

# Vegetation, fire, and climate history of the northwestern Great Basin during the last 14,000 years

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## Abstract

The northwestern Great Basin lies in the transition zone between the mesic Pacific Northwest and xeric intermountain West. The paleoenvironmental history based on pollen, macroscopic charcoal, and plant macrofossils from three sites in the northwestern Great Basin was examined to understand the relationships among the modern vegetation, fire disturbance and climate. The vegetation history suggests that steppe and open forest communities were present at high elevations from ca 11,000 to 7000 cal yr BP, and were replaced by forests composed of white fir, western white pine, and whitebark pine in the late Holocene. Over the last 11,000 years, fires were more frequent in mid-elevation forests (10–25 fire episodes/1000 years) and rare in high-elevation forests (2–5 fire episodes/1000 years). Applying modern pollen–climate relationships to the fossil pollen spectra provided a means to interpret past climate changes in this region. In the past 9000 years summer temperatures decreased from 1 to 4 °C, and annual precipitation has increased 7–15%. These results indicate that the millennial-scale climate forcing driving vegetation changes can be quantified within the intermountain West in general and northwestern Great Basin in particular. In addition, fire can be considered an important component of these ecosystems, but it does not appear to be a forcing mechanism for vegetation change at the resolution of these records.

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## 1. Introduction

The semi-arid Great Basin of the intermountain West is bordered by the Sierra Nevada and southern Cascade Range on the west, the Rocky Mountains on the east, and the Columbia River and Colorado River drainages on the north and south (Fig. 1A). The region is characterized by north–south trending mountain ranges separated by large alluviated basins, many of which contained pluvial lakes during the last glacial period (Morrison, 1991). These physiographic characteristics make the Great Basin one of the more diverse landscapes in North America in terms of topographic and biogeographic complexity.

Present-day climate in the Great Basin is controlled by seasonal variations in the configuration of the jet stream

and the strength and position of the Aleutian Low and eastern North Pacific subtropical high-pressure systems (Hirschboeck, 1991; Minckley et al., 2004). Winter precipitation dominates the region, with most of the precipitation occurring in the mountains and the valleys experiencing drier, rain shadow-like conditions (Mock, 1996). Summer precipitation is associated with airflow from the southeastern North Pacific coincident with the development of the North American monsoon (Houghton, 1969; Adams and Comrey, 1997).

The diversity of Great Basin plant communities results from steep climate gradients associated with latitude, longitude, and elevation. Regional patterns of effective moisture and available energy position desert-steppe, conifer forest, and alpine tundra conditions in close proximity (Fig. 1B) (Merriam, 1894; Billings, 1951). The west-to-east precipitation gradient creates a floristically unique region that combines mesic species of the Sierra Nevada and Cascade Range forests, with xeric species from the western interior mountains and Rocky Mountain

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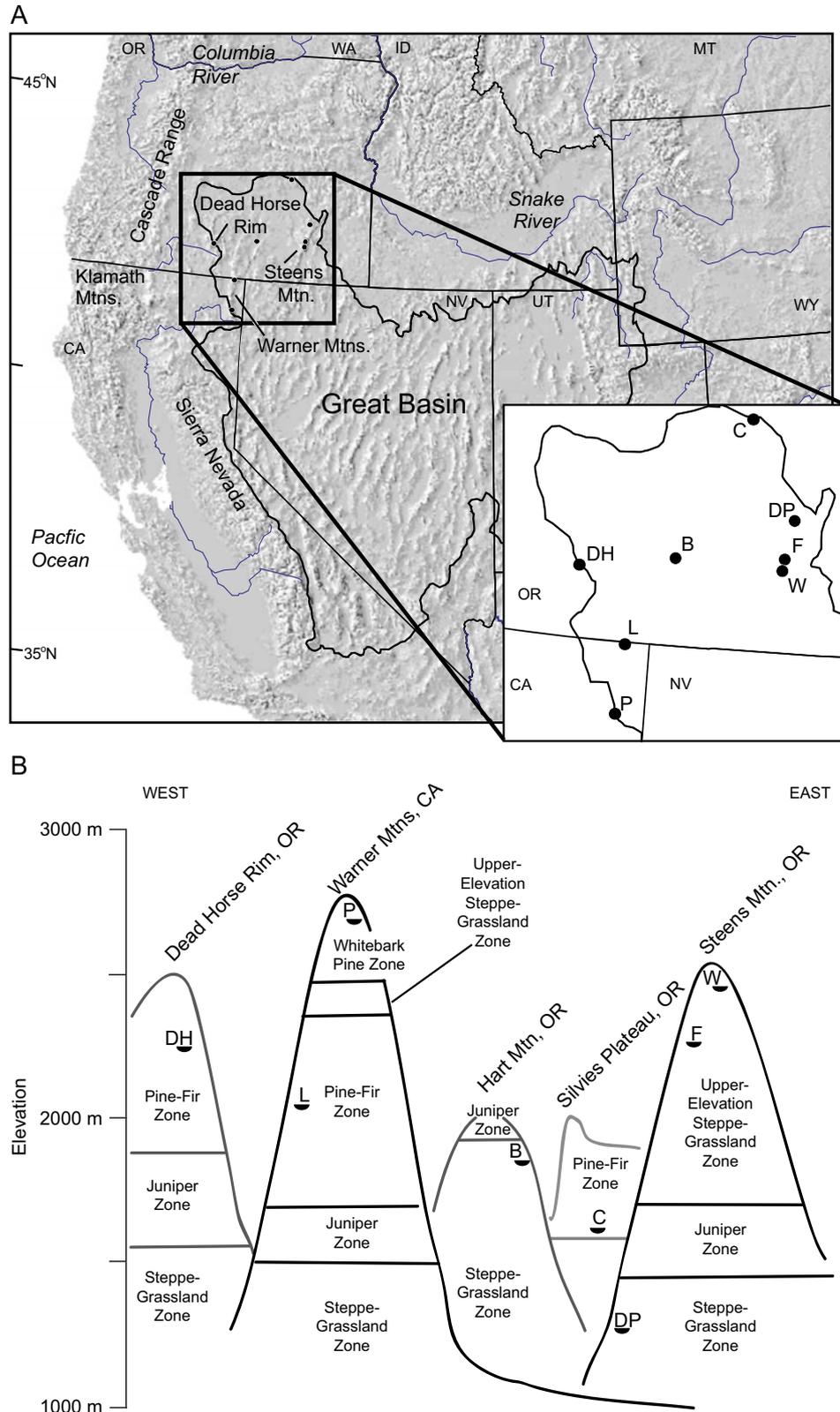


Fig. 1. (A) Map showing the location of the northwestern Great Basin and other physiographic features discussed in text (Morrison, 1991). Inset shows the location of sites in the northwestern Great Basin including (P) Patterson Lake, (L) Lily Lake, (DH) Dead Horse Lake, (B) Bicycle Pond (Wigand and Rhode, 2002), (C) Craddock Meadow (Wigand, 1989), (DP) Diamond Pond (Wigand, 1987), (F) Fish Lake (Mehringner, 1985), and (W) Wildhorse Lake (Mehringner, 1985). (B) A schematic transect of physiographic features and present-day vegetation zonation of the northwestern Great Basin based on Billings (1951) and Franklin and Dynnes (1988). Shown are the elevational positions of the sites discussed in this paper.

forests whose distribution is governed by slope, aspect, and elevation in the Great Basin (Critchfield and Allenbaugh, 1969; Harper et al., 1978; Charlet, 2007).

In this paper, pollen and high-resolution charcoal data from three sites were examined to reconstruct the vegetation, fire, and climate history of the northwestern Great Basin. Reconstructions of past climate variations were based on modern pollen–climate relations for western North America (Minckley, 2003; Whitmore et al., 2005; Williams et al., 2006). The records presented here were compared with published data to help disclose the nature of regional environmental changes during the last 14,000 cal yr BP (e.g., Mehringer, 1985; Whitlock, 1992; Thompson et al., 1993; Davis, 1999; Mensing, 2001; Wigand and Rhode, 2002).

### 1.1. Study sites

Patterson Lake and Lily Lake in the Warner Mountains of northeastern California, and Dead Horse Lake in south-central Oregon lie in different environmental settings and elevations (Fig. 1). Patterson Lake (41.3867°N, 120.2236°W, 2743 m elevation, 12 ha) is located within a late-Pleistocene cirque with no inflowing stream and an ephemeral outflowing stream. Annual precipitation is 415 mm/yr, most of which is received between November and March. Annual temperatures average 3.5 °C ranging from –47 to 36 °C. The lake lies within the whitebark pine (*Pinus albicaulis*) zone of the Sierran vegetation series (Billings, 1951). The rim and western flank of the cirque support open whitebark pine forest with gooseberry/currant (*Ribes* sp.) and rock spirea (*Holodiscus microphyllus*) in openings and understory. The north and east slopes of the basin are covered by an open meadow with Sierra willow (*Salix eastwoodiae*), and alpine herbs, sedges (*Carex* spp.), showy silkweed (*Frasera speciosa*), American bistort (*Polygonum bistortoides*), valerian (*Valeriana californica*), and grasses (Hickman, 1993).

Lily Lake (41.9758°N, 120.2097°W, 2042 m elevation, 3.7 ha) was dammed by a landslide in the northern Warner Mountains. The lake has an inflowing, but no outflowing stream. Annual precipitation is 490 mm/yr, most of which is received between November and March. Annual temperatures average 6.3 °C, and range from –39 to 39 °C. The lake lies in the pine-fir zone of the Sierran vegetation series (Billings, 1951), and forests around Lily Lake are composed of white fir (*Abies concolor* var. *lowiana*), ponderosa pine (*Pinus ponderosa*), lodgepole pine (*Pinus contorta*), western white pine (*Pinus monticola*), incense cedar (*Calocedrus decurrens*), and western juniper (*Juniperus occidentalis*). Also present in the watershed are silver wormwood (*Artemisia ludoviciana*), serviceberry (*Amelanchier pallida*), wood rose (*Rosa woodsii*), rock spirea, and gooseberry.

Dead Horse Lake (42.5606°N, 120.7781°W, 2248 m elevation, 15 ha) is located behind a late-Pleistocene moraine on Dead Horse Rim, one of the western-most

fault block structures of the northern Great Basin. The lake has no inflowing or outflowing stream. Annual precipitation is 650 mm/yr, most of which is received between November and March. Annual temperatures average 3.6 °C and range between –42 and 37 °C. The lake is in the lodgepole pine-mountain hemlock (*Tsuga mertensiana*) zone of the Sierran vegetation series (Billings, 1951) (lodgepole pine zone in Franklin and Dyrness, 1988), and is surrounded by lodgepole pine, western white pine, and whitebark pine with little understory vegetation (Hopkins, 1979). Mountain mahogany (*Cercocarpus ledifolius*), big sagebrush (*Artemisia tridentata*), rock spirea, and gooseberry are present on rocky slopes near the southeastern lakeshore. The lodgepole pine zone is unique to this part of the Great Basin, where pumaceous soils from the eruption of Mount Mazama predominate (Franklin and Dyrness, 1988).

## 2. Methods

### 2.1. Field

Sediment cores were collected with a modified 5-cm diameter Livingstone piston corer (Wright et al., 1983). A 2.23-m-long core was recovered from Patterson Lake, a 9.95-m-long core from Lily Lake, and a 2.33-m-long core from Dead Horse Lake. Cores were described, wrapped in plastic and aluminum foil, and transported to the University of Oregon where they were refrigerated. Short-cores were collected using a polycarbonate plastic-tube with a piston, which recovered the mud–water interface and top ~30 cm of sediment from each lake. These cores were field sampled at 1-cm intervals.

### 2.2. Lithologic analyses

Loss-on-ignition (LOI) and magnetic susceptibility (MS) analyses provided information on the sedimentation history for each lake, particularly the relative contribution of inorganic detrital sediment (Thompson and Oldfield, 1986; Geyde et al., 2000). For LOI analysis, samples of 1 cm<sup>3</sup> were taken at 5-cm intervals, dried for 24 h at 90 °C and weighed. Sample weights were measured after combusting samples at 550 °C for 2 h to determine percent of organic content and at 900 °C for 2 h to determine percent of carbonate content (Dean, 1974).

Sediment MS (measured in electromagnetic units [emu]) was measured in contiguous 1-cm samples. At Patterson Lake and Lily Lake, 8 cm<sup>3</sup> of sediment was placed in plastic containers and measured in a Sapphire cup-coil magnetic-susceptibility instrument. At Dead Horse Lake, MS was measured prior to LOI analysis on entire core segments by taking measurements at 1-cm intervals with a Sapphire ring-coil magnetic-susceptibility instrument (Oldfield et al., 1979). MS data were plotted on a log-scale to better compare results using these two methods because of the relative high MS values obtained from Dead Horse Lake.

### 2.3. Charcoal analysis

Charcoal analysis was used to reconstruct past fire activity within each watershed following methods of Long et al. (1998) and Whitlock and Larsen (2002). Samples of 5 cm<sup>3</sup> from the Patterson and Lily lake cores and 4 cm<sup>3</sup> from the Dead Horse Lake core were taken at contiguous 1-cm intervals. Samples were washed through metal sieves with mesh sizes of  $\geq 125 \mu\text{m}$ . Other studies suggest that such large particles are not transported great distances from their source and consequently represent local fire events (Whitlock and Millspaugh, 1996; Clark et al., 1998; Gardner and Whitlock, 2001).

Charcoal particles  $> 125 \mu\text{m}$  in minimum diameter were counted at 25–50 $\times$  magnifications. Charcoal concentration (particles/cm<sup>3</sup>) was calculated for each sample, and interpolated into pseudo-annual values that were then averaged over 10-year intervals. These values were divided by deposition times (yr/cm) (calculated from the age–depth models for each site) to obtain time series of charcoal accumulation rates (CHAR) (particles cm<sup>2</sup>/yr). CHAR records were broken into two components: (1) a slowly varying “background” component representing long-term changes in charcoal production and transport (either primary or secondary) into the lake sediments, and (2) a rapidly varying “peaks” component representing short-term changes of CHAR as well as large peaks caused by “fire episodes” (i.e., one or more fire events during the time span of the sample) (Long et al., 1998; Whitlock and Larsen, 2002). Fire episodes were identified as periods when the charcoal peaks component exceeded the background component by an iteratively determined threshold. A locally weighted mean with a 900-year smoothing window was used to define background at all three sites. CHAR threshold-ratio values were set at 1.05 for Patterson Lake and 1.15 for Lily Lake. At Dead Horse Lake, the CHAR threshold-ratio value was set at 1.00 (i.e., all peaks above the background level were identified as fire episodes). These values were selected as the lowest thresholds that did not identify fires in the past 100 years at Patterson and Lily lakes, which is consistent with the known fire history. No information was available for Dead Horse Lake so threshold value of unity was used, which could lead to an inflated estimate of the number of peaks throughout the record. Higher sedimentation rates at Lily Lake relative to Patterson and Dead Horse lakes resulted in a high-resolution CHAR record and the detection of more fire events. At all sites, a locally weighted mean fire frequency (fire episodes/1000 years) was obtained with a 1000-year weighted averaging window.

### 2.4. Plant macrofossil analysis

Plant macrofossils provided species identifications and confirmed local occurrences of taxa. Plant remains found in the  $> 125 \mu\text{m}$  sieved residues were identified by comparisons with reference material and published keys (e.g., Fassett,

1975; Hickman, 1993). Macrofossils of ponderosa pine (*P. ponderosa*), Jeffrey pine (*Pinus jeffreyi*), and Washoe pine (*Pinus washoensis*) are indistinguishable from each other and were grouped as *P. cf. ponderosa*. Macrofossil occurrence was plotted on the pollen diagrams.

### 2.5. Pollen analysis

Pollen analysis was undertaken at 3-cm intervals at Patterson Lake, 10-cm intervals for Lily Lake, and 8-cm intervals at Dead Horse Lake to reconstruct the vegetation history. Methods followed Faegri et al. (1989). At least 300 terrestrial pollen grains were identified for each sample at magnifications of 500–1250 $\times$ . Pollen counts were converted to percentages based on a denominator of total terrestrial pollen.

Pollen grains were identified at the lowest taxonomic level possible through comparisons with reference collections at the University of Oregon and published atlases (e.g., Erdtman, 1969; Bassett et al., 1978; Moore et al., 1991; Kapp et al., 2000). *Pinus* pollen was identified as *Pinus* subgen. *Pinus* or *P. subgen. Strobus* if the distal membrane remained intact, otherwise, *Pinus* was counted as undifferentiated. Possible *P. subgen. Strobus* species include western white pine and whitebark pine. *P. subgen. Pinus* pollen could come have from lodgepole pine, ponderosa pine, Jeffrey pine, and Washoe pine. *Abies* pollen was attributed to white fir, based on modern phytogeography and plant macrofossils. Cupressaceae pollen was assigned to western juniper, but incense cedar (*C. decurrens*) and common juniper (*Juniperus communis*) were also possibilities. *Cercocarpus*-type pollen was attributed to mountain mahogany, which grows today near each lake, but also could have come from antelope bush (*Purshia tridentata*) (Anderson and Davis, 1988). Unknown and unidentifiable pollen grains were included in the terrestrial pollen sum. Stratigraphic zonation was determined by constrained cluster analysis (CONISS) (Grimm, 1988).

#### 2.5.1. Climate reconstruction

A compilation of 2013 surface and core-top pollen spectra from western North America were assigned modern climate values (Minckley, 2003; Whitmore et al., 2005; Williams et al., 2006). These modern climate values were attributed to fossil pollen spectra using the modern analog approach (Prentice, 1980; Overpeck et al., 1985; Bartlein and Whitlock, 1993). Climate reconstructions were based on a weighted average of modern climate values from the location of the 10 closest pollen analogs for each fossil spectra. Results were presented as anomalies from present-day climate.

## 3. Results

### 3.1. Lithology and chronology

The sediments at Patterson Lake consisted of basal sands with a gravel unit (2.18–1.96 m depth) (Fig. 2) overlain by

clayey silt gyttja. One-cm thick tephra layers at 1.27 and 1.22 m depth were identified as the Llaio and Mazama ashes (A.M. Sarna-Wojcicki, pers. commun. 2001). Organic-carbon content was low (4–8%) suggesting continual terrestrial material input into the lake. MS was relatively high ( $10^{-4}$  emu) between 2.18 and 1.80 m depth (Fig. 2), but declined by an order of magnitude around 1.80 m depth and remained low to the top of the core. A 3rd-order polynomial regression was used to develop an age–depth model at Patterson Lake based on six  $^{210}\text{Pb}$  ages, eight  $^{14}\text{C}$  AMS ages converted to calendar years using Stuiver et al. (1998), and ages of the volcanic ashes (Table 1, Fig. 3).

Lily Lake cores consisted of basal inorganic clay (9.95 and 8.91 m depth) overlain by coarse detritus gyttja and interbedded clays from 8.91 to 7.74 m depth. Between 7.74 and 7.29 m depth was a thick tephra unit, assumed to be Mazama ash. A massive inorganic clay unit was present between 5.86 and 4.80 m depth. Organic carbon content ranged between 4% and 6% below 8.91 m depth and between 4% and 17% from 8.90 m depth to the top of the core. MS values were variable and relatively high ( $10^{-5}$ – $10^{-4}$  emu) between the depth of Mazama ash (9.95 and 9.00 m depth), and at the mid-core clay unit (5.86–4.80 m depth) (Fig. 2).

The chronology of Lily Lake was based on a 3rd-order polynomial regression developed from six  $^{14}\text{C}$  ages

converted to calendar years using Stuiver et al. (1998), and the age of Mazama ash (Table 1). Four additional  $^{14}\text{C}$  ages in and above the clay unit were out of stratigraphic sequence, approximately contemporaneous, and considered evidence of sediment mixing and rapid deposition (Table 1, Fig. 3). The clay unit was likely the distal facies of a mass movement into Lily Lake. The lithologic data for Lily Lake indicate that the lake formed at ca 10,000 cal yr BP and, prior to ca 8800 cal yr BP, sediment input was rapid and lake productivity was low. The clay unit between 5.86 and 4.80 m depth and the Mazama ash were subtracted from the core length prior to calculating age–depth relationships.

At Dead Horse Lake, a basal sand unit was overlain by clay between 2.33 and 2.02 m depth (Fig. 2). Above 2.02 m depth, sediments were fine-detritus gyttja. A tephra between 1.40 and 1.13 m depth was attributed to the Mazama ash. Organic content was relatively low (6–8%) from 2.33 m depth to the top of the core. MS in the Dead Horse Lake core was relatively high ( $10^{-3}$  emu) between 2.33 and 2.16 m depth. MS values declined two orders of magnitude to  $10^{-5}$ – $10^{-4}$  emu between 2.16 and 1.40 m depth and remained low to the top of the core. MS measurements for Dead Horse Lake were likely higher than at Patterson and Lily lakes because whole-core measurements were used, which integrate MS across a larger core segment.

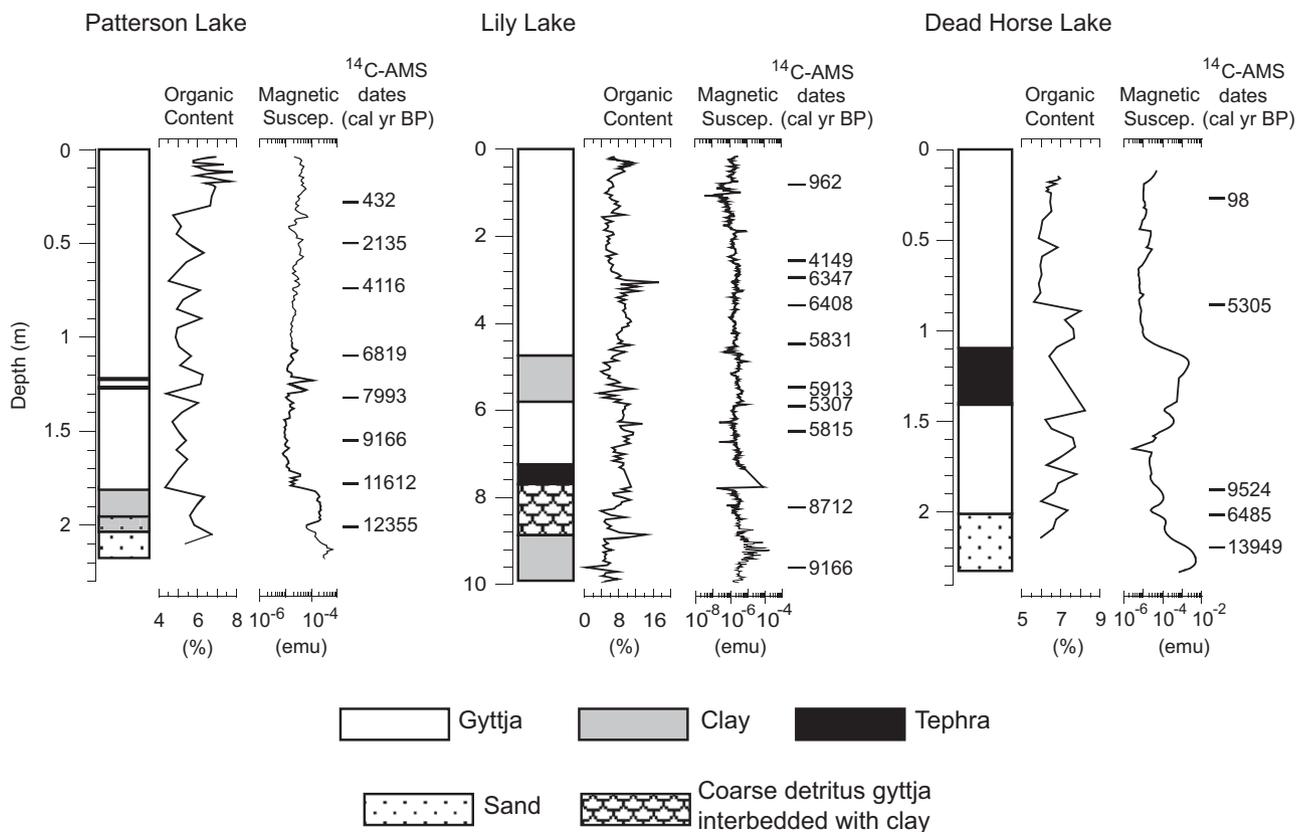


Fig. 2. Generalized lithology, magnetic susceptibility, and organic content of the Patterson Lake, Lily Lake, and Dead Horse Lake cores. Also shown are the locations of  $^{14}\text{C}$ -AMS dates converted to calendar years.

Table 1  
Information on age–depth relations at the fossil study sites<sup>a</sup>

Sample depth below mud surface (cm)	Adjusted sample depth (cm) <sup>b</sup>	Dates ( <sup>210</sup> Pb, <sup>14</sup> C, volcanic tephra)	Calibrated age (cal yr BP) <sup>c</sup>	Material dated	Laboratory number
<i>Patterson Lake, Modoc Nat. Forest, Warner Mtns., CA: Core 2KC</i>					
4–5	4–5	37.04	–13	Lake sediment <sup>d</sup>	...
5–6	5–6	49.61	0	Lake sediment <sup>d</sup>	...
6–7	6–7	64.11	15	Lake sediment <sup>d</sup>	...
7–8	7–8	91.65	42	Lake sediment <sup>d</sup>	...
8–9	8–9	141.75	92	Lake sediment <sup>d</sup>	...
9–10	9–10	192.34	142	Lake sediment <sup>d</sup>	...
27–28	27–28	400 ± 35	482	Conifer twig	CURL-5738
49–50	49–50	2160 ± 30	2135	Conifer twig	CURL-5739
73–74	73–74	3760 ± 35	4116	Conifer twig	CURL-5740
109–110	109–110	6030 ± 40	6819	Whitebark pine needle	CURL-5741
122–123	122	Mazama ash	7627 <sup>e</sup>	...	...
127–128	126	Llao tephra	7798 <sup>e</sup>	...	...
131–132	129–130	7190 ± 60	7993	Conifer twig	CURL-5742
154–155	152–153	8260 ± 45	9166	Conifer twig	CURL-5743
177–178	175–176	10080 ± 55	11612	Charcoal	CURL-5962
205–208	203–206	10380 ± 100	12355	Charcoal	AA-55857
Age–depth model: Age (cal yr BP) = $-0.0018 \times \text{depth}^3 + 0.5591 \times \text{depth}^2 + 20.091 \times \text{depth} - 50$				$R^2 = 0.9961$	
<i>Lily Lake, Modoc Nat. Forest, Warner Mtns., CA: Core 99B</i>					
80–81	80–81	990 ± 30	962	Lodgepole pine needle	CURL-5744
255–256	255–256	3781 ± 56	4149	Charcoal	AA-51462
292–293	292–293	5580 ± 80	6347 <sup>f</sup>	Charcoal	CURL-5960
357–358	357–358	5635 ± 96	6408 <sup>f</sup>	White fir needle	AA-51463
445–446	445–446	5038 ± 43	5831 <sup>f</sup>	Wood	AA-51464
547–548	460–461	5150 ± 50	5913 <sup>f</sup>	Conifer needles	CURL-5745
586–587	463–464	4580 ± 50	5307	Charcoal	AA-51466
628–629	505–506	5073 ± 61	5815	Wood	AA-51467
729–774	606	Mazama ash	7627 <sup>e</sup>	...	...
813–814	645–646	7920 ± 45	8712	White fir/white pine needle	CURL-5746
959–960	791–792	8260 ± 150	9166	Pine needle	AA-37485
Age–depth model: Age (cal yr BP) = $0.000005 \times \text{depth}^3 - 0.008 \times \text{depth}^2 + 15401 \times \text{depth} - 49$				$R^2 = 0.9656$	
<i>Dead Horse Lake, Fremont Nat. Forest, Dead Horse Rim, OR: Core 99B</i>					
26–27	26–27	108 ± 45	98	Charcoal	AA-53492
82–84	82–84	4593 ± 46	5305	Lake sediment	AA-55855
113–139	113	Mazama ash	7627 <sup>e</sup>	...	...
178–180	152–154	8566 ± 72	9524	Lake sediment	AA-55856
202–203	176–177	5690 ± 100 <sup>f</sup>	6485	Charcoal	AA-52494
219–221	194–195	11 910 ± 120	13949	Lake sediment	AA-52489
Age–depth: model Age (cal yr BP) = $-0.0009 \times \text{depth}^3 + 0.3685 \times \text{depth}^2 + 32.746 \times \text{depth} - 49$				$R^2 = 0.976$	

<sup>a</sup>Radiometric ages given as means ± 1 SD.

<sup>b</sup>Used in the age–depth determinations.

<sup>c</sup>Based on Stuiver et al. (1998).

<sup>d</sup><sup>210</sup>Pb data provided by James Budahn, US Geological Survey, Denver.

<sup>e</sup>Identified by A.M. Sarna-Wojcicki, US Geological Survey, Menlo Park. Age of Mazama ash from Zdanowicz et al., 1999.

<sup>f</sup>Denotes samples not used in the age depth calculations.

The chronology of Dead Horse Lake was based on the age of the Mazama ash, four <sup>14</sup>C ages converted to calendar years using Stuiver et al. (1998), and described by a 3rd-order polynomial regression (Table 1). An out-of-sequence age at 203 cm depth, presumably caused by down-core contamination, was not used in the age–depth model. The 0.27 m thickness of the tephra was subtracted from the core length to create an adjusted depth for calculating age–depth relationships (Table 1, Fig. 3). The lithologic

information implies high sediment input and low lake production prior to ca 11,600 cal yr BP. Autochthonous organic input increased although the lake remained relatively unproductive.

### 3.2. Fire history records

A dendrochronological fire history was not available for the study area, thus the fire-frequency reconstructions

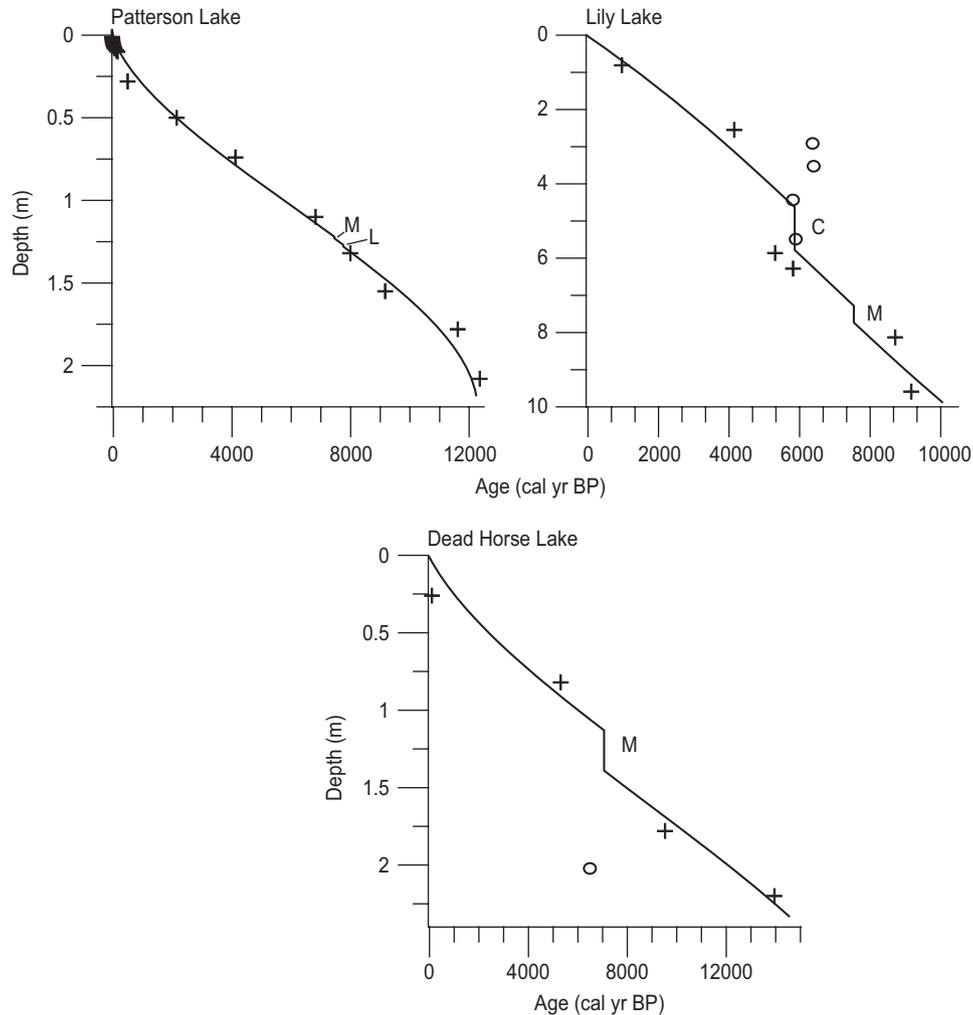


Fig. 3. Age–depth determinations for Patterson, Lily, and Dead Horse lakes. Also shown are the locations of the dates used in the age–depth calculations. Circles denote dates not used in age–depth calculations. Age–depth curve was calculated using adjusted depths (Table 1). Vertical drops represent the instantaneous deposition of (M) Mazama ash bed, (L) Llaol tephra bed (Bacon, 1983), and clay (see Fig. 2).

based on charcoal data could not be calibrated against independent records. In open forests, such as those at the study sites, fire regimes at present are characterized by moderate to infrequent (30–300 years), mixed-intensity fires and may not produce much charcoal (Agee, 1993). In addition, lakes like Patterson and Dead Horse, with slow sedimentation rates, may have incorporated charcoal from multiple fire events in each centimeter of core sediment, thereby reducing the resolution of the fire history reconstruction. Fires were not recorded in the watersheds of Patterson Lake and Lily Lake in the last 100 years (Sydney Smith, Modoc National Forest, US Forest Service, pers. commun., 2002), and no fire information was available for Dead Horse Lake, but fire intervals (time between fires) of 65–350 years has been suggested for other lodgepole pine forests in south-central Oregon (Agee, 1993). Selection of the threshold-ratio parameter, as mentioned previously, produced recent peaks-frequency values consistent with these known fire intervals.

At Patterson Lake, background CHAR was low before ca 11,150 cal yr BP (Fig. 4), and increased to 1.38 to

particles  $\text{cm}^2/\text{yr}$  at 9850 cal yr BP. Background CHAR decreased to 0.35 particles  $\text{cm}^2/\text{yr}$  at 8100 cal yr BP, increased at 6600 cal yr BP, decreased at 6000 cal yr BP, and then increased to present. Inferred fire-episode frequency was generally low at this site (1.5 and 6.3 episodes/1000 years). Fire-episode frequency was lowest between ca 6700 and 5600 cal yr BP and highest from 2200 to present (>6 episodes/1000 years).

At Lily Lake, background CHAR increased from 0.06 to 0.29 particles  $\text{cm}^2/\text{yr}$  between 10,000 and 9300 cal yr BP (Fig. 4). Background CHAR increased to 4.47 particles  $\text{cm}^2/\text{yr}$  at 5900 cal yr BP, decreased to 1.11 particles  $\text{cm}^2/\text{yr}$  from 5900 to 4600 cal yr BP, increased to 3.41 particles  $\text{cm}^2/\text{yr}$  at 4100 cal yr BP, and generally decreased to 0.5 particles  $\text{cm}^2/\text{yr}$  at present. Inferred fire-episode frequency showed large variations through the Lily Lake record, ranging between 7 and 27 episodes/1000 years. High fire activity (generally >17 episodes/1000 years) was evident from 8900 to 7400, 5700 to 4700, 1900 to 1000 cal yr BP. Long periods with no fire episodes were also evident from 9150 to 8750, 4760 to 4220, and 3040 to 2680 cal yr BP.

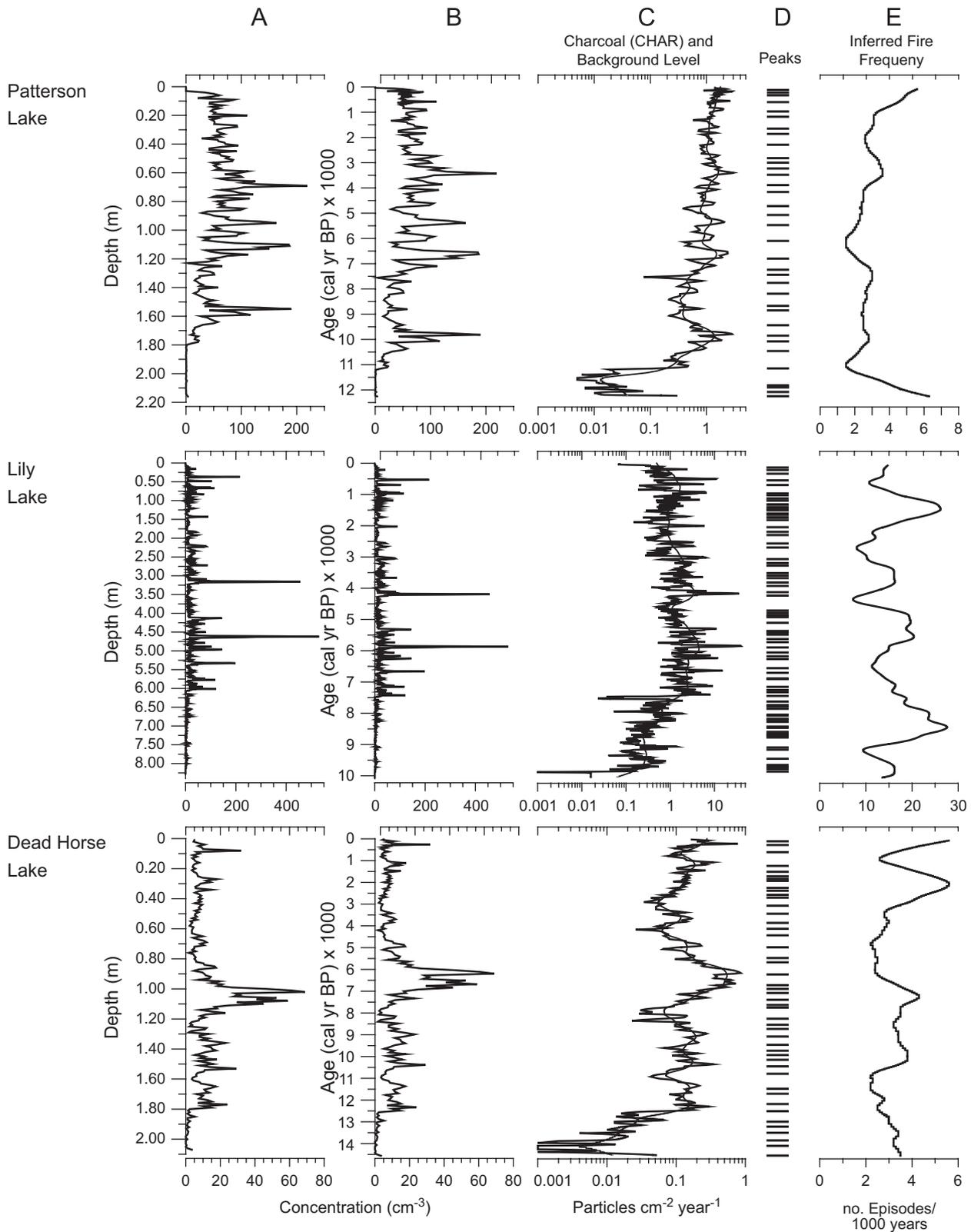


Fig. 4. Decomposition of the Patterson, Lily, and Dead Horse lakes charcoal data. (A) Concentration of raw charcoal counts plotted by adjusted depth. (B) Concentration of raw charcoal counts plotted by time (cal yr BP). (C) Decomposed accumulation rates (CHAR) calculated at 10-year time intervals, and background levels calculated with a weighted running mean of 900 years. By plotting these data on a log-scale both the low- and high-frequency variations of the data can be examined. (D) Peaks above background CHAR based on a 1.05, 1.15, and 1.00 threshold-ratio value, respectively, used to identify peaks above background. (E) Inferred fire-episode frequency based on a weighted 1000-year average.

At Dead Horse Lake, background CHAR was low ( $<0.10$  particles  $\text{cm}^2/\text{yr}$ ), between ca 14,000 and 13,500 cal yr BP (Fig. 4). Background CHAR increased to 0.16 particles  $\text{cm}^2/\text{yr}$  at 12,200 cal yr BP, and fluctuated between 0.07 and 0.19 particles  $\text{cm}^2/\text{yr}$  until 7200 cal yr BP. A peak of 0.55 particles  $\text{cm}^2/\text{yr}$  occurred at 6300 cal yr BP followed by a decline. In the last 3000 cal yr, background CHAR increased from 0.05 to 0.28 particles  $\text{cm}^2/\text{yr}$  at present. Between 14,000 and 10,900 cal yr BP, fire-episode frequency decreased from 3.5 to 2.2 episodes/1000 years. A long peaks-free interval occurred from ca 11,500 to 10,700 cal yr BP. Fire-episode frequency was relatively constant between 10,500 and 3000 cal yr BP with 2–4.5 episodes/1000 years. Fire frequency was high (5.6 episodes/1000 years) between 2100 and 2000 cal yr BP, decreased until ca 900 cal yr BP, and then increased to 5.9 episodes/1000 years to the present.

Similar trends in fire-episode frequency were evident on millennial time scales at all three sites. Patterson and Dead Horse lakes showed low fire-episode frequency prior to 11,000 cal yr BP. Between 10,000 and 7000 cal yr BP, fire-episode frequency was relatively constant at Patterson and Dead Horse lakes while Lily Lake had high fire-episode frequency centered on 8500 cal yr BP that decreased by 6000 cal yr BP. Between 6000 and 3500 cal yr BP, trends in fire-episode frequency were more variable, with records from Patterson and Dead Horse lakes showing increasing frequencies, while the Lily Lake record generally decreased. However, peaks in fire-episode frequency  $\sim$ 3500 cal yr BP occurred at all sites. From ca 3000 cal yr BP to present, fire-episode frequency increased at Patterson and Lily lakes with peaks in fire frequency ca 1500 cal yr BP. Dead Horse Lake had high fire-episode frequency at ca 2000 cal yr BP. Between 900 and 700 cal yr BP, Lily and Dead Horse lakes registered low fire-episode frequency not evident in the Patterson Lake record.

### 3.3. Pollen and macrofossil records

#### 3.3.1. Patterson Lake

The pollen record from Patterson Lake was divided into four zones (Fig. 5): Zone P-1 (2.18–1.80 m depth; 12,200–11,350 cal yr BP) was characterized by relatively high percentages of *Pinus* ( $\bar{x} = 70\%$ ). *Pinus* subgen. *Pinus* ( $\bar{x} = 10\%$ ) dominated the differentiated *Pinus*. *Artemisia* percentages were moderate ( $\bar{x} = 17\%$ ), as were Cupressaceae ( $\bar{x} = 2\%$ ), Poaceae ( $\bar{x} = 5\%$ ), Chenopodiaceae ( $\bar{x} = 3\%$ ), and *Sarcobatus* ( $\bar{x} = 2\%$ ). *Abies* and *Cercocarpus* pollen were rare. Modern analogs from western North America for zone P-1 are found east and north of Patterson Lake in the Ruby Mountains, Nevada (Thompson, 1984), and Bighorn Mountains, Wyoming (Burkart, 1976) above 2200 m elevation, suggesting subalpine forest and grassland surrounded Patterson Lake prior to 11,500 cal yr BP.

Zone P-2 (1.80–1.07 m depth, ca 11,350–6250 cal yr BP) featured relatively low *Pinus* percentages ( $\bar{x} = 64\%$ ) and high *Artemisia* percentages ( $\bar{x} = 25\%$ ). *Pinus* subgen. *Strobilus*, *Abies*, and Rosaceae percentages increased during

this period. *Cercocarpus* increased in this zone and reached its maximum percentage (4%). Poaceae percentages were relatively high ( $\bar{x} = 5\%$ ), but decreased in this zone. Whitebark pine was locally present at the lake ca 8550 cal yr BP, evidenced by a needle fragment. Modern pollen analogs for zone P-2 are found east and north in the Ruby Mountains, Nevada (Thompson, 1984) and Chuska Mountains, New Mexico (Bent and Wright, 1963) in locations above 2000 m elevation, and in the Columbia Basin, Washington (Mack and Bryant, 1974) suggesting steppe, grassland and western pine forest.

Zone P-3 (1.07–0.85 m depth; ca 6250–4600 cal yr BP) was characterized by increasing average *Pinus* percentages (70–80%). Percentages of *Pinus* subgen. *Strobilus*, probably from whitebark pine, was relatively stable ( $\bar{x} = 7\%$ ). *Abies* percentages increased to an average of 4% in this zone, and percentages of *Cercocarpus* were  $\sim$ 2%. *Artemisia* percentages are moderate ( $\bar{x} = 16\%$ ). Modern pollen analogs for zone P-3 matched modern pollen spectra from Patterson Lake; the Ruby Mountains, Nevada (Thompson, 1984); Yellowstone region (Whitlock, 1993; Millspaugh et al., 2000); and Chuska Mountains, New Mexico (Bent and Wright, 1963). All modern analogs occur above 2000 m elevation in pine and subalpine forests.

Zone P-4 (0.85–0 m depth, 4600 cal yr BP to present) had high average *Pinus* (75%) and *Abies* (5%), low average *Artemisia* (12%), and increasing Poaceae ( $\bar{x} = 5\%$ ). Whitebark pine needles were abundant in the top 0.32 m of the sediment core. Modern analogs for this zone include Patterson Lake; high-elevation ( $>2000$  m elevation) localities in the central and northern Rocky Mountains (Whitlock, 1993; Lynch, 1998; Millspaugh et al., 2000); and Chuska Mountains, New Mexico (Bent and Wright, 1963).

To summarize, prior to 11,350 cal yr BP, the environment at Patterson Lake resembled present-day upper-elevation steppe and grasslands. *Pinus*, Cupressaceae, and *Sarcobatus* were likely present at lower elevations and their pollen was transported upslope, similar to modern pollen rain in the mountains of southeastern Oregon (Minckley and Whitlock, 2000). Cold, windy conditions and sparse vegetation cover are suggested by the lithology, low fire activity, extralocal pollen, and modern analogs that indicate subalpine environments. Increased *Artemisia* and Poaceae percentages from 11,350 to 6250 cal yr BP, along with higher CHAR suggest increased shrub cover prior to early forest development. Whitebark pine forests formed ca 8550 cal yr BP during an interval of moderate fire activity. Increased forest cover, suggested by higher *Pinus* subgen. *Strobilus* percentages after ca 6250 cal yr BP, coincided with low fire-episode frequency. Fire-episode frequency has increased to present, but the forest composition has remained constant since 4300 cal yr BP.

#### 3.3.2. Lily Lake

The pollen record from Lily Lake was divided into three zones (Fig. 6). The closest modern analogs for fossil pollen assemblages at Lily Lake often includes the site itself,

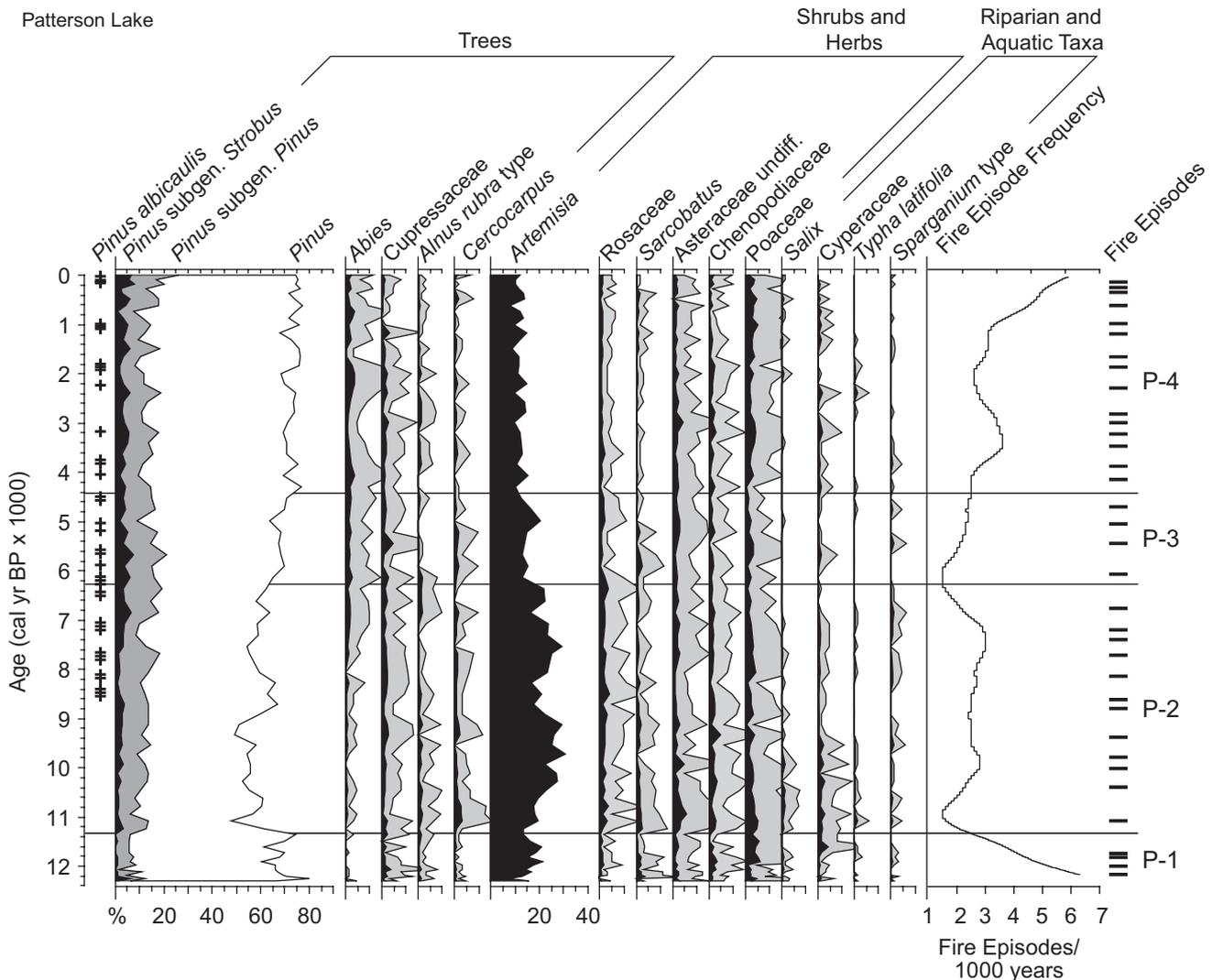


Fig. 5. Pollen percentage diagram for Patterson Lake showing macrofossil data, selected pollen taxa, and fire-episode frequency, and fire episodes (see Fig. 5). Light gray shading indicates  $5 \times$  exaggeration of pollen percentages.

suggesting this forest has not significantly changed over the length of the record.

Zone L-1 (9.95–7.15 m depth, ca 10,000–7400 cal yr BP) featured moderate *Pinus* ( $\bar{x} = 60\%$ ), *Artemisia* ( $\bar{x} = 16\%$ ), *Abies* ( $\bar{x} = 5\%$ ) and Cupressaceae ( $\bar{x} = 3\%$ ) percentages. Macrofossils indicated that *P. monticola*, *P. cf. ponderosa*, and *A. concolor* were locally present.

Zone L-2 (7.15–2.57 m depth, ca 7400–3450 cal yr BP) was characterized by high *Pinus* percentages ( $\bar{x} = 65\%$ ), increasing *Abies* ( $\bar{x} = 7\%$ ), and low *Artemisia* ( $\bar{x} = 11\%$ ) values. Cupressaceae ( $\bar{x} = 4\%$ ) and Poaceae ( $\bar{x} = 3\%$ ) percentages were low. Needle fragments of *P. monticola*, *P. cf. ponderosa*, *Pinus concolor*, *A. concolor*, *C. decurrens* and *Juniperus occidentales* were recovered.

Zone L-3 (2.57–0 m depth, 3450 cal yr BP to present) had high percentages of *Pinus* ( $\bar{x} = 70\%$ ), and moderate-to-high values of Cupressaceae ( $\bar{x} = 6\%$ ), Poaceae ( $\bar{x} = 5\%$ ), and *Abies* ( $\bar{x} = 8\%$ ). *Artemisia* ( $\bar{x} = 9\%$ ) percentages were moderate to low in this zone. Macrofossils of *P. monticola*, *P. cf. ponderosa*, and *A. concolor* were identified.

The pollen and macrofossil records of Lily Lake and their comparison with modern analogs suggest that the forest changed little in the past 10,000 years. Changes in fire-episode frequency apparently did not alter the general composition of the vegetation.

### 3.3.3. Dead Horse Lake

The pollen record from Dead Horse Lake was divided into two zones (Fig. 7). Zone D-1 (2.33–1.12 m depth, ca 14,500–7000 cal yr BP) was characterized by high percentages of *Artemisia* ( $\bar{x} = 15\%$ ) and low percentages of *Sarcobatus* ( $\bar{x} = 2\%$ ), Asteraceae undiff. ( $\bar{x} = 2\%$ ), and Chenopodiaceae ( $\bar{x} = 3\%$ ). *Pinus* percentages ( $\bar{x} = 65\%$ ) were moderate in this zone. *Abies* pollen appeared in the record ca 10,000 cal yr BP. Modern pollen analogs for this zone come from elevations above 1900 m in the northern Rocky Mountains and Yellowstone region (Burkart, 1976; Millsbaugh et al., 2000), south in the Warner Mountains (Patterson Lake), and Sierra Nevada (Anderson and Davis, 1988). These locations suggest that Dead Horse Lake was

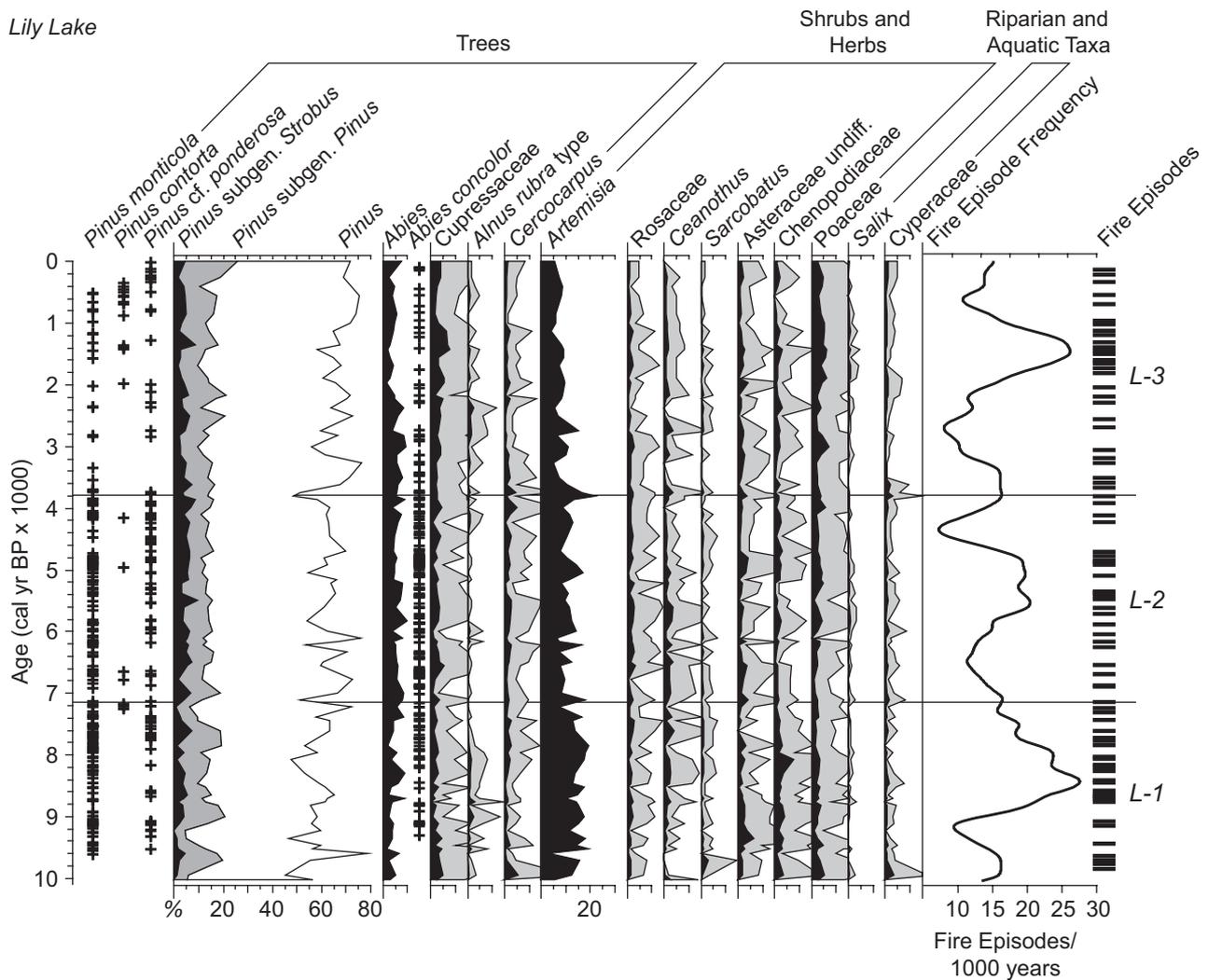


Fig. 6. Pollen percentage diagram for Lily Lake showing macrofossil data, selected pollen taxa, fire-episode frequency, and fire episodes (see Fig. 5). Light gray shading indicates  $5 \times$  exaggeration of pollen percentages.

predominantly surrounded by a western pine forest prior to 7000 cal yr BP but some subalpine and mesic forest elements were also present.

Zone D-2 (1.12–0 m depth, 7000 cal yr BP to present) was dominated by high *Pinus* ( $\bar{x} = 90\%$ ) and low *Artemisia* percentages ( $\bar{x} = 5\%$ ). *Abies* percentages ( $\bar{x} = 2\%$ ) were also low. Other shrub and herbaceous pollen types were poorly represented. The dominance of *Pinus* pollen reflects the development of pine forests following the deposition of ash from the eruption of Mount Mazama, 7626 cal yr BP. *P. contorta* and *P. cf. ponderosa* needle fragments were present at ca 1400 cal yr BP suggesting mixed conifer forests. Modern pollen analogs for this zone come from sites above 2500 m elevation in the Sierra Nevada (Anderson and Davis, 1988; Anderson, 1990), suggesting subalpine pine forest.

The pollen and macrofossil record from Dead Horse Lake indicates a lodgepole or ponderosa pine dominated forest with few (<3 fire episodes/1000 years) fires prior to 7000 cal yr BP. The deposition of the Mazama ash created

dry low-nutrient soils that favored lodgepole pine (see Franklin and Dyrness, 1988). This change in forest composition was not accompanied by greater fire-episode frequency, until ca 4500 cal yr BP. Fire frequency since then has widely varied, despite little change in vegetation.

The three fossil-pollen records from the northwestern margin of the Great Basin show similar patterns in the timing and direction of vegetation change. Prior to 11,300 cal yr BP, high-elevation steppe grassland grew near Patterson Lake while at Dead Horse Lake subalpine conditions prevailed. Forests developed ca 10,000 cal yr BP at Dead Horse Lake and ca 8550 cal yr BP at Patterson Lake. Since ca 6800 cal yr BP, forests have expanded upslope and become more dense at all three locations.

### 3.4. Climate reconstructions based on modern pollen analogs

Pollen data from Patterson, Lily, and Dead Horse lakes provide quantitative estimates of climate changes over the past 12,500 years. Three temperature and two moisture

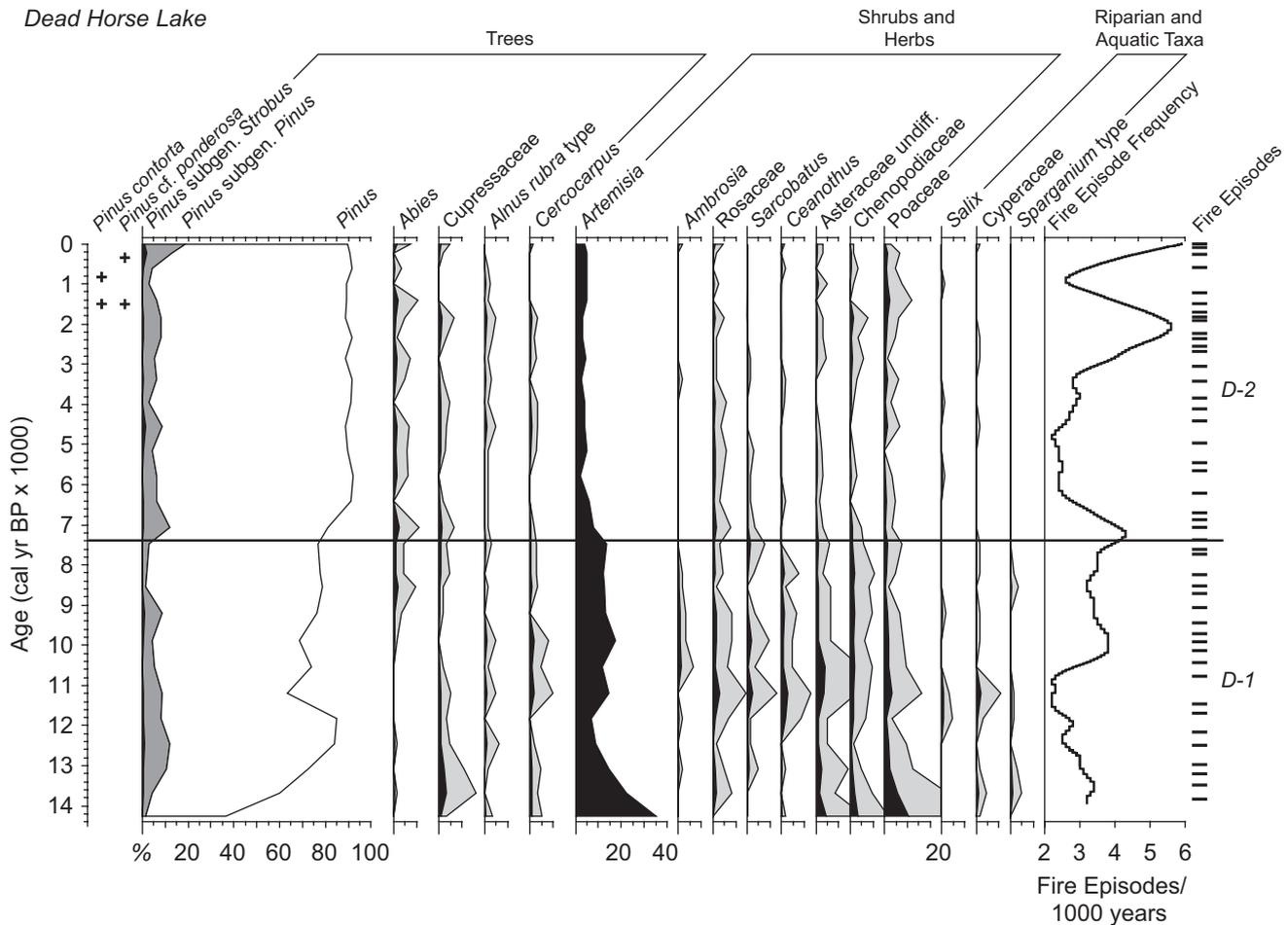


Fig. 7. Pollen percentage diagram for Dead Horse Lake showing macrofossil data, selected pollen taxa, fire-episode frequency, and fire episodes (see Fig. 5). Light gray shading indicates  $5 \times$  exaggeration of pollen percentages.

variables were selected for climate reconstruction because they relate to the establishment, growth, and distribution of plants (Prentice et al., 1992; Sykes et al., 1996; Thompson et al., 1999). The selected variables were growing degree-days base  $5^{\circ}\text{C}$  (GDD5) (i.e., growing season length and intensity), mean temperature of the coldest month (MTCO), mean temperature of the warmest month (MTWA), the ratio of actual evapotranspiration to potential evapotranspiration (AE/PE)—calculated as in Prentice et al. (1992), and annual precipitation (ANNP). Climate reconstructions for Dead Horse Lake were undertaken from 12,500 to ca 7000 cal yr BP. After this time, vegetation patterns around Dead Horse Lake are assumed to have been edaphically controlled by thick deposits of volcanic ash-rich soils (Franklin and Dyrness, 1988). Maximum dissimilarity values for each fossil spectrum were less than 0.15, suggesting good modern analogs for the fossil pollen spectra presented here (Anderson et al., 1989; Davis, 1995; Minckley, 2003).

Reconstructed GDD5 suggest more-than-present GDD5 from 12,500 to 10,500 cal yr BP at Patterson and Dead Horse lakes (Fig. 8). Patterson Lake data imply more GDD5 prior to 9000 cal yr BP, while the Dead Horse Lake

reconstruction suggests less GDD5 at ca 10,500 cal yr BP and GDD5 was generally similar to present-day values from ca 9300 to 7000 cal yr BP. Reconstructed GDD5 was more-than-present at Patterson Lake from 9300 to 1000 cal yr BP. However, reconstructed GDD5 values were less-than-present at Lily Lake over the same time interval. During the past 1000 years, GDD5 synchronously varied near present day values at both Patterson and Lily lakes.

Reconstructed MTCO from 12,500 to 7000 cal yr BP for Patterson Lake was out of phase with the reconstructed MTCO of Dead Horse and Lily lakes (Fig. 8). At Patterson Lake, MTCO was up to  $3^{\circ}\text{C}$  higher-than-present, whereas at Dead Horse and Lily lakes, MTCO reconstructions were up to  $3^{\circ}\text{C}$  lower-than-present conditions. From ca 6800 to 5200 cal yr BP, MTCO was within  $2^{\circ}\text{C}$  of present-day values at both Patterson and Lily lakes. Both records show similar trends in the timing and duration of increases and decreases in MTCO. MTCO reconstructions for 5200 cal yr BP to present were out of phase for Patterson and Lily lakes. MTCO for Patterson Lake was generally higher than present, while MTCO at Lily Lake was generally lower than present. However, both records

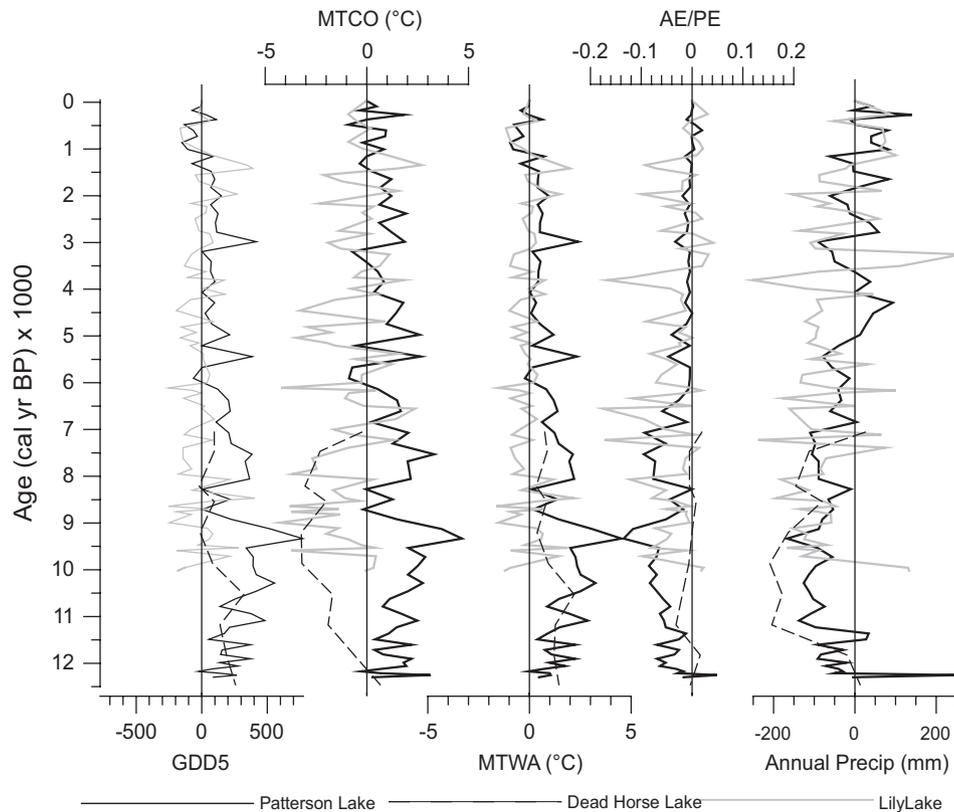


Fig. 8. Climate reconstructions Patterson (solid black line), Dead Horse (dashed black line), and Lily (solid gray line) lakes. Five climate variables are shown, growing degree days-base 5 °C (GDD5), mean temperature of the coldest month (MTCO), mean temperature of the warmest month (MTWA), the ratio of actual evapotranspiration to potential evapotranspiration (AE/PE) and Annual Precipitation (ANNP). Climate anomalies relative to reconstructed present-day climate for each site are based on the weighted average climate values for the 10 closest analogs determined from the squared-chord distances of 2013 surface samples from western North America (Minckley, 2003). Vertical lines show present-day climate values.

indicated similar-to-present MTCO (i.e.,  $\pm 1$  °C) from 4200 to 3200 cal yr BP and 1000 cal yr BP to present.

Reconstructed MTWA was 1–3 °C higher than present from 12,500 to 10,000 cal yr BP at Patterson and Dead Horse lakes (Fig. 8), and increased at Patterson Lake to a peak of  $\sim 4$  °C higher than present at 9300 cal yr BP. Between 9300 and 1500 cal yr BP, MTWA was higher than present at Patterson Lake. Reconstructed MTWA was similar to present at Lily Lake during the same interval. Between 1000 and 500 cal yr BP, MTWA was 1 °C lower than present at both Patterson and Lily lakes.

Reconstructed AE/PE was lower than present at Patterson and Dead Horse lakes between 12,500 and 10,000 cal yr BP (Fig. 8). AE/PE was generally lower than present but trending towards present-day values between 9200 and 3500 cal yr BP at both Patterson and Lily lakes, although the amplitude of variations in AE/PE was lower at Patterson Lake. From 9200 to 7000 cal yr BP, reconstructed AE/PE was similar to present at Dead Horse Lake. At Lily Lake, AE/PE was  $> 10\%$  lower than present ca 8100, 7250, 6600, 6300, and 3850 cal yr BP. From 3500 to present, AE/PE was generally within 2% of present day values at both Patterson and Lily lakes.

Reconstructed ANNP was similar to present at 12,500 cal yr BP but was 100–200 mm/yr lower than present by 10,000 cal yr BP at Patterson and Dead Horse lakes (Fig. 8). ANNP was lower than present between 10,000 and 5000 cal yr BP at Patterson, Dead Horse and Lily lakes. From 5000 to 3500 cal yr BP, ANNP was higher than present at Patterson Lake. The largest reconstructed anomalies in precipitation values ( $\pm 250$  mm/yr) were evident from 3900 to 3300 cal yr BP at Lily Lake. Between ca 3000 and 300 cal yr BP, ANNP increased from 100 mm/yr lower than present to 100 mm/yr higher than present at both Patterson and Lily lakes and then decreased to present values.

Pollen-based climate reconstructions from Patterson, Dead Horse and Lily lakes suggest similar or more-than-present GDD5, lower-than-present MTCO, and higher-than-present MTWA from 12,500 to ca 9000 cal yr BP. During this time, moisture availability was lower than present in terms of AE/PE and ANNP. After 9000 cal yr BP, decreasing temperature (GDD5, MTWA) and increasing moisture (AE/PE, ANNP) suggest shared responses among these variable to decreased summer insolation, and weakening of the eastern North Pacific Subtropical High (COHMAP members, 1988; Bartlein

et al., 1998). Since 9000 cal yr BP, higher ANNP has offset decreases in AE/PE.

#### 4. Discussion

Comparison of this study with five other pollen studies from central and southeastern Oregon provides a picture of environmental change for the northern Great Basin over the last 14,000 years (Fig. 9). These sites are Bicycle Pond, OR (Wigand and Rhode, 2002); Craddock Meadow, OR (Wigand, 1989); Diamond Pond, OR (Wigand, 1987); Fish Lake, OR (Mehring, 1985); and Wildhorse Lake, OR (Mehring, 1985). Bicycle Pond and Diamond Pond are located near the lower forest/steppe border (Fig. 1B) (Wigand, 1989; Wigand and Rhode, 2002). Craddock Meadow lies in the pine-fir zone near the ecotone with sagebrush-grassland (Wigand, 1989). Fish Lake and Wildhorse Lake are located in the upper-elevation sagebrush-grassland zone (Mehring, 1985). Craddock Meadow and Fish Lake span the last 14,000–15,000 cal yr (Mehring, 1985; Wigand, 1989). Wildhorse Lake has a 10,500-cal-yr-long record, Bicycle Pond has a 9500-cal-yr-long record, and Diamond Pond spans the last 6000 cal yr (Mehring, 1985; Wigand, 1987; Wigand and Rhode, 2002). The original  $^{14}\text{C}$  ages of the pollen zone boundaries for these sites were converted to cal yr BP using INTCAL04.14C (Stuiver et al., 1998) but new age–depth models were not calculated.

Pollen and macrofossil evidence suggests that most of the trees and shrubs that currently grow in the Warner Mountains and Dead Horse Rim have been in the region for the last 14,000 years. The early appearance of white-bark pine and white fir implies proximal glacial distributions, which allowed these species to respond rapidly to climate warming. At Patterson Lake and Dead Horse Lake, pollen and charcoal data suggest upper-elevation sagebrush grassland and subalpine forest prior to 11,000 cal yr BP. Late-glacial vegetation at Patterson Lake, Dead Horse Lake, Craddock Meadow, and Fish Lake was more open than at present (Mehring, 1985; Wigand, 1989; Wigand and Rhode, 2002). The quantitative climate reconstruction for Patterson Lake and qualitative inferences from other sites indicate warmer winters than present and slightly drier than present conditions during the late-glacial period (Mehring, 1985; Mehring and Wigand, 1990; Thompson et al., 1993; Wigand and Rhode, 2002). Fire activity was low during this period, likely because of low biomass around these lakes after deglaciation.

Warmer and drier than previous conditions from ca 11,000 to 9000 cal yr BP allowed open forests and steppe conditions to develop at higher elevations (Fig. 9) (Mehring, 1986). Mehring (1986) proposed that warming temperatures from ~10,000 to 8000 yr BP may have been associated increases in summer rainfall. Upper-elevation steppe-grassland formed at Patterson Lake, and subalpine pine forest with steppe was present at Dead Horse Lake from 11,000 to 9000 cal yr BP. At Lily Lake,

the mixed conifer forest was more open than present at this time. Warm, dry conditions between 11,000 and 9000 cal yr BP likely created patchy forest cover at all elevations and restricted forests to north facing slopes (Thompson, 1990; Nowak et al., 1994a,b; Wigand and Rhode, 2002). After ca 9000 cal yr BP, white fir expanded upslope and became prominent in the mid-elevation forests in the Warner Mountains and on Dead Horse Rim, whereas elsewhere in the region, sagebrush steppe persisted until 6000 cal yr BP (Mehring, 1985, 1986; Wigand, 1989; Wigand and Rhode, 2002). This vegetation change in the middle Holocene indicates, increases in effective moisture, relative to the early Holocene, in the mountains of the northwestern Great Basin, consistent with qualitative climate interpretations (Mehring, 1986). However, quantitative climate estimates from Patterson, Lily, and Dead Horse lakes indicate warmer-than-present summers and drier-than-present conditions persisted from the early Holocene to ca 7000 cal yr BP.

Other paleobotanical evidence from the Great Basin also suggests dry conditions from 9000 to 8000 cal years BP. For example, upper treeline of bristlecone pine (*Pinus aristata*) was over 150 m higher than present at ca 8000 cal yr BP in the White Mountains, California (LaMarche, 1973). At Ruby Marshes in northeastern Nevada, an increase in Chenopodiaceae pollen also implies warm dry conditions ca 8000 cal yr BP (Thompson, 1992).

Present-day vegetation patterns developed after 7000 cal yr BP across the northwestern Great Basin. Forest cover increased, beginning with an increase of pines, followed by increases in white fir at Patterson, Lily and Dead Horse lakes (Figs. 5–7). The gradual progression to modern conditions in the Warner Mountains and on Dead Horse Rim over the past 7000 years suggests that middle and late Holocene climate changes have led to only subtle changes in vegetation in the northwestern Great Basin. Alternatively, it could be that the climate changes themselves were subtle. Fire-episode frequency fluctuated with the shift to closed forests, particularly at Lily and Dead Horse lakes. Non-forested sites generally showed increases in grass (Fish Lake, Wildhorse Lake) and juniper (Bicycle Pond, Diamond Pond) and reductions of sagebrush starting between 5000 and 4000 cal yr BP (Mehring, 1985; Wigand, 1989; Mehring and Wigand, 1990; Wigand and Rhode, 2002).

The quantitative and qualitative climate reconstructions inferred from the pollen data suggest that effective moisture increased after ca 7000 cal yr BP in the northwestern Great Basin. This increase in effective moisture likely resulted from decreasing summer insolation, and its effect on atmospheric circulation in the region (Bartlein et al., 1998). Concomitant increases in winter insolation may have led to greater atmospheric water holding capacity, which would account for the higher annual precipitation suggested by the climate reconstructions (Fig. 8). Increasing annual precipitation since 7000 cal yr BP may represent the reduced strength of the

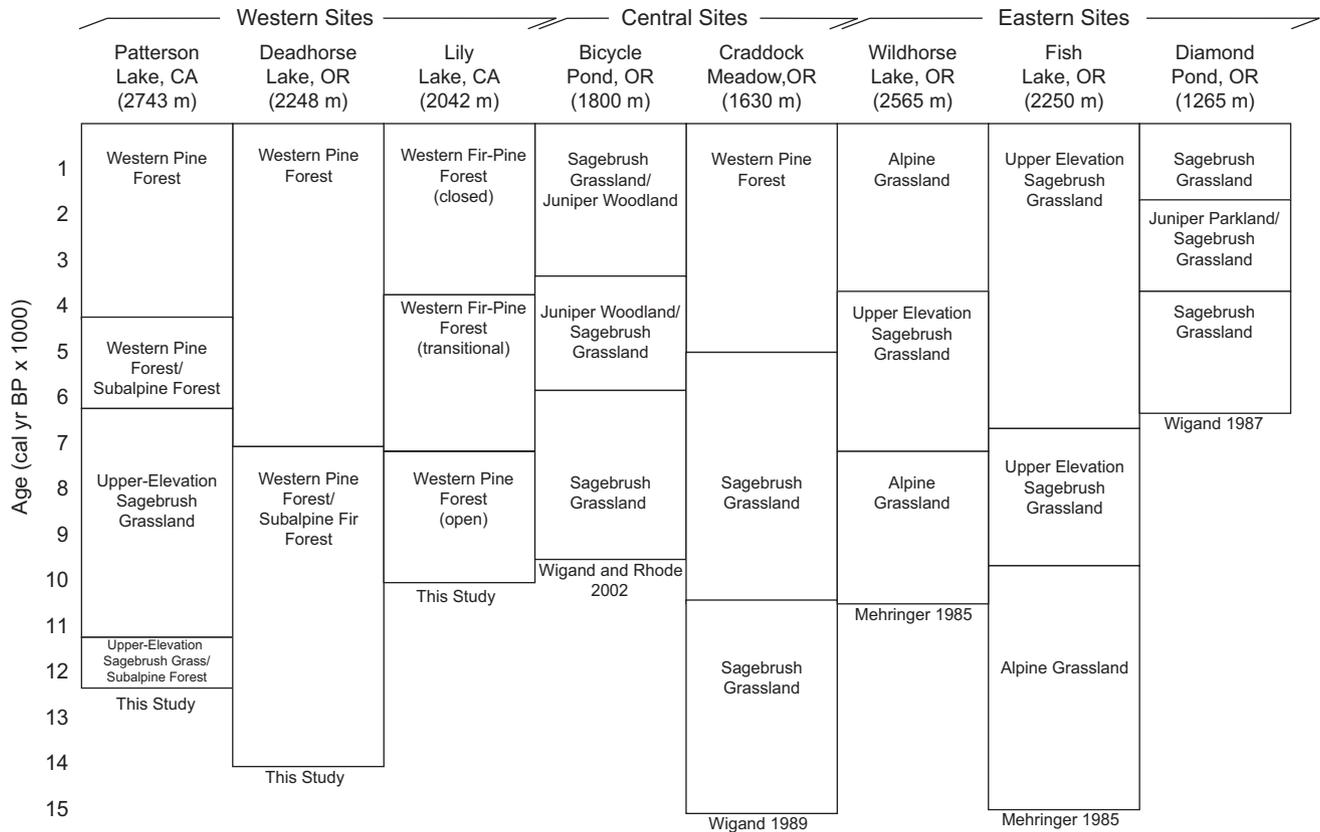


Fig. 9. Summary of regional vegetation changes in the northwestern Great Basin. Sites are arranged by elevation in three groups: west (Patterson Lake, Dead Horse Lake, Lily Lake), central (Bicycle Pond, Craddock Meadow), and east (Wildhorse Lake, Fish Lake, Diamond Pond) (see Fig. 1). Zone descriptions based on Billings (1951).

Aleutian Low, which coincides with the progressive weakening of the summer sea-level pressure anomaly, leading to an earlier onset of winter, and delayed onset of the summer drought (Bartlein et al., 1998).

Millennial-scale trends in fire-episode frequency at Patterson, Lily, and Dead Horse lakes show some similarities in their fire history (Fig. 4), but changes in fire activity were independent of vegetation composition. The general synchronicity of fire-episode variations but differences in the number of fires/1000 years over the past 6000 years suggests widespread differences in actual fire occurrences. For example, Patterson Lake and Dead Horse Lake register low fire-episode frequency over their entire record, whereas Lily Lake registers more variable fire-episode frequency, particularly in the past 6000 years. Lily and Dead Horse lakes indicate that the last period of low fire frequency occurred from 800–700 cal yr BP. The difference in fire activity between Lily Lake and other sites is partly explained by the greater temporal resolution and higher fire frequency of the record.

The difference in fire activity between these records may also reflect different intensity of anthropogenic burning. In the northwestern Great Basin, human populations increased over the past 5000 years (Jenkins, 1994; Oetting, 1999), but population density remained relatively low, and

little is known about the use of fire by Great Basin peoples (Griffin, 2002). Between 2000 and 600 cal yr BP, the last period of relatively high fire activity, peoples of the northwestern Great Basin began to intensify their use of upland resources (Jenkins, 1994; Oetting, 1999) and it is possible that greater upland utilization by these people led to increased fire activity after 2000 cal yr BP.

## 5. Conclusions

This study examined the vegetation, fire, and climate history of the northwestern Great Basin for the last 14,000 years to understand the evolution of the modern landscape. The regional vegetation history indicates a greater-than-present areal extent of low and high elevation steppe grassland from ca 11,000 to 7000 cal yr BP, followed by an expansion of forest and increase of grasses in steppe regions. The dominant conifers of the region (i.e., white fir, western white pine, and whitebark pine) have been present in the region since ca 12,000 cal yr BP. Fire reconstructions show that in high-elevation forests, fire episodes were infrequent (2–5 fire episodes/1000 years) during the late-glacial and Holocene. Fire episodes in mid-elevation forests have been variable (10–25 fire episodes/1000 years) over the last 10,000 years. Applying modern pollen–climate

relationships to the fossil pollen spectra provides a means for interpreting past climate changes in this region. Climate reconstructions indicate that in the past 9000 years winters have generally become warmer, summers have generally become cooler, and over the last 6000 years, annual precipitation has increased. Combined these analyses provide a quantification of Holocene climate and environmental change in the transition zone between the mesic ecosystems of the Pacific Northwest and xeric ecosystems of the intermountain West.

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