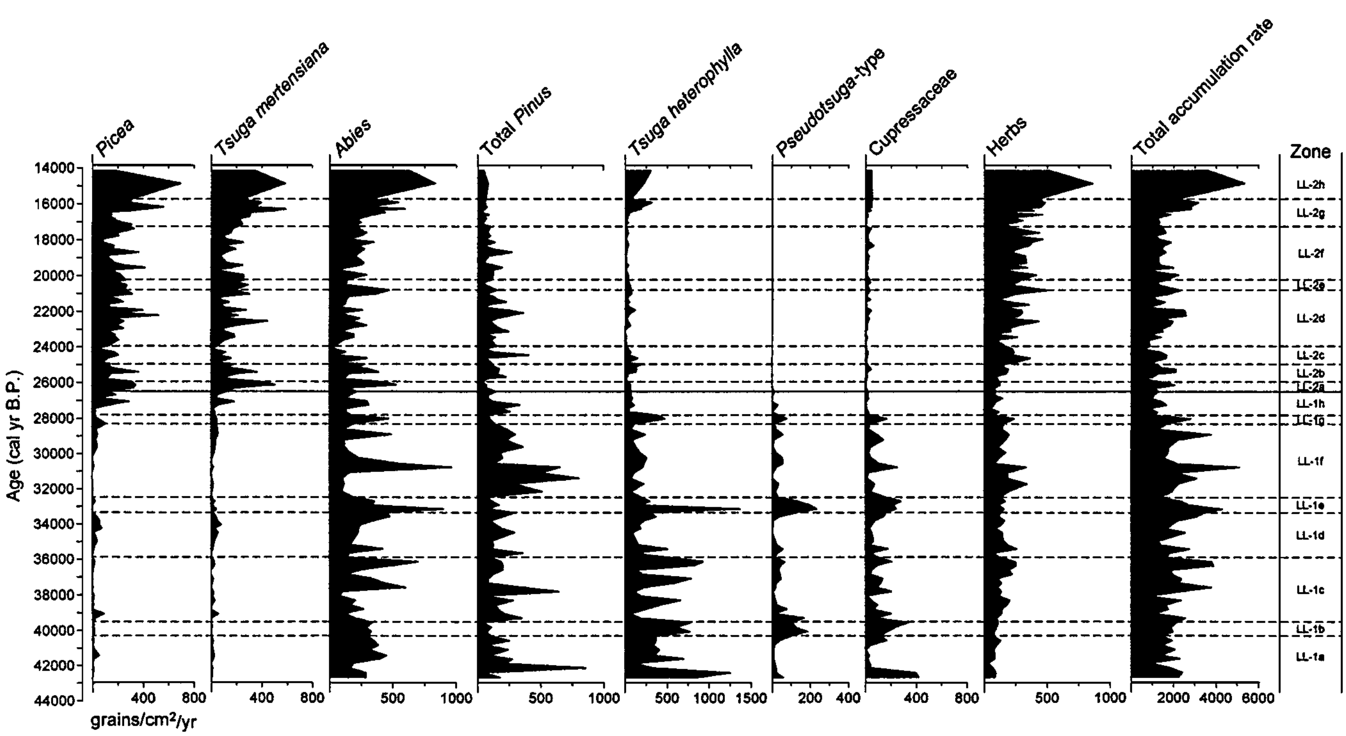

FIG. 5. Pollen accumulation rates for selected Little Lake taxa during zones LL-1 and LL-2.

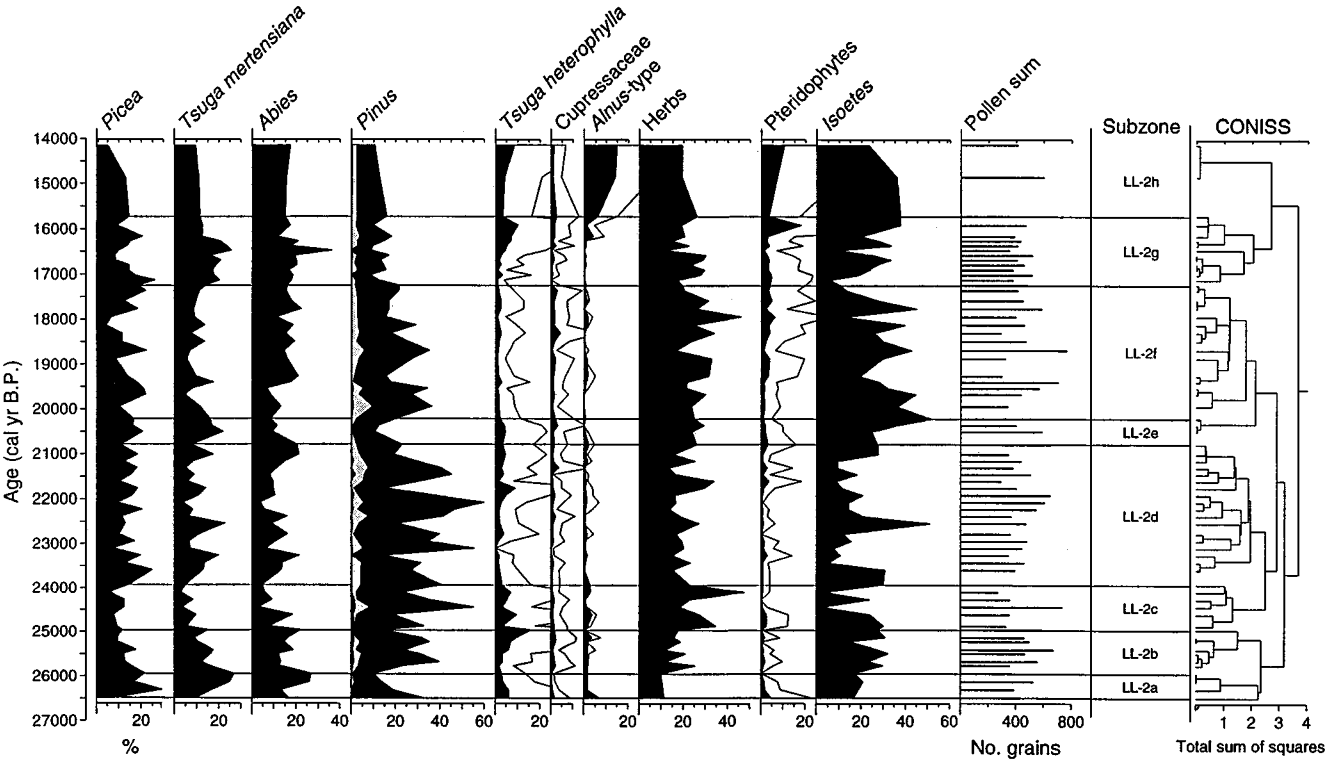

FIG. 6. Pollen percentage diagram for selected Little Lake taxa during marine isotope stage 2. Open curves represent fivefold exaggeration of black curves. *Pinus* includes haploxylon-*type* (*P. monticola* or *P. albicaulis*) and is shown with a stripped curve, diploxylon-type (*P. contorta* or *P. ponderosa*) is represented by an open curve, and total *Pinus* is shown with a black curve. The results of a constrained cluster analysis are illustrated.

LL-1b was probably similar to modern montane or low-elevation temperate forests based on high pollen percentages of *T. heterophylla*, *Pseudotsuga*-type, and Cupressaceae. These taxa imply some of the warmest conditions of Zone LL-1 (mean annual temperature estimated at 5°–6°C). Low magnetic susceptibility and high percentages of OC indicate a period of increased anoxia.

Subzone LL-1c is characterized by a series of fluctuations between *Pinus* (haploxylon-type) and *T. heterophylla* pollen that probably represent shifts between montane forests of *T. heterophylla* and those of *Pinus (P. monticola)*. Although *T. heterophylla* and *P. monticola* have overlapping climate ranges, *P. monticola* tends to grow in areas with colder winter temperatures and less summer precipitation than *T. heterophylla* (Thompson *et al.*, 1999). The high frequency fluctuations between *T. heterophylla* and *P. monticola* therefore imply that the climate oscillated three times from cool, wet conditions (mean annual temperatures estimated at 4°–5°C), to colder and possibly drier conditions (mean annual temperatures estimated at 3°–4°C). Peaks in OC and S and low magnetic susceptibility suggest increased anoxia during the *T. heterophylla*-dominated intervals.

Subzone LL-1d contains high percentages of *Pinus* (haploxylon-type) and *Abies* pollen, and minor amounts of *Picea* and *T. mertensiana* pollen, which indicate the establishment of a subalpine forest and one of the coldest periods of Zone LL-1 (annual temperatures estimated at 2°–4°C). Precipitation was probably equivalent to that of the present day, although more of it would have been in the form of snow. Subzone LL-1d is also characterized by low values of OC and S, which suggest relatively oxic conditions.

During Subzone LL-1e, *Pseudotsuga*-type, Cupressaceae, and *T. heterophylla* pollen increase significantly. This shift suggests a transition to a montane or low-elevation temperate forest and to warmer conditions than the previous subzone (mean annual temperatures estimated at 5°–6°C). At the base of this subzone, OC and S percentages increase and magnetic susceptibility values decrease, indicating a period of increased anoxia.

Pinus percentages reach their peak during Subzone LL-1f and imply a montane forest probably dominated by *P. monticola* and colder conditions than the previous subzone (mean annual temperatures estimated at 3°–4°C). Increased pollen percentages of *Abies* and *T. heterophylla* and minor amounts of *Pseudotsuga*-type and Cupressaceae at 30,500 cal yr B.P. indicate the establishment of a mixed montane and temperate forest and slightly warmer conditions. A peak in OC and S at 30,500 cal yr B.P. suggests anoxic conditions and a brief warming. PAR show an increase in *T. heterophylla*, *Pseudotsuga*-type, and Cupressaceae at 29,000 cal yr B.P. that coincides with a second peak in OC and S (Fig. 5). Thus, PAR and geochemical data imply another brief warm interval at 29,000 cal yr B.P. The return of *Pinus* pollen in the top of this subzone reflects the reestablishment of *P. monticola* and colder conditions. The latter part of Subzone LL-1f and Subzone LL-1g are characterized by an overall decrease in temperate and montane species (*T. heterophylla*, *Pseudotsuga*-type, Cupressaceae, and *Alnus*-types) and an increase in subalpine types (*Picea*, *T. mertensiana*, and *Abies*). This shift marks the transition from zones LL-1 to LL-2 and an estimated 2°–4°C decrease in temperatures. Superimposed on this larger-scale trend is an increase in pollen percentages of *T. heterophylla* during Subzone LL1-g, which indicate a change to montane forest and warmer conditions than before. By the end of Subzone LL-1h, temperate and montane taxa are replaced by subalpine taxa.

Abies pollen in Zone LL-2 could be from several species common to montane and subalpine forests in the Pacific Northwest (*A. amabilis*, *A. lasiocarpa*, or *A. procera*). *Picea* pollen is probably from *P. engelmannii* based on the identification of a macrofossil needle at 11.70 m (ca. 16,700 cal yr B.P.). Diploxylon-type *Pinus* pollen could either be from *P. contorta* (lodgepole pine) or *P. ponderosa* (ponderosa pine). However *P. contorta* is more commonly associated with subalpine species, such as *T. mertensiana*. The fossil pollen spectra from Zone LL-2 are similar to modern pollen assemblages from subalpine forests (ca. 1300–2000 m elevation) and upper treeline (ca. 1400–2200 m elevation) in the Pacific Northwest (Franklin and Dyrness, 1988; Pellat *et al.*, 1997; Table 2; Figs. 5 and 6). Climate estimates based on these modern analogues suggest that temperatures were 7°–11°C colder and precipitation was 250–500 mm less than present, values consistent with previous estimates (Worona and Whitlock, 1995). Dry conditions are also reflected in C/N values that indicate an autochthonous source for the OC and that imply less erosion. OC and S percentages are considerably lower than in Zone LL-1, whereas magnetic susceptibility is higher, reflecting more oxic conditions (Fig. 3).

Subzones LL-2a through LL-2e are characterized by repeated fluctuations between high pollen percentages of *Picea*, *Abies*, and *T. mertensiana* and those of *Pinus* (mostly diploxylon-type). The spectra dominated by *Picea*, *Abies*, and *T. mertensiana* pollen are similar to those from modern mesic subalpine forests in coastal British Columbia (Pellat *et al.*, 1997). The climate of this area is characterized by abundant year-round precipitation and cold temperatures. Intervals with abundant *Pinus* pollen are similar to *P. contorta*-dominated subalpine forests in Oregon and Washington and suggest a more xeric or seasonably wet subalpine forest. These fluctuations reflect a series of oscillations between cold-wet and cold-dry conditions. *T. heterophylla* pollen increases during subzones LL-2b and LL-2d (ca. 25,000 and 22,000 cal yr B.P.), suggesting more montane forests and slightly warmer and wetter conditions than before. Minor peaks in the percentages of OC and S overlap with these peaks in *T. heterophylla* and suggest relatively anoxic conditions. Increased herbaceous taxa at ca. 25,000 cal yr B.P. indicate the establishment of a more open forest and the onset of cold-dry conditions. This increase occurs at about the same time as levels of Fe, Mn, and magnetic susceptibility increase, implying that the lake was better mixed and more oxygenated, probably as a result of the more open forest and increased wind.

Few modern pollen assemblages are similar to Subzone LL-2f. However, based on the composition of modern alpine communities, this assemblage implies a subalpine parkland dominated by *A. lasiocarpa* and Poaceae (grass family), similar to those growing at present on the eastern slopes of the Washington Cascades and Olympic Mountains (Franklin and Dyrness, 1988). These environments suggest the coldest and driest conditions of this zone. Subzone LL-2g marks the return of high pollen percentages of *Picea*, *T. mertensiana*, and *Abies* and a decline in herbaceous taxa, which indicate the development of a closed and mesic subalpine forest. A sharp decline in Fe and Mn implies increased lake stratification and anoxia that probably resulted from a more protective forest cover around the lake and from warmer conditions. Subzone LL-2h shows the first major increase in warm-adapted taxa (*T. heterophylla* and *Alnus*-type) and a decrease in subalpine taxa. These changes suggest a shift to montane forest and the onset of the warm-wet conditions that mark the end of the last glacial period.

DISCUSSION

Summary of Changes at Little Lake and Comparison with Other Pacific Northwest Records

The sum of nonarboreal pollen percentages is used as a measure of changes in vegetation cover near Little Lake (Fig. 7). The establishment of a relatively open forest between 25,000 and 17,000 cal yr B.P. is concurrent with an increase in oxidizing conditions in the hypolimnion, perhaps as a result of less protective forest cover around the margins of the lake. Effective precipitation is represented by the sum of the wet-adapted taxa, *Abies* and *T. mertensiana*, which are both currently associated with high annual precipitation and actual/potential evaporation (Minckley and Whitlock, 2000). After the establishment of a more open forest, intervals of high effective precipitation are associated with peaks in magnetic susceptibility, perhaps indicating increased erosion within the watershed. Millennial-scale changes in temperature are reflected by the sum of warm-adapted taxa (*Pseudotsuga*-type, Cupressaceae, and *T. heterophylla*) over the sum of cold-adapted taxa (*Picea, T. mertensiana*, and Poaceae). In Zone LL-1, fluctuations in *Pinus* (probably *P. monticola*) parellel the cold-adapted taxa. However, *Pinus* pollen was not included in the cold sum because the taxon reflects dry conditions during Zone LL-2. The warm intervals shown by this ratio are usually supported by corresponding periods of increased anoxia.

Many of the millennial-scale climate changes identified at Little Lake are evident in other records from the Pacific Northwest (Fig. 1). A series of peaks in arboreal pollen at Carp Lake suggests four warm and/or wet periods (39,500, 36,000, 32,500, and 29,000 cal yr B.P.), which correlate with warm intervals during Zone LL-1 (Whitlock and Bartlein, 1997). A record from the Kalaloch sea cliffs on the Olympic Peninsula, Washington,

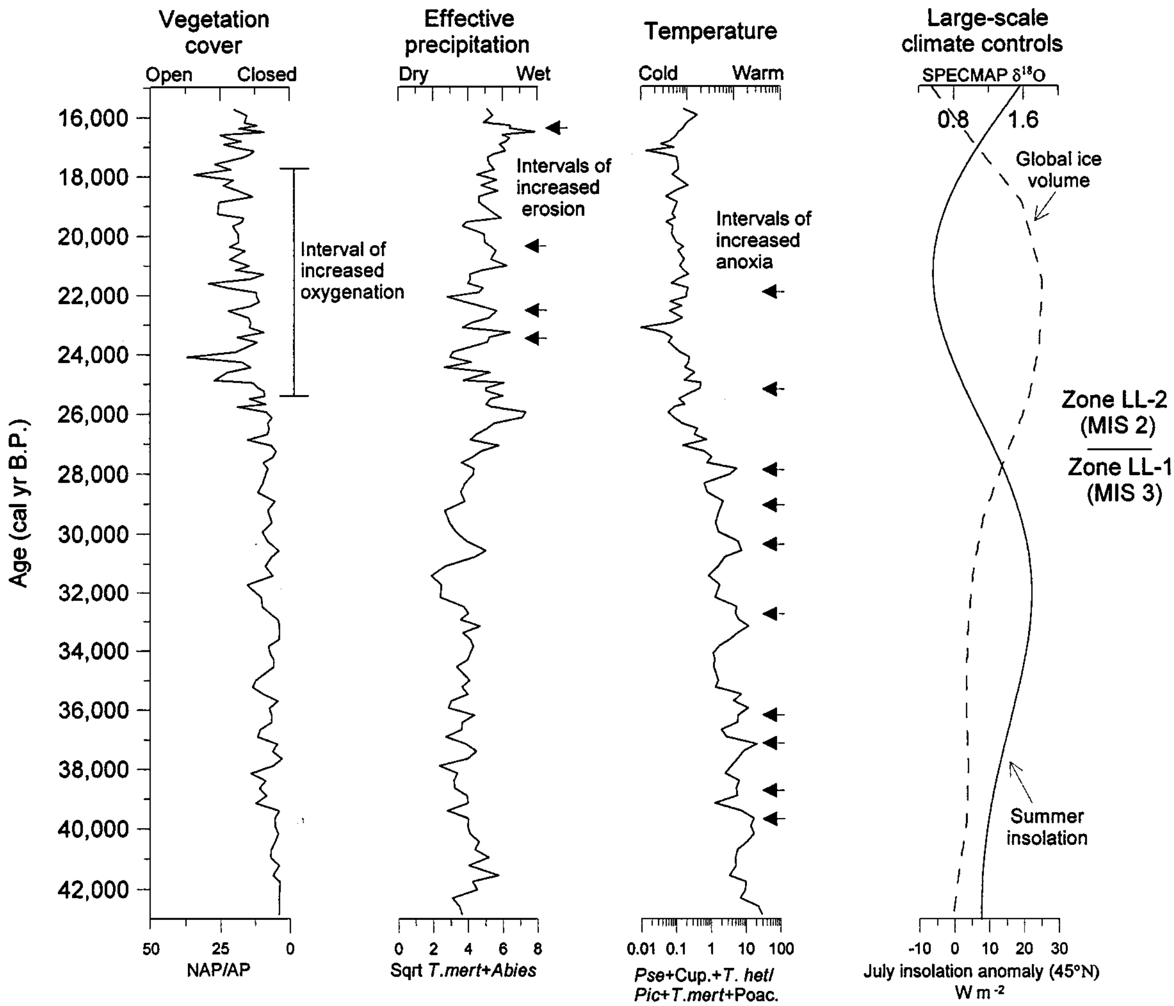

FIG. 7. Summary of changes in vegetation and climate at Little Lake during marine isotope stages 2 and 3 based on vegetation and climate indexes developed from the pollen data. Significant environmental changes inferred from the geochemical and magnetic susceptibility data are also shown. In the far right graph, variations in northern hemisphere summer insolation (Berger, 1978) are represented by a solid line. Global ice-volume, as inferred from the SPECMAP stacked and smoothed oxygen isotope data, is shown by the dashed line (Imbrie *et al.*, 1984).

shows five fluctuations between arboreal and nonarboreal pollen taxa between ca. 40,000 and 27,000 cal yr B.P. (Heusser, 1972). Increased percentages of *Picea* pollen at Carp Lake suggest cold, wet periods at ca. 26,000, 23,000, and 17,500 cal yr B.P., which are concurrent with the wet periods at Little Lake during Zone LL-2. Advances of the Cordilleran ice sheet in southwestern British Columbia also indicate cold, wet conditions between ca. 18,000 and 16,000 cal yr B.P. (Vashon Stade) and between ca. 24,500 and 22,000 cal yr B.P. (Coquitlam Stade; Hicock *et al.*, 1999). The ca. 22,000 cal yr B.P. warm period at Little Lake is registered at other sites in the Pacific Northwest between 23,000 and 22,000 cal yr B.P. (Hicock *et al.*, 1999; Mathewes, 1991; Whitlock and Grigg, 1999). These similarities in the direction and timing of vegetation changes throughout the Pacific Northwest imply that they represent a response to regional variations in climate.

Influence of Orbital-Scale Climate Controls

The generally moderate climate of the Pacific Northwest during MIS 3 (Alley, 1979; Barnosky, 1981, 1985; Heusser *et al.*, 1999; Warner *et al.*, 1984; Whitlock and Bartlein, 1997) corresponds with intermediate levels of summer insolation and global ice-volume. These orbital-scale controls fluctuated during MIS 3, but the highs and lows were not as dramatic as those of MIS 2 or the Holocene (Fig. 7). A comparison of MIS 3 climate changes at Little Lake with summer insolation and ice-volume suggests that variability in the magnitude of climate change partially reflects orbital-scale variations (Fig. 7). Two of the warmest periods of Zone LL-1 (at 33,000 and 30,500 cal yr B.P.) occurred during a period (33,000 to 30,500 cal yr B.P.), of high summer insolation and moderate ice volumes whereas the coldest intervals occurred before and after this time. Between 38,000 and 35,000 cal yr B.P.

a series of moderate shifts in climate imply a generally cool and seasonably equable climate, which coincides with relatively low summer and moderate winter insolation. However, some of the warmest conditions occurred (40,000 cal yr B.P.) when summer insolation was low and ice-volumes were moderate, suggesting that variations in millennial-scale climate controls, independent of orbital-scale forcings, also influenced the degree of climate change at Little Lake during MIS 3.

During MIS 2 (27,600–14,100 cal yr B.P.), the presence of subalpine forest and parkland in many low-elevation areas of the Pacific Northwest indicates colder, drier conditions than at present (Alley, 1979; Barnosky, 1981, 1985; Heusser *et al.*, 1999; Worona and Whitlock, 1995). Paleoclimate model simulations show that such conditions were related to the size of the Laurentide ice-sheet (Bartlein *et al.*, 1998; Thompson *et al.*, 1993) with the climate of the Pacific Northwest being affected by (1) the large extent of the ice-sheet causing a decrease in northern hemisphere temperatures; (2) the height of the ice-sheet displacing the jet stream to the south of its present position and; (3) the presence of a large ice-sheet resulting in a midcontinental, glacial anticyclone that increased easterly flow across western North America. The coldest and driest conditions occurred at Little Lake between ca. 21,000 and 17,000 cal yr B.P., which corresponds to a low in summer insolation and a maximum in global ice-volume.

Causes of Millennial-Scale Climate Change at Little Lake

A comparison of the Little Lake data with a record of sea-surface temperatures and sediment bioturbation from the Santa Barbara Basin (Behl and Kennett, 1996; Hendy and Kennett, 1999) provides an opportunity to evaluate current explanations of millennial-scale climate change in western North America and the Northeast Pacific. The pollen-based temperature proxy from Little Lake, when compared with the bioturbation index from the Santa Barbara Basin, shows that low values coincide with warmer sea-surface temperatures (Fig. 8). The Santa Barbara record, like that from Little Lake, has the greatest frequency and magnitude of millennial-scale climate change during the latter part of MIS 3. Warm intervals are concurrent at both sites with the exception of the 38,500 cal yr B.P. warm period at Little Lake. This event is registered by an increase in *T. heterophylla* pollen and not by the warmest-adapted taxa (*Pseudotsuga*-type or Cupressaceae). Consequently, it may reflect a modest climate warming or one of local extent.

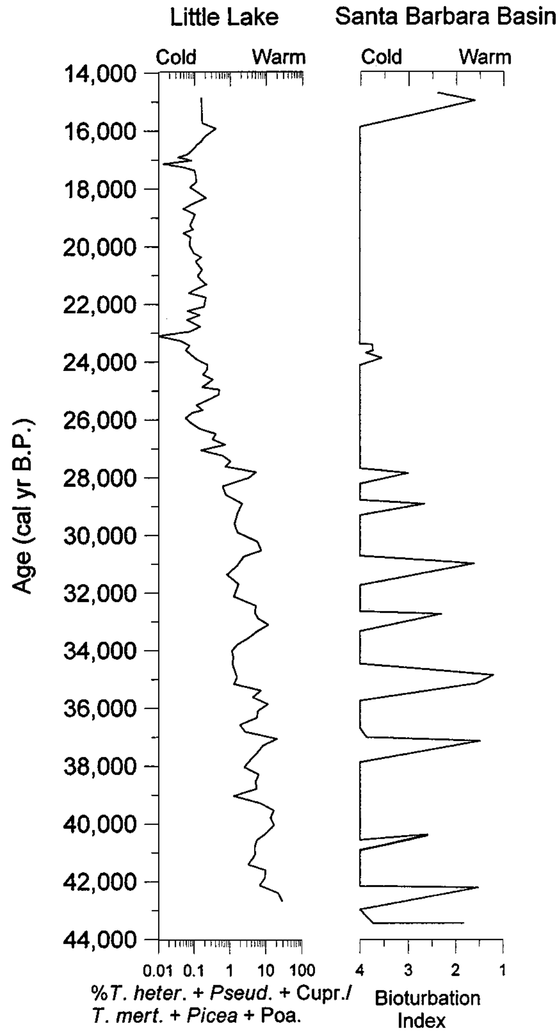

FIG. 8. A comparison of the pollen-based temperature ratio from Little Lake with the bioturbation index from Santa Barbara Basin (Behl and Kennett, 1996). A bioturbation index of four indicates periods of laminated sediments. Low values of the bioturbation index reflect intervals that correlate with warmer sea-surface temperatures (Hendy and Kennett, 1999).

Warmer sea-surface temperatures in the Santa Barbara Basin during MIS 3 have been attributed to northward shifts in the position of the subtropical high, the Aleutian low, and the jet stream (Hendy and Kennett, 1999). A significant northward shift of North Pacific atmospheric circulation patterns would have led to warmer year-round conditions in the Pacific Northwest and to a change from relatively dry to wet winters. Although this explanation is consistent with evidence for warming at Little Lake during the latter part of MIS 3, the lack of a strong precipitation signal implies little change in the position of the jet stream. In addition, strong upwelling along the central California margin suggests that the jet stream and subtropical high were close to their present positions (Gardner *et al.*, 1997). An alternative explanation invokes variations in the *strength* of the subtropical high and the jet stream, without significant shifts in their *position*. A stronger subtropical high would explain warmer conditions in both the Santa Barbara Basin and the Pacific Northwest, while an increase in the strength of the jet stream would have caused a shift in the Pacific Northwest from wet to wetter winter conditions. This more subtle change may explain the evidence of only small increases in precipitation during warm intervals at Little Lake.

The correlation of MIS 3 sea-surface temperatures in the Santa Barbara Basin with changes in temperatures and surges in the Laurentide ice-sheet implies that variations in the strength of the subtropical high and jet stream may have been related to fluctuations in North Atlantic atmospheric and oceanic circulation and ice-sheet size. Other correlations do exist between western North America, Northeast Pacific and North Atlantic paleorecords (Benson, 1999; Hicock *et al.*, 1999; Lin *et al.*, 1998; Lund and Mix, 1998; Whitlock and Grigg, 1999). However, model simulations for western North America and the Northeast Pacific during MIS 2 and the late-glacial period show only small variations in temperature in response to changes in the height of the ice sheet and/or temperatures over the North Atlantic (Hostetler and Bartlein, 1999; Mikolajewicz *et al.*, 1997). Shifts in Pacific sea-surface temperatures that were concurrent with North Atlantic changes may explain the degree of warming evident in the geologic record (Hostetler and Bartlein, 1999; Peteet *et al.*, 1997).

Fluctuations between cold-wet and cold-dry conditions at Little Lake are not apparent in the Santa Barbara Basin during MIS 2 (Fig. 8). Changes in precipitation in the Pacific Northwest may be explained by variations in the strength of the glacial anticyclone and the latitude of the jet stream. A recent model simulation, which describes the potential responses of western North America to large surges in the Laurentide ice-sheet during Heinrich events, shows that a reduction in the height of the ice sheet results in a weakened glacial anticyclone and a less prominent split in the jet stream (Hostetler and Bartlein, 1999). As a result, westerly flow and precipitation increases in the Pacific Northwest, while southern California shows little change in climate. These results are supported by the correlation of cold, wet intervals in the Pacific Northwest at 23,500 and 17,000 cal yr B.P. with Heinrich events 1 and 2 (Bond and Lotti, 1995; Hicock *et al.*, 1999; Whitlock and Grigg, 1999). In addition, cold wet intervals at 26,000 and 21,000 cal yr B.P. at Little Lake correspond to smaller and less established peaks in ice-rafted debris from the North Atlantic.

The warming at Little Lake at 22,500 cal yr B.P. is associated with widespread evidence in the Pacific Northwest for moderate warming between 23,000 and 22,000 cal yr B.P. (Mathewes, 1991; Whitlock and Grigg, 1999) and may correspond with a small increase in sea-surface temperatures in the Santa Barbara record at 23,500 cal yr B.P. However, an additional warm period at Little Lake at 25,000 cal yr B.P. is not matched in the Santa Barbara Basin record. Warming along the west coast of North America during MIS 2 may have resulted from a brief expansion of the subtropical high. Small changes in the subtropical high have been attributed to variations in Laurentide ice-sheet size and in North Atlantic atmospheric and oceanic circulation (Clark and Bartlein, 1995; Hendy and Kennett, 1999). However, model simulations of the last glacial maximum do not show warmer conditions along the coast in response to a reduction in ice-sheet height or an increase in sea-surface temperatures in the North Atlantic (Hostetler and Bartlein, 1999). Additional factors, such as an increase in sea-surface temperatures in the North Pacific, may have contributed to warming in western North America and the Northeast Pacific during this time (Hostetler and Bartlein, 1999; Peteet *et al.*, 1997).

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