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A paleoclimate record for the past 250,000 years from Summer Lake, Oregon, USA: II. Sedimentology, paleontology and geochemistry

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Abstract

We have obtained a detailed paleoenvironmental record in the Summer Lake Basin, Oregon (northwestern Great Basin, US) spanning from 250 ka–5 ka. This record is derived from core and outcrop sites extending from a proximal deltaic setting to near the modern depocenter. Lithostratigraphic, paleontologic (ostracodes and pollen) and geochemical indicators all provide evidence for hydroclimate and climate change over the study interval.

Lithostratigraphic analysis of the Summer Lake deposits allows subdivision into a series of unconformity – or paraconformity-bound lithosomes. The unconformity and facies histories indicate that the lake underwent several major lake-level excursions through the Middle and Late Pleistocene. High stands occurred between ~200 and ~165 ka, between ~89 and 50 ka and between ~25 and 13 ka. Uppermost Pleistocene and Holocene sediments have been removed by deflation of the basin, with the exception of a thin veneer of late Holocene sediment. These high stands correspond closely with Marine Oxygen Isotope Stages 6, 4 and 2, within the margin of error associated with the Summer Lake age model. A major unconformity from ~158 ka until ~102 ka (duration varies between sites) interrupts the record at both core and outcrop sites.

Lake level fluctuations, in turn are closely linked with TOC and salinity fluctuations, such that periods of lake high stands correlate with periods of relatively low productivity, fresher water and increased water inflow/evaporation ratios. Paleotemperature estimates based on palynology and geochemistry (Mg/Ca ratios in ostracodes) indicate an overall decrease in temperature from ~236 ka–165 ka, with a brief interlude of warming and drying immediately after this (prior to the major unconformity). This temperature decrease was superimposed on higher frequency variations in temperature that are not evident in the sediments deposited during the past 100 ka. Indicators disagree about temperatures immediately following the unconformity (~102–95 ka), but most suggest warmer temperatures between ~100–89 ka, followed by a rapid and dramatic cooling event. Cooler conditions persisted throughout most of the remainder of the Pleistocene at Summer Lake, with the possible exception of brief warm intervals about 27–23 ka. Paleotemperature estimates for the proximal deltaic site are more erratic than for more distal sites, indicative of short term air temperature excursions that are buffered in deeper water.

Estimates of paleotemperature from Mg/Ca ratios are generally in good agreement with evidence from upland palynology. However, there is a significant discordance between the upland pollen record and lake indicators with respect to paleoprecipitation for some parts of the record. Several possibilities may explain this discordance. We favor a direct link between lake level and salinity fluctuations and climate change, but we also recognize the possibility that some of these hydroclimate changes in the Summer Lake record may have resulted from episodic drainage captures of the Chewaucan River between the Summer Lake and Lake Abert basins.

Introduction

High resolution paleoclimate records from western North America for the Late Pleistocene period prior to 50 ka previously have been difficult to obtain because of the combined effects of inadequate age control prior to the radiocarbon 'era' and a limited number of sites with continuous paleoclimate records. This has resulted in numerous attempts to fit poorly dated but climatically diagnostic deposits into a model chronology based on assumed correlation with the marine oxygen isotope record of glacial advance and retreat (Negrini, submitted). This approach may produce erroneous interpretations when the indicator records do not fit preconceptions of the timing of climate change. There is a need for climate chronologies in the Great Basin that do not rely on such assumptions, but rather can be dated directly.

The interval between 250–50 ka is of particular interest, because there is abundant evidence in western North America for glacial/interglacial and pluvial/interpluvial events spanning oxygen isotope stages 7–4 (e.g. Smith & Bischoff, 1993) . For the earlier part of this interval there are still relatively few dated records of any kind in the region (Negrini, submitted; Wigand & Rhode, submitted). The climate transitions surrounding oxygen isotope stage 5 (i.e. the last interglacial) remain poorly constrained, in part because many of the potential repository basins for sediments were deflated and/or eroded during this time period. Many more records are available for the Middle Wisconsin. This interval in the western US appears to have been punctuated by at least one and perhaps several episodes of cooler and wetter climates, and by several glacial advances (Fullerton, 1986; Richmond, 1986; Phillips et al., 1996; Bischoff et al., 1997; Menking et al., 1997), although detailed chronologies often remain controversial (Richmond, 1986; Berry, 1994). Climate records obtained from lake, groundwater and spring deposits from the central and southern Great Basin all support episodes of increased precipitation at various times between 80–60 ka (Smith et al., 1983; Bischoff et al., 1985; Oviatt et al., 1987; Winograd et al., 1992; Szabo et al., 1994; Quade et al., 1995, 1998). Drift deposits from the Puget Lowlands suggest glacial activity in that area ~80 ka (Easterbrook, 1986) and the ~70 ka Pumice Castle tephra set has been found in lacustrine sediments at three locations in the northwest Great Basin (Bradbury, 1991; Botkin & Carembelas, 1992a and b; Negrini, submitted). Studies from Carp Lake, located in the eastern Cascade Mountains, document climate change over the past 125 ka, or longer, in the northern Intermountain West (Whitlock & Bartlein, 1997), but the authors were compelled to base parts of their age model on assumed fits of vegetation change to the global oxygen isotope record. A record of the entire 250 ka time period in the northwestern Great Basin which uses a locally-derived age model has not been previously available.

Here we report a well-dated record of limnologic, vegetational and climatic change from Summer Lake, located in the south-central part of Oregon, at the northwestern edge of the Basin and Range (Figure 1). This record fills an important geographic and chronologic gap in the history of Pleistocene climate change following the last interglacial but prior to the extreme cooling of the late Wisconsin. We have used multiple lines of evidence (sedimentology, ostracodes, pollen, stable isotopes and trace elements) to obtain this paleoenvironmental record. Furthermore this record can be constrained chronologically at a relatively high level of resolution (~2 ka) based on absolute age dates, interpolations between tephra correlations, magnetic susceptibility and paleomagnetic stratigraphy.

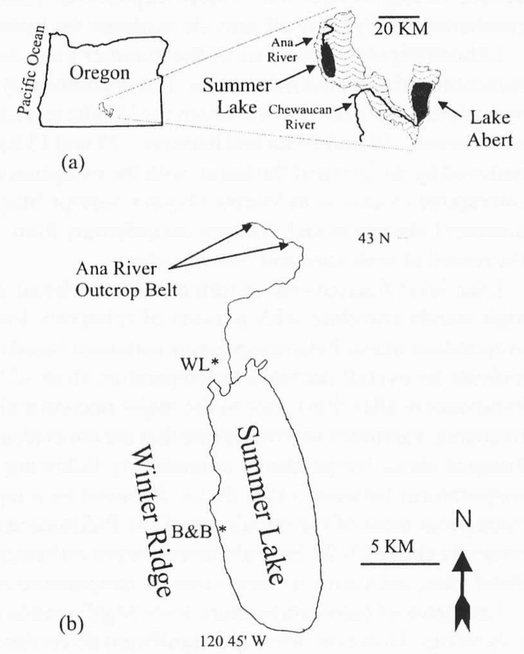


Figure 1. Location map of modern Summer Lake, Oregon. 1a) The maximum highstand of Late Wisconsin pluvial Lake Chewaucan is shown in the wave-shaded pattern on the intermediate scale inset. 1b) Relative locations of the 3 study sites: Ana River Canyon outcrop area, Wetlands Levee (WL) core site and the Bed & Breakfast (B&B) core site.

This paper concentrates on the local record from the Summer Lake area. Regional paleoclimate indicator comparisons and broader paleoclimatic significance are discussed at length elsewhere (Negrini, submitted; Wigand & Rhode, submitted).

Study area

Summer Lake is one of the northernmost remnants of the numerous pluvial lakes that occupied much of the Great Basin during the late Pleistocene. It lies on a physiographic transition zone between the Great Basin to the south and east and the Cascade Volcanic Province to the northwest. The surface elevation of Summer Lake at present averages 1264 meters above sea level (masl).

The Summer Lake area today is characterized by a semi-arid climate. Paisley, Oregon, a town close to the south end of Summer Lake, has recorded a mean (30 yr. record) annual precipitation of 273 mm, most of which falls during the winter months. Temperatures over this same time interval (1951–1980) range from a January mean of -1 °C (range: -6 ° min. to +5 ° max) to a July mean of 20 °C (range: +10 ° min. to +30 ° max.). Fronts derived from westerlies off the North Pacific Ocean are primarily responsible for the relatively mild temperatures and winter precipitation in this area. Less commonly, continental air masses from the north and east produce very cold winter or very hot summer conditions. Potential evapotranspiration exceeds precipitation in the Summer Lake area by approximately 710 mm/yr (Oregon Climate Service, 1993).

The vegetation in the Summer Lake area today, as in other portions of the northern Great Basin, consists of several elevationally stratified plant communities (Billings, 1951; Faegri, 1966). In the immediate vicinity of the lake are desert scrub plant communities dominated by salt-tolerant desert shrubs, notably the shadscale (*Atriplex confertifolia*) and black greasewood (*Sarcobatus vermiculatus*). A lower sagebrush zone dominated by big sagebrush (*Artemisia tridentata*) and rabbitbrush (*Chrysothamnus* spp.) lies to the north and west of Summer Lake above the desert shrub zone. Slightly higher patches of semi-arid woodland dominated by western juniper (*Juniperus occidentalis*) occasionally mixed with mountain mahogany (*Cercocarpos ledifolius*) is restricted to small stands along the east side of Winter Ridge. Western juniper (*Juniperus occidentalis*) also occurs commonly along the Abert Rim, east of Lake Abert and approximately 40 km southeast of Summer Lake. Above this lies a mixed conifer zone characterized by ponderosa pine (*Pinus ponderosa*) and douglas fir (*Pseudotsuga menziesii*), but includes grand fir (*Abies grandis*) and white pine (*Pinus monticola*) in more mesic localities, and white-bark pine (*Pinus albicaulis*) in the higher, rockier reaches of Winter Ridge. Quaking aspen (*Populus tremuloides*) is abundant in areas where seeps are common. This forest dominates the Winter Ridge area, immediately to the west and south of, and above Summer Lake. Lodgepole pine (*Pinus contorta*) is very common in the forests immediately to the west of Winter Ridge. Scattered areas of shallow, stony soils, especially near the top of Winter Ridge, are dominated by low sagebrush (*Artemisia arbuscula*), arrow-leafed balsamroot (*Balsamorhiza sagittata*), various biscuitroots (*Lomatium* spp.), and sego lilies (*Calochortus* spp.). Finally, seeps and small lakes in the Winter Ridge area are characterized by a variety of aquatic plants including wide-leafed cat-tail (*Typha latifolia*), sedges, water smart weeds (*Polygonum* spp.), and pond weeds (*Potamogeton* spp.).

Modern Summer Lake is a shallow ($z_{avg} < 2$m), closed basin lake with an area of approximately 180km^2. Variation in modern surface area is controlled both by seasonal precipitation and waterflow control behind artificially-maintained levees. The modern lake occupies a N-S trending half-graben within the northwestern corner of the Basin and Range Province. The lake is fed by several small springs and snow-melt streams, most notably the Ana River. The Chewaucan River enters the combined Summer Lake/Lake Abert Basin via the Chewaucan marshes. At present this river flows south upon entering this marsh area to Lake Abert, but has entered the Summer Lake sub-basin earlier in the Pleistocene as a result of delta aggradation (Allison, 1982; Davis, 1985) (Figure 1).

Summer Lake is highly alkaline and saline (pH 9.5–9.6, 40–50,000 mg l^{-1} Total Dissolved Solids) and dominated by Na^++ Cl^- + HCO_3^- + CO_3^{-2} + K^+ + SO_4^{-2} (Palacios-Fest et al., 1993). Like most saline-alkaline lakes, it is depleted in Ca^{+2} and Mg^{+2}. Lake chemistry in Summer Lake varies considerably both seasonally and yearly, as a result of both short-term fluctuations in precipitation and seasonal water extraction from the Ana River for agricultural and waterfowl management purposes.

Summer Lake and its neighbor Lake Abert are Holocene relicts of the much larger Pleistocene pluvial Lake Chewaucan (Allison, 1945, 1982: Antevs, 1948) (Figure 1). Lake Chewaucan at its greatest extent (~1250km^2) occupied several adjacent structural basins. The highest Late Pleistocene shorelines (1380 masl) indicate that the lake reached a depth of approximately 115 m (Allison,1982).

Prior investigations of the geology and paleolimnology of Lake Chewaucan have mostly been focused on the shoreline terraces surrounding the lake (Allison, 1982), and a series of lake-margin outcrops exposed along the Ana River Canyon, north of Summer Lake (Conrad, 1953; Allison, 1982; Davis, 1985; Negrini et al., 1988; Negrini & Davis, 1992; Freidel, 1993; Palacios-Fest et al., 1993; Negrini et al., 1994). Over the past 10 yrs, a highly detailed geochronology has been constructed for the exposed Summer Lake deposits, mainly using tephrochronology (both directly dated and correlated tephras) and paleomagnetic secular variation records (Davis, 1985; Negrini et al., 1988, 1994; Berger, 1991; Negrini & Davis, 1992). Within this chronologic framework, a general history of the lake was reconstructed from ostracode and geochemical data by Palacios-Fest et al. (1993). Unfortunately, both the Ana River Canyon and the terrace exposures represent condensed sections with significant unconformities that preclude paleolake (and therefore paleoclimate) reconstructions from parts of these records, particularly for the low-stand intervals of the lake's history.

In 1986, the US Geological Survey and the US Forest Service retrieved several cores from the margins of Summer Lake, Lake Abert and the Chewaucan Marshes (D. Adam, pers. comm. to D.E., 1991). We decided to obtain long cores from more central parts of Summer Lake basin, where stratigraphic condensation should be minimal. Here we present results from these previously undescribed cores, and compare these results with prior work at the Ana River Canyon.

Our primary record is from the Wetlands Levee Site, located ~10 km. south of the Ana River Canyon site, and ~10 km to the north of the modern Summer Lake depocenter. An additional core was collected at the Bed & Breakfast Site, located close to both the basin's eastern margin and its modern depocenter (3 km west of the core site), and ~20 km south of the Ana River Canyon outcrops (Figure 1). Both of these sites would be underwater today in the absence of drainage diversion and levee construction. Here we concentrate on comparisons between the core and outcrop record from the various study sites and their significance for regional paleoclimate over the past 250 ka. A companion paper (Negrini et al., 2000) presents paleomagnetic and tephrochronologic data used to develop the age model upon which the interpretations presented here are based.

Methods

Our samples come from both outcrop and core localities, requiring slightly different sampling procedures for each. Both outcrop and cores were studied with a variety of proxy records and correlation tools, including granulometry, sediment composition, fossil pollen and ostracode stratigraphy, trace elements and stable isotopes, secular magnetic variation and sediment magnetic properties. We collected outcrop samples in 1987 and 1989 from the Ana River Canyon, at Davis' (1985) section localities C, D, E, F, and G. Details of drilling operations are given in Negrini et al. (2000). Outcrop sampling methodologies for ostracodes and paleomagnetic samples are given in Negrini et al. (1988) and Palacios-Fest et al. (1993).

We chose coring sites which provided access for the drilling rig to the middle portions of Summer Lake, along artificial levee or wetland maintenance roads. We cored at two sites; Ana River Canyon Section Locality C, at 1280masl (to duplicate outcrop observations), and Wetlands Levee, near the northern margin of modern Summer Lake at the west end of Gold Dike (1265masl). A ~30 m composite core was extracted from the Wetlands Levee site. An additional coring site (the 'Bed & Breakfast' locality) was sampled using a modified Livingston piston corer. This core penetrated ~12 m of sediments.

Following core length measurement, we split and field-logged each core segment for color, texture, ash occurrence and presence of bedding features, including unconformities. Approximately 50 gm of sediment was sampled every 20 cm (sample thickness ~2 cm) for ostracodes and 2.3 cc of sediment was sampled every 5 cm (sample height ~1cm) for pollen. In the lab, we subsampled for granulometry, organic and inorganic carbon content, tephra content, sediment magnetic properties and paleomagnetism. Details of sampling, analytical, correlation and age modelling procedures for tephras and paleomagnetism are given in Negrini et al. (2000).

Granulometry samples for cores were collected at 20–40 cm intervals and analyzed using a CILAS 715 laser particle size analyzer at the University of California-Davis. Approximately 0.3 gm of each sample was wet sieved to 188 μm and placed in the analyzer's suspension liquid circulation system. Ana River Canyon outcrop sample granulometry was done on samples prepared for ostracode analysis, using a single100 μm wet sieve to separate clay+silt+fine sand from coarser fractions.

Total organic carbon (TOC) was estimated by loss on ignition of dried sediment samples in a muffle furnace at 600 °C. Percent calcium carbonate was measured in 2 ways. For Ana River Canyon samples,% $CaCO_3$ was estimated by sample ignition at 925 °C. $CaCO_3$ from Wetlands Levee and B&B core samples was measured using the CO_2 Evolution-Coulometric Titration method (ASTM, 1982) at the US Geological Survey (Menlo Park).

Fossil ostracodes were processed from 54 samples from the Wetlands Levee Core and 76 from the Ana River Canyon outcrops. Sample preparation and data analytical methods follow those described in Palacios-Fest et al. (1993). Fossil ostracode shells were also analyzed for minor element and stable isotopic composition as indicators of lake hydroclimate, again following Palacios-Fest et al. (1993) for methods and instrumentation. To avoid complications of vital effects and variable trace element uptake among juveniles vs. adults, only adult valves of *Limnocythere ceriotuberosa* (a common ostracode in the Summer Lake fossil record) were used in the elemental ratio (Mg/Ca and Sr/Ca) and stable isotope ($\delta^{18}O$ and $\delta^{13}C$) studies reported here. We analyzed 162 individual valves from 34 stratigraphic horizons from the Ana River Canyon locality and 220 valves from 44 stratigraphic horizons from the Wetlands Levee Core for minor element (Mg and Sr) content. Stratigraphic spacing between minor element samples ranged from ~10–40 cm, with 2–5 replicate samples per horizon.

We obtained $\delta^{18}O$ and $\delta^{13}C$ results on ostracodes from 16 stratigraphic horizons from the Ana River Canyon section and 35 horizons from the Wetlands Levee core. As a result of the small size of the ostracodes in many of these samples, it was necessary to combine 3–5 valves for each sample. Stratigraphic spacing for stable isotope samples was coarser than for the Ca, Mg and Sr analyses, on average every 50 cm for Ana River Canyon and every 35 cm for the Wetlands Levee core.

One hundred and forty-six samples were processed for pollen from the Summer Lake Basin (68 from Ana River Canyon, 25 from Bed and Breakfast Locality, and 53 from Wetlands Levee). Processing of samples for pollen, spores and acid-resistant algae generally followed Mehringer (1967). The number of grains counted for each sample was determined by either the degree of preservation and/or the abundance of pine. All samples were counted until enough grains were counted to provide an adequate statistical estimate of the pollen in each. Raw counts of pine were converted to relative percentages of total terrestrial pollen, whereas other terrestrial pollen percentages were calculated using total terrestrial pollen minus pine. Percentages of aquatic pollen types were calculated relative to total pollen, whereas percentages of spore types were calculated relative to total spores. Algal percentages are based upon the total algae present in the sample. Ratios generated from abundant pollen types were used for further interpretation of the data, as were pollen zones generated using a constrained single-linked dendrogram program (Birks & Gordon, 1985).

Results and interpretation

Lithostratigraphy

Summer Lake and Lake Chewaucan sediments are dominated by laminated to massive ostracodal silts interspersed with thin to thick-bedded, rhyolitic and basaltic tephras (Figure 2). Some sediments are sandy, diatomaceous or oolitic. Details of tephrochronology and extrabasinal correlation are presented in Davis (1985), Negrini et al. (1994), and Negrini et al. (2000).

Sediment accumulation rates for the 3 study sites increase basin-ward, in common with other lake systems. The Ana River Canyon site apparently has been a lake-marginal site for all of its recorded history, and retains only a highly condensed stratigraphic section with numerous depositional hiatuses during low lake stands. In contrast, the Bed & Breakfast site, and to a lesser extent the Wetlands Levee site, have been sites of sediment focusing and probable enhanced depositional rates during low lake stands (Lithosomes II, III, VI and VII, discussed below).

The oldest sediments recovered to date are found in the lower part of the Ana River Canyon section, and are unrepresented in the more distal cores. These ostracodal muds and silty sands range in age from about 250–174 ka, although some aspects of their dating are problematic (Negrini et al., 2000). Eight unconformity-bounded lithosomes can be recognized and correlated between the Ana River Canyon section and the Wetlands Levee core (Figure 2). In some cases an unconformity is not evident in the Wetlands Levee core and lithologic criteria indicating shallower water conditions are used to identify correlative conformities. The upper 3 lithosomes can also be recognized in the Bed & Breakfast core. From oldest to youngest these include:

Lithosome I (~174–158 ka).
This unit comprises the section above a basal unconformity (expressed as a gravel lag at Ana River

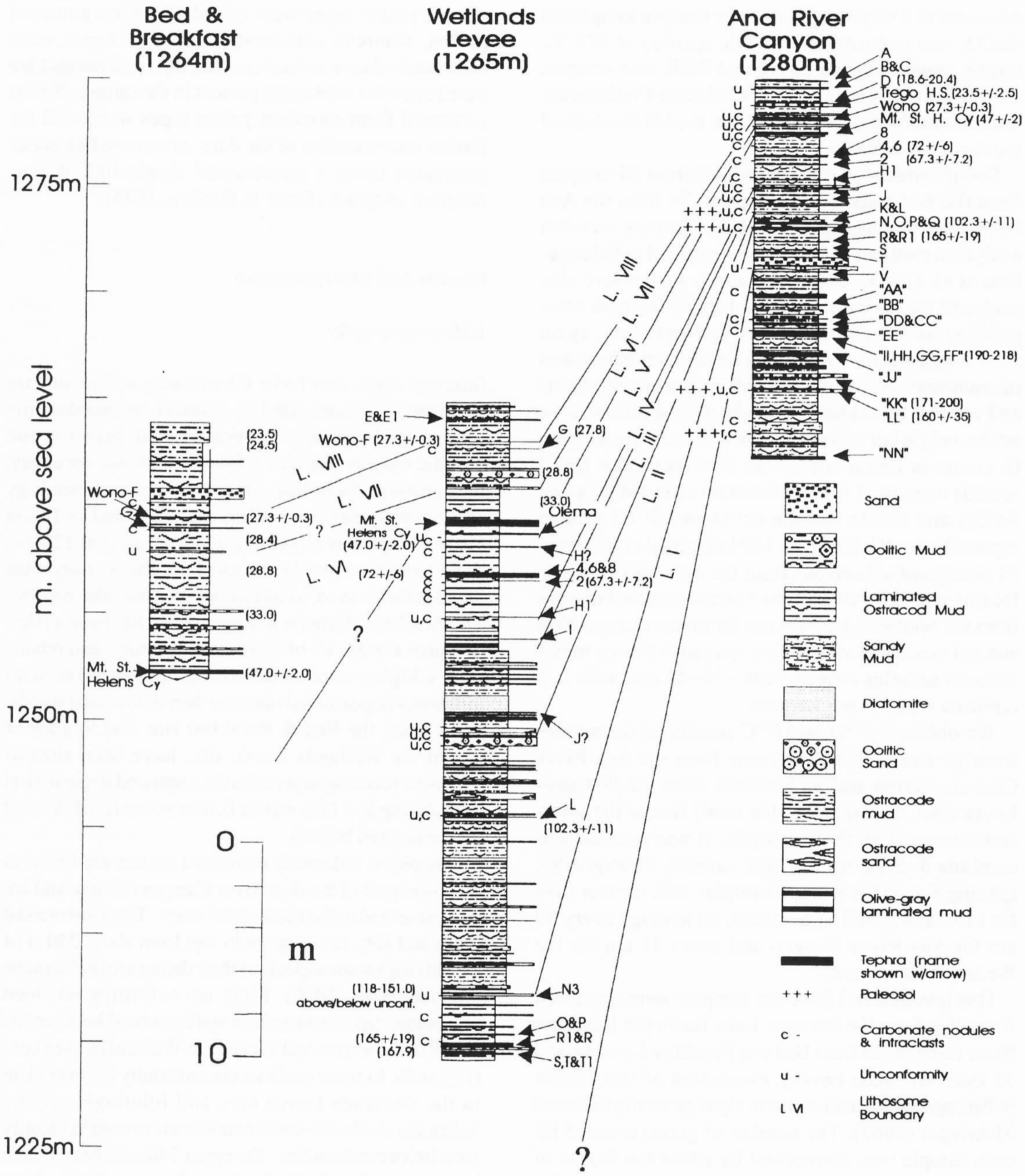


Figure 2. Stratigraphy of the 3 study sites. Key tephras for intra- and interbasinal correlations are labeled by their local letter designation or (when more widely recognized) by their common name. Major lithosome boundaries are shown as tie lines between sections and the lithosomes themselves are indicated L. I, L.II, etc.. Radiometric, tephrochronologic or paleomagnetic age constraints are given in brackets.

Canyon) directly above Summer Lake Tephra 'V (~174 ka)'. The basal unconformity of this lithosome, however, is only observed in the Ana River Canyon outcrops. Tephra R, from the Ana River Canyon, has been TL-dated at 165+/-19 ka (Berger, 1991). Although this tephra has not been identified at Wetlands Levee, Tephra S (which lies ~50cm below R at Ana River Canyon) occurs at both localities, suggesting that the correlative horizon to Tephra R at Wetlands Levee is approximately 1m above the base of the core. At Wetlands Levee the upper 3m of this lithosome was penetrated by the core. At Ana River Canyon the lower 0.5m of Lithosome I is sandy, otherwise at both Ana River Canyon and in the Wetlands Levee core it consists of thinly-bedded or laminated olive grey muds. The lithofacies in this unit suggests deepening water towards the top of the lithosome at Wetlands Levee, with the upper portion of the Wetlands Levee section having been deposited below storm wave base. However, magnetic susceptibility indicators of shallower water conditions exist in the uppermost portion of the Ana River Canyon section (between~165–158 ka).

Lithosome II (~118–95 ka).

This unit comprises the section above a major unconformity recognized by Allison (1945) and Davis (1985). In the Ana River Canyon outcrop it is marked by a 10 cm-thick, ostracode-rich horizon that includes flat carbonate intraclasts and some reworked gastropod shells. In the Wetlands Levee core it is marked by an abrupt color transition in overlying and underlying muds, separated by a 2–3 cm ostracode/tephra sand, that also contains reworked gastropod shells. The basal unconformity predates Tephra 'N' (TL dated at 102.3+/-11 ka, Berger et al., 1990) and postdates the 165+/-19 ka age for Tephra 'R'. Based on extrapolated sediment accumulation rates above and below the unconformity, Negrini et al. (2000) estimate its duration as approximately 151–118 ka at Wetlands Levee (i.e. placing the base of Lithosome II there at ~118 ka). At Ana River Canyon, Section C Negrini et al. (2000) estimate a somewhat longer duration for the unconformity (162–102 ka). However the erosion surface is locally irregular, and at the nearby Ana River Canyon Section D, younger sediments that have not been stripped away by subsequent erosion during the unconformity interval probably date to about 158 ka. Elsewhere, Davis (1985), Berger (1991), Palacios-Fest et al. (1993) and Negrini (submitted) have argued that this major unconformity represents an extended period of desiccation and probable deflation associated with Marine Oxygen Isotope Stage 5, and new data in this paper support that correlation. Lithosome II comprises a thin (~60cm) sequence of sandy muds containing a distinct vesicular 'A' soil horizon at the top. This soil consists of a leached zone of coarse material, primarily ostracodal sands, with millimeter-scale vertical voids and a vesicular texture. At Wetlands Levee, Lithosome II is a much thicker (12m), slightly coarsening-upwards sequence, comprising a basal, laminated mud, giving way to a thick sequence of ostracodal muds, and finally a ~2m sequence of oolitic muds. The presence of soil horizons, reworked shells and intraclasts, and thin sandy horizons all support a shallow water environment of deposition for Lithosome II.

Lithosome III (~95–88.9 ka).

This unit lies above the prominent Ana River Canyon 'A' horizon soil, dated at 95 ka by interpolation between the ages of Tephras N and 2/6. At Ana River Canyon it consists of ~60cm thick sequence of ostracodal muds that become sandier towards the top. In the Wetlands Levee core this unit is considerably thicker (~3.5m), and is composed of interbedded ostracodal, diatomaceous and sandy muds. The common occurrence of sandy laminae and oolitic beds suggests that this unit formed in shallow water.

Lithosome IV (~88.9–84 ka)

The base of this unit at the Ana River Canyon is a carbonate-rich sandy interval, that directly overlays Tephra J. Neither the unconformity surface nor Tephra J are present in the Wetlands Levee core. We suspect that the tephra was either lost in a missing segment between coring drives, or the unconformity is not present here. Interpolation between TL and K/Ar dated horizons from the Ana River Canyon outcrop suggests an age of 89,000 yr for the base of this lithosome. At both Ana River Canyon and Wetlands Levee this lithosome is marked by an absence of the sandy laminae and oolitic horizons common in Lithosomes II and III. At Ana River Canyon Lithosome IV consists of a thin (~50 cm) sequence of laminated ostracodal muds. Our tentative placement of the base of this lithosome in the Wetlands Levee core suggests it is ~1.3m thick here, and also comprises laminated ostracodal muds. We interpret Lithosome IV to have accumulated in deeper water, below storm wave base.

Lithosome V (~84-47 ka)

The base of this unit is marked in the Ana River Canyon outcrop by a carbonate pebble conglomerate lying below Tephra H1. The conglomerate horizon can be correlated with an unconsolidated lime mud that lies between Tephras H1 and I in the Wetlands Levee core. Interpolation between bracketing TL-dated horizons suggests an age of ~84 ka for this boundary. The Ana River Canyon Lithosome V sequence is ~2.5 m thick, increasing to ~4 m at Wetlands Levee. In both areas it has a similar lithology to Lithosome IV, consisting of laminated to massive ostracodal muds with occasional micritic interbeds, but again never with sub-aerial exposure features. It appears to have been deposited in deep water conditions, below storm wave base.

Lithosome VI (~47ka–30,900^{14}C yr B.P.)
The base of this unit comprises a pebble-rich sandy paraconformity with abundant carbonate that lies directly below Summer Lake Tephra 12 (= Mount St. Helens Tephra Cy) dated at 47+/-2 ka). Tephra WL-A-7-2, which occurs at the paraconformity surface in the Wetlands Levee core, can be correlated with the Olema Ash (Sarna-Wojcicki et al., 1988), and has an estimated age of 55–50 ka, based on its position relative to the Pumice Castle and Mt. St. Helens Cy tephra layers. (Negrini et al., 2000). The basal pebbly horizon is not present in the Bed & Breakfast core, which bottomed out at Summer Lake Tephra 12. Lithosome VI increases substantially in thickness between the three sections, going from ~50cm at the Ana River Canyon, to ~2.2 m at Wetlands Levee, to at least 6m at the Bed & Breakfast site. At all three localities Lithosome VI consists of light to dark grey massive ostracodal muds. At Wetlands Levee the Lithosome changes into an oolitic mud near its top. The common occurrence of oolitic horizons and sand laminae suggest that Lithosome VI was deposited under relatively shallow water conditions, above storm wave base.

Lithosome VII (~30,900–27,300 ^{14}C yr B.P.)
The base of this unit is a prominent, 2 cm-thick oolitic sand, traceable between all 3 sites. Based on paleomagnetic secular variation correlation between all 3 records and the Lake Russell record of Lund et al. (1988), the base of Lithosome VII can be dated at 30,900 ^{14}C yr B.P.. As with Lithosome VI, this unit thickens distally away from the Ana River Canyon, going from ~50 cm at Ana River Canyon to 2.2 m at Wetlands Levee and 1.7 m at Bed & Breakfast. At all 3 localities this lithosome is dominated by massive ostracodal muds, with occasional sand laminae, suggestive of wave reworking. At Wetlands Levee, oolitic lenses are present in several horizons near the base of the lithosome. Overall the setting suggested by all the above features is shallow water.

Lithosome VIII. (~27,300 ^{14}C yr B.P.-Holocene?)
The base of this unit is marked by a lag deposit of the Wono Tephra (27,300 ^{14}C yr B.P, Benson et al., 1997). This lithosome comprises the composite outcrop section of Negrini & Davis (1992), including all of the youngest sediments at their Section E above the Wono Tephra. Several additional unconformities are evident in the Ana River Canyon section above the Wono, but are not expressed at the Wetlands Levee or Bed & Breakfast localities. At both Ana River Canyon and Wetlands Levee, Lithosome VIII is approximately 1.8m thick, increasing to 4.3 m at the Bed & Breakfast site. At all 3 sites Lithosome VIII is dominated by massive gray ostracodal muds. Sandier intervals are also evident both in the upper part of the section at Wetlands Levee and in the lower portion of the Bed & Breakfast core. The presence of abundant lag and sand laminae suggests that Lithosome VIII was deposited in relatively shallow water conditions. At all 3 sites sediments near the top of Lithosome VIII are late Pleistocene in age (~22–16 ka). Latest Pleistocene deposits were either never laid down or deflated prior to our sampling. The uppermost ~60–80 cm in the Wetlands Levee core is almost certainly of late Holocene age (i.e. post-5500 yr B.P.), based on the presence of warm climate indicators in the pollen, ostracode and geochemistry records, and the widespread occurrence of similar-aged lacustrine interludes in the intermountain west. However, there is no evident lithologic break indicative of an unconformity in the Wetlands Levee record.

We can reconstruct lake level history, based on the presence of depth-diagnostic indicators in the stratigraphic records from the 3 study sites, and earlier data on shoreline elevations for the latest Pleistocene (Allison, 1982; Negrini & Davis, 1992; Freidl, 1993; Negrini, submitted) (Figure 3). Exposure surfaces and paleosols indicate that the lake surface was below the outcrop or core sample elevation at the time of formation. Given the size and probable fetch of Lake Chewaucan we interpret wave reworking features, such as clean oolite, cross bedded sands or ostracode shell lags to indicate water depths of 0–2 m, bioturbated and massive ostracodal muds to indicate depths of approximately 2–10 m, and well-laminated, fine grained muds to indicate deposition in greater than about 10 m depth. Given the elevational offset between the various core and outcrop sites it is thereby possible to constrain lake depths for the sampling sites. Three intervals of high lake stands during the past ~190 ka are apparent in our records. An early high stand, corresponding to Lithosome I and sediments immediately underlying Lithosome I at Ana River Canyon, spans the interval ~190–165 ka, with indications of declining lake levels thereafter up to the unconformity. Although lake stands were certainly very low for some period (perhaps most?) represented by the unconformity, it is possible that the lake was high for part of this interval as well.

A second high stand interval, corresponding to Lithosomes IV and V (~89–47 ka), had comparable lake surface elevations with the earlier high stand. A third, latest Pleistocene high stand (~21–13 ka), is unrepresented in either the core or Ana River Canyon records, probably as a result of widespread Holocene deflation of the Summer Lake Basin. However, well preserved and dated terraces surrounding the basin and well above the Ana River Canyon outcrops (most of which are indicative of relatively shallow water) demonstrate that this Late Wisconsin lake (which appears to have actually been several pulses of high lake stands) was considerably deeper than the earlier high stand lakes.

Granulometry, organic and inorganic carbon

At the more proximal Ana River Canyon locality fine-grained sedimentation is associated with Lithosomes I, IV and V, all inferred to have been deposited at times of relatively high lake stands (Figure 4). Considerably sandier sediments were deposited here during Lithosome VI, a time of much lower lake level. For the remaining lithosomes textures are quite variable and may be more strongly related to variable depositional processes along the delta front, such as episodic flooding and gravity flow events, rather than lake level per se.

Textural variability as expressed in the silt to clay ratio at the distal deltaic Wetlands Levee site shows a long-term pattern superimposed on the higher frequency variability in grain size within and between lithosomes (Figures 4 and 17). Very low silt-clay ratios occur in the lowermost lithosome (Lithosome I). Above the major unconformity there is a long term trend from coarser to finer sediments, extending from Lithosome II to Lithosome VIII. These quantitative variations in grain size are not clearly diagnostic of short term lake level fluctuation, as with the qualitative changes in bedding character and sedimentary structures that define the lithosomes. Clays are generally diluted by lacustrine silts, which come from a variety of unrelated sources, including ostracode valve fragments, diatom frustules, silicic tephras and probably wind-derived dust.

Total organic carbon (TOC) content shows a strong and consistent inverse relationship with both magnetic susceptibility and lake level, indicative of increasing productivity during periods of low lake stand (Negrini et al., companion paper). TOC content is quite high (~5–10%) in the older Ana River Canyon outcrop sediments

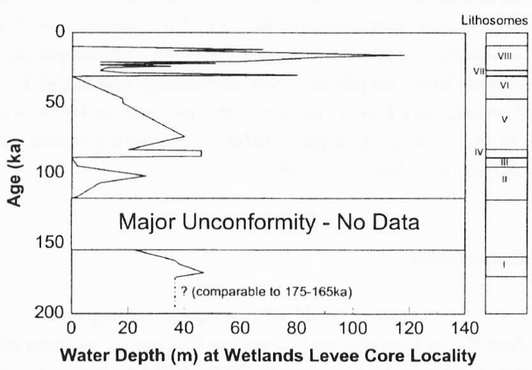


Figure 3. Lake Chewaucan lake level history for the past 200,000 yr. The curve shown is an estimate of water depth at the Wetlands Levee site, using lithologic and paleoecologic indicators at this locality and contemporaneous indicators at the other sites. This method of plotting lake level fluctuations was chosen over lake surface elevation because there has been significant sedimentary aggradation of the lake floor over the study interval, complicating comparisons of absolute lake surface elevations. Lake level estimation methods are discussed in the text. The Late Pleistocene (post 30,000 yr) record is based on a combination of data from this study and earlier work by Negrini & Davis (1992) and Negrini (submitted), and incorporates terrace elevation as well as core data.

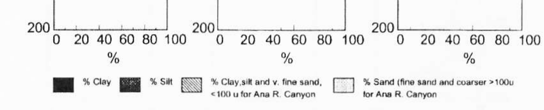


Figure 4. Granulometry data (% dry weight) for Ana River Canyon (AR-proximal deltaic site), Wetlands Levee core (WL, distal deltaic site) and Bed & Breakfast core (WL, distal deltaic site). The Ana River Canyon outcrop samples are subdivided by different size categories than the core sites because they were only measured as a byproduct of wet sieving for ostracodes.

(pre ~230 ka) (Figure 5). Values rapidly decline after this, and, aside from a brief increase between ~210–200 ka, remain relatively constant (averaging about 3.5%) throughout the remainder of the record. The rapid decline in TOC early in the Ana River Canyon record mirrors a similar abrupt change in palynological and carbonate indicators over this time period, and may reflect a real trend of rapidly declining productivity (and temperatures) after 230 ka. Slightly more erratic values of TOC occur throughout the measured portion of the Wetlands Levee and Bed & Breakfast cores. A slight shift towards higher values moving towards the more distal and deeper water sites probably results from enhanced organic matter preservation in deeper water settings, rather than reduced clastic dilution, since total sedimentation rates are higher near the depocenter.

$CaCO_3$ percentage in lake sediments is regulated by several factors (temperature, primary productivity, changes in drainage basin chemistry, salinity, dissolution and clastic dilution), complicating its interpretation. $CaCO_3$ percentages are low in both core and outcrop sediments, with no evident onshore/offshore trends in concentration (Figure 6). Similarities in major trends of $\%CaCO_3$ with paleotemperature indicators (in particular pollen and Mg/Ca, both discussed below) suggest that temperature is a dominant control in the Ana River Canyon and Wetlands Levee records. $CaCO_3$ tends to be higher during warmer intervals particularly for the early part of the Ana River Canyon record (pre

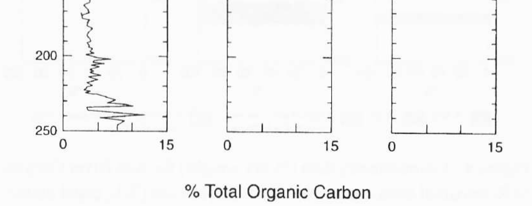


Figure 5. Total Organic Carbon (% dry weight) data for Ana River Canyon (AR-proximal deltaic site), Wetlands Levee core (WL, distal deltaic site) and Bed & Breakfast core (WL, distal deltaic site). Z notations on right side of figure indicate ostracode faunal zones.

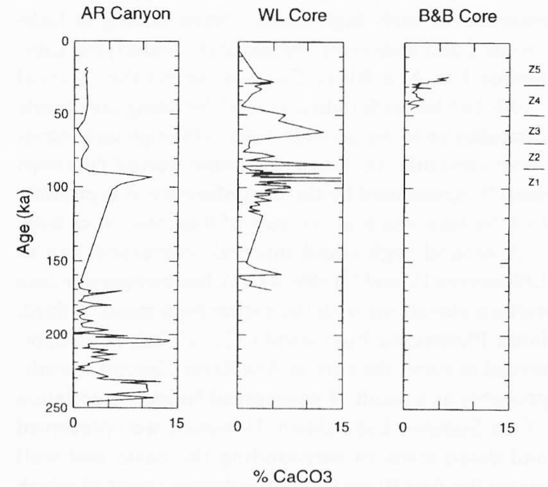


Figure 6. Calcium carbonate content (% dry weight) data for Ana River Canyon (AR-proximal deltaic site), Wetlands Levee core (WL, distal deltaic site) and Bed & Breakfast core (WL, distal deltaic site). Z notations on right side of figure indicate ostracode faunal zones.

230 ka), probably the indirect result of greater primary productivity and reduced solubility, both of which are correlated with higher temperatures. A brief, warm interlude after about 27 ka is suggested by the $CaCO_3$ record from the Bed & Breakfast site, consistent with Mg/Ca data from the same horizon discussed below. No consistent relationship exists between paleodepth estimates based on physical sedimentology and $\%CaCO_3$. A weak positive relationship also exists between $\%CaCO_3$ algae and paleosalinity estimates based on ostracodes, discussed below.

Ostracodes

Low resolution records of ostracode stratigraphy for Ana River Canyon and covering the earlier portion of the Summer Lake record (230–100 ka) have been published elsewhere (Palacios-Fest et al., 1993). We have combined these data with new collections from Ana River Canyon, using the new age model of Negrini et al. (2000) (Figure 7). Paleosalinity estimates for this interval suggest relatively saline lake conditions prior to ~200 ka, followed by an abrupt and brief increase in salinity and alkalinity, with a decline in salinity thereafter (Figure 8). Freshwater conditions are evident in the Ana River Canyon outcrop record almost up to the Lithosome I/II unconformity (~165 ka), with a brief interlude of slightly elevated salinities immediately

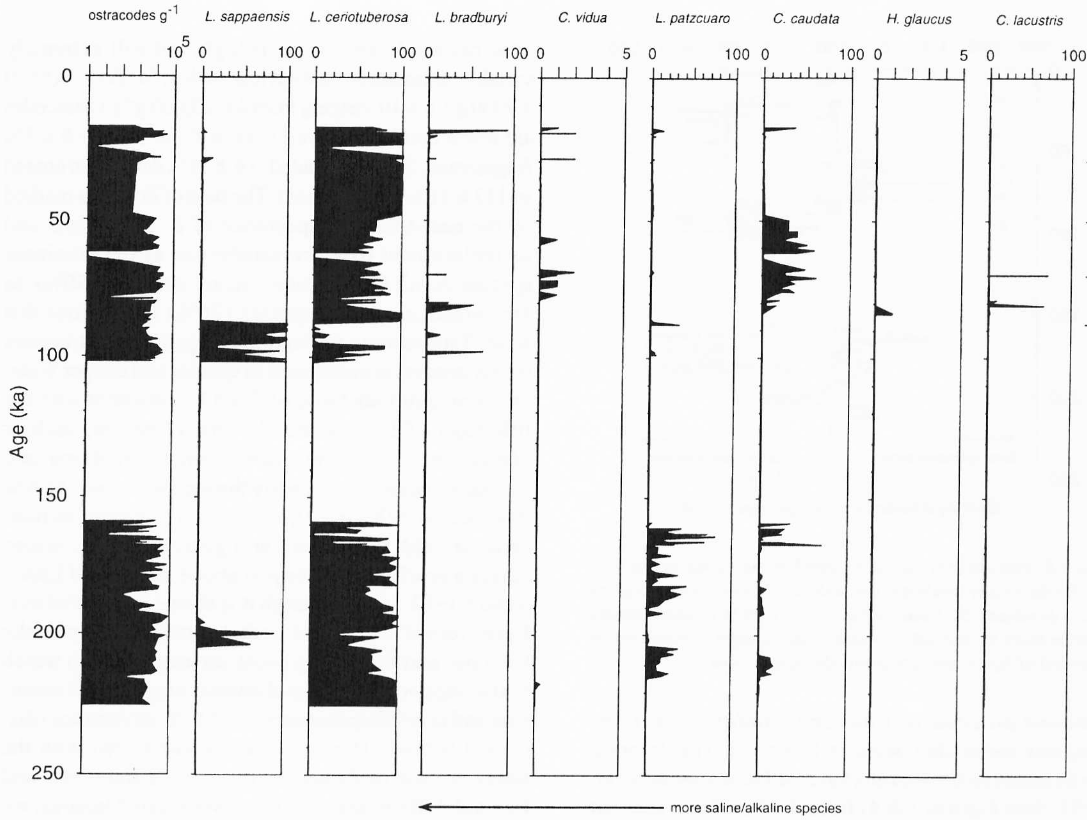


Figure 7. Ostracode stratigraphy for the Ana River Canyon outcrop site. Species are arranged with generally more saline/alkaline tolerant taxa on the left. The data are derived from Palacios-Fest et al. (1993), but are plotted using the new age model and refined tephrochronology of Negrini et al. (submitted). Z notations on right side of figure indicate ostracode faunal zones.

prior to the major unconformity (158–102 ka). Ostracodes directly above the Lithosome I/II unconformity indicated that extremely saline conditions existed during the early refilling of the lake around 102 ka. However the timing or rapidity of salinity increase and lake level fall cannot be estimated given the likelihood of lake floor deflation at some time between 158–102 ka.

For this study we made a detailed analysis of the critical post–100,000 yr interval, to document changes associated with the Late Pleistocene lake level fluctuations (Figures 9 and 10). We found a total of 12 species of ostracodes in these sediments (only those species with more than a single occurrence are plotted), all of which are common taxa in other western North American pluvial lakes. Ostracode abundance is high but erratic throughout most of the Wetlands Levee core ($\bar{x}$ = 9700 ± 14,200 ostracodes g^{-1}, ranging from 0–72,000 g^{-1}). Throughout the core, ostracode preservation is good, with abundant adults (indicating consistent population maturation) and few damaged valves.

We distinguish 5 faunal zones in our Wetlands Levee core data. Zone 1 extends from 14.76 m below the surface to 11.44 m, closely corresponding to Lithosome III. Here ostracodes are moderately abundant, mostly 10^3–10^4 g^{-1} Ostracode fossils in this interval are moderately fragmented (percentage of fragmented individual fossils=14 ± 10%, n = 17 samples of 300 individuals), abraded (13 ± 10%), and oxide stained (21 ± 23%), all consistent with the lithologic indications in Lithosome III of shallow, agitated water. The fauna is dominated by *Limnocythere sappaensis*, a species indicative of highly saline and alkaline conditions, consistent with the high frequency of carbonate-coated fossils (32 ± 27%). Adults (44%) and whole carapaces (22%) are abundant among fossils of this species. Smaller numbers of *Limnocythere ceriotuberosa*,

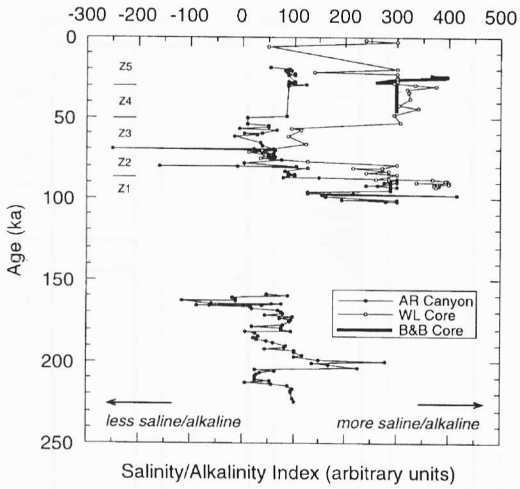


Figure 8. Paleosalinity estimates based on ostracode assemblage data for the 3 study localities. The scale is a dimensionless weighted index, developed by Palacios-Fest et al. (1993) to qualitiatively describe more vs. less saline-tolerant assemblages. Z notations on upper left of figure indicate ostracode faunal zones.

Candona patzcuaro, *Candona caudata* and *Cypridoposis vidua* also occur in this zone. This interval can be correlated with Interval 2 of Palacios-Fest et al. (1993, their Figures 3 & 4) from the Ana River Canyon outcrops, north of Summer Lake, although we now recognize that our earlier estimates for the age of this interval are too young. Based on the correlation of ostracode zones with the Ana River Canyon section, and the tephrostratigraphy discussed earlier, we estimate that Zone 1 extends from ~94 to ~89 ka.

Zone 2 extends from 11.44–8.90 m, equivalent to Lithosome IV and the lowest part of Lithosome V. In this zone ostracode abundance is high, but still extremely variable, although somewhat less so than for Zone 1 (8800 ± 7000 g^{-1}, n = 10, ranging from 45–23,000 g^{-1}). Ostracodes are much better preserved than in Zone 1 (only 6 ± 3% fragmented, 5 ± 5% abraded, 14 ± 31% carbonate coated and 12 ± 31% oxide stained). The base of Zone 2 is marked by the near-total disappearance of *L. sappaensis* and its replacement by *L. ceriotuberosa* as the dominant species. Adult *L. ceriotuberosa* are common (50%), as are carapaces of this species (26%) throughout this zone. Taphonomic indicators suggest that ostracodes in this interval accumulated in quieter and deeper water environments than those of Zone 1, consistent with the lithology of Lithosome IV. Other species, such as *Limnocythere platyforma*, *Limnocythere bradburyi*, and *C. caudata* occur occasionally through this zone. At Ana River Canyon (Figure 7), *Cytherissa lacustris*, an indicator of cold, very fresh, and generally deep, water, makes a brief appearance just above the base of Lithosome V (~83.5 ka), although it is absent in the Wetlands Levee record. This event probably records a rapid lake level rise and freshening event, an interpretation which is also supported by a simultaneous negative ^{18}O excursion and brief disappearance of $^{18}O/^{13}C$ covariance (discussed below). This zone can be correlated with the lower portion of Palacios-Fest et al.,'s Interval 3 and Interval 4. By interpolation between dated tephras, we estimate this zone to have formed ~89–70 ka.

Zone 3 extends from 8.90–6.63 m., equivalent to most of the remainder of Lithosome V. In this zone ostracodes are consistently abundant (27,000 ± 17,000 g^{-1}, n = 10, ranging from 2400–58,000 g^{-1}), and very well preserved (7 ± 2% fragmented, 5 ± 5% abraded, 8 ± 16% carbonate-coated and 2 ± 3% oxide stained). Like the previous zone, both taphofacies and lithofacies in-

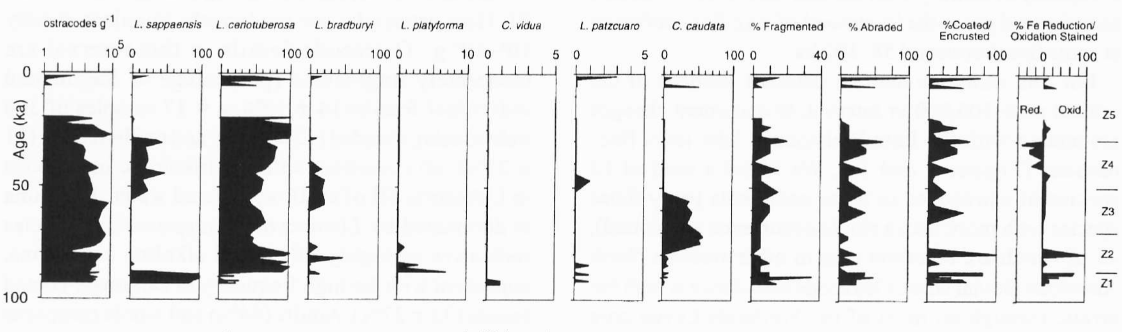


Figure 9. Ostracode stratigraphy for the upper portion of the Wetlands Levee core. Species are arranged with generally more saline/alkaline tolerant taxa on the left. Z notations on right side of figure indicate ostracode faunal zones.

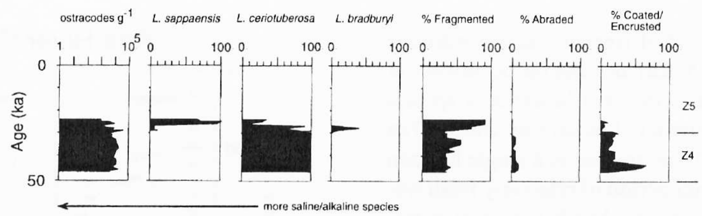


Figure 10. Ostracode stratigraphy for the Bed & Breakfast core. Z notations on right side of figure indicate ostracode faunal zones.

dicators suggest deposition in deep, calm water. This zone is characterized by the codominance of *C. caudata* and *L. ceriotuberosa*. Other *Limnocythere* species are rare to absent through this interval, except for moderately abundant *L. bradburyi* near the top of the interval. This interval can be correlated with a similar *C. caudata*-rich portion of the Ana River Canyon section (see Palacios-Fest et al., 1993, their lower-middle Interval 5). The age of Zone 3 is ~70–50 ka.

Zone 4 (6.63–3.37 m) corresponds to uppermost Lithosome V, Lithosome VI and the lower part of Lithosome VII. In this zone ostracodes are less abundant and more poorly preserved than in Zone 3 (2300 ± 3100 g^{-1}, n = 9, ranging from 93–9900 g^{-1}, 11 ± 5% fragmented, 13 ± 7% abraded, 30 ± 18% carbonate coated, and 15 ± 22% reduction stained). The combination of moderate fragmentation and abrasion, with common coated grains and reduction staining probably results from accumulation in shallow water, perhaps under conditions of high biological oxygen demand or productivity. This interpretation is also supported by the lithologic character of Lithosome VI-lower VII. This interval is marked by the disappearance of *C. caudata* and the reappearance of *L. sappaensis* and *L. bradburyi*. The latter 2 species commonly occur as both adults and whole carapaces, suggesting good environmental conditions for maturation of these alkalophile species. Based on the disappearance of *C. caudata* and the occasional appearance of *L. bradburyi* at Ana River Canyon, this zone probably correlates with most of the upper part of Palacios-Fest et al.,'s Interval 5. No obvious correlation can be drawn between this zone and the contemporary ostracode stratigraphy at the Bed & Breakfast (distal) core site (where the fauna is almost exclusively *L. ceriotuberosa* during this time), although both tephrochronology and magnetic susceptibility records indicate that the two records overlap in time (Negrini et al., 2000). Based on the bracketing ages of the Mt. St. Helens Cy Tephra and the Mono Lake geomagnetic excursion, the age of Zone 4 is ~50–29.7 ka.

Zone 5 (3.37–0 m, 29.7 ^{14}C yr B.P-Holocene) corresponds with the upper portion of Lithosome VII and all of Lithosome VIII. Ostracodes occur very erratically through this zone (8900 ± 20,000 g^{-1}, n = 13), ranging from 72,000 to 0 g^{-1} near the top of the core. Ostracode fossils are rarely fragmented (7 ± 4%) or abraded (4 ± 3%) in this interval, although whole carapaces are relatively uncommon. Carbonate-coated fossils are common in this zone (23 ± 22%). In the lower part of the zone reduction-stained ostracodes occur commonly (similar to those of Zone 4) but these disappear above 2.4 m below the surface, replaced by unstained or lightly oxidized fossils. The taphonomic indicators suggest quieter water conditions and decreasing organic matter accumulation or BOD relative to Zone 4, although the presence of abundant sandy lags suggests that water depth fluctuated considerably during this time. The base of this zone is marked by the disappearance of *L. sappaensis*. Through most of this zone the ostracode fossils are almost exclusively *L. ceriotuberosa* but *C. caudata* reappears abundantly near the top of the core. An abrupt change in the assemblage at the Wetlands Levee core top (*L. bradburyi*, *L. sappaensis*, and *C. patzcuaro*) is consistent with the interpretation that these sediments are Holocene in age.

Paleosalinity estimates based on ostracode faunal assemblages during the 230–162 ka interval, only recorded in the Ana River Canyon site, shows moderate to high salinities prior to 190 ka (very high at about 200 ka), giving way to decreasing salinities between 190–165 ka (Figure 8). A brief interval of rising salinity occurs just prior to the major unconformity (165–162 ka). Extremely saline values are indicated between 102–95 ka at Ana River Canyon. For the post-95 ka interval, the 2 records show a good agreement in major trends, with indications of elevated salinities during the deposition of Zone 1, declining salinities within Zone 2, low salinities within Zone 3, higher and stable salinities within Zone 4, and highly variable salinities within Zone 5. The absolute differences between the 2 records reflects the proximity of the Ana River Canyon outcrops to the paleoinflow relative to the more distal Wetlands Levee site.

The prolonged period of fresher water conditions evident in Zone 4 (70–50 ka) includes the occurrence at Ana River Canyon of *Cytherissa lacustris*, a species that is restricted to extremely low salinity waters. The extreme abundance of this species at a single horizon is consistent with a brief period of relatively fresh water input to the lake from the Ana River early in this interval. However, no specimens of this species were found in the Wetlands Levee core. This observation, taken with the trace element data discussed below, suggests 2 possibilities. Lake waters may have been both cold and relatively fresh near the Ana River outlet, whereas the open lake water near the Wetlands Levee site, although also cold, remained too saline to support this species. Alternatively, this very fresh and cold interlude was quite brief and was simply missed in the Wetlands Levee samples.

Ostracode paleoecology and taphonomy are in very good agreement with the lithostratigraphic record of lake level change. A ~95–89 ka interval of high salinity and shallow water was followed by a deeper phase with indications of declining salinity (89–~70 ka). The freshest water indications, (70–50 ka) correspond with apparent high stand conditions. A return to more saline lake conditions after 50 ka corresponds to the general drop in lake levels indicated by both lithology and ostracode taphonomy.

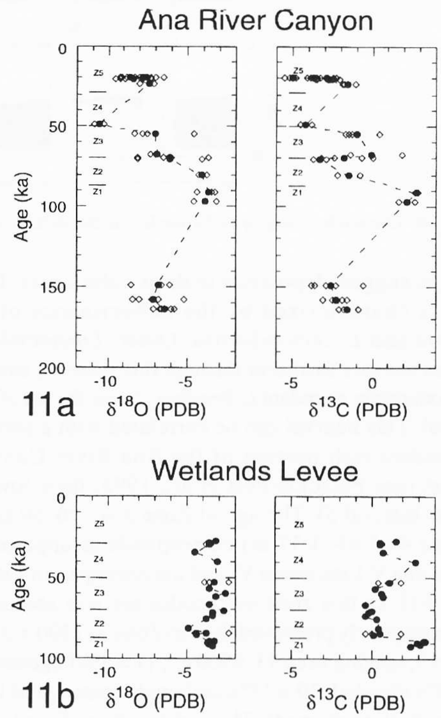


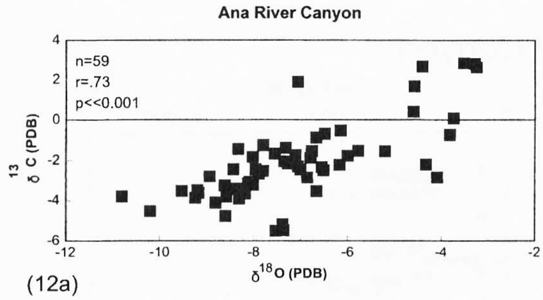
Figure 11. Stable isotope $\delta^{18}O$ and $\delta^{13}C$) stratigraphy from ostracode valves for the Ana River Canyon (11a) and Wetlands Levee (11b) sites.

Stable isotopes

Stable isotope ($\delta^{18}O$ and $\delta^{13}C$) records from ostracode valves from the Wetlands Levee and Ana River Canyon sites are quite erratic, with short-term (< 1 ka) fluctuations of up to several ‰, as shown by both between – and within – horizon variance (Figure 11). This variability is probably the result of short-term variability in evaporation/precipitation ratios within the basin. The Ana River Canyon values for both $\delta^{18}O$ and $\delta^{13}C$ are significantly lower than those of Wetlands Levee, reflecting the local influence of river inflow at the former site, and the subsequent evaporative enrichment of the main lake water mass at the latter site (Figures 11 & 12). The Ana River Canyon values are also more variable than those of Wetlands Levee, perhaps because the Ana River inflows included runoff and groundwater sources that only later became homogenized within the main lake. This pattern of isotopic variability has been documented elsewhere by Sharpe (1998).

$\delta^{18}O$ values at both sites are high between ~94–89 ka. Wetlands Levee oxygen isotope values remain relatively constant through the period from ~94–27ka, suggesting that the central portion of the basin did not experience major variation in residence times. However, at the more proximal Ana River locality, major excursions were noted, which may correspond to variations in shallow groundwater input. From ~80 – 50 ka there is a shift to more negative values. The uppermost isotopic samples (~22 ka) have intermediate compositions between the very negative 50 ka and very positive (94–80 ka) samples. The much greater variability and evident temporal trends at Ana River versus Wetlands Levee may suggest that Ana River water (sourced from shallow groundwater) is responding to long term variability in the isotopic composition of precipitation (perhaps changes in precipitation source area or the proportions of summer rain vs. winter snow). Subsequent isotopic evolution within Summer Lake caused waters to evolve towards a fairly uniform composition, which may indicate that Ana River is not the predominant source of moisture for the basin over most of its history.

A statistically significant covariance between $\delta^{18}O$



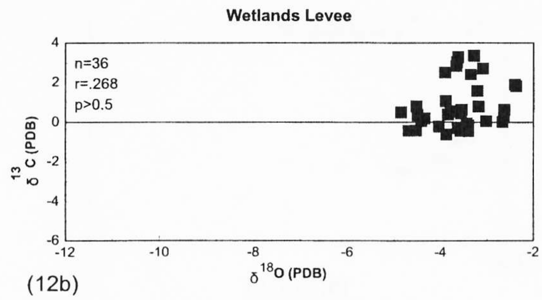


Figure 12. Cross plots of $\delta^{18}O$ vs. $\delta^{13}C$ for both Ana River Canyon (12a) and Wetlands Levee (12b) data.

and $\delta^{13}C$ absolute values is evident at Ana River Canyon (Figure 12a). At Wetlands Levee, there is no statistically significant covariance between $\delta^{18}O$ and $\delta^{13}C$ absolute values (Figure 12b). Talbot & Kelts (1990) have argued that long-term covariant patterns in $\delta^{13}C$ and $\delta^{18}O$ values support the presence of hydrologically closed-basin conditions. However this cannot explain the pattern we observe at Summer Lake, where covariance is evident at the deltaic (Ana River Canyon) locality, but not in the main lake (Wetlands Levee) locality. A more likely explanation for the Ana River covariance is that intervals of more positive $\delta^{18}O$ are reflecting warmer periods when precipitation was dominated by increased summer rainfall derived from lower latitudes (more isotopically positive sources) and primary productivity would have been higher, reflected in more positive $\delta^{13}C$ values.

Mg/Ca And Sr/Ca results

Mg/Ca and Sr/Ca ratio profiles were measured on ostracodes from all three localities (Figure 13). Under conditions of uniform solute input, Mg/Ca and Sr/Ca ratios are presumed to reflect ambient paleosalinity and paleotemperature during the ostracode growing season (late spring-summer). However, research disagrees on the primary environmental controls on the uptake of these elements. Chivas et al. (1983, 1986) suggested that Mg/Ca_{valve} ratios in ostracodes are dependent both on temperature and the Mg/Ca_{water} ratio (the latter frequently strongly correlated with Total Dissolved Solids (TDS) in evolving brines when Ca and Mg are dominant cations), whereas Sr/Ca_{valve} ratios are dependent solely on Sr/Ca_{water} (also frequently correlated with TDS within individual lakes). Chivas et al. (1986) showed that $Sr/Ca_{valve}/Sr/Ca_{water}$ (their K_D) varies between species, apparently in a phylogenetically significant fashion, since related ostracode species had similar values. Somewhat different results have been obtained by other authors (e.g. Engstrom & Nelson, 1991; Palacios-Fest et al., 1993; Palacios-Fest, 1996). In experiments with *Limnocythere staplini*, Palacios-Fest (1996) found that Mg/Ca_{valve} was highly correlated with the ambient growth temperature of the water. For the same species there was little or no correlation between Mg/Ca_{valve} and Mg/Ca_{water}, whereas Sr/Ca_{valve} ratios are correlated with Sr/Ca_{water}.

Thermodynamic considerations would suggest that ambient Mg/Ca_{water}, which in both the Summer Lake and culturing systems, is correlated with salinity, should strongly influence the Mg/Ca_{shell} ratio. The fact that Palacios-Fest's culturing experiments did not produce this result strongly suggests that biokinetic controls, as opposed to thermodynamics, are most critical in regulating ostracode shell Mg uptake, at least within the cultured taxon. Although little is known about the biokinetics of ostracode shell formation in general, a number of prior studies show the importance of examining Mg uptake rates, as opposed to simply free-metal ratios, in examining inorganic carbonate precipitation (Given & Wilkinson, 1985), organically-formed carbonate shell material (Rosenthal & Katz, 1989, for freshwater molluscs), and physiologic Mg-uptake (De Lagarra et al., 1985 for the crayfish *Procambrus clarkii*). The latter study is particularly intriguing, because it demonstrates that ambient temperature had a strong effect on Mg uptake in a crustacean (albeit distantly related to ostracodes). Bodergat et al. (1993) showed that metabolic rates (strongly correlated with temperature) can have an important influence on Mg-uptake in the cytheracean ostracode *Leptocythere psammophilia*. Finally, other multi-indicator studies have shown strong relationships between Mg/Ca_{shell} in fossil material and alternative indicators of paleotemperature (but not with paleosalinity) (e.g. Rasmussen et al., 1999). Disagreements between authors as to the relative roles of thermodynamics and biokinetics in controlling ostracode shell formation may reflect their variable importance for

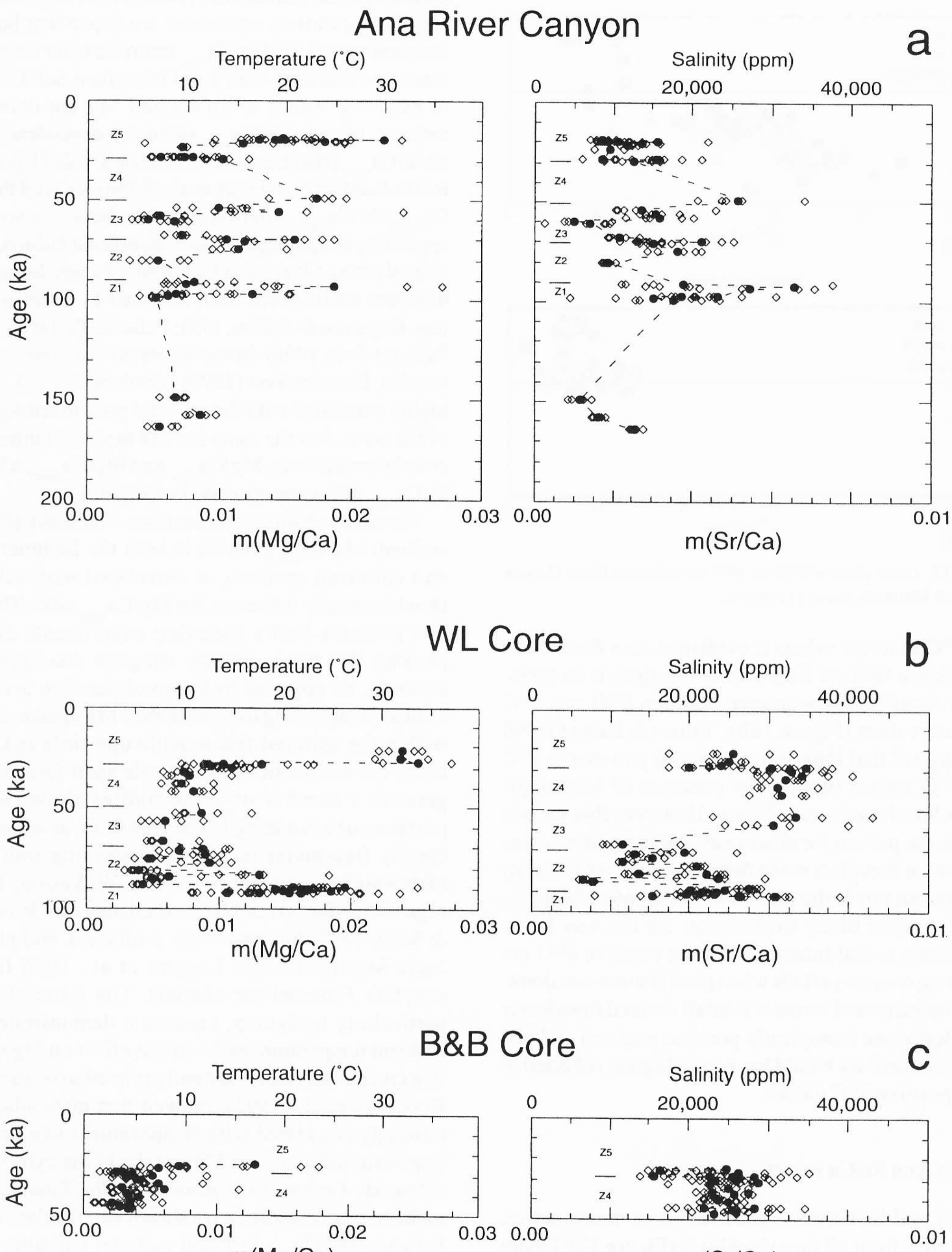


Figure 13. Elemental ratio (Mg/Ca and Sr/Ca) data for the ostracode *Limnocythere ceriotuberosa* from the Wetlands Levee core. Temperature and salinity estimates are based on application of the multiple regression models of Palacios-Fest (1996), based on results from culturing experiments of the species *Limnocythere staplini*, a closely related species to the fossils used in this study. Absolute values of temperature and salininty estimates are only interpretable for the Mg/Ca and Sr/Ca ratio ranges over which culturing experiments provide data (15–25 degrees C. and 10,000–30,000 ppm), and it is likely that the relationships beyond these ranges (particularly for higher Mg/Ca and temperature estimates) are nonlinear. Open symbols are measured values and filled symbols are means for that stratigraphic horizon.

different taxa. Our application of the culturing results from *L. staplini* to *L. ceriotuberosa* reflects the fact that the former species is the most closely related species we have to serve as a model for comparison with the fossil material. It is entirely possible that *L. ceriotuberosa* Mg-uptake behaves by its own rules, but those rules are currently unknown, pending future culturing experiments.

From these experiments Palacios-Fest developed multiple regressions that empirically relegate Mg/Ca_{valve} and Sr/Ca_{valve} ratios to temperature and TDS. These models produced realistic paleosalinity and paleotemperature estimates for Late Holocene Hohokham irrigation canals in Arizona. More recently, Palacios-Fest et al. (in review) applied this model to Late Pleistocene-Holocene deposits of Laguna Babicora, Chihuahua, Mexico and also obtained realistic paleotemperature and paleosalinity estimates (i.e. results for modern ostracodes were consistent with instrumentally recorded temperature and salinity). No analogous experiments have yet been performed for *L. ceriotuberosa*, the ubiquitous *Limnocythere* species in the Lake Chewaucan record. However, based on the close phylogenetic relationship between the species in question, we have calculated paleotemperature and paleosalinity estimates for the *L. ceriotuberosa* record at Wetlands Levee, using the *L. staplini*-based regressions. It is important to note, however, that Palacios-Fest's original experiments were conducted over narrow temperature and salinity ranges (15–25 °C and 10,000–30,000 ppm TDS, respectively). Therefore we cannot make linear extrapolations beyond the culturing experimental ranges. For example the extremely high Zone 5 paleotemperature estimates that would result from such an extrapolation are almost certainly too high. It is interesting in this regard that the previously-cited study by Rasmussen et al. (1999) also suggests significant nonlinearity in Mg uptake at higher temperatures (i.e. Mg uptake faster than a linear relationship with temperature would predict), based on unreasonably high paleotemperature estimates.

In the Ana River Canyon record (Figure 13a) low values of Mg/Ca and Sr/Ca occur in Lithosome I below the major unconformity (~164–158 ka). Low values of both ratios also occur immediately above the unconformity, but both ratios rise after ~100 ka in Zone 1 and then quickly fall. This pattern of spiky, covariant fluctuations in Mg/Ca and Sr/Ca repeats several times between 90–19 ka.

Short term (< 10 ka) variability in the Wetlands Levee record (Figure 13b) is more subdued than at Ana River Canyon (presumably a result of the former's more distal setting, where lake waters were more homogenous). However, large-scale fluctuations are still evident. Like the Ana River Canyon record, Mg/Ca ratios initially rise and then are high for most of ostracode Zone 1 (~94–89 ka). Sr/Ca ratios are generally high throughout this interval. In Zone 2 and most of Zone 3 (~89–~60 ka) both Mg/Ca and Sr/Ca ratios are much lower. In upper Zone 3 Sr/Ca ratios abruptly rise and are very high throughout Zone 4 (~60–~29 ka), but Mg/Ca ratios remain relatively constant and low throughout this interval. In Zone 5 (< ~27 ka) Mg/Ca ratios rise to high values. This increase corresponds with a correlative rise in Mg/Ca at Bed and Breakfast (Figure 13c), but is unrepresented in the Ana River Canyon record as a result of an unconformity. Very high Mg/Ca values at the top of the core may be Holocene (post 5.5 ka) in age. Sr/Ca ratios initially decline in the 27 to ~23 ka interval, but in the Holocene they again rise to very high values.

The Bed and Breakfast record (Figure 13c) resembles that of Wetlands Levee, in that low values of Mg/Ca occur prior to ~30 ka, followed by a dramatic rise in this ratio at about 27 ka, coincident with the Mg/Ca rise at Wetlands Levee. However, no trend in Sr/Ca is evident over this interval. Low and invariant paleotemperature estimates (averaging ~4–6 degrees C from 50–27 ka) at the Bed and Breakfast site reflect the fact that this locality was at or very close to the lake's depocenter, and was probably normally below the summer thermocline.

Using Palacios-Fest's laboratory observations we interpret the Mg/Ca and Sr/Ca records as follows. From 164–158 ka Lake Chewaucan had relatively low TDS and was cool and similar conditions occurred immediately above the major unconformity, about 100 ka. No evidence for the brief reversal towards warmer and drier conditions after 165 ka (indicated by the ostracode assemblage, lake level and pollen data) is apparent in the Mg/Ca and Sr/Ca records. During the deposition of the lower portion of ostracode Zone 1, temperatures initially increased slightly (between ~95–94 ka) and salinity declined slightly over the same interval. A period of relative stability in both temperature and salinity ensued through the remainder of Zone 1 (94–89 ka). At Ana River Canyon, a spiky, covariant record of Mg/Ca and Sr/Ca between ~95–19 ka reflects short-term variability in the Ana River inflow. Periodic high temperatures probably reflect intervals when the outcrop site was persistently above the thermocline of the lake during summer months.

During these times, when the lake was somewhat shallower, inflow from the Ana River must have been reduced and the salinities of this distal-deltaic area more closely approximated those of the open lake. This inflow variability, coupled with abundant unconformities in the Ana River Canyon record, makes Mg/Ca and Sr/Ca trend correlations with the other two records difficult. At Wetlands Levee, at about 89 ka, both temperature and salinity appear to have dropped rapidly and remained low throughout Zone 2 and lower Zone 3 (89–60 ka). Approximately 60 ka salinity increased abruptly, while temperatures remained low and relatively constant. This situation continued until near the top of the Pleistocene section ~27–23 ka. A major rise in paleotemperature estimates is recorded at that time at Wetlands Levee and Bed and Breakfast, whereas this interval is marked by an unconformity at Ana River Canyon. High paleotemperature estimates and modest decline in salinity are evident in Holocene sediments at Wetlands Levee.

An alternative interpretation of this record based on the Chivas et al., model (1983, 1986) would agree with the above reconstruction except that the intervals of covarinat change in Sr/Ca and Mg/Ca (e.g. at the end of Zone 1) do not indicate any necessary change in temperature (see De Deckker & Forester, 1988, Figure 6 for an example). Thus, this model does not support the extended interval of lowered temperatures between 89–25 ka produced by the Palacios-Fest model.

Palynology

Pollen abundance was highly variable and, except in a few cases, clearly a function of preservation rather than of accumulation rates. Pollen from sediments deeper in the basin was less frequently subjected to oxidation and better preserved. Pollen preservation was best at Bed & Breakfast and worst at Ana River Canyon. Pollen abundance also varied vertically, with poorer pollen preservation beneath unconformities and better preservation above these surfaces. Pine (*Pinus*) pollen predominates all three records (Figures 14a, 14b & 14c). Of the remaining arboreal pollen types, spruce (*Picea*), fir (*Abies*), and juniper (*Juniperus*) are the

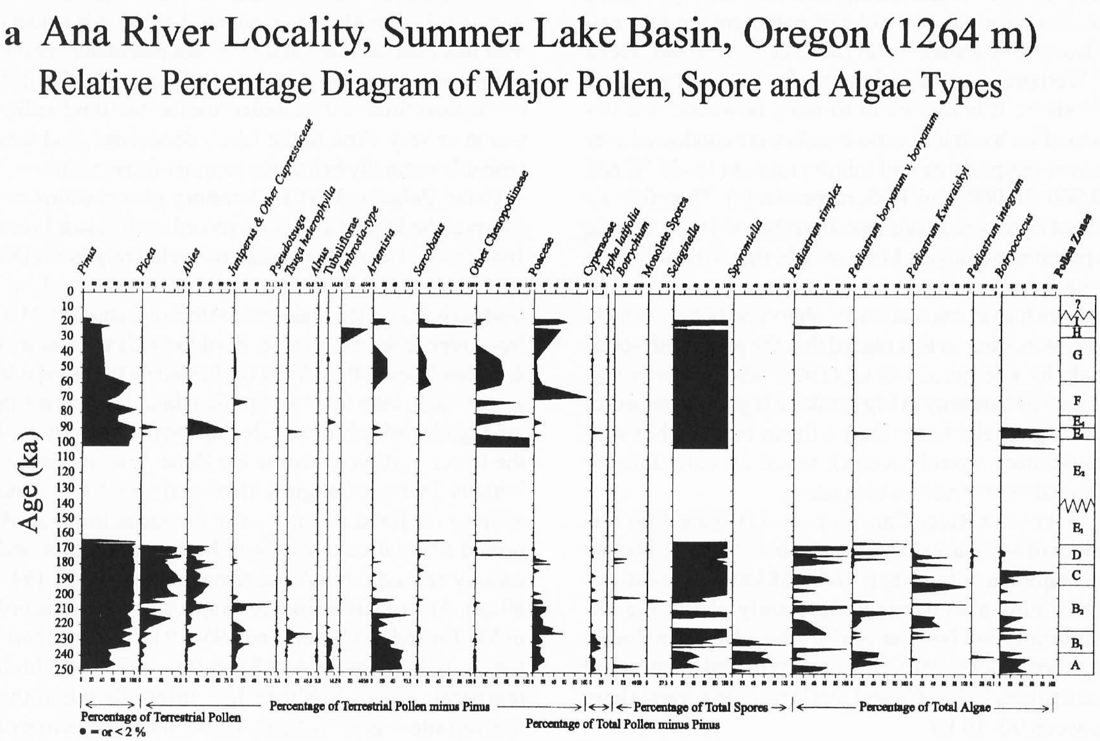
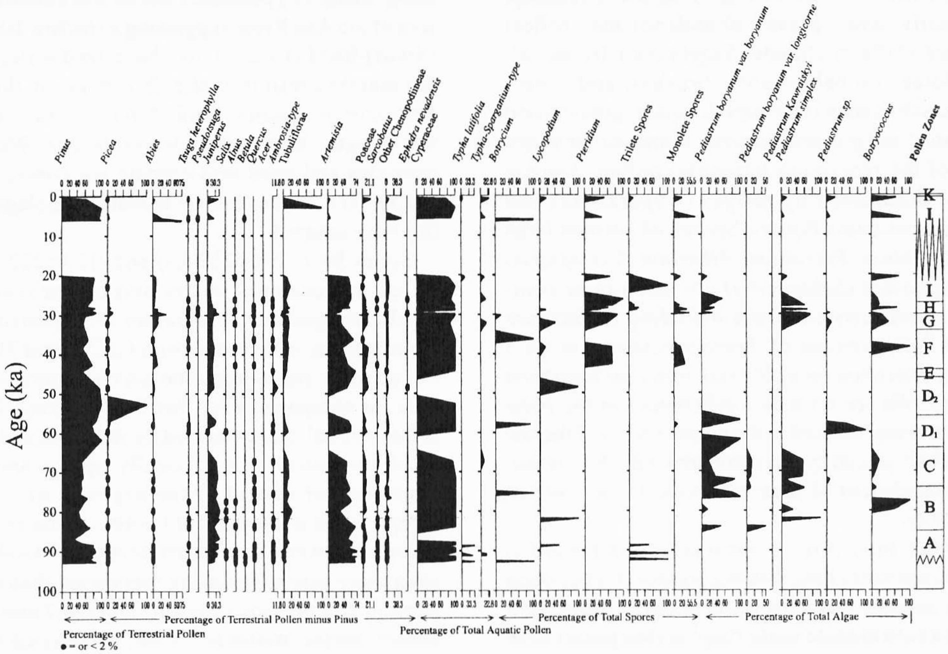


Figure 14. Relative percentage diagram of major pollen, spore and algae types. (a) Pollen stratigraphy for the Ana River Canyon outcrops; (b) Pollen stratigraphy for the Wetlands Levee core; (c) Pollen stratigraphy for the Bed & Breakfast core.

b Wetlands Levee Locality, Summer Lake Basin, Oregon (1264 m)

Relative Percentage Diagram of Major Pollen, Spore and Algae Types



c Bed and Breakfast Locality, Summer Lake Basin, Oregon (1264 m)

Relative Percentage Diagram of Major Pollen, Spore and Algae Types

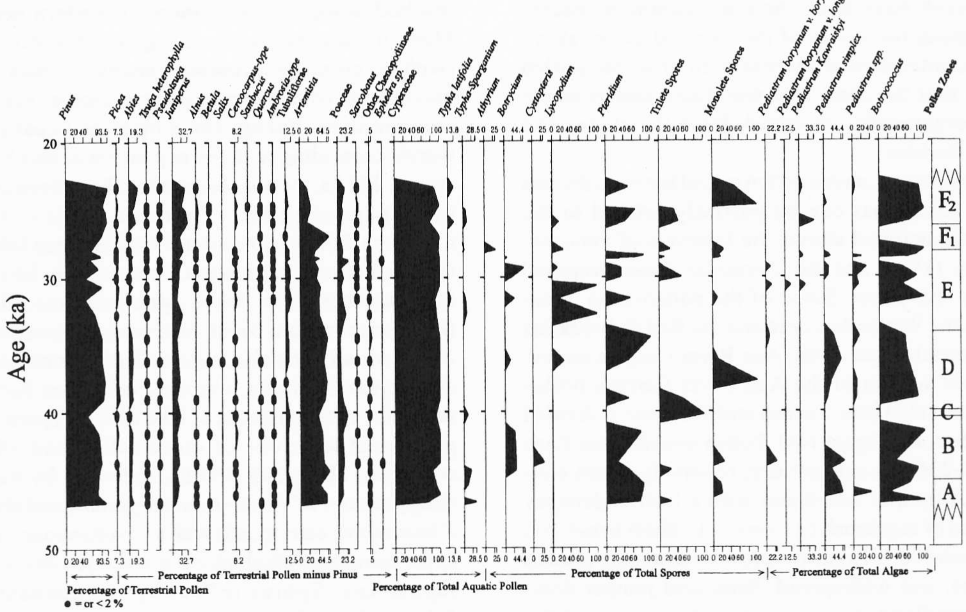


Figure 14. Continued.

most abundant. Douglas fir (*Pseudotsuga* sp.) and western hemlock (*Tsuga heterophylla*) are rare, except occasionally, when greater abundance may reflect significant shifts in climate. Sagebrush (*Artemisia*), Tubuliflorae (probably rabbit brushes, and horsebrushes), saltbushes (Chenopodiineae), greasewood (*Sarcobatus*) and grasses (Poaceae) comprise the major portion of the remaining terrestrial pollen. Aquatic pollen is dominated by sedges (Cyperaceae) and occasional cat-tails (*Typha*). Spores of various ferns (e.g. *Botrychium, Pteridium, Athyrium, Cystopteris*) and spikemosses (*Selaginella*) are much rarer components. Acid resistant algae including *Pediastrum* spp. and *Botryococcus* cf. *braunii* occurred at all 3 localities, providing an additional proxy for lacustrine conditions. We see no major differences in the plant communities represented in the pollen record of the last one and a half glacial cycles compared with the present, but the abundances of their components have varied considerably.

Although pine pollen dominates the record it represents primarily long distance transport. The pollen evidence indicate that both warm climate (diploxylon pines) and cold climate pines (haploxylon pines) have also varied considerably in abundance during the last 250 ka in the area reflecting significant regional changes in temperature, and precipitation. Although plants that comprise the current sagebrush steppe and desert scrub have been the most common species present along the shores of the lake and slopes above the lake, intermittent increases in arboreal pollen indicate that the semi-arid woodland and montane forest occasionally advanced down the slopes and reached the lake.

The Ana River Canyon pollen record has been divided into 13 zones that can be partially related to the lithosomes outlined above, the Intervals of Palacios-Fest et al. (1993) and the Ostracode Zones proposed earlier in this paper. Some of the pollen-zone boundaries in the Wetlands Levee and the Bed & Breakfast records match those in the Ana River Canyon record. The lower portion of the Ana River Canyon pollen record is divided into 5 zones and 1of these is divided into 2 subzones (Figure 14a). Pollen records from Zone A (~250–236 ka) suggest dry, relatively warm conditions. A juniper woodland with a lush understory comprised of sagebrush (*Artemisia*), rabbit brush and horsebrush (both Tubulifloraes) and abundant grasses (Poaceae), was widespread. Semi-arid juniper dominated woodland was also a significant part of the landscape. Pines were still important in the region and probably dominated the Winter Ridge as they do today. Sedge (Cyperaceae) marsh was extensive in the area of the Ana River suggesting a shallow lake. Alder (*Alnus*)-lined channels may have feed water through the marshes into the lake. Presence of the algae, *Pediastrum simplex* and *Botryococcus*, indicates warm water with eutrophic conditions. Warm temperatures and semi-arid climate are consistent with $CaCO_3$, TOC and ostracode faunal assemblage data for this time interval.

Zones B1 (~236–222 ka) and B2 (~222–192 ka) reflect increasingly wetter and cooler conditions. Cooler temperature indicators are consistent with simultaneous indications from $CaCO_3$ and TOC data, but no other paleoprecipitation indicators are available for comparison for this time interval. Zone B1 is transitional, characterized by regional, and perhaps local increases in pine. Locally, spruce and fir and even juniper spread at the expense of the shrub steppe. Most diagnostic of cooler, wetter conditions is the sudden rise in western hemlock. Marshes were even more extensive during this period than they had been in the previous period. During Zone B2, the cooler, mesic, montane forest, dominated by pine, spruce and fir, became fully established. While juniper woodland with a grassy sagebrush shrub understory retreated before moisture loving montane forest species, much colder winter conditions may have resulted in local retrenchment of western hemlock. Short-lived expansions of sagebrush shrub steppe (without coincident juniper expansion) may reflect movements of the upper, or near alpine, sagebrush communities. Except for a brief re-expansion of marsh, coinciding with the expansion of shrub steppe around 228 ka, the marshes seem to have been covered by an expanding lake. Dramatic shifts in the abundance of the various algae suggest deep lake conditions, but with significant fluctuation in lake levels and lake chemistry. Based upon other records from the Intermountain West, increased abundances of *Pediastrum kwaraiskyi*, must reflect a contracting lake and increased salinity, whereas *Pediastrum boryanum* seems to reflect episodes of lake filling. A particularly prominent interlude of more saline and alkaline conditions about 205–195 ka, indicated by the near-disappearance of *Pediastrum boryanum*, and abundant *P. kwaraiskyi*, corresponds with a simultaneous increase in the halophilic/alkalophilic ostracode *Limnocythere sappaensis*. Spores reflecting the abundance of *Selaginella* (spikemoss), probably track a decrease in local temperature. *Selaginella* abundance reached a maximum by 200 ka. Indications of cooler and usually wetter climate in Zone B relative to Zone A are consistent with all available paleolimnological evidence, which supports a colder and fresher lake for this time.

Zones C (~192–178 ka) and D (~178–165 ka) reflect the coolest conditions of Isotope Stage 6 at Summer Lake, consistent with paleolimnological indications of high lake stand, and geochemical indications of low summer temperatures. Shifts in the abundance of spruce versus fir comprise the main difference between zones C and D with grasses becoming more important in the zone D. The end of this period is climaxed by the coolest, most mesic periods of the last 200 ka at Summer Lake.

Zones E1 (~ 165–163 ka), E2 (~102–95 ka), E3 (~95–87 ka), and E4 (~87–80 ka) reflect relatively similar vegetation conditions in the Summer Lake Basin. Zones E1 and E2, although they stradle a major unconformity, were more similar to each other than the post-unconformity subzones to the vegetation in Zone D (Figures 14a and 15a). Higher abundances of saltbushes and declines in pine and other cool, moist montane forest species, as well as of grasses, indicate much drier climatic conditions. There is no evidence for marsh conditions in the Ana River Canyon area at this time. A dramatic decrease in *Selaginella* abundance indicates much warmer conditions. Greasewood was abundant in Zone E1 but absent in Zone E2, where other saltbushes were abundant. Increasing pine pollen abundance in the record probably reflects the decline in local pollen production and the increasing importance of pollen derived from long distances. Marshes were absent or rare, but the abundance of the algae *Botryococcus* reflects eutrophic lake conditions. Indications of drier conditions during Zone E1 and E2 are consistent with paleolimnological indications of precipitation (reduced lake depth and area, and the ostracode assemblage data). However paleotemperature estimates both immediately prior to and after the unconformity interval are inconsistent.

Zone E3 (~95–87 ka) of the Ana River Canyon sequence approximately corresponds to Zone A (~ > 93–85 ka) of the Wetlands Levee record (Figures 14a & 14b). From this point on comparison of records from several localities in the Summer Lake Basin highlights differences in the pollen record of a large pluvial lake basin resulting from differences in depositional environment and post-depositional sub-aerial exposure. Particularly crucial were the composition of the plant communities closest to each site (or distance from shore if the site was underwater), and how often the sites were subjected to sub-aerial weathering (poor pollen preservation is indicated by greatly reduced pollen type diversity). From about 100 ka until the latest Pleistocene, the Ana River Canyon site was in the proximal parts of the Ana River delta, and was probably subaerially exposed for most of this time, as indicated by unconformities, soils and shallow water indicators. The pollen record indicates that the area surrounding the Ana River Canyon locality was more often dominated by desert scrub vegetation than by marsh. The Wetlands Levee area lies deeper in the basin and less water would be needed to create marsh or to flood the area. Therefore, the pollen record from the Wetlands Levee area contains much more marsh pollen than that of the Ana River Canyon. In addition, the pollen of vegetation communities up-wind of the Wetlands Levee locality, is more common than pollen from the lower ridges and shallow basins that lie upwind of the Ana River Canyon locality.

During the Ana River Canyon Zone E3 and Wetlands Levee Zone A period terrestrial vegetation records from the Ana River Canyon locality indicates an advance of upper or near alpine sagebrush-dominated vegetation (Figure 14a & 14b). Significant amounts of pine pollen were still blowing in from the west and dominating the local pollen record, although local pine forests were also clearly extensive. The greater abundance of fir and pine pollen, as well as western hemlock, suggests that these 3 species were more important in the montane forests to the west than during the previous period. Spruce and fir pollen is more abundant in the Ana River Canyon record because more of these heavy grains would have settled out closer to shore than the Wetlands Levee Locality. However, the Wetlands Levee record also indicates that mixed conifer montane forest comprised of pine, spruce and fir grew on the slopes west of the lake, and western hemlock may have been growing nearby. The Wetlands Levee pollen mirrors the expansion of sagebrush steppe revealed in the Ana River Canyon record. A juniper woodland with a sagebrush understory was common on the lake-side slopes, and willow (*Salix* spp.), alder, and birch (*Betula* sp.) grew either in places along the lake-shore or along the streams that fed the lake. Increased sedge pollen at both localities indicates that marsh was expanding. In addition, wide-leafed cat-tails increase in proportion to sedges indicating an input of fresher water (Figures 14b & 15b see the Cat-tails/Sedges ratio). Increasingly cool-moist climatic conditions seem to have prevailed. This is confirmed by the spikemoss spore record and a general increase in fern spores (Figures 14b &15a).

Paleolimnological indicators for this time interval also support a cooling trend, with initially dry and relatively warm conditions that became increasingly mesic, accompanied by lake expansion after 89 ka.

Zone E4 (~87–80 ka) in the Ana River Canyon record overlaps the lower half of Zone B (~85–74 ka) in the Wetlands Levee record. Ana River Canyon pollen records the beginning of an expansion of desert scrub vegetation, primarily saltbushes (Chenopodiineae) during this period. The concurrent rise of rabbitbrush, horsebrush, and grass pollen suggests a decline in annual precipitation and a shift from winter towards summer precipitation. Increasing saltbush pollen during this period overwhelmed the long distance pollen contribution, so that pine declined, and juniper and sagebrush even disappeared from the local record. Algae also drop from the record. The Wetlands Levee record indicates that regionally elements of the cool-moist montane forest (pine, spruce, fir and western hemlock) declined. However, juniper and its associated understory expanded. This may in part reflect the response of vegetation along the Winter Ridge to the shift towards summer rainfall. Increased saltbush pollen in the Wetlands Levee record reflects the expansion of desert scrub in the Ana River delta area. The Cool-Moist Index constructed from cool-moist

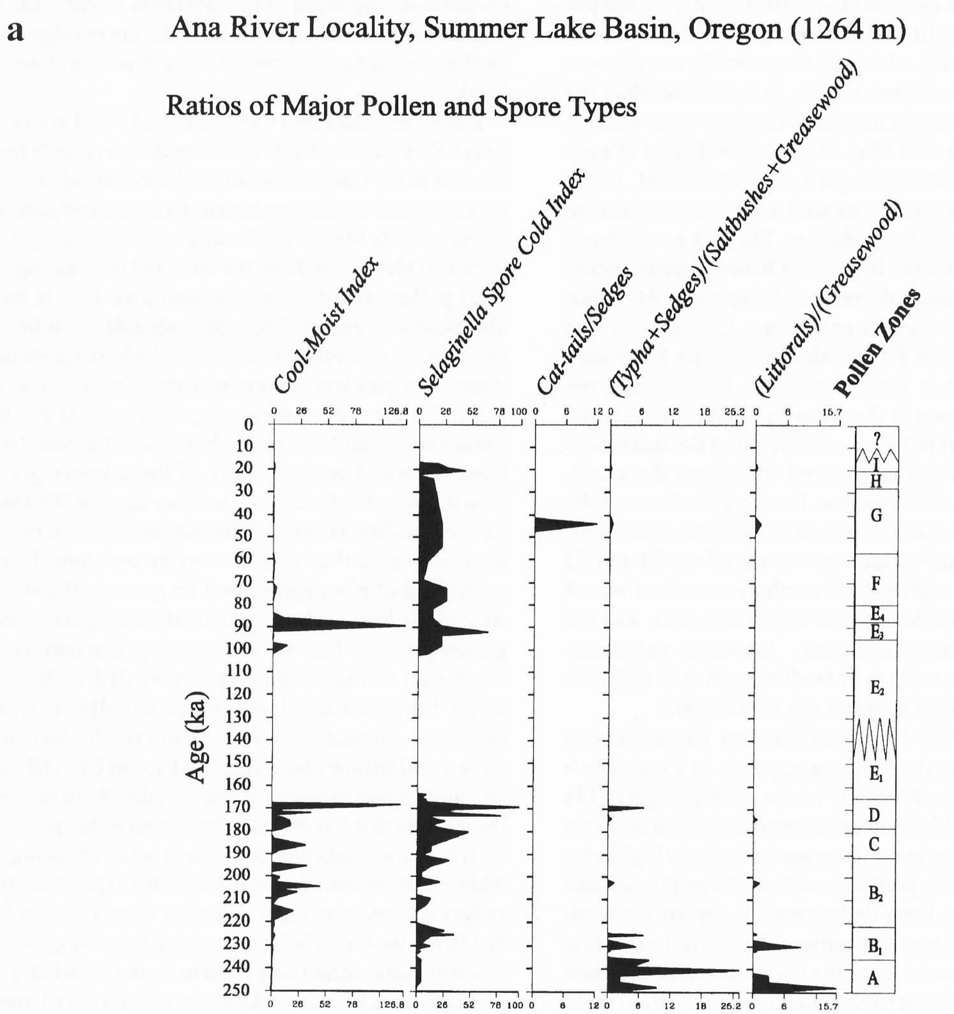
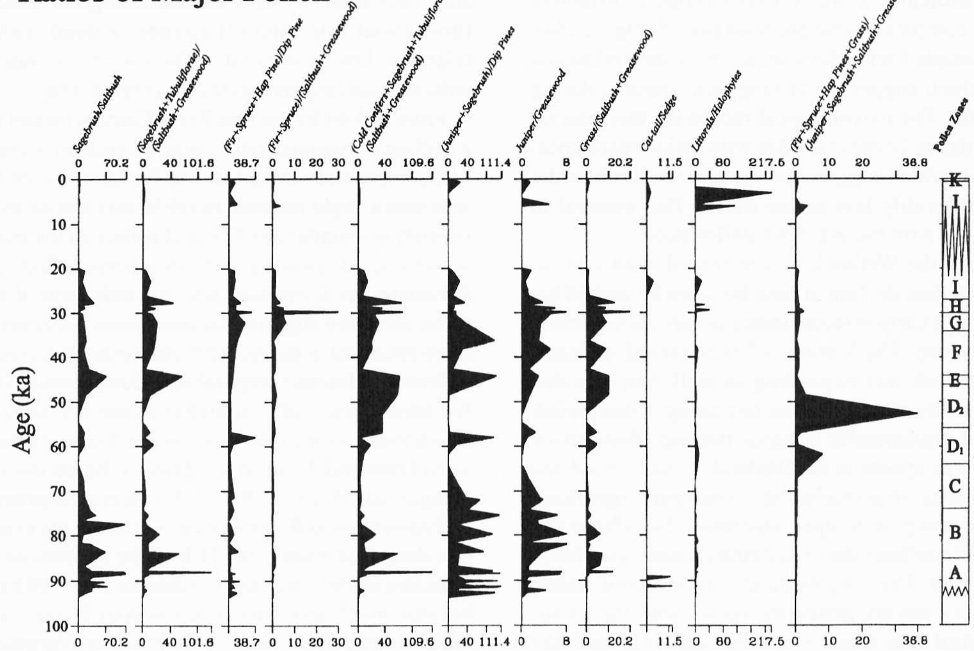


Figure 15. Ratios of major pollen and spore types. (a) Ana River Canyon core; (b) Wetlands Levee core; (c) Bed & Breakfast core.

b Wetlands Levee Locality, Summer Lake Basin, Oregon (1264 m)
Ratios of Major Pollen



c Bed and Breakfast Locality, Summer Lake Basin, Oregon (1264 m)
Ratios of Major Pollen and Algae

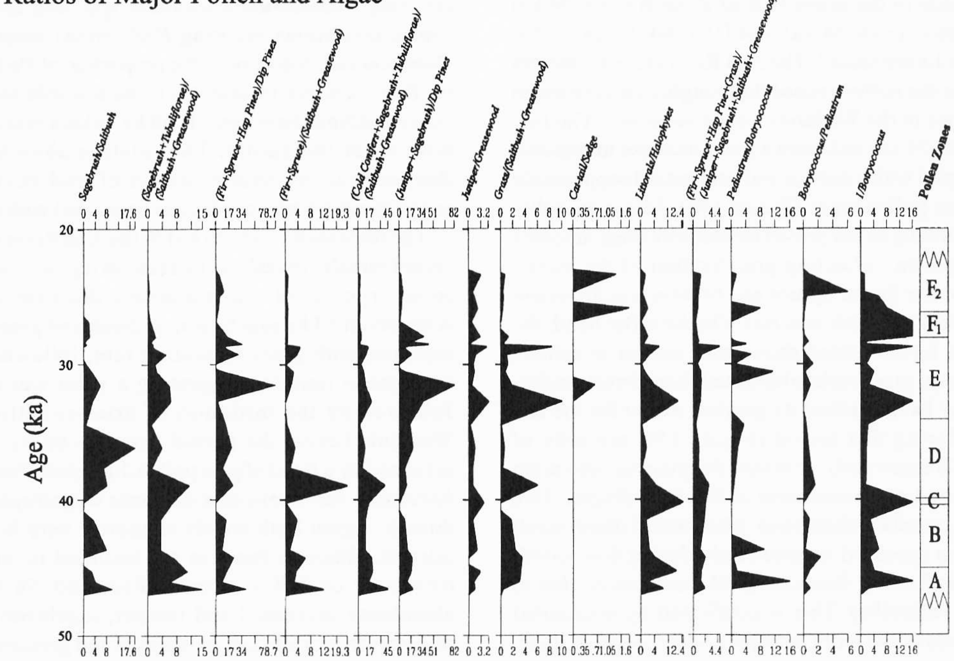


Figure 15. Continued

plant pollen versus warm-dry plant pollen at Ana River Canyon indicates a precipitous decline in effective moisture compared to the previous period (Figure 15a). This is coupled with the indication of the spikemoss record which suggests warming temperatures during that period. The palynological record for this interval from Wetlands Levee is at odds with paleolimnological indicators, which suggest that the lake was both colder and considerably less saline during this interval in comparison with the previous pollen zone.

Whereas the Wetlands Levee record shows an increase and then decline in pine between 83 and 80 ka, the Ana River Canyon record indicates only the beginning of an increase. The Wetlands Levee record indicates that sagebrush was expanding as well. Marshes that were relatively restricted at the beginning of this period, expanded dramatically towards the end. *Pediastrum boryanum* re-appear in the Wetlands Levee record and indicate cooler, oligotrophic lake conditions. Significant fluctuations in pine, juniper, sagebrush, Tuuliflorae and grass pollen reflects the underlying climatic variability of this period. Thus, the Wetlands Levee record, unlike Ana River Canyon, generally agrees with the paleolimnological indications of a colder and less saline lake at this time.

Zone F (~80–57 ka) in the Ana River Canyon record corresponds to the upper half of Zone B (~85–74 ka) and Zones C (~74–64 ka) and D1 (~64–56 ka) in the Wetlands Levee record. The Ana River record indicates 3 shifts in the pollen record that roughly correspond to the 3 zones in the Wetlands Levee sequence. The first shift (~80–74 ka) indicates a great increase in regional pine coupled with a decline and near-total disappearance of all other pollen types (Figure 14a). This is probably due to flooding of the pollen record with huge amounts of pine pollen, reflecting proliferation of the mesic, mixed conifer forest. Spruce and fir were less important and western hemlock was rare. On the other hand, the Wetlands Levee record shows that juniper woodland with grassy, sagebrush/rabbitbrush/horsebrush understory may have reached its greatest extent for the last 100 ka during this period (Figure 15b, see ratio of Juniper/Greasewood) . Marshes fringing the lake seem to have been more extensive at this time (Figure 15b, ratio of Littorals/Halophytes). Elements of desert scrub vegetation appeared intermittently during this period, perhaps reflecting fluctuating lake conditions due to climatic variability. This is confirmed by occasional abundances of the more eutrophic lake condition algae, *Botryococcus* and *Pediastrum simplex*. However, the proportions of *Botryococcus* and *Pediastrum* indicate a trend from shallow lake (80-77 ka) to generally deep, cold lake after 77 ka. Paleolimnological indicators throughout this interval suggest a deep, cold and relatively low-salinity lake. The spikemoss index also indicates cooling temperatures (Figure 15a).

From ~75-64 ka the Ana River Canyon record shows a decline in regional pine, coupled with an increase in both juniper and saltbushes. TheWetlands Levee record indicates a slight increase in saltbushes and an increase in relative abundance of fir and spruce in the montane woodland. Beginning with this period *Pediastrum boryanum* var. *boryanum*, which is indicative of fresher water and more oligotrophic conditions, becomes much more abundant in the record from Wetlands Levee. This reflects conditions comparable to those found in snow-fed lakes about 1000 m higher in the northern Great Basin today, (e.g., Fish Lake in the Steens Mountains or to Diamond Pond in the Harney Basin during the Neoglacial (Wigand,1987)). An increased proportion of *Pediastrum* to *Botryococcus* indicates the existence of a deep lake from ~74–71 ka. The spikemoss index indicates slightly warmer conditions. After 70 ka grass became much more abundant at Ana River Canyon, suggesting an increase in precipitation. Variability in water chemistry is indicated in the alternation of lacustrine algae between more oligotrophic conditions favoring *Pediastrum boryanum* spp. and more eutrophic conditions favoring *Pediastrum simplex* and *Botryococcus*. A decline in the proportion of *Pediastrum* to *Botryococcus* indicates a trend towards shallower lake conditions between 71–65 ka, which is consistent with other indicators. The relative abundance of *Botryococcus* (an inverse measure of productivity) also suggests colder lake conditions coincided with this fall.

The third shift (~65–56 ka) in the Ana River Canyon record initially reveals an increase in regional pine (and some fir) coupled with a dramatic decrease in grass. After about 63 ka sagebrush, saltbush and greasewood increase, with grass remaining rare, followed by an increase in juniper, suggesting a short wet interval followed by the initiation of drier conditions. At Wetlands Levee, the period from 64–60 ka is characterized by a flood of pine pollen. Increased *Pediastrum boryanum* var. *boryanum* suggests oligotrophic conditions. Again both trends support a very brief wet interval, although this was not recorded in either the ostracode or Sr/Ca records. From 60–56 ka pine abundance decreased and juniper, sagebrush, grass, rabbitbrush, horsebrush, saltbushes and greasewood all increased in abundance, suggesting the beginning of drier conditions. Very poor preservation of *Pediastrum* fossils during this period suggests that they may have been redeposited from older sediments, consistent with sedimentological evidence for continued lake level decline. A declining proportion of *Pediastrum* to *Botryococcus* also indicates the end of deep lake conditions around 60–57 ka. The spikemoss index indicates warmer temperatures with gradually cooling temperatures at the end of this period, consistent with the Mg/Ca record.

Zone G (~57–~44 ka) in the Ana River Canyon record is equivalent to Zone D2 (~56–46 ka) and E (~46–44 ka) in the Wetlands Levee record, and overlaps into Zone A (> 47–46 ka) in the Bed & Breakfast Locality. The Ana River Canyon record indicates that this period is one of expansion of juniper woodland and its sagebrush understory, and of desert scrub vegetation. About 47–44 ka fir signals a brief episode of wetter conditions. A peak in the Cat-tail/Sedge, (Cat-tail+Sedge)/(Saltbushes+Greasewood), and the Littorals/Greasewood ratios at this time reflects a rebirth of marshes in the area. However this increasing precipitation signal is not observed in any of the lithologic, geochemical or ostracode indicators. In the Wetlands Levee record an anomalous major expansion of spruce is evident at the beginning of Zone D2. Above this sample the expansion of juniper and sagebrush reflects the continued importance of juniper woodland and sagebrush steppe. Increased fir corresponds to the brief wet episode seen in the Ana River Canyon record around 47–44 ka. Marsh pollen and lacustrine algae disappear from both the Ana River and Wetlands levee records. The Bed & Breakfast Locality record is similar to the Wetlands Levee sequence of terrestrial vegetation, i.e. regional dominance of montane pine forest and lower elevation juniper woodland with grassy sagebrush understory (Figure 14c). Abundant sedge and rarer cat-tail pollen at Bed & Breakfast, but its absence from the Wetlands Levee, suggest that a marsh may have occurred sporadically along the shore of pluvial Lake Chewaucan. Presence of *Pediastrum simplex* and *Botryococcus* suggests slightly warmer, more eutrophic lake conditions. Geochemical data for this time interval however do not provide evidence for a warming trend.

Sediments equivalent to Wetlands Levee Zones F (~46–34 ka), G (~34–30 ka), and H (~30–27 ka) are absent from the Ana River Canyon record. However, Zone F from the Wetlands Levee locality corresponds to Zones B (~46–41 ka), C (~41–40 ka), and D (~40–34 ka) in the Bed &Breakfast Locality record. Zone F of Wetlands Levee records less pine pollen than earlier and declining fir and spruce in the montane forest. Sagebrush-rich juniper woodland was still wide-spread, but sagebrush, rabbit brush, horsebrush and grasses were more abundant than in earlier periods. Increasing proportions of *Botryococcus* to *Pediastrum* indicate shallow lake conditions ~38 ka, and declining *Botryococcus* abundance indicates cool lake conditions. Spores of bracken ferns (*Pteridium* sp.), indicating open meadow, appear. Other fern spores (Monolete Spores) appear as well. The Bed & Breakfast record from 46–34 ka is similar to that of Wetlands Levee. The Bed and Breakfast core algal record during this period is quite variable, but generally indicates slightly less fresh water (*Pediastrum simplex*) and more eutrophic (*Botryococcus*) conditions. The spikemoss record from the Ana River Canyon suggests cool, but not cold conditions. Expansion of greasewood at the expense of sedge during this period suggests drier conditions. These data are consistent with sedimentological indications of declining lake levels at this time compared with the previous period, as well as both geochemical and faunal data.

Wetlands Levee Zones G (~34–30 ka) and H (~30–27 ka) correspond to Zone E (~34–27 ka) in the Bed & Breakfast record. The earlier decline in pine in the Wetlands Levee core (the lower portion of Zone E at the Bed & Breakfast Locality) ends in Zone G. A regional expansion of the montane forest suggests a cooler, moister climate. This period is typified by increasing spruce, fir and cold-climate pines with respect to warm-climate pines (Figure 15b, see (Fir+Spruce+Hap Pine)/Dip Pine ratio and (Fir+Spruce)/(Saltbush+Greasewood) ratios). Juniper and grass also increase during this period (Figure 15b, see Juniper/Greasewood and Grass/(Saltbush+Greasewood) ratios). These ratios also indicate a sharp decline in their regional abundance. The record from the Bed & Breakfast locality shows these same trends (Figure 15c, see the (Fir+Spruce+Hap Pine)/Dip Pine and the (Juniper+Sagebrush)/Dip Pine ratios). Marshes with a cat-tail expansion followed by sedge expansion also characterize this period. Declining *Botryococcus* suggests colder temperatures as well. The growth of a deep lake followed this event closely as is registered by an increasing proportion of *Pediastrum* to *Botryococcus* at both the Wetlands Levee and Bed & Breakfast localities.

Zone H in the Wetlands Levee core (the upper portion of Zone E at the Bed & Breakfast Locality) marks a severe regional drought. Regional decline in the montane forest, and grassy, juniper woodland coincides with major expansion of desert scrub vegetation, dominated by saltbushes and greasewood.

Grasses in particular declined dramatically. This event can be traced up to 250 km to the southeast near Pyramid Lake where dramatic increases and decreases of grass pollen mirror the pattern in the Summer Lake Basin (Wigand & Rhode, submitted). The marsh record of the Wetlands Levee locality indicates that a sedge-dominated marsh shrank as well in response to the drought (Figure 15b, see Cyperaceae).

Zone H (~27–21 ka) in the Ana River Canyon record and Zone I (~27– ~ < 21 ka) in the Wetlands Levee record correspond to Zones F1 (~ 27–26 ka) and F2 (~26–23 ka), the 2 upper most zones in the Bed & Breakfast record. This zone is typified by an increase in cool-moist montane forest species. Pine expansion is recorded at all three localities. Increased fir and to a lesser extent spruce is recorded both at the Wetlands Levee and Bed & Breakfast localities. However, the regional increase in juniper recorded in both the Wetlands Levee and Bed & Breakfast localities seems to be relatively greater (see the Juniper/Greasewood ratio, Figure 15b). Although an expansion of sagebrush appears to be recorded at the Wetlands Levee and Ana River Canyon localities, such evidence does not appear in the Bed & Breakfast locality record. A regional increase in grass recorded at all three localities suggests wetter conditions just prior to the onset of the Glacial Maximum. A brief initial increase in Cat-tail/Sedge and Litoral/Halophyte ratios ~24 ka (Figure 15b) followed by an increase in the proportion of *Botryococcus* to *Pediastrum* by ~21 ka (Figure 15c) records first marsh expansion and then expansion of a shallow lake. This is consistent with paleolimnological indications of declining lake salinity. The 1/*Botryococcus* ratio suggests that this shallow lake was cold (Figure 15c). The spikemoss ratio suggests initially cool terrestrial conditions ~27 ka, followed by a warm episode contemporary to the initial rise of the lake ~23 ka, followed by a strong switch to cold conditions by ~21 ka (Figure 15a). However, palynological indications of cooler conditions from 27–23 ka are at odds with geochemical indications, which suggest warmer summer water temperatures throughout this interval (and continuing during the 23–21 ka period, when all records are concordant for warmer conditions).

Zone I (~21– < 17 ka), the top zone in the Ana River Canyon record, has no equivalent in the Wetlands Levee record. The vegetation record from this sample suggests a very cold-dry Glacial Maximum. The pine dominated montane forest had contracted, and the grassy, sagebrush steppe had expanded at its expense. An expansion of greasewood vegetation may have been drowned by a re-expansion of the lake, indicated by the sudden appearance in the record of the eutrophic lake indicator *Pediastrum simplex*. This is consistent with the ostracode records. Pollen from the adjacent shadscale shrub community (saltbushes and greasewood) predominates. By the end of this period it appears that a sagebrush – and grass-dominated, shrub-steppe community covered the nearby slopes or delta, perhaps mixed with some saltbushes. The spikemoss spore record indicates extremely cold conditions.

Zones J and K in the Wetlands Levee record have no equivalents in either the Ana River Record or in the Bed & Breakfast locality. Given the appearance of the pollen (Figure 14b), ratio (Figure 15b), and the zoning diagrams (Figure 14b)one of us (P.W.) suspects that the top 4 samples of the Wetlands Levee core should be assigned to the middle to late Holocene. Samples 2,3 and 4 appear to continue trends such as the increase in fir and spruce that typify the onset of the Glacial Maximum but they probably owe their nature to the climatic conditions present in the Summer Lake Basin during the Neoglacial and Little Ice Age (Figures 14b and 15b). It is clear that the top pollen sample, which clearly reflects the current vegetation in the area, is Holocene and probably recent in age.

The vegetation record at Summer Lake bears some resemblance to the record from Carp Lake in eastern Washington (Whitlock & Bartlein, 1997). However, the conditions frequently do not correspond. This is unsurprising, since storm tracks deflected far enough northward to make the Carp Lake area wetter than today are probably too far north to affect the Summer Lake area. When the Summer Lake area was subjected to more humid conditions, the Carp Lake area probably was subjected to drier conditions. Temperature is more likely to have been similar at the two areas. An additional problem in comparing the lower portion of both the Summer Lake and Carp Lake records arises from conflicting age assignments for volcanic ashes and ages extrapolated from these deposits.

Synthesis

Lake high stands, declining salinity, and palynological indications of progressively moister climate occurred at Summer Lake between ~236 and ~165 ka (with a brief reversal towards more saline conditions around 200 ka). A cooling trend is also indicated by all proxies from 236 until 165 ka. Indications of drier and warmer conditions are apparent in most or all records from 165 until at least 158 ka, but this episode is of uncertain duration, as the record is truncated by a major unconformity above this (158–118 ka). The duration of the lake level lowstand associated with the major unconformity is unknown, since it is evident from irregular topography across the unconformity surface that significant erosion and or deflation of pre-Stage 5 sediments has occurred. Dry conditions prevailed from 118–90 ka. However from 118–89 ka. there is little consistency between paleotemperature proxy records. All records indicate moister climates from ~89–~50 ka. Most records suggest cold conditions prevailed throughout much of the 89–17 ka interval, although significant deviations from this are indicated by several warming excursions in the palynological data. Strong indications of warm interludes, evident in all lacustrine signals between ~100 and 89 ka, and again from 27–23 ka, are only partially supported by the pollen data. Drier conditions are consistently indicated for the period 40–32 ka. Latest Pleistocene and most Holocene sediments have been removed from the core and outcrop sites by deflation of the basin, but Late Wisconsin lake deposits are well preserved in high elevation beach terraces around the basin margin.

The paleoenvironmental history of lake level, salinity fluctuations and vegetational change at Summer Lake over the past 250 ka bears intriguing similarities to the global record of ice volume recorded in the marine oxygen isotope record (Figure 16) (Martinson et al., 1987). The timing of all indicators of freshwater conditions, deep lakes and more mesic vegetation generally coincide with oxygen isotope Stages 6, 4 & 2. Warm climate indicators above a major unconformity correspond with Stage 5. Note that these age estimates from Summer Lake are based solely on data completely independent of the marine record. No theoretical preconceptions of when Pleistocene pluvials or vegetation types 'should have occurred' based on curve matching or expectations based on orbital forcing models were used in assembling this record. However, it is difficult to correlate the shorter-term variability in the Summer Lake record with global trends in temperature or precipitation. Deviations in the Summer Lake record from the global record of ice volume can be used to deduce climatic processes that are acting on a more regional to local scale, such as regional deflections in the jet stream. As more records in the region become independently dated it should be possible to refine scenarios of regional climate variability in the Great Basin during recent glacial and interglacial cycles without reference to the global pattern of climate change.

Palynological records do not fully agree among themselves or with lake records with respect to the paleotemperature and paleoprecipitation histories of

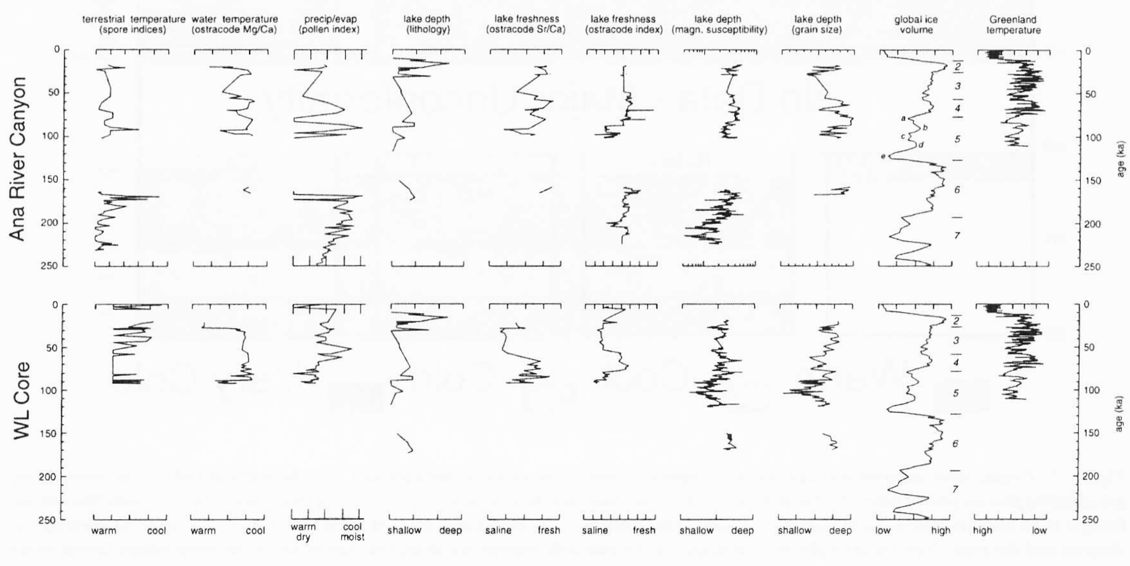


Figure 16. Key paleoenvironmental proxies from the Lake Chewaucan record, compared with the marine oxygen isotope record of Martinson et al. (1987) and the Greenland Ice Sheet isotopic record (Grootes et al., 1993).

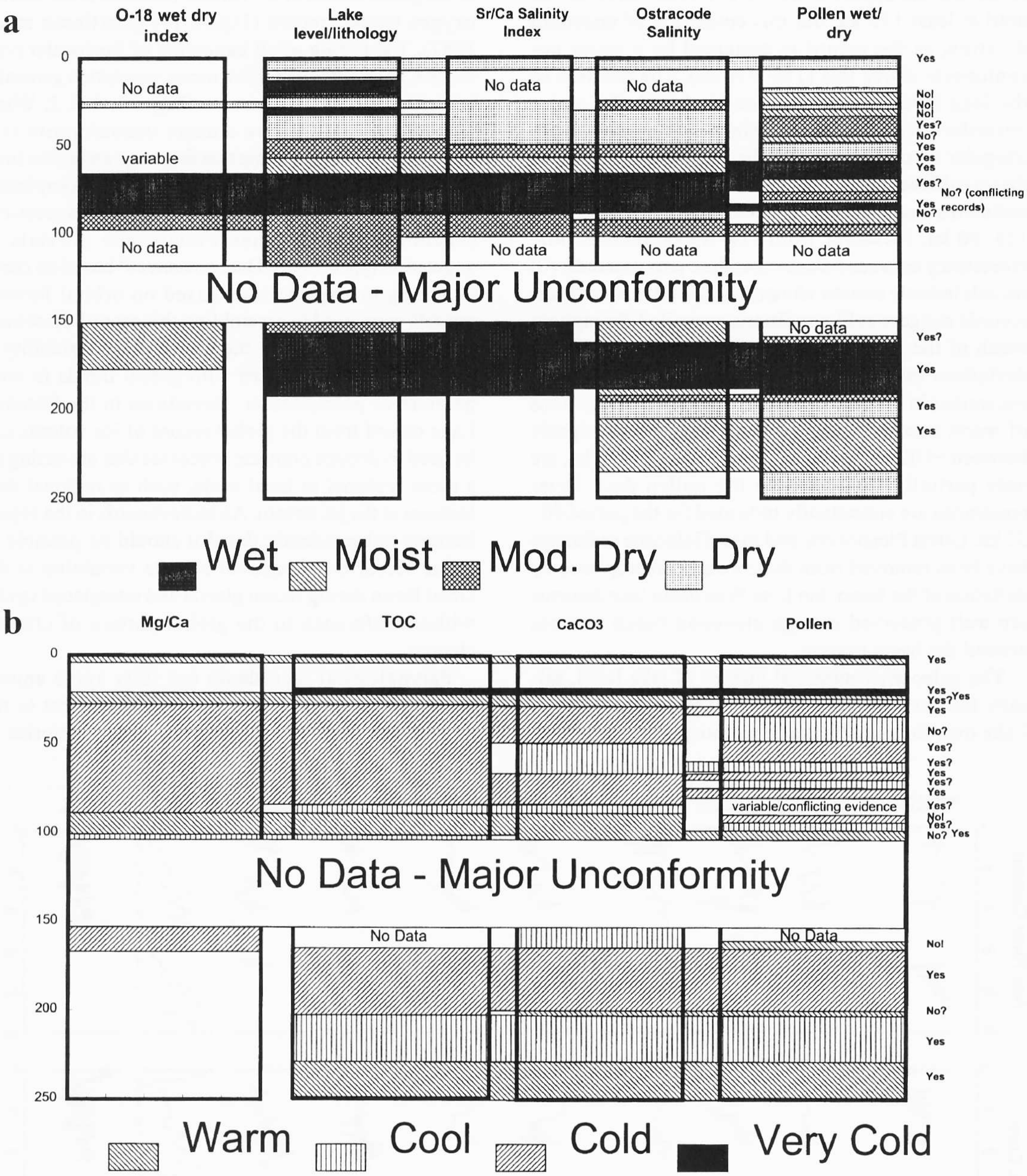


Figure 17. Concordance diagram between major paleoprecipitation and paleotemperature proxies from the Summer Lake basin. Some of the geochemical proxies plotted here (^{18}O, %$CaCO_3$ and %TOC) are known to be related to multiple forcing mechanisms in many contexts, but are thought to be strongly controlled by climate variables in the Summer Lake record as discussed in the text. Continuous patterns crossing the diagram and the term 'Yes' on the right side, indicate that all available proxies are in general agreement as to paleoclimate (either in an absolute sense, or in terms of trends) for that time interval. A break in the horizontal pattern and the term 'No!' on the right indicates that proxies are not in good agreement for the time interval. '?' indicates that the record is ambiguous for key proxies at that time, and 'No?' and 'Yes?' indicate our variable level of confidence in matches between general trends. (a) Paleoprecipitation proxies; (b) Paleotemperature proxies.

Summer Lake (Figure 17). Differences among palynological records between the three coring localities have been affected through time in different ways by their location in the basin. For example, because deposition rate was much greater at the Bed & Breakfast locality than at either the Wetlands Levee and Ana River Canyon localities, the relative position of the lake bottom at that point was changing relative to the other two localities, making them susceptible to very different rates and types of pollen flux. On the other hand the Ana River Canyon locality made the progression through the last 250 ka between a deep lake and marsh to a situation in which it varied between a delta or a desert scrub flat.

Wigand & Rhode (submitted) have discussed the common problem in western North American Pleistocene records of discordance in paleoprecipitation proxies between lake (lithostratigraphy, ostracodes, geochemistry and aquatic palynomorphs) vs. upland pollen records. Their possible explanations for this problem include:

1) Differing senstitivity or response rate of different systems to the same climatic factor. For example, as the Summer Lake Basin became infilled by sediment the amount of water required to attain a given lake level elevation would have varied. Thus inundation of lake margin localities during Isotope Stage 6 high stands may have required more precipitation than during Stage 4. Increasing subsidence rates over this time period would have the opposite effect.
2) Variable responses of different systems to different parameters within 'climate space'. For example, during the more extreme periods of the late Glacial, much of the upper portion of the Winter Ridge was probably covered with snow. Gearhart Mountain (2549 m), which lies south of Summer Lake and served as the source of the Chewaucan River, had glaciers on it down to at least 1830 m, only 566 m (1800 ft) above the current lake surface. This does not leave much space for vegetation between high lake stand and lowest snow line (even less than 566 m, which is based upon the current lake stand). Regardless of precipitation/evaporation levels during such intervals (which would determine lake level and salinity) vegetation records would reflect a substantial reduction in vegetational cover (and presumably pollen flux).
3) Presence of local or regional factors affecting 1 system but not the other (e.g. drainage capture).
4) Data gaps in 1 or more proxy records.
5) Uncertainties about relationships between proxy records and underlying climate controls, or
6) Some combination of the above.

An additional possibility not discussed by Wigand & Rhode (submitted) is that lapse rates or relative evaporation/precipitation ratios across elevational gradients varied through the Pleistocene. This may have resulted in upland areas (where lake-bound moisture would have been derived) being considerably wetter at some times than the areas immediately around the lake (from which most of the pollen signal would have been derived).

None of these possibilities can be eliminated in the specific case of Summer Lake. Because of the broad coincidence between the episodes of high lake levels and lowered salinity with the global glacial/interglacial record over the past 250 ka (Figure 16), we believe that this hydroclimate record is probably a good reflection of actual regional precipitation patterns. However, alternative possibilities cannot be ruled out. The third possibility listed above is particularly intriguing in light of two observations.

First, the Chewaucan River, a major basin inflow, periodically is diverted across the Upper Chewaucan marshes into either the Summer Lake or Lake Abert basin. Given the present elevation of that drainage split (~1315 m) and our interpretation that lake levels were unlikely to have been that high in Summer Lake basin prior to the late Wisconsin, a drainage capture scenario cannot be eliminated as a contributing factor in determining lake salinity and level. We will need a comparable stratigraphic record from Lake Abert before the Summer Lake depositional history can be confidently interpreted as resulting from climate rather than drainage diversion history.

Second, during various parts of the Pleistocene, various authors have postulated changes in the prevailing winds for the northern Great Basin (COHMAP Members, 1988; Thompson et al., 1993). When continental and Cordilleran ice was in place prevailing winds may have come from the east-northeast rather than from the west-southwest. This would change the complexion of the regional pollen record. During periods when the winds were from the east-northeast sagebrush pollen would be more prominent in the pollen record. When it was blowing from the west-southwest across the Cascades pine pollen would have been more prominent in the pollen record.

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