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Reconstruction of Holocene Climate Change Using Multiproxy Analysis of Sediments from Pyramid Lake, British Columbia, Canada

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Abstract

Sediment cores from Pyramid Lake, an alpine tarn in the Cassiar Mountains of northwestern British Columbia, were investigated for changes in pollen, plant macrofossils, charcoal, and clastic sediment, which are used to infer changes in climate throughout the Holocene. Radiometric dating has yielded a chronology of high-magnitude rainstorm events and timberline migration for the Pyramid Lake basin since deglaciation at about 10600 B.P. Fifteen distinct minerogenic layers represent material delivered to the lake by runoff events. The frequency of minerogenic layer deposition, and by analogy of storms, has changed throughout the Holocene. Four large-magnitude rainstorm events occurred between 4400 and 5100 B.P. During this period white spruce (*Picea cf. glauca*) was likely present near the lake, although a closed forest stand did not develop around the lake at any point during the Holocene. The macrofossil record indicates that subalpine fir (*Abies lasiocarpa*) has been present, likely as krummholz, above the elevation of the lake since at least 9400 B.P. Pollen of western hemlock (*Tsuga heterophylla*) is represented from ca. 1500 B.P. to the present and may be a consequence of changes in regional air-mass circulation patterns.

Introduction

Lake sediment can be used as a paleoclimatic proxy in two ways. First, sediment contains biogenic material such as macrofossils, pollen, algae (e.g., diatoms, dinoflagellates, chrysophytes), ostracodes, molluscs, and other invertebrates that can be used to infer paleoenvironmental conditions. Second, the physical properties of the sediments, such as bulk density; x-ray transparency; grain size; magnetic susceptibility; and organic, inorganic, and charcoal carbon content, can be used to determine depositional environments, which may be linked to paleoclimate. This study utilizes both biogenic and physical characteristics of sediment obtained from Pyramid Lake (informal name) in northwestern British Columbia to interpret climate change within the Cassiar Mountain region of the northwestern Cordillera.

The existing paleoclimatic record of the northwestern Cordillera has been largely based on reconstructions of vegetation using fossil pollen (Miller and Anderson, 1974; Banner et al., 1983; Friesen, 1985; Cwynar, 1988; Stuart et al., 1989; Gottesfeld et al., 1991; Cwynar, 1993; Hebda, 1995; Clague and Mathewes, 1996; Spear and Cwynar, 1997; Spooner et al., 1997, 2002) and geomorphic evidence (Ryder, 1987; Ryder and Maynard, 1991; Clague and Mathewes, 1996; Spooner et al., 1997). The climate of many of these sites is dominated by Pacific air (Banner et al., 1983; Cwynar, 1993; Clague and Mathewes, 1996; Spear and Cwynar, 1997) and is likely not comparable to the more continental climate of the interior mountains (Miller and Anderson, 1974). To date, only one study (Spooner et al., 1997) has been carried out in the western portion of the interior mountains of the northwestern Cordillera.

To gain insight into the variability of Holocene climate in the northwestern Cordillera, this study utilizes lacustrine sedimentology, macrofossils, charcoal, and fossil pollen as proxies for past environmental conditions, which in turn are used to infer climate. Because timberline is controlled primarily by temperature (Tranquillini, 1979) and responds rapidly to changes in climate (Reasoner and Hickman, 1989), Pyramid Lake sediments should record past fluctuations in timberline because the lake is located in the alpine zone near the present timberline. This paper also demonstrates that some alpine tarns have the potential to record high-magnitude precipitation events. Past studies have identified minerogenic lacustrine sediments resulting from large-magnitude rainstorms, which may trigger slope runoff (Brown et al., 2000), landslides (Eden and Page, 1998), and debris flows (Rodbell et al., 1999). Minerogenic layers within Pyramid Lake sediment most likely originated from slope runoff events triggered by large-magnitude rainstorms. It is hypothesized that within the study area, the frequency of large storm events may be related to changes in summer atmospheric circulation patterns. The climate of the northwestern Cordillera is strongly influenced by the incursion of moist Pacific air into dry stable arctic air masses (Meidinger and Pojar, 1991). Changes in the extent and intensity of maritime incursions or the stability of continental air dramatically affect the climate of the region. The frequency of rainstorm events is compared to palynological records that suggest that the presence of exotic pollen represents increased summer penetration of coastal air into interior regions (Miller and Anderson, 1974; Cwynar, 1993; Spooner et al., 1997; Cwynar, 1998; Spooner et al., 2002).

Study Area

Pyramid Lake, located in the Cassiar Mountains of the northwestern Cordillera (58°53′N, 129°50′W; Fig. 1), is a small tarn about 50 m higher than present-day timberline, with a maximum fetch of 480 m and a maximum measured depth of 9 m. Based on lake size and alpine setting with strong winds, the lake is likely polymictic (Gorham and Boyce, 1989) with minimal sediment mixing due to slumping or bioturbation (Larson and MacDonald, 1993). The lake is situated in the transition between the spruce-willow-birch and alpine tundra biogeoclimatic zones and within the zone of sporadic permafrost (Meidinger and Pojar, 1991). The alpine tundra immediately surrounding the lake consists of grass, sedge, shrubs of alder (*Alnus sinuate*), willow (*Salix arctica* and *S. planifolia*), dwarf birch

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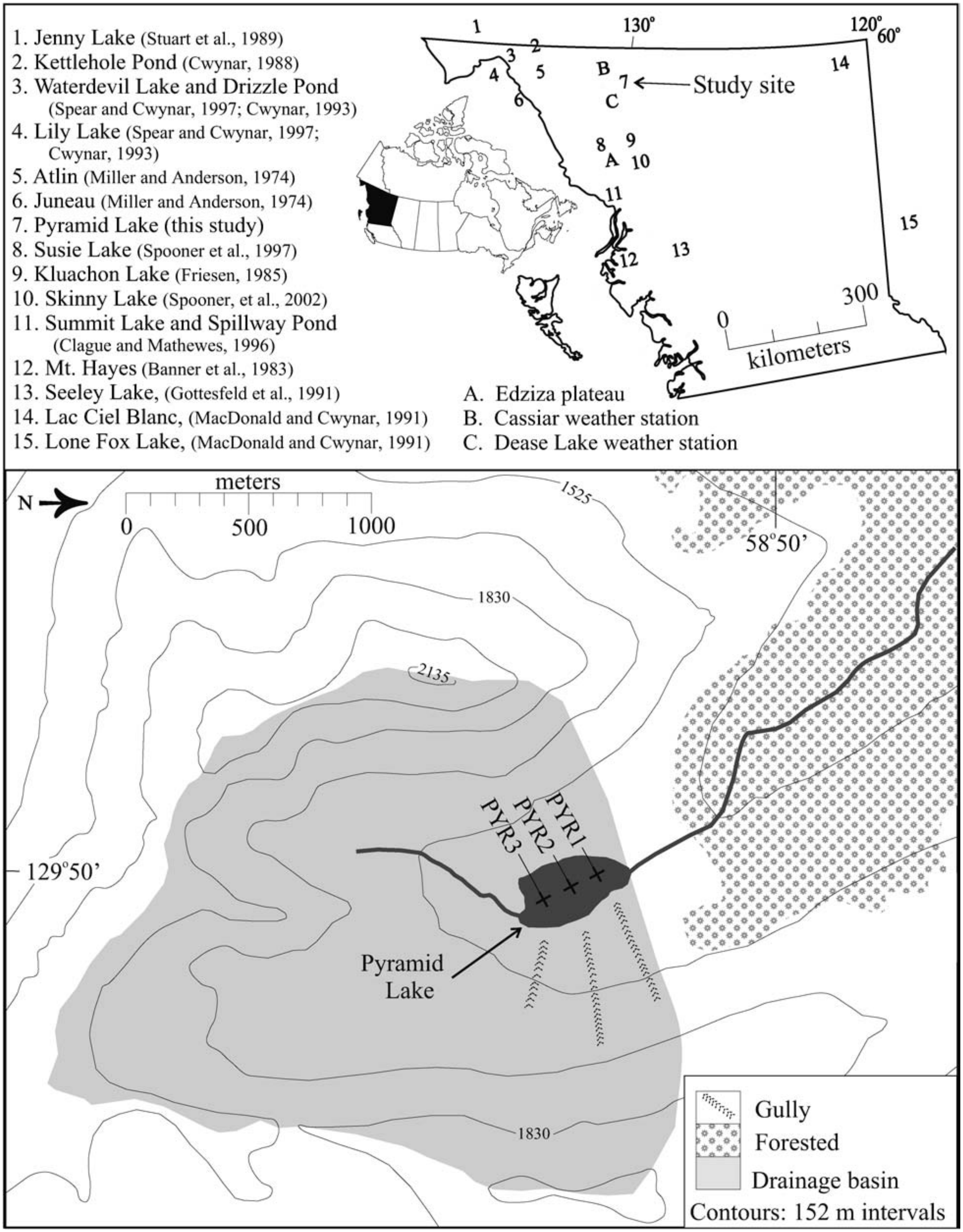

FIGURE 1. Map showing sites of paleoenvironmental investigation and relevant weather stations in the northwestern Cordillera as well as the topography of the Pyramid Lake drainage basin (from NTS map sheet 104 I/13W), timberline, gullies, and locations of sediment cores PYR1, PYR2, and PYR3.

(*Betula glandulosa*), and krummholz subalpine fir. On the steep slopes and ridges above the lake, vegetation is sparse, with occasional patches of deciduous shrubs, grasses, and sedges. Three vegetated gullies are incised into the eastern slope above the lake, and steep talus forms the western shore. A stream fed by groundwater enters the lake at its south end, and water exits the lake through and over volcanic and metamorphic bedrock (Gabrielse, 1998). The region is seismically (Basham et al., 1985) and volcanically (Souther, 1977) active.

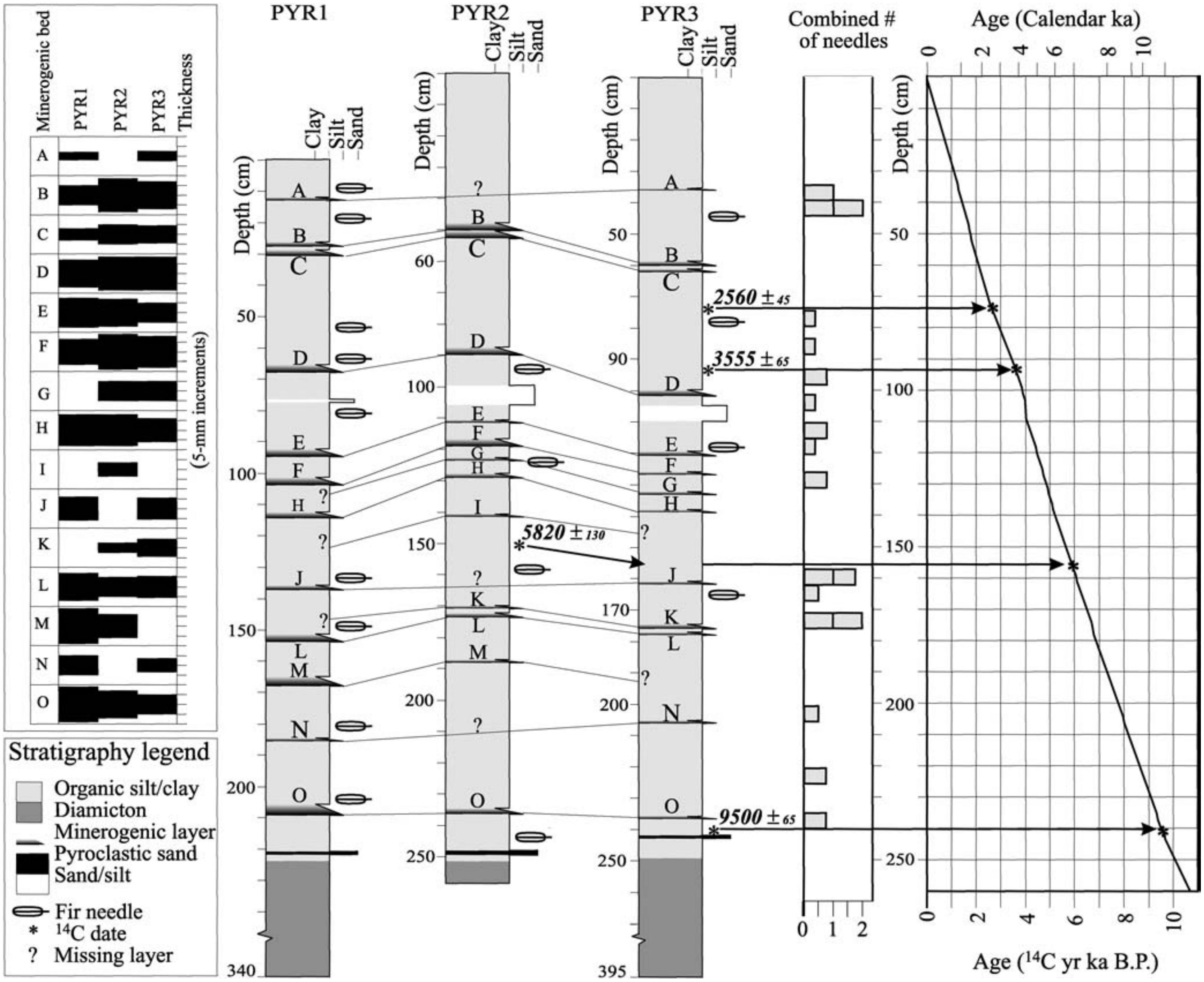

FIGURE 2. Stratigraphy of cores from Pyramid Lake, age-depth relation, and minerogenic bed thickness.

Long-term climate data available from Dease Lake, about 55 km southwest and 630 m lower (C in Fig. 1), and Cassiar, about 40 km north and 750 m lower (B in Fig. 1), suggest that average annual precipitation within the region is 700 mm, with 60% falling as snow. The mean annual air temperature is –3.2°C, with 7 mo below 0°C and 1 mo above 10°C (Meidinger and Pojar, 1991). Pyramid Lake is wetter and cooler than the listed stations due to the effect of elevation. During the summer, rain is generally the product of surface heating, which leads to sudden, often violent convective showers (Meidinger and Pojar, 1991).

Methods

SAMPLE COLLECTION AND PREPARATION

Using a modified percussion corer (Gilbert and Glew, 1985; Reasoner, 1993), we retrieved three 2.7- to 3.5-m-long continuous cores that all bottomed in diamicton. The cores were x-rayed using a Pickard Industrial x-ray unit and then split lengthwise. The stratigraphy of the three cores is summarized in Figure 2. At 5-cm intervals and directly above and below stratigraphic changes, 1-cm^3 samples were extracted for pollen, magnetic susceptibility (Sapphire Instruments–2B™), dry bulk density, and loss on ignition (LOI 550°C and LOI 950°C after Dean, 1974), and 2-cm^3 samples were used for macroscopic charcoal analysis following the procedures of Waddington (1969). Core PYR3 was selected for detailed analysis because the upper sediments appeared to be the least disturbed. The microsedimentology of individual inorganic layers in this core was observed in thin sections prepared using techniques modified from Lamoureux (1994), and grain size was determined using a Coulter™ LS200 laser particle-size analyzer.

Following sampling, contiguous <5-cm half-core sections were removed, avoiding disturbed portions along the core tube, and washed through a 250-μm mesh to recover plant macrofossils. Macrofossils were identified using modern reference specimens and a reference key (MacKinnon et al., 1992). A combined needle record (Fig. 2) was created by visually correlating macrofossil depths of PYR1 and PYR2 to that of PYR3 by their occurrences relative to the top of the core and the two sand layers.

Pollen influx estimates were made by the addition of 2 tablets of exotic *Lycopodium* spores to each sample (Stockmar, 1971), following the methods outlined by Birks and Birks (1980). Pollen was prepared as outlined by Faegri and Iversen (1975), followed by sieving (Cwynar et al., 1979).

POLLEN COUNTS AND ZONATION METHODS

Palynomorphs and spores were identified and counted using reference slides and reference texts (Kremp, 1965; McAndrews et al.,

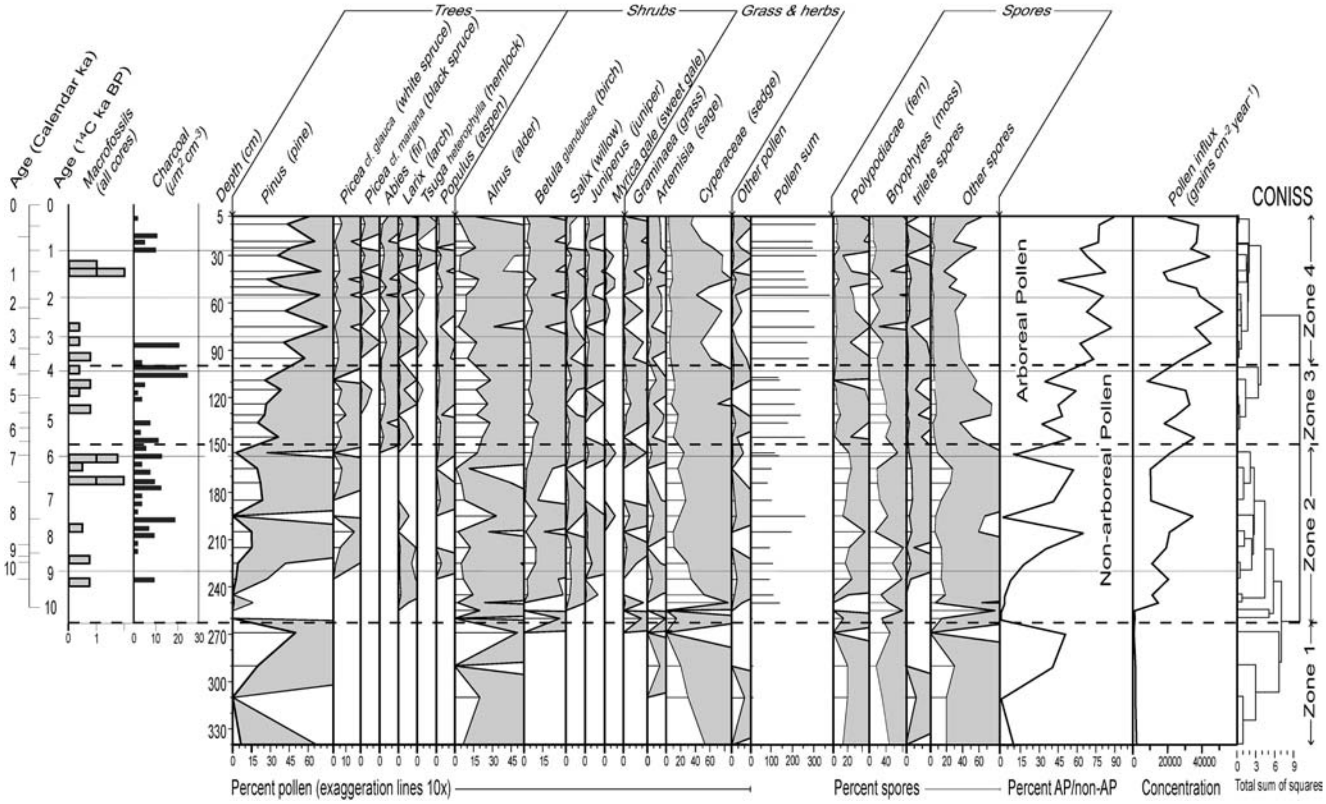

FIGURE 3. Percent pollen diagram for core PYR3, charcoal counts, and combined macrofossils from all cores. Gray horizontal lines are 1000-radiocarbon-yr intervals; thick dashed lines are zone boundaries.

1973; Bassett et al., 1978). Pollen sums include all terrestrial pollen, with at least 300 pollen grains identified per sample, apart from depths having low pollen concentration, for which counting ceased when 100 exotic *Lycopodium* spores were tallied. All small (ca. 10 µm) psilate spores were recorded as Bryophytes, and percent spores are relative to the pollen sum. No attempt was made to distinguish between varieties of trilete spores or between types of *Pinus*. *Picea* pollen was separated into species given the criteria of McAndrews et al. (1973). Due to uncertainties in separating *Picea*, the estimated percentages carry the designation "cf." (Cushing, 1963). Taxa occurring in most samples or those with concentrations greater than 1% are displayed on the pollen percentage diagram (Fig. 3). Values are rounded to the nearest whole percent in the text. Pollen included in the pollen sum but not displayed on the pollen diagram include *Ambrosia*, *Chenopodiaceae*, *Dryas*, *Epilobium*, *Eupayorium*, *Fraxinus*, *Oxyria*, *Potamogeton*, *Quercus*, *Sheperdia canadensis*, and *Thuja*. Pollen data were plotted using CANPLOT (Campbell and McAndrews, 1992) and divided into zones using cluster analysis, accomplished with CONISS (Grimm, 1987).

DATING AND SEDIMENTATION RATES

One beta-counting and 3 accelerator mass-spectrometer radiocarbon dates were obtained from wood within PYR2 and PYR3, respectively (Table 1). Two sand layers provide isochrons for stratigraphic correlations among cores. The date from PYR2 was correlated to PYR3 by its position relative to the uppermost sand bed.

Fifteen inorganic graded layers and two sand beds in PYR3 were likely deposited instantaneously (cf. Eden and Page, 1998). Sedimentation rate curves were constructed graphically following the methods of Berglund and Raslska-Jasiewiczowa (1986) and reveal a relatively uniform sedimentation rate (Fig. 2). Dates within the text are in uncalibrated radiocarbon years (B.P.) to simplify comparison with published records. Appropriate diagrams have both uncalibrated radiocarbon and calendar time scales based on conversions according to Stuiver and Reimer (1993).

Results

PHYSICAL PROPERTIES

X-radiography revealed the presence of concentrations of granule- and pebble-sized clasts not visible on the surface of the split core (Fig. 4). These pebbly zones are not present below about 200 cm in the record and increase in frequency upcore (Fig. 5). Granule and pebble concentration may be related to eolian (McKenna-Neuman, 1990) or snow avalanche processes (Luckman, 1975) that transport sediment onto lake ice. The absence of pebbled layers within the lower portion of the sediments is enigmatic.

Charcoal abundance is low throughout the core, which is to be expected given the low amounts of fuel immediately available within the alpine zone. Most of the charcoal was likely transported by wind from lower-elevation forest fires. Concentrations of charcoal occurring at about 1, 4, and 7.6 ka B.P. (Fig. 5) record periods of extensive, likely drought-related fire. Charcoal is absent from the record between about 1 and 3.2 ka B.P., suggesting that wetter and perhaps cooler summer conditions prevailed.

Magnetic susceptibility (MS), loss-on-ignition, bulk density, and grain-size data exhibit coincident, distinct variations that correspond to visible changes in the sediment (Fig. 5). Generally, from bottom to top of the cores, MS and bulk density decrease, LOI 550°C increases, and LOI 950°C remains low throughout. The largest divergence in MS and LOI 550°C occurs at a sand layer near the base of the cores. This sand layer occurs at about 9600 B.P. in all three cores, is massive in structure, is well sorted, contains no clay, and has a predominantly mafic mineralogy consistent with basalt according to lithoprobe analysis (S. Barr, personal communication, 2000). As there is no

TABLE 1

Radiocarbon dates for Pyramid Lake, northwestern British Columbia

Core	Depth (cm)	Laboratory number and material	Age (yr B.P.)	Calendar B.P. age ranges (2 sigma)	Relative area under distribution
PYR 3	73	AA36379	2560 ± 45	2765–2684	0.988
		wood		2487–2479	0.012
PYR 3	93	AA36380	3555 ± 65	4071–4046	0.021
		wood		4034–4034	0.001
				3988–3685	0.962
				3662–3642	0.016
PYR 3	241	AA36381	9500 ± 65	11091–10934	0.368
		wood		10932–10919	0.009
				10910–10634	0.557
				10631–10578	0.066
PYR 2	160	BGS2158	5820 ± 130	6907–6379	0.964
		wood		6373–6311	0.036

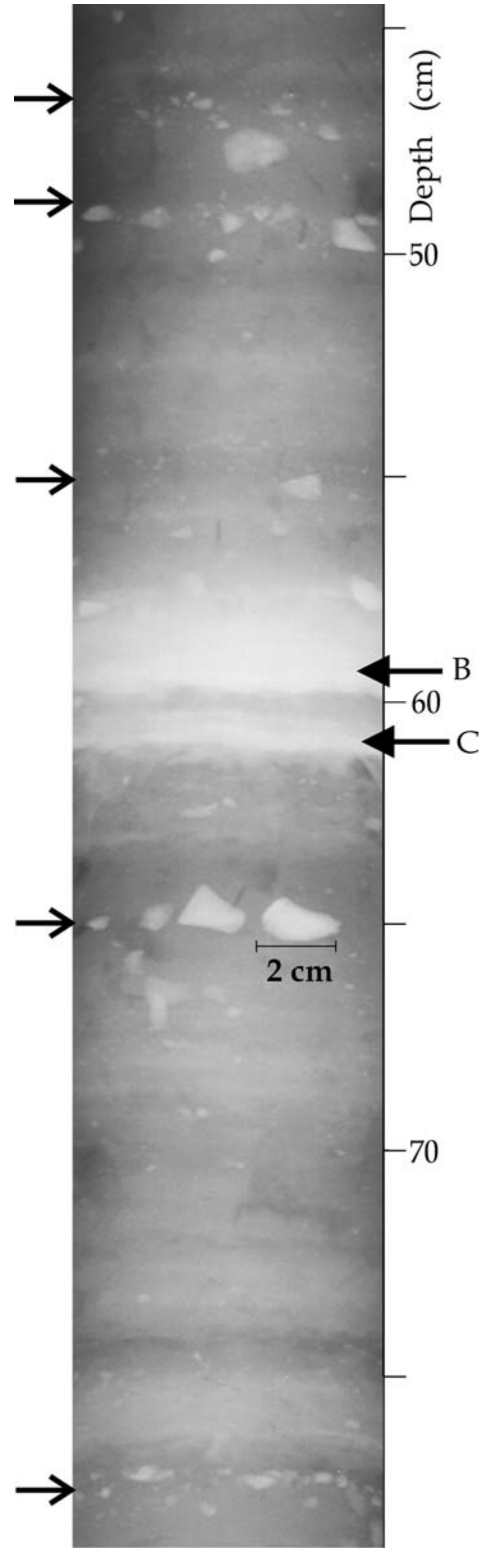

FIGURE 4. X-radiographic negative of a section of core PYR3 showing minerogenic layers B and C and pebble zones.

basalt within the drainage basin, the sand is likely pyroclastic ejecta previously undocumented within the region, perhaps originating from the Edziza plateau (A in Fig. 1).

The bottom sediment within the three cores is light gray, massive, inorganic silt and clay, with random, angular, sand- to cobble-sized clasts. This diamicton was likely deposited in the ice-proximal environment of a late Pleistocene retreating glacier. The remainder of the cores consists of a dark, organic silt/clay, which represents slow accumulation of biological material and fine clastic sediment. Interspersed within this organic silt/clay are discrete beds and laminae of light-colored, minerogenic sediment that typically demonstrate high MS and bulk density and low LOI 550°C. The bottoms of the minerogenic layers are in sharp contact with the organic silt/clay and normally grade from sandy silt to fine silt and clay (Fig. 6). The minerogenic layers must be allochthonous, as greater MS and grain size at the bases of the layers cannot be explained by redistribution of autochthonous material.

Minerogenic layers are scattered through the organic portions of the cores, occurring on average every 16 cm or every 630 yr. The largest spacing between minerogenic layers is 48 cm (about 1900 yr), and a few are separated by less than a centimeter (about 40 yr). The highest concentration of minerogenic layers is fourwithin 20 cm, identified by a shaded bar on Fig. 5. This concentration of layers formed between ca. 4400 and 5100 B.P. and can be correlated between all three cores (Fig. 2).

Correlation of minerogenic layers was done visually by their position relative to the top of the core, the two sand layers, and each other (Fig. 2). Minerogenic layer deposition occurred a maximum of 15 times throughout the record. Eight of these minerogenic layers (B, C, D, E, F, H, L, O; Fig. 2) occur in all three cores; one layer (I in Fig. 2) occurs in only one core. Minerogenic layers range in thickness from 4 to 18 mm. Average thickness is 12.5 mm within cores PYR1 and PYR2 and 10.9 mm in core PYR3. Individual layers range in thickness between cores (Fig. 2); three layers (E, H, O) become thinner toward the proximal end of the lake.

TIMBERLINE POLLEN ASSEMBLAGES

Because the coarse-grained event deposits likely contain high percentages of pollen detritus, we exclude samples from those layers from the cluster analysis. Zone 1 (bottom—268 cm, unknown age to ca. 10600 B.P.) has a very low average pollen concentration of 920 grains $cm^{-2}\ yr^{-1}$ (Fig. 3). Zone 2 (268 to 150 cm, ca. 10600 to 5600 B.P.) is designated the "white spruce-alder-birch-sedge-fern" zone. Pollen concentration generally increases upward through the zone from 14,370 to 22,950 (average 17,630) grains $cm^{-2}\ yr^{-1}$. Nonarboreal pollen and high percentages of spores dominate the beginning of this zone. Pine and white spruce fluctuate markedly through the zone but are nowhere more than 24% and 17%, respectively. Willow and grass (Gramineae) pollen average 3% and 2%, respectively; birch pollen averages 5%. Aspen (*Populus*) and larch (*Larix*) occur intermittently and average about 1% throughout the zone.

Zone 3 (150 to 100 cm, 5600–3800 B.P.) is designated the "pine-alder-white spruce-birch" zone. Pollen concentrations generally increase through the zone, from 7650 to 32,780 (average 22,560) grains

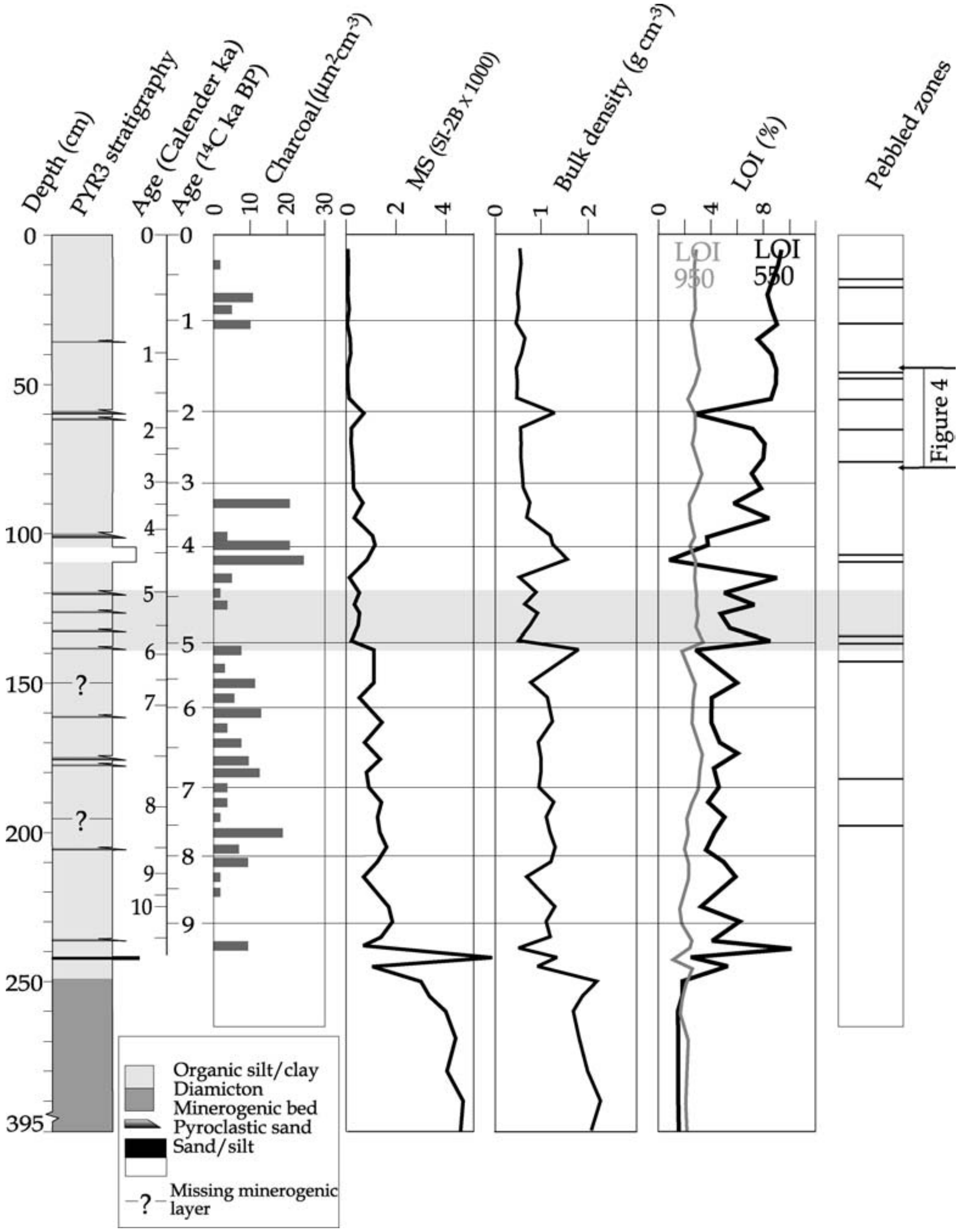

FIGURE 5. Stratigraphy, age, magnetic susceptibility (MS), bulk density, loss on ignition, charcoal, and location of pebble zones in sediment core PYR3. Shaded bar represents time of frequent deposition of minerogenic layers, and horizontal lines are 1000-^{14}C-yr intervals.

cm^{-2} yr^{-1}. Pine and alder pollen dominate this zone, averaging 28% and 22%, respectively. White spruce averages 5%, less than in zone 2; birch averages 5%. Fir pollen first appears at the base of the zone and averages 3%. Aspen and willow average 2%; larch averages <1%.

Zone 4 (100 cm to the top, 3800 B.P. to present) is designated the "pine-alder-alpine fir" zone. Pollen influx is high, ranging from 17,570 to 45,100 (average 34,450) grains cm^{-2} yr^{-1}. Pine averages 52%, and alder averages 10%. White spruce declines significantly from zone 3 (average 3%). Birch and aspen average 5% and 2%, respectively. Willow and larch average 1% and 2%, respectively. Fir pollen is present throughout, averaging 3%.

MACROFOSSIL ANALYSIS

Fir needles first appear about 9400 B.P. (base of zone 2) and are present until about 1200 B.P. (middle of zone 4). No other conifer needles were found in the sediment. The highest concentration of needles occurs at the top of zone 2 from about 5900 to 6600 B.P., with three full needles and two needle fragments within about 20 cm. The longest section without needles is at the top of the sequence from about 1200 B.P. to present.

Interpretation and Discussion

VEGETATION PROPERTIES

The very small amounts of pollen within zone 1 are likely not representative of local vegetation. Rather, the pollen probably was transported from afar or reworked from late Wisconsinan or older sediment. The paleoenvironmental significance of pollen percentages within zone 1 cannot be inferred reliably.

The beginning of zone 2 records colonization of the site by trees, shrubs, and herbs. Upon deglaciation before 9700 B.P., grass and sedge rapidly colonized the site. High percentages of sedge pollen and fern and moss spores indicate that moss and sedge probably grew around the lake.

Alder pollen increases in the lower part of zone 2, which is not surprising given that alder is a colonizer of nitrogen-poor, barren terrain (Hosie, 1990). Willow was established near the shore of the lake by about 9000 B.P. Decreasing percentages in nonarboreal pollen through zone 2 reflect the relative increase in pine pollen.

Moderate spruce pollen concentrations in zone 2 suggest that spruce was likely near the site by about 8200 B.P. Spooner et al. (2002) propose that spruce was established about 200 km to the south at Skinny Lake (10 in Fig. 1) by 8200 B.P., and Spooner et al. (1997) show that mixed spruce and fir forests were present about 150 km to the southwest at Susie Lake (8 in Fig. 1) as early as 7800 B.P.

Relatively low percentages of pine pollen in zone 2 suggest that pine was not established in the area until after 5600 B.P. This conclusion is consistent with findings by MacDonald and Cwynar (1991), who propose that pine first reached 56°N in western Alberta at 5100 B.P. (Lone Fox Lake; 15 in Fig. 1), although Gottesfeld et al. (1991) record the presence of pine at 55°N in British Columbia (Seeley Lake; 13 in Fig. 1) as early as 9200 B.P. The marked declines in pine and spruce at ca. 5800 and 7400 B.P. correspond to elevated levels of charcoal, suggesting that forest fires may have been common at these times and removed arboreal vegetation throughout the region. These declines are followed by rapid increases in pine pollen, implying that the pine is lodgepole pine (*Pinus contorta* spp. *latifolia*), a seral species. MacDonald (1987) also records an increase in fires in northern Alberta (Lac Ciel Blanc; 14 in Fig. 1) from 8000 to 6000 B.P., and Gottesfeld et al. (1991) document increased fire frequency at Seeley Lake at about 6000 B.P.

Fir pollen is absent in zone 2, yet fir needles were found throughout the zone, suggesting that fir was growing above the lake as early as 9400 B.P. The lack of fir pollen throughout zone 2 most likely results from delays between the first arrival of the plant at the site and a population large enough to be detectable in the pollen record (cf. Birks, 1989). Fir pollen is first noted at the beginning of zone 3.

The moderate increase in pine in zone 3 suggests that pine was likely growing in the region by ca. 5400 B.P. An increase in charcoal at the top of zone 3 implies that there were fires near the drainage basin around 4300 B.P. The general decline in sedge pollen and fern and moss spores in zone 3 may reflect the expansion of alder and willow.

The decrease in white spruce pollen from zone 3 to zone 4 suggests that white spruce may have retreated downvalley from the lake and/or became less extensive in the region. This reduction, which begins at 4200 B.P., is consistent with the disappearance of spruce macrofossils at 4000 B.P. at Susie Lake (Spooner et al., 1997).

Moderately high concentrations of pine pollen in zone 4 suggest that pine was well established near the site by ca. 3700 B.P. The increase in pine pollen is likely associated with the northerly migration of lodgepole pine documented by MacDonald and Cwynar (1985, 1991), who inferred that this taxa reached a similar latitude to the east (Lac Ciel Blanc; 14 in Fig. 1) by ca. 3000 B.P. and Susie Lake (Spooner et al., 1997) by 4000 B.P.

The presence of western hemlock pollen from about 1500 B.P. to the present is not likely evidence for its growing near the site, as concentrations are low and hemlock is not present near the site today. This pollen was probably transported by wind from the west and may indicate higher pollen productivity or increased incursion of Pacific air into the northwestern Cordilleran interior, or both. Increases in western hemlock pollen at other interior sites have been noted by Friesen (1985), Gottesfeld et al. (1991), Cwynar (1993), Spear and Cwynar (1997) and Spooner et al. (1997, 2002). Pyramid Lake is farther inland than these sites, attesting to the scale of exotic western hemlock pollen transport by Pacific air masses.

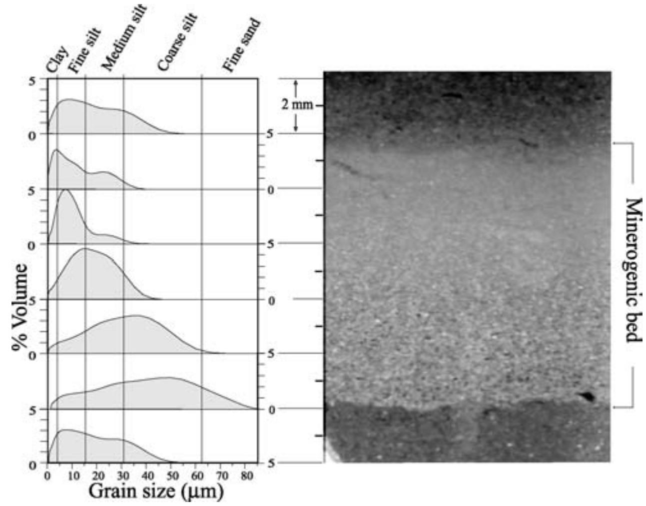

FIGURE 6. Thin section of minerogenic bed B (Fig. 4) and grain-size distributions at 2-mm intervals. Bright white spots are air bubbles in the epoxy resin.

PHYSICAL PROPERTIES

The fifteen thin minerogenic layers from core PYR3 record terrestrial erosional events that delivered coarse sediment to the lake. There are three processes that could have repeatedly transported terrestrial sediment to Pyramid Lake: removal of hill-slope vegetation by fire or disease, earthquakes, or large rainstorms.

Fire may remove hill-slope vegetation, resulting in slope failure or slopewash transporting sediment into a lake (Clark, 1988; Millspaugh and Whitlock, 1995). However, charcoal fragments were not seen in thin sections of the minerogenic layers, and charcoal concentration does not increase below or within these layers, nor does it decrease when frequency of minerogenic layers is highest (Fig. 5). Pollen data suggest that extensive deforestation occurred within the region at 5800 and 7800 B.P., likely due to drought-related fires. These fires may have removed vegetation within the drainage basin, although minerogenic layers do not occur at 5800 or 7800 B.P. Removal of slope vegetation due to disease likely did not occur within the basin because alder and birch pollen is present throughout the record. Alder and birch pollen are not transported long distances (Moore and Webb, 1978) and hence must have been continuously present within the basin since the early Holocene. Disease may have occurred for periods less than the resolution of the pollen record, but root structures would have likely remained during these periods, resulting in minimal slope destabilization. In summary, it is unlikely that the minerogenic layers are the result of fire or disease.

Blais et al. (1997) identify earthquake-triggered, subaqueous debris flow deposits within sediments of Saanich Inlet, British Columbia. The minerogenic laminae and beds within Pyramid Lake are unlike the earthquake deposits of Blais et al. (1997), as they are not massive and have high MS levels that preclude autochthonous origin. Minerogenic layers are also too low in organic carbon to be comparable to the resuspended gyttja deposits described by Doig (1991). Earthquakes have the potential to trigger terrestrial slump or slide sediment directly into the lake, but it is unlikely that the minerogenic layers originate from such mass movements as the grain size at the bottom of the minerogenic layers is consistently too fine and homogeneous. However, the sand bed at 105–110 cm (PYR3) may be the product of such a mass movement.

The most likely process to repeatedly supply terrestrial sediments to Pyramid Lake is slope erosion triggered by large rainstorms. Specifically, we propose that sediment was eroded from the gullies on the east slope and transported to the lake by surface runoff. It may have reached the center of the lake by flowing out on lake ice (Luckman, 1975) or in turbidity currents (Weirich, 1986). Thinning of the minerogenic layers toward the south end of the lake suggests that the gullies on the east slope, rather than the stream, were the likely sources of sediment.

The transport of sediment by surface overflow requires that a threshold in the water infiltration capacity of the hill-slope sediment be crossed, which may be accomplished through high levels of rainfall over a short period. The Pyramid Lake record may provide a proxy for the frequency and timing of such high-intensity rainstorms. However, due to the absence of rainfall records for the period of minerogenic bed deposition, we can only speculate on the magnitudes of such storms. From 1954 to 1999, the maximum precipitation recorded at Dease Lake (55 km SW of the study site) was 46 mm, and at Cassiar (40 km N) 85 mm (Environment Canada, 2000). A minerogenic bed was not deposited during this rainstorm; hence, we suggest the minerogenic deposits were produced by larger daily precipitation events.

Four minerogenic layers occur between ca. 4400 and 5100 B.P. (Fig. 5). This time frame is within the "mesothermic interval" of increased effective moisture at the end of the hypsithermal proposed by Hebda (1995). Pollen-based studies that record warmer and wetter climate than present conditions during this period include Miller and Anderson (1974), Cwynar (1988), and Gottesfeld et al. (1991). Cwynar (1993) suggests that high levels of exotic hemlock pollen at Waterdevil Lake prior to 4500 B.P. may be the result of increased Pacific air at the site. More frequent inland incursion of moist Pacific air, combined with warm hypsithermal temperatures, may have amplified the magnitude and frequency of orographic rainstorm activity at Pyramid Lake.

Although Brown et al. (2000) and Eden and Page (1998) relate storm magnitude to minerogenic layer thickness, we attempt no such correlation as thickness is a product of not simply storm magnitude but also storm intensity and duration, nature of the land surface erodability, lacustrine conditions, and sediment availability. The ability of a rainstorm to exceed the infiltration capacity of a slope is controlled by the intensity of the storm and the degree of slope saturation prior to the storm. High levels of saturation achieved by abundant precipitation prior to peak storm intensity or latent soil moisture from previous precipitation may allow less intense storms to exceed threshold levels.

The time of the rainstorm relative to summer active-layer development also affects the impact of the storm on the slope. Erosion would be minimal if there is no active layer. A thin active layer would be highly susceptible to erosion, whereas a thick active layer may require substantially larger quantities of precipitation before saturation occurs.

Missing minerogenic layers and different thicknesses of layers among the three cores are likely the result of lacustrine processes unevenly distributing sediment within the lake. Well-defined lake stratification may result in even dispersion of sediments through interflow. Conversely, lack of lake stratification may result in turbidity currents traveling as narrow plumes, and, depending on the entry location(s), sediment deposition may be restricted to a limited portion of the lake floor. The presence of ice cover would result in sediment accumulation on ice. The sediment would subsequently be rafted and rained out elsewhere, but likely not evenly (Luckman, 1975).

Sediment availability is determined by environmental conditions and length of time since the last major storm event. Periglacial processes, such as small-scale gelifluction, may increase the amount of sediment available for transport by deepening the active layer into new sources of fine-grained sediment (Carson and Kirkby, 1972). Conversely, large-scale gelifluction or other mass wasting processes such as rock or snow avalanches may reduce sediment availability by transporting sediment to more stable locations. Frost shattering and chemical weathering reduce grain size, increasing susceptibility to transport by overflow. Antecedent earthquakes may, through ground shaking, increase the susceptibility of sediment to later storm erosion (Eden and Page, 1998). The length of time since the last sediment-transporting storm would likely be a major influence on the amount of sediment available for transport.

Rates of mass wasting and periglacial processes have varied throughout the Holocene, as temperature and moisture regimes have not been constant (Hebda, 1995). Increased temperatures during the hypsithermal likely reduced periglacial effects, whereas during the Little Ice Age, these effects were likely pronounced. As indicated by the pollen record, changes in vegetation types and density have occurred that would affect the water absorption capacity of the east slope above the lake.

Conclusions

The Pyramid Lake record illustrates the utility of tarn sediments as a proxy for estimating past rainstorm frequency, timberline migration, and climate change. This study focused on obtaining a better understanding of the variability in Holocene climate in the Cassiar Mountain region of the northwestern Cordillera. Sediments from Pyramid Lake provide a record of landscape response to extreme rainfall events since deglaciation before 9700 ± 200 B.P. Graded minerogenic layers record deposition of eroded hill-slope sediment during high-intensity rainstorms. Correlation of minerogenic layer thickness with paleostorm magnitude is problematic due to uncertainties in sediment availability, storm characteristics, and lacustrine processes.

Krummholz subalpine fir has been present above Pyramid Lake since at least 9400 B.P., but mature forest did not reach the elevation of the lake at any time during the Holocene. Climate within the area was drier than at present from 5800 to 7400 B.P. High-magnitude rainstorms may have been more frequent from ca. 4400 to 5100 B.P., possibly as a result of increased summer temperatures. An elevated spruce habitat during that time supports this idea. An increase in moisture likely occurred throughout the region from about 1000 to 3200 B.P., and significant changes in regional air-mass circulation may have occurred at about 1500 B.P.

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