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Chronology of latest Pleistocene lake-level fluctuations in the pluvial Lake Chewaucan basin, Oregon, USA

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ABSTRACT: New accelerator mass spectrometer radiocarbon ages from gastropods in shore deposits within the pluvial Lake Chewaucan basin, combined with stratigraphical and geomorphological evidence, identify an abrupt rise and fall of lake level at ca. 12 ^{14}C ka. The lake-level high is coeval with lake-level lows in the well-dated records of palaeolakes Bonneville and Lahontan, and with a period of relatively wet conditions in the more southerly Owens Lake basin. This spatial pattern of pluvial lake levels in the western USA at 12 ^{14}C ka indicates a variable synoptic response to climate forcing at this time. Copyright © 2001 John Wiley & Sons, Ltd.

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KEYWORDS: pluvial lakes; Summer Lake; palaeoclimatology; geomorphology; Great Basin.

Introduction

Palaeolake-level fluctuations are one of the best sources of high-resolution palaeoclimate data in arid regions of the western USA (Smith and Street-Perrott, 1983). Numerous workers have catalogued lake-level data and proposed lake-level chronologies for various lake systems in the Great Basin (e.g. Mifflin and Wheat, 1979; Smith and Street-Perrott, 1983; Benson and Thompson, 1987; Benson *et al.*, 1990; Oviatt *et al.*, 1992; Negrini, in press). These high-resolution records of palaeolake-level fluctuations provide strong evidence that large and rapid climate oscillations occurred at millennial time-scales in the western USA during the last glaciation (Allen and Anderson, 1993; Phillips *et al.*, 1994; Benson *et al.*, 1995, 1996a, 1998a,b; Oviatt, 1997; Benson, 1999). Efforts to correlate these lake histories with climate oscillations elsewhere are currently limited by uncertainties in age control (Benson, 1999). Nevertheless, an investigation of possible linkages between climate records in the western USA and other regions is crucial for further evaluating mechanisms of climate change in these regions (Clark and Bartlein, 1995; Benson *et al.*, 1998a; Hostetler and Bartlein, 1999).

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At its maximum Pleistocene highstand of 1378 m, pluvial Lake Chewaucan in south-central Oregon was 114 m deep and covered 1244 km^2, filling four sub-basins now occupied by modern Summer Lake, Upper Chewaucan Marsh, Lower Chewaucan Marsh and Lake Abert (Fig. 1) (Allison, 1982). Using elevations of prominent shoreline features, Allison (1982, and references therein) found evidence for at least six distinct stages in the history of Lake Chewaucan following the maximum highstand at 1378 m. The Chewaucan basin has continued to be the site of numerous palaeoclimate and related studies because of its excellent exposures of palaeolake sediments and numerous tephra layers, its long and well-preserved record of sediment magnetism and the abundance of fossils (e.g. Davis, 1985; Palacios-Fest *et al.*, 1993; Freidel, 1993; Cohen *et al.*, 2000; Negrini *et al.*, 2000). Despite the impressive wealth of palaeoclimatic data that have emerged from studies in the Chewaucan basin, few data constrain the timing of lake-surface fluctuations during the latest Pleistocene and Holocene.

This paper presents new radiocarbon data combined with stratigraphical and geomorphological evidence that help constrain the latest Pleistocene lake-level history of pluvial Lake Chewaucan. Refining the lake-level history in the Chewaucan basin is an important research objective because its peripheral location in the extreme northwest part of the Great Basin provides a significant geographical extension of lake histories required to test models of climate change across the western USA. Details of the lake history also have a critical bearing on archaeological (e.g. Pettigrew, 1985; Oetting, 1994) and tectonic (e.g. Simpson, 1990; Pezzopane and Weldon, 1993) studies of the region.

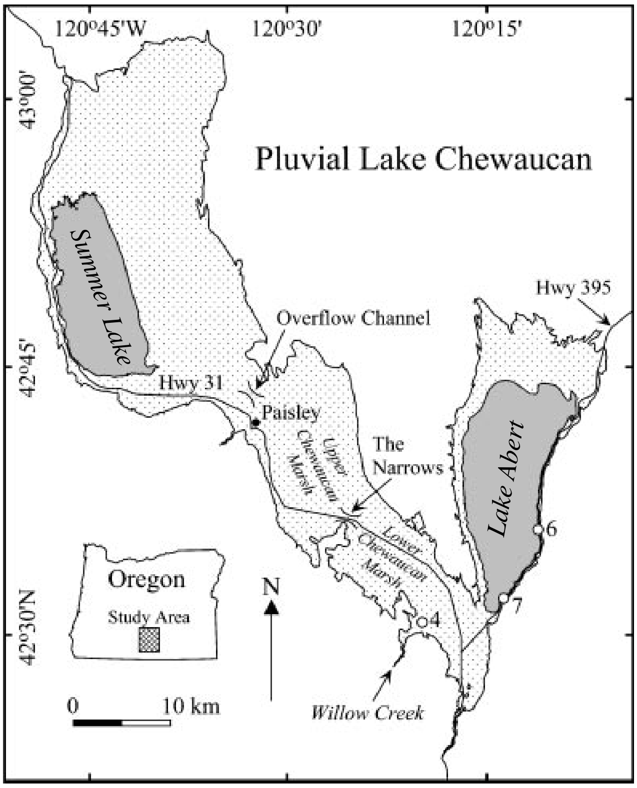

Figure 1 Location map for the Chewaucan basin. The extent of pluvial Lake Chewaucan at its maximum 1378 m highstand is shown as light stipple, and the modern remnants Summer Lake and Lake Abert are shown as darker stipple. Stratigraphical section locations are marked with numbered circles

Methods

Field work for this study consisted of sedimentological description and measurement of lake sediment exposures at various localities throughout the basin (Fig. 1). Several of the exposures contain fossiliferous shore deposits interbedded with shallow and deeper water sediments. Accelerator mass spectrometer (AMS) radiocarbon analyses were performed on fossil gastropods collected from these shore deposits and from other sedimentary facies associated with former lake-levels. Elevations of shore deposits and fossil collection sites were measured with reference to modern lake-surface elevations or nearby benchmarks, using a Jacob staff and clinometer, and probably are accurate to within 1 m.

Results

The four AMS radiocarbon ages reported in this paper (Table 1) identify the age of latest Pleistocene shoreline features. Additional sample details are compiled in the Appendix.

Latest Pleistocene fan-delta in Lower Chewaucan Marsh

Sample 4F (11 930 ± 90 ^{14}C yr BP) is a collection of gastropods from a 5.25 m high exposure (‘section 4’) of fine-grained sediments in a bluff along Willow Creek in the Lower Chewaucan Marsh sub-basin (Fig. 1). The basal zone of the section measured is dominated by massive to weakly laminated clay and silt, and coarsens upward into thinly bedded silt and sand. Gastropods and pelecypods are sparse throughout the section. The gastropods chosen for radiocarbon analyses were sampled from a silt bed approximately 1.65 m from the top of the section at an elevation of 1325 m. The bluffs along Willow Creek in the vicinity of section 4 contain exposures of sand units with climbing-ripple cross-lamination, normally graded gravely channel fill units that truncate the laminated silts and sands, and alluvial gravels. Principal bedding planes within the units are inclined gently (>10°) to the northeast (toward the basin), and the deposits underlie a low-gradient fan surface that extends into the Lower Chewaucan Marsh.

The geomorphological expression and facies associations of these deposits are interpreted as an assemblage of lacustrine and alluvial deposits within a fan-delta that prograded into a former lake at the mouth of Willow Creek. In this interpretation, the finer grained lacustrine strata formed in the distal prodelta, and the coarser sands and silts were deposited in the proximal prodelta and delta-front regions. The channel-fill facies mark the location of distributary channels within the fan-delta complex, or may have formed during subsequent dissection of the deltaic and lacustrine units as the lake-level fell and the fan-delta was abandoned. The development of a prograding fan-delta at the emergence of Willow Creek into the former lake is a probable consequence of high sedimentation rates at the creek mouth during a stable lake-level. The surface elevation of 1335 m near the apex of the fan-delta suggests that the lake-level stabilised near this elevation during much of the period of fan-delta construction, hence the radiocarbon age is considered to date a lake-level that is close to 1335 m.

Table 1 Radiocarbon data from Lake Chewaucan

Field sample number	^{14}C Laboratory number	Elevation (m)	^{14}C yr BP[a]
4F	AA13588	1325	11 930 ± 90
6AV	AA13589	1310	11 670 ± 90
7EG	AA13590	1307	11 560 ± 120
7FG	AA13591	1310	12 030 ± 90

[a] Errors are $\pm 1\sigma$.

Latest Pleistocene shorelines at Lake Abert

Samples 6AV, 7EG and 7FG are gastropods found in shore and nearshore deposits in the Lake Abert sub-basin that document a lake-level at about 1310 m. Sample 6AV (11 670 ± 90 ^{14}C yr BP) is from a 2.9 m high road-cut exposure (‘section 6’) through the outer edge of a wave-formed terrace on the east shore of Lake Abert, which contains two stratigraphical units (Figs 1 and 2). The basal 1.1 m of the section measured (unit Qs) consists of well-rounded medium sand with occasional rounded cobbles and boulders, and is interpreted as a nearshore facies composed of sand that was transported offshore by wave action. Gastropods, pelecypods and ostracods are concentrated in pockets and layers that generally are present throughout the unit. Unit Qs has a sharp upper contact with a 1.8 m thick unit (Qg_2) of well-rounded, clast-supported boulders, cobbles and pebbles, with a sandy matrix. Unit Qg_2 represents a beach gravel composed of wave-modified colluvial material derived from the Abert Rim scarp directly uphill from the section. Sample 6AV was obtained from a layer

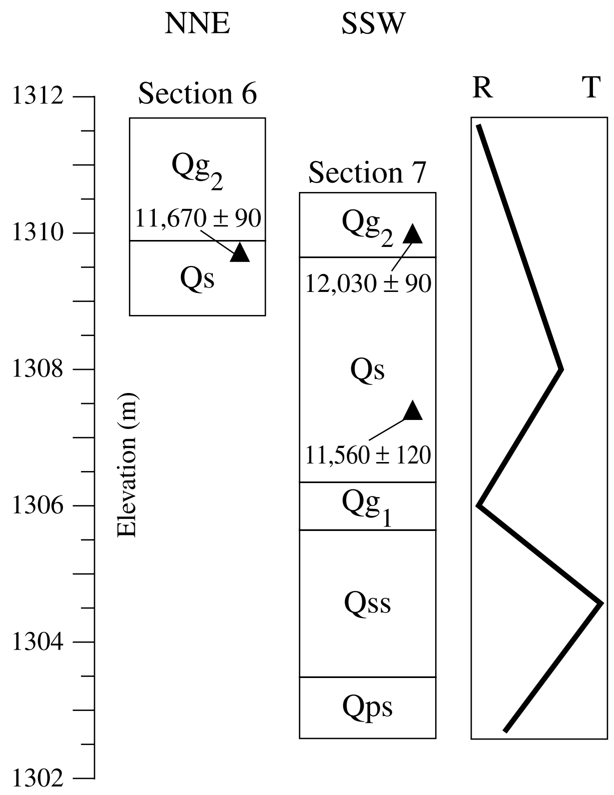

Figure 2 Measured stratigraphical sections 6 and 7 (see Fig. 1 for locations), with a schematic curve of regressive (R) and transgressive (T) lake cycles. Labelled stratigraphical units are described in the text. Black triangles mark the location of radiocarbon-dated gastropods. Ages are in ^{14}C yr BP

of gastropods in unit Qs located about 0.2 m below the base of the Qg_2 beach gravel.

Samples 7EG (11 560 ± 120 ^{14}C yr BP) and 7FG (12 030 ± 90 ^{14}C yr BP) are from an 8.0 m high road-cut exposure ('section 7') through the outer edge of a wave-formed terrace located approximately 8 km south-southwest of section 6 along the east shore of Lake Abert (Fig. 1). Section 7 contains five distinct stratigraphical units that may represent two transgressive–regressive cycles of pluvial Lake Chewaucan (Fig. 2). The basal 0.9 m of section 7 (unit Qps) consists of rounded to subrounded medium to fine sand interbedded with imbricated pebbly strata and occasional cobbles. Unit Qps is interpreted as evidence for a wave-worked nearshore or beach environment with a northward longshore current, as indicated by the pebble imbrication. The poorly sorted character of unit Qps suggests storm-influenced episodic deposition of coarser and finer strata. This unit is overlain by 2.15 m of thinly bedded silt and fine sand with sparse pebbles (unit Qss), which is interpreted as offshore sediment that records a lake transgression. The occasional pebbles in unit Qss may be dropstones that were carried offshore by lake ice. Unit Qss is in sharp contact with an overlying 0.7 m thick unit (Qg_1) of well-rounded, clast-supported boulders, cobbles and pebbles, which is interpreted as a beach gravel facies. Above unit Qg_1 is a 3.3 m thick unit (Qs) dominated by well-rounded, fine- to medium-grained sand with occasional pebbles, which is interpreted as a nearshore sand facies. Unit Qs is overlain by a gravel unit (Qg_2) that forms the uppermost 0.95 m of the section, and has a sharp lower contact with the underlying sands. Unit Qg_2 is matrix-supported for the lowermost 0.4 m, and grades upward into clast-supported and increasingly angular material. The matrix-supported basal portion of the gravel contains a sand matrix with occasional fossils. The unit Qg_2 represents another beach gravel that grades upward into colluvium-dominated material. Units Qs and Qg_2 in section 7 are presumed to be correlative with units Qs and Qg_2 in section 6 on the basis of their very similar sedimentary character and stratigraphical position (Fig. 2).

Primary bedding planes throughout section 7 are inclined gently (ca. 10°) toward the lake, suggesting a primary depositional gradient in the foreshore and shoreface environments. Sample 7EG was collected from a fossil-rich bed in unit Qs approximately 2.25 m below the base of the Qg_2 gravels, and sample 7FG was collected from the sandy matrix of unit Qg_2, approximately 0.35 m above its base.

The elevation of terrace shoreline angles above sections 6 and 7 are 1318 m and 1320 m, respectively (±1 m). The terrace segment containing section 6 is discontinuous with that which contains section 7, but these two segments are considered correlative on the basis of their nearly equivalent elevations. The shoreline angles may mark the approximate level of the lake during construction of the wave-formed terrace (cf. Gilbert, 1890). Stratigraphical relationships suggest that the beach gravels capping the terrace segments are regressive deposits (Fig. 2), and the ca. 1310 m elevation of their exposure at the edge of the terrace probably marks the lake-level near the time of terrace abandonment. The radiocarbon ages associated with the gravels therefore are considered to date a lake-level that is close to 1310 m.

Discussion

Sources of uncertainty in gastropod ages

Although a large body of previous work has shown that gastropod shells can in many cases provide reliable radiocarbon ages for constraining palaeolake chronologies, shell ages are subject to a variety of potential sources of error that requires careful consideration in interpretations. These include incorporation of old carbon introduced by groundwater that mixes with modern lake water (the so-called hardwater or reservoir effect), contamination by young carbon when modern carbon is added to a porous carbonate by recrystallization of the original shell material, and biological fractionation of carbon isotopes (Benson *et al.*, 1990, 1995; Oviatt *et al.*, 1992; Benson, 1993).

Numerous workers have considered the radiocarbon reservoir effect in detail (e.g. Benson, 1978, 1993; Benson *et al.*, 1990; Oviatt *et al.*, 1992; Lin *et al.*, 1998), yet the precise magnitude of the effect in Great Basin palaeolakes remains poorly understood, and the necessary age correction may vary widely from basin to basin. Benson (1978) estimated that the correction for Lake Lahontan is less than 200 yr, or comparable to analytical uncertainties. Subsequent calculations performed by Benson (1993) suggest that the magnitude of the reservoir effect was greatly minimized relative to the present in large Pleistocene palaeolakes, and indicate that reservoir corrections were probably negligible during the Lake Lahontan highstand. Lin *et al.* (1998) prefer a larger reservoir correction for carbonate ages in the Lahontan and Searles basins (300–700 yr), based on different assumptions for ^{14}C-free CO_2 fluxes from springs and CO_2 exchange rates. Discrepancies between their paired ^{14}C–^{230}Th ages (the calibrated ^{14}C ages are generally ca. 1000 yr younger than the ^{230}Th ages), however, suggest that overestimated reservoir corrections may be contributing to the age disagreement, although other factors such as young carbon contamination or uncertainties in the ^{230}Th ages also could be responsible for the discrepancies. Oviatt *et al.* (1992) considered that errors resulting from the reservoir effect in the Bonneville basin probably do not exceed analytical errors in most cases, and they did not apply a correction to their carbonate ages. Because of the difficulties in establishing precise late Pleistocene reservoir ages, and in view of work from these other Great Basin lakes that suggest palaeoreservoir effects most probably fall within the range of analytical error, I have chosen to not correct the radiocarbon ages reported here for a reservoir effect. Application of a correction would result in systematically younger ages.

Post-depositional contamination with young carbon has the potential to cause significant non-systematic age errors (Benson *et al.*, 1990, 1995; Oviatt *et al.*, 1992; Benson, 1993). In gastropods, this type of contamination commonly occurs via replacement of the original aragonite with secondary calcite. Analyses of shells from each dated collection site by X-ray diffraction detected negligible amounts of calcite, which provides evidence that contamination is unlikely to be a problem for these samples. Even for shells that have been properly assessed for recrystallization and corrected for fractionation, however, gastropod ages from other sites have been observed to yield anomalously old or young ages in some cases (Oviatt *et al.*, 1992). Indeed, the stratigraphical reversal in age between samples 7EG and 7FG observed in section 7 (Table 1) is difficult to explain. Aside from this minor age reversal, however, the relative coherency of the ages argues against serious problems with young carbon contamination.

Tectonic concerns

Vertical displacement and warping of palaeolake shorelines in the tectonically active Central Oregon fault zone (Pezzopane and Weldon, 1993) has the potential to complicate the reconstruction of past lake levels. Pezzopane (1993) found evidence for as much as 14 m of vertical difference in the elevations of the highest shorelines along the base of Abert Rim and its associated fault with respect to the maximum highstand shoreline in other parts of the Chewaucan basin. The documentation here of a shoreline that occurs at a consistent 1310 m elevation over an extensive reach of the Lake Abert sub-basin would appear to allay such concerns for the lower shorelines along Abert Rim, or at least suggests that these lower shorelines are all located in the hanging wall of the Abert Rim fault.

Hypsometric and geomorphological considerations

The 'Paisley fan', a fan-delta at the mouth of the Chewaucan River near the town of Paisley (Fig. 1), was presumably built at, or near, the maximum 1378 m level of Lake Chewaucan (Allison, 1982), and subsequently formed a dam between the lake in the Summer Lake sub-basin ('Winter Lake' of Allison, 1982) and the coalesced lake in the three eastern sub-basins ('ZX Lake' of Allison, 1982). Construction of the Paisley fan diverted the Chewaucan River to the eastern sub-basins, thus cutting off the Summer Lake sub-basin from pluvial Lake Chewaucan's principal source of water. Allison (1982) identified an overflow channel from ZX Lake into Winter Lake across the Paisley fan with an upper intake elevation of 1336 m (Fig. 1). The 1338 m elevation of the bank beside the channel intake suggests that overflow was initiated at this level. A small delta with an upper surface elevation of 1323 m was built at the mouth of the overflow channel into the Summer Lake sub-basin, indicating that ZX Lake was about 15 m higher than Winter Lake at the initiation of overflow. From these geomorphological observations, Allison (1982) argued that:

1 both lakes had receded to lower levels sometime after their separation by the Paisley fan;
2 ZX Lake rose faster than Winter Lake during a subsequent lake transgression, probably owing to a larger water supply via the diverted Chewaucan River.

As noted in a previous section, the surface elevation near the apex of the fan-delta in the Lower Chewaucan Marsh suggests that the lake-level stabilised near 1335 m during much of the fan-delta construction. The observation that the elevations of the upper surface of the fan-delta (1335 m) and of the overflow channel across Paisley flat (1336–1338 m) are nearly the same strongly suggests that construction of the fan-delta was coeval with a stabilisation of lake-level during overflow of ZX Lake into Winter Lake in a transgressive cycle of the lake. If this scenario is correct, then the radiocarbon age from the fan-delta (sample 4F) provides an approximate age for the overflow event.

The 1310 m elevation of the shoreline gravels on the east shore of Lake Abert corresponds almost exactly to the present elevation of the narrow gap (The Narrows) between the Upper and Lower Chewaucan Marsh sub-basins (Fig. 1). The floor of the expansive Upper Chewaucan Marsh basin is very flat, and increases almost imperceptibly in elevation from 1310 m at the outlet in the southeast to 1313 m in the northwest. Given a steady increase in lake volume, this hypsometry would result in a marked decrease in the rate of lake-level rise above 1310 m, owing to a substantial increase in the lake surface-area/volume ratio once all three eastern sub-basins became linked. Moreover, the increased surface-area/volume ratio would enhance the efficacy of evaporation in retarding the rate of lake-level rise above 1310 m. Given a steady *decrease* in lake volume, the same hypsometric and evaporative influences would act to decrease the rate of lake-level recession below 1310 m, and therefore also would promote the occurrence of a stillstand at 1310 m during overall recession. Hypsometric arguments therefore predict that a lake-level oscillation above the 1310 m geomorphological threshold would result in occupation, followed by submergence, then reoccupation of a ca. 1310 m shoreline. This scenario is similar, though not strictly analogous, to the stabilising influence of the bedrock sill at Red Rock Pass, which may have resulted in reoccupation of the Provo shoreline of Lake Bonneville during the latest deep-lake cycle (Sack, 1999).

One must now ask whether the 1335 m and 1310 m shorelines represent stillstands in overall lake recession, or if they are associated with a fluctuation involving a rise and fall of lake level. A resolution to this question is not trivial because the latter scenario implies a more significant climate forcing affecting the Chewaucan basin. Although it is possible that the 1335 m shoreline marks a stillstand in overall lake recession just prior to the occupation of the 1310 m shoreline, geomorphological observations show that its formation is likely to have occurred in response to a climatically induced lake-level rise that culminated in an overflow of ZX Lake into Winter Lake. As for the 1310 m shoreline, geomorphological considerations provide an explanation for how this shoreline could have developed largely as a consequence of hypsometric effects during a steady increase or decrease in lake volume. The stratigraphical sequence in section 7, however, provides evidence that the capping beach gravels (unit Qg_2) at ca. 1310 m are associated with the recessional phase of a lake-level fluctuation (Fig. 2). As discussed in the following section, correlation with lake-level data from the adjoining Fort Rock basin further supports the occurrence of a rise and fall in lake level at ca. 12 ^{14}C ka.

Latest Pleistocene lake-level history of pluvial Lake Chewaucan

The age of 11 930 ± 90 ^{14}C yr BP from the gastropods in the fan-delta (sample 4F) indicates that the lake-level was at least as high as 1325 m at this time, although the upper surface elevation of the fan-delta suggests that the radiocarbon age constrains a lake-level that stabilised closer to 1335 m. At this level, all three eastern sub-basins of Lake Chewaucan would have held a single coalesced lake, and any further rise above 1335 m would have resulted in overflow of the lake across the Paisley fan into the Summer Lake sub-basin. Geomorphological observations summarised in the previous section indicate that such an overflow event probably occurred at this time.

The three radiocarbon ages of samples 6AV, 7EG and 7FG from the shorelines in the Lake Abert sub-basin have a mean of 11 750 ± 180 ^{14}C yr BP, which is adopted as the best age for the 1310 m shoreline. The lake at this level would have constituted a coalesced body of water in the Lake Abert and Lower Chewaucan Marsh sub-basins (Fig. 1). Tufa mounds overlying rounded gravels in the Sawed Horn area at the northeast end of Lake Abert occur in a narrow elevation range centred on 1310 m (Langridge *et al.*, 1995, 1996; Jellinek *et al.*, 1996), suggesting that their formation marks the same event responsible for the beach gravels in sections 6 and 7 further south.

Taken at face value, the radiocarbon ages therefore indicate a stabilisation of lake-level at ca. 1335 m, accompanied by overflow into the Summer Lake sub-basin, and followed by a rapid recession to the 1310 m shoreline. Because the ages of the 1335 m and 1310 m shorelines overlap within uncertainties, however, it is not possible to firmly establish the relative timing between the two shoreline occupations, although it is certainly reasonable to assume that these events were closely spaced in time.

Published lake-level data with age control from the Chewaucan basin are sparse, and only two other radiocarbon ages help constrain the lake-level history for the past 20 kyr. Allison (1982) obtained a shell age of 17 500 ± 300 ^{14}C yr BP from nearshore sediments at an elevation of 1349 m, at a time when the lake was ca. 75% full. This shell age may somehow be associated with the timing of the maximum late Pleistocene highstand of Lake Chewaucan, but because the stratigraphical context (transgressional or regressional) of the shell is not clear, it is equally possible that the maximum highstand occurred before or after ca. 17.5 ^{14}C ka. A radiocarbon age of 9390 ± 45 ^{14}C yr BP reported on tufa-coated beach gravel found at an elevation of 1318 m in the Lake Abert sub-basin (S. Robinson, personal communication cited in Gehr, 1980) suggests an early Holocene transgression that briefly submerged the 1310 m shoreline, but it is possible that the tufa age is too young (cf. Benson (1978) for discussion of problematic tufa ages). Difficulties in reconciling the tufa age with conflicting data for low lake-levels in the Fort Rock basin at this time further suggest that the age may be in error.

The synthesised latest Pleistocene lake-level curve for the basin (Fig. 3; cf. Negrini, in press) is a composite that relies on correlation of Chewaucan data to lake-level data from the nearby Fort Rock basin (summarized in Freidel, 1993, 1994). The validity of the composite approach draws from the assumption that these two basins were hydrologically connected through subterranean conduits (Allison, 1982), or that they have experienced synchronous lake-level fluctuations under the same regional climate forcing, or a combination of both. Given this argument for correlating between the Fort Rock and Chewaucan basins, evidence for low lake-levels in the Fort Rock basin prior to 12 ^{14}C ka indicates that the lake-level high in the Chewaucan basin represents an abrupt rise and fall of lake-level at ca. 12 ^{14}C ka (Fig. 3). Inherent uncertainties in interbasin correlation and age control prevent conclusive documentation of this proposed lake-level oscillation, and allow for the possibility that the ca. 12 ^{14}C ka Chewaucan shorelines represent temporary stillstands in the overall late Pleistocene recession of Lake Chewaucan. The combined direct and indirect evidence from the radiocarbon ages, stratigraphical relationships, geomorphological observations, and interbasin correlation, however, provides strong support for the probable occurrence of a lake-level oscillation of Lake Chewaucan at ca. 12 ^{14}C ka.

Comparisons with other Great Basin palaeolake records

The lake-level high in the Lake Chewaucan basin at ca. 12 ^{14}C ka occurred during a period of low lake-levels in the Bonneville and Lahontan basins, and post-dates the highstands of these lakes by at least 1000 yr (Fig. 3). Recent mapping and radiocarbon dating of shorelines in the Jessup embayment of Lake Lahontan places the last major (Sehoo) highstand age at 13.1 ^{14}C ka, and suggests that the highest level was maintained for a brief period of years to decades (Adams and Wesnousky, 1998). The period between ca. 12.4 and 11 ^{14}C ka is interpreted as a time of relatively low lake-levels in the Lahontan basin, as indicated by dolomite precipitation at low elevations and a gap in tufa deposition (Benson *et al.*, 1995), and consistent with relatively heavy δ^{18}O and δ^{13}C stable isotope values at this time (Benson *et al.*, 1996b). Lake Bonneville rose to its Bonneville shoreline highstand sometime after ca. 15.2 ^{14}C ka, and remained at this shoreline until the lake-level catastrophically dropped 108 m during the Bonneville Flood at about 14.5 ^{14}C ka (Oviatt *et al.*, 1992; Oviatt, 1997). It is not possible to estimate how long the lake-level would have remained at the highstand had the Bonneville Flood not occurred, but age data clearly show that the lake had entered a major recessional phase after 14 ^{14}C ka that culminated in a low stage at about 12 ^{14}C ka, during which lake-levels may have been lower than average modern levels of Great Salt Lake (Oviatt *et al.*, 1992).

In contrast, the lake-level high in the Lake Chewaucan basin at ca. 12 ^{14}C ka coincides with a period of relatively wet conditions in the Owens Lake basin. A prominent hiatus between ca. 15 and 13.5 ^{14}C ka in several sediment cores from the Owens Lake basin indicates a period of complete lake desiccation during this time interval (Benson *et al.*, 1996a, 1998b; Benson, 1999). This was followed by a period of generally wetter conditions consisting of four wet–dry oscillations between ca. 13.6 and 9.0 ^{14}C ka, as indicated by δ^{18}O proxy data (Fig. 3) (Benson, 1999). At 12 ^{14}C ka, the palaeolake was midway through the longest of the wet oscillations (‘W_1’ of Benson, 1999), which lasted from about 12.9 to 11.2 ^{14}C ka (Fig. 3).

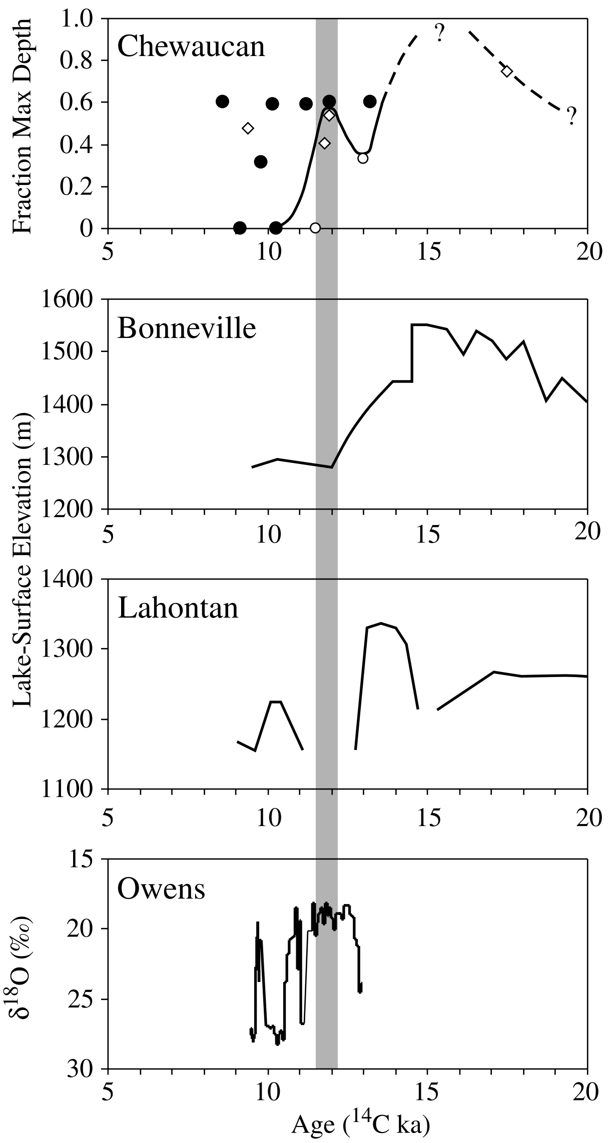

Figure 3 Lake-level curves and lake-size proxies of Great Basin pluvial lakes from 20 to 5 ^{14}C ka, arranged generally north (top panel) to south (bottom panel). The Lake Chewaucan curve is a composite record of lake-level data, normalised to the fraction of maximum lake depth, from the Chewaucan and Fort Rock basins (after Negrini, in press). Open diamonds denote material formed below lake-level in the Chewaucan basin. Open circles denote material formed below lake-level in the Fort Rock basin. Solid circles denote archaeological sites and other material formed above lake-level in the Fort Rock basin. The question mark obscuring the precise timing of the maximum highstand indicates the conjectural nature of its age, and dashed lines indicate poorly constrained portions of the curve. The stippled column identifies the approximate timing of the Chewaucan lake-level oscillation. The curves for lakes Bonneville and Lahontan are generalised from published data of Benson *et al.* (1995, 1997), Oviatt *et al.* (1992), Oviatt (1997), Adams and Wesnousky (1998), and references therein. The Owens Lake δ^{18}O proxy data are from core OL84B using the age model of Benson (1999); δ^{18}O minima (plotted as peaks) are proxies of wet oscillations in the basin

Implications for mechanisms of climate change

Shifts in the position of the polar jet stream, which serves as a boundary between warm tropical air masses and cold polar air masses, have long been implicated as the dominant control on fluctuations in the size of Great Basin palaeolakes and glaciers (Antevs, 1948; Hostetler and Benson, 1990; Benson *et al.*, 1995). The conceptual model of a migrating jet stream predicts a northward progression in the timing of lake-level highs in response to the northerly retreat of the polar jet stream owing to the gradual collapse of the Laurentide Ice Sheet during the deglaciation. Simulations with atmospheric general circulation models (AGCMs) support the jet stream hypothesis (Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; Broccoli and Manabe, 1987). Regional climate models (RegCMs), on the other hand, simulate a complex and spatially heterogeneous pattern of responses across the western USA to prescribed climate forcing (Hostetler and Bartlein, 1999), suggesting that the overall influence of a migrating jet stream may be obscured by regional-scale climatic effects in individual basins.

Although migration of the polar jet stream may have been an important control on the size of Great Basin palaeolakes at other times during the late Pleistocene, the lake-level high in the Lake Chewaucan basin at ca. 12 ^{14}C ka, which constrains a pattern of wet–dry–wet along a north–south transect (Chewaucan–Lahontan/Bonneville–Owens) in the Great Basin at this time (Fig. 3), is not explained by this hypothesis. Relative to the Last Glacial Maximum, the influence of the polar jet stream in western North America was much diminished by 12 ^{14}C ka (Thompson *et al.*, 1993; Kutzbach *et al.*, 1998).

The ca. 12 ^{14}C ka lake levels indicate a spatially variable response to some climate forcing, consistent with RegCM simulations of a complex spatial pattern of climate change across the western USA (Hostetler and Bartlein, 1999). Mechanisms controlling modern climate variability in the western USA may be viable analogues for similar mechanisms that operated during the late Pleistocene. The modern climate of western North America is influenced predominantly by the juxtaposition of the eastern Pacific subtropical high and the southwestern monsoonal circulation from the Gulf of California, each with spatially and seasonally varying impacts on the region (Bryson and Hare, 1974; Mitchell, 1976). The interplay between these air masses and their imprint on regional climate is modulated by El Niño–Southern Oscillation (ENSO), and by changes in the intensity of monsoonal circulation. The ENSO-type variability observed in modern climate may have operated during the late Pleistocene (Heusser and Sirocko, 1997; Cane and Clement, 1999). The ENSO phenomenon commonly causes spatially variable responses in the modern climate, hence by analogy it may have had the potential to cause similar effects in the glacial-age climate across western North America (Cane and Clement, 1999). With regard to the palaeolake record, this would require sustained increases or decreases in the frequency of ENSO of sufficient duration to cause prolonged lake-level fluctuations.

Variations in the intensity of monsoonal precipitation patterns appear to be driven by orbitally-induced changes in insolation (e.g. Thompson *et al.*, 1993). Net summer insolation in the Northern Hemisphere at 12 ^{14}C ka was higher than at the Last Glacial Maximum (ca. 18 ^{14}C ka) or at present (Berger, 1978), which would have amplified the seasonal cycle of solar radiation and driven an intensification of monsoonal circulation at this time. The greater summer radiation would have simultaneously strengthened the eastern Pacific subtropical high, thereby increasing drought conditions in regions dominated by this air mass (Whitlock and Bartlein, 1993; Thompson *et al.*, 1993). The deglaciation of the Laurentide Ice Sheet may have led to a shift in the climatological regime of the western USA, from one that was dominated by the influence of the ice sheet through its effect on the polar jet stream and depression of temperatures in the Northern Hemisphere, to a new regime influenced primarily by the combination of an intensified southwestern monsoon and a strengthened subtropical high, with more sharply delimited and more spatially variable responses to climate forcings than previously experienced during the Last Glacial Maximum (Thompson *et al.*, 1993). The lake-levels observed at 12 ^{14}C ka may be a reflection of this change to a new climate regime.

Variable response times, correlated to lake size, could be another important cause of asynchronous behaviour among Great Basin palaeolakes over short time-scales. Lake Chewaucan and Owens Lake, for example, were likely to respond more quickly than the much larger lakes Bonneville and Lahontan to the same climate forcing. Topographic controls on the amount and type (rain or snow) of precipitation represent another important source of spatial variability of lake-level fluctuations on more local scales across the geographically diverse western USA (cf. Whitlock and Bartlein, 1993; McCabe, 1994).

In summary, the viability of the jet stream hypothesis cannot be rejected based only on a snapshot of pluvial lake-levels at 12 ^{14}C ka. Because these lake-levels reflect spatial variability of climate at this time, however, they emphasize the importance of searching for other mechanisms of climate change in western North America that may be implicated for synoptic climate responses. The importance of these various mechanisms relative to the influence of a migrating jet stream on glacial-age climate, and their effect on the jet stream position itself, remains uncertain. A more rigorous evaluation of mechanisms of climate change in the western USA will require a substantially more complete knowledge of the spatial and temporal pattern of palaeolake-level oscillations across the Great Basin.

Conclusions

1 Direct evidence from radiocarbon ages and stratigraphical relationships, combined with indirect evidence from geomorphological observations and correlation to lake-level data in the Fort Rock basin, indicates that a fluctuation involving a rise and fall of lake-level occurred in the eastern sub-basins of pluvial Lake Chewaucan at ca. 12 ^{14}C ka.

2 Geomorphological observations suggest that the ca. 12 ^{14}C ka Chewaucan lake-level oscillation is associated with an overflow of the coalesced lake in the three eastern sub-basins into the Summer Lake sub-basin.

3 The influence of hypsometry and the consequences of river diversion on lake-level and shoreline development in the Chewaucan basin illustrate important non-climatic mechanisms for modulation of lake-levels in Great Basin palaeolakes.

4 The ca. 12 ^{14}C ka Chewaucan lake-level high is coeval with lowstands in the Bonneville and Lahontan basins and relatively wet conditions in the more southerly Owens Lake basin. This observed spatial pattern of pluvial lake-levels is not explained by the conceptual model of a migrating polar jet stream as a dominant control on palaeolake levels in the Great Basin, and may represent an important example of synoptic responses to some climate forcing.

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Appendix: additional sample details

Sample 4F: *Valvata humeralis* (Burch, 1989; all identifications performed by the author), section 4 on Willow Creek, Lower Chewaucan Marsh sub-basin, Coglan Buttes SE quadrangle, NW1/4 section 26 T35S R20E, latitude 42.5129°N, longitude 120.3296°W.

Sample 6AV: *Valvata humeralis*, section 6 on east shore of Lake Abert, Lake Abert sub-basin, Lake Abert South quadrangle, NE1/4 section 25 T34S R21E, latitude 42.5990°N, longitude 120.1845°W.

Sample 7EG: *Vorticifex (Parapholyx) effusa*, section 7 on east shore of Lake Abert, Lake Abert sub-basin, Lake Abert South quadrangle, NW1/4 section 15 T35S R21E, latitude 42.5351°N, longitude 120.2280°W.

Sample 7FG: *Vorticifex (Parapholyx) effusa*, section 7 on east shore of Lake Abert, Lake Abert sub-basin, Lake Abert South quadrangle, NW1/4 section 15 T35S R21E, latitude 42.5351°N, longitude 120.2280°W.

References

Adams KD, Wesnousky SG. 1998. Shoreline processes and the age of the Lake Lahontan highstand in the Jessup embayment, Nevada. *Geological Society of America Bulletin* **110**: 1318–1332.

Allen BD, Anderson RY. 1993. Evidence from western North America for rapid shifts in climate during the last glacial maximum. *Science* **260**: 1920–1923.

Allison IS. 1982. *Geology of Pluvial Lake Chewaucan, Lake County, Oregon*. Oregon State Monographs, Studies in Geology Number Eleven. Oregon State University Press: Corvallis; 79 pp.

Antevs E. 1948. The Great Basin with emphasis of glacial and postglacial times: climatic changes and pre-white man. *University of Utah Bulletin* **38**: 167–191.

Benson LV. 1978. Fluctuations in the level of pluvial Lake Lahontan during the last 40,000 years. *Quaternary Research* **9**: 300–318.

Benson LV. 1993. Factors affecting ^{14}C ages of lacustrine carbonates: timing and duration of the last highstand lake in the Lahontan basin. *Quaternary Research* **39**: 163–174.

Benson LV. 1999. Records of millennial-scale climate change from the Great Basin of the western United States. In *Mechanisms of Global Climate Change at Millennial Time Scales*, Clark PU, Webb RS, Keigwin LD (eds). Geophysical Monograph 112, American Geophysical Union; Washington, DC; 203–225.

Benson LV, Thompson RS. 1987. The physical record of lakes in the Great Basin. In *North America and Adjacent Oceans During the Last Deglaciation*, Ruddiman WF, Wright HE Jr (eds). Geology of North America, Geological Society of America: Boulder; 241–260.

Benson LV, Currey DR, Dorn RI, Lajoie KR, Oviatt CG, Robinson SW, Smith GI, Stine S. 1990. Chronology of expansion and contraction of four Great Basin lake systems during the past 35,000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* **78**: 241–286.

Benson LV, Kashgarian M, Rubin M. 1995. Carbonate deposition, Pyramid Lake sub-basin, Nevada: 2. lake levels and polar jet stream positions reconstructed from radiocarbon ages and elevations of carbonates (tufas) deposited in the Lahontan basin *Palaeogeography, Palaeoclimatology, Palaeoecology* **117**: 1–30.

Benson LV, Burdett JW, Kashgarian M, Lund SP, Phillips FM, Rye RO. 1996a. Climatic and hydrologic oscillations in the Owens Lake basin and adjacent Sierra Nevada, California. *Science* **274**: 746–749.

Benson LV, White LD, Rye RO. 1996b. Carbonate deposition, Pyramid Lake subbasin, Nevada: 4. Comparison of the stable isotope values of carbonate deposits (tufas) and the Lahontan lake-level record. *Palaeogeography, Palaeoclimatology, Palaeoecology* **122**: 45–76.

Benson LV, Smoot JP, Kashgarian M, Sarna-Wojcicki A, Burdett JW. 1997. Radiocarbon ages and environments of deposition of the Wono and Trego Hot Springs tephra layers in the Pyramid Lake subbasin, Nevada. *Quaternary Research* **47**: 251–260.

Benson LV, Lund SP, Burdett JW, Kashgarian M, Rose TP, Smoot JP, Schwartz M. 1998a. Correlation of late-Pleistocene lake-level oscillations in Mono Lake, California, with North Atlantic climate events. *Quaternary Research* **49**: 1–10.

Benson LV, May HM, Antweiler RC, Brinton TI, Kashgarian M, Smoot JP, Lund SP. 1998b. Continuous lake-sediment records of glaciation in the Sierra Nevada between 52,600 and 12,500 ^{14}C yr B.P. *Quaternary Research* **50**: 113–127.

Berger AL. 1978. Long-term variations of caloric insolation resulting from Earth's orbital elements. *Quaternary Research* **9**: 139–167.

Broccoli AJ, Manabe S. 1987. The influence of continental ice, atmospheric CO_2, and land albedo on the climate of the last glacial maximum. *Climate Dynamics* **1**: 87–99.

Bryson RA, Hare FK. 1974. The climate of North America. In *Climates of North America*, Bryson RA, Hare FK (eds). Elsevier: Amsterdam; 1–47.

Burch JB. 1989. *North American Freshwater Snails*. Malacological Publications: Hamburg, Michigan; 365 pp.

Cane M, Clement AC. 1999. A role for the tropical Pacific coupled ocean–atmosphere system on Milankovitch and millennial timescales. Part II: Global impacts. In *Mechanisms of Global Climate Change at Millennial Time Scales*, Clark PU, Webb RS, Keigwin LD (eds). Geophysical Monograph 112, American Geophysical Union: Washington, DC; 373–383.

Clark PU, Bartlein PJ. 1995. Correlation of late-Pleistocene glaciation in the western United States with North Atlantic Heinrich events. *Geology* **23**: 483–486.

Cohen AS, Palacios-Fest MR, Negrini RM, Wigand PE, Erbes DB. 2000. A paleoclimate record for the last 250,000 years from Summer Lake, Oregon, U.S.A.: II. sedimentology, paleontology, and geochemistry. *Journal of Paleolimnology* **24**: 151–181.

Davis JO. 1985. Correlation of late Quaternary tephra layers in a long pluvial sequence near Summer Lake, Oregon. *Quaternary Research* **23**: 38–53.

Freidel DE. 1993. *Chronology and climatic controls of late Quaternary lake-level fluctuations in Chewaucan, Fort Rock, and Alkali basins, south-central Oregon*. Unpublished PhD dissertation, University of Oregon; 244 pp.

Freidel DE. 1994. Palaeolake shorelines and lake-level chronology of the Fort Rock basin, Oregon. In *Archaeological Researches in the Northern Great Basin: Fort Rock Archaeology since Cressman*, Aikens CM, Jenkins DL (eds). Anthropological Papers 50, University of Oregon Department of Anthropology and State Museum of Anthropology: Eugene; 21–40.

Gehr KD. 1980. Late Pleistocene and Recent archaeology and geomorphology of the south shore of Harney Lake, Oregon. Unpublished MA thesis, Portland State University; 170 pp.

Gilbert GK. 1890. *Lake Bonneville*. Monograph 1, US Geological Survey: Washington, DC; 438 pp.

Heusser LE, Sirocko F. 1997. Millennial pulsing of environmental change in southern California from the past 24 k.y.: a record of Indo-Pacific ENSO events? *Geology* **25**: 243–246.

Hostetler SW, Bartlein PJ. 1999. Simulation of the potential responses of regional climate and surface processes in western North America to a canonical Heinrich event. In *Mechanisms of Global Climate Change at Millennial Time Scales*, Clark PU, Webb RS, Keigwin LD (eds). Geophysical Monograph 112, American Geophysical Union: Washington, DC; 313–327.

Hostetler SW, Benson LV. 1990. Paleoclimatic implications of the high stand of Lake Lahontan derived from models of evaporation and lake level. *Climate Dynamics* **4**: 207–217.

Jellinek AM, Madin IP, Langridge RM. 1996. Field and stable isotope indicators of geothermal resource potential, central Lake County, Oregon. *Oregon Geology* **58**: 3–9.

Kutzbach JE, Guetter PJ. 1986. The influence of changing orbital para meters and surface boundary conditions on climate simulations for the past 18,000 years. *Journal of Atmospheric Science* **43**: 1726–1759.

Kutzbach JE, Gallimore R, Harrison S, Behling P, Selin R, Laarif F. 1998. Climate and biome simulations for the past 21,000 years. *Quaternary Science Reviews* **17**: 473–506.

Langridge RM, Madin IP, Jellinek AM. 1995. Late Pleistocene 'hydrothermal' tufa mounds of Lake Chewaucan, central Oregon: interaction of structure, hydrology, and lake-level in tufa-genesis. *Geological Society of America Abstracts with Programs* **27**: 59.

Langridge RM, Weldon RJ, Pezzopane SK, Jellinek AM. 1996. Active faulting and tufa formation at Sawed Horn, central Oregon: a possible kinematic link between the Abert Rim and Viewpoint faults. *Geological Society of America Abstracts with Programs* **28**: 84.

Lin JC, Broecker WS, Hemming SR, Hajdas I, Anderson RF, Smith GI, Kelley M, Bonani G. 1998. A reassessment of U–Th and ^{14}C ages for late-glacial high-frequency hydrological events at Searles Lake, California. *Quaternary Research* **49**: 11–23.

Manabe S, Broccoli AJ. 1985. The influence of continental ice sheets on the climate of an ice age. *Journal of Geophysical Research* **90**: 2167–2190.

McCabe GJ Jr. 1994. Relationships between atmospheric circulation and snowpack in the Gunnison River basin, Colorado. *Journal of Hydrology* **157**: 157–175.

Mifflin MD, Wheat MM. 1979. Pluvial lakes and estimated pluvial climates of Nevada. *Nevada Bureau of Mines and Geology Bulletin* **94**: 57 pp; University of Nevada, Leno.

Mitchell VL. 1976. The regionalization of climate in the western United States. *Journal of Applied Meteorology* **15**: 920–927.

Negrini RM. In press. Pluvial lake sizes in the northwestern Great Basin throughout the Quaternary period. In *Great Basin Aquatic Systems History*, Hershler R, Currey DR, Madsen DB (eds). Smithsonian Press; Washington, DC.

Negrini RM, Erbes DB, Faber K, Herrera AM, Roberts AP, Cohen AS, Wigand PE, Foit FF Jr. 2000. A paleoclimate record for the last 250,000 years from Summer Lake, Oregon, U.S.A.: I. Age control and magnetic proxies for lake level. *Journal of Paleolimnology* **24**: 125–149.

Oetting AC. 1994. Chronology and time markers in the northwestern Great Basin: the Chewaucan basin cultural chronology. In *Archaeological Researches in the Northern Great Basin: Fort Rock Archaeology since Cressman*, Aikens CM, Jenkins DL (eds). Anthropological Papers 50, University of Oregon Department of Anthropology and State Museum of Anthropology: Eugene; 41–62.

Oviatt CG. 1997. Lake Bonneville fluctuations and global climate change. *Geology* **25**: 155–158.

Oviatt CG, Currey DR, Sack D. 1992. Radiocarbon chronology of Lake Bonneville, eastern Great Basin, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* **99**: 225–241.

Palacios-Fest MR, Cohen AS, Ruiz J, Blank B. 1993. Comparative paleoclimatic interpretations from nonmarine ostracodes using faunal assemblages, trace elements, shell chemistry, and stable isotope data. In *Climate Change in Continental Isotopic Records*, Swart PK, Lohmann KC, McKenzie J, Savin S (eds). Geophysical Monograph 78, American Geophysical Union: Washington, DC; 179–190.

Pettigrew RM. 1985. *Archaeological Investigations on the East Shore of Lake Abert, Lake County, Oregon*, Vol. 1. Anthropological Papers 32, University of Oregon Department of Anthropology and State Museum of Anthropology: Eugene; 229 pp.

Pezzopane SK. 1993. *Active faults and earthquake ground motions in Oregon*. Unpublished PhD dissertation, University of Oregon; 208 pp.

Pezzopane SK, Weldon RJ. 1993. Tectonic role of active faulting in central Oregon. *Tectonics 12*: 1140–1169.

Phillips FM, Campbell AR, Smith GI, Bischoff JL. 1994. Interstadial climatic cycles: a link between western North America and Greenland? *Geology* **22**: 1115–1118.

Sack D. 1999. The composite nature of the Provo level of Lake Bonneville, Great Basin, western North America. *Quaternary Research* **52**: 316–327.

Simpson GD. 1990. *Late Quaternary tectonic development of the northwestern part of Summer Lake basin, south-central Oregon*. Unpublished MS thesis, Humboldt State University; 121 p.

Smith GI, Street-Perrott FA. 1983. Pluvial lakes of the western United States. In *Late Quaternary Environments of the United States*, Porter SC (ed.). University of Minnesota Press: Minneapolis; 190–211.

Thompson RS, Whitlock C, Bartlein PJ, Harrison SP, Spaulding WG. 1993. Climatic changes in the western United States since 18,000 yr B.P. In *Global Climates Since the Last Glacial Maximum*, Wright HE Jr, Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, Bartlein PJ (eds). University of Minnesota Press: Minneapolis; 468–513.

Whitlock C, Bartlein PJ. 1993. Spatial variations of Holocene climatic change in the Yellowstone region. *Quaternary Research* **39**: 231–238.

