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Lake-size variations in the Lahontan and Bonneville basins between 13,000 and 9000 ^{14}C yr B.P.

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ABSTRACT

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Recessions of Lakes Lahontan and Bonneville that commenced ~13,500 ^{14}C yr B.P. were interrupted at ≥11,500 ^{14}C yr B.P. in the Lahontan basin and ~12,200 ^{14}C yr B.P. in the Bonneville basin by relatively large perturbations in lake level that persisted for ~2000 years. Minor glacial readvances in the Sierra Nevada and White Mountains of California–Nevada occurred during the latter half of this interval (between 11,000 and 9700 ^{14}C yr B.P.). The hydrologic response of Lakes Lahontan and Bonneville and the mountain glacial advances were concurrent with the Allerød/Younger Dryas climatic intervals recorded in vegetational and glacial records of western and central Europe.

Introduction

Between 13,000 and 10,000 ^{14}C yr B.P., Europe south of the Weichselian ice sheet experienced major changes in climate. Based primarily on vegetational change, this period has been subdivided into three intervals: Bølling (13,000–12,000 ^{14}C yr B.P.), Allerød (12,000–11,000 ^{14}C yr B.P.), and Younger Dryas (11,000–10,000 ^{14}C yr B.P.) (Mangerud et al., 1974). During the Allerød interval, Europe was the site of a transition from tundra to tree birch; but at the end of the Allerød interval, the inferred trend in climatic warming was reversed by a return to tundra conditions—the Younger Dryas interval—which lasted about 1000 years (Wright, 1989).

During the last few years, a debate has developed over whether the Younger Dryas event was confined to the periphery of the North Atlantic and the European continent or whether the event was global in scope. In North America, only sites adjacent to the North Atlantic indicate unequivocal evidence of a climatic perturbation between 11,000 and 10,000 ^{14}C yr B.P. (Livingstone and Livingstone, 1958; Anderson, 1983; Mott et al., 1984; Mott, 1985). Shane (1987, 1990) has documented the recurrence of spruce (*Picea*) at 14 sites in the Till Plains of Indiana, Michigan, and Ohio which were interpreted as indicating a climatic cooling between 11,000 and 10,000 ^{14}C yr B.P.; however, whether this cooling was due to a global Younger Dryas event or to the discharge of cold glacial meltwater to lakes north of the Till Plains remains unresolved. Hence, the effect of the Younger Dryas event on the climate of the North American continent west of the Atlantic border remains uncertain.

The overall objective of this paper is to present the record of lake-level change in the northern Great Basin for the 4500 yr that followed the highstands of Lakes Lahontan and Bonneville. In focusing on this period, we also present some

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evidence on the nature of climatic change that occurred in the western United States during the Bølling/Allerød/Younger Dryas intervals.

Lake Lahontan and Lake Bonneville (Fig. 1) may have had nearly synchronous recessions commencing ~13,500 ^{14}C yr B.P. (Benson et al., 1990). Until recently, however, not much was known about lake-size variation in the Lahontan and Bonneville basins subsequent to 13,000 ^{14}C yr B.P. In this paper, model chronologies of lake-level variation are presented for 13,000–9000 ^{14}C yr B.P. The chronologies are based on radiocarbon-

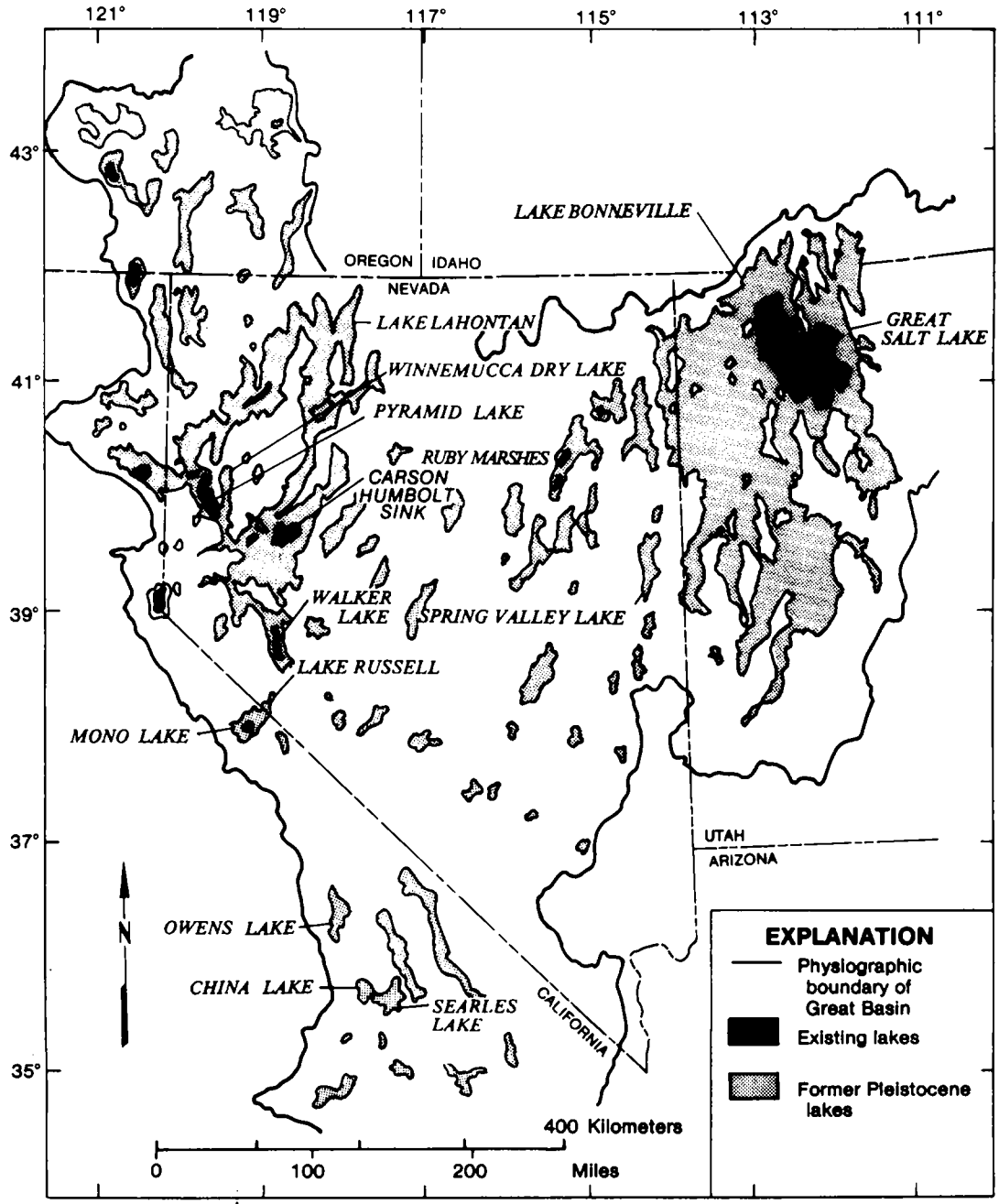

Fig. 1. Principal Great Basin paleolake systems of the late Quaternary.

age estimations of tufas, shells, plants, and organic material associated with rock varnish. Most of the data on which the Lahontan chronology is based has not been published previously; and the data on which the Bonneville chronology is based has been previously published in journals and reports that may not be accessible to the interested reader.

Lake-level variation in the Lahontan basin between 13,000 and 9000 ^{14}C yr B.P.

The Lahontan basin contains seven subbasins separated by sills of varying elevation (Fig. 2). Radiocarbon ages[1] of carbonate materials from the Carson Desert subbasin (Table 1, Fig. 2) indicate the existence of a shallow lake between ~11,500 and 10,500 ^{14}C yr B.P. Currey (1988, 1989) correlated these carbonate materials with shoreline features that occur at an elevation of 1203 m; therefore, the surface area of the lake in the Carson Desert subbasin ~11,000 ^{14}C yr B.P. was ~2.3 times the present-day calculated surface areas (PDCSA) of lakes in the Carson Desert[2]. Given a corresponding increase in the hydrologic balance of the Truckee River–Pyramid Lake, surface-water system, the calculated combined surface areas of Pyramid and Winnemucca Lakes would have exceeded 1140 km^2, and the elevation of this lake would have been stabilized at 1207 m by spill across the Emerson Pass sill (Fig. 2).

A search for datable materials deposited at an elevation of ~1207 m in the Winnemucca Lake subbasin (Fig. 2) led to the discovery of two tufas located at elevations (1205 and 1206 m, Table 1) only slightly below the elevation of the Emerson Pass Spill. Thin-section examination of samples of these tufas indicated that one sample (WDL89-5) contained small pockets of brown secondary material along its base; therefore, the surface 0.5 cm of both samples was removed prior to radiocarbon-age dating. We believe these tufas were formed in shallow water when the level of the lake was maintained at ~1207 m by spill across the Emerson Pass sill.

Carbonate materials deposited at an elevation of 1207 m in the Pyramid Lake subbasin were not found; instead, the distal margin of a wave-cut platform was discovered at approximately this elevation (1208 m) on Anaho Island, Marble Bluff (Fig. 2) and elsewhere[3]. On rocky areas such as Marble Bluff and Anaho Island, the surface of the platform steepens with increasing elevation. The lower (1208 m) and upper limits (1225 1230 m) of the platform are defined by thick ($\leqslant$2.1 m) accumulations of carbonate capped by dendritic tufa. Carbonates from the inner surface of the 2.1-m thick tufa found at the upper boundary of the platform have a ^{14}C age of 18,880 $\pm$240 yr B.P. and the dendritic carbonate that forms the outer surface of the tufa has a ^{14}C age of 13,240$\pm$80 yr B.P. (L.V. Benson and Y. Lao, unpublished data). However, no evidence of the 13,240 $\pm$80 ^{14}C yr B.P. dendritic tufa can be found on the wave-cut platform, although this type of tufa is found above and below the wave-cut platform throughout the Pyramid Lake subbasin. Instead, the platform is composed of erosionally-exposed dense carbonate that cements dolomite talus weathered from the face of Marble Bluff. Carbonate from near the surface of the platform has a detrital ^{230}Th-corrected uranium-series age of 61,800 $\pm$2300 yr B.P. (L.V. Benson and Y. Lao, unpublished data).

We believe that the formation of the 1208–1225-m wave-cut platform can best be attributed to a rise of lake level that removed the 13,240 $\pm$80 ^{14}C yr B.P. dendritic tufa material from the area of

[1] Radiocarbon-age dates used in this paper are not absolute with respect to time but are based on a Libby half-life of 5568 yr and corrected for isotopic fractionation ($\delta^{13}C$ values when not available were set to a value typical of Lahontan tufas; i.e. 3.2‰ relative to PDB).

[2] In this paper, we refer to the mean historical, reconstructed surface area as present-day calculated surface area (PDCSA). In the Lahontan basin, PDCSA was derived using reconstructed mean annual streamflow-discharge data, precipitation data (Benson and Thompson, 1987), an evaporation rate of 1.25 m/yr and the water-balance equation for a closed-basin lake. Discharge data for the four major rivers entering the Lahontan basin is available for $\geq$31 yr. Discharge data were reconstructed to pristine conditions by multiplying irrigated areas by 1.00 m (the estimated evaporation rate on irrigated lands) and adding that amount of water to the measured discharge.

[3] The terrace was cut into sediments around the margins of present-day Pyramid Lake and Winnemucca Dry Lake playa. Because of the unconsolidated nature of the sedimentary material, the width of the terrace often exceeds several tens of meters.

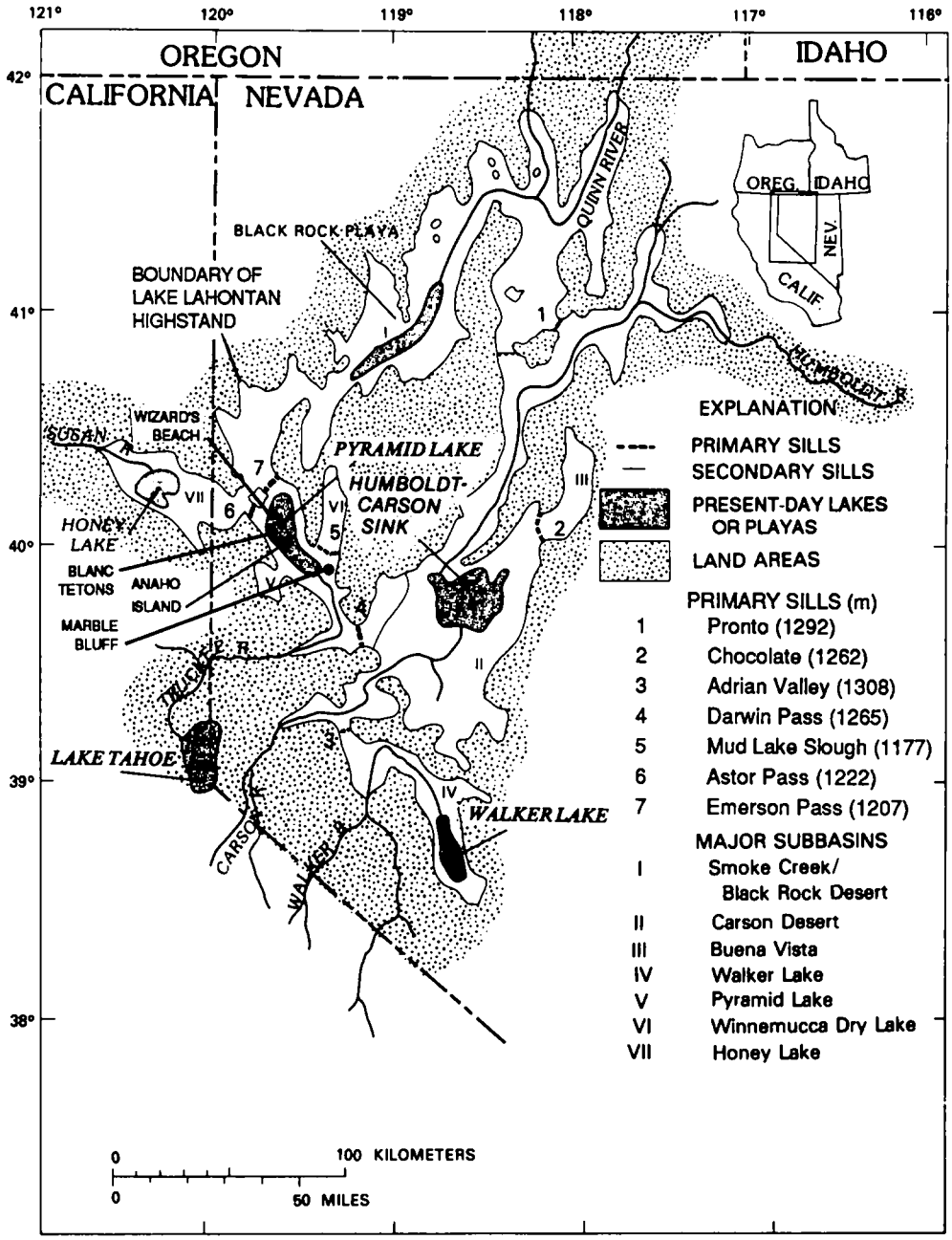

Fig. 2. Rivers, subbasins, sills, and sample localities in the Lahontan basin.

the platform. A lake that is receding or held stable against a steeply inclined shoreland is capable of eroding only a narrow wave-cut platform (Stine, 1987). Stabilization in lake level might account for partial removal of carbonate that coated the distal margin of the platform when the level of the lake was constrained by spill across Emerson Pass sill, but stabilization cannot account for removal of the carbonate from the wave-cut platform from above 1207 m. In addition, the outer surfaces of dendritic tufa from between 1330 and 1230 m are not eroded indicating that removal of tufa from this interval did not occur during recession of the lake from its highstand.

TABLE 1

Number, type, locality, elevation, and radiocarbon ages of samples from the Lahontan basin

Sample number	Sample type	Locality information	Elevation (m)	^{14}C age (yr B.P.)
Pyramid Lake subbasin				
PL90-1a	top half laminated tufa	Blanc Tetons	1163	9020 ± 70
Beta-31065	organic matter in rock varnish	Anaho Island	1220	9620 ± 90
PL90-2a	top half laminated tufa	Blanc Tetons	1161	9660 ± 80
GX13744[a]	bark fishing line	Wizard's Beach	1154	9660 ± 170
WIS-377[b]	wood	Truckee River Delta	1169	9720 ± 100
I-8194[c]	wood	Truckee River Delta	1177	9780 ± 130
PL90-1b	bottom half laminated tufa	Blanc Tetons	1163	9930 ± 80
I-8193[c]	wood	Truckee River Delta	1177	9970 ± 140
PL90-2b	bottom half laminated tufa	Blanc Tetons	1161	10,180 ± 80
TO-167	organic matter below rock varnish	Anaho Island	1220	10,460 ± 180
Winnemucca Dry Lake subbasin				
LDGO-1743L (WDL89-5)	tufa (algal nodules)	3 km NW Coyote Canyon	1205	10,080 ± 60
LDGO-1743K (WDL89-4)	tufa (algal coating)	3 km NW Coyote Canyon	1206	11,380 ± 120
Carson Desert subbasin				
Beta-29024[d]	*Anodonta* shells	N side of Humboldt Bar, 0.5-m-thick shore-zone sand	1198	10,380 ± 80
Beta-24290[e]	tufa (dense nodules)	E side of Upsal Hogback, top of Carson River paleodelta	1203	11,100 ± 110
Beta-24291[e]	marl	Stillwater Slough, elevation of lake surface not known	1198	11,300 ± 130

[a]previously reported in Tuohy (1988).
[b]previously reported in Born (1972).
[c]previously reported in Prokopovich (1983).
[d]previously reported in Currey (1990).
[e]previously reported in Currey (1988).

As the outer surfaces of dendritic tufa from elevations below 1208 m (the distal margin of the terrace) also indicate evidence of erosion, it is possible that the lake receded below this altitude prior to its subsequent rise. The fact that only the outer tufa surfaces were removed can be interpreted to indicate that the rise in lake level to 1208 m was relatively rapid. With respect to the maximum extent of the transgression, radiocarbon ages of organic material from water-soluble packrat middens found in the Winnemucca Dry Lake subbasin (Thompson et al., 1986) indicate that lakes in the western Lahontan subbasins could not have not risen above an altitude of 1230 m since 12,070 ± 210 ^{14}C yr B.P.

Rock-varnish samples were collected from 17 boulders located at an elevation of 1220 m on the upper part of the erosional terrace on the west side of Anaho Island. At least five cross-sections of varnish from each boulder were analyzed by light and electron microscopy (Krinsley et al., 1990) to determine which samples were suitable for age dating. Based on these analyses, varnishes from two boulders penetrated by fungi were rejected for AMS radiocarbon-age dating. Varnishes were scraped and composited from two sets of the remaining boulders and the composites processed for AMS radiocarbon dating as described in Dorn et al. (1989). The Beta-31065 (ETH-5265) composite yielded a radiocarbon age of 9620 ± 95 ^{14}C yr B.P. and the TO-167 composite yielded a radiocarbon age of 10,460 ± 180 ^{14}C yr

B.P. (Table 1). Examination of varnish cross-sections from one TO-167 boulder indicated the presence of abundant pockets of organic debris trapped in subvarnish cavities. Some of this organic material consists of pollen grains but most of the material is unidentifiable. Most of the organic matter extracted from TO-167 boulders may have come from subvarnish pockets and the 9% difference in the ages of the two varnish composites may, therefore, be due to differences in the locations of organic matter; i.e, the TO-167 sample contained organic matter that was deposited within the upper surface of the terrace $\geqslant$10,500 ^{14}C yr B.P. prior to accumulation of an amount of basal varnish organic matter sufficient for AMS ^{14}C age dating. Radiocarbon-age dates of organic material from basal varnish layers provide a *minimum* surface exposure age for the underlying landform; i.e., the erosional terrace formerly occupied by the lake. Dorn et al. (1989) have demonstrated, that in desertic areas, basal varnish ^{14}C-age dates lag the true ages of landform exposure by $\leqslant$10%[1]. In a more precisely calibrated study, Dorn et al. (1989) have shown that radiocarbon age dates of basal varnish layers formed on lava flows on the semi-arid side of Hualalai Volcano, Hawaii, lag radiocarbon age dates on charcoals formed during emplacement of the basalts by ~13%. In this paper, we will apply the 13% lag-time correction to the AMS ^{14}C age date of Beta-31065 to estimate the maximum age of terrace exposure (10,850 ^{14}C yr B.P.). The uncorrected AMS ^{14}C date of Beta-31065 provides an estimate of the minimum age of terrace exposure (9620 ^{14}C yr B.P.). Therefore, abandonment of the 1220-m shoreline is calculated to have occurred sometime between 10,850 and 9,600 ^{14}C yr B.P. If the TO-167 composite accumulated in the upper surface of the erosional terrace soon after recession of the lake, the TO-167 age date (10,460 $\pm$ 180 ^{14}C yr B.P.) may closely approximate the actual time of shoreline abandonment.

On the west side of Pyramid Lake (Blanc Tetons site, Fig. 2), a 2-cm thick laminated tufa coats the outer face of a tufa complex. Samples of the laminated tufa were taken from material exposed on the present-day beach (PL90-2a and b) and the top of a tufa sphere (PL90-1a and b) that stands ~3 m above the beach. The surfaces of the samples were removed and the top and bottom halves of the remaining samples submitted for radiocarbon-age determinations (Table 1). Thin-section examination of the samples indicated that the top half of PL90-1a contains void-filling, clear, carbonate cement. The estimated age of this sample (9000$\pm$80 ^{14}C yr B.P.) is, therefore, a minimum. Significant amounts of secondary infill were not found in the remaining three samples; however, the textural complexity of these samples makes detection of diagenetic features difficult. Although secondary carbonate was not observed in thin sections of sample PL90-2a, the apparent age difference between PL90-2a and PL90-2b (~500 yr) could be due to addition of 1.7% modern carbon to sample PL90-2a.

The above data, together with radiocarbon ages of paleo-Indian fishing line (Tuohy, 1988) and wood from the ancestral Truckee River delta (Prokopovich, 1983)(Table 1) have been used to create a model chronology of lake-level variation (Fig. 3) for the western Lahontan subbasins (Pyramid Lake, Winnemucca Dry Lake, and Smoke Creek/Black Rock Desert) for 13,000–9000 ^{14}C yr B.P. The timing of the initial recession of Lake Lahontan is based on data tabulated in Benson et al. (1990). The line depicting the recession was drawn using a date of 13,500 ^{14}C yr B.P. and an elevation of 1330 m. Between 12,800 and 9700 ^{14}C yr B.P., the following considerations were used in constraining the shape of the lake level curve: (1) Between ~12,500 and 11,000 ^{14}C yr B.P., age dates of twigs from soluble packrat middens provide an upper limit on the placement of lake level (Thompson et al., 1986); (2) During the post-highstand recession, lake level receded to $\leqslant$1208 m; (3) The indicated stabilization of lake level at 1207 m is based on the existence of tufa that is interpreted to have formed in shallow water as the

[1]The true ages of landform exposures used in Dorn et al. (1989) were obtained from ^{14}C-age dates of carbon-bearing materials associated with the landform. In many cases, these age dates are only approximations of the true age of a landform. For example, ^{14}C-age dates of tufas that cement the Lahontan highstand terrace sometimes yield age dates that are minimum estimates of the true age of the landform (Benson et al., 1990). Therefore the $\leq$10% lag time assigned by Dorn et al. (1989) is a minimum value.

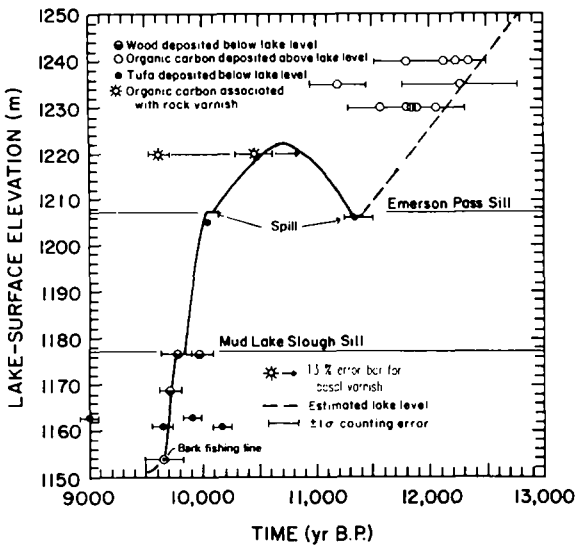

Fig. 3. Model chronology for Lake Lahontan 13,000–9000 ^{14}C yr B.P. Open circles plotted at elevations >1225 m indicate samples of organic carbon extracted from water-soluble packrat middens (Thompson et al., 1986). The upper end of the line depicting the recession was drawn from an elevation of 1330 m and a ^{14}C age of 13,500 yr B.P. Between 12,800 and 9700 ^{14}C yr B.P., the following considerations were used in constraining the shape of the lake level curve: (1) Between ~12,500 and 11,000 ^{14}C yr B.P., age dates of twigs from soluble packrat middens provide an upper limit on the placement of lake level (Thompson et al., 1986); (2) During the post-highstand recession, lake level retreated to at least 1208 m—the base of the platform cut during the following rise in lake level; (3) The indicated stabilization of lake level at 1207 m is based on the existence of tufa that is interpreted to have formed in shallow water as the lake spilled to the Smoke Creek/Black Rock Desert subbasin ~11,300 ^{14}C yr B.P.; (4) The maximum level achieved during the transgression could not have exceeded the lower elevational limit of the soluble packrat middens (1230 m) but must have been higher than the elevation of the rock varnish (1220 m) that formed after the lake receded; (5) Recession from the 1220-m level has been shown as occurring somewhat prior to 10,600 ^{14}C yr B.P—the age of organic matter entrapped in the upper surface of the terrace eroded into Anaho Island; (6) The indicated stabilization of lake level at 1207 m is based on the existence of tufa that is interpreted to have formed in shallow water as the lake spilled to the Smoke Creek/Black Rock Desert subbasin ~10,100 ^{14}C yr B.P.; (7) The recession in lake level between 10,000 and 9700 ^{14}C yr B.P. is tightly constrained by ^{14}C age dates on both tufas and wood. The arrow drawn from the rock-varnish organic carbon sample indicates the amount of time that may have elapsed since the exposure of the terrace to the atmosphere (≤13% of the ^{14}C age of the sample). An similar arrow has not been attached to the other organic carbon sample since it may represent material deposited within the terrace surface soon after exposure of the terrace to the atmosphere.

lake spilled to the Smoke Creek/Black Rock Desert subbasin ~11,400 ^{14}C yr B.P.; (4) The maximum level achieved during the transgression could not have exceeded the lower elevational limit of the soluble packrat middens (1230 m) but must have been higher than the elevation of the rock varnish (1220 m) that formed after the lake receded; (5) Recession from the 1220-m level has been shown as occurring somewhat prior to 10,600 ^{14}C yr B.P—the age of organic matter entrapped in the upper surface of the terrace eroded into Anaho Island; (6) The indicated stabilization of lake level at 1207 m is based on the existence of tufa that is interpreted to have formed in shallow water as the lake spilled to the Smoke Creek/Black Rock Desert subbasin ~10,100 ^{14}C yr B.P.; (7) The recession in lake level between 10,000 and 9700 ^{14}C yr B.P. is tightly constrained by ^{14}C age dates on both tufas and wood.

Lake-level variation in the Bonneville basin between 13,000 and 9000 ^{14}C yr B.P.

The Bonneville basin (Fig. 1) consists of three main subbasins (Great Salt Lake, Great Salt Lake Desert, and Sevier Desert) separated by sills of varying elevation (Fig. 4). Today, the sill between the Great Salt Lake and Great Salt Lake Desert subbasins is only ~2 m higher than the bottom of the Great Salt Lake Desert subbasin (1283 m), and during Pleistocene time, this sill may not have existed. During the late Pleistocene, the sill separating the Great Salt Lake and Sevier Desert subbasins (Old River Bed sill, Fig. 4) was at an elevation of ~1390 m. Because of postpluvial isostatic rebound, the Old River Bed sill is now at an elevation of 1400 m (Fig. 5)(Currey, 1982).

At the west Public Shooting Grounds, located near the northeast shore of Great Salt Lake (Fig. 4), the oldest exposed sediments are red beds unconformably overlain by gastropod-bearing organic muds and gastropod-rich sands (Fig. 15, Currey, 1990) that have an age of ~12,000 ^{14}C yr B.P. (Beta-16912, Table 2). The red sediments probably were derived from iron sulfide bearing, anoxic facies of Lake Bonneville deep-water sediments that were reddened by oxidation as receding brines reworked them basinward across mudflats

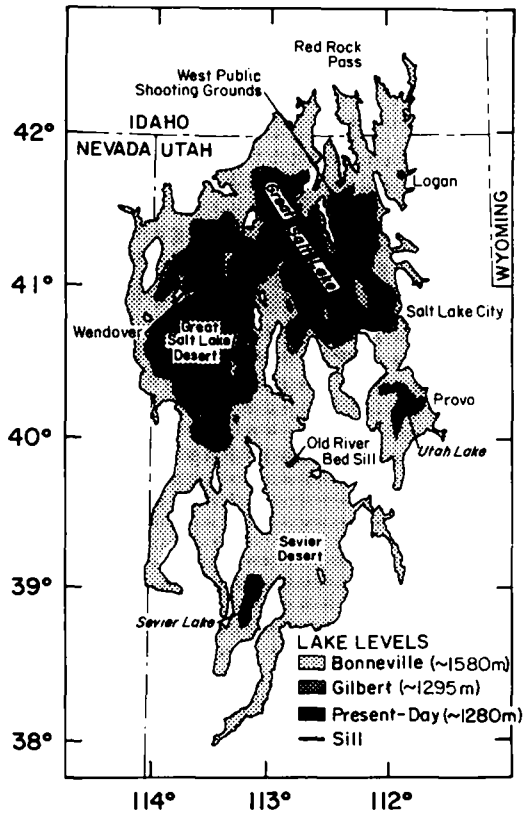

Fig. 4. Map of the Lake Bonneville region depicting the open-basin (Bonneville) stage prior to the Bonneville Flood, the highest (Gilbert) stage of Great Salt Lake subsequent to the final regression of Lake Bonneville, and the areas of present-day Great Salt Lake and Utah Lake (adapted from Currey, 1990).

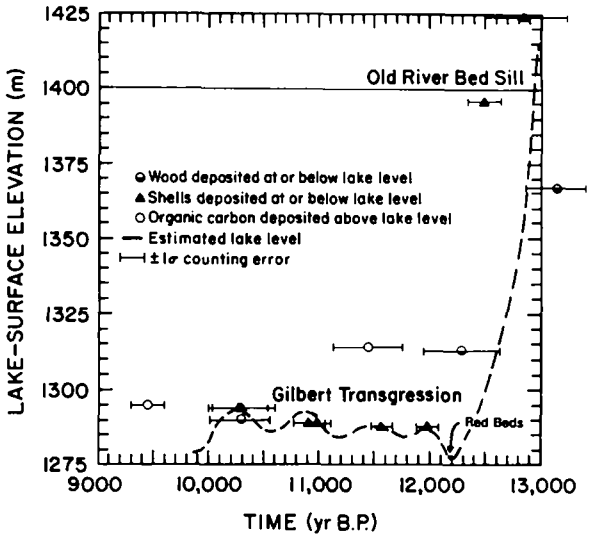

Fig. 5. Model chronology for Lake Bonneville 13,000–9000 ^{14}C yr B.P. Data for elevations > 1300 m taken from Currey and Oviatt (1985). The upper end of the line depicting the recession was projected from an elevation of 1580 m and a ^{14}C age of 13,500 yr B.P. Between 1300 and 1400 m, the path of the line is constrained by ^{14}C ages of woody material. The wiggles in lake level between 12,200 and 10,000 ^{14}C yr B.P. represent the minor transgression–regression couplets. Elevation of Old River Bed sill that separates Great Salt Lake and Sevier Desert subbasins indicated by a solid line at 1400 m. Elevations have not been corrected for isostatic rebound.

at the margins of a dwindling lake (Currey, 1990). The offshore correlative of the red beds is the base of a thick sequence of mirabalite ($Na_2SO_4 \cdot 10 H_2O$) and mud that underlies the deepest part of Great Salt Lake (Eardley, 1962). The organic muds and sands which form the base of the Gilbert transgression sequence are conformably overlain by several couplets, each consisting of a mud overlain by a sand lens. The sands are interpreted as having been deposited during minor transgressions and the organic muds are considered to represent marsh deposits formed during subsequent minor regressions during the transgressive phase of the Gilbert-age lake (Murchison, 1989; Currey, 1990).

Radiocarbon ages of one wood and six gastropod samples extracted from sediments deposited during the Gilbert transgression together with the radiocarbon age of a peaty mud that overlies the Gilbert transgression (Table 2) have been used to create a model lake-level chronology (13,000–9000 ^{14}C yr B.P.) for the Great Salt Lake subbasin (Fig. 5). The gastropod death assemblages were formed when saline lake water transgressed into their fresh-water habitats. Data from Currey and Oviatt (1985) also were used to constrain the early part of the model chronology. The data indicate that the Gilbert transgression began ~12,200 ^{14}C yr B.P. and terminated ~10,000 ^{14}C yr B.P. The wiggles in lake level schematically depicted in Fig. 5 between 12,200 and 10,000 ^{14}C yr B.P. represent the minor transgression–regression couplets discussed in the preceding paragraph. Since radiocarbon ages of gastropod shells used to determine the recession of Lake Bonneville have been shown to be internally inconsistent (Benson et al., 1990), the overall timing of the gastropod-based

TABLE 2

Number, type, locality, elevation, and radiocarbon ages of samples from the Bonneville basin

Sample number	Sample type	Locality information	Elevation (m)	^{14}C age (yr B.P.)
Great Salt Lake subbasin				
Beta-21807[a]	peaty mud	Juke Box trench, mud overlies Gilbert beach gravel	1295	9450 ± 150
GX-6614[b]	gastropod shells	Magna spit, lagoonal marsh at Gilbert shoreline	1294	10,285 ± 265
GX-6949[b]	gastropod shells	Magna spit, lagoonal marsh at Gilbert shoreline	1294	10,300 ± 310
I-696[c]	wood	Willard Canal, overlain by sand of Gilbert transgression	1290	10,300 ± 275
W-4395[d]	gastropod shells	Shooting Grounds, marsh predating rise to Gilbert shoreline	1289	10,920 ± 150
Beta-22431[e]	gastropod shells	Shooting Grounds, transgressive sand in marsh	1289	10,990 ± 110
Beta-16913[f]	gastropod shells	Shooting Grounds, transgressive sand in marsh	1288	11,570 ± 100
Beta-16912[f]	gastropod shells	Shooting Grounds, transgressive sand in marsh	1288	11,990 ± 100
Sevier Desert subbasin				
Beta-12987[g]	gastropod shells	alluvial mud overlying lacustrine sediment	1387	9570 ± 530
Beta-19455[h]	*Anodonta* shells	lacustrine mud adjacent to shoreline	1385	10,070 ± 130
GX-6776[i]	*Anodonta* shells	lacustrine mud adjacent to shoreline	1385	10,360 ± 225
Beta-17883[h]	*Anodonta* shells	lacustrine mud adjacent to shoreline	1385	11,270 ± 110

[a] ^{13}C set to −25.
[b] previously reported in Currey et al. (1983b).
[c] previously reported in Trautman and Willis (1966).
[d] previously reported in Miller (1980).
[e] previously reported in Murchison (1989).
[f] previously reported in Currey (1990).
[g] previously reported in Simms (1985).
[h] previously reported in Oviatt (1987).
[i] previously reported in Currey (1980).

Gilbert transgression also may be shifted in time; i.e., some of the gastropod ages may be too young.

In the Sevier Desert subbasin (Fig. 4), radiocarbon ages of molluscs (Table 2) indicate that Sevier Lake spilled across the Old River Bed sill to the Great Salt Lake subbasin between ~11,300 and 10,100 ^{14}C yr B.P. Oviatt (1988) indicates that Sevier Lake was full and spilling when Lake Bonneville receded below the Old River Bed sill ~13,000 ^{14}C yr B.P. However, lake-level data for the Sevier Desert subbasin are lacking between 13,000 and 11,300 ^{14}C yr B.P. and it is possible that Sevier Lake receded below sill level (Fig. 5) at the same time the lake in the Great Salt Lake subbasin was undergoing redbed formation and mirabalite deposition.

Surface areas of lakes in the Lahontan and Bonneville basins 11,000 ^{14}C yr B.P.

In this paper, surface area is used as an indicator of change in the hydrologic balance. We follow the procedure suggested by Benson and Paillet (1989), whereby the surface area of a paleolake is normalized by dividing its paleosurface area by its mean historical, reconstructed surface area (PDCSA). In the Lahontan basin, PDCSA (Table 3) was derived using the reconstructed mean

TABLE 3

Surface areas of past and present-day lakes in the Bonneville and Lake Lahontan basins.

Surface areas of the Bonneville and Lahontan basins
Bonneville basin = 139,500 km^2 (Currey and Oviatt, 1985)
Lahontan basin = 114,700 km^2 (Mifflin and Wheat, 1979)
Surface areas of highstand Lakes Bonneville and Lahontan
Lake Bonneville highstand $\geqslant$51,300 km^2 (Currey et al., 1983a)
Lake Lahontan highstand = 22,200 km^2 (Benson and Mifflin, 1986)
Calculated surface areas of present-day lakes in the Bonneville basin (PDCSA)
Great Salt Lake = 6170 km^2 [obtained using reconstructed surface areas of Great Salt Lake from 1851 to 1984 (Stauffer, 1985) and linear piecewise regression of Great Salt Lake area-elevation curve (Fig. 16, Currey, 1990)]
Utah Lake = $\geqslant$400 km^2 (obtained from Currey et al., 1983a)
Calculated surface areas of present-day lakes in the Lahontan basin
[obtained using river discharge and precipitation data listed in Tables 2 and 3 of Benson and Thompson (1987) and an evaporation rate of 1.25 m/yr].

River	Subbasin/s	Surface area
Susan River	Honey Lake	100 km^2
Truckee River	Pyramid Lake, Winnemucca Dry Lake	670 km^2
Carson River	Carson Desert	405 km^2
Walker River	Walker Lake	360 km^2
Quinn River	Smoke Creek/Black Rock Desert	35 km^2
Humboldt River	Carson Desert	905 km^2
	Total surface area =	2475 km^2

Ratio of highstand surface areas to PDCSA
Lake Bonneville = 51,300 km^2/6570 km^2 > 7.8 (minimum due to downcutting that occurred during Bonneville Flood)
Lake Lahontan = 22,200 km^2/2475 km^2 = 9.0
Ratio of PDCSA of lakes to drainage basin areas in the Bonneville and Lahontan basins
Bonneville system = 6570 km^2/139,500 km^2 = 0.047
Lahontan system = 2475 km^2/114,700 km^2 = 0.022

annual streamflow-discharge and precipitation data and an evaporation rate of 1.25 m/yr (Benson and Thompson, 1987) in the following form of the water-balance equation for a closed-basin lake:

$$A_l = D/(E_l - P_l) \quad (1)$$

where A_l = calculated surface area of the closed-basin lake, D = volume of streamflow discharge, E_l = lake-evaporation rate, and P_l = precipitation on the lake surface. The PDCSA is, therefore, a hypothetical area that would exist if the bottoms of closed-basin lakes were truly impermeable.

In the Bonneville basin, the mean annual surface area of Great Salt Lake was calculated using reconstructed annual elevations of Great Salt Lake from 1851 to 1984 (Stauffer, 1985). The mean elevation for each year of record was converted to surface area using the elevation-area data listed in Kay and Diaz (1985, appendix 1). The area of Utah Lake was taken from Currey et al. (1983a) and the Sevier Lake subbasin was assumed to have, on the average, remained dry since 1850.

Because of the topographic complexity of the Lahontan basin and the limited number of subbasins for which lake-level chronologies are available, total lake-surface areas for 11,000 ^{14}C yr B.P can only be approximated (Table 4, Fig. 6). A lake-surface elevation of 1220 m for the western Lahontan subbasins was used in the surface area calculations. River diversion has played a role in regulating lake levels (surface areas) in Lahontan subbasins and any discussion of past change in the hydrologic balance of the Lahontan basin based on lake level data from a subbasin must account for the possibility that river diversion was responsible for part or all of the observed change in lake level in that subbasin (Benson and Paillet, 1989). Calculations that account for the possibility of river diversion at 11,000 ^{14}C yr B.P. (Table 4) indicate that the combined surface area of lakes in the Lahontan basin was >2.9 PDCSA. Given present-day river configurations, calculations indicate that the surface area of the lake in the Carson Desert subbasin was ~2.3 times PDCSA and the combined surface area of lakes in the western Lahontan subbasins was 5.7 times PDCSA at 11,000 ^{14}C yr B.P. (Table 4).

For the Bonneville basin, calculations (Table 5) indicate that ~11,000 ^{14}C yr B.P the total surface area of lakes was ~17,300 km^2 (~2.6 times PDCSA). The similarity of surface area increases of lakes in the Carson Desert subbasin and the Bonneville basin (2.3 and 2.6 times PDCSA) and the magnitude of these increases relative to surface-area increases of lakes in western Lahontan subbasins (5.7 times PDCSA) indicate that change in the effective wetness was not homogeneous over the northern Great Basin area 11,000 ^{14}C yr B.P.

TABLE 4

Calculated present-day and 11,000 yr B.P. elevations (m) and surface areas (km^2) of lakes in the Lake Lahontan basin. [PL = Pyramid Lake subbasin; WDL = Winnemucca Dry Lake subbasin; SC = Smoke Creek–Blackrock Desert subbasin; HL = Honey Lake subbasin; CD = Carson Desert subbasin; WL = Walker Lake subbasin; TSA = total surface area of lakes in the six subbasins; ele = elevation; WR = Walker River; HR = Humboldt River; ->, river diversion]

	PL	WDL	SC	HL	CD	WL	TSA
Present-day simulation							
ele (m)	1177	1153	1171	1217	1183	1259	
area (km^2)	570	100	35	100	1310	360	2475
11,000 yr B.P. simulation							
ele (m)	1220	1220	1220	> 1217	1203	> 1259	
area (km^2)	715[a]	360[a]	2950[a]	> 100[b]	2990[c]	> 360	> 7475
11,000 yr B.P. simulation, WR->CD							
ele (m)	1220	1220	1220	> 1217	1203	0	
area (km^2)	715[b]	360[b]	2950[b]	> 100	2990[d]	0	> 7115
11,000 yr B.P. simulation, HR->SC							
ele (m)	1220	1220	1220	> 1217	1203	> 1259	
area (km^2)	715[e]	360[e]	2950[e]	> 100	2990[f]	> 360	> 7475
11,000 yr B.P. simulation, HR->SC, WR->CD							
ele (m)	1220	1220	1220	> 1217	1203	0	
area (km^2)	715[e]	360[e]	2950[e]	> 100	2990[g]	0	> 7115

[a]The total surface area of lakes in the PL, WDL, and SC was 5.7 times their PDCSA.
[b]If change in the surface area of the lake in the HL corresponded to change in the surface areas of lakes in PL, WDL, and SC, Honey Lake would have spilled contributing 35 km^2 to SC.
[c]The surface area of the lake in the CD was 2.3 times its PDCSA; however, if change in the surface area of the lake in the WL corresponded to change in the surface areas of lakes in PL, WDL, and SC, and Walker Lake spilled to CD, the surface area of the lake in the CD was 1.3 times the PDCSA of the lake in CD associated with discharge of the Carson and Humboldt Rivers.
[d]If the Walker River diverted to the CD, the surface area of the lake in CD was 1.8 times the PDCSA of lakes in CD and WL.
[e]If the Humboldt River diverted to the SC, the total surface area of lakes in PL, WDL, and SC was 2.5 times their PDCSA associated with discharge of the Quinn and Truckee Rivers.
[f]If the Humboldt River diverted to the SC, the surface area of the lake in CD was 7.4 times the PDCSA associated with discharge of the Carson River to the CD. However, if change in the surface area of the lake in the WL corresponded to change in the total surface area of lakes in PL, WDL, and SC, Walker Lake would have overflowed to CD and the surface area of the lake in the CD was 7.1 times the PDCSA associated with discharge of the Carson River to the CD.
[g]If the Walker River diverted to CD, and the Humboldt River diverted to the SC, the surface area of the lake in CD was 3.9 times the PDCSA associated discharge of the Carson River to the CD.

TABLE 5

Surface areas of lakes in the Bonneville basin at ~11,000 yr B.P.

Great Salt Lake subbasin =	16,300 km^2 (1295-m ele)
Sevier Lake subbasin =	~600 km^2
Utah Lake =	~400 km^2
Total surface area (TSA) =	~17,300 km^2
TSA/PDCSA =	~2.6

Inhomogeneity of effective wetness may have been due to a number of factors including difference in the latitudinal distribution of storm systems (headwaters of the western Lahontan subbasins are located between 38 and 40°N latitude and the headwaters of both the Bonneville and Humboldt River basins are located between 40 and 42°N latitude) and orographic blocking of westerly storms by the Sierra Nevada. By 9500 ^{14}C yr B.P.,

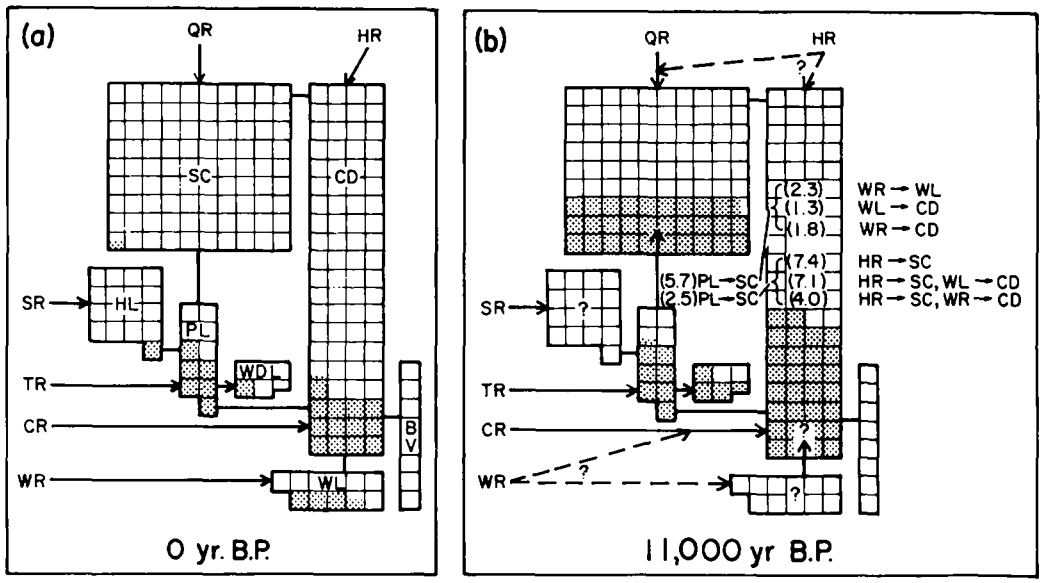

Fig. 6. Schematic representation of surface areas (stippled pattern) of lakes in Lahontan subbasins at 0 and ~11,000 ^{14}C yr B.P. (see Table 4 for subbasin abbreviations; *QR*= Quinn River, *BV*= Buena Vista subbasin, *SR*= Susan River). Solid lines connecting subbasins indicate sills and arrows indicate intersubbasin spill or streamflow discharge. Numbers in parentheses indicate the surface area/PDCSA ratio for the time of interest. Question marks centered in subbasins indicate that lake-size information is lacking, and question marks/dashed lines from catchment areas indicate the possibility of river diversion. Each square indicates an area of 100 km^2. See footnotes in Table 4 for discussion of surface-area calculations.

lakes in both Lahontan and Bonneville basins had receded, achieving surface areas similar to their PDCSAs (Figs. 3 and 5).

Mountain glaciation in the Great Basin between 13,000 and 9000 ^{14}C yr B.P.

Dorn et al. (1990) have found evidence for a young glaciation in the Chiatovich Cirque area of the White Mountains of Nevada–California (Elliot-Fisk, 1987). Basal layers of rock varnish coating boulders at this locality have a radiocarbon age of 9740±210 ^{14}C yr B.P., an age consistent with the measured in situ accumulation of ^{14}C (Jull et al., 1989) and ^{36}Cl (Zreda et al. 1991). The estimated time of till deposition (between 11,100 and 9700 ^{14}C yr B.P.), calculated assuming that the measured age is within 13% of the time of till deposition, is consistent with the data of Mezger and Burbank (1986) who obtained radiocarbon ages of 10,640±160, 10,320±110, and 10,060±130 ^{14}C yr B.P. on basal material from a bog formed behind a moraine in the southeastern Sierra Nevada.

Summary and discussion

Recessions from highstand levels in the Lahontan and Bonneville basins were reversed when lakes began to rise ⩾11,500 ^{14}C yr B.P. in the Lahontan basin and ~12,200 ^{14}C yr B.P. in the Bonneville basin. The perturbations in lake level lasted for ~2000 yr, and between 10,000 and 9500 ^{14}C yr B.P., lakes in both basins abruptly receded. The total surface areas of lakes in the Lahontan and Bonneville basins were >2.9 and 2.6 PDCSA ~11,000 ^{14}C yr B.P. Given present-day river configurations, surface areas of lakes in western Lahontan subbasins increased about twice as much as lakes in the Bonneville and Carson Desert basins. The inhomogeneous distribution of change in effective wetness may have been due to difference in the latitudinal distribution of storm systems or to orographic blocking of westerly storm systems by the Sierra Nevada. During the last phase of lake transgressions in the Lahontan and Bonneville basins, minor glacial advances occurred in the Sierra Nevada and White Mountains.

Within the accuracy of the chronologies, the

12,000–10,000 ^{14}C yr B.P. perturbations in lake level that occurred in the Lahontan and Bonneville basins were concurrent with the European Allerød and Younger Dryas intervals. Advances of mountain glaciers in the Sierra Nevada and White Mountains appear to be concurrent with the Younger Dryas interval. However, we do not believe that the reliability and accuracy of the ^{14}C age dates on which the lacustrine chronologies are based are sufficient to assert that the lakes rose during the Younger Dryas interval and fell during the Allerød interval or that their rise and fall was confined to only one interval. Given the inherent inaccuracies of the chronologies, we could just as well assert that the lake-level perturbation began in the middle of the Allerød interval and ended after the termination of the Younger Dryas interval.

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